INTRODUCTION TO HISTORICAL GEOLOGY



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TABLE OF CONTENTS

Licensing

1: What is Historical Geology?

2: An Introduction to Earth

2.1: The Origin of Earth and Our Solar System

2.2: Earth's Structure and the Origin of the Continents and Oceans

2.3: Key Points

3: Plate Tectonics

3.1: Plates Defined

3.2: Tectonics Defined

3.3: Plate Boundaries and Interiors

3.4: Geophysical Phenomena

3.5: Driving Forces

3.6: The Beginning of Plate Tectonics and the Continents

3.7: The Story of Plate Tectonics as Recorded by the Rocks

3.8: Key Points

4: Earth Materials - Minerals

4.1: What is a Mineral?

4.2: Building Blocks of Rock

4.3: Chemistry of Minerals

4.4: Formation of Minerals

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

4.6: Other Important Rock Forming Minerals

4.7: Key Points

5: Earth Materials - Rocks

5.1: Igneous Rocks

5.2: Metamorphic Rocks

5.3: Sediments and Sedimentary Rocks

5.4: The Rock Cycle

5.5: Key Points

6: Sedimentology and Stratigraphy

6.1: Sediment Deposition and Bedding

6.2: Other Sedimentary Structures

6.3: Fossils and Fossilization

6.4: Facies

6.5: Depositional Environments

6.6: Introduction to Strata and Basins

6.7: Sequence Stratigraphy and Walther's Law
6.8: Stratigraphic Correlation
6.9: Key Points
6.10: Further Reading

7: Geologic Structures

7.1: Tectonic Movement
7.2: Geological Maps and Cross-Sections
7.3: Stress, Strain, and Deformation
7.4: Folds
7.5: Faults
7.6: Key Points

8: Geologic Time

8.1: Earth's Age and Abbreviating Time Intervals
8.2: Uniformitarianism and Relative Age Dating
8.3: Unconformities
8.4: Fossils and Age Dating
8.5: Absolute Age Dating
8.6: The Geological Time Scale
8.7: Key Points

9: Evolution

9.1: What is Evolution?
9.2: Development of the Theory of Evolution
9.3: A Modern View of Evolution
9.4: Evolutionary Patterns in Fossils
9.5: Key Points
9.6: Further Reading

10: Hadean and Archean

10.1: Hadean Eon
10.2: Earth's Oldest Rocks
10.3: Archean Eon
10.4: When Did Life Begin?
10.5: Key Points
10.6: Further Reading

11: Proterozoic Eon

11.1: Proterozoic Geography
11.2: Proterozoic Glaciations
11.3: Proterozoic Atmosphere
11.4: Proterozoic Life
11.5: Key Points

11.6: Further Reading

12: Phanerozoic Eon - Paleozoic Era

12.1: Geography and Tectonics of the Early Paleozoic

12.2: Evidence for the Taconian Orogeny

12.3: Geography and Tectonics of the Middle Paleozoic

12.4: Geography and Tectonics of the Late Paleozoic

12.5: Early Paleozoic Life - The Cambrian and Ordovician

12.6: Middle Paleozoic Life - The Silurian and Devonian

12.7: Late Paleozoic Life - The Carboniferous and Permian

12.8: Key Points

13: Phanerozoic Eon - Mesozoic Era

13.1: Paleography and Tectonics of the Mesozoic

13.2: Mesozoic Evolution - Reptiles and Amphibians

13.3: Mesozoic Evolution - Amniotes

13.4: Mesozoic Evolution - Dinosaurs

13.5: Mesozoic Mass Extinctions

13.6: Key Points

14: Phanerozoic Eon - Cenozoic Era

14.1: Paleogeography and Tectonics of the Cenozoic

14.2: An In-Depth Look at the San Andreas Fault and the Basin and Range

14.3: Cenozoic Biology

14.4: Key Points

15: Glossary

Index

Glossary

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1: What is Historical Geology?

Introduction

Historical geology seeks to tell the story of the earth. Recall some of your favorite stories. They may have been told through a movie, a show, a book, or by a person.

The planet you live on has some stories to tell. Earth has had 4.55 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 to read it, we first need to learn the language in which it is recorded. We must learn to "speak Rock."

Stories from Stone

Earth has written countless stories in the language of Rock. This cobble from Montana provides a great example.



Figure 1.1: What is the history of a rock like this? What did it take for this thing to exist? (Callan Bentley photo.)

The texture of this rock contains features 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 initially formed as inorganic carbonate mud, laid down (1) in calm water, as evidenced by its very fine grain size. It sat still long enough to stick together into a semi-coherent mass. But then it must have been torn up to make (2) mud-chip pieces entombed in (3) quartz sand. The event was energetic enough to rip up the carbonate mud and transport coarse sand particles. Perhaps a hurricane provided the energy. Eventually, the sediment was buried and sank further underground, smothered beneath the crushing weight of thousands of feet of sediment piled atop it and was (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 for us to pick it up and hold it, we know it must have been (6) uplifted and brought up to Earth's surface; 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 1.2: 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 and 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.

Decoding the handful of past processes encoded into a single cobble is powerful. However, more power still comes from looking at sedimentary rock in outcrop (see figure below), and comparing the conditions of one layer with those below and above it. The common language of historical geology is sedimentary rock, and vast epic tales have been inscribed in the sequence the rock layers. Here is an example, from Inuyama, Japan:

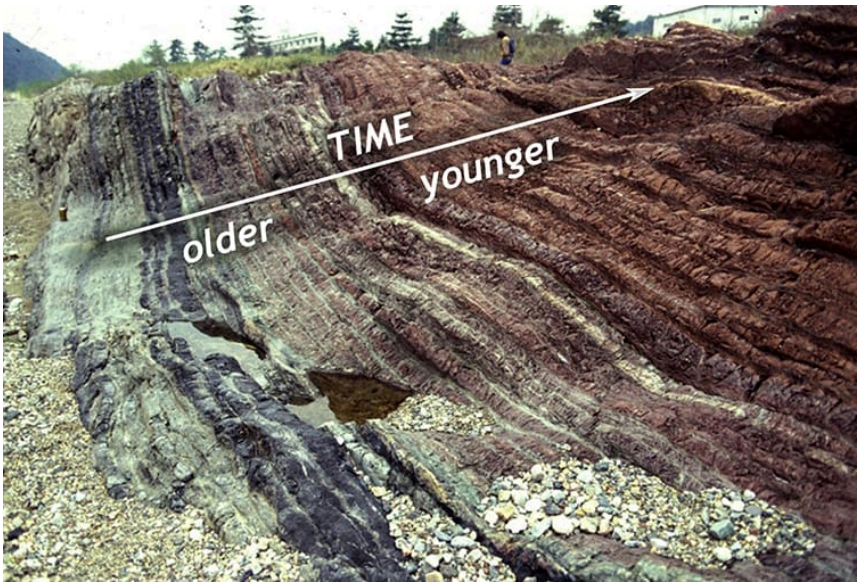


Figure 1.3: Outcrops in Inuyama, Japan, show a Triassic sequence of sedimentary strata that record changes in deep ocean oxygen levels after the end-Permian mass extinction event. The strata have been tectonically rotated due to accretion with the Japanese mainland. (Photo by Yukio Isozaki; reproduced with permission.)

Here, the strata are not in their original horizontal orientation. They have been rotated during tectonic accretion (adding) of these deep ocean strata to the Japanese mainland. The layers get younger to the right.

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 none (black), then plenty (red). This is interesting because a major mass extinction occurs just before this sequence at the end of a period of time called the Permian: the end-Permian mass extinction (also called “The Great Dying”). These strata are 247 to 242 Ma, early middle Triassic (i.e. the period after the Permian), showing recovery after the mass extinction. Evidence such as this sequence of strata suggests that ocean anoxia (low oxygen levels) was associated with this major die-off in animals, either as a cause or a consequence. And most everything on the planet died: 95% of species. It was the worst thing to ever happen to life on our planet, and this sequence of black layers transitioning into red layers is trying to tell us why it happened. As we see here, no single layer or rock tells the whole story – we must look at the succession of layers over time.

Why should you care about any of this?

You are an Earthling. You were born on this planet and will live 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, neurons, and fingertips was derived from food you ate, and the plants that captured that carbon pulled it from the atmosphere. It got there perhaps from volcanic emissions, 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.



Figure 1.4: We instinctively connect with the Earth from a very young age. (Callan Bentley photo.)

Understanding Historical Geology deepens our appreciation of the planet we interact with throughout our lives. It provides a framework for the existence of all living things, an explanation for how life has evolved over time, and the ability to make predictions for our future Earth.

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2: AN INTRODUCTION TO EARTH

Introduction

The changes that have occurred since the inception of Earth are vast and significant. From the introduction of oxygen into 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.

How and when does the story of Earth begin? A logical place to start is with the formation of our planet. However, the formation of our planet is part of a larger story, the story of the formation of our solar system.

This chapter summarizes the origin of our solar system and introduces the earliest chapters of the Earth's story.

2.1: The Origin of Earth and Our Solar System

2.2: Earth's Structure and the Origin of the Continents and Oceans

2.3: Key Points

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2.1: The Origin of Earth and Our Solar System

The prevailing scientific explanation for the origin of the Earth can be applied to the formation of the Sun and all the other planets. So, it's not only Earth's story but the story of the origin of the whole solar system. Not only that, but our Sun is 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 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 present is the key to the past.

Using powerful telescopes, we observe many other stars, but we also observe other things, including fuzzy-looking "stars" called nebulae. A **nebula** is a massive cloud of gas and dust in space that is the byproduct of a previous star that suffered a supernova explosion - its leftover material. Orion's "belt" is part of a constellation of stars easily identifiable in the northern hemisphere's winter night sky. Below the row of three stars of the belt is a smaller trio of light spots; this is Orion's sword scabbard. Binoculars and a dark enough sky 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.

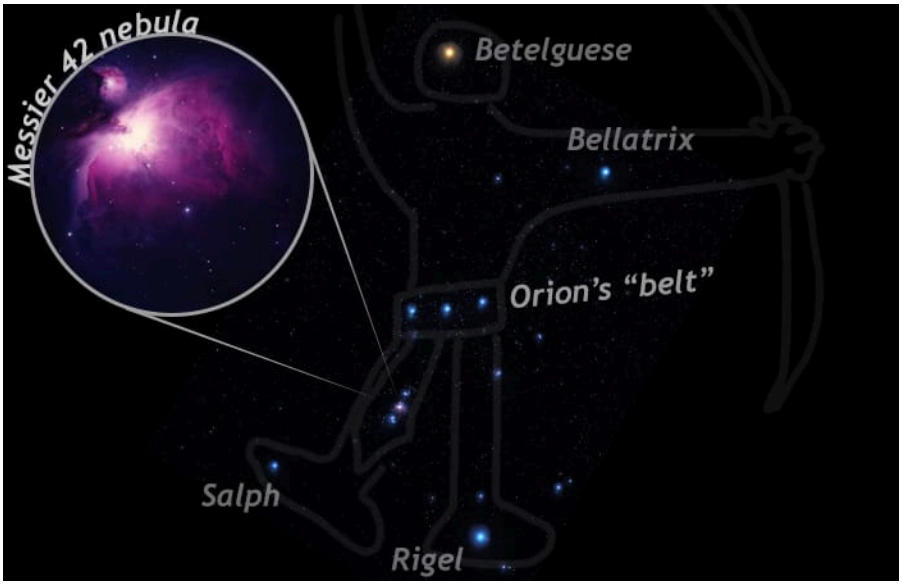


Figure 2.1.1: The Messier 42 nebula, shown in the context of the "scabbard" of the constellation Orion.

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. Particles attract one another under the influence of forces such as electrostatic attraction or "static cling". This is the same force that turns bits of dust or fabric into larger "dust bunnies" under your couch!



Figure 2.1.2: Three dust bunnies under the couch: a household hassle, or a glimpse into the birth of the solar system?

Electrostatic forces are effective at attracting small particles over short distances. However, if you want to make big things like planets and stars out of a nebula, gravity needs to take over at some point. Although gravity is a relatively weak force - you overcome the force of gravity every time you pick up your foot to take a step - gravity can work very efficiently over distance if the masses involved are large enough. So static cling initially organized matter into "space dust bunnies" that once large enough, were pulled together by gravity. The net result is that the countless tiny pieces of the nebula were drawn together, swirling into a denser and denser amalgamation (clump). The nebula began to spin, flattening from top to bottom and forming a rotating disk, like a fried egg with the yolk surrounded by the white.

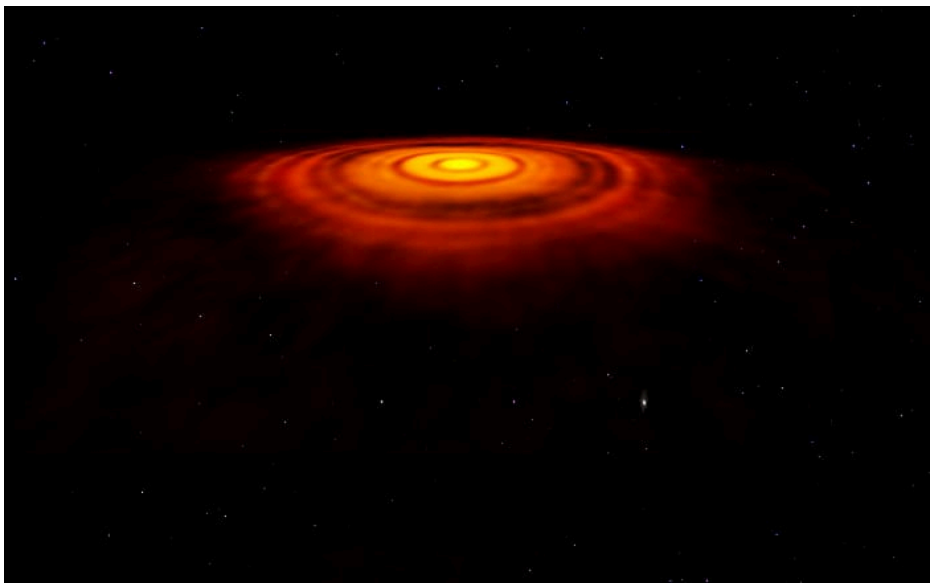


Figure 2.1.3: An artist's conception of an oblique view of a protoplanetary disk, using actual photoimagery of the HL Tauri disk, originally gathered by the European Southern Observatory.

Once a star forms in the center, astronomers call the ring of debris around it a **protoplanetary disk**. For our solar system, this happened 4.55 billion years ago.

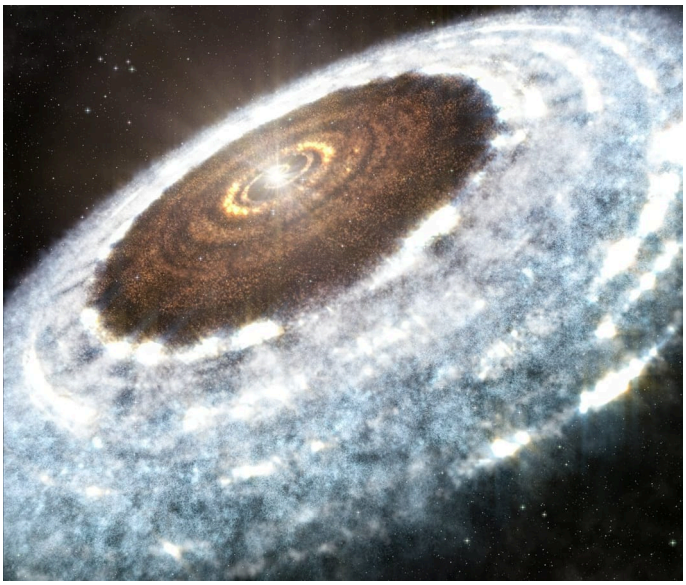


Figure 2.1.4: This artist's impression of the water snowline around the young star V883 Orionis, as detected with ALMA.

Two important processes that helped further organize the protoplanetary disk and form planets were condensation and accretion.

Condensation is a process that allows gaseous matter to stick 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 today, entombed inside certain meteorites.)

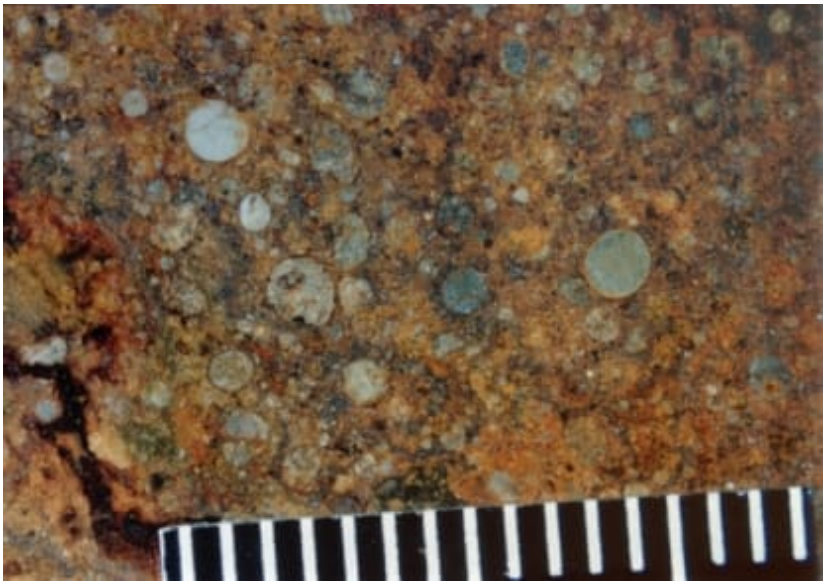


Figure 2.1.5: Chondrules in a meteorite, with a scale in mm.

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 have sufficient gravity to pull their mass into a round shape, forming **planetesimals**, the nuclei of planets. Planetesimals gobbled up nearby asteroids and smashed into other planetesimals, merging and growing through time through **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, **differentiation** could happen, where denser material sank to their centers (or cores), and lighter-weight elements and compounds rose to their surface (or outer layers).

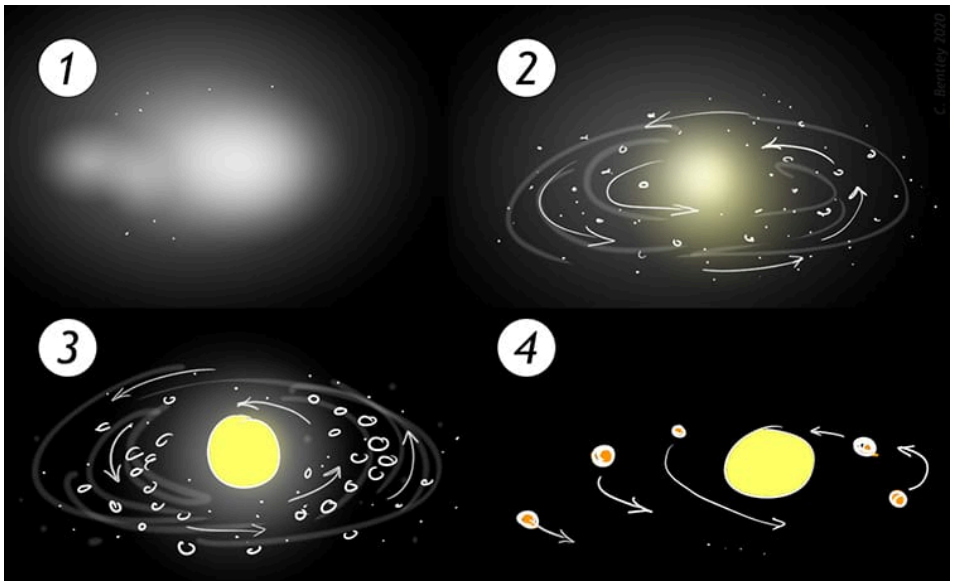


Figure 2.1.6: A cartoon model showing the evolution of our solar system from a pre-solar nebula.

Metallic iron meteorites, such as the Tamentit meteorite below, are used as evidence of the differentiation of planetesimals into layered bodies. Meteorites with metallic compositions represent the “core” material from these planetesimals.



Figure 2.1.7: Tamentit Iron Meteorite, found in 1864 in the Sahara, weighing about 500 kg (1,100 lb). From Ji-Elle, Public domain, via Wikimedia Commons.

The process of accretion continues into the present day, though at a slower pace than the earliest days of the solar system. You can observe this in the asteroid belt, where certain asteroids are simply a clump of space rocks, held together by their gravity. Consider the asteroid called Itokawa 25143, for instance which is made up of numerous uniquely-sized boulders rather than of a single solid body.

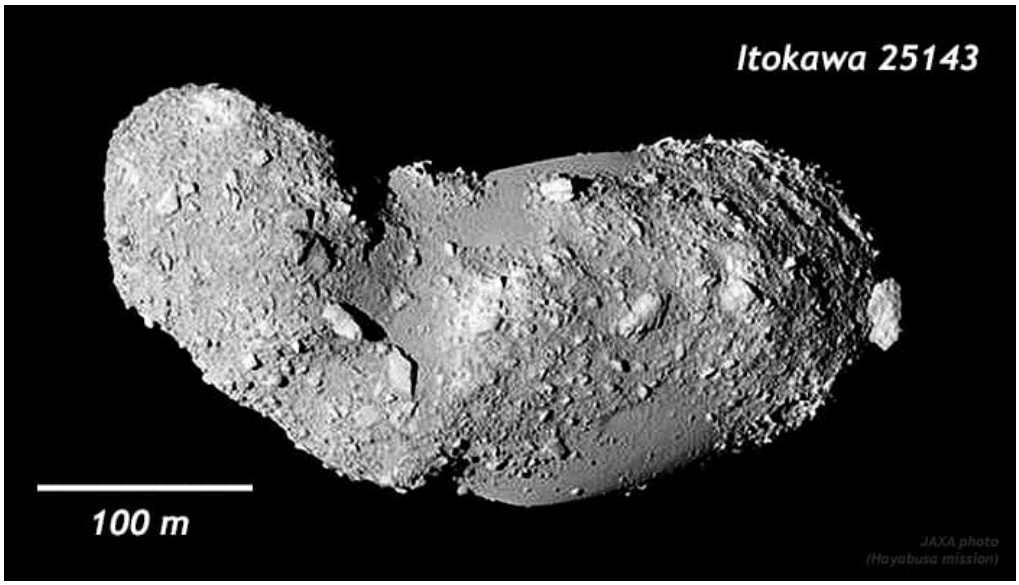


Figure 2.1.8: The asteroid 25143 Itokawa.

Another modern example of accretion 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 when Jupiter's immense gravity broke the comet into chunks, 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 ^{o}C fireball and enormous impact scars that were as large as the entire Earth. These scars lasted for months.

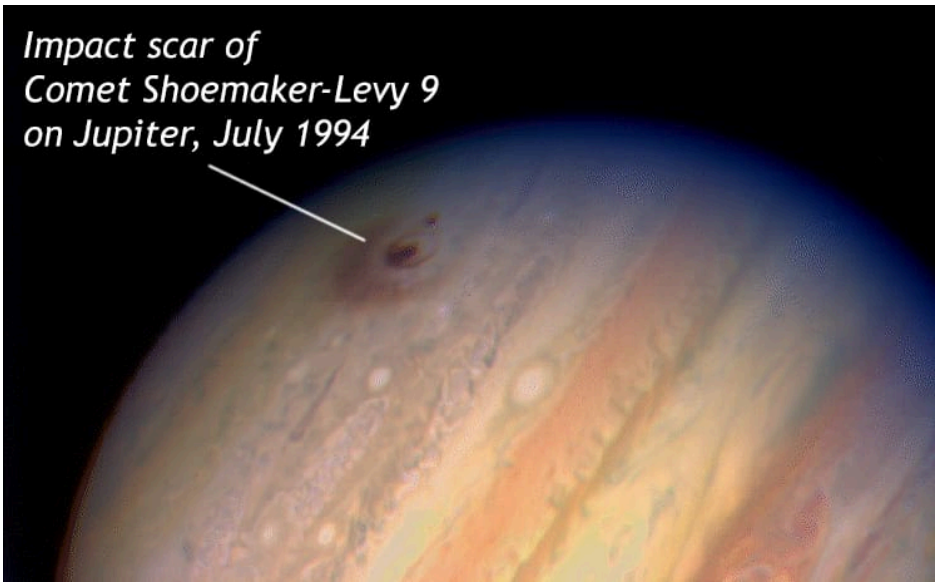


Figure 2.1.9: Accretion in action: scar from the July 1994 collision between a fragment of Comet Shoemaker-Levy 9 and the surface of the atmosphere of the planet Jupiter.

This incredibly dramatic event reminded Earthlings that they are not safe from accretionary impacts, just as the dinosaurs discovered some 66 million years ago. In this case, an object from the solar system was removed while adding a bit to Jupiter's mass.

In summary, Earth is part of a solar system centered on the Sun. This solar system, with its star, planets, dwarf planets, and "leftover" comets and asteroids, formed from a nebula full of elements in the form of gas and dust. Over time, these very small pieces stuck together to make larger concentrations of mass, eventually culminating in a star and its orbiting planets. Asteroids (and asteroids that fall to Earth, called meteorites) are leftovers from this process. The original nebula formed from the destruction

of some previous star that exploded in a supernova - a galactic-sized example of recycling!

Key Terms

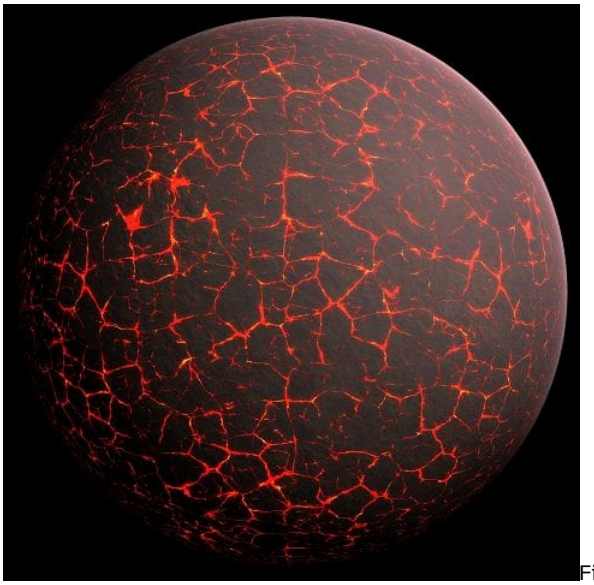
- **accretion** - increasing the mass of an object by adding material to it
- **chondrules** - small grains found in certain types of meteorites formed in the early solar nebula as molten droplets quickly cooled and crystallized
- **differentiation** - material settling out according to density
- **nebula** - a massive cloud of gas and dust in space that is the byproduct of a previous star that suffered a supernova explosion
- **planetesimals** - the nuclei of planets formed in the early solar system from collisions with other objects in the solar system
- **protoplanetary disk** - ring of debris around a star

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2.2: Earth's Structure and the Origin of the Continents and Oceans

Forming Earth's Structure

Around 4.55 billion years ago, the brand new Earth was extraordinarily hot and mostly molten, resembling a ball of lava floating in space.


Figure 2.2.1: Earliest Earth.

As Earth cooled from its molten state, minerals crystallized and settled according to their densities - differentiation transformed a homogenous sphere of magma into a heterogeneous planet with a distinct structure. Metals and radioactive elements have the greatest density and sank to the center to make the core, while silicate compounds floated to the surface. Iron and magnesium-rich silica compounds of moderate density took up the space between. This resulted in three primary layers: the core, mantle, and crust. The terms **lithosphere** and **asthenosphere** are also used to describe the structure of Earth's outer layers. The lithosphere includes the solid uppermost mantle and crust, while the asthenosphere describes the semi-solid upper mantle below. These terms are particularly important in the context of **plate tectonics** which is discussed in detail elsewhere in this book. For now, consider it the theory that Earth’s rigid outer shell, the lithosphere, is broken up into individual pieces called plates, that move and interact with one another along plate boundaries.

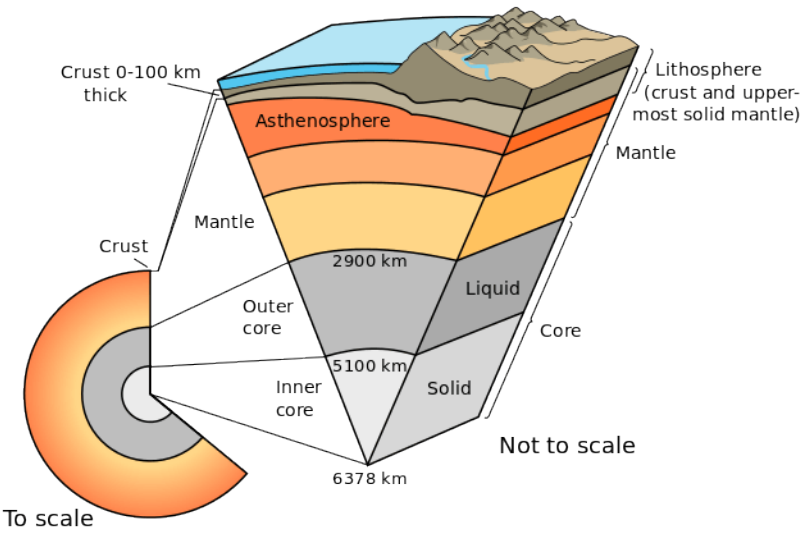


Figure 2.2.2: The structure of Earth. The crust, mantle, and core are layers with distinct compositions. Lithosphere and asthenosphere have different physical properties.

Origin of the Continents

The first continental crust formed shortly after the earth's origin, but continents, as we know them today, and the plate tectonic processes that build and shape them did not happen until much later. Initially, crustal pieces were continually split apart by heat rising from Earth's very hot core. Earth's surface may have resembled the lava lakes we can see today within the crater of a few volcanoes.

Lava Lake Video

Watch the video from 1:31 to 2:03.

YouTube Video

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Origin of Earth's Water

The origin of surface water on Earth is still uncertain. Explanations for its origin 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.

Where Did Our Water Come From? Video

Watch the video from 4:47 to 10:03.

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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 probably that all three sources contributed to the origin of Earth's water.

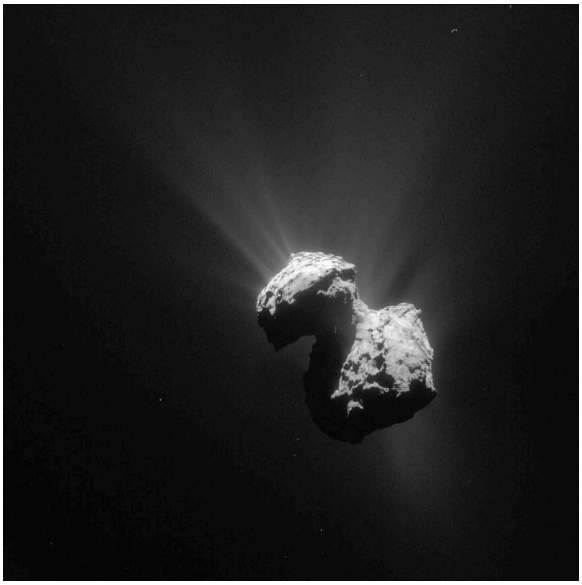


Figure 2.2.3: Water vapor leaves comet 67P/Churyumov–Gerasimenko.

Key Terms

- **asthenosphere** - the layer beneath the lithosphere, which is in a plastic, semi-solid state
- **lithosphere** - the outermost layer of Earth that includes the crust and solid upper mantle; this solid, rock-layer, is broken up into individual pieces called plates
- **plate tectonics** - the theory that Earth's rigid outer shell, the lithosphere, is broken up into individual pieces called plates, that move and interact with one another along plate boundaries, driving Earth's constructive geologic processes

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2.3: Key Points

- Earth and our solar system formed 4.55 billion years ago from a nebulae
- The Earth's layered structure is the result of differentiation, the separation of materials according to density
- Earth's crust formed shortly after the origin of the planet but continents as we know them didn't happen until much later
- Oceans developed from the outgassing of volcanoes and meteor and comet impacts

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3: PLATE TECTONICS

Introduction

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 (continents move around on top of the ocean floor) with new data showing seafloor spreading (seafloor is created at "cracks" in the ocean floor as the two sides separate). The new theory was a revolutionary idea that made the distribution of volcanoes, earthquakes, continents, and oceans 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.

3.1: Plates Defined

3.2: Tectonics Defined

3.3: Plate Boundaries and Interiors

3.4: Geophysical Phenomena

3.5: Driving Forces

3.6: The Beginning of Plate Tectonics and the Continents

3.7: The Story of Plate Tectonics as Recorded by the Rocks

3.8: Key Points

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3.1: Plates Defined

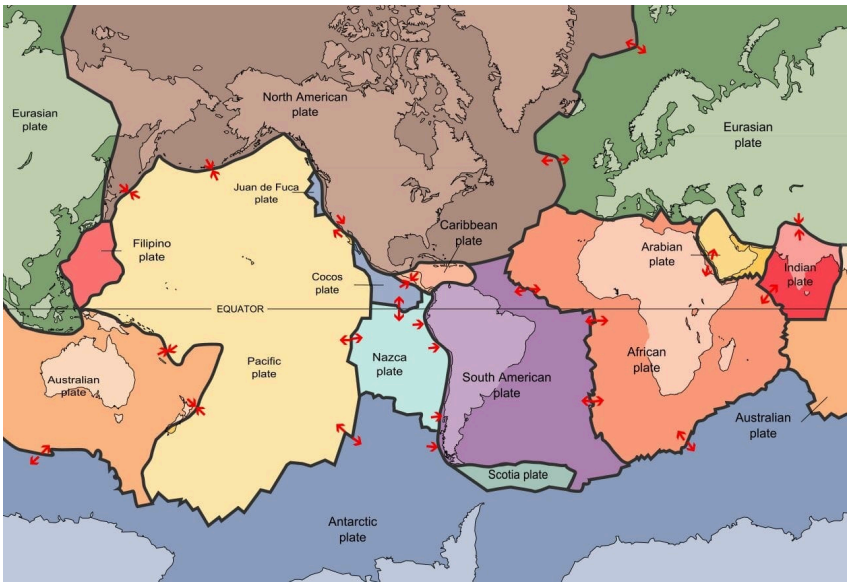


Figure 3.1.1: Plate boundary map. Arrows at the plate boundaries indicate the direction of movement along that particular boundary. (Public domain; Wikipedia.)

Earth's outermost rocky surface is broken into about 15 pieces (called **plates**) that move about relative to each other. The plates are 50-280 km thick and are comprised of the **lithosphere** which in turn is made up of the solid upper mantle and crust. The crust that sits on top of the solid upper mantle comes in one of two varieties: **continental crust** or **oceanic crust**.

Oceanic lithosphere, composed of oceanic crust and the underlying solid upper mantle, is different than continental lithosphere which has a cap of continental crust on the mantle instead. Oceanic lithosphere is a lot thinner (around 10 km thick) and more dense. The continental crust tends to be about 30 km thick, but can get close to 100 km thick under new mountain belts, and is thinner beneath **rift zones** (weak spots where the crust is stretching). The mantle portion of the lithosphere is the same composition beneath the continental 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. The mantle below the lithosphere, called the **asthenosphere**, is a weak layer in that is mostly solid but perhaps 0.5% molten. This tiny amount of melt dramatically lowers the strength of the asthenosphere. Movement in the asthenosphere contributes to the motion of the lithospheric plates that sit on top of it.

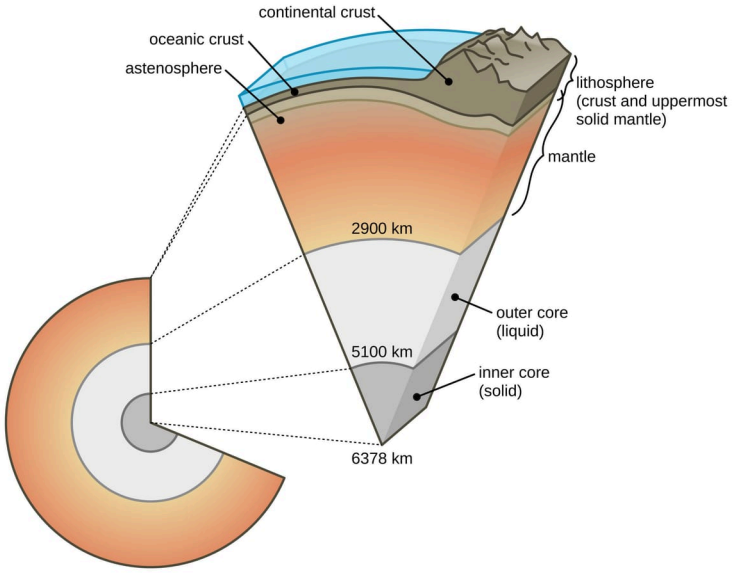


Figure 3.1.2: The layered interior of the Geosphere. Note that Lithosphere includes both the crust and uppermost portion of the solid mantle. This layer comprises Earth's tectonic plates. (U.S. Geological Survey.)

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.

The ordered layering of crust, mantle, and core is a reflection of a mature planet, that has differentiated into distinct horizons of varying density which increases toward Earth's center:

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 Plate and Nazca Plate, too. The smallest plate consisting primarily of continental lithosphere is the Arabian Plate, at 5,000,000 km^2.

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.

Key Terms

- **asthenosphere** - the layer beneath the lithosphere, which is in a plastic, semi-solid state
- **continental crust** - thicker, less dense crust that when paired with the solid upper mantle creates continental landmasses
- **lithosphere** - the outermost layer of Earth that includes the crust and solid upper mantle; this solid, rock-layer, is broken up into individual pieces called plates
- **oceanic crust** - thinner, more dense crust that when paired with solid upper mantle creates the ocean floors
- **plates** - pieces of lithosphere composed of oceanic and/or continental crust and the solid upper mantle
- **rift zones** - weak spots in Earth's crust where stretching is occurring creating low spots

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3.2: Tectonics Defined

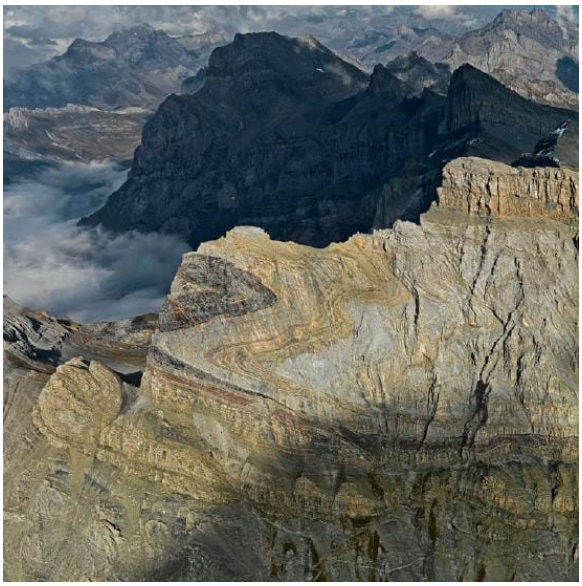


Figure 3.2.1: The huge fold in the flank of the 2969 m high Dent de Morcles is the most impressive witness to the tectonic collision between Africa and Europe. The darker layers are younger than the light greyish and yellowish rock. As the Alps were shoved upward, these layers were inverted as the rock folded. (Photograph by Bernhard Edmaier; reproduced with permission.)

Much of the attention we devote to 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.

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 volcanically active chains of islands, ocean formation, earthquake locations, and more. There are three types of plate boundaries based on the relative motion between the neighboring plates.

- **convergent boundary** - two plates move toward one another
- **divergent boundary** - two plates move away from one another
- **transform boundary** - two plates slide past one another

The dynamics of their relative motion end up determining which suite of geological processes and phenomena results.

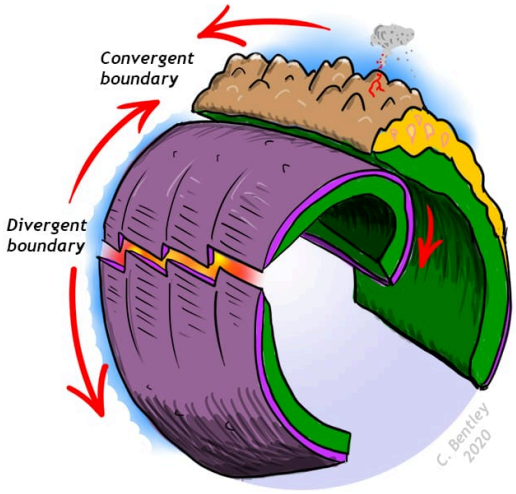


Figure 3.2.2: Three plates interact along two boundaries. The middle plate, of thin oceanic lithosphere, moves away from the left plate, and moves toward the thick continental plate to the right. (Drawing by Callan Bentley.)

Key Terms

- **convergent boundary** - plate boundary type at which the two plates move toward one another
- **divergent boundary** - plate boundary type at which the two plates move away from one another
- **transform boundary** - plate boundary type at which the two plates slide past one another

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3.3: Plate Boundaries and Interiors

Transform Boundaries

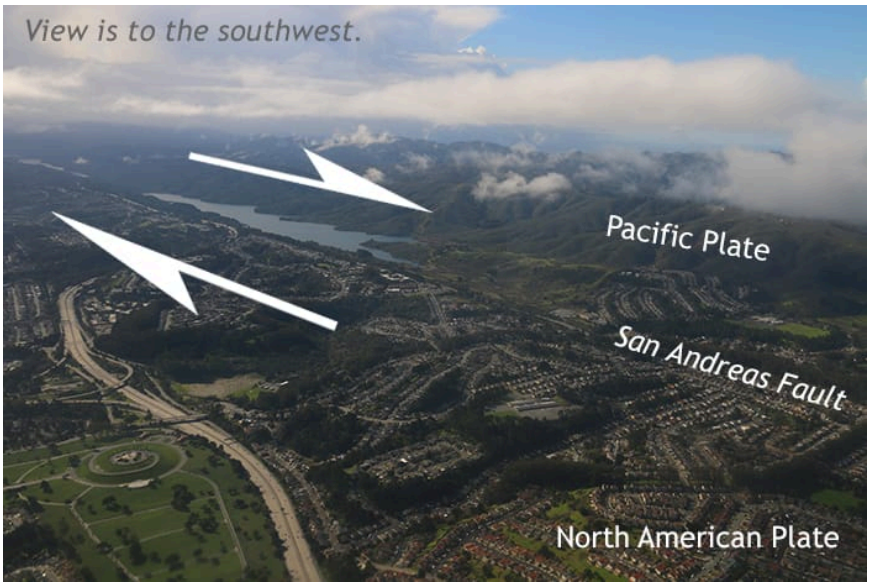


Figure 3.3.1: The modern plate boundary in the San Francisco region formed as the plates slide past one another, seen here south of San Francisco along the trace of the San Andreas Fault. (Callan Bentley photo.)

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. 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. This difference in magnitude occurs as rock is weaker under shear stress (stress generated when materials try to slide past one another) than under compressional stress (stress generated when materials are pushed together), so less stress builds up at transform boundaries.

Transform boundaries can be recognized by offset features that cross the plate boundary such as bodies of rock, landscape features such as stream valleys, or human-built structures such as roads and fences.

Watch this 50-second video as a geologist shows where the San Andreas Fault offset a stream. You can see the offset where the stream cuts to the right in the video.

YouTube Video

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Figure 3.3.2: 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.)

As a result of the stress and subsequent breakage of rock along the boundary, fault zones are created. 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.

In places where the trace of the faults on a transform boundary bend, the fault is no longer parallel to the plates' motion. This results in localized areas of compression (pushing forces) or tension (pulling forces). When compression results, rock is forced upward into a pressure ridge. Pressure ridges can be small enough to see in a parking lot pavement, or the size of mountains. When tension results, land drops down. If it is a small feature and it fills with water it is called a sag pond. If it is a bigger feature and fills with water, it can make a larger trough like the Salton Sea.

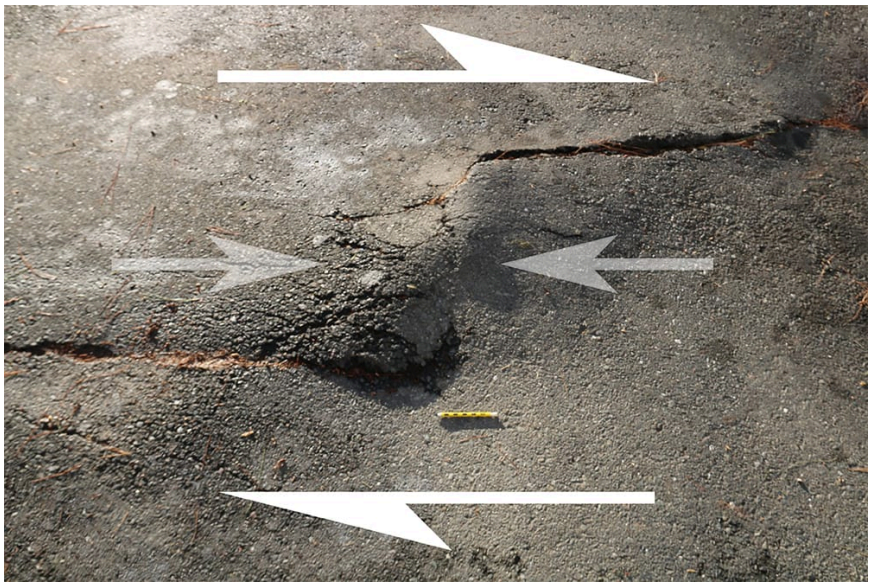


Figure 3.3.3: A small pressure ridge developed in parking lot asphalt at Central Park, Fremont, California. The plate motions are indicated by the large arrows. The bend in the fault and subsequent compression is indicated with the smaller arrows. Perspective is looking to the southwest across the fault. (Callan Bentley photo.)



Figure 3.3.4: A sag pond developed along the North Anatolian Fault in Turkey. (Callan Bentley photo.)

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.

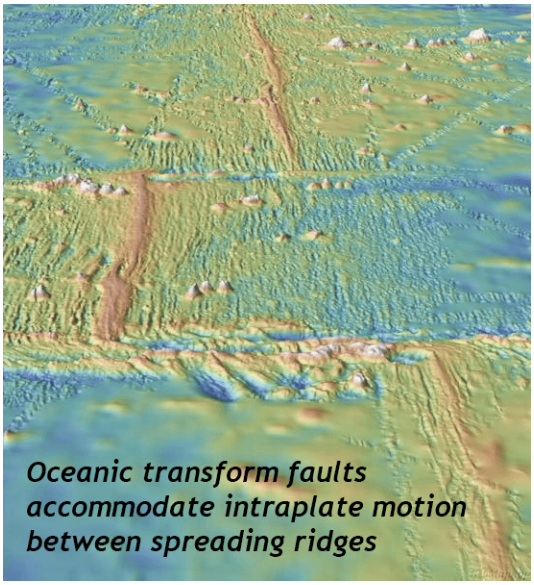


Figure 3.3.5: The oceanic ridge system comes in many segments, offset along perpendicular fracture sets. The big arrows show the divergence at the plate boundary. The smaller red arrows show where slipping is occurring along transform boundaries.

Divergent Boundaries

Divergent boundaries are sites where two plates move away from one another. In this extensional setting the crust thins while it stretches out laterally. Divergent boundaries come in two principal varieties that are relevant to geology:

- rift valleys in continental crust
- oceanic ridges in oceanic crust

Rift Valleys in Continental Crust

When a divergent boundary develops in continental crust, blocks of crust drop down into the gap created producing an elongated low spot called a **rift valley**. The classic modern example is the Great Rift Valley of east Africa (pictured here).



Figure 3.3.6: A fresh divergent boundary developing in continental lithosphere, east Africa. (NASA photo, annotated by Callan Bentley.)

The edges of the rift valley are high in elevation, which encourages sediment (loose material) to drop from the surrounding highlands into the rift valley. This sediment does not travel far, or for very long so material of various sizes and compositions pile up along these rift valley edges, smoothing out the highest relief areas. Smaller sediment gets carried a bit further, and creates deposits of sand. Water accumulate in the lowest regions, furthest from the rift valley edge, forming lakes that grow salty as they evaporate.

A very similar distribution of rocks and sediment 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 periods (150-250 million years ago), as the supercontinent Pangaea was breaking apart and the Atlantic Ocean was first forming.

Crustal thinning also allows molten material to rise to the surface. As the plates move apart at the divergent boundary, mantle material rises up. As the pressure is reduced, melting occurs without the addition of any new heat. This molten material may rise to the surface to erupt as “floods” of liquidy lava. This is seen as a large black patches in the satellite image above, showing the Afar Triangle region of northeastern Africa.



Figure 3.3.7: Volcanic ash from rift valley volcanism preserves trace fossils of hominids that walked upright.

As the rising molten material interacts with the crust on its ascent, it can transfer some of its heat and produce larger bodies of molten material, resulting in large volcanoes such as Mt. Kenya, Mt. Kilimanjaro, or Mt. Meru. It was ash erupted at one of these large volcanoes in the East African Rift Valley 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, a process called seafloor spreading takes over. This can be seen just north of East Africa, where a narrow body of oceanic crust separates Africa from Arabia. These gaps are filled with water creating the Red Sea and the Gulf of Aden. They are just a bit more 'tectonically mature' than the Great Rift Valley, which is not thinned enough to allow ocean water to flood it creating a sea.

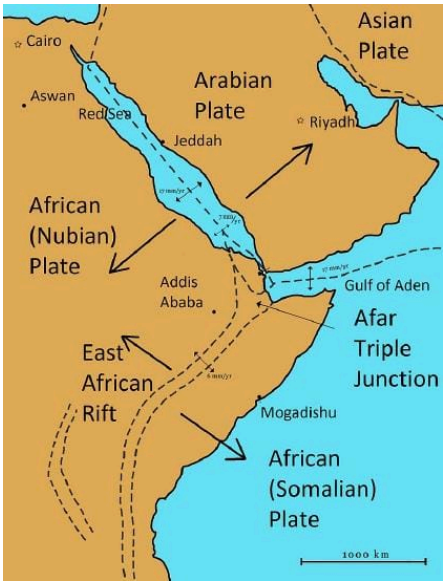


Figure 3.3.8: Junction of the African (Nubian), African (Somalian), and Arabian plates showing how the East African Rift has not dropped low enough for water to flood it like what has occurred to form the Red Sea and Gulf of Aden. Used under the CC BY 3.0 license.

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**.

Oceanic Ridges in Oceanic Crust

When oceanic lithosphere diverges and seafloor spreading occurs, mantle rises up and the pressure lowers triggering the formation of melt, a process called **decompression melting**. This melt cools at various depths to produce new oceanic crust. When freshly formed, this oceanic crust makes a high feature on the seafloor, an **oceanic ridge** (also called a "mid-ocean ridge," since many of them are in the middle of the ocean basins).

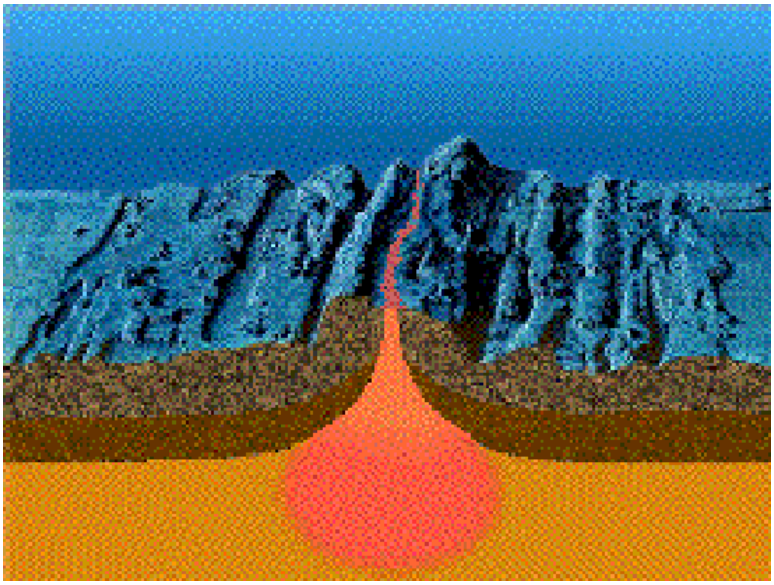


Figure 3.3.9: The formation of sea floor at an oceanic ridge. (USGS, Public Domain)

As the plates continue to diverge, this melt squirts into the cracks. Some of the melt may make it all the way into the water and the bottom of the ocean, erupting to produce pillow basalts (see figure below), while some magma cools rapidly and solidifies in the feeder crack itself, making a what is called a dike (see figure below).

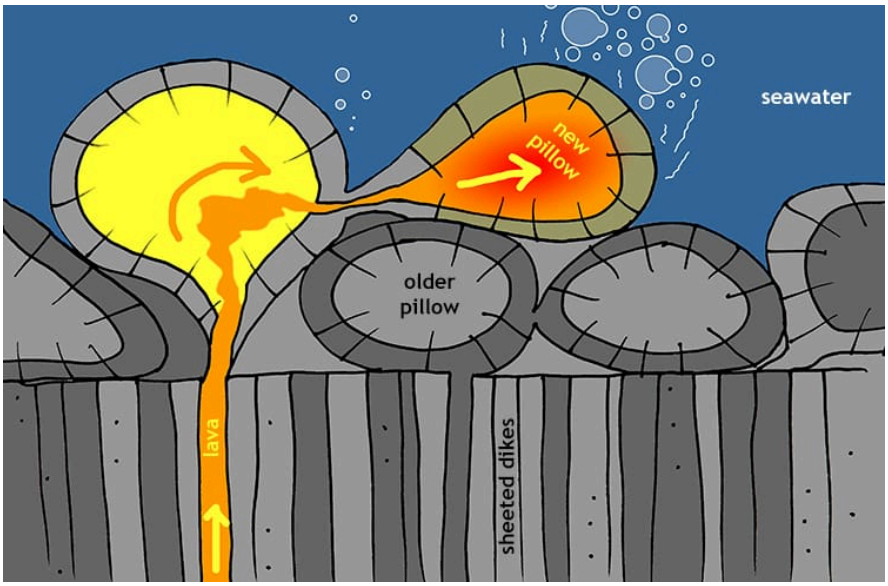


Figure 3.3.10: The formation of pillow structures occurs when melt erupts underwater. (Callan Bentley cartoon.)

This all produces a rock called basalt, which is explained elsewhere in this text. Still more melt cools slowly, deeper in the new oceanic crust, producing bodies of a rock called gabbro which is also explained elsewhere in the text. This sequence of rock types is an ophiolite sequence, an important feature discussed below in the convergent boundaries section.

Seafloor spreading happens at different rates in different ocean basins. The East Pacific Rise in the Pacific Ocean generates

roughly twice as much new seafloor in a given amount of time as the Mid-Atlantic Ridge in the Atlantic Ocean. This is seen in the diagram below as young seafloor indicated using red, orange, and yellow covers much more of the Pacific Ocean than the Atlantic.

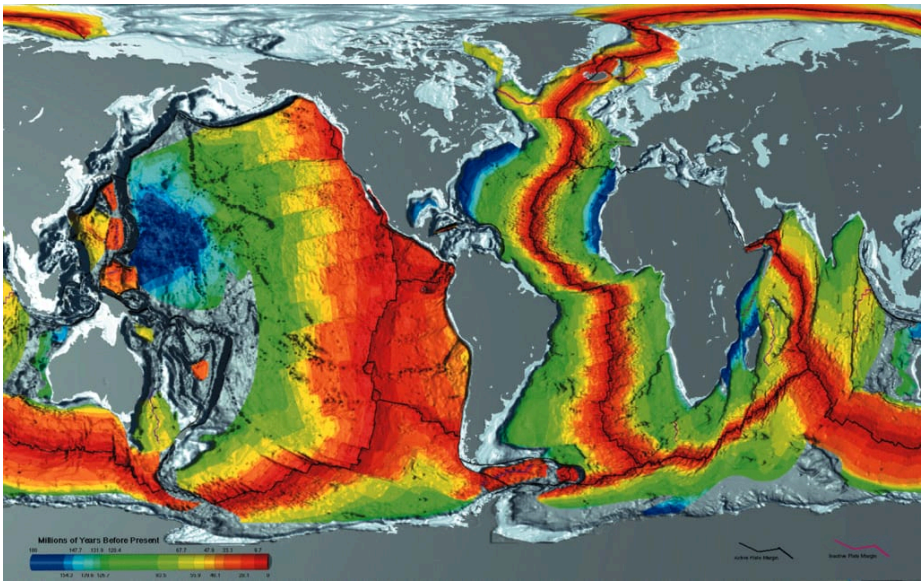


Figure 3.3.11: Age of the oceanic crust on Earth. (NOAA map.)

Oceanic crust also records the changes in Earth's magnetic field over time, as magnetic north (currently roughly lined up with the Earth's North Pole) flip flops back and forth at irregularly scheduled intervals. The melt that cools to make oceanic crust contains minerals such as magnetite, which will rotate into alignment with Earth's magnetic field before the melt around them solidifies. The generation of new oceanic crust through seafloor spreading therefore records changes in the polarity of Earth's magnetic field at any given time and creates a continuous record.

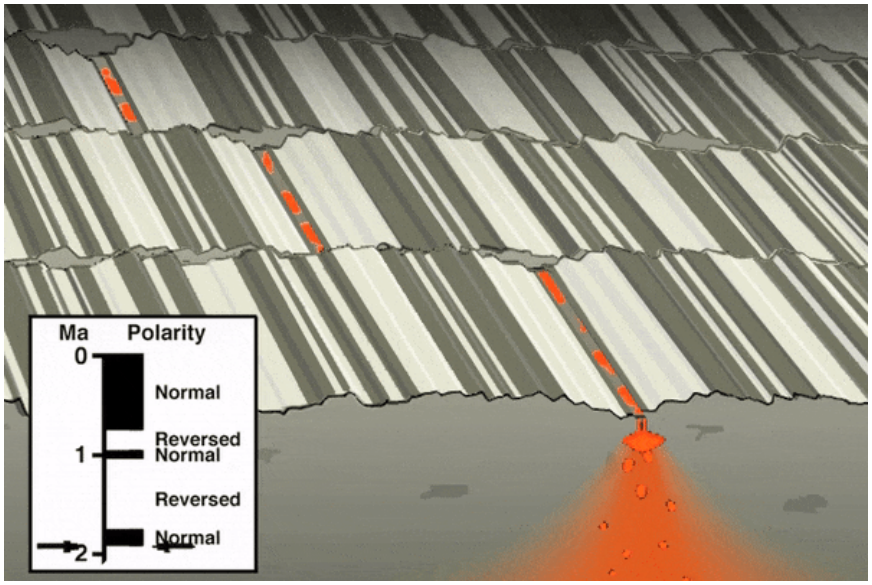


Figure 3.3.12: Spreading along several mid-ocean ridges, showing magnetic striping symmetry. (Animation by Tanya Atwater, UCSB.)

Convergent Boundaries

Convergent boundaries are sites where two plates move toward one another. Depending on whether the leading edge of these two plates consists of oceanic lithosphere or continental lithosphere, several different situations can result.

Oceanic-Continental Convergence

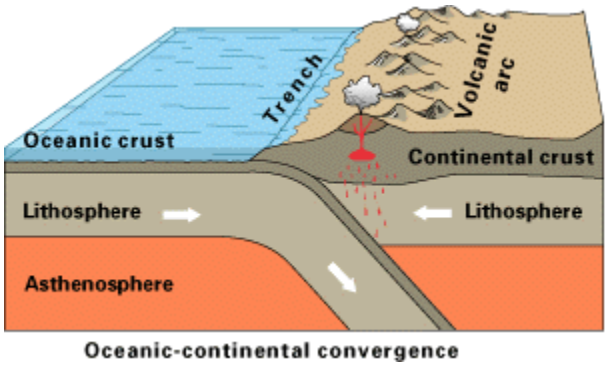


Figure 3.3.13: Key features in an oceanic-continental subduction zone. (Public domain via USGS.)

When oceanic lithosphere converges with a continent, **subduction** happens. Subduction is when the plate of oceanic lithosphere descends down into the mantle beneath the plate of continental lithosphere. This subduction is marked by many phenomena: an oceanic trench (see figure above) 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 boundary type can be among the largest known. At the trench, any sediment or other features atop the plate may be scraped off, building up in a thick jumbled pile called an **accretionary wedge**.



Figure 3.3.14: Osorno (left) and Calbuco (right) are classic Andean volcanoes in Chile marking a oceanic-continental convergent boundary. (Callan Bentley photo.)

As it descends, the subducted plate releases water into the surrounding mantle. This water enables the melting of the mantle rock, generating 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 the creation of a mixed melt, which continues to ascend through the crust. They may lodge at depth, crystallizing to make solid rock bodies called plutons (smaller) and batholiths (bigger), or they may find their way to the surface, where they erupt as volcanoes. The development of a **continental volcanic arc** (chain of volcanic mountains found along a coastline) parallel to the trench is a sure indication of oceanic-continental convergence.

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 (continental lithosphere), where subduction of the Nazca Plate (oceanic lithosphere) has resulted in the Andes volcanic arc as well as some of the largest earthquakes ever recorded.



Figure 3.3.14: Osorno (left) and Calbuco (right) are classic Andean volcanoes in Chile marking a oceanic-continental convergent boundary. (Callan Bentley photo.)

Interactive Content

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Figure 3.3.15: Shaded relief map of Earth showing concentration of earthquakes (black dots) and volcanoes (red dots) along plate boundaries which are often along coastlines. (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. When geologists find these complexes they know subduction occurred there sometime in the past at a convergent boundary.

Oceanic-Oceanic Convergence

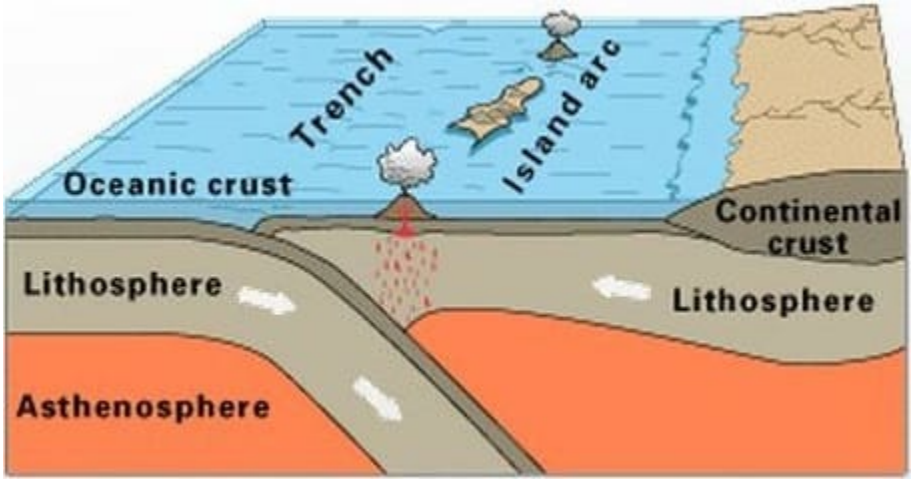


Figure 3.3.16: Key features in an oceanic-oceanic subduction zone. (Public domain via USGS.)

When two plates of oceanic lithosphere converge, subduction results when one of the two plates descends beneath the other. Because they are both composed of the same plate type, 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.

When old oceanic lithosphere subducts, a trench (see figure above) and an accretionary wedge results, just like when subduction occurs under continental lithosphere. Again the rock of the subducted slab releasing water into the mantle which triggers melting of the mantle. The melt 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):

Interactive Content

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Figure 3.3.17: There is a strong coincidence of deep sea trenches and parallel belts of volcanoes. (Dale Sawyer's "Discovering Plate Boundaries" maps.)

Continental-Continental Convergence

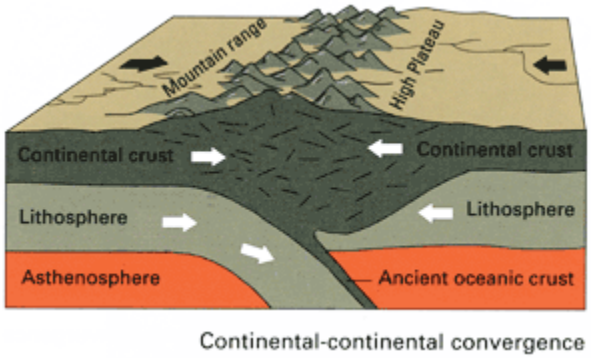


Figure 3.3.18: Key features in an continent-continent convergent boundary. (Public domain via USGS.)

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 root 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 since melt is not produced in this process.



Figure 3.3.19: Aerial view over the raw mountains formed as India and Asia collide. (Photograph by Susan Hough; reproduced with permission.)

If the compressional stresses are strong, rocks on the flanks of the mountain belt will be deformed through folding (bending) and faulting (breaking). This zone of compressional deformation is distinct and pervasive in mountain belts across time and space. 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.



Figure 3.3.20: Zerrissene turbidites, folded by the Damara Orogeny, Namibia. (Photograph by Jay Kaufman; reproduced with permission.)

Here are a few images highlighting folds created in this way around the world:



Figure 3.3.21: Folded sedimentary layers are a signature of compressional forces associated with mountain-building.



Figure 3.3.22: Google Earth Image of the Appalachian Mountain fold belt in western Pennsylvania at 200 km eye altitude.



Figure 3.3.23: Cape Liptrap (southeast of Melbourne in Victoria, Australia). (Photo by Sandra McLaren.)

Plate Interiors



Figure 3.3.24: 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 single block, traveling in a common direction or rotating around some axis of motion. **Passive margins** are the "trailing" edges of continents within a plate such that they are marked by coastlines without a plate boundary. 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** where the coastline does mark the plate boundary (with Cascadia subduction and two giant transform fault zones). 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 sediment. These sediments 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, as plate boundaries do not exist at these coastlines.

Key Terms

- **accretionary wedge** - structure near an oceanic trench formed by the accumulation of sediment scraped off the subducting oceanic lithosphere and shoved onto the overriding oceanic or continental plate
- **active margin** - a continental coastline that is marked by a plate boundary
- **continental volcanic arc** - a chain of volcanoes formed on a continental plate as a result of the subduction of oceanic lithosphere
- **convergent boundary** - plate boundary type at which the two plates move toward one another
- **decompression melting** - melting of rock triggered as rock rises and pressure is lowered
- **divergent boundary** - plate boundary type at which the two plates move away from one another
- **mountain belt** - a mountain chain formed as a result of the collision of two continental plates
- **oceanic ridge** - mountain chain at the bottom of the ocean, often in the middle of the ocean, created at divergent boundaries where new crust is created
- **orogeny** - mountain building event occurring due to compression at a convergent plate boundary
- **passive margin** - a continental coastline that is not marked by a plate boundary
- **rift valley** - elongated low spot created when a divergent boundary develops in continental lithosphere
- **seafloor spreading** - process where new oceanic crust is formed at oceanic ridges and moves away from the ridge
- **subduction** - process by which dense oceanic lithosphere dives down beneath another tectonic plate
- **transform boundary** - plate boundary type at which the two plates slide past one another
- **volcanic island arc -** a chain of volcanic islands formed on an oceanic plate as a result of the subduction of oceanic lithosphere

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3.4: Geophysical Phenomena

Earthquakes

Most earthquakes are associated with plate boundaries. Examine this map showing more than a century of epicenters to get an immediate sense of this pattern:

Figure 3.4.1: GigaMacro image taken from https://opengeology.org/historicalge...ate-tectonics/

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 the resistance created by friction. Most (90-95%) of the places where rock is forced into the uncomfortable position of scraping over other rock are located at plate boundaries. The exceptions (5-10%) 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.

Volcanoes



Figure 3.4.2: 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 Mountains)
2. Subduction of oceanic lithosphere under oceanic lithosphere (example: the Philippines or Aleutian Islands in Alaska)
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 (areas where heat is concentrated in the mantle). The idea here is that, independent of plate motions, the mantle has isolated places where rising warm material moves upward, and in so doing partially melts. The melt 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)

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3.5: 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 move? **Convection** in Earth's mantle is relevant, but probably greater momentum comes from the forces of **ridge push** and **slab pull**. Watch the video to explore convection, ridge push, and slab pull.

YouTube Video

This interactive element is not available in this format. Visit the link or scan the QR code to access it online.

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IRIS video.

Key Terms

- **convection** - the movement of the mantle as it transfers heat from the hot core to the brittle lithosphere
- **ridge push** - the elevation of ocean ridges, caused by rising magma, pushes the plates away from the high spot down toward the deeper ocean floor causing the plate to move
- **slab pull** - the pulling force exerted by an oceanic plate subducting into the mantle due to its own weight

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3.6: The Beginning of Plate Tectonics and the Continents

Introduction

Understanding the origin of Earth's first rocks is fairly straightforward, even though no direct evidence of their existence has ever been found. At the start, Earth was basically a hot ball of magma with the densest (metallic) material sinking to form the core and lighter (silicate) material floating upwards and cooling enough to form the first crust, most likely a very mafic (high in iron and magnesium and relatively low in silica) veneer on a magma ocean. Earth may have resembled modern Venus with a crust of mafic volcanic rock, no plate tectonics, and volcanism driving the constructive processes shaping the surface.

As for the start of plate tectonics, the story is not so straightforward. We do know that modern subduction zones didn't exist in the earliest chapters of Earth's story because subduction-zone rocks, such as blueschist (a metamorphic rock covered elsewhere in this text), are missing from the rock record of the Archean and earlier. Today, blueschists form under high pressure and relatively low-temperature conditions along subduction zones. Using **uniformitarianism**, we can infer that a lack of old blueschists indicates that the tectonic conditions needed to form them didn't yet exist. Consequently, determining the transition from more "Venus-like" conditions to the type of planetary geology we have today, with distinct slabs of interacting continental and oceanic crust, is not so straightforward.

Subduction and Continents

The problem with determining the start of plate tectonics is that we don't know when the lithosphere, with distinct continental and ocean crusts, was established.



Figure 3.6.1: A cross-sectional view of Earth. From top to bottom are the lithosphere, asthenosphere, mesosphere, and inner and outer cores. Image from Volcan26, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons

This is crucial because an outer layer composed of distinct continental and oceanic crust enables subduction, a key component of plate tectonics. In fact, subduction may be the most important component in making plate boundaries and driving plate tectonics. Once subduction begins, pieces of crust (terranes) could accrete (smash together like globs of play-dough) to build continents through collisions.

Watch this video to learn about how accretion works.

YouTube Video

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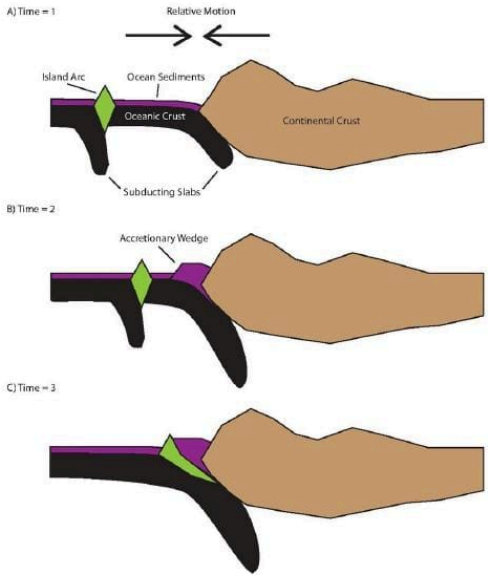


Figure 3.6.2: Cartoon showing the general process of accretion in growing continents. (Graphic: BZenith via Wikimedia.)

In contrast, where the ocean crust is stretched, it can eventually rift apart into two pieces, allowing seafloor spreading.

So, how did plate tectonics start (before there was distinct continental material)? And how can continental crust be made without plate tectonics? In modern settings, making the felsic rock that composes the continental crust components of plates involves the differentiation of partially melted mantle material through the subduction process and partial melting. This is summarized in the illustration below, where (1) water released from the subducting plate reduces the melting temperature of the overriding plate. This causes partial melting (2) of the ultramafic (very high in metals) mantle and generation of basaltic (high in metals) magma, which rises to the base of the continental crust where it stalls out (3) due to it being more dense. The heat from this magma partially melts the continental crust (4), making intermediate-felsic (lower in metals; higher in silica) magma that rises and further differentiates as it incorporates the continental crust and mixes with other plutons (5 & 6).

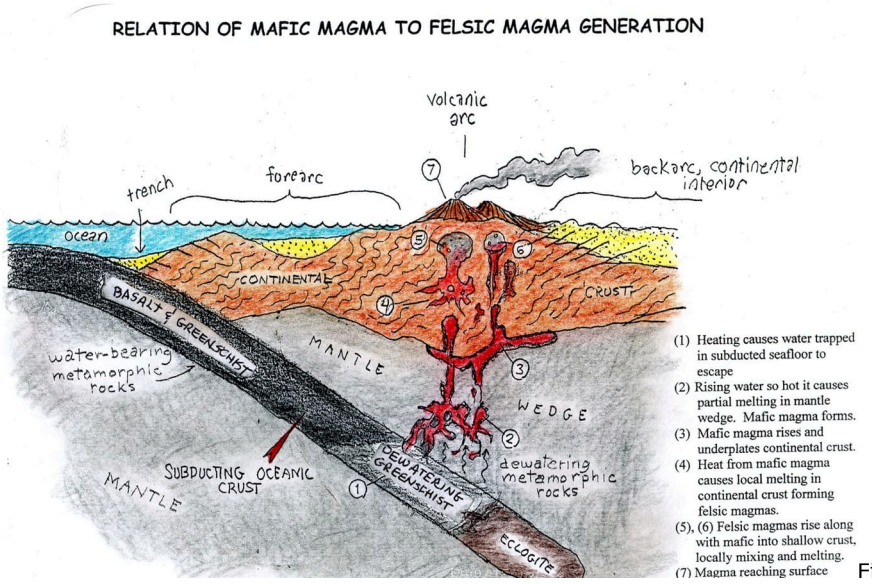


Figure 3.6.3: Relation of mafic magma to felsic magma differentiation by Dee Trent.

We arrive at a chicken-and-the-egg scenario, where we can't be certain which came first: distinct masses of continental crust or plate tectonics. There likely was a gradual transition from slabs of ultramafic-mafic (high in metals) crust to distinct plates composed of oceanic and continental crust, with some subduction-like process, or "proto-subduction," as the first step.

Proto-Subduction

Even without plate tectonics as we know it today, it's possible to make some felsic (low in metals; high in silica) material, or at least material that is more felsic than the average crust at the time, in various ways. Perhaps that is all that is needed; a slight increase in felsic material in a section of the crust might have been enough to make it buoyant, start the more complex processes described above, and make full-fledged continents.

The internal heat of the early Earth was much higher than today (as much as 300 degrees Celsius), so there was higher heat flow out of the planet and, thus, a warmer lithosphere. Modeling indicates that this heat would have weakened the lithosphere and made it more prone to breaking. A trigger, such as the extra heat associated with rising mantle plumes (Greya *et al.*, 2015), the "lava" in a lava lamp, or even a very large impact (*e.g.*, O'Neil et al., 2019), may have weakened the lithosphere enough to initiate a subduction-like process.

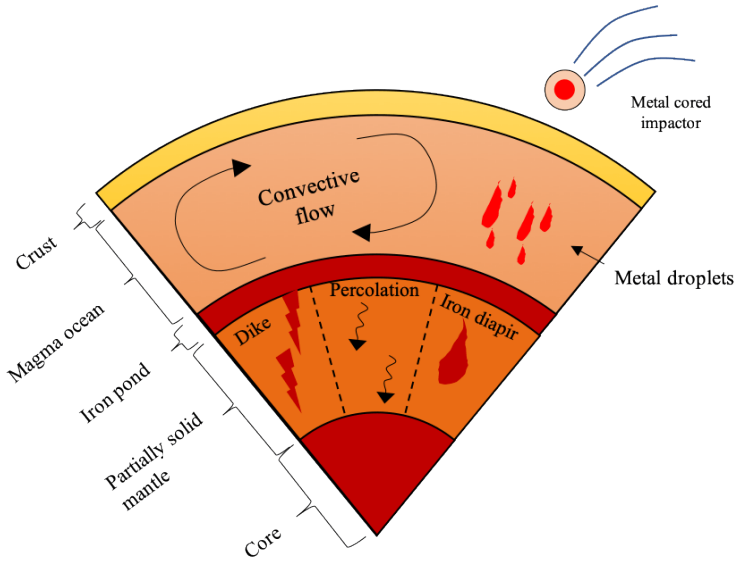


Figure 3.6.4: Hypothetical core-mantle differentiation processes: Percolation, diking, and diapirism. After Rubie et al. (2015). (CC BY-SA 4.0; By AlexInMetal – Own work, https://commons.wikimedia.org/w/inde...curid=76607390)

What was this earliest form of tectonics like, and what is the evidence for it? The oldest observable features that can answer this question are called **greenstone belts**. These are linear or branching zones of distinctive low-grade metasedimentary (metamorphosed sedimentary rocks) and metavolcanic (metamorphosed volcanic rocks) rocks. Notable greenstone belts on Earth include examples from the oldest parts of the shields ("continental nuclei") of Australia, South Africa, Europe, Siberia, Brazil, Antarctica, India, Greenland, China, and north-central North America. All are included in an otherwise fairly homogenous felsic crust of similarly ancient age.

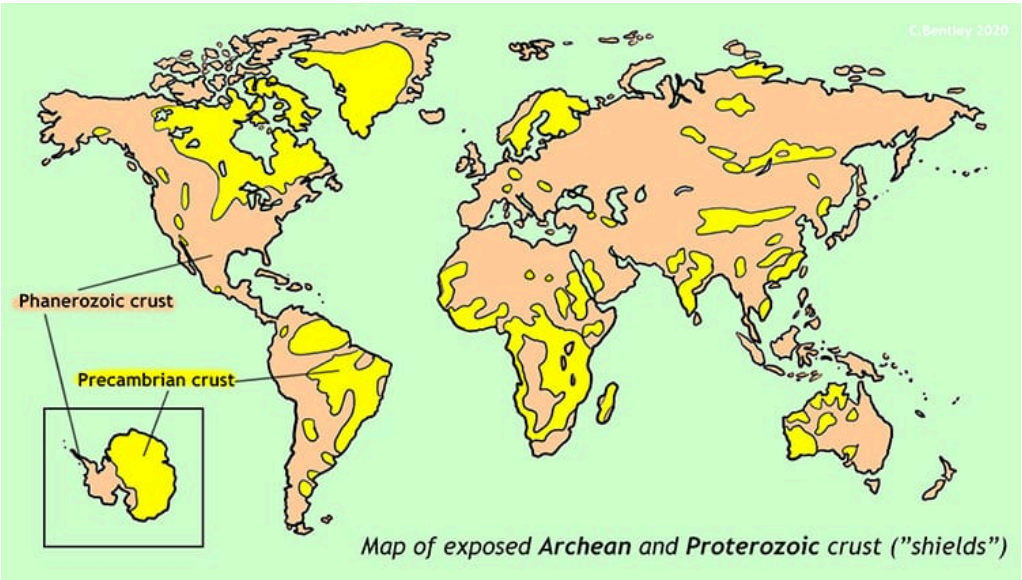


Figure 3.6.5: Map showing the distribution of Precambrian (Archean + Proterozoic) crust on Earth. (Callan Bentley graphic)

The greenstone belts are typically sandwiched between large bodies of cratonic crustal rocks such as granite and gneiss. Not all greenstone belts are Archean; some are also known from the Proterozoic.

The greenstone belts are interpreted as remnants of ancient ocean basins that were squeezed between ancient proto-continental terranes, in the same sense that, in modern plate tectonics, ophiolites mark the positions of former ocean basins that have since been uplifted. The process of partial melting repeatedly "distilled" the planet's silicate crust 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 ("keel" in the illustration below), 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 remained stuck together (accretion), forming a granite/greenstone belt/granite horizontal granite sequence. This geometry is called “dome and keel,” and it’s thought to represent a kind of vertical tectonics that dominated during the Archean. Below is a diagram showing the relative motion of the plutonic domes and the surface strata folded into the thin keels.

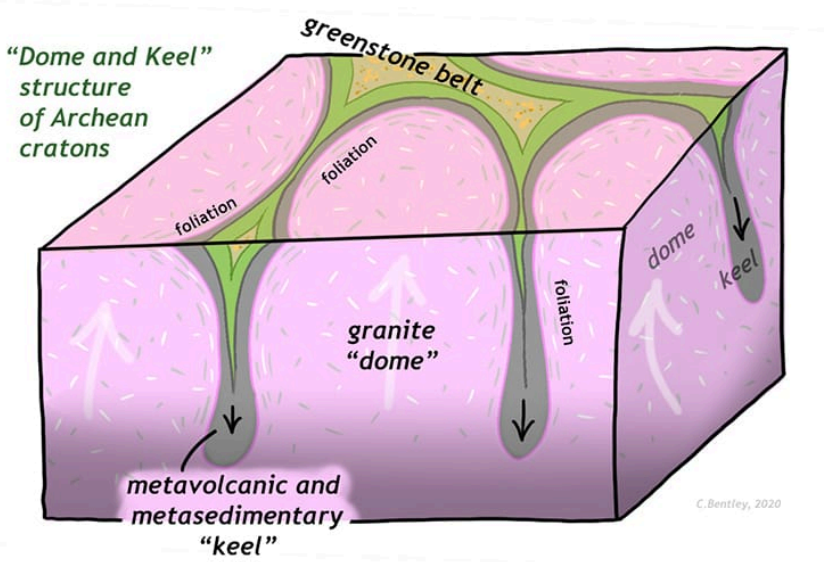


Figure 3.6.6: Block diagram showing the structure of a typical Archean craton, such as the Pilbara. Arrows show relative motion of the granite “domes” relative to the metavolcanic & metasedimentary “keels.” (Callan Bentley cartoon)

The idea 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 formed through 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.

Early models for the movement of material within Earth have focused on "vertical tectonics" due to the structure of greenstone belts, implying upward-moving **mantle plumes** and **diapirs** and downward-moving keels. The extra heat in the mantle likely drove a more active plume system across the planet, with each plume causing heating and weakening of the lithosphere - perhaps similar to the hot spots of today but far greater in concentration. This is known as "plume-lid tectonics" - the rising material from plumes was "trapped" by the young, hot, but continuous lithosphere. Another part of this plume-lid model involves a rock known as eclogite. Models predict a thicker oceanic crust during this time, much thicker than today’s. At a thickness of about 40 km, basalt metamorphoses into eclogite, a much denser rock. This dense lower part of the crust would “peel off” in a process known as delamination and "drip" downward. This proto-subduction is a possible mechanism for material cycling before the onset of plate tectonics. It would also trigger mountain building and volcanism, with the hot, less dense asthenosphere replacing the dense eclogite and driving uplift and the partial melting of the overlying crust (Foulger, G. R. (2011). Plates vs plumes: A geological controversy. John Wiley & Sons.).

With time, accretion and partial melting would result in larger masses of more felsic, buoyant crust; meanwhile, Earth's interior cooled, slowing the rate of rising plumes and diapirs and that frequently disrupted the early continental crust. As crustal pieces became more rigid and permanent, their interaction may have resulted in a slightly denser piece sinking below a less dense one, which in turn could initiate convection currents in the mantle. Eventually, Earth's surface would be defined by distinct oceanic and continental crust, with denser oceanic lithosphere subducting beneath less dense continental crust. Early on, though, subduction is thought to have been shallower than it is today, because higher core heat flow would have prevented oceanic lithospheric slabs from remaining intact deep into the mantle, as is the case today in modern plate tectonics.

The video below, The World Before Plate Tectonics, starts at 2:42. Watch until at least 4:20.

YouTube Video

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View this YouTube Video online

As is typically the case in geology, the younger the rocks, the more detailed their story is because they have had less time to "lose information". In North America, the sedimentary rock record tells a more precise story about the start of the plate tectonics cycle.

Key Terms

- **diapir** - blobs of magma that rise towards Earth's surface
- **greenstone belt** - a linear or branching zone of metamorphosed sedimentary and igneous rock within Precambrian cratons; the name comes from the presence of green minerals like chlorite, actinolite, and eclogite
- **mantle plume** - long-lived narrow conduits of magma rising from Earth's core to make hot spots
- **uniformitarianism** - the recognition that the processes operating on Earth today must have been operating on Earth in the past as well

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3.7: The Story of Plate Tectonics as Recorded by the Rocks

Introduction

While the picture of early plate tectonics is still fuzzy, more recent rocks paint a clearer one. As mentioned elsewhere in this text, rocks record the history of the Earth. Sedimentary rocks are especially useful because their "pages" record the geologic and biologic activity happening on Earth's surface at the time they were made. The Appalachian region of the United States east coast provides an excellent record of 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 large interior 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 North American craton from the late Proterozoic Era through the end of the Paleozoic Era. The oldest, the Sauk Sequence, is dominated by carbonate rock formations. 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.

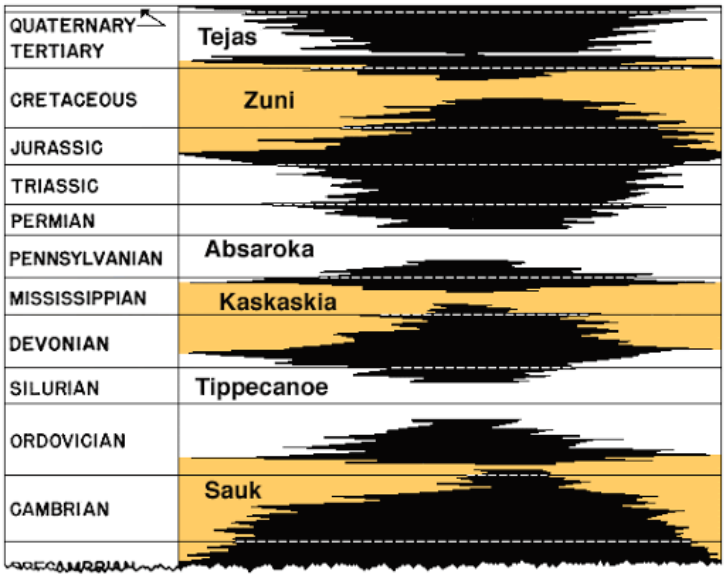


Figure 3.7.1: Sedimentary “Sloss” Sequences of North America. In the Mid-Atlantic region, the Sauk, Tippecanoe, and Kaskaskia are well exposed. White and yellow areas represent major marine transgressions onto the craton while black areas represent major unconformities (sequence boundaries). The Sauk 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 became famous because of a paper in which he identified six sequences of sedimentation that occurred along the margins and interior of ancient North America, known as Laurasia, from the late Neoproterozoic through the early Cenozoic Eras, a time span of close to a billion years. 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. This sequence of rock layers records the history of this sea at that location. As the Tippecanoe Sea surrounded all of Laurasia, there are many places in North America where you can see a progression of this sequence in the stratigraphic record, including, for example, the Grand Canyon.

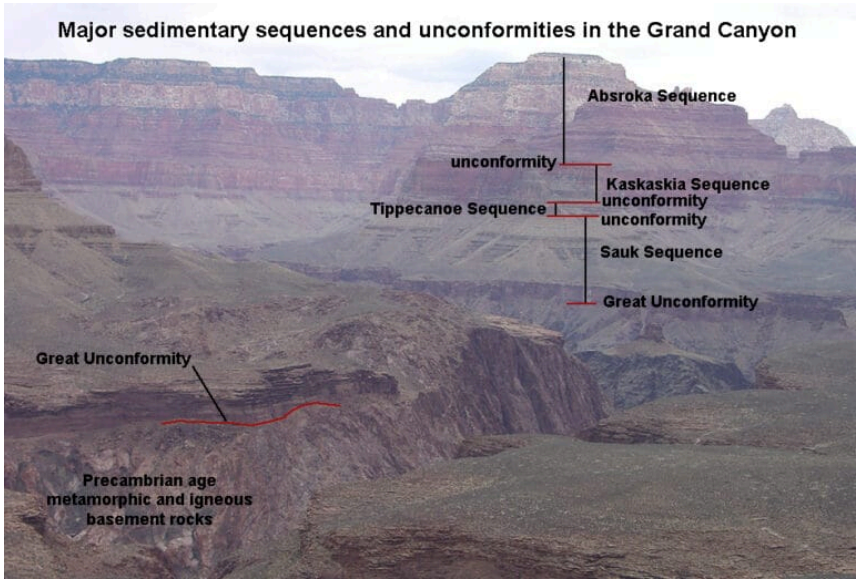


Figure 3.7.2: Cratonic Sequences, or Sloss Sequences, as they exist in the Colorado Plateau and Grand Canyon region in the American southwest.

Sloss (1964) identified these sequences through the development of curves based on data on land subsidence (lowering), cratonic interior uplift, and sedimentation. The curve explores global (eustatic) sea level changes related 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 a coastline progradation (coastline extending further out to sea) and then ends with a major transgression. In the middle of all of these sequences are major orogenic (mountain-building) events.

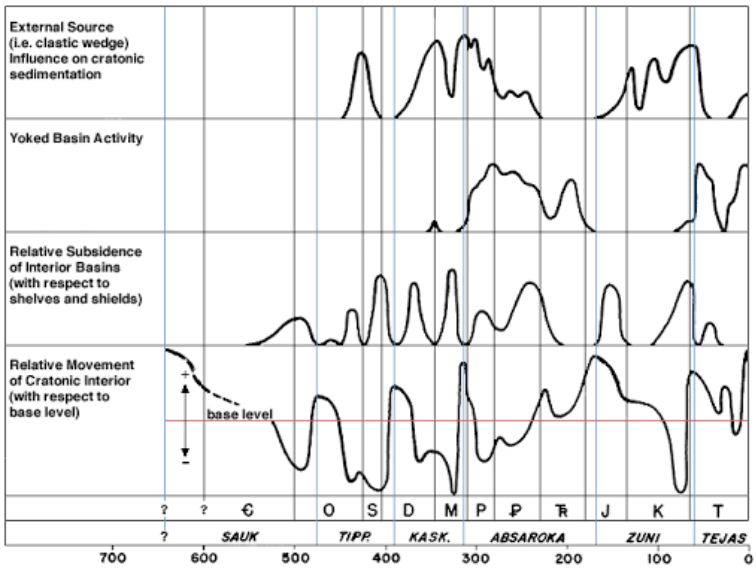


Figure 3.7.3: Four plots depicting relative tectonic activity across ancient North America. From the top down, Sedimentation fluctuations, Yoked basin activity, Relative land subsidence, and Relative movement of the cratonic interior. Tectonic, sedimentation, and global (eustatic) sea level activity 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 two-minute 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.

YouTube Video

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The Cycle of Plate Tectonics - 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 one-minute schematic video of the Wilson Cycle below.

YouTube Video

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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. The Tippecanoe Sequence takes place in this model between stages E through G.

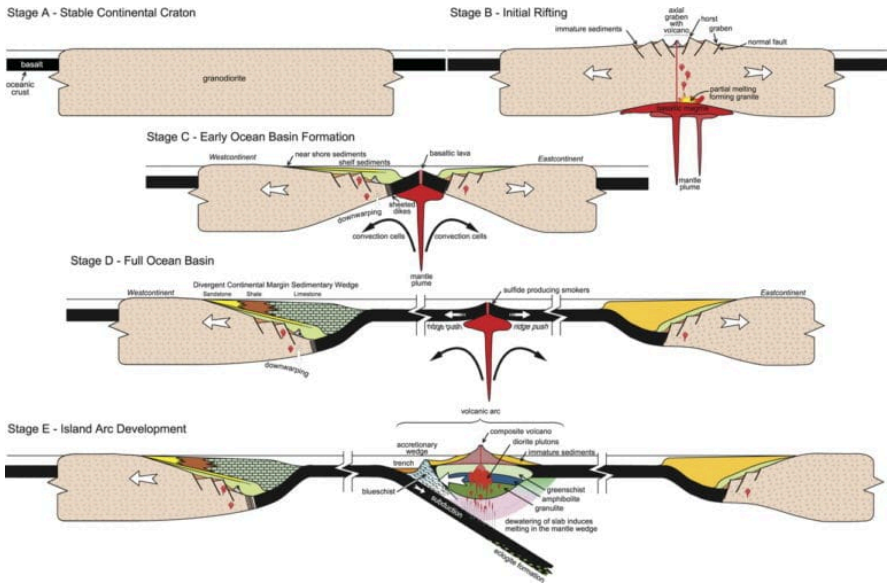


Figure 3.7.4: Stages A through E of a Wilson Cycle (Whitmeyer et al., 2007)

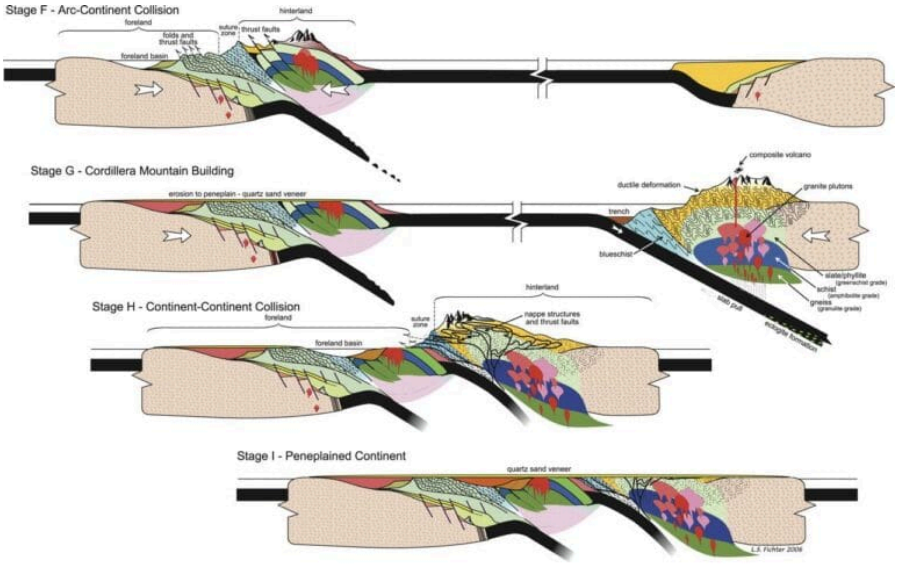


Figure 3.7.5: Stages F through I of a Wilson Cycle (Whitmeyer et al., 2007).

The Tippecanoe Sequence begins with missing rock, a 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 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 various important body fossils.

Key Terms

- **regression** - a time period when sea levels drop relative to land masses; caused by sea level drop or land level rise
- **transgression** - a time period when seas spread over land areas; caused by sea level rise or land level drop

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3.8: Key Points

1. Earth's outermost solid layer, the lithosphere, is broken into about 15 plates that move about relative to each other on the asthenosphere
2. Oceanic lithosphere is thinner and more dense than continental lithosphere
3. The three types of plate boundaries where the plates interact with one another are convergent (two plates collide) with compressional stress, divergent (two plates move away from one another) with tensional stress, and transform (two plates slide past one another) with shear stress
4. Divergent boundaries start as rift valleys and ultimately result in the formation of an ocean; volcanic activity is present during all phases
5. Convergent boundaries occur between two continental plates and result in the formation of mountains, between two oceanic plates and result in the formation of a volcanic island arc, and between one oceanic and one continental plate and result in the formation of a continental volcanic arc
6. Passive margins are coastlines without a plate boundary and is a site with no tectonic activity
7. Active margins are coastlines with a plate boundary where volcanism, earthquake activity, and/or mountain building occur
8. The majority of earthquakes and volcanic activity occur on plate boundaries

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4: EARTH MATERIALS - MINERALS

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).

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, as are fossil fuels and water which 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 the minerals largely found in the outermost solid layer of Earth, the lithosphere.

Minerals serve as the building blocks of rocks. If rocks tell the different stories of Earth's history, then minerals act as the chapters of those stories. Geologists look at the various properties of these minerals to identify them. The specific minerals in a particular rock give clues as to the conditions under which the rock formed. Since some rocks form in many ways, these minerals differentiate which story that rock tells.

4.1: What is a Mineral?

4.2: Building Blocks of Rock

4.3: Chemistry of Minerals

4.4: Formation of Minerals

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

4.6: Other Important Rock Forming Minerals

4.7: Key Points

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4.1: What is a Mineral?



Figure 4.1.1: 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 $\mathrm{CaCO_3}$, 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.”

Key Terms

- **mineral** - naturally occurring, inorganic solid with an ordered atomic structure and definite chemical composition

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4.2: Building Blocks of Rock

A **rock** is a solid substance that is made of one or more minerals or mineraloids. There are three families of rock composed of minerals: igneous, sedimentary, and metamorphic.

The detail behind the rock families is as follows:

- Igneous rock forms when molten material crystallizes.
- Sedimentary rock is composed of the products of mechanical weathering, (sand, gravel, etc.) and/or chemical weathering (minerals and mineraloids precipitated from solution).
- Metamorphic rock is produced by the chemical and physical reorganization of other rock under conditions induced by elevated heat and/or pressure.

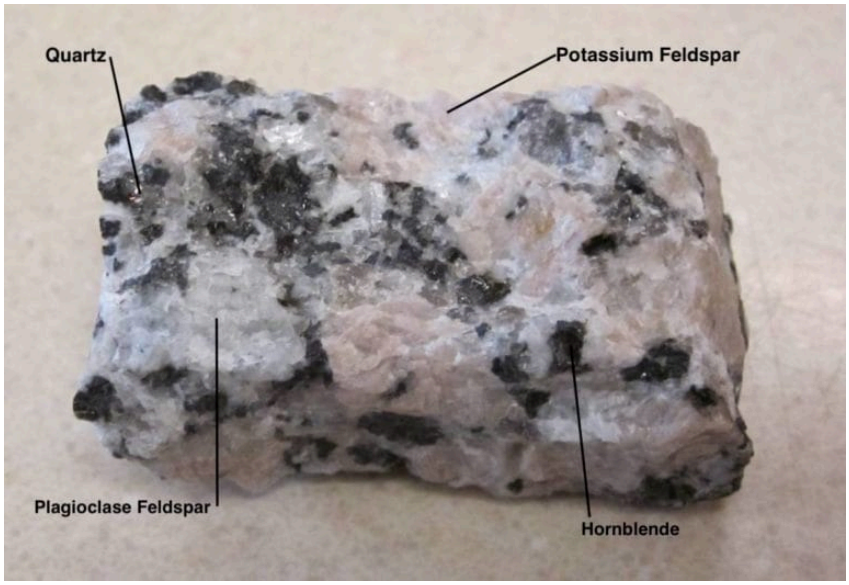


Figure 4.2.1: The common igneous rock, granite. The different minerals that compose the granite are labeled. (Creative Commons Attribution 3.0; Ralph L. Dawes, Ph.D. and Cheryl D. Dawes.)

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 (attached to the Earth; not separated like a pebble in a stream). 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 2025, there were over 6,000 minerals 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).

“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.

These ten minerals, plus a handful of others, will be our focus and will recur in many discussions in this text.

Key Terms

- **rock** - a solid composed of grains of one or more minerals or mineraloids

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4.3: Chemistry of Minerals

Atoms

To understand mineral chemistry, it is essential to examine the fundamental unit of all matter, the atom.

Matter is made of atoms. Atoms consist of subatomic particles: **protons**, **neutrons**, and **electrons**. A simple model of the atom has a central **nucleus** composed of protons, which have 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.

Note

To summarize:

The subatomic particles in an atom are:

- protons - positively charged, mass = 1 amu, located in the nucleus
- neutrons - no charge, mass = 1 amu, located in the nucleus
- electrons - negatively charged, mass = 0 amu, located outside of the nucleus

This means that all of an atom's mass is in the nucleus. The charge of the atom depends on the balance of protons and electrons such that if the number of electrons equals the number of protons, the element has no charge.

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. 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.

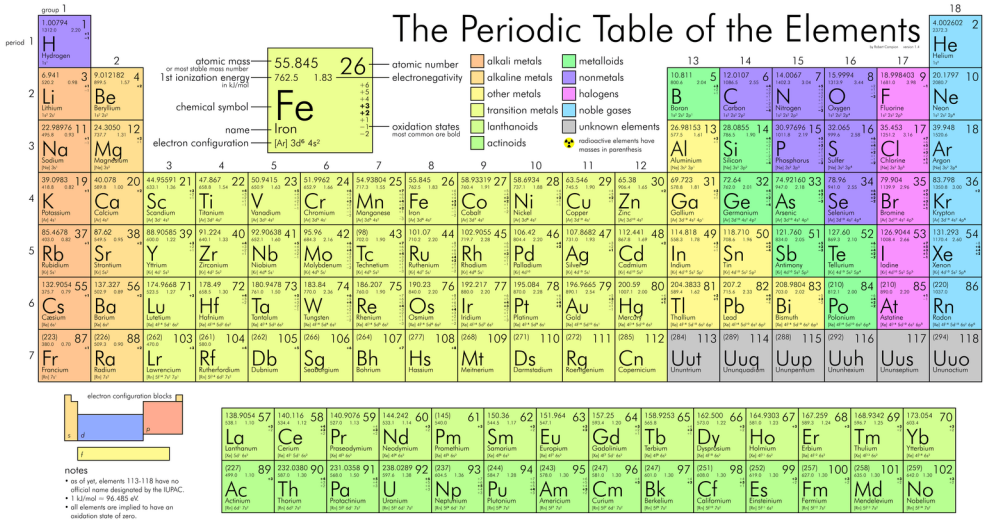


Figure 4.3.1: The periodic table of the elements of the world, showing element names, oxidation states, electron negativities, first ionization energies, and electron configurations. (Public domain, CC0 1.0 Universal; Wikimedia)

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, with the 0.01 being the mass of the electrons.

Note

To summarize:

Just like there are brands of cars there are elements of atoms. So Toyota and Honda are brands (or a type of car), just like Oxygen and Silicon are elements (or a type of atom). The element an atom is, is based on the number of protons. Oxygen is always 8; Silicon is always 14. To find the mass of an atom you add the number of protons and neutrons (since electrons carry negligible mass). Silicon is therefore 28 (14 protons + 14 neutrons).

Isotopes

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 geology. Later you see how they are used to give ages to certain rocks.

Note

To summarize and give an example:

To add to our car analogy from above, isotopes are varieties of elements just like there are different models of cars. For example Honda makes many models including Accords and Civics. Oxygen has several isotopes as the number of protons remains at 8 (that is what makes it Oxygen) and the number of neutrons varies between 8, 9, and 10. Accordingly the mass of the isotopes varies as follows:

- Oxygen isotope with 8 neutrons has a mass of 16 (8 protons + 8 neutrons); we call this O-16
- Oxygen isotope with 9 neutrons has a mass of 17 (8 protons + 9 neutrons); we call this O-17
- Oxygen isotope with 10 neutrons has a mass of 18 (8 protons + 10 neutrons); we call this O-18

Common Elements of the Earth's Crust

Among the 118 known elements appearing on the periodic table, 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.

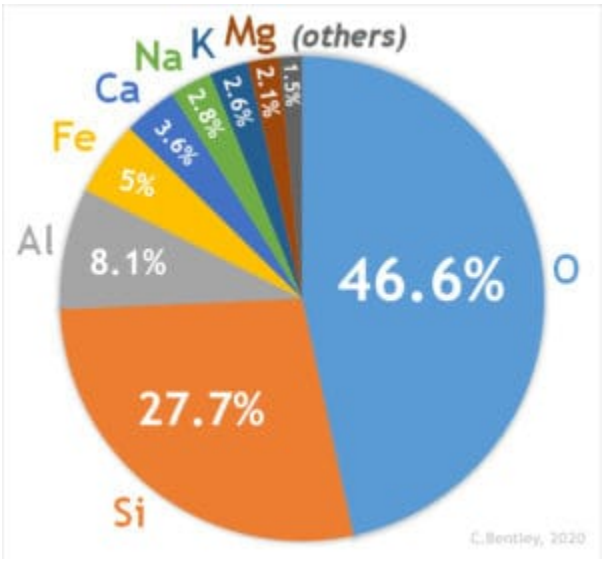


Figure 4.3.2: A pie chart showing the elemental composition of the crust.

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. Halite (table salt) forms as sodium (Na) and chlorine (Cl) combine with the chemical formula: NaCl. There are some minerals that are not compounds as they are composed of only one element. Examples include native Copper (Cu), native Gold (Au), or native Silver (Ag). These single elements bond to each other to form molecules.

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 and the electrons here are involved in chemical bonding. Atoms are more stable when this valence shell is full, so most elements tend to gain or lose electrons to achieve this. If an atom gains or loses electrons in its valence shell, the number of protons and electrons are no longer equal and it gains an either positive or negative charge. This charged atom is called an ion. If protons outnumber electrons the charge is positive and it is called a **cation**. If electrons outnumber protons the charge is negative and it is called an **anion**.

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. Currently, one atom of sodium (Na) has an equal number of protons and electrons so it has no change. When it loses the one electron, it becomes positively charged (more protons than electrons) and is a cation designated as Na^{+}. Coincidentally, chlorine (Cl) has one spot left for an electron in its valence shell. Chlorine (Cl) gains an electron, and becomes negatively charged (more electrons than protons) and is an anion designated as Cl^{-}.

Note

To summarize:

- ions form as atoms lose or gain electrons resulting in a positively or negatively charged atom
- ions that lose electrons have more protons than electrons and have a positive charge; they are called cations
- ions that gain electrons have more electrons than protons and have a negative charge; they are called anions

A chemical bond 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. If this happens repeatedly, the mineral halite forms and is known as NaCl.

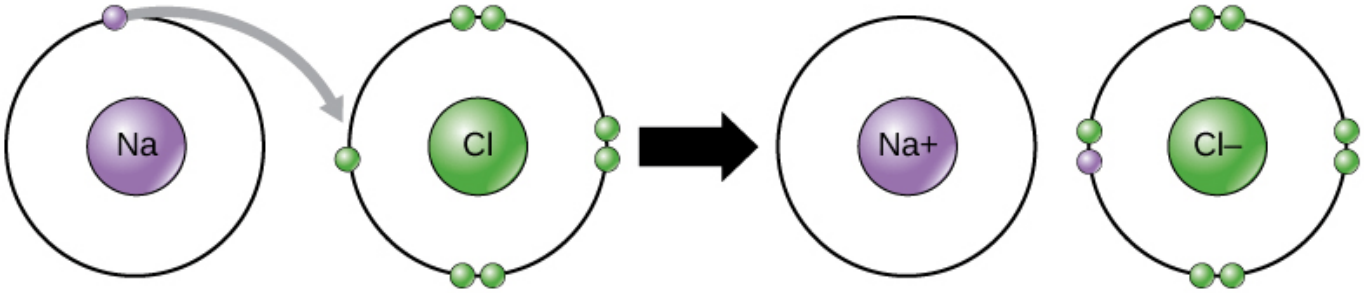


Figure 4.3.3: The formation of an ionic chemical bond by electron transfer. (CC BY 4.0; OpenStax, https://louis.oercommons.org/courseware/module/601/student/?task=9)

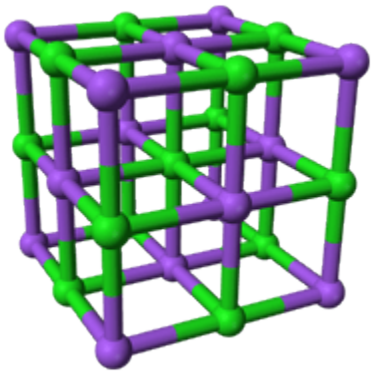


Figure 4.3.4: The unit cell for the mineral halite. (Public domain, CC0 1.0; Benjah-bmm27, Wikimedia)

Explore the following 3D molecular crystal structure of the mineral halite, below. Roll your cursor over to grab and rotate the structure. The grey dots are Na atoms and the green dots are Cl atoms.

Interactive Content

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Figure 4.3.5: 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.

Interactive Content

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Figure 4.3.6: 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 shape! We will look at this in more detail later.

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. Oxygen is missing 2 electrons to be stable (that shell wants 8 electrons) and hydrogen is missing one electron (that shell wants 2 electrons). Having two hydrogen atoms and one oxygen atom share their electrons, the valence shells will contain 8 electrons for oxygen and 2 for the hydrogens. Both configurations are very stable. This type of bond is called a **covalent bond** and is stronger than an ionic bond.

Interactive Content

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Figure 4.3.7: 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 which are the two most common elements in the Earth's crust. 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 tetrahedron (or **silica tetrahedron)** where four oxygen atoms pack very tightly around one silicon atom. The resulting molecular structure is shaped like a tetrahedron.

Interactive Content

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Figure 4.3.8: 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 center (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.)

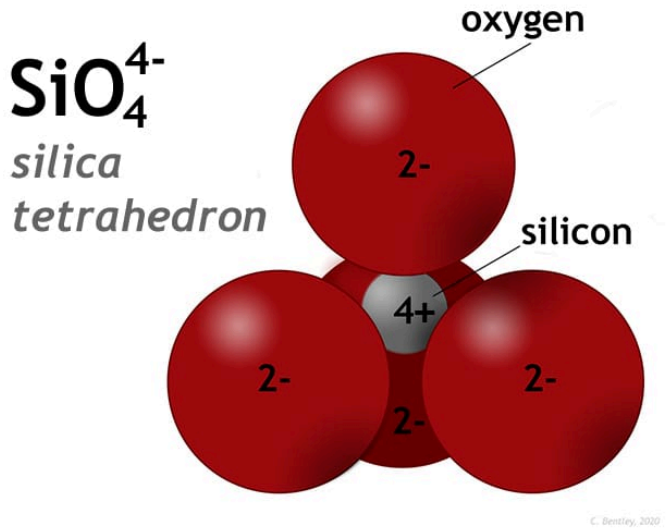


Figure 4.3.9: The atomic structure of the silica tetrahedron, a foundational molecular structure in all silicate minerals. (Callan Bentley)

Since the the overall charge of the silica tetrahedra is -4 the molecule acts as an anion. As a result, this 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.

Olivine most commonly forms as a magma cools and minerals begin to crystallize. The chemical formula for the mineral olivine is: $(Mg, Fe)_2SiO_4$ 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.

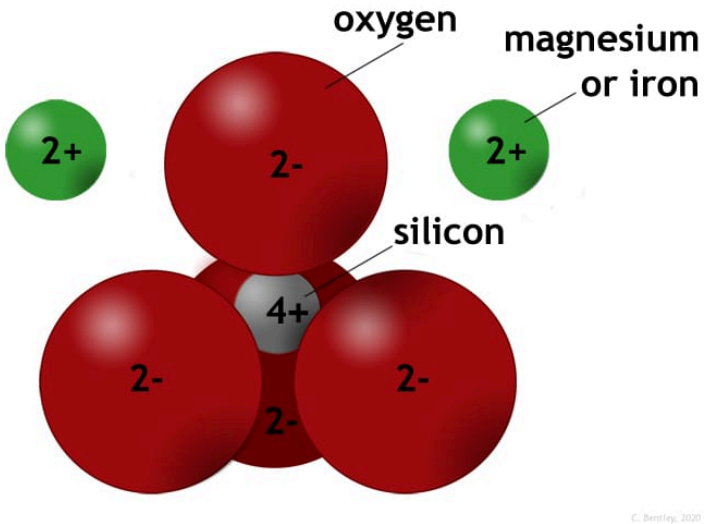


Figure 4.3.10: The unit cell of the mineral olivine features atoms of oxygen, silicon, and iron or magnesium (either will do just fine) in this arrangement. (Callan Bentley.)

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.

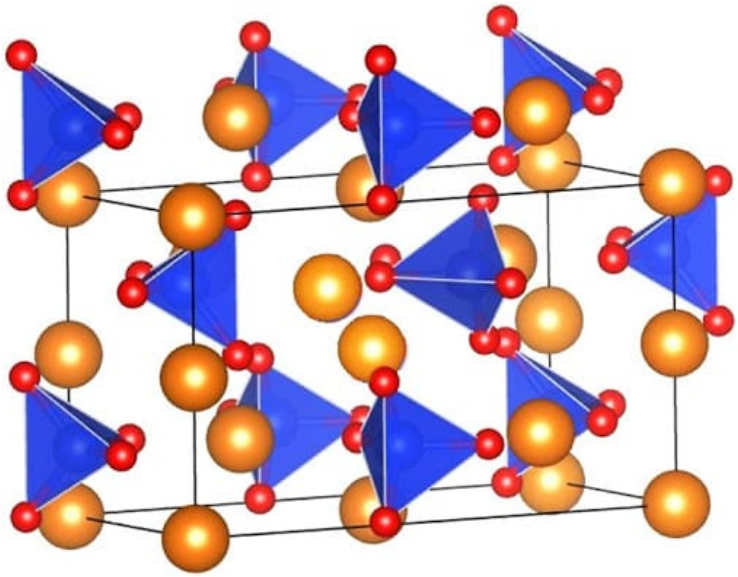


Figure 4.3.11: One unit cell of the mineral olivine. Silica tetrahedra represented in blue with oxygen atoms in red. Iron (Fe^{+2}) and/or magnesium (Mg^{+2}) ions are represented in gold. (Creative Commons Attribution 3.0; S Mahendran et al 2017 Modelling Simul. Mater. Sci. Eng. 25 054002)

Key Terms

- **anion** - a negatively charged ion formed as the number of electrons exceeds the number of protons
- **atomic mass** - the mass of an atom determined by adding the number of protons and neutrons in the atom
- **atomic number** - the number of protons in an atom which defines what element the atom is
- **cation** - a positively charged ion formed as the number of protons exceeds the number of electrons
- **covalent bond** - a strong type of bond in which atoms share electrons to complete their valence shells rending them stable
- **electron** - a negatively charged subatomic particle with essentially no mass that resides outside of an atom's nucleus
- **ionic bond** - a type of bond in which atoms lose or gain electrons to become negatively or positively charged and thus attracted to one another
- **isotopes** - varieties of a given element that vary in their number of neutrons and therefore in their atomic mass
- **neutron** - a subatomic particle with no charge having a mass of 1 atomic mass unit that resides inside an atom's nucleus
- **nucleus** - the central portion of an atom in which protons and neutrons reside
- **proton** - a positively charged subatomic particle having a mass of 1 atomic mass unit that resides inside an atom's nucleus
- **silica tetrahedron** - a molecule composed of one silicon atom and four oxygen atoms arranged in a tetrahedron with the silicon in the middle

[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)

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4.4: 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, the physical and chemical conditions must be favorable, and there must be 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.

Common Processes of Mineral Formation

Here is a list of how mineral form.

- Cooling: crystallization from molten rock material (magma or lava); the majority of minerals in the crust formed this way
- Organic formation: formation of minerals by organisms within shells (primarily calcite) and teeth and bones (primarily apatite)
- Precipitation: formation 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: 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.

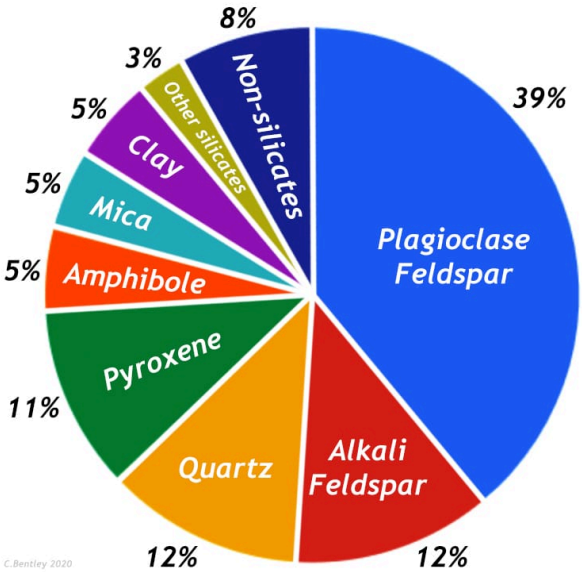


Figure 4.4.1: The most abundant minerals in Earth's crust. The crust is dominantly composed of just a few mineral groups, the vast majority of them silicates. (Callan Bentley.)

Crystallization from Magma and the Formation of the Silicate Minerals

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 of molten rock with freely moving ions. If molten rock below the surface (**magma**) is allowed to cool or if it is erupted onto the

surface (as **lava**), mineral crystals will form as temperature decreases.

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.

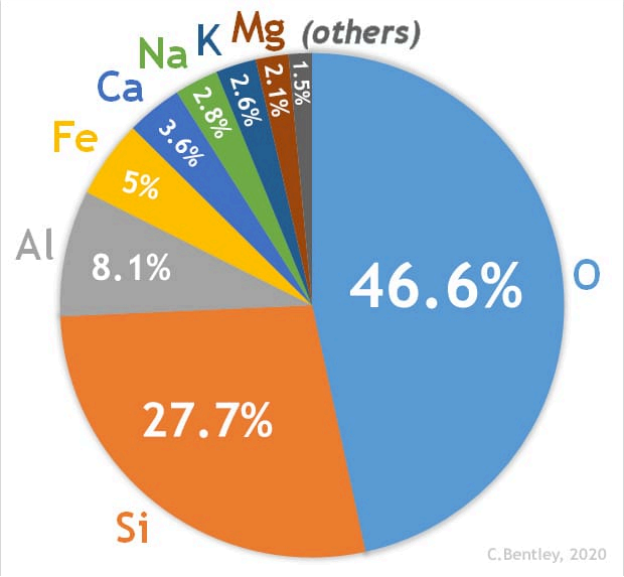


Figure 4.4.2: The elemental composition of the crust. (Callan Bentley.)

Bowen's Reaction Series

Most of the minerals of Earth's crust formed through the cooling of molten rock (magma or lava). The crystallization of minerals follows a very specific sequence. This sequence was first determined experimentally by the scientist Norman L. Bowen (1887-1956) in the early 1900's such that the order is referred to as "**Bowen's Reaction Series**". This specific process of crystallization leads us to the formation of nine of "The Big Ten" minerals listed earlier and presented in the diagram below.

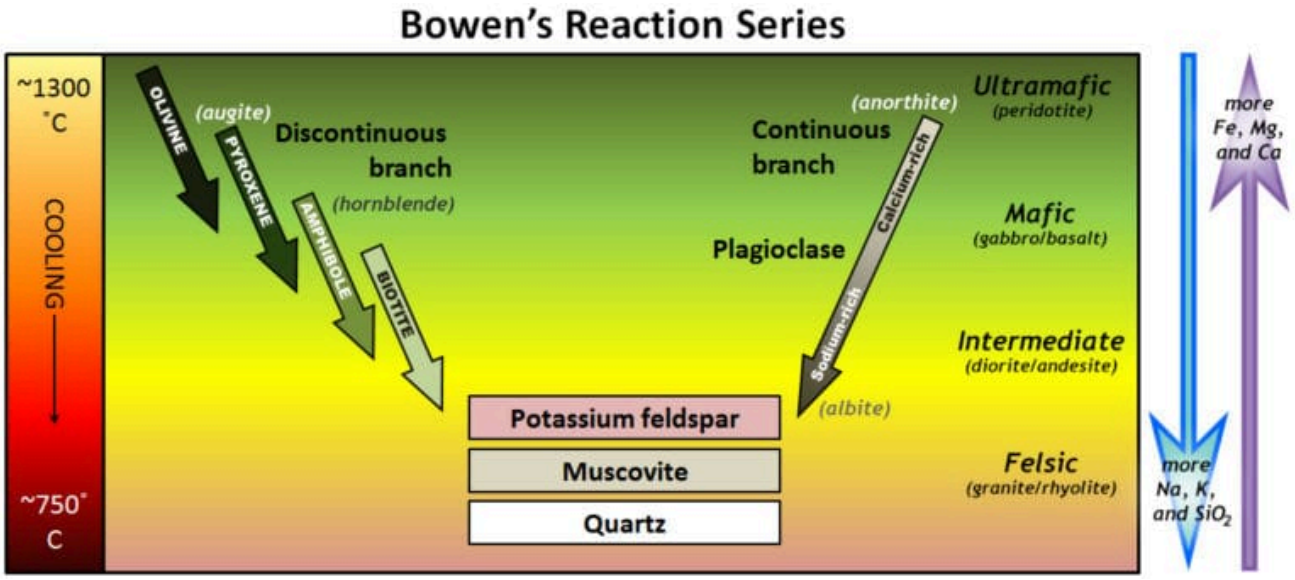


Figure 4.4.3: Bowen's Reaction Series. Developed through laboratory experimentation by Norman L. Bowen in the early 1900's. It establishes the order of crystallization of minerals from a silicate magma. (Creative Commons Attribution 4.0 International License; Modified after Steven Earle, https://opentextbc.ca/physicalgeology2ed/)

The left side of Bowen's Reaction Series 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. 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.

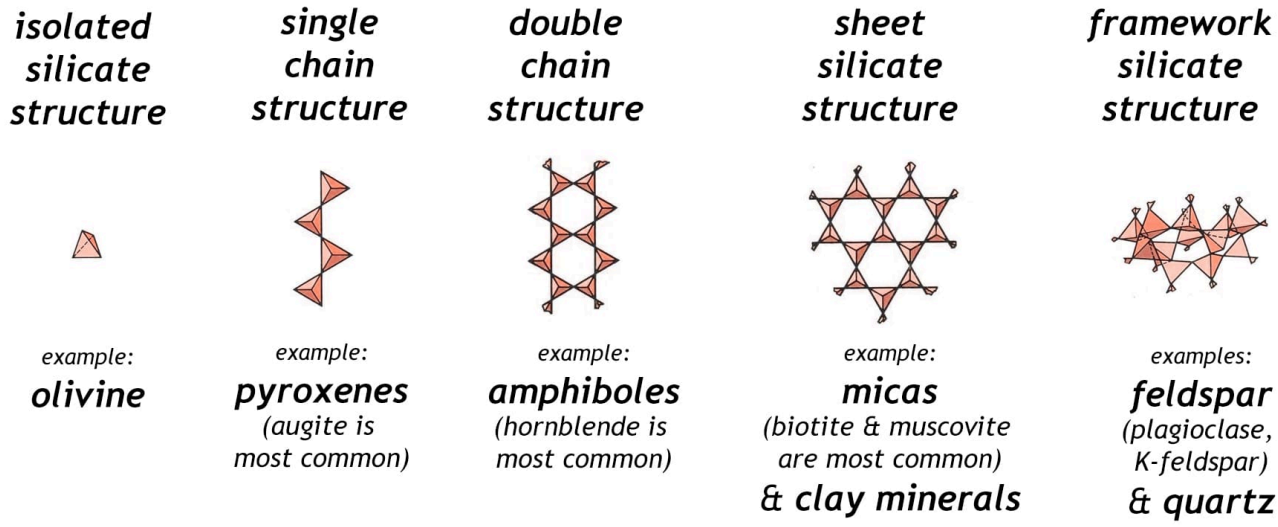


Figure 4.4.4: The various crystal structures of silicate minerals. (Callan Bentley.)

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.

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. 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.

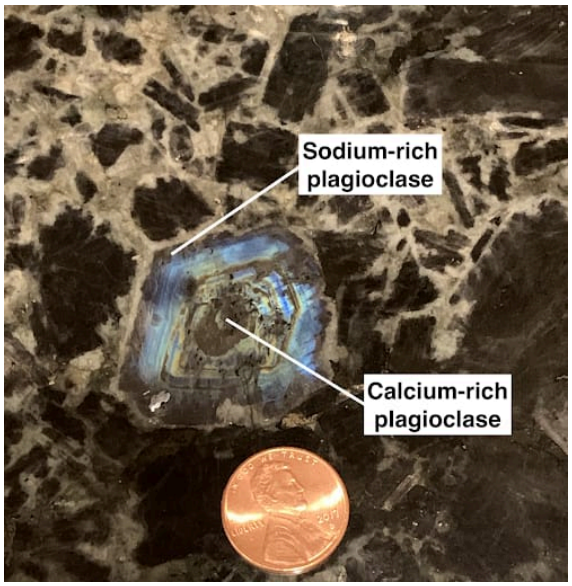


Figure 4.4.5: "Zoning" in a large crystal of plagioclase formed during the slow cooling of magma in an intrusive igneous body. The most calcium-rich concentration is located in the center that formed during the early cooling stages. The plagioclase became progressively more sodium-rich as cooling and crystal growth continued. The blue coloration is called "schillerization" due to microscopic inclusions common to this variety of plagioclase. (Shelley Jaye.)

Open and select play on this cooling 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.

Key Terms

- **Bowen's Reaction Series** - a diagram that depicts the order in which minerals crystallize as molten material cools
- **lava** - molten material on Earth's surface
- **magma** - molten material below Earth's surface

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

Olivine

Starting at the top of Bowen's Reaction Series's discontinuous branch, you find the most simple minerals. The first one is **olivine** which is mostly silica, iron, and magnesium and typically green in color when not weathered. Olivine is the primary mineral component in mantle rock (called peridotite) and ocean floor rock (called basalt). The chemical formula is $(\mathrm{Fe}, \mathrm{Mg})_2\mathrm{SiO}_4$. The comma between iron (Fe) and magnesium (Mg) indicates these two elements can substitute for one another. That "spot" can host either of those elements.

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Figure 4.5.1: 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 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: $(\mathrm{Ca}, \mathrm{Na})(\mathrm{Mg}, \mathrm{Fe}, \mathrm{Al}, \mathrm{Ti})(\mathrm{Si}, \mathrm{Al})_2\mathrm{O}_6$.

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Figure 4.5.2: 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 $(\mathrm{Ca}, \mathrm{Na})_2(\mathrm{Mg}, \mathrm{Fe}, \mathrm{Al})_5(\mathrm{Al}, \mathrm{Si})_8\mathrm{O}_{22}(\mathrm{OH})_2$.

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Figure 4.5.3: 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).

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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.

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Figure 4.5.4: 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.)

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Figure 4.5.5: 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. 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 found on the right side of Bowen's Reaction Series. 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.

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Figure 4.5.6: 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.)

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Figure 4.5.7: 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.)

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Figure 4.5.8: 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, SiO_2 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 SiO_2 however, impurities consisting of atoms within this framework give rise to many varieties of quartz among which are gemstones like amethyst (purple), rose quartz (pink), and citrine (yellow).

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Figure 4.5.9: Quartz, a framework silicate mineral composed of almost entirely SiO_2. 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.)

Key Terms

- **augite** - a common single-chain silicate mineral rich in iron and magnesium though other elements often substitute in making its formula complex; the most common mineral of the pyroxene family
- **biotite** - a sheet silicate rich in iron and magnesium making it light in color
- **hornblende** - a common double-chain silicate mineral; it has a complex formula due to its many possible substitututions
- **feldspar** - a framework silicate mineral with a wide variety of compositions as potassium, sodium, and calcium fit into the structure
- **muscovite** - a sheet silicate rich in potassium making it light in color
- **olivine** - a common silicate mineral rich in iron and magnesium making up much of the ocean floor and the mantle
- **quartz** - a framework silicate mineral composed entirely of silicon and oxygen though small amounts of other elements can cause color changes

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4.6: Other Important Rock Forming Minerals

Calcite

The final "Big Ten" mineral is **calcite** and was not previously discussed as it is not a silicate mineral. Calcite formation is most typically associated with sedimentary processes.

Calcite belongs to a family, or group of minerals, called the carbonates. The basic chemical and crystalline foundation for 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.

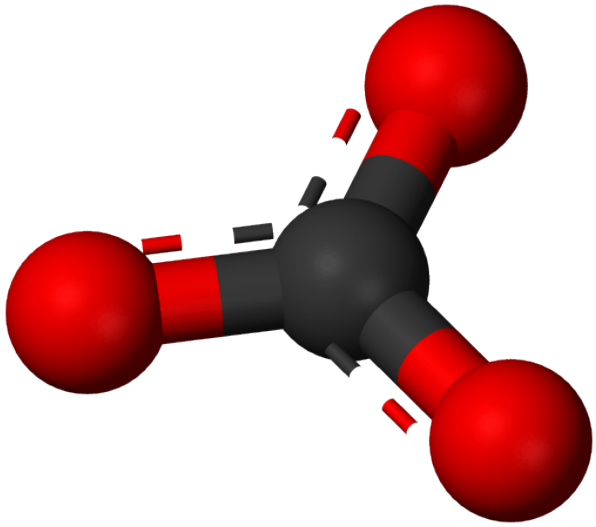


Figure 4.6.1: The complex carbonate ion $(CO_3)^{-2}$ where one carbon cation (brown, center) is surrounded by three oxygen anions (red) in a triangular pattern. This is the basic framework of “The Big Ten” mineral, calcite. (Public domain, CC0 1.0; Benjah-bmm27, Wikimedia)

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.

Crystallization of Calcite

Calcite most commonly forms 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

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Figure 4.6.2: 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.)

Dolomite is another common carbonate mineral closely related to calcite often occurring in similar environments of formation. Dolomite has many similar characteristics to calcite including its chemical formula $\mathrm{CaMg(CO_3)_2}$.

Clay Minerals

Clay is both a size of material and also a whole group of silicate minerals 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, $\mathrm{H_2CO_3}$. 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.

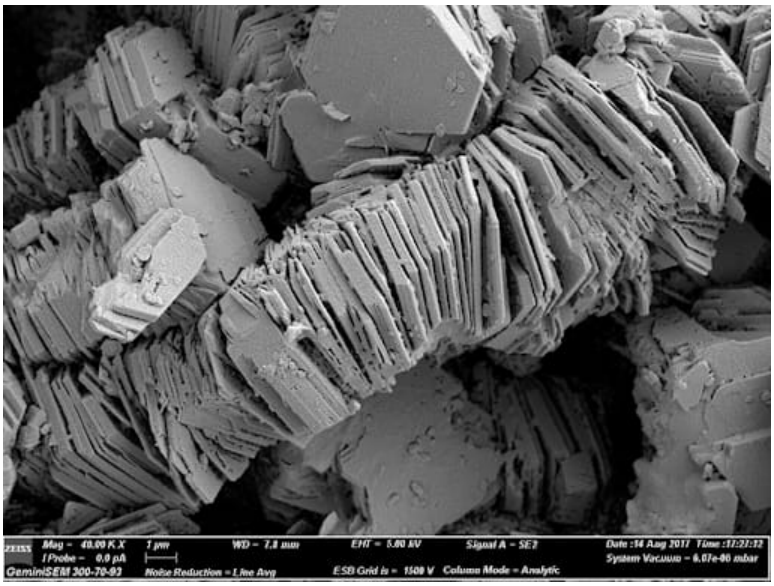


Figure 4.6.3: Scanning electron micrograph of the small sheets of the clay mineral kaolinite. Note the scale is in microns (µm). A micron is 1/1000 of a millimeter! Just to appreciate that scale, a human hair is approximately 75 microns in width! (CC Attribution 3.0; ACEMAC Nano Scale Electron Microscopy 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 are important to the study of geology because they are the main components of fine-grained sediment (mud) and form a very common sedimentary rock, shale.

Other Rock-Forming Silicates

The silicate minerals that fall into the group of “other” are largely formed 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 change in physical form and chemical composition as they recrystallize in response to these altered conditions.

One common mineral formed this way is garnet. Garnets are largely formed through the alteration 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 land and the formation of mountains.

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Figure 4.6.4: 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 4.6.1: 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 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).

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Figure 4.6.5: 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 4.6.6: The "Three Patriarchs" in Zion National Park, Utah. (Public domain, CC0 1.0; Daniel Mayer, Wikimedia)

Halides

The **halide** minerals contain the element chlorine ionically bonded with sodium or other cations. These minerals include halite (NaCl; table salt).

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Figure 4.6.7: 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 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.

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"), chalcopyrite (iron-copper sulfide), galena (lead sulfide), and 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.

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Figure 4.6.8: 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 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.

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Figure 4.6.9: 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.

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Figure 4.6.10: 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 4.6.11: 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.)

Key Terms

- **calcite** - the only "Big Ten" mineral that is not a silicate; composed of calcium, carbon, and oxygen; typically associated with sedimentary processes
- **clay** - a group of silicate minerals formed through the weathering of pre-existing silicate minerals; also a very small size of material
- **halides** - minerals consisting of chlorine bonded with sodium or other cations
- **native elements** - minerals that are composed of one element, existing in a pure or nearly pure state
- **oxides** - minerals consisting of metal ions covalently bonded with oxygen
- **phosphates** - minerals consisting of a $(PO_4)^{-2}$ cation bonded with various ions
- **sulfates** - minerals consisting of metal ions bonded to a sulfate ion
- **sulfides** - minerals consisting of metal ions bonded with sulfur

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4.7: Key Points

1. A mineral is a naturally occurring, inorganic solid with an ordered atomic structure and definite chemical composition.
2. Minerals are important as they are the building blocks of rocks which give us clues as to an area's geologic setting in the past.
3. The most important minerals in identifying Earth's common rocks are dubbed "the Big Ten" and are:
    1. olivine,
    2. augite,
    3. hornblende,
    4. biotite,
    5. calcium-rich plagioclase (anorthite),
    6. sodium-rich plagioclase (albite),
    7. potassium-rich feldspar (commonly orthoclase),
    8. muscovite,
    9. quartz, and
    10. calcite.
4. The silica tetrahedron is composed of one silicon atom and four oxygen atoms arranged in a tetrahedron. It serves as the framework for the silicate minerals which are the most common minerals in Earth's crust. The first nine of "the Big Ten" minerals are silicates.
5. Minerals form through cooling of molten rock, organic processes related to animal activity, precipitation from solution, and metamorphism.
6. Bowen's Reaction Series is a useful diagram depicting the order in which silicate minerals crystalize from a melt as molten material cools.

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5: EARTH MATERIALS - ROCKS

Introduction

There are three types of rocks, each composed of minerals formed in very different ways. Igneous rocks form as minerals crystallize when molten material cools. Metamorphic rocks form when elevated temperatures and pressures, plus the introduction of fluids, cause minerals to become unstable and change (or "morph") into new minerals without going through the melting phase. Sedimentary rocks are formed in several ways including through the collection, compaction, and cementation of sediment (loose material) and through the precipitation of mineral material from solution.

A foundational concept in Geology explores how these rock types are related. A diagram called the "Rock Cycle" shows these relationships as Earth materials are constantly in a state of change and one rock type turns into another.

5.1: Igneous Rocks

5.2: Metamorphic Rocks

5.3: Sediments and Sedimentary Rocks

5.4: The Rock Cycle

5.5: Key Points

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5.1: Igneous Rocks

Defined

Igneous rocks are those that form as molten material cools. When the molten material is found below the surface, we call it **magma**. The temperature of magma can range between about 800° and 1600 °C depending on where within the mantle the magma forms. Magma cools slowly within the crust (over centuries to millions of years)—forming **intrusive** igneous rock (also called **plutonic** rock), and crystallizes at depths of hundreds of meters to tens of kilometers below the surface. When molten material makes it to the surface, we call it **lava**. It cools quickly (within seconds to years)—forming **extrusive** igneous rock (also called **volcanic** rock).



Figure 5.1.1: Extrusive igneous rock. Basalt lava flow on Old Chain of Craters Road, Hawaii. (With permission by: Garry Hayes and Susan Hayes)

To summarize:

- Molten material below the surface is magma. It cools on the "interior" of the Earth it forms intrusive or plutonic rocks.
- Molten material on the surface is lava. It cools on the "exterior" of the Earth it forms extrusive or volcanic rocks.

Classification of Igneous Rocks

Igneous rocks are classified on two characteristics: composition and texture. The composition of igneous rocks is based on the minerals present. The texture refers to the size of the mineral grains.

Igneous Rock Composition

For the classification based on composition, we use Bowen's Reaction Series which outlines the crystallization sequence of silicate minerals forming from a cooling magma.

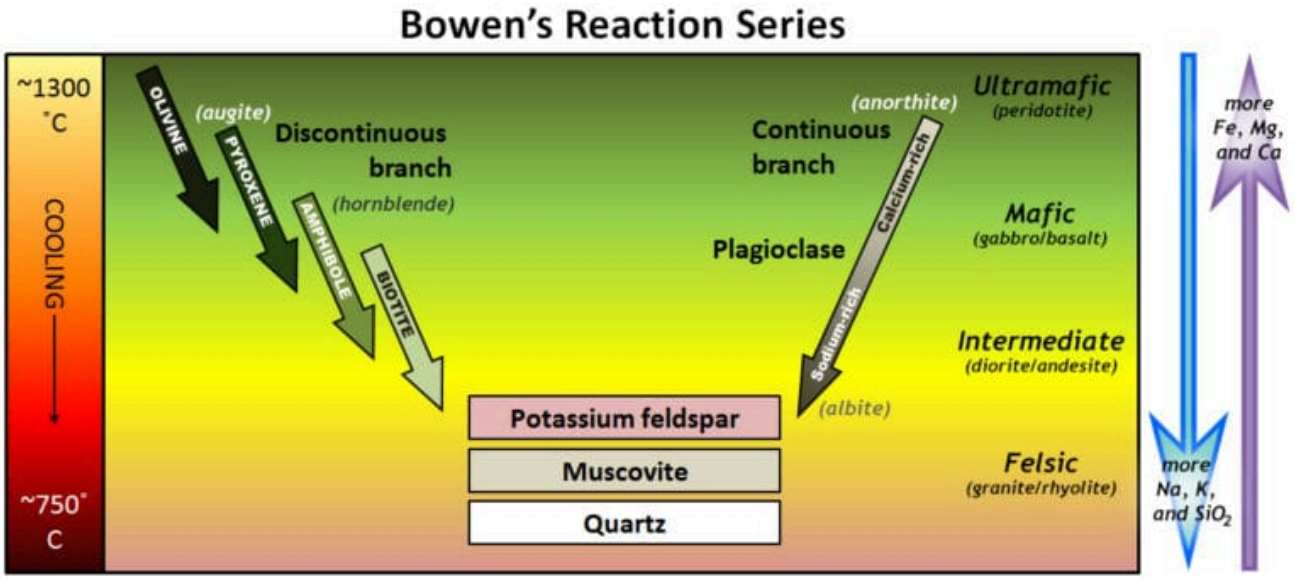


Figure 5.1.2: Bowen's Reaction Series. Developed through laboratory experimentation by Norman L. Bowen in the early 1900's. It establishes the order of crystallization of minerals from a silicate magma. (Creative Commons Attribution 4.0 International License; Modified after an original by Steven Earle)

As the magma cools from 1300° to 750°C (as noted on the left side of the diagram), minerals form in the order indicated from top to bottom. Looking at the left branch, olivine forms first, then pyroxene (augite), then amphibole, and so on and so forth until quartz forms. This is called the discontinuous branch because the minerals that form are unique silicate minerals. Looking at the right branch, note it is called the continuous branch. Here, different versions of one silicate (plagioclase) forms. At higher temperatures, the version of plagioclase with calcium crystallizes (anorthite). At lower temperatures, the version with sodium crystallizes (albite). In the middle of that series about half the spots are taken up with calcium, and half with sodium. As you move along the branch there are an infinite number of possible versions formed as the amount of calcium decreases, and sodium increases.

Igneous rock composition is divided into four categories: **ultramafic**, **mafic**, **intermediate** and **felsic**. You can see these labeled on the right side of Bowen's Reaction Series such that we read these igneous rock compositions as horizontal strips that run from right to left. 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 magnesium (Mg) and iron (Fe) and typically contain calcium-rich plagioclase plus varying amounts of olivine, pyroxene (augite) and amphibole (hornblende). Felsic refers to the overall composition being enriched in the feldspar minerals of sodium-rich plagioclase and potassium feldspar plus silica (SiO_2) in the form of quartz and typically contain sodium-rich plagioclase, biotite, potassium feldspar, muscovite and quartz.

Reading Bowen's Reaction Series above, we see that:

- 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 compositions with an abundance of dark iron- and magnesium-rich minerals of olivine and augite plus gray calcium-rich plagioclase feldspar. The rocks making up the oceanic crust are primarily mafic. 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 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.

Looking at the diagram below, you see the felsic rocks (granite and rhyolite) are light in color. As you move to the right toward ultramafic, the rocks get darker as the amount of iron and magnesium increases.

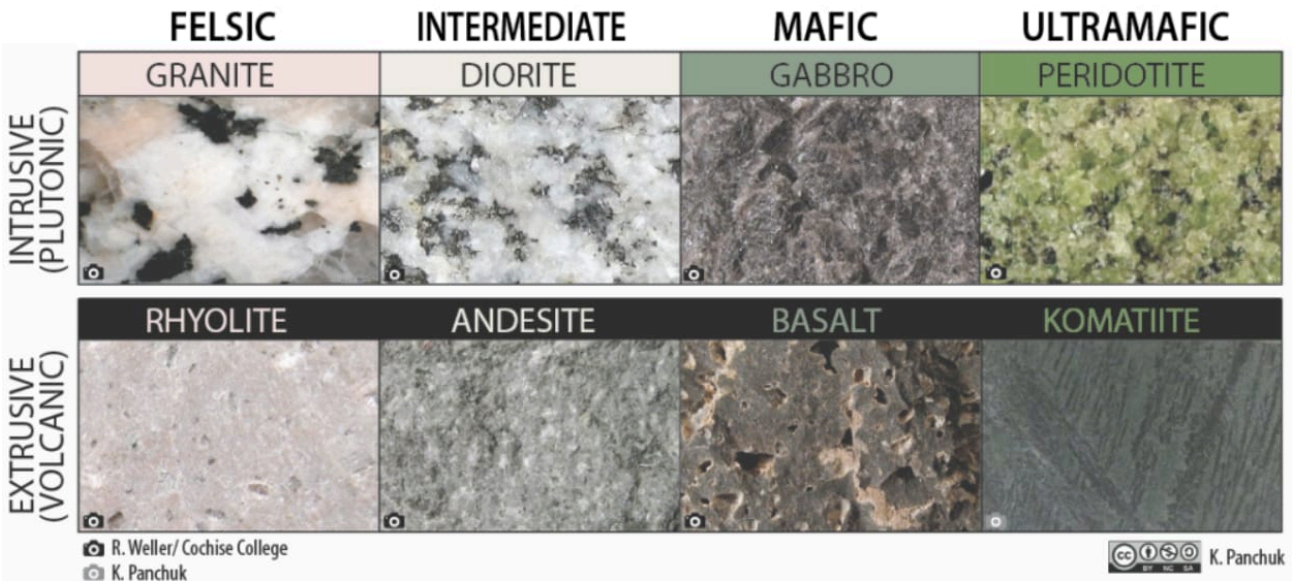


Figure 5.1.3: Igneous Rock Classification Diagram. The diagram is read in columns from the composition at the top to the example textural types along the side.

To summarize:

Compositionally, igneous rocks are categorized as follows:

- Ultramafic rocks are very low in silica and very high in iron and magnesium. They are very dark in color due to very high iron/magnesium.
- Mafic rocks are low in silica and high in iron and magnesium. They are dark in color due to high iron/magnesium.
- Intermediate rocks are medium in silica and medium in iron and magnesium. They are medium in color due to high iron/magnesium.
- Felsic rocks are high in silica and low in iron and magnesium. They are light in color due to low iron/magnesium.

As rocks contain more iron and magnesium, they get darker.

Igneous Rock Texture

If magma cools slowly, deep within the crust on the "interior" of the Earth, the 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 visible to the unaided eye.

If lava cools quickly, on the surface or on the "exterior" of the Earth, the rock is called extrusive or volcanic. The quick cooling process does not allow crystals to grow large, giving extrusive igneous rocks a fine-grained or **aphanitic** texture. The individual mineral crystals are too small to see with the unaided eye. These tiny crystals can be viewed using a special 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.

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Figure 5.1.4: 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.)

YouTube Video

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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.)

Naming Igneous Rocks

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 is either granite or rhyolite. If it is an intrusive (plutonic) rock, it will have coarse, visible grains and it is given the name **granite**. If it is an extrusive (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** and **komatiite** are the intrusive and extrusive names for ultramafic igneous rocks. Komatiite has been a rare rock because for most of Earth history, volcanic material that comes directly from the mantle is not common.

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Figure 5.1.5: 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.)

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Figure 5.1.6: 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.)

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Figure 5.1.7: 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.)

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Figure 5.1.8: 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.)

Key Terms

- **andesite** - an aphanitic (intrusive), intermediate igneous rock
- **basalt** - an aphanitic (intrusive), mafic igneous rock
- **diorite** - a phaneritic (intrusive), intermediate igneous rock
- **extrusive** - igneous rock formed as molten material cools on the surface
- **felsic** - a composition in which there are high (~70%) concentrations of silica; resultant rocks are typically light in color
- **gabbro** - a phaneritic (intrusive), mafic igneous rock
- **granite** - a phaneritic (intrusive), felsic igneous rock
- **igneous rock** - a rock that forms as minerals crystallize when molten material cools

- **intermediate** - a composition in which there are moderate (~60%) concentrations of silica; resultant rocks are typically medium in color
- **intrusive** - igneous rock formed as molten material cools below the surface
- **komatiite** - an aphanitic (intrusive), ultramafic igneous roc
- **lava** - molten material on the surface
- **mafic** - a composition in which there are low (~50%) concentrations of silica; resultant rocks are typically dark in color
- **magma** - molten material below the surface
- **peridotite** - a phaneritic (intrusive), ultramafic igneous rock
- **plutonic** - igneous rock formed as molten material cools below the surface
- **rhyolite** - an aphanitic (intrusive), felsic igneous rock
- **volcanic** - igneous rock formed as molten material cools on the surface
- **ultramafic** - a composition in which there are very low (~40%) concentrations of silica; resultant rocks are typically very dark in color

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5.2: Metamorphic Rocks

Defined

Metamorphic rocks form when one rock changes, or "morphs", into another rock without going through the melt phase but instead through the recrystallization of minerals. This occurs when a body of rock is subjected to conditions that are different from those in which it formed. We call this process, metamorphism.

In most cases 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.

Types of Metamorphism

There are two primary environments that produce metamorphism that are important in geology:

Regional metamorphism 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 lithosphere (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 and result in the baking of the surface rock through the volcanic eruption of lava.

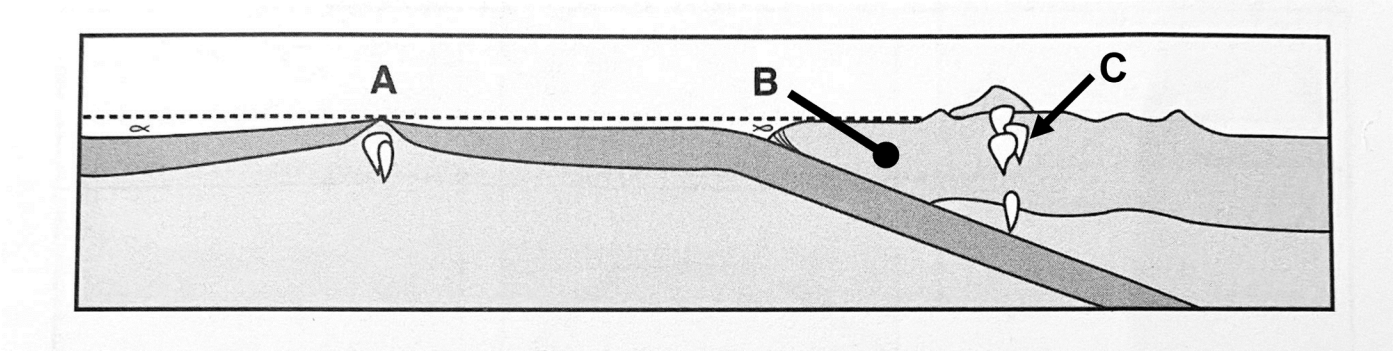


Figure 5.2.1: Illustration of two plate tectonic boundaries. At Site B, regional metamorphism would occur as a result of the elevated temperatures and pressures associated with the subduction of oceanic crust at a convergent boundary. Rock in that entire area becomes metamorphic rock. At Site C, contact metamorphism occurs as a result of the elevated temperatures associated with the cooling of magma bodies. The rock around the magma body is what becomes the metamorphic rock.

Metamorphic Processes

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

- the mineral composition of the protolith (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.

Protolith (Parent Rock)

The protolith can be any type of rock 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 or stress and directed stress. **Stress** is a scientific term indicating a force. **Strain** is the result of this stress, including metamorphic changes within minerals.

Confining Pressure

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

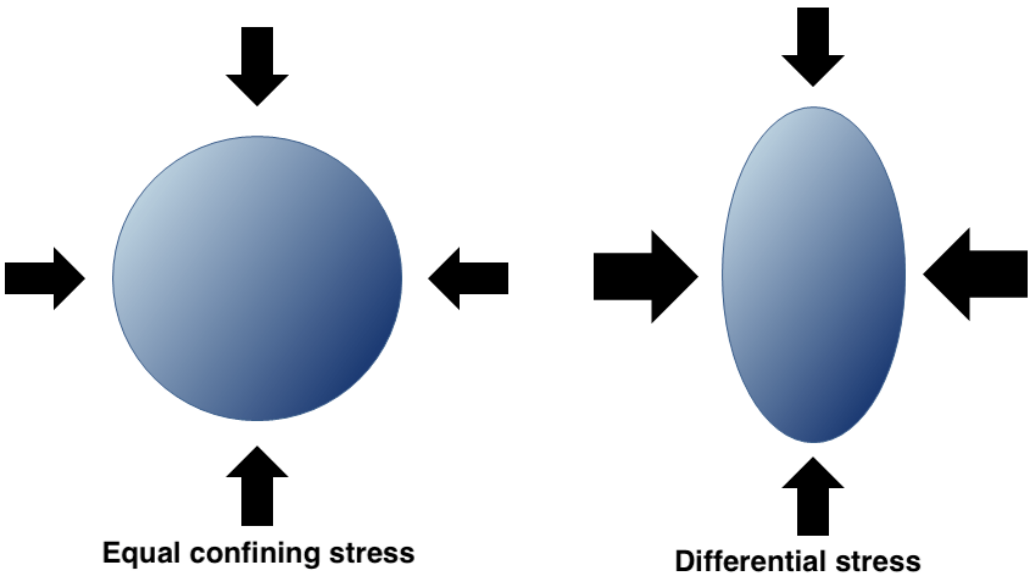


Figure 5.2.2: An illustration of different types of pressure on rocks. Confining pressure, where the pressure is essentially equal in all directions, and differential stress, where the pressure from the sides is greater than that from the top and bottom. In both diagrams, there is also pressure in and out of the page. (CC BY; Steven Earle.)

Differential Stress

Differential stress is an unequal balance of forces on the protolith in one or more directions (see figure above). Differential stress is most commonly associated with the tectonic movement of plates during mountain building. 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 the repetitive layering found in metamorphic rock. Sometimes foliation results in the banding you see in the figure below. Sometimes it is more subtle and results in the rock breaking into flat sheets as you will see further below.

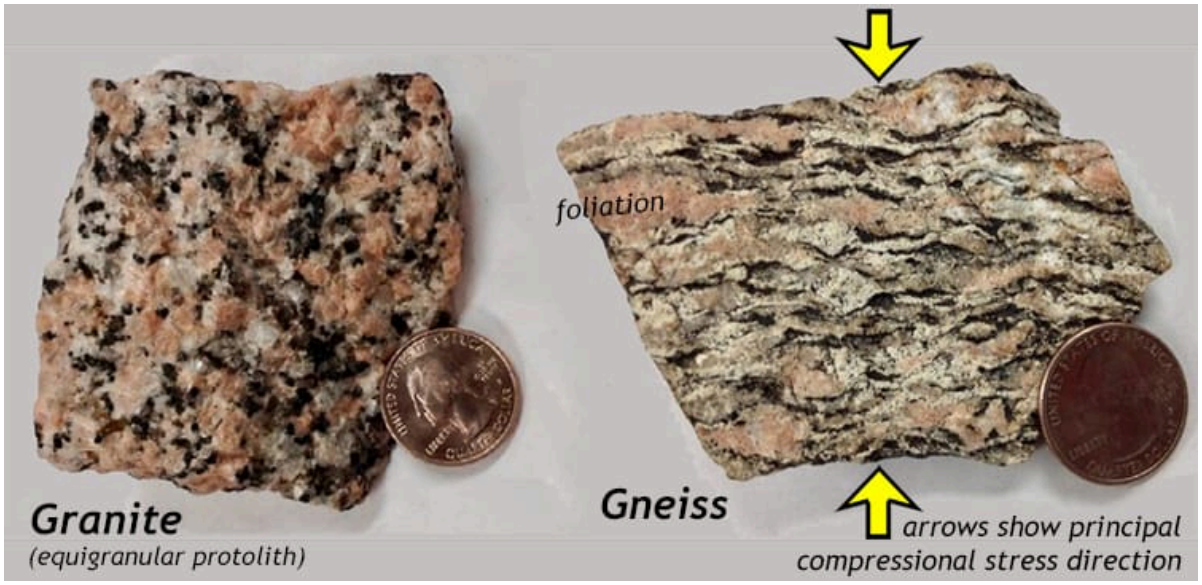


Figure 5.2.3: A comparison of granite and gneiss. The texture of the protolith granite is the same in every direction (equigranular) but the gneiss has a pronounced foliation (or preferred orientation), which we infer to have developed from squeezing (compressional stress) in a direction that was perpendicular to the resulting foliation. (Peter Davis. Modified by Callan Bentley.)

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).

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Figure 5.2.4: Laminated micrite/limestone (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 millimeter 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 the first characteristic observed in the identification process. There are two main types of metamorphic rocks, differentiated by texture:

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. Simply put, it shows a preferred orientation.

Non-foliated: either not subjected to directed pressure or, the dominant mineral does not tend to display any alignment. Simply put, it does not show a preferred orientation.

Foliated Metamorphic Rocks



Figure 5.2.5: 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 5.2.6: A slate boulder on the side of Mt. Wapta in the Rockies near Field, BC. Bedding of the protolith 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 immediately above). 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 (or a small amount of) 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.

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Figure 5.2.7: 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 is a metamorphic that displays a “schistose” 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 like biotite and muscovite) or garnet mica schist (mica schist with garnets).

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Figure 5.2.8: 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 is a metamorphic rock with mineral banding that 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. 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 or pockets within the migmatite.

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Figure 5.2.9: (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.

Quartzite and Marble

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Figure 5.2.10: 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.

Other Non-foliated Rocks

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.

Figure 5.2.11: Amphibolite (CC BY Attribution 3.0; Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

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 right to get a feel for the metamorphic fabric and overall green color of the rock.

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Figure 5.2.12: 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.)

Metamorphic Rock Identification Chart

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

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

Key Terms

- **amphibolite** - non-foliated metamorphic rock made of amphibole (hornblende) formed from a mafic igneous protolith
- **confining pressure** - pressure coming from all directions equally

- **contact metamorphism** - metamorphism that occurs in rock exposed to high temperatures resulting from the proximity to a magma body or lava flow
- **differential stress** - an unequal balance of forces in one or more directions
- **foliated/foliation** - the characteristic of a metamorphic rock by which there is a distinct alignment of minerals
- **gneiss** - a foliated metamorphic rock having minerals separated into light and dark bands
- **greenstone** - non-foliated metamorphic rock made of chlorite formed from a basalt protolith
- **marble** - non-foliated metamorphic rock made of calcite or dolomite formed as limestone was metamorphosed
- **metamorphic rock** - rock formed when elevated temperatures and pressures cause minerals to change into new minerals without going through the melting phase
- **migmatite** - a metamorphic rock that is partially melted showing the transition from metamorphic and igneous activity
- **non-foliated** - the characteristic of a metamorphic rock by which there is no alignment of minerals
- **phyllite** - 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
- **porphyroblast** - a large crystal of a particular mineral in a metamorphic rock surrounded by small grains
- **protolith** - the parent rock or the rock from which the metamorphic rock forms
- **quartzite** - non-foliated metamorphic rock made of quartz formed as quartz sandstone was metamorphosed
- **regional metamorphism** - metamorphism that occurs over a wide area resulting from the collision and compression of tectonic plates
- **schist** - a foliated metamorphic rock where minerals are aligned and visible as individual crystals
- **slate** - a fine-grained foliated metamorphic rock which breaks into flat sheets
- **strain** - the changes in a object that occur as a result of the stress applied
- **stress** - force applied to an object

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5.3: Sediments and Sedimentary Rocks

Sedimentary rocks are arguably the most important to study and understand in this class as they contain a level of detail about Earth's history not always found in other rock types. They are also have the widest variation in their formation such that we classify them into four different types:

- clastic sedimentary rock (detrital)
- chemical sedimentary rock
- biochemical sedimentary rock
- organic sedimentary rock

Weathering

In most cases the process of sedimentary rock formation begins with the exposure of pre-existing rock to weathering and erosion. **Weathering** involves exposure of rock to Earth's surface weather conditions. Wind, water and ice, plus gravity, play a role in the breakdown of the materials. **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 the process of physical disintegration: the breaking of solid rock into smaller pieces without changing its chemical composition, primarily caused by external forces such as temperature fluctuations, ice expansion, or abrasion by wind and water.



Figure 5.3.1: Boulder of igneous rock from Iceland displaying the effects of physical weathering. (Bridget Wade)

One example of physical weathering occurs when water in cracks in rock freezes and expands. This expansion places stress on the rock, and it breaks into smaller pieces. This process exposes an ever-increasing number of surfaces to additional weathering.

Chemical Weathering

Chemical weathering alters the chemistry of minerals in the pre-existing rock. These changes may result in new mineral formation by adding or subtracting elements from the original mineral structure, or in the further disintegration of pre-existing minerals through chemical decomposition.



Figure 5.3.2: 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)

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 exposed minerals and physically removing weathered rock material and ions lost through decomposition.

Products of Weathering

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 of sediment together.
- 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 are discussed in more detail elsewhere in this text. The simple classification of sedimentary rock is presented below.

Clastic Sedimentary Rocks (Detrital)

Clastic sedimentary rock (also called 'detrital' sedimentary rock) consists of sediment pieces (clasts/rock detritus) that come from weathered (broken down) rock, cemented together to make a new rock. A more comprehensive explanation of this process is represented in the diagram immediately below in which source rock is weathered to create sediment, eroded (removed from its original location), transported (carried by wind, water, or ice), deposited (its journey ends), buried under other sediment such that it is exposed to small amounts of pressure and temperature resulting in the formation of the sedimentary rock.

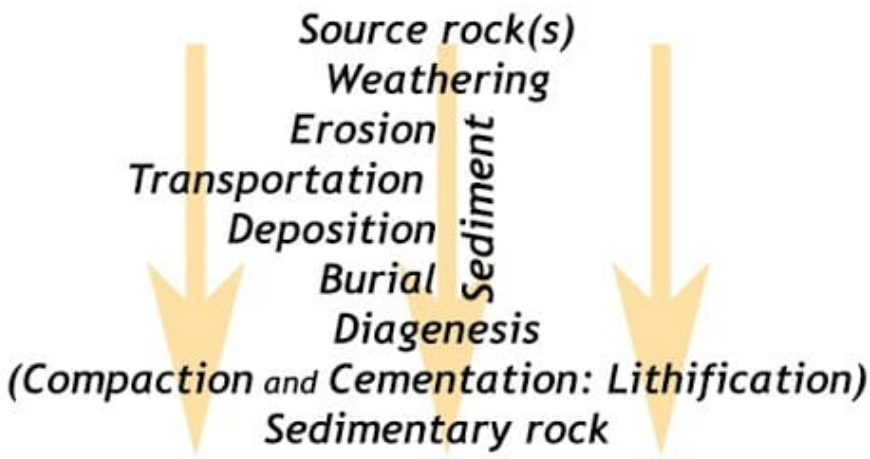


Figure 5.3.3: The steps in the Sedimentary Rock Cycle. (Callan Bentley (2020))

Clastic (or detrital) rock is classified and named based on the size and composition of the clasts. Their various sizes, from boulders (large) to sand to clay (very small), is referred to as the sediment grain size. The classification and description of the various clastic rocks appears in the chart at the bottom of this page. 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.

Common Clastic Sedimentary Rocks

Conglomerate is a coarse-grained clastic sedimentary rock composed of a substantial amount of rounded gravel-size (larger than 2 mm in diameter) clasts. Conglomerates form by the lithification (compaction and cementation) of gravel. Conglomerates are typically poorly sorted (contain many different grain sizes), having finer-grained sediment, e.g., either sand, silt, and/or clay in between the gravels.

Figure 5.3.4: Conglomerate. *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.*

Breccia is a coarse-grained clastic sedimentary rock composed of a substantial amount of angular gravel-size (larger than 2 mm in diameter) clasts. Breccias are typically poorly sorted (contain many different grain sizes), having finer-grained sediment, e.g., either sand, silt, and/or clay in between the gravels.

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Quartz sandstone is a clastic sedimentary rock composed mainly of sand-sized (1/16 to 2 mm in diameter) quartz mineral grains. They are most commonly white or red in color. The red sandstones have iron oxide (essentially rust) cementing the quartz grains together. The shape of the sand grains (rounded or angular) and the sorting (all one size or variable in size) varies and helps geologists determine in what geologic setting (beach vs sand dune vs river) the sand was collected.

Figure 5.3.6: Quartz Sandstone. *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 is a clastic sedimentary rock composed mainly of sand-sized (1/16 to 2 mm in diameter) feldspar mineral grains. Pink (potassium feldspar) and white (plagioclase feldspar) are most common. The sand grains tend to be angular and they tend to be poorly sorted (contain many different grain sizes).

Figure 5.3.7: Arkose Sandstone. *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 is a clastic sedimentary rock that is a variety of sandstone. It is generally characterized by its dark color, and poorly sorted (contain many different grain sizes), angular grains of quartz, feldspar, and small rock fragments set in a matrix of clay.

Figure 5.3.8: Greywacke. *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 is a clastic sedimentary rock primarily composed of silt-sized (1/16-1/256 mm in diameter) particles.

Figure 5.3.9: Siltstone. *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 in the sample above), which is the most common sedimentary rock, is a clastic sedimentary rock primarily composed of silt- and clay-sized (< 1/16 mm) minerals. Shale is characterized by breaks along parallel layering or bedding less than one centimeter in thickness, called fissility.

Figure 5.3.10: Shale. *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 is 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 as it does not break easily into layers like shale.



Figure 5.3.11: Claystone. *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 Rocks

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).

Common Chemical, Biochemical and Organic Sedimentary Rocks

Fossiliferous limestone is any type of limestone, made mostly of calcium carbonate ($\mathrm{CaCO_3}$) in the form of the mineral 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.

Figure 5.3.12: Fossiliferous Limestone. *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: 3.0 Attribution.*

Oolitic Limestone is a type of limestone formed from ooids, small, spherical grains composed internally of calcium carbonate ($\mathrm{CaCO_3}$ – calcite) in concentric layers.

Figure 5.3.13: Oolitic Limestone. *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: 3.0 Attribution.*

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.



Figure 5.3.14: Micrite.
"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: 3.0 Attribution.

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.

Figure 5.3.15: Chalk. *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: 3.0 Attribution.*

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, below. Grab and rotate the sample to see it in 3D). Travertine can be white, tan, cream-colored, or somewhat rusty in color.

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Figure 5.3.16: **Speleothem** by **Marissa Dudek** on **Sketchfab**

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.



Figure 5.3.17: Crystalline Limestone. *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: 3.0 Attribution.*

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 (whereas limestones effervesce without being powdered).



Figure 5.3.18: Dolostone. *Photo credit: "Dolostone (Silurian; midwestern USA)" by James St. John is licensed under CC BY 2.0. Definition credit: Wikipedia licensed under CC BY: 3.0 Attribution.*

Chert is a hard, fine-grained sedimentary rock composed of crystals of quartz (silica $\mathrm{SiO_2}$) 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.



Figure 5.3.19: Chert. *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 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.



Figure 5.3.20: Rock Gypsum. *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.

Figure 5.3.21: Rock Salt. *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 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. Consequently, there is no such thing as “clean coal".



Figure 5.3.22: Bituminous 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: 3.0 Attribution.*

Sedimentary Rock Identification Chart

Table 5.3.1: 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.	Gravel (> 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 banding of browns, grays, whites, and blacks.	Travertine

		Coarse crystals easily visible.	Crystalline Limestone
	Dolomite	Fizzes with acid only when powdered; fine to coarsely crystalline; may contain fossils.	Dolostone
	Microcrystalline Quartz	Microcrystalline[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.

Key Terms

- **arkose sandstone** - clastic sedimentary rock composed of sand-sized feldspar particles
- **biochemical sedimentary rock** - sedimentary rock formed as shells and the body parts of underwater organisms lithify
- **bituminous coal** - sedimentary rock formed by the breakdown of peat bog material
- **breccia** - poorly sorted clastic sedimentary rock composed of angular gravel-sized particles
- **chemical sedimentary rock** - sedimentary rock formed as minerals precipitate from ions dissolved in solution
- **chalk** - sedimentary rock made of very small shells made of calcite
- **chert** - sedimentary rock made of very small crystals of quartz
- **clastic sedimentary rock** - sedimentary rock composed of sediment compacted and cemented to form a new rock
- **claystone** - clastic sedimentary rock composed of clay-sized particles
- **conglomerate** - poorly sorted clastic sedimentary rock composed of rounded gravel-sized particles
- **crystalline limestone** - a sedimentary rock formed as calcite precipitates
- **dolostone** - a sedimentary rocks formed as magnesium replaces the calcium in limestone before lithification
- **erosion** - the removal of weathered rock material from its original location

- **fossiliferous limestone** - sedimentary rock composed of calcite that contains an abundance of fossils
- **greywacke** - clastic sedimentary rock composed of sand-sized particles that vary in their composition
- **micrite** - sedimentary rock made of very fine grains of calcite
- **oolitic limestone** - sedimentary rock formed from small grains of calcite
- **organic sedimentary rock** - sedimentary rock formed as the remains of organic materials undergo lithification
- **quartz sandstone** - clastic sedimentary rock composed of sand-sized quartz particles
- **rock gypsum** - sedimentary rock made of gypsum
- **rock salt** - sedimentary rock made of halite
- **shale** - clastic sedimentary rock composed of silt- and clay-sized particles that breaks into sheets
- **siltstone** - clastic sedimentary rock composed of silt-sized particles
- **travertine** - form of limestone deposited by mineral springs
- **weathering** - breakdown of materials due to exposure to the atmosphere and water

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5.4: 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.

As paths through the rock cycle are described in the text below, trace where in the rock cycle (depicted in the figure below) the material being described is.

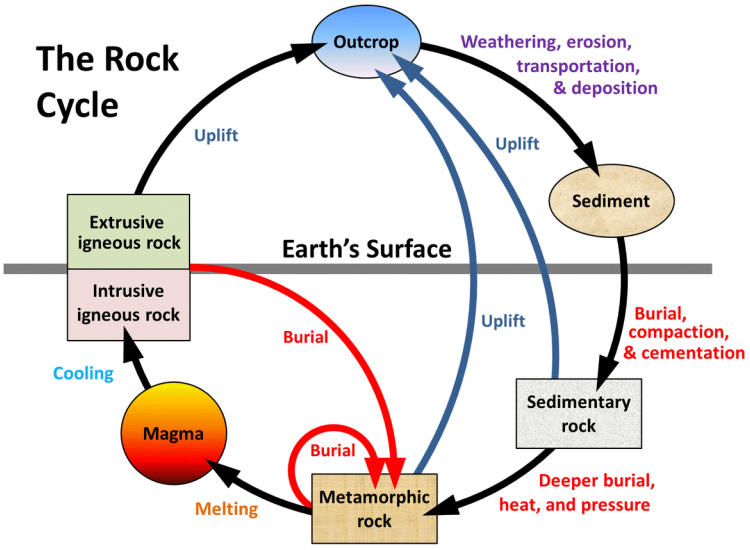


Figure 5.4.1: The Rock Cycle. (Creative Commons Attribution 4.0 International License; Steven Earle)

In describing the rock cycle, while you can begin anywhere since the cycle is continuous, let's start with magma because from magma, new minerals and igneous rock are created. 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).

Interactive Content

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Figure 5.4.2: Granite, an example of an igneous rock that cooled from magma. Zoom in close to notice the interlocking fabric of minerals crystals of different size and shape. (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.

Interactive Content

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Figure 5.4.3: 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. (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.

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Figure 5.4.4: 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. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Key Terms

- **rock cycle** - a diagram that describes the processes through which the three main types of rocks – igneous, metamorphic, and sedimentary – are created, transformed, and recycled over geological time

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5.5: Key Points

1. Igneous rock forms as magma cools below the surface to form coarse-grained (phaneritic) intrusive rocks or lava cools on the surface to form fine-grained (aphanitic) extrusive rocks.
2. Igneous rocks are classified based on composition and texture.
3. Igneous rock compositions are ultramafic (very low amounts of silica ~40%) which results in very dark-colored rocks, mafic (low amounts of silica ~50%) which results in dark-colored rocks, intermediate (medium amounts of silica ~60%) which results in medium-colored rocks, and felsic (high amounts of silica ~70%) which results in light-colored rocks.
4. Metamorphic rock forms as existing rock (a protolith) is exposed to elevated temperature and pressure resulting in a recrystallization of the minerals without going through the melt phase.
5. The two types of metamorphism are regional metamorphism in which a large area is affected as two plates collide and contact metamorphism in which rocks is exposed to high temperatures adjacent to magma bodies or volcanic eruptions.
6. The type of metamorphic rock that forms depends on the composition of the protolith, the temperature, pressure (confining pressure or directed stress), and fluids the protolith is exposed to, and time.
   Metamorphic rocks are either foliated (show a preferred orientation) or non-foliated (do not show a preferred orientation).
7. There are two types of weathering. Physical weathering causes a reduction in size through mechanical interactions. Chemical weathering causes a change in composition due to exposure to the atmosphere and water.
8. The products of weathering are sediment and ions in solution.
9. Clastic sedimentary rock is formed as sediment is compacted and cemented together and is classified based on the size and composition of the clasts.
10. Chemical sedimentary rock is formed as minerals precipitate from ions in solution.
11. Biochemical sedimentary rock is formed as shells collect and are cemented together.
12. Organic sedimentary rock is formed as organic material lithifies.
13. The rock cycle is used to visualize how the components of the crust are constantly being changed from one form into another through various processes such as weathering, erosion, transportation, uplift, cooling, burial, compaction, cementation, melting, etc.

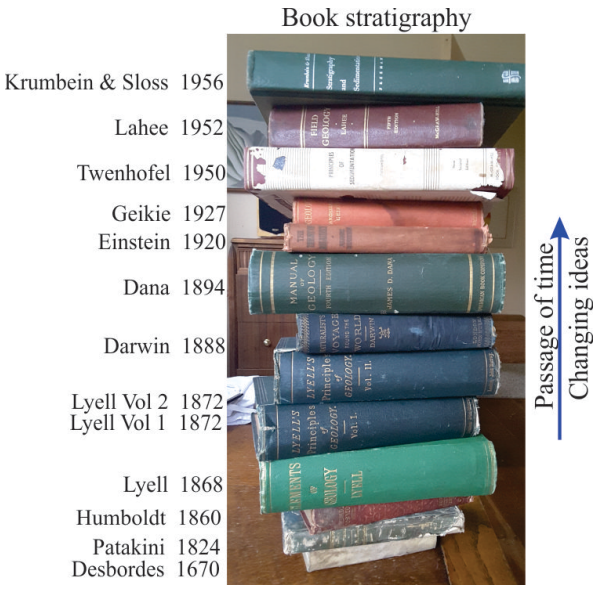
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6: SEDIMENTOLOGY AND STRATIGRAPHY

Introduction

> "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



Rocks to a geologist are like books and ancient manuscripts to a historian as they contain a record of past events, environments, climates, and life. But the rock record is incomplete, like a book with missing chapters or pages. In order to reconstruct the history of the Earth from the rock record, geologists have to try and figure out what is missing. Is it time? How much time? Have some of the rocks been eroded and 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. 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?

6.1: Sediment Deposition and Bedding

6.2: Other Sedimentary Structures

6.3: Fossils and Fossilization

6.4: Facies

6.5: Depositional Environments

6.6: Introduction to Strata and Basins
6.7: Sequence Stratigraphy and Walther's Law
6.8: Stratigraphic Correlation
6.9: Key Points
6.10: Further Reading

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6.1: Sediment Deposition and Bedding

Sedimentary Structures

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. Let's look at some common sedimentary structures that provide insight into how sediments were transported across a substrate and deposited.

Bedding

Layered - or stratified - rocks are organized into **beds**: centimeter- to meter-thick units of sedimentary rock that were deposited approximately horizontally and are separated by approximately horizontal planes (bedding planes); the rocks typically weather more along these planes. Beds are usually fairly uniform or change gradationally in composition. Bedding planes usually represent breaks in sedimentation or changes in grain size. In other words, they usually represent changes in flow characteristics. (from https://geo.libretexts.org/Courses/U...06%3A_Bedforms)

Bedding is therefore a sedimentary structure. Sediment moved by currents or falling from suspension through water or air will accumulate in a bed; this process will continue in a depositional environment until the supply of sediment is terminated. The 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.



Figure 6.1.1: Example of beds and bedding planes in stratified rocks, southern Turkey. (Bedding by Ykr300, public domain.)

The form of a sedimentary bed can be observed at the outcrop level. In some cases, individual layers may maintain their thickness across the entire outcrop, creating **tabular beds**, or parallel beds. This type of bedding is common when sediment supplies are constant and the depositional environment is laterally pervasive. Keep in mind that descriptive words like tabular or lenticular bedding depend on the scale of your observations. For example, beds that are tabular at the outcrop scale, may "pinch out" or thicken in exposures farther afield.



Figure 6.1.2: Tabular bedding can form in many environments. This example is from a turbidite succession but parallel beds can form in almost any sediment type, from conglomerate to mudstone; siliciclastic, carbonate, volcaniclastic. (Brian Ricketts.)

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.



Figure 6.1.3: Lenticular bedding is common in many environments. In this example of a channel sandstone from West Virginia, note how the thickest light tan layer in the center of the image thins and pinches out into the foreground and background. It can form in any sediment type, from conglomerate to mudstone; siliciclastic, carbonate, volcaniclastic. (K. Layou.)

Ripples and Dunes



Figure 6.1.4: 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.

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 are referred to as "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
- There is a critical flow velocity at which grain movement will begin; this 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.

Most of our knowledge about bedforms and how they form has been garnered from studies of modern environments. After all, if on your walks across a tidal flat or desert 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 Principle of Uniformitarianism in action!).

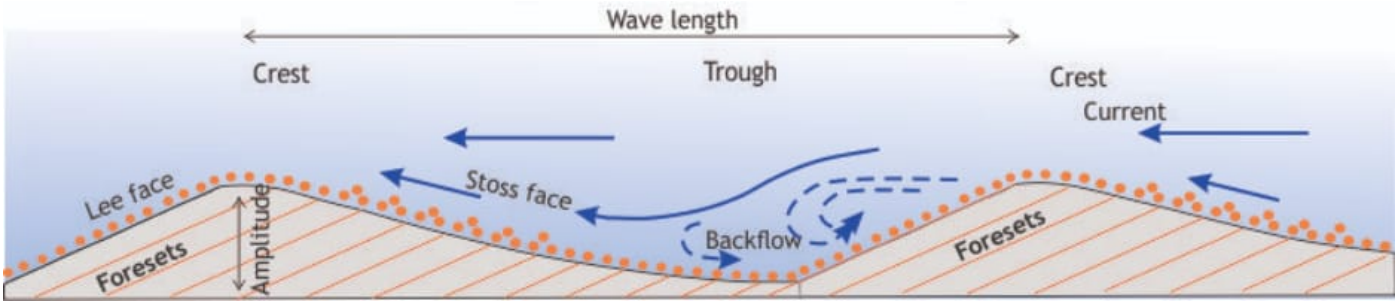


Figure 6.1.5: Ripple formation and terminology. The current moves sand grains up the stoss slope. At the crest of the ripple the grains enter a region of low velocity backflow where gravity takes over and the grains tumble or avalanche down the lee face. In this fashion the ripple moves down-stream with the current.

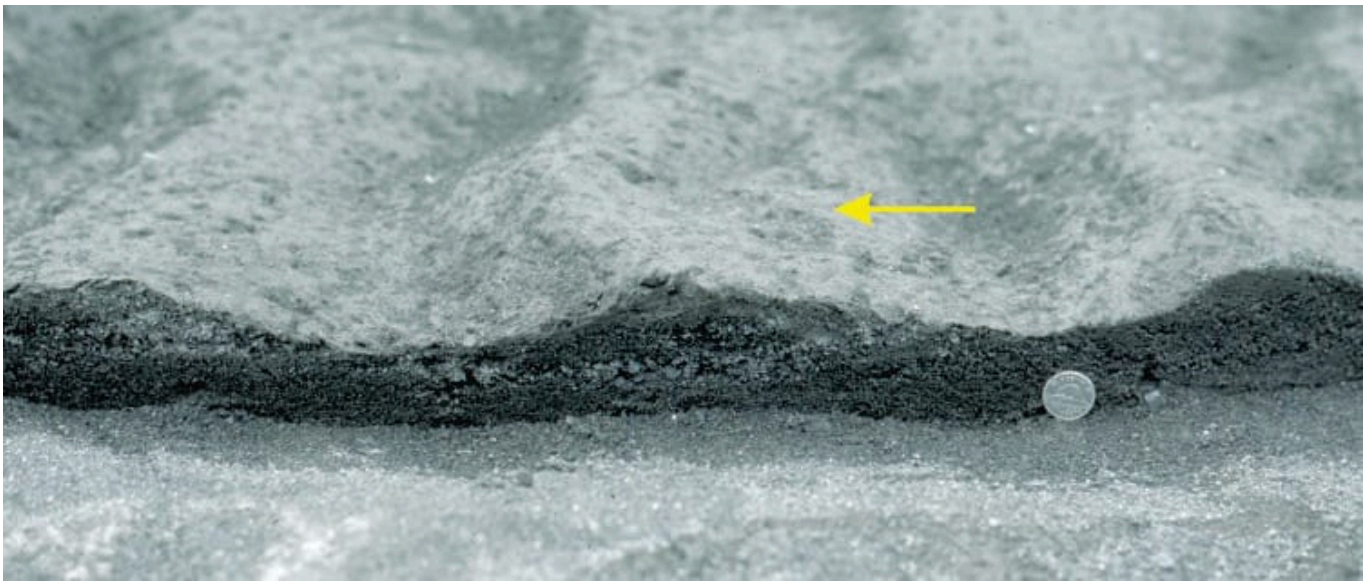


Figure 6.1.6: A slice through modern ripples in soft river sand. Current flow indicated. Compare the example with the diagram above for details of terminology and mechanism of formation. (Brian Ricketts.)

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. Ripples can be asymmetrical, formed from a unidirectional current (like a flowing river) or symmetrical, formed from oscillating currents (like waves). Figure 6.1.6 above shows an asymmetrical ripple form, with the right side of the ripples more gently sloped than the steeper right sides; this indicates that the right side of the ripples are upstream (stoss) while the left side of the ripples are downstream (leeward).



Figure 6.1.7: 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. (Brian Ricketts.)

Laminae and Crossbedding

Laminae are color, composition, or grain size variations defining surfaces within a bed. They typically represent variations in flow velocity, sediment supply, sediment composition, etc. Planar laminae are parallel to bedding: planar and deposited approximately horizontally.

Cross lamination, cross stratification, or crossbedding are laminations or layers that are oriented obliquely to bedding. They

truncate older laminae and are truncated by younger laminae. The erosional surfaces that separate sets of similarly oriented laminae are called “bounding surfaces”. There are lots of subdivisions of cross stratification; different types represent different types of bedforms and different flow conditions.

(from https://geo.libretexts.org/Courses/U...06%3A_Bedforms)



Figure 6.1.8: Sandstone dome in the Dry Fork of Coyote Gulch, park of the Canyons of the Escalante in Southern Utah. This dome exhibits crossbedding, remnant of its origins as sand dunes. (G. Thomas)

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).

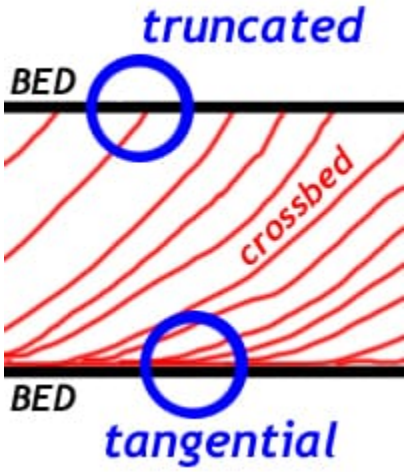


Figure 6.1.9: A cartoon cross-section showing the different crossbed/bed relationships at the upper and lower portion of the bed. (Image by Callan Bentley (2020).)

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.

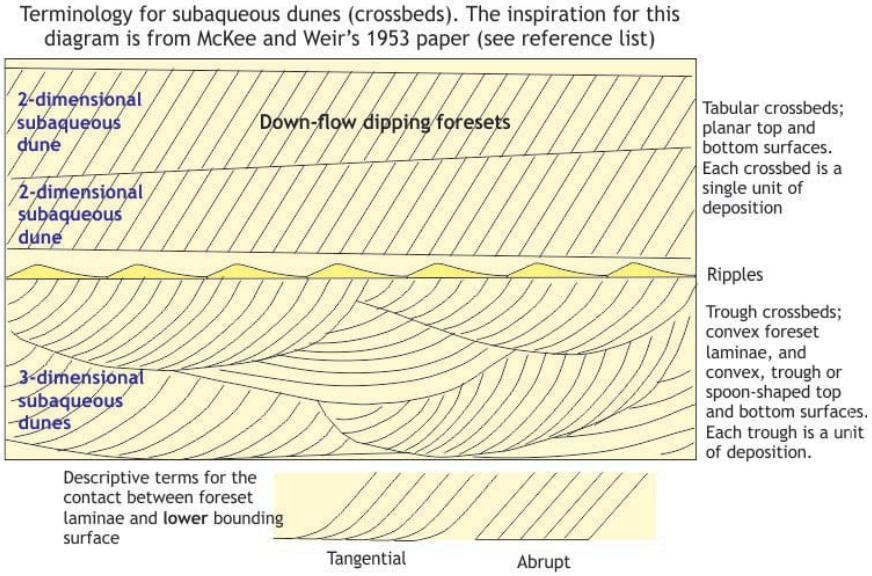


Figure 6.1.10: Terminology for common forms of subaqueous crossbeds (ripples, dunes). (The inspiration for this Brian Ricketts diagram is from McKee and Weir, 1953.)

Laminae in crossbedding 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 foresets in Figure 6.1.11 below are indicated by dashed yellow lines.

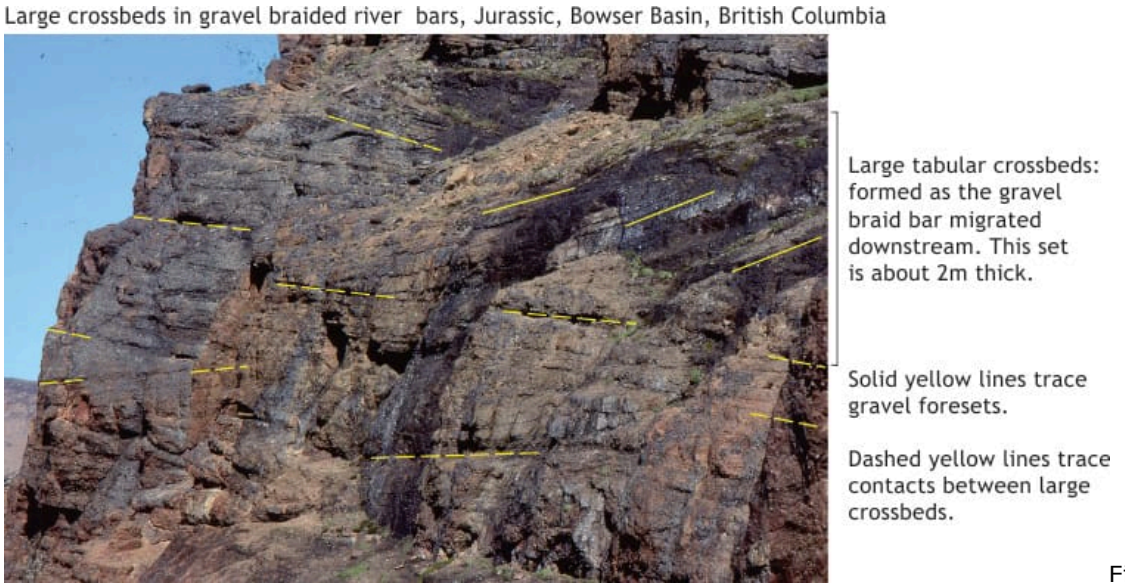


Figure 6.1.11: Tabular crossbeds up to 2m thick. The crossbeds formed as part of large gravel bars within a Jurassic river channel. (Brian Ricketts.)

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 taking into account other criteria - like body

and trace fossils, sediment composition, and stratigraphic trends - will help us pinpoint specific depositional settings.

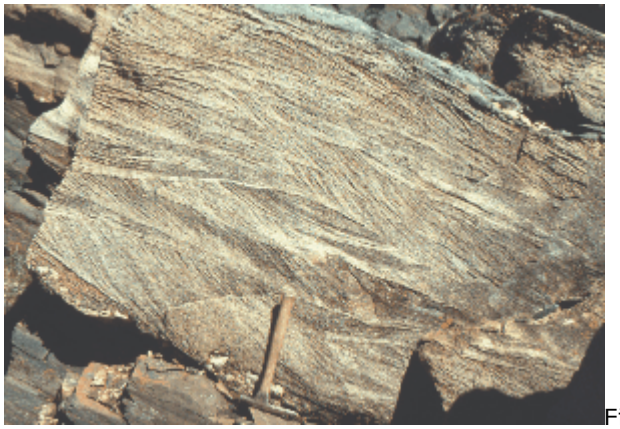

Figure 6.1.12: Trough crossbeds from an ancient tidal channel. In 2D exposures, trough crossbeds have convex-up basal contacts and foresets tend to be tangential. Troughs commonly cross-cut earlier-formed structures, resulting in a complex array of partly preserved bedforms. Common in channelized (confined) flow; found in tidal, estuarine, fluvial, and alluvial fan channels. (Brian Ricketts.)


Figure 6.1.13: 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 way-up indicator. (Callan Bentley.)

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).

Dune-Ripple Formation

YouTube Video

This interactive element is not available in this format. Visit the link or scan the QR code to access it online.

View this YouTube Video online

The above video shows subaqueous dunes forming in a flume (movie presented by Michael Calzi, SUNY Geneseo Dept. of Geological Science). 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 modeling engineering problems such as the flow of water around bridge foundations. The sequence begins by showing water flow in front of the migrating lee face; look carefully and you will notice:

- 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 (Figure 6.1.10).

Key Terms

- **bed** - a layer of sediment or sedimentary rock that is distinguishable from adjacent layers due to differences in its composition, texture, or other physical properties
- **bedform** - a geological feature developed when a moveable bed interfaces with a fluid resulting in the material being shaped by the flow
- **cross-bed** - a sedimentary structure where layers of sediment are deposited at an angle to the main, horizontal bedding plane
- **discontinuous bed** - a sediment ary bed that pinches out in one direction
- **foreset** - an inclined layer of sediment in a bedform that dips in the direction of current flow, forming the steep part of the structure
- **lenticular bed** - a sedimentary bed that pinches out in both directions
- **tabular bed** - a broad continuous layer of sedimentary material characterized by a consistent thickness and planar bounding surfaces

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6.2: Other Sedimentary Structures

In the previous section, we examined common descriptive categories of bedding and bedforms that inform us about current flow and sediment deposition. Identifying sedimentary structures and deciphering how they formed are two of the most important tasks for any sedimentological study. They provide important clues to interpreting ancient environments.

Describing Stratified Rocks

The utility of sedimentary structures to unravel the past becomes even more powerful when used in conjunction with other rock properties, like sediment texture, fabric, color, and fossil content. Basic steps for describing stratified rocks - at the outcrop, core, or hand sample scale - are listed below.

1. Examine the **bedding**.
    1. How thick are the beds?
    2. What is the bed geometry (parallel, lenticular, lensoidal)?
    3. How do the beds weather? Are the rocks more or less resistant to weathering, forming cliffs or recesses?
2. Examine the color of the strata. Is it uniform or variable?
3. Examine the **texture** and **fabric** of the rocks.
    1. What is the maximum grain size?
    2. What is the range of grain sizes (i.e. sorting of clasts)?
    3. What is the clast shape (round vs. angular, platy vs. spherical)?
    4. What is the clast framework (clast-supported vs. matrix-supported)?
    5. Is there a preferred clast orientation (alignment/imbrication)?
    6. Is there a preferred clast distribution (e.g. graded, bimodal, etc.)?
    7. What is the rock classification (i.e. mudrock, sandstone, conglomerate/breccia)?
4. Examine the **composition** of the rocks.
    1. Is it a carbonate, siliciclastic, or volcaniclastic?
    2. What are the main clast types (quartz, feldspar, lithics, bioclasts, etc.)?
    3. What is the cement (silica, carbonate, iron oxide)?
    4. How hard is the rock vs. how easily does it break apart (i.e. induration: does it ring when you hit it with a hammer or does it crumble)?
5. Examine the **fossil** content of the rocks.
    1. Are there body fossils, trace fossils, cast or molds, etc.?
    2. Are there any preferred faunal or sedimentary associations?
    3. Do body fossils show signs of transportation (e.g. disarticulation, breakage) or are they in living position?
    4. What are the fossilized organisms' relationship to the substrate/bedding (i.e. infaunal vs. epifaunal)?
    5. How are the fossils preserved?

In this section, we will explore more sedimentary structures that can aid our interpretation of depositional environments and Earth's history.

Structures Formed By Sediment Gravity Flows

Graded bedding refers to a sequence of increasingly coarse- or fine-grained sediment layers. Graded bedding often develops when sediment deposition occurs in an environment of decreasing energy.

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**. Bouma sequence beds are formed by offshore sediment gravity flows, which are underwater flows of sediment. These subsea density flows begin when sediment is stirred up by an energetic process and becomes a dense slurry of mixed grains.

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

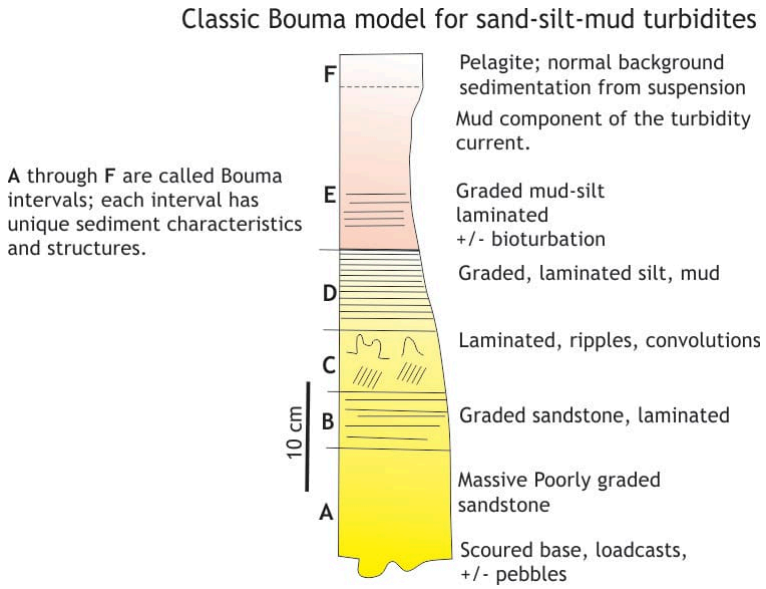


Figure 6.2.1: Bouma sequence for turbidite deposition, modeling classic graded bedding. A=coarse- to fine-grained sandstone, possibly with an erosive base. B=laminated medium- to fine-grained sandstone. C=rippled fine-grained sandstone. D=laminated siltstone grading to mudstone. E=mud mixed with background suspended sediment from the water column. F=return to normal background sedimentation. (Brian Ricketts.)

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).

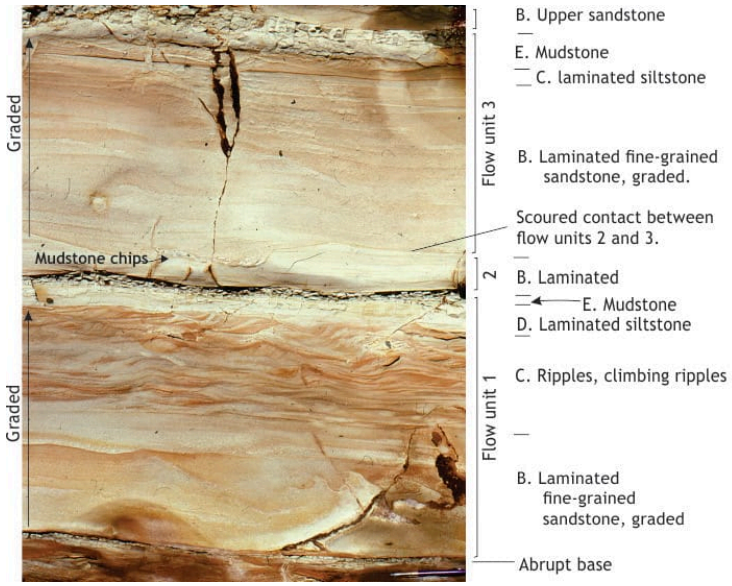


Figure 6.2.2: Three turbidite flow units: A lower B-C-D-E unit with excellent examples of ripples and climbing ripples in the C interval; a middle B unit, the top of which has been eroded by the upper B-C-D-E unit. Units 1 and 3 are graded. A lag of mudchips formed during scouring of the middle unit. (Brian Ricketts.)

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.



Figure 6.2.3: 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 in debris flows tend to be pebble through boulder size, and are extremely useful for identifying ancient sediment sources. (Brian Ricketts.)

Sole Structures

Sole structures form on the base of some beds 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 formed are filled by new sediment so 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. Like the sole is the bottom of a shoe, sole structures are 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. Gutter casts are narrow, elongate scours in the sea floor 15-20 cm deep that form during coastal storm surges.

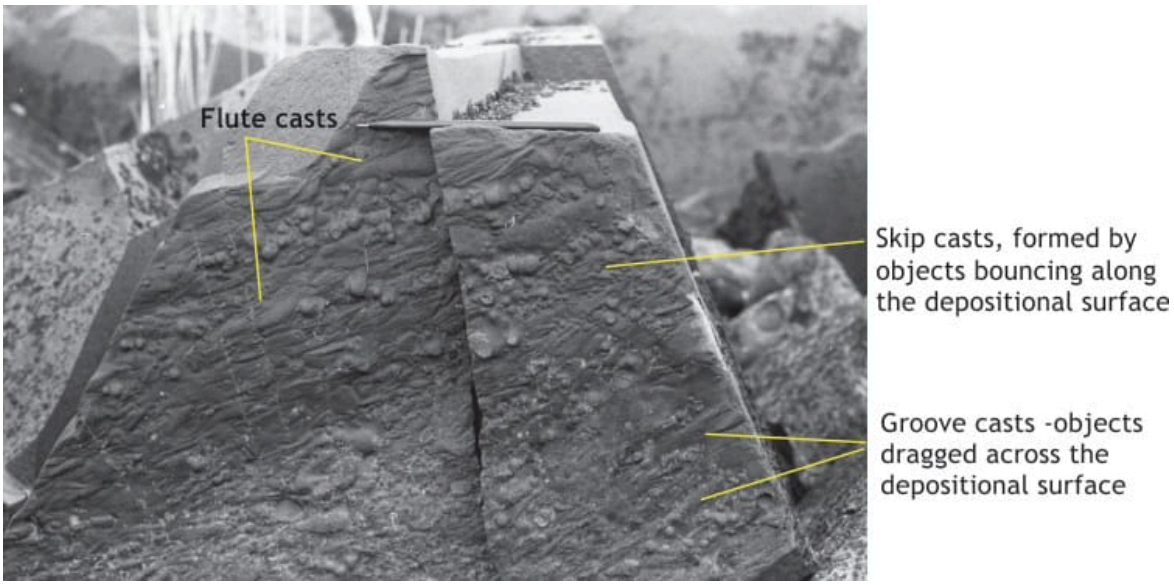


Figure 6.2.4: A nice collection of skip, groove, and flute casts at the base of a turbidite bed. The flutes indicate flow to the right. (Brian Ricketts.)

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, clasts like these will tend to be aligned according to their most stable hydrodynamic orientation. This preferred alignment of the clasts is referred to as **imbrication**. Identification of preferred clast orientations provides another useful tool for measuring paleoflow directions.



Figure 6.2.5: Imbrication of oblong or platy gravel clasts can serve as an indicator of paleo-current flow direction. This example comes from Death Valley, California. (Callan Bentley.)

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.



Figure 6.2.6: 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. (Brian Ricketts.)

When preserved in the rock record, mud cracks may indicate:

- falling sea levels,
- changes in the location of river channels and adjacent floodplains,
- arid climate.

Trace Fossils & Bioturbation



Figure 6.2.7: Bioturbated sandstone from the Lower Cambrian Sellick Hill Formation in coastal South Australia. Burrowed textures in sedimentary rocks are referred to as bioturbation. (Trace Fossils 15 by James St. John, licensed under CC BY 2.0.)

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; trace fossils, or lack of them, also reflect these events.

Most animals produce more than one kind of trace depending on what they are doing. This 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 (for the last ~540 million years or so) have enjoyed the munching-burrowing efforts of myriad nameless critters.

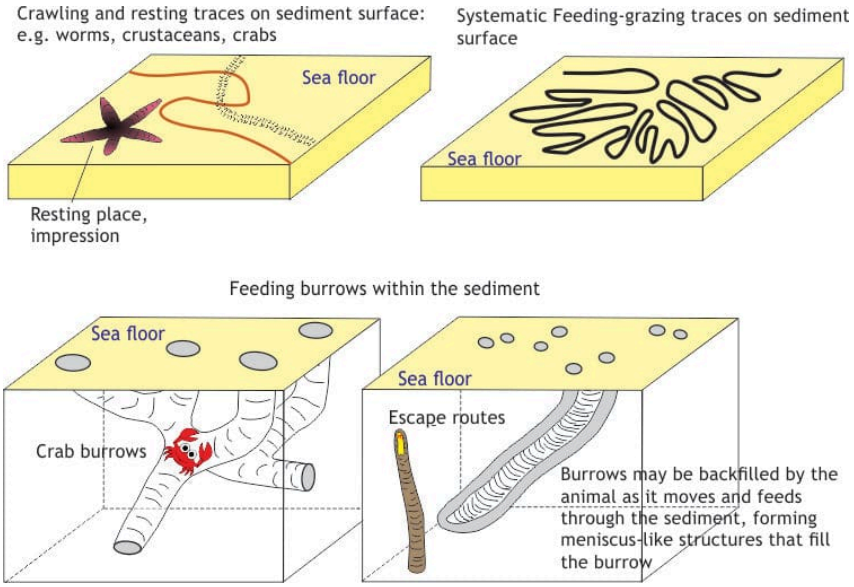


Figure 6.2.8: Trace fossils can be divided into groups according to the behavior of the animals that produced them. Here are four schematics of resting, crawling grazing, feeding, dwelling, and escape activities. (Brian Ricketts.)

The images below are organized into six ethological (i.e. behavioral) structures:

1. Resting traces *(Cubichnia):* the animal is taking a break or escaping a marauding predator.



Figure 6.2.9: 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. (Brian Ricketts.)

2. Crawling traces *(Repichnia):* Moving from one point to another, so the emphasis is on body and limb movement.



Figure 6.2.10: 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. (Brian Ricketts.)

3. Grazing traces *(Pascichnia):* The search for food at the sediment surface or just beneath it – similar to humans strip mining.



Figure 6.2.11: *Helminthopsis,* a grazing trace fossil from the Logan Formation (Lower Carboniferous) of Wooster, Ohio. (Wikimedia Commons by Mark A. Wilson.)

4. Feeding structures *(Fodichnia):* Like grazing, but in this group the critters burrow (like subsurface mining), forming simple temporary burrows or branched burrow complexes.


Figure 6.2.12: 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. (Brian Ricketts.)

5. Dwelling structures *(Domichnia):* These are more permanent burrows that the animals call home.



Figure 6.2.13: 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 environments (e.g. fossils, crossbedding). (Brian Ricketts.)

6. Escape structures *(Fugichnia):* 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.



Figure 6.2.14: Simple vertical structures that begin from the base of a bed, reflecting the critter's desire to escape to the sea floor. (Brian Ricketts.)

Key Terms

- **bedding** - the layered structure of sedimentary rocks, where layers, are separated by planes
- **bioturbation** - the physical churning and mixing of sediments by the activities of living organisms (such as burrowing)
- **Bouma Sequence** - a geological model describing the idealized vertical succession of sedimentary structures and lithologies in a turbidite
- **composition** - the materials something is made of
- **debris flow** - mixtures of mud, water, and coarse debris, moving down a slope due to gravity that lack fluid turbulence
- **fabric** - the arrangement and organization of a rock's constituent parts and features

- **fossil** - a remnant or evidence of past animal or plant activity
- **graded bedding** - a sedimentary layer characterized by progressively finer-grained particles from the bottom upward
- **imbrication** - a primary depositional fabric where elongated or flat clasts are arranged in a overlapping pattern indicating flow direction
- **texture** - the size, shape, and arrangement of mineral grains, crystals, or fragments within a rock
- **turbidite** - a type of sedimentary rock composed of layered particles that grade upward from coarser to finer sizes thought to have originated from ancient fast-moving, gravity-driven underwater currents carrying a high concentration of suspended sediment in the oceans

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6.3: Fossils and Fossilization

Fossils Defined

Consider this photo of *Archaeopteryx*. This image likely represents what you might think of when you 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 6.3.1: Archaeopteryx fossil

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.



Figure 6.3.2: Fossil ant in Baltic amber, about 50 million years Eocene

Taphonomy: From Death to Fossil

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.

Decay of Soft Tissue

When a living organism dies, soft tissues (e.g. 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, because these factors inhibit bacterial growth. Therefore, some of the best fossils that include unaltered soft tissues come from deposits that represent rapid burial that 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.



Figure 6.3.3: Otzi is a mummified man, who had been preserved in ice, and found high in the Alps between Austria and Italy.

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 6.3.4: 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.

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Figure 6.3.5: Articulated Mercenaria sp. specimen

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Figure 6.3.6: Disarticulated Mercenaria specimen



Figure 6.3.7: 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 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.

Body Fossils: 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 6.3.1

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 6.3.8: 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.

Body Fossils: Altered Remains

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 6.3.9: 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 in a process referred to as **carbonization**. Intricate details of delicate soft tissues may be preserved, including individual cell structure in plants, fish scales, insect wings, and even internal organs.

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Figure 6.3.10: Crane fly, Florissant Fossil Beds National Park

Replacement

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 in a process called **replacement**. 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.



Figure 6.3.11: Pyritized trilobite, Triarthrus eatoni, Upper Ordovician, New York

Recrystallization

Recrystallization is an important process that preserves fossils made of aragonite - a common but chemically unstable mineral - over geologic time. Many marine invertebrates (molluscs, modern corals) have a metabolism that allows them to crystallize aragonite as their skeletal material. Aragonite is a polymorph of $\mathrm{CaCO_3}$ 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.

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Figure 6.3.12: Iridescent mother of pearl on nautiloid shell

Internal Molds

When the interior space of an organism is buried and filled with sediment that is then lithified, the interior shape is preserved as an . 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.



internal mold

Figure 6.3.13: Internal mold of snail. Note shell is dissolved away and only infilling sediment remains.

External Molds

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.

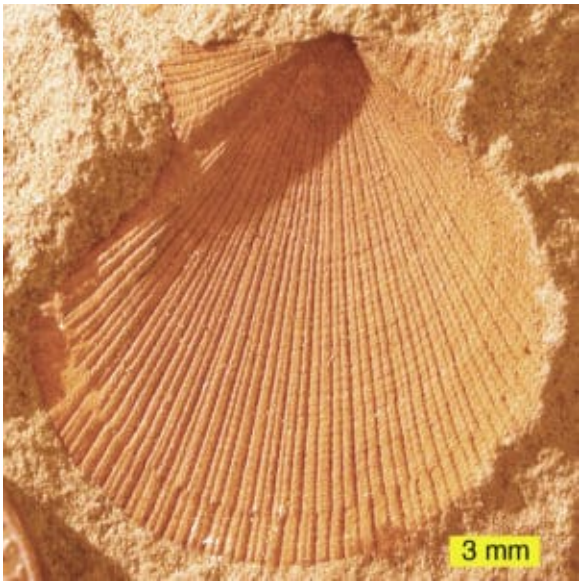


Figure 6.3.14: External mold of a bivalve from the Logan Formation, Lower Carboniferous, Ohio

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.



Figure 6.3.15: Dinosaur trackways, Dinosaur Ridge, Colorado.

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 6.3.16: 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.

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Figure 6.3.17: External mold (left) of a trilobite body fossil in a concretion

Casts

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.



Figure 6.3.18: Sue, the most complete (and famous) Tyrranosaurus rex fossil at the Field Museum, Chicago.

Trace Fossils

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.

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.



Figure 6.3.19: Stromatolite fossil from the Fort Laclede Bed, Laney Member, Green River Formation of south-western Wyoming in USA. It is the preserved fossilized remains of a stromatolite that lived in the ancient Lake Gosuite. This is a cross-sectional slice through the fossil, showing its internal layered structure. (Stromatolite Fossil from Wyoming by James St. John, CC BY 2.0)

Classification of Trace Fossils

Trace fossils are often classified by an ecological method that focuses on the behavior of the organism (referred to as ethologic classes, after Seilacher 1967). 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.



Figure 6.3.20:

Skolithos linearis burrows in quartzose sandstone, frequently referred to as "piperock" by geologists because of the linear vertical burrows (Clinch Formation, Lower Silurian; Clinch Mountain, Tennessee, USA). (Skolithos linearis burrows in sandstone 6 by James St. John, CC BY 2.0)

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. 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. Organisms burrow now just as they did 400 million years ago; though the organisms may be different, the action is the same (Seilacher, 1967). In rare instances, organisms such as trilobites have been preserved within a trace they created (e.g., *Cruziana sp.*); however, it is usually very difficult to associate a particular organism with a particular trace. The real utility of trace fossils is in their association with particular environments.

Some traces are characteristic of beaches, while other different traces are typical of deep-water muds. 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. Thus, trace fossils are excellent facies indicators and provide fabulous insights into the organisms and environments that formed them.

Further Readings

University of California Museum of Paleontology Fossils

Ichnofacies descriptions

Seilacher, A. (1967). “Bathymetry of trace fossils”. *Marine Geology*. **5**: 413–428. doi:10.1016/0025-3227(67)90051-5.

Key Terms

- **body fossil** - a preserved remnant or part of an actual organism, such as bones, teeth, shells, leaves, or wood

- **carbonization** - a specific form of fossilization where organic matter (like plants) is converted into a thin, dark, flattened film of carbon due to heat and pressure, driving off volatile elements and preserving detailed impressions of the original organism
- **casts** - replicas of organisms, essentially recreating their original shape
- **external mold** - an impression of an organism's outside shape
- **fossil** – any preserved remains, impression, or trace of an organism that lived in the past; must be older than 11,700 years old
- **internal mold** – a fossil formed when sediment fills the inside of a shell or other hollow structure; after the shell dissolves, the hardened sediment shows the internal shape
- **permineralization** – a fossilization process where minerals carried by water fill the pores and empty spaces in bones, wood, or shells, preserving fine details of the original structure
- **recrystallization** – a preservation process where the original minerals of a fossil change into more stable forms without changing the overall shape of the fossil
- **replacement** – a fossilization process where the original material of an organism is gradually dissolved and replaced molecule by molecule with minerals, creating a detailed copy
- **trace fossil** – evidence of an organism's activity such as footprints, burrows, feeding marks, or feces

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6.4: Facies

Facies Defined

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.



Figure 6.4.1: 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.)

Describing Sedimentary Facies

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 6.4.1: 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

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 6.4.2: Weathering rind showing difference in color on external surfaces versus inner portion of rock. (Photo by C. Bentley.)

Sedimentary Rock Composition

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

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 (QFL) fragments in the rock, 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.

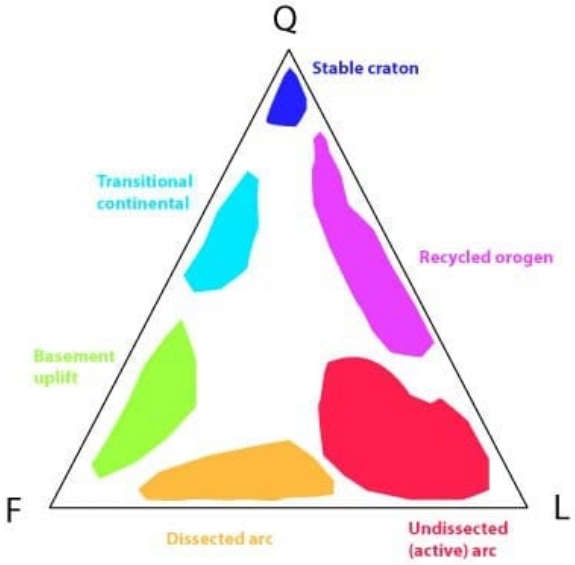


Figure 6.4.3: QFL (Quartz, feldspar, and lithic) ternary diagram for clastic sedimentary rocks.

Carbonate Rocks

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).

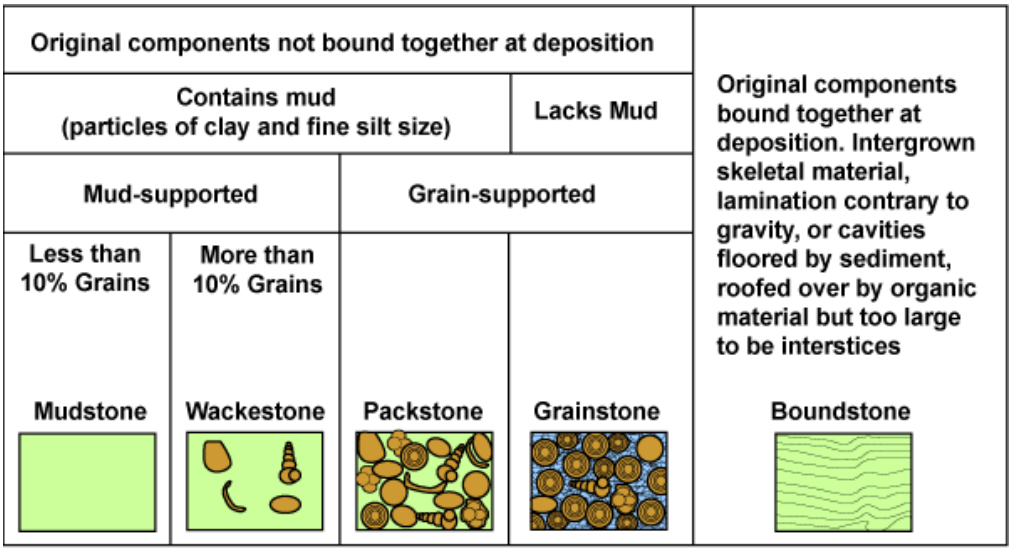


Figure 6.4.4: Dunham classification scheme for carbonate rocks.

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.

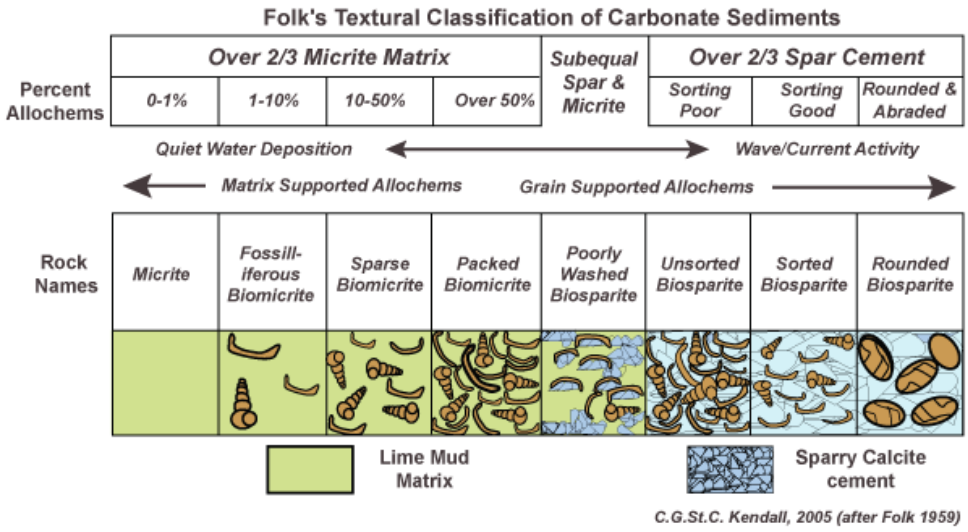


Figure 6.4.5: Folk classification scheme for carbonate rocks

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 6.4.6: 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))

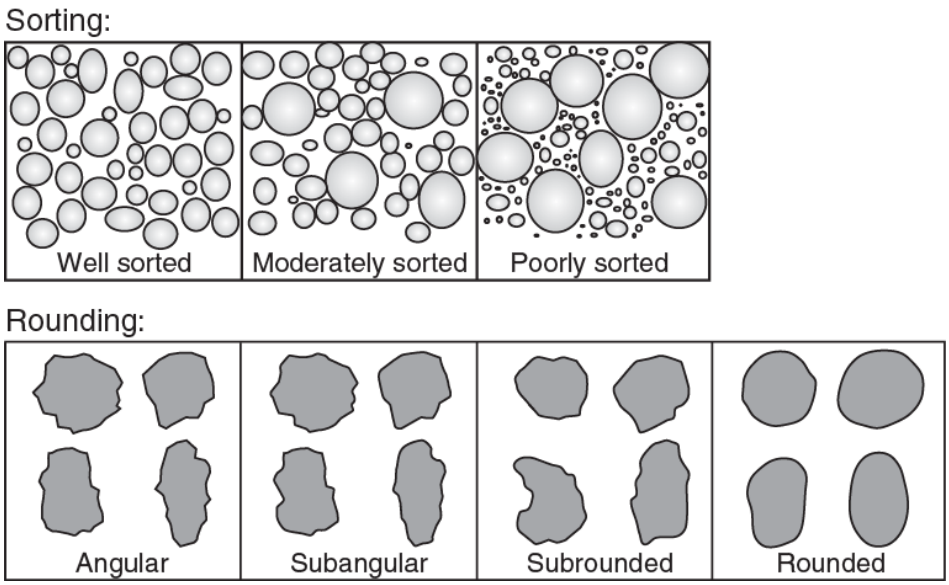


Figure 6.4.7: Top: Sorting of grains, ranging from well sorted where the grains are similar in size, to poorly sorted, where the grains vary greatly in size. Bottom: Rounding refers to how smooth or rough the edges of a clast are. Clasts with sharp edges and corners are angular. Clasts with smooth surfaces are rounded. Clasts that fall in between are sub-angular or sub-rounded. (CC BY 3.0; Reagan et al. (2015))

Sedimentary Structures

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.

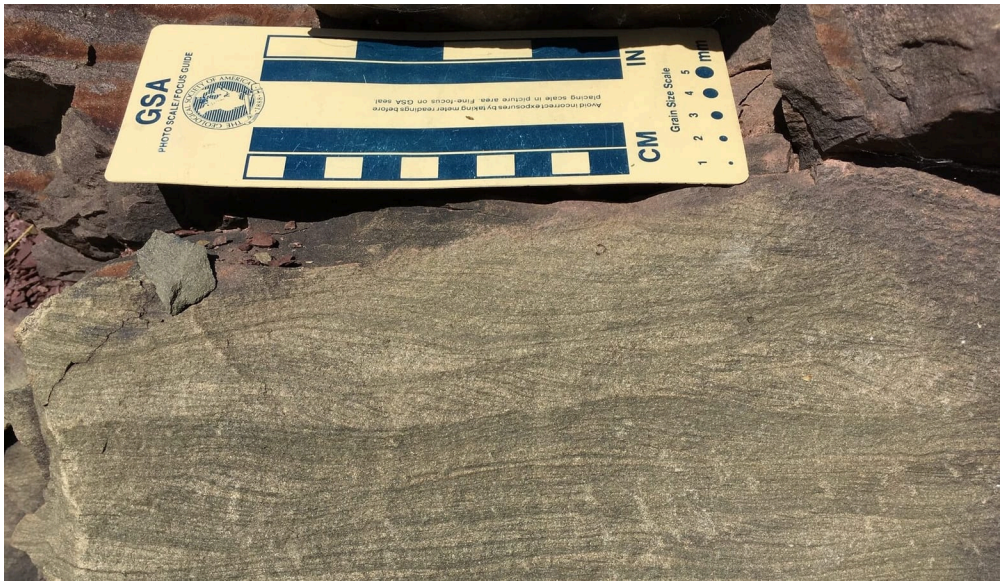


Figure 6.4.8: Ripples in cross section, Corridor H, West Virginia. (Photo by K. Layou.)

Fossils

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 6.4.9: Fossiliferous layers of the Yorktown Formation at Cobham Wharf, Virginia. (Photo by K. Layou.)

Association

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.



Figure 6.4.10: Conformable Ordovician limestone layers near Carthage, Tennessee. (Photo by K. Layou.)

Form

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.



Figure 6.4.11: 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.)

Key Terms

- **carbonate rocks** – sedimentary rocks made mostly of carbonate minerals (like calcite and dolomite), commonly formed in marine environments (e.g., limestone, dolostone)
- **facies** – the set of characteristics (such as rock type, fossils, sedimentary structures, and color) that reflect the conditions in which a sedimentary rock was deposited
- **sedimentary structures** – features that form in sediments at or shortly after deposition, providing clues to the depositional environment (e.g. a current, alternating wet and dry environment, etc.)
- **siliciclastic rocks** – sedimentary rocks made mainly from silicate mineral fragments formed from the weathering and erosion of preexisting rocks

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6.5: Depositional Environments

Sedimentary facies reflect the environment in which the sediments were deposited. Here are some 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. For each sedimentary environment, we'll highlight the facies characteristics that are most observable in hand samples: color and composition, texture, sedimentary structures, fossils, and any other distinctive characteristics.

Terrestrial/Continental Depositional Environments

Alpine Glaciers

When glaciers melt, they deposit their sediment load. Glacial sediment is called till.



Figure 6.5.1: 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

Alluvial fans are hand-fan-shaped piles of alluvium (sediment deposited by streams) that form where streams exit their mountain channels. Common in arid environments with growing mountains.

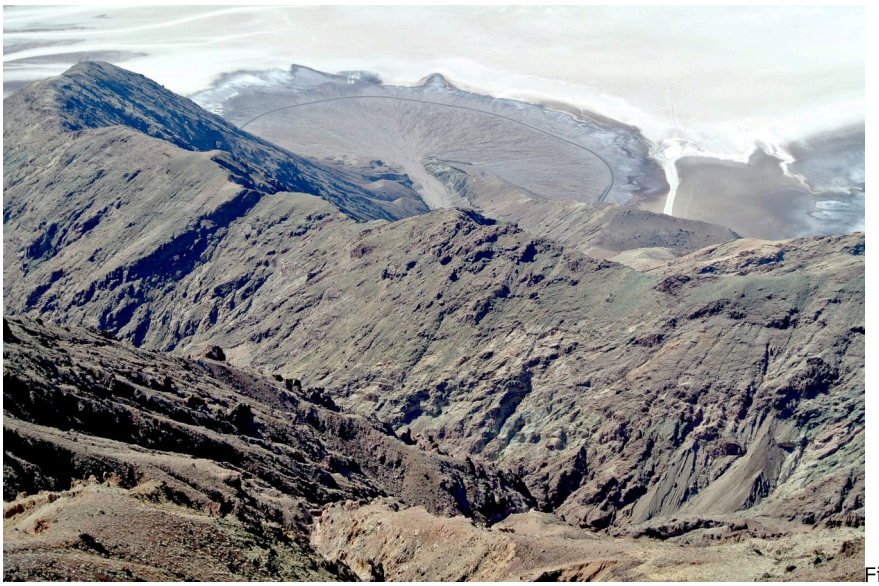


Figure 6.5.2: Alluvial fan in Death Valley, California, seen from source area of Black Mountains at Dante's View.

- 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

Braided rivers are characterized by lobe-shaped piles of alluvium across the stream's channel. These channel bars divert stream flow, causing the river to split into many intertwining channels.



Figure 6.5.3: 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)

Meandering Rivers

A river with a single, sinuous channel characterized by large bends (meanders) that migrate laterally across the floodplain.



Figure 6.5.4: 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)

Lakes/Lacustrine

A body of freshwater with little or no current.



Figure 6.5.5: 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)

Deserts/Aeolian

Landforms produced or transported by wind, such as sand dune deposits.



Figure 6.5.6: 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

Transitional/Coastal Depositional Environments

Deltas

Wedge-shaped deposits of sediment formed at the mouth of a river where it enters a standing body of water, such as a lake or ocean.



Figure 6.5.7: Yukon River Delta entering the Bering Sea, creating a lobe shaped protrusion formed from the sediments carried by the river. (Yukon Delta, Alaska by NASA Earth Observatory, CC BY 2.0)

- 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

Gently sloping shoreline environment composed of unconsolidated sediment (sand, gravel, or pebbles) shaped by waves, tides, and currents.



Figure 6.5.8: 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

Tidal Flats

Low-lying coastal areas that are periodically flooded and exposed to tides; often composed of mud and sand.



Figure 6.5.9: 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, symmetrical ripples
- Fossils: stromatolites, snails, crustaceans and burrows

- Other notes: may see laminations if burrows not present

Marine Depositional Environments

Shallow Subtidal (Upper Shoreface)

Shallow marine settings where sand to clay-sized terrestrial sediment along with biogeneous sediment is shaped by waves, tides, and currents.



Figure 6.5.10: Symmetrical ripples in biogenic, aragonitic sands; shallow subtidal seafloor just offshore from Sand Dollar Beach, northwestern San Salvador Island, Bahamas. (Symmetrical ripples in biogenic, aragonitic sands 2 by James St. John, CC BY 2.0)

- Color: gray to tan
- Composition: quartz and clay minerals OR calcite
- Texture: subrounded sand, silt-clay, well-sorted
- Sedimentary structures: burrows, wave-formed ripples, crossbedding
- 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)

A transition zone between the high-energy upper shoreface and low-energy offshore environments.



Figure 6.5.11: Laminated mudrock from the Kinzers Formation, Lower Cambrian, Pennsylvania. Well-preserved laminations indicate lack of post-depositional disturbance by burrowing organisms. (Laminated mudrock 2 by James St. John, CC BY 2.0)

- Color: dark gray to dark brown
- Composition: quartz and clay minerals OR calcite
- Texture: silt to clay
- Sedimentary structures: laminations, hummocky cross stratification
- Fossils: planktonic microfossils, ammonites, vertebrates
- Other notes: area of shelf below storm wavebase

Submarine Fan and Turbidites

Large, fan-shaped bodies of sediment deposited on the deep ocean floor by turbidity currents, usually at the base of continental slopes.



Figure 6.5.12: A prominent submarine dune with large ripples, possible result of a turbidity current, Gulf of Mexico. (expn2486 by NOAA OKEANOS EXPLORER Program, Gulf of Mexico 2014 Expedition, CC BY 2.0)

- 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

Vast, flat, and deep region of the ocean floor, typically found at depths of 3,000–6,000 meters, covered by fine-grained sediments.

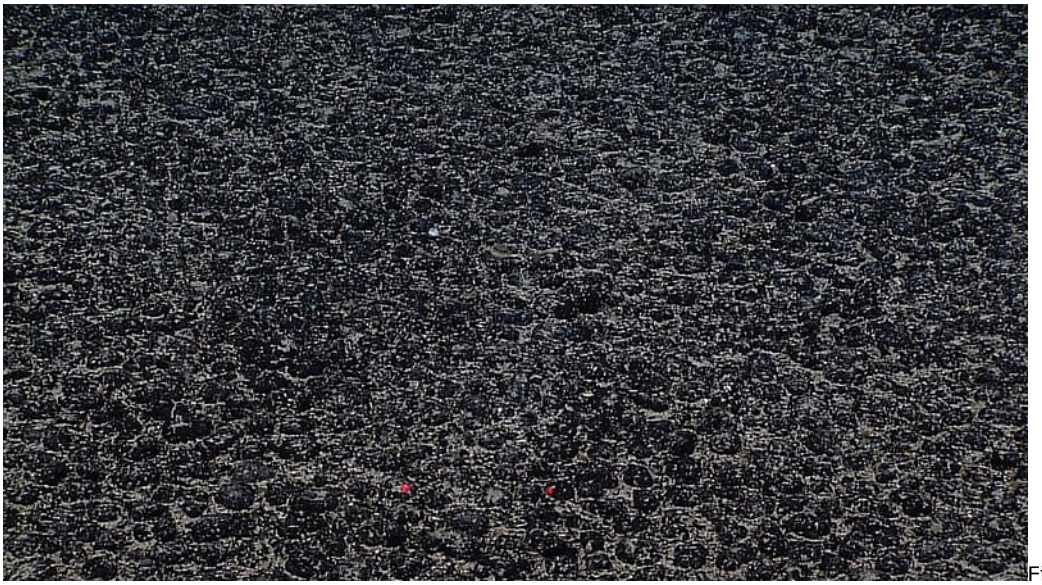


Figure 6.5.13: A pavement of manganese nodules on the deep sea floor; each nodule appears to be about 1 inch across (2.5 cm). Hawaii, West Northampton Seamount Ridge. (expn4910 by NOAA Office of Ocean Exploration and Research, 2015 Hohonu Moana, CC BY 2.0)

- 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

Key Terms

- **abyssal plain** – the vast, flat, and deep region of the ocean floor, typically found at depths of 3,000–6,000 meters, covered by fine-grained sediments
- **aeolian** – refers to processes, sediments, or landforms produced or transported by wind, such as sand dunes deposits
- **alluvial fan** – a fan- or cone-shaped deposit of sediment that forms where a river or stream flows out of a mountainous area onto a flatter plain
- **alpine glacier** – a glacier that forms in mountainous regions and flows down valleys under the influence of gravity
- **beach** – a gently sloping shoreline environment composed of unconsolidated sediment (sand, gravel, or pebbles) shaped by waves, tides, and currents
- **braided river** – a river system with multiple intertwining channels separated by bars or islands of sediment, usually formed where sediment supply is high and water amount and velocity varies
- **delta** – a wedge-shaped deposit of sediment formed at the mouth of a river where it enters a standing body of water, such as a lake or ocean
- **desert** – a region that receives very little precipitation (typically less than 25 cm per year), characterized by sparse vegetation and dominance of wind and occasional water erosion
- **lacustrine** – pertaining to lakes; describes sediments, deposits, or environments associated with lake basins

- **meandering river** – a river with a single, sinuous channel characterized by large bends (meanders) that migrate laterally across the floodplain
- **submarine fan** – a large, fan-shaped body of sediment deposited on the deep ocean floor by turbidity currents, usually at the base of continental slopes
- **tidal flat** – a low-lying coastal area that is periodically flooded and exposed by tides, often composed of mud and sand
- **turbidite** – a sedimentary deposit formed by an underwater landslide, typically showing graded bedding with coarser material at the base and finer material at the top

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- 14.3: Putting it all together - facies models by Callan Bentley, Karen Layou, Russ Kohrs, Shelley Jaye, Matt Affolter, and Brian Ricketts is licensed CC BY-NC 4.0. Original source: https://opengeology.org/historicalgeology/.

6.6: Introduction to Strata and Basins

Overview

Plates move and collide, mountains rise, sea level rises and falls. 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. Many variables determine the outcome for these sediments: latitude and climate, the composition of parent rock that weathered, the degree of physical and chemical weathering, life forms that interacted with the grains, etc., ultimately determine the story written in the layers of sediment. As these layers accumulate, so do the chapters in the history book of the Earth.

Strata

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.



Figure 6.6.1: 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.)

Depending on the distance traveled, some sediments 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 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**.

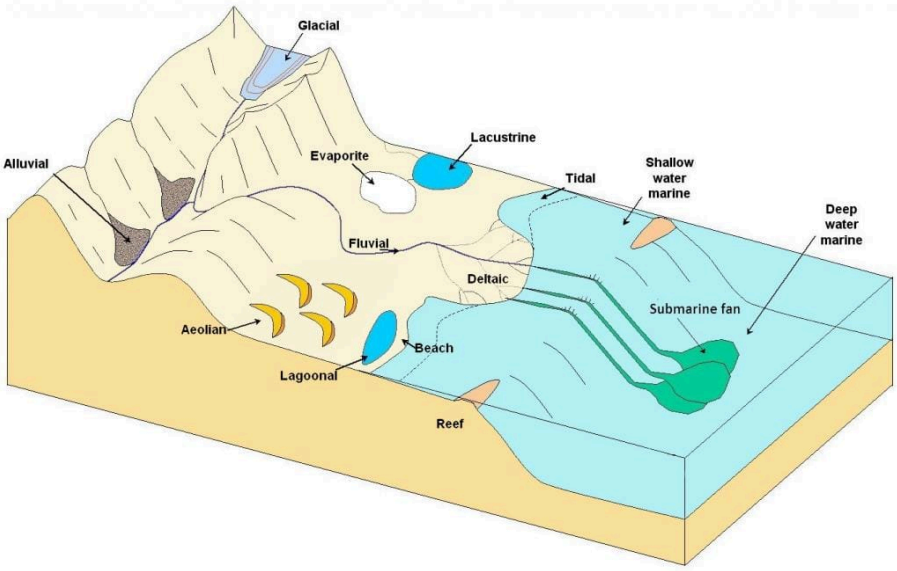


Figure 6.6.2: Sedimentary facies commonly found in a variety of basins.

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 as discussed previously. Real-time facies migrate, or change, at a location as conditions change. For example, when sea level rises and an area is submerged, the type of sediment that deposits changes. 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.

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.

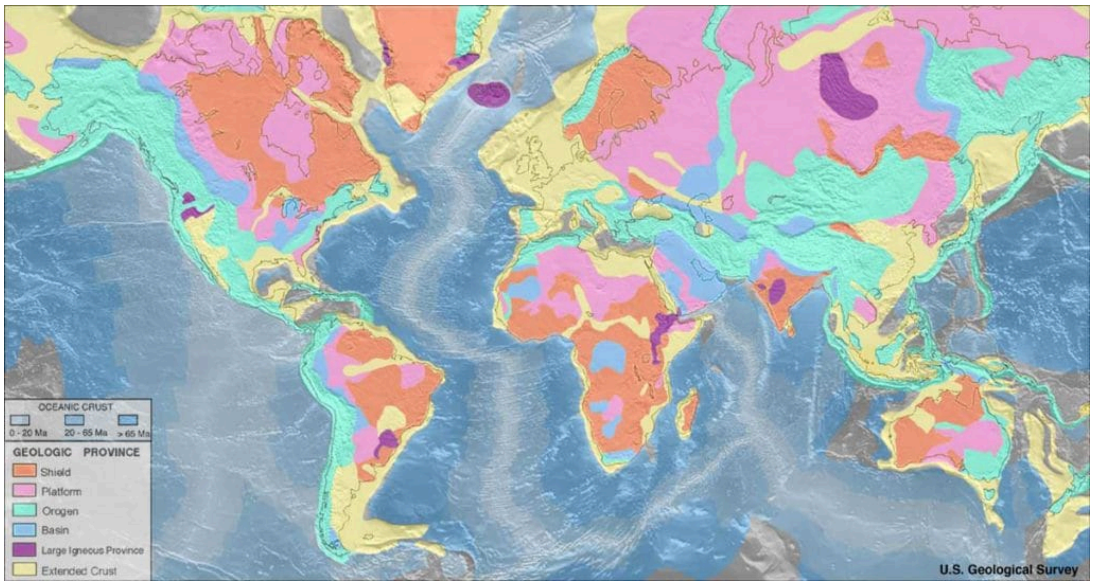


Figure 6.6.3: USGS image depicting shields, platforms, basins, recent orogens (mountain belts), and other features. The cratons are made up of the shield and platform areas. The basins, in blue, are just some of the many sedimentary basins that exist globally. In modern as in ancient times, the oceans themselves are basins.

On shorter time scales, 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.

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 available volume within a basin to accommodate influxes of sediment. 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.

Stratigraphy in 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, some 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.

(Note: the limitation of this metaphor is that most of the "stratigraphy" you see here is vertical, implying a great deal of geologic alteration. 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.

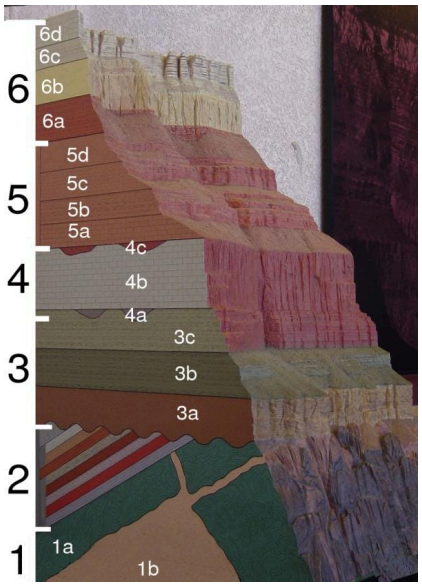


Figure 6.6.4: A geologic model of the Grand Canyon. Formations are indicated with numbers and letters (ie 6a through 6d are all formations, or books of rock). Formations are then assembled into Groups, where appropriate (ie 1 through 6). Groups are like shelves of similar books. (CC BY-SA 3.0; The original 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.

Key Terms

- **bed** – a small recognizable layer of rock sedimentary rock
- **craton** – a large, stable block of the Earth's crust that forms the old, tectonically inactive core of a continent
- **formation** – a distinct, mappable body of rock with unique physical characteristics, such as color, composition, or particle size, that differentiates it from the surrounding rock layers
- **platform** – a portion of a craton that is covered by relatively undeformed, horizontally lying or gently tilted sedimentary rocks, overlying older crystalline basement

- **sedimentary basin** – a low area of the Earth's crust where sediment accumulates over time to form thick, laterally extensive sequences
- **shield** – a large, exposed area of a craton composed of old crystalline igneous and metamorphic rocks, typically tectonically stable and deeply eroded
- **strata** – distinct layers of sedimentary rock or volcanic deposits that can be distinguished by composition, texture, or other features
- **stratigraphy** – the branch of geology concerned with the description, correlation, and interpretation of layered rocks, including their age relationships and depositional environments

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6.7: Sequence Stratigraphy and Walther's Law

Understanding Earth's History Through Sequence Stratigraphy

In the stratigraphic record of a basin, there can be packages of beds that repeat in a cyclic fashion: these related and predictable packages of rock are referred to as **sequences**. Sometimes, sequences 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.

YouTube Video

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These cycles 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.

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 first-order megasequence, which has a period of several hundred million years, a number of second-order supersequences will occur. Within a supersequence, a number of smaller sequences occur, and so on. Each of these different packages of strata are bounded by unconformities, created by erosional flooding surfaces, that can be traced over distances. 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.

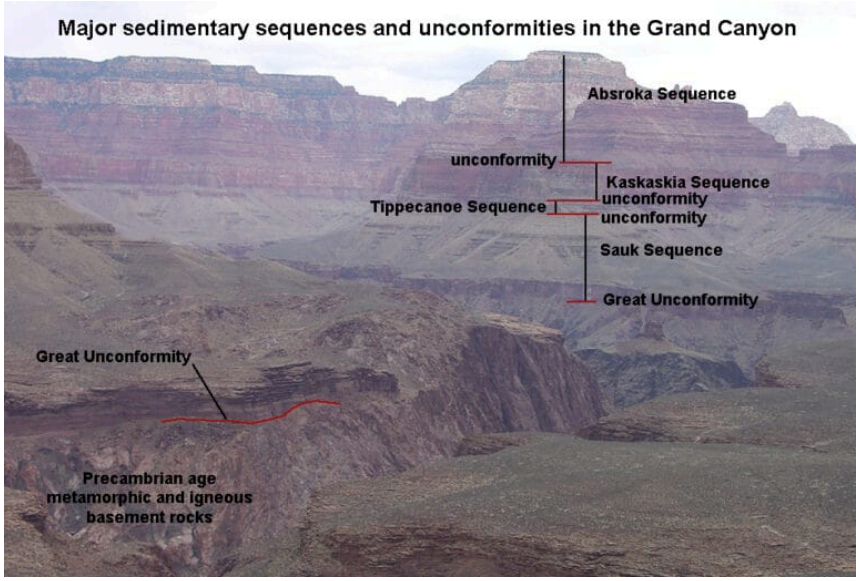


Figure 6.7.1: Cratonic Sequences, or Sloss Sequences, as they exist in the Colorado Plateau and Grand Canyon region in the American southwest.

The causes of sequences of various hierarchical levels is subject to debate. First-order megasequences are known to reflect global phenomena, such as the formation and breakup of supercontinents like Rodinia and Pangaea. “The Great Unconformity”, viewable at the Grand Canyon, 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.

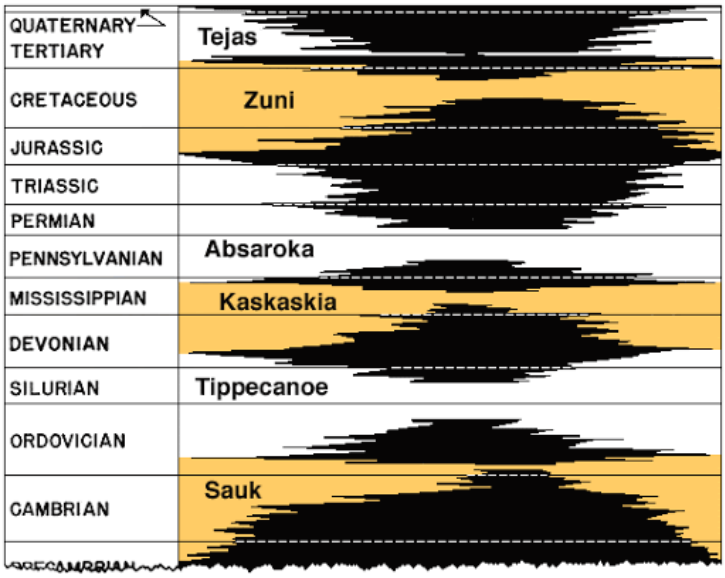


Figure 6.7.2: Sedimentary “Sloss” Sequences of North America. White and yellow areas represent major marine transgressions onto the craton while black areas represent major unconformities (sequence boundaries). The Sauk is bounded on the bottom by the “Great Unconformity”, the boundary between the Sauk 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)

Second-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 second-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, third-order sequences. Within those sequences, fourth-order and fifth-order parasequences, by which level, the data is entirely within a single stratigraphic formation and could reflect a

cyclicity of just days. These shortest parasequences can be driven by a number of factors, but Milankovitch orbital variations are likely drivers.

The ultimate goal of sequence stratigraphy is to understand the Earth's past as recorded in the stratigraphic history of a basin. We can use 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.

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, move 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).

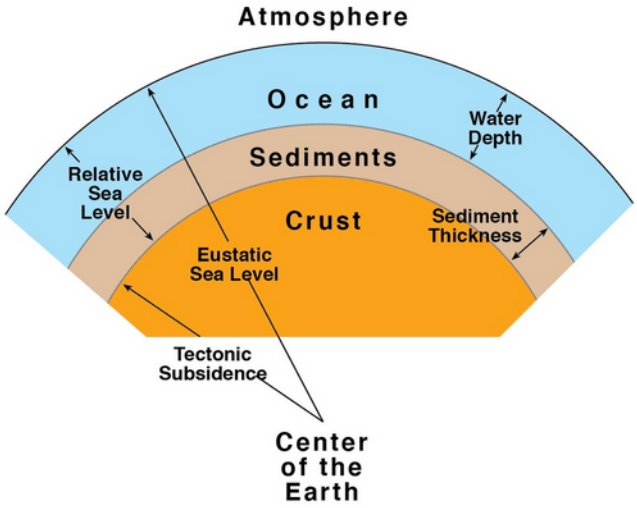


Figure 6.7.3: Cross-section of the Earth showing important sequence stratigraphic definitions (from Holland, 2018).

T: Tectonic Changes

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. 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.

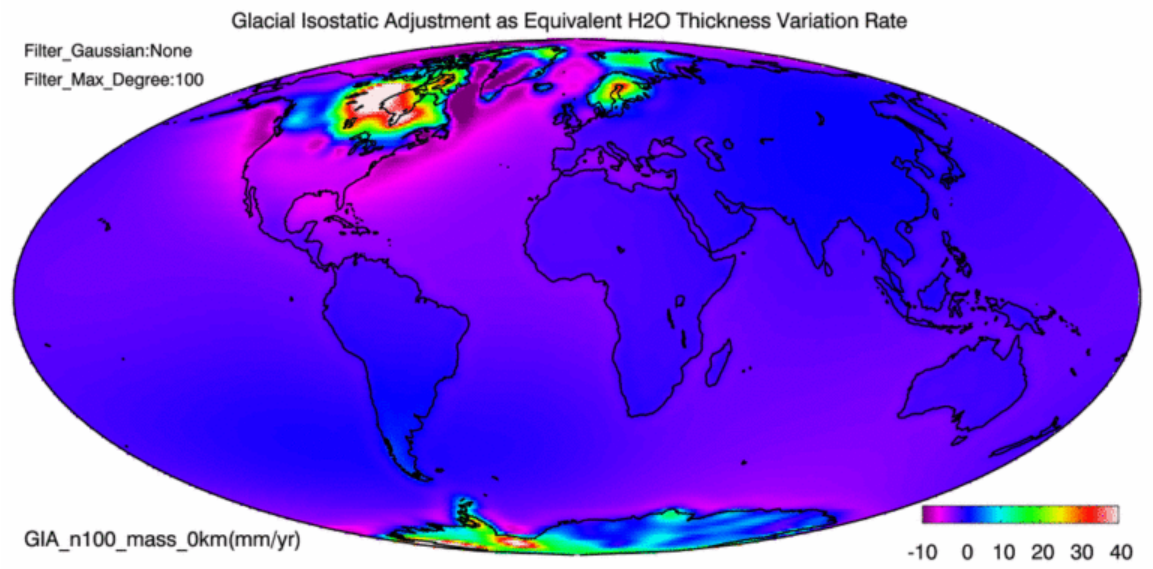


Figure 6.7.4: Glacial Isostatic Adjustment as equivalent H_2O thickeness variation rate. In other words, gravitational anomalies (changes) as a result of tectonic rebound from the last glacial advance of the Pleistocene. Note the high levels of adjustment over areas of northern North America, Europe, 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

E: Absolute Sea Level Changes

Absolute sea level (ASL), also known as eustatic sea level, is the distance between the center of the Earth and the global average sea level. Eustatic sea level fluctuates as the volume of water in the ocean changes, due to changes in glaciation, ocean temperatures, and glacial ice volume. 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.

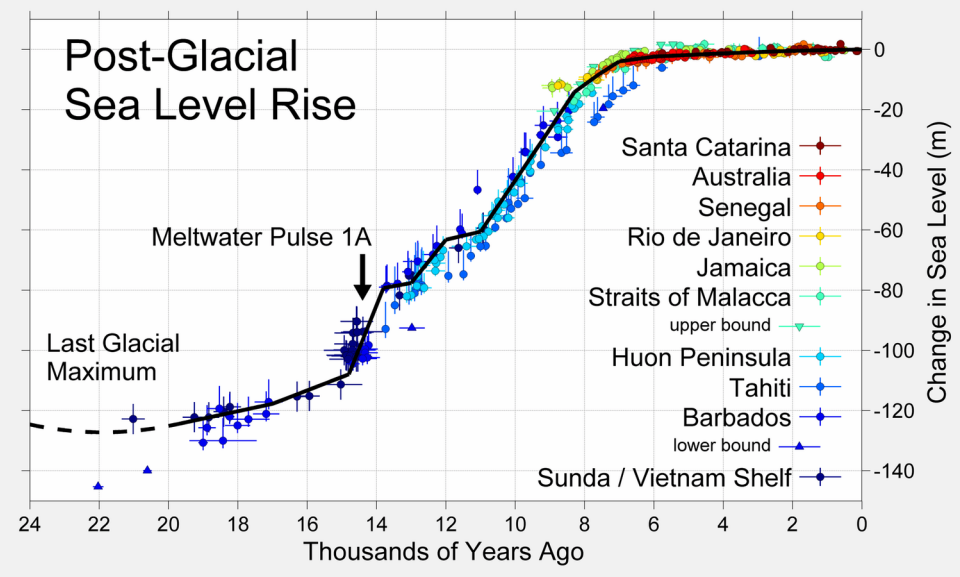


Figure 6.7.5: Eustatic sea level rise resulting from the melting of glacial ice over the past 24,000 years (Post-Glacial Sea Level by Robert A. Rhode, CC BY-SA 3.0).

S: Sedimentation Rates

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 $\mathrm{CaCO_3}$ directly from seawater.

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.

W: Relative Sea Level Changes

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 1) increases in water mass (from melting glacial ice), 2) thermal expansion of water with a rise in temperature, 3) changes in ocean currents or seawater salinity, and 4) isostatic rebound adjustments of continents.

When we look at NOAA tidal gage data today, we can see that the eastern shore of Virginia is experiencing rising seas, while the Gulf coast of Alaska is experiencing a net drop in sea level. What explains the difference in relative sea level trends?

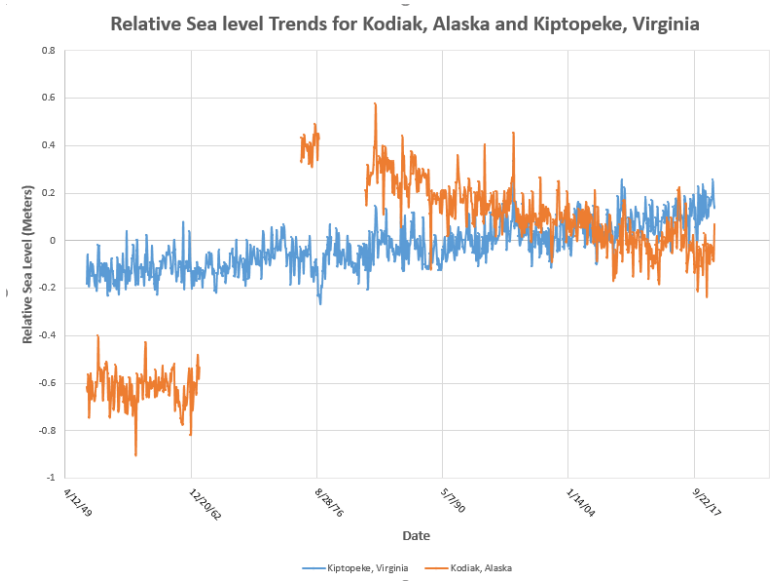


Figure 6.7.6: Relative sea level (RSL) trends for two sites on opposite sides of the North American craton. Kiptopeke, Virginia, is experiencing a relative rise in sea level of 3.97 mm/yr where Kodiak, Alaska (southern coast) is experiencing a drop in relative sea level of -9.92 mm/yr. While global (eustatic) 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.

Virginia's relative sea level reflects the eustatic trend in increasing global sea level due to thermal expansion of the oceans, but the tidal gauge data for Alaska nicely illustrates 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.

Transgressions, Regressions, and Walther's Law

As mentioned above, when sediments are deposited into a basin, they are affected by and also affect the basin. Where sediments ultimately come to rest depends on 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).

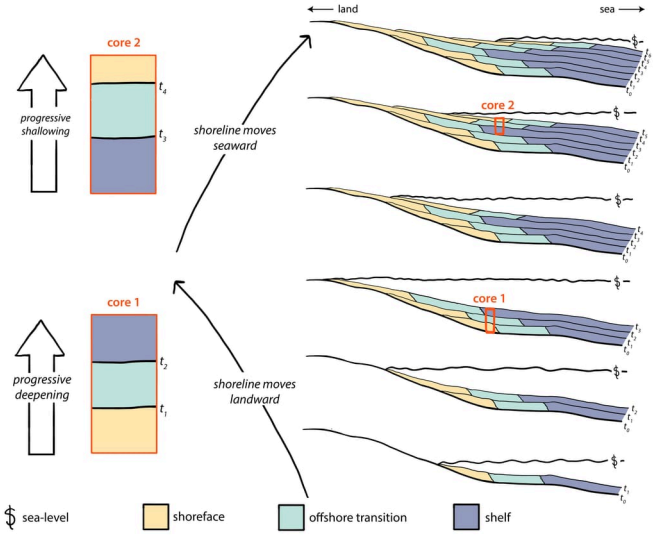


Figure 6.7.7: Overview of transgression (bottom three time slices) and regression (top three time slices) as recorded by changes in water depth, shoreline position, and changes in depositional environments through time (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

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, **Walther's Law** proposes that facies in continuous vertical successions were also deposited in laterally adjacent environments. In the diagram provided above, we have three laterally adjacent environments: a shoreface (likely sandstone), offshore transition (likely mudrock), and shelf (likely carbonate). If a geologist were given Core 1, they'd see these facies stacked on top of one another, from sandstone, to mudrock, to carbonate, and conclude that the water got progressively deeper through time to account for the facies changes. It would be reasonable to interpret this conformable succession as recording a transgression. In Core 2, they would see the opposite trend: a facies succession that records progressive shallowing, reasonably interpreted as a regression.

Please watch this video below for an excellent explanation.

YouTube Video

This interactive element is not available in this format. Visit the link or scan the QR code to access it online.

View this YouTube Video online

As seen in the video above, Walther's Law 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.

Key Terms

- **regression** – a geologic process in which the sea level lowers, exposing previously submerged areas and leading to the deposition of progressively shallower-water sediments over deeper-water sediments
- **sequence** – a relatively conformable package of sedimentary strata bounded above and below by unconformities (erosional or non-depositional surfaces) that record a cycle of sea-level change and sediment deposition
- **transgression** – a geologic process in which the sea level rises, resulting in progressively deeper-water sediments being deposited over shallower-water sediments
- **Walther's Law** – a principle stating that the vertical order of sedimentary facies in a stratigraphic sequence reflects the lateral changes that existed at the time of deposition, facies that are found adjacent to one another in space will succeed one another vertically in the rock record

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6.8: Stratigraphic Correlation

What is 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.

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, smaller 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.

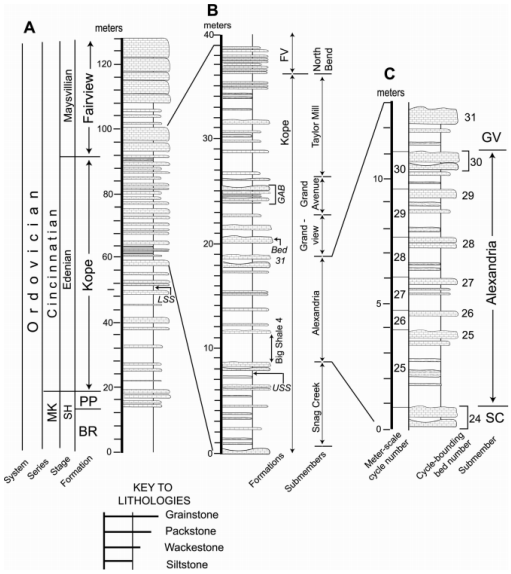


Figure 6.8.1: Upper Ordovician stratigraphy of the Cincinnati, Ohio region. A) Generalized stratigraphy of uppermost Mohawkian and lower Cincinnatian Series. BR Bromley Shale; LSS lower strophic Sowerbyella rugosa interval; MK Mohawkian Series; PP Point Pleasant Formation; SH Shermanian Stage. B) Stratigraphy 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 by 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

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.

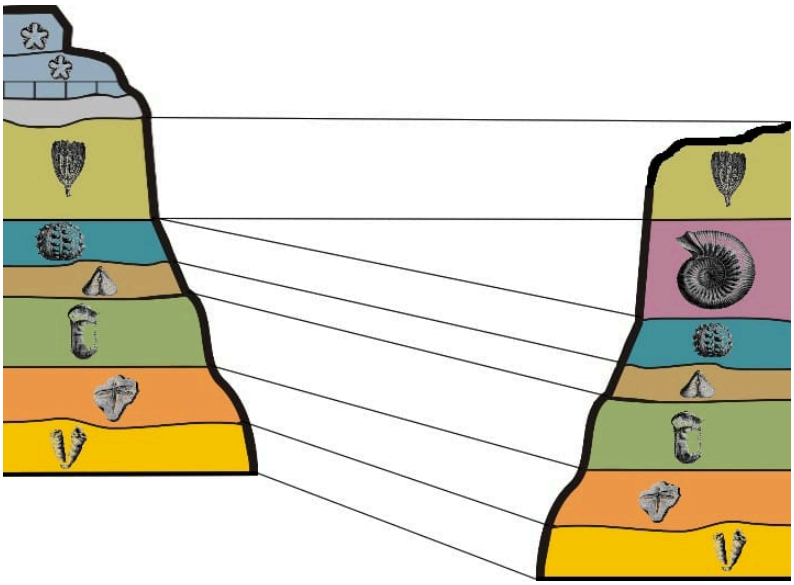


Figure 6.8.2: A simple example of biostratigraphic correlation (CC BY-SA 3.0; By דקי – אני יצרתי, https://commons.wikimedia.org/w/index.php?curid=17935309)

Index Fossils

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 on Wikipedia.

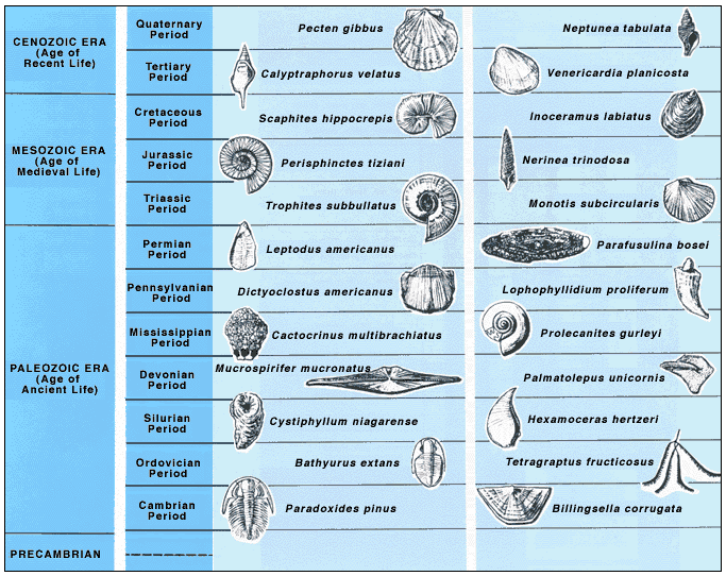


Figure 6.8.3: Some common index fossils (Source: USGS)

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.

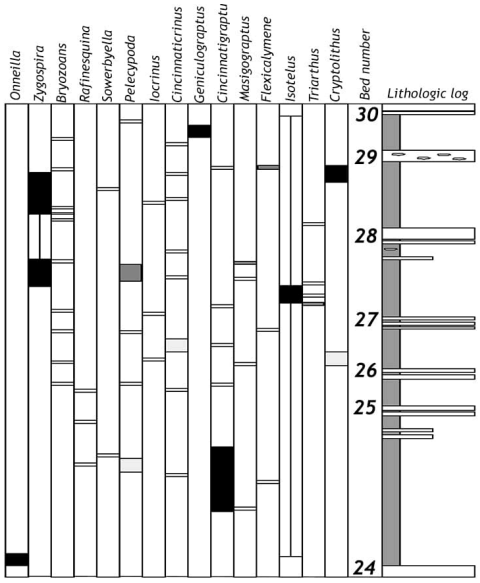


Figure 6.8.4: Fossil occurrences, stratigraphically, within the Alexandria Submember of the Kope Formation. This data was compiled from studies at six sites. Fossil occurrences at 5-6 sites are indicated with black bars. Gray bars identify fossils found at 3-4 sites. Light gray bars indicate fossils 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.

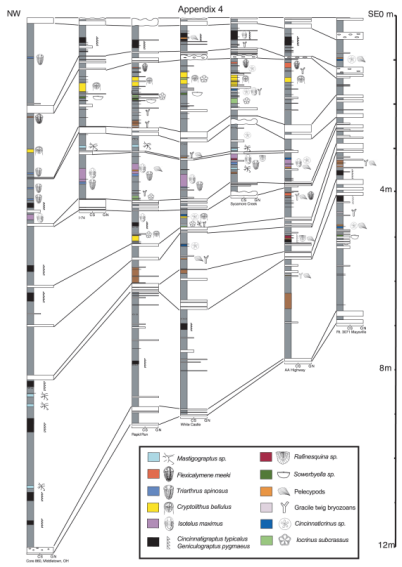


Figure 6.8.5: A lithostratigraphic and biostratigraphic interpretation of mudstone (gray) and limestone (white) layers within the Alexandria Submember of the Kope Formation in the Cincinnati/northern Kentucky region. Each column represents a different location across a transect covering over 50 miles 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).

Abundance Zones or Lack Thereof: Fossil Lagerstätten, Epiboles, and Outages



Figure 6.8.6: 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 6.8.7: 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 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.

Key Terms

- **biostratigraphy** – a branch of stratigraphy that uses fossil content and fossil assemblages to divide and correlate rock layers, helping to establish their relative ages
- **chronostratigraphy** – the study of rock strata in relation to geologic time, focusing on assigning absolute or relative ages to rock units and organizing them into a standardized time framework
- **correlation** – the process of demonstrating equivalence or continuity between rock units or strata at different geographic locations, based on lithology, fossil content, age, or other stratigraphic markers
- **epibole** – the sudden, widespread appearance (or disappearance) of a fossil species or assemblage within the stratigraphic record, often used as a marker horizon for correlation
- **formation** – a fundamental lithostratigraphic unit of rock that is mappable, has a distinct set of characteristics (lithology, composition, etc.), and is thick and extensive enough to be recognized at the Earth’s surface or in the subsurface
- **group** – a lithostratigraphic unit consisting of two or more related formations that share similar geologic features and are grouped together for convenience in mapping and description
- **lithostratigraphy** – the branch of stratigraphy that classifies rock units based on their physical and lithologic characteristics, without direct reference to age
- **magnetostratigraphy** – a stratigraphic method that uses the record of Earth’s magnetic field reversals preserved in rocks to establish ages and correlate strata
- **member** – a lithostratigraphic subdivision of a formation, smaller than a formation but still mappable, that has distinctive properties
- **outage** – in stratigraphy, more commonly referred to as a hiatus, it is a gap in the geologic record representing a period of non-deposition or erosion between rock layers
- **supergroup** – a lithostratigraphic unit that consists of two or more groups, representing an even larger-scale rock package, useful for describing very thick or regionally extensive stratigraphic successions

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6.9: Key Points

1. By studying bedforms in modern environments we are able to look at bedforms in sedimentary beds and determine what that past depositional environment was like.
2. To describe stratified rocks geologists examine the bedding, color, texture/fabric, composition, and fossil content of the rocks.
3. Fossils are remnants of organisms (actual organism parts or signs of activity) over 11,700 years old and are used to help date the rocks they are found in.
4. Facies is the set of characteristics that reflect the conditions in which a sedimentary rock was deposited.
5. Facies can be described using observable characteristics such as color, composition, texture, sediment structures, fossils, association, and form.
6. There are three categories of depositional environments: terrestrial/continental, transitional/coastal, and marine.
7. Each depositional environment produces sediment and/or rocks with unique color, composition, texture, structure(s), fossil content, and more.
8. Stratigraphy, layers of sediment accumulating over time, is a main focus of sedimentary geology.
9. Basins accumulate sediment over time. The sediment captures the geologic history of the area as beds combine to form formations, formations combine to form groups, and groups combine to form the stratigraphic record of that basin.
10. Predictable patterns in sedimentary rocks are called sequences and recognizing them in a stratigraphic section allows for better understanding of the geologic history of the area.
11. The amount of space in a past or current basin is calculated as you consider the rate of tectonic subsidence, rate of sea level change, rate of sedimentation, and rate of change in water depth in the basin.
12. Transgressions indicate a drop in relative sea level and are reflected as sedimentary beds coarsen as you go upward in a stratigraphic section.
13. Regressions indicate a rise in relative sea level and are reflected as sedimentary beds become more fine grained as you go upward in a stratigraphic section.
14. Correlation, or the connecting of layers across distances can be done using rock type to connect depositional environments, or fossils, radiometrically dated materials, or magnetozones to connect by age.

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6.10: Further Reading

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Brett, C. E. & Baird, G. C. (eds) (1997). Paleontological Events. Stratigraphic, Ecological, and Evolutionary Implications. xvii + 604 pp. New York: Columbia University Press.

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7: GEOLOGIC STRUCTURES

Introduction

Crustal deformation occurs when applied forces exceed the internal strength of rocks, physically changing their shapes. These forces are called stress, and the physical changes they create are called strain. Forces involved in tectonic processes as well as gravity and igneous pluton emplacement produce strains in rocks that include folds, fractures, and faults. When rock experiences large amounts of stress and breaks, energy is released in the form of seismic waves, commonly known as an earthquake. Whether a rock respond by folding or faulting (breaking) tells geologists a lot about the conditions it was under at the time stress was applied and helps them construct the geologic history of an area.

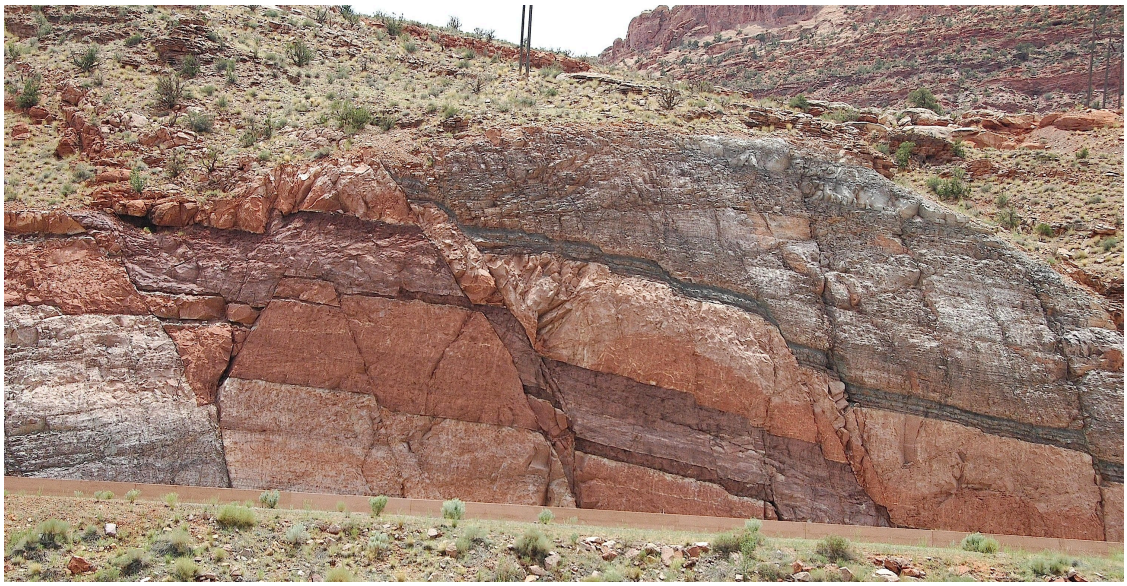


Figure 7.1: Example of normal faulting in an outcrop of the Pennsylvanian Honaker Trail Formation near Moab, Utah.

7.1: Tectonic Movement

7.2: Geological Maps and Cross-Sections

7.3: Stress, Strain, and Deformation

7.4: Folds

7.5: Faults

7.6: Key Points

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7.1: Tectonic Movement

Motion at Plate Boundaries

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.

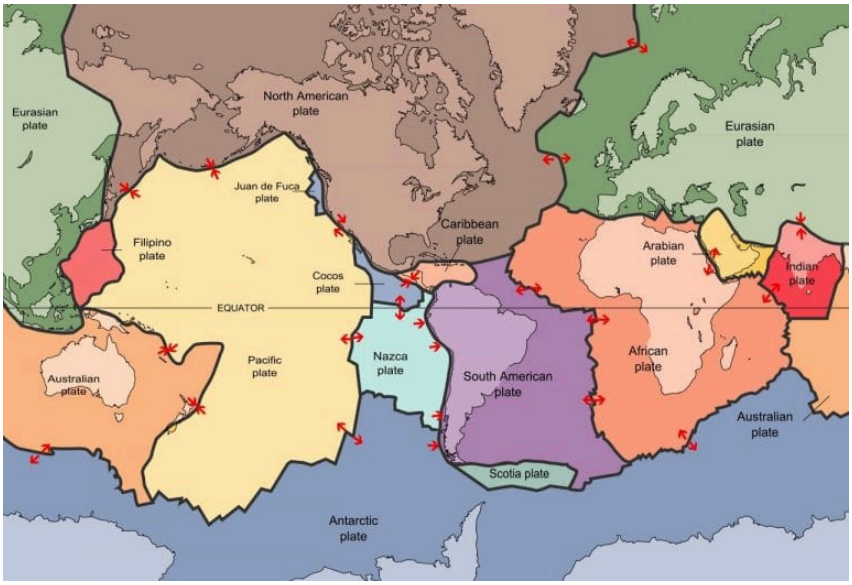
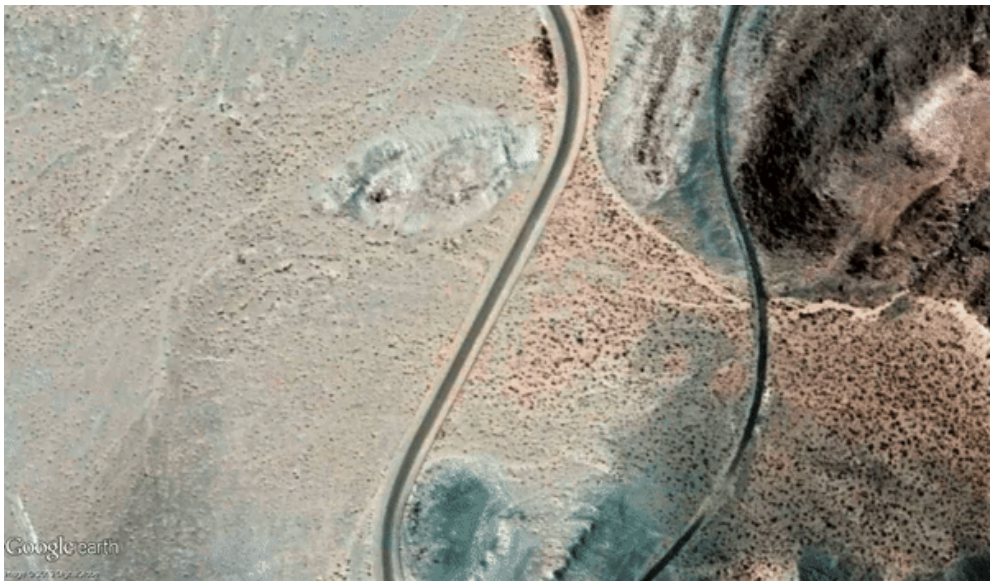


Figure 7.1.1: Tectonic Plate Boundary Map. Arrows at the plate boundaries indicate the direction of movement along that particular boundary. (Public domain; from Wikipedia.)

Displacement in Ridgecrest, California

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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7.2: Geological Maps and Cross-Sections

Maps

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

Cross-sections

Cross-sections are subsurface interpretations made from surface and subsurface measurements. Maps display geology in the horizontal plane while cross-sections show subsurface geology in the vertical plane.

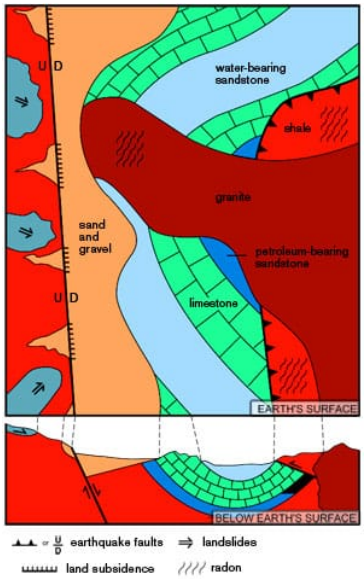


Figure 7.2.1: Example of a geologic map (top) and its corresponding geologic cross-section. USGS (After R. L. Bernknopf et al., 1993).

Strike and Dip

Geologists use a special symbol called strike and dip to represent inclined beds. Strike and dip map symbols look like the capital letter *T*, with a short trunk and extra-wide top line. The short trunk represents the dip and the top line represents the strike. **Strike** is the direction the bed travels. **Dip** has two components.

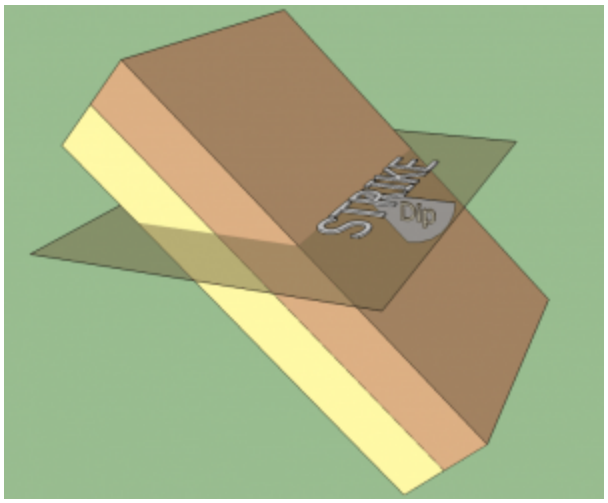


Figure 7.2.2: “Strike” and “dip” are words used to describe the orientation of rock layers with respect to North/South and Horizontal.

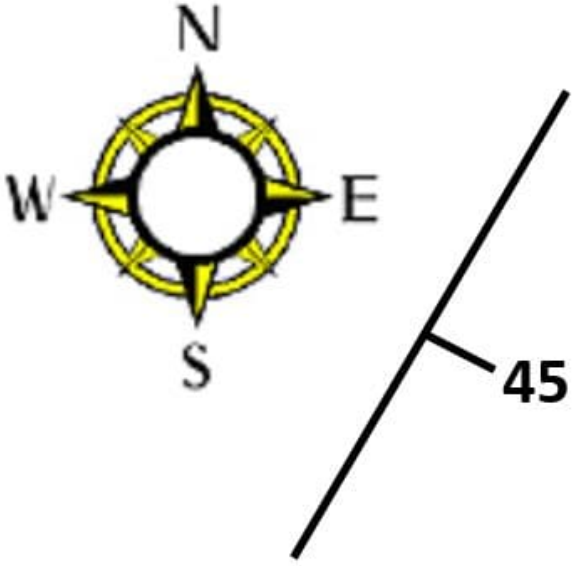


Figure 7.2.3: Attitude symbol on a geologic map (with compass directions for reference) showing strike of N30^{O}E and dip of 45 to the SE.

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

Key Terms

- **cross section** – a vertical slice or profile through the Earth, showing the arrangement and relationships of rock layers, structures, and features beneath the surface
- **dip** – the angle at which a rock layer, fault, or other planar feature is inclined from the horizontal, measured downward in the direction of maximum slope
- **map** – a two-dimensional representation of the Earth’s surface used to show the distribution of rock units, structures, and landforms
- **strike** – the compass direction of the line formed by the intersection of a rock layer or planar feature with a horizontal surface

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7.3: Stress, Strain, and Deformation

Stress

Stress is the force exerted per unit area. There are three types of stress: **tensional**, **compressional**, and **shear** [1]. Tensional stress involves forces pulling in opposite directions, which results in strain that stretches and thins rock. Compressional stress involves forces pushing together, and the compressional strain shows up as rock folding and thickening. Shear stress involves transverse forces; the strain shows up as opposing blocks or regions of the material moving past each other.

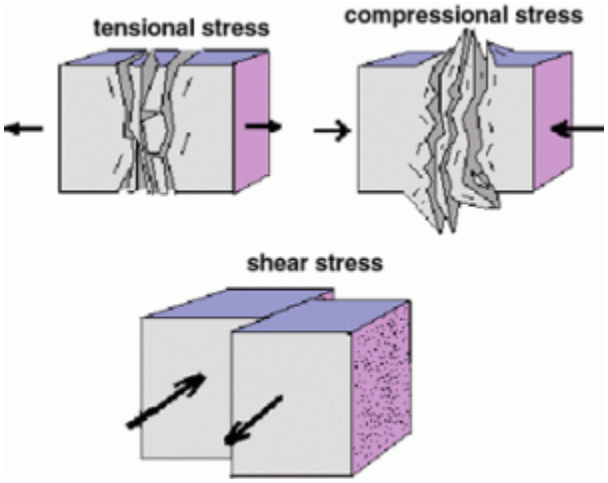


Figure 7.3.1: Types of stress. Clockwise from top left: tensional stress, compressional stress, and shear stress, and some examples of resulting strain.

Strain

Strain is the physical change that results in response to stress. When the applied stress is greater than the internal strength of rock, strain results in the form of deformation of the rock caused by the stress. Strain in rocks can be represented as a change in rock volume and/or rock shape, as well as fracturing the rock and is described as elastic, ductile, or brittle deformation.

Elastic deformation is a strain that is reversible after the stress is released. For example, when you stretch a rubber band, it elastically returns to its original shape after you release it. **Ductile deformation** occurs when enough stress is applied to a material that the changes in its shape are permanent, and the material is no longer able to revert to its original shape. For example, if you bend a metal bar too far, it can be permanently bent out of shape. The point at which elastic deformation is surpassed and strain becomes permanent is called the **yield point**. **Brittle deformation** is another critical point of no return when rock integrity fails and the rock fractures under increasing stress.

The type of deformation a rock undergoes depends on pore pressure, strain rate, rock strength, temperature, stress intensity, time, and confining pressure. For example, applying stress slowly makes it is easier to bend a piece of wood without breaking it. Rock strength measures how easily a rock deforms under stress. Shale has low strength and granite has high strength. Decreasing the temperature makes materials more rigid and susceptible to brittle deformation. On the other hand, heating materials makes them more ductile and less brittle. For example, heated glass can be bent and stretched.

Table showing the relationship between factors operating on rock and the resulting strains:

Factor	Strain Response
Increase Temperature	More Ductile
Increase Strain Rate	More Brittle

Increase Rock Strength	More Brittle

Table showing types of stress and resulting strain. Fault types are discussed later in this chapter but are still documented in this table.

Type of Stress	Associated Plate Boundary type (see Ch. 2)	Resulting Strain	Associated fault types
Tensional	divergent	Stretching and thinning	Normal
Compressional	convergent	Shortening and thickening	Reverse
Shear	transform	Tearing	Strike-slip

References

1. Earle S (2015) Physical geology OER textbook. BC Campus OpenEd

Key Terms

- **brittle deformation** – the breaking or fracturing of rocks a a result of exposure to stress
- **compressional stress** – a type of stress that squeezes rocks together, shortening and thickening the crust
- **ductile deformation** – the bending, flowing, or folding of rocks without breaking when subjected to stress, usually at high temperature and pressure
- **elastic deformation** – a temporary change in the shape or size of a rock that is reversible; the rock returns to its original form once the stress is removed
- **shear stress** – a type of stress that causes parts of a rock body to slide past each other in opposite directions
- **strain** – the measure of change in a rock's shape or volume in response to applied stress
- **stress** – the force applied to a rock per unit area, which can cause deformation
- **tensional stress** – a type of stress that pulls rocks apart, stretching and thinning the crust
- **yield point** – the limit beyond which a rock will no longer deform elastically; permanent deformation (ductile or brittle) occurs once stress passes this point

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7.4: Folds

Fold Creation

Compressional stress is the dominant force in mountain building. As plates collide, compressional stress causes the crust to buckle 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 7.4.1: 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 three general categories: monoclines, anticlines and synclines.

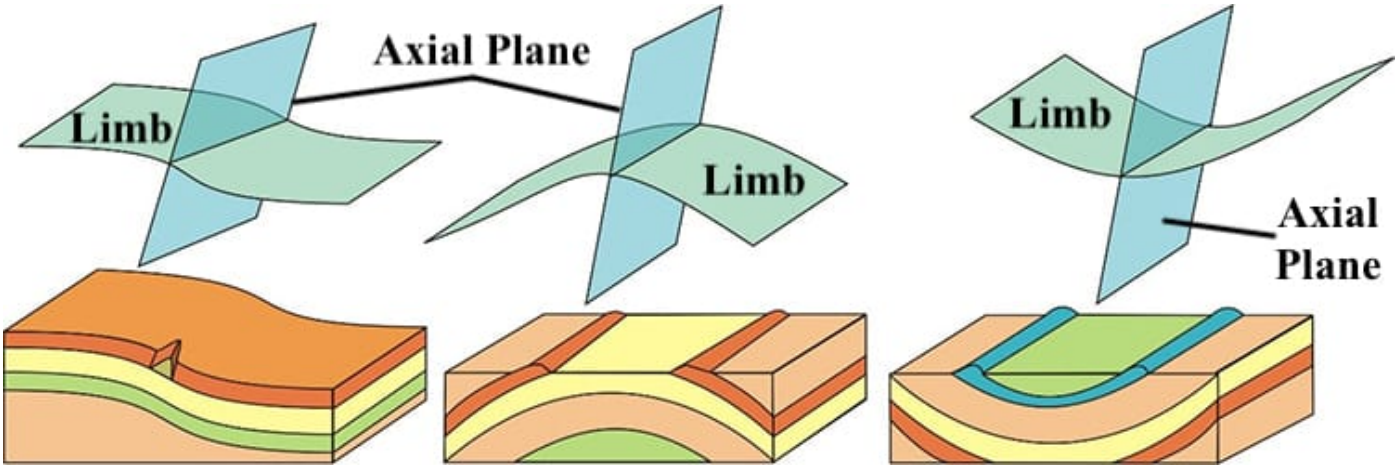


Figure 7.4.2: Monoclines, anticlines, and synclines are the three basic types of folds. Here they are shown in block diagrams, as well as in diagrams showing their essential geometry: the relationship of the strata to the **axial plane**.

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 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 7.4.3: 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 and be called **overturned**. A fold will become increasingly asymmetric as the axial plane tips toward 90° from vertical at which time it will be classified as **recumbent**.

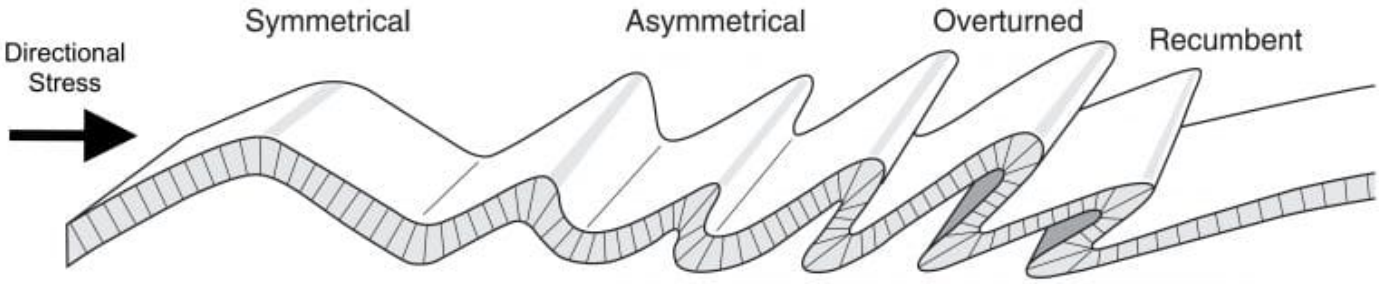


Figure 7.4.4: Directionally applied stress will result in folds becoming more asymmetrical as the axial plane dips away from the vertical. (Modified after unknown source.)

Overturned Fold



Figure 7.4.5: Overturned anticline-syncline pair in marble and quartzite. (By: Marli Miller Photography.)

Recumbent Fold



Figure 7.4.6: Recumbent Fold, Port au Port Peninsula, Newfoundland. (CC BY-NC 4.0; from: https://openeducationalberta.ca/introductorystructuralgeology/chapter/e-folded-surfaces/)

Key Terms

- **anticline** – a fold in rock layers that is arch-shaped, with the oldest rocks located at the core of the fold and the limbs dipping away from the center
- **axial plane** – an imaginary surface that divides a fold into two roughly symmetrical halves, passing through the center of each layer
- **limb** – the side or arm of a fold, consisting of rock layers that dip away from (or toward) the center of the fold
- **monocline** – a simple fold in which otherwise horizontal rock layers bend in one direction, creating a single step-like feature
- **overturned fold** – a fold in which one or both limbs are tilted beyond vertical, so that the rock layers on one limb are upside down.

- **recumbent fold** – a fold that has been rotated so much that its axial plane is nearly horizontal, causing the limbs to be almost parallel
- **syncline** – a fold in rock layers that is trough-shaped, with the youngest rocks located at the core of the fold and the limbs dipping inward toward the hinge

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7.5: Faults

Fault Creation

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. Folding, as discussed previously, 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/thrust fault is caused by compressional stress at convergent plate boundaries. In reverse/thrust faults, the hanging wall has moved up relative to the footwall.

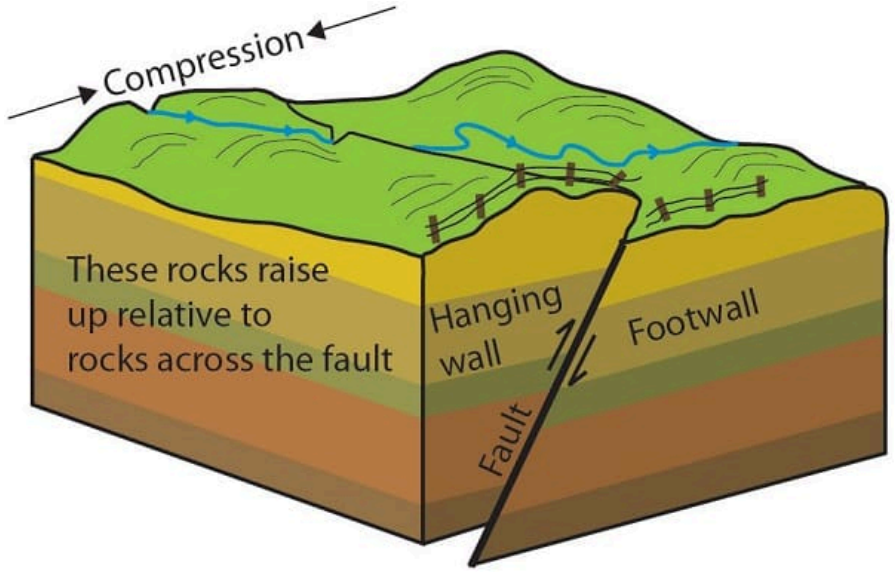


Figure 7.5.1: A reverse fault formed by the tectonic force of compression along a convergent plate boundary. (Public Domain; By: Trista L. Thornberry-Ehrlich, Colorado State University, From: National Park Service.)

The difference between a reverse and thrust fault is the angle of the fault plane. A **reverse fault** 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.

Interactive Content

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Figure 7.5.2: Reverse Fault

Interactive Content

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Figure 7.5.3: Thrust Fault

Compare the examples of a reverse and thrust fault below, with the 3D images above.

Reverse Fault



Figure 7.5.4: 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 7.5.5: 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 normal faults, the hanging wall has moved down relative to the footwall.

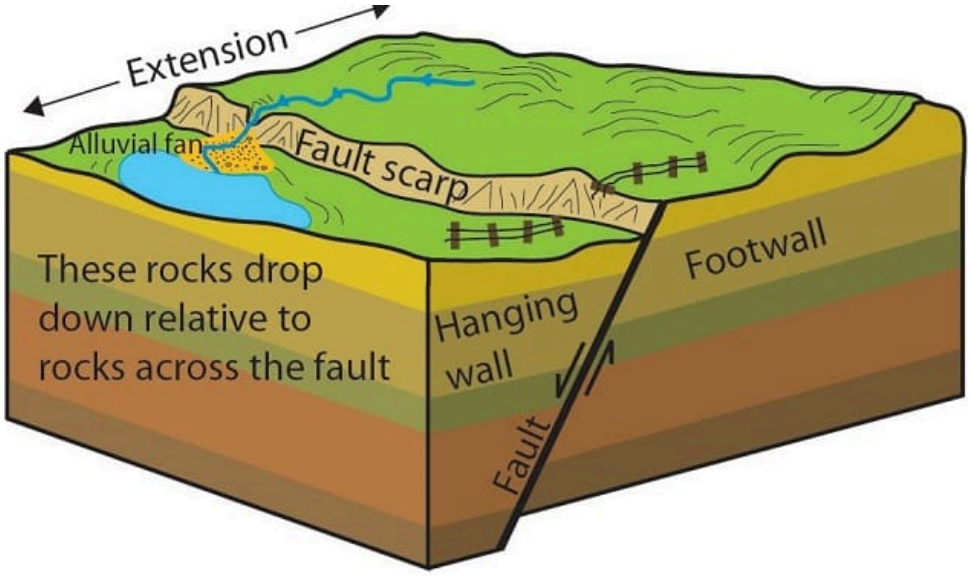


Figure 7.5.6: A normal fault formed by the tectonic force of extension (tensional stress) at a divergent plate boundary. (Public Domain; By: Trista L. Thornberry-Ehrlich, Colorado State University. From: National Park Service.)

Explore this 3D block structure of a normal fault. Once the models load, you can grab them to rotate in 3D. Then compare the 3D block structure to the images of normal faults below it.

Interactive Content

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View this Interactive Content online

Figure 7.5.7: Normal Fault



Figure 7.5.8: 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 7.5.9: 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.

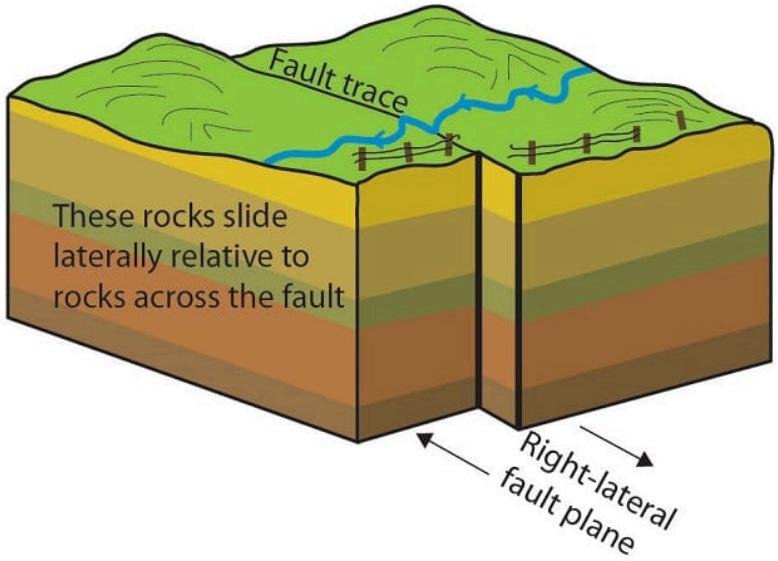


Figure 7.5.10: A strike-slip fault formed by shear tectonic force along a transform plate boundary. (Public Domain; By: Trista L. Thornberry-Ehrlich, Colorado State University. From: National Park Service.)

Let's take a look at another one of these 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.



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 7.5.11: 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.)

Key Terms

- **footwall** – the block of rock that lies beneath an inclined fault plane
- **hanging wall** – the block of rock that lies above an inclined fault plane
- **normal fault** – a type of dip-slip fault where the hanging wall moves downward relative to the footwall, caused by tensional stress
- **reverse fault** – a type of dip-slip fault where the hanging wall moves upward relative to the footwall, caused by compressional stress
- **strike-slip fault** – a fault in which the main movement is horizontal and parallel to the fault's strike, caused by shear stress
- **thrust fault** – a low-angle reverse fault (typically dipping less than 30°) where older rocks are pushed up and over younger rocks due to strong compressional forces

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7.6: Key Points

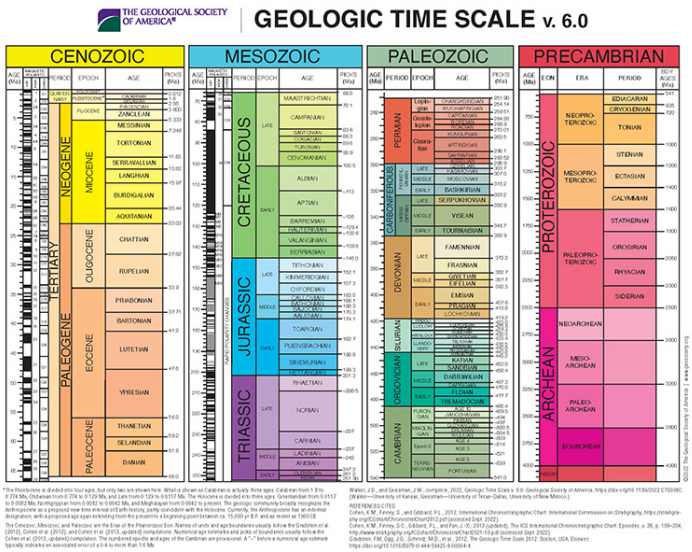
1. Stress generated at plate boundaries results in major changes to Earth's landscape over long periods of time.
2. Geologic maps and cross sections show rock types and geologic structures from the top and side view respectively.
3. Stress is the force placed on an object and strain is the physical change that results from stress, often called deformation.
4. Compressional stress occurs at convergent plate boundaries. Tensional stress occurs at divergent plate boundaries. Shear stress occurs at transform boundaries.
5. Ductile deformation results in the bending of rock and occurs when rock is warm.
6. Brittle deformation results in the breaking of rock and occurs when rock is cold.
7. Types of deformation include the folding (bending) of rock materials which occurs when the rock is warmer at depth and the faulting (breaking) of rock materials which occurs when the rock is cooler near the surface.
8. The fault type generated is a result of the type of stress materials are exposed to which often relates to the type of plate boundary present. Normal faults result from tensional stress at divergent boundaries. Reverse faults result from compressional stress at convergent boundaries. Strike-slip faults result from shear stress at transform boundaries.

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8: GEOLOGIC TIME

Introduction

Geologic time is the framework scientists use to understand Earth's 4.6-billion-year history and the sequence of events that have shaped the planet. Geologists measure time in two main ways: relative time and absolute time. Relative time places rocks and events in a sequence without assigning specific numerical ages, relying on principles to determine which layers are older or younger. Absolute time, in contrast, provides a numerical age, often determined through radiometric dating, which measures the decay of elements in minerals to calculate when that mineral, and therefore the rock, formed. Together, relative and absolute dating allow geologists to reconstruct Earth's history producing the geologic time scale pictured above to correlate events and interpret past environments.



8.1: Earth's Age and Abbreviating Time Intervals

8.2: Uniformitarianism and Relative Age Dating

8.3: Unconformities

8.4: Fossils and Age Dating

8.5: Absolute Age Dating

8.6: The Geological Time Scale

8.7: Key Points

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8.1: Earth's Age and Abbreviating Time Intervals

Earth's Age

The Earth is 4.6 billion years old.

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.

Have a look at the image below (or download this PDF file), which illustrates 5,000 individual zeros.

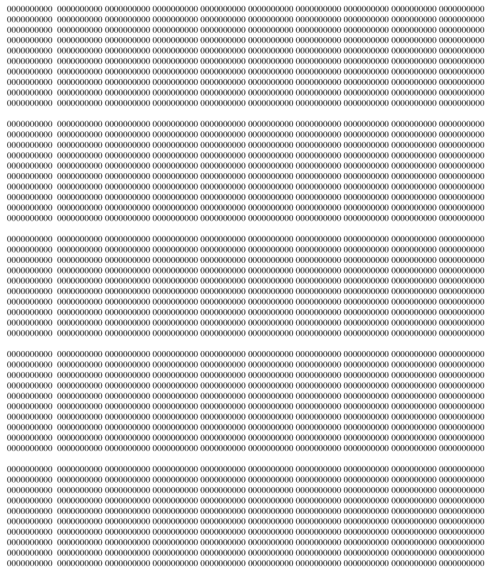


Figure 8.1.1: Image by Jonathan R. Hendricks. (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License.)

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...

Abbreviating Time Intervals

Geologists measure events in Earth's history in years before the present date. They use certain conventions for abbreviating intervals of time.

- 1 thousand years = kilo-annum, abbreviated "ka"
- 1 million years = mega-annum, abbreviated "Ma"
- 1 billion years = giga-annum, abbreviated "Ga"
- 1 million years ago is abbreviated "mya"
- 1 thousand years ago is abbreviated "kya"

Geologic Time Scale

The Geologic Time Scale on the previous page, as well as others in this text, does a great job organizing the different time periods and events of Earth's history, but does not show these events at their proper scale, making it challenging to conceptualize the magnitude of these time periods relative to one another. The sprial time scale below is drawn to scale. Take a few minutes to study it and gain a greater appreciation for geologic time.

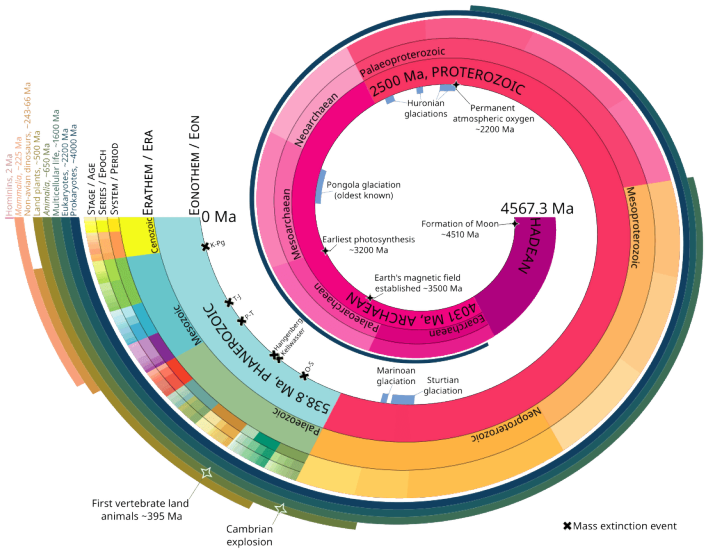


Figure. Sprial Geologic Time Scale to scale. Image by Jarred C Lloyd, CC BY 4.0 <https://creativecommons.org/licenses/by/4.0>, via Wikimedia Commons.

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8.2: Uniformitarianism and Relative Age Dating

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 8.2.1: 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 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.



Figure 8.2.2: 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 million years ago). 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**.



Figure 8.2.3: 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 (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.

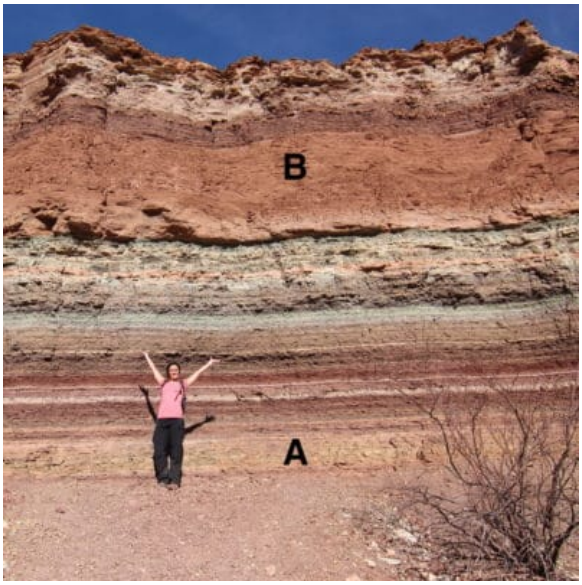


Figure 8.2.4: Sedimentary rock layers exposed at Quebrada de Cafayate, Salta (Argentina). (CC BY-SA 2.0, https://commons.wikimedia.org/w/index...curid=24779211; by travelwayoflife https://www.flickr.com/photos/travelwayoflife/6164348161/in/photostream/)

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 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 (in figure 8.2.5 below) live before or after the trilobite from Oklahoma (in figure 8.2.6 below)?



Figure 8.2.5: 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 8.2.6: 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 8.2.7: 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. 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.



Figure 8.2.8: 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.)

Figure 8.2.8 demonstrates this principle as it shows a sequence of Devonian-aged (~380 Ma) rocks exposed at the 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.

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 8.2.9: 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

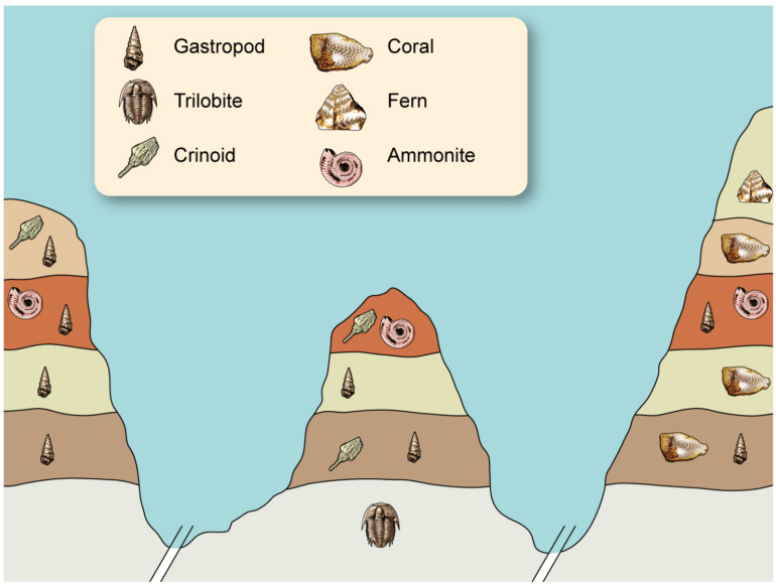


Figure 8.2.10: Sequence and correlation of fossils from different locations. (CC BY-SA; https://commons.wikimedia.org/wiki/File:Fossils.png by: Kurt Rosenkrantz)

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 8.2.11: 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 8.2.12: 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 (e.g. the block on the right dropped down). 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 8.2.13: 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)



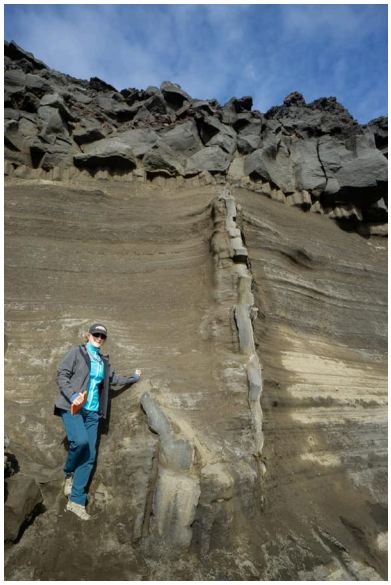


Figure 8.2.14: Above: A thin basaltic dike cross-cutting volcanic sediments of Paleogene age at Lagganulva, on the Isle of Mull, Scotland. (By permission, John Faithful, University of Glasgow, Scotland.) *Below:* Feeder dike intruding tephra with basaltic lava flow above, Iceland. (By: Marli Miller from 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 Figure 8.2.14, 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 8.2.15: What sequence of steps were needed to make this Connecticut outcrop? (Photograph by Gregory Simones; reproduced with permission)



Figure 8.2.16: The sequence of steps needed to make the Connecticut outcrop. (Original photograph by Gregory Simones; reproduced with permission)

Principle of Inclusions

The **principle of 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** (fragment of rock trapped in another type of rock) 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 8.2.17: 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 8.2.18: 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.)

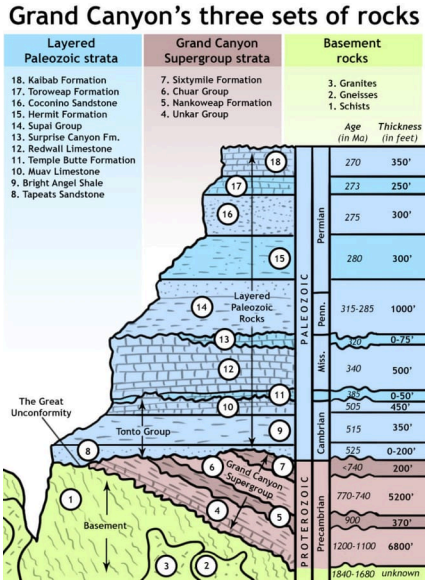


Figure 8.2.19: A schematic cross-section of the strata exposed in the Grand Canyon. (United States Public Domain; Callan Bentley after National Park Service from: https://www.nps.gov/articles/age-of-rocks-in-grand-canyon.htm.)

Figure 8.2.19 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.

Key Terms

- **principle of cross-cutting relationships** – a geologic principle stating that a rock or geologic feature (such as a fault or intrusion) that cuts across another rock body must be younger than the rock it cuts through
- **principle of faunal succession** – the concept that fossils succeed one another in a definite, recognizable order, allowing rock layers to be dated and correlated by the fossils they contain
- **principle of inclusions** – states that any rock fragments (inclusions) contained within another rock must be older than the rock in which they are enclosed.
- **principle of lateral continuity** – sedimentary layers originally extend in all directions until they thin out or are interrupted by a barrier, meaning that separated layers on opposite sides of a valley were once continuous
- **principle of original horizontality** – states that layers of sediment are originally deposited in horizontal or nearly horizontal layers so tilting must have occurred after deposition
- **principle of superposition** – in an undisturbed sequence of sedimentary rocks, the oldest layers are at the bottom and the youngest layers are at the top
- **relative dating** – a method of determining the chronological order of past events by comparing rock layers and fossils, without determining their exact numerical age
- **uniformitarianism** – the principle that the geologic processes we see shaping the Earth today (such as erosion, sedimentation, and volcanism) have operated in the same way throughout Earth's history
- **xenolith** – a fragment of rock enclosed within another, usually a piece of country rock trapped within an igneous intrusion such that the fragment is older than what it is trapped in

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8.3: Unconformities

Unconformities – Missing Time

An **unconformity** represents an interruption in the process of depositing sedimentary layers. Sometimes deposition just stops. Sometimes the area is being actively eroded. Either way, sediment is not depositing so there is nothing recording the geologic events of the time.

There are three types of unconformities. More details are provided below using real-life examples, but these simple diagrams may be useful for initial understanding.

- disconformity - a gap in the geological record that occurs between parallel layers of sedimentary rock
- nonconformity - a gap in the geologic record where sedimentary rock overlies igneous or metamorphic rock
- angular unconformity - a geological feature where younger, horizontal sedimentary rock layers are deposited on top of older, tilted or folded sedimentary rock layers, creating an angular discordance between the two sets of strata

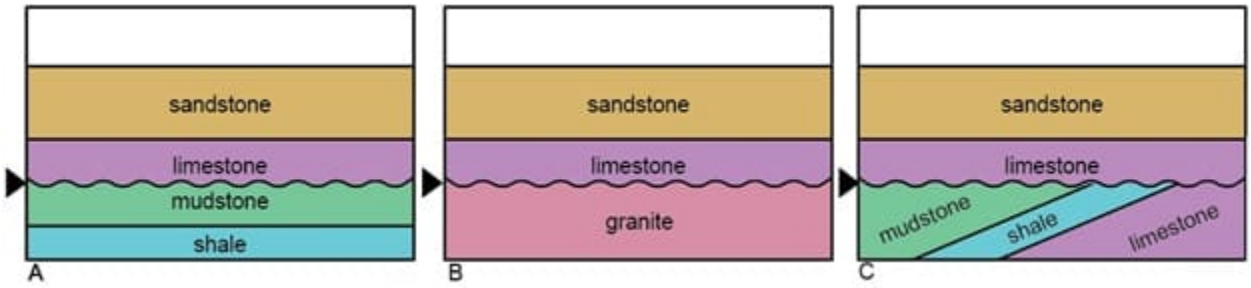


Figure \({8{.}3{.}\eatSpaces{1\): Cross-sections showing the three types of unconformities. A) Arrow indicates a disconformity between two parallel sedimentary layers. B) Arrow indicates a nonconformity between an underlying igneous rock and sedimentary layer. C) Arrow indicates an angular unconformity between underlying tilted sedimentary layers and overlying sedimentary layers. Authored by Kelly Ruppert.

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.

Angular Unconformity

An example of one of the most iconic unconformities is shown in Figure Extra close brace or missing open brace 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.

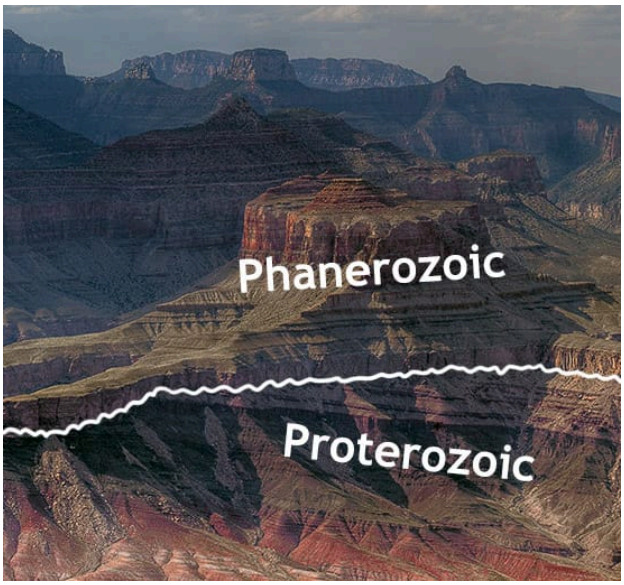


Figure 8.3.2: The great angular unconformity in the Grand Canyon, Arizona. The tilted rocks at the bottom are part of the Proterozoic Grand Canyon Supergroup (aged 825 to 1,250 Ma). The flat-lying rocks at the top are Paleozoic (540 to 250 Ma). The boundary between the two represents a time gap of nearly 300 million years. (Credit: Doug Dolde / Doug Dolde at en.Wikipedia / Public domain. Annotated by Callan Bentley.)

Nonconformity

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 Figure 8.3.3, you see 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.

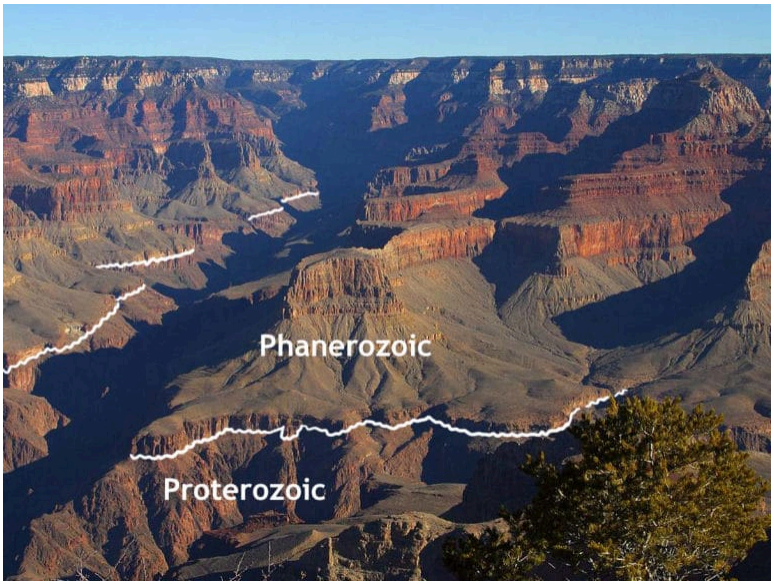


Figure 8.3.3: The view is centered on Sumner Butte, at southeast terminus of Bright Angel Canyon, intersecting with Granite Gorge. (The original uploader was Szumyk at Polish Wikipedia. – http://pdphoto.org/PictureDetail.php?mat=pdef&pg=5506, Public Domain, https://commons.wikimedia.org/w/inde...?curid=5883465. Annotated by Callan Bentley.)

Disconformity

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 Figure 8.3.4.

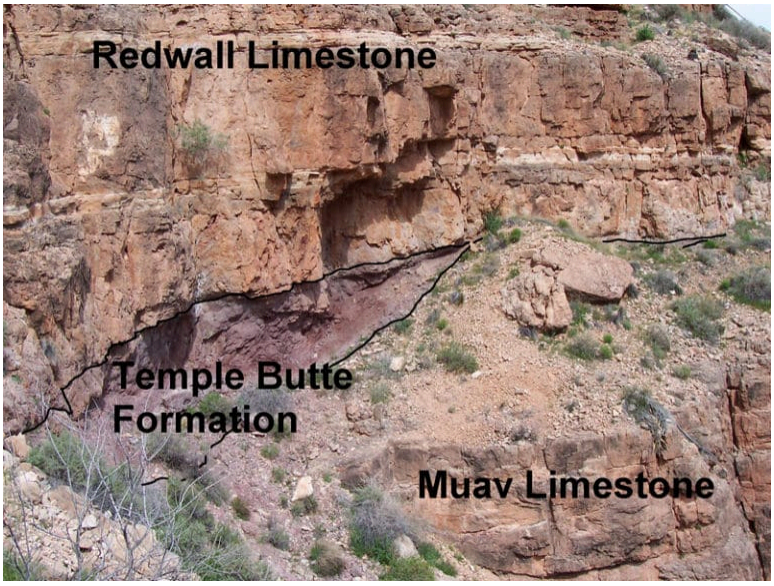


Figure 8.3.4: Redwall, Temple Butte and Muav formations in the Grand Canyon. An erosional surface (disconformity) exists between both the Muav Limestone and the Temple Butte formations as well as between the Temple Butte and the Redwall Limestone formations. (United States Public Domain; National Park Service from: https://upload.wikimedia.org/wikipedia/commons/e/eb/Redwall%2C_Temple_Butte_and_Muav_formations_in_Grand_Canyon.jpg.)

Example 8.3.1: Unconformity Examples

Let's take a look at some additional examples of the different types of unconformities.

Disconformity (sedimentary layers above and below the line are parallel to one another):

Angular unconformity (layers below the unconformity are tilted, layers above the unconformity are horizontal):

Nonconformity (sedimentary layers on top of an igneous unit):

More unconformities:



Figure 8.3.5: Angular unconformity, Telheiro Beach, Portugal, showing Triassic sandstone atop folded, eroded, Carboniferous graywacke and shale. (Photo by Joao Duarte, 2019.)

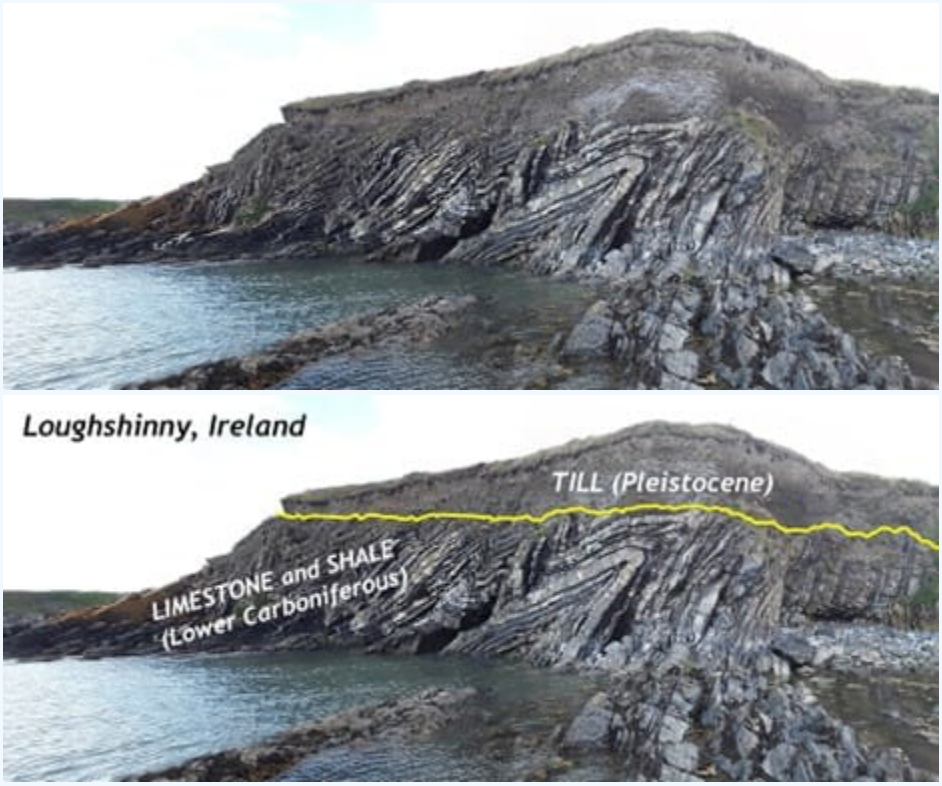


Figure 8.3.6: Angular unconformity in Loughshinny, Ireland, showing flat-lying Pleistocene till overlying folded limestone and shale. (Photo by Jason Loxton, 2018.)



Figure 8.3.7: Carboniferous / Cretaceous angular unconformity at Ense-Bremen, North Rhine-Westphalia, Germany. (Photo by Henning Bergermann; reproduced with permission.)

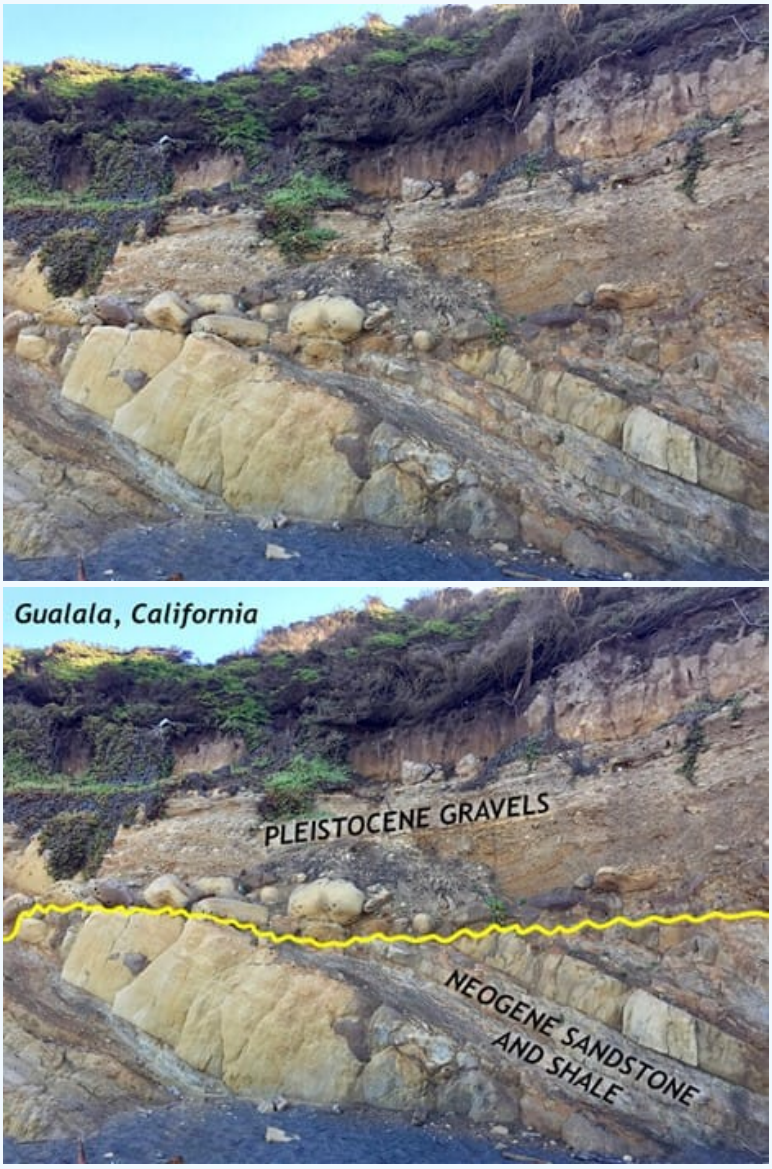


Figure 8.3.8: Angular unconformity marked where tilted layers of Neogene-aged sandstone and shale overlain by Pleistocene gravels (with clam-bored boulder bed just above the unconformity surface). Bonus: a layer of recent colluvium overlies the gravels! (Photo by Seth Finnegan; reproduced with permission.)



Figure 8.3.9: Angular unconformity marked where Jurassic-aged Morrison Formation (shale and sandstone) is folded and eroded, overlain by Quaternary sandstone and conglomerate. (Photo by Elizabeth Bollen; reproduced with permission.)



Figure 8.3.10: Nonconformity between Neoproterozoic andesite (~570 Ma) and overlying Triassic sandstone (~210 Ma). (Photo by Jack Matthews; reproduced with permission.)

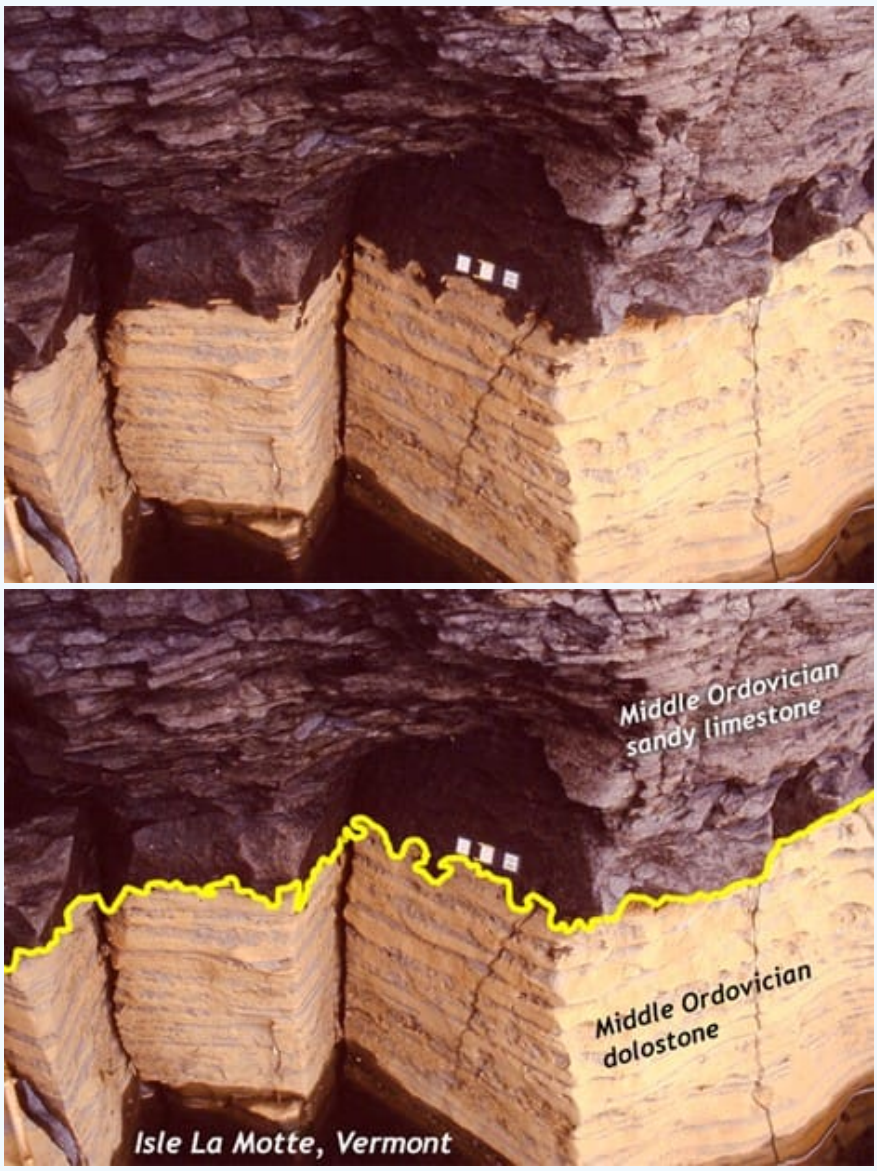


Figure 8.3.11: The Knox unconformity (a disconformity bounding the Sauk and Tippecanoe epeiric sequences) exposed on Isle La Motte, Lake Champlain, Vermont. Providence Island Formation of the Beekmantown Group (tidal dolomite of the Middle Ordovician) is overlain by lower Day Point Formation (also Middle 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.)

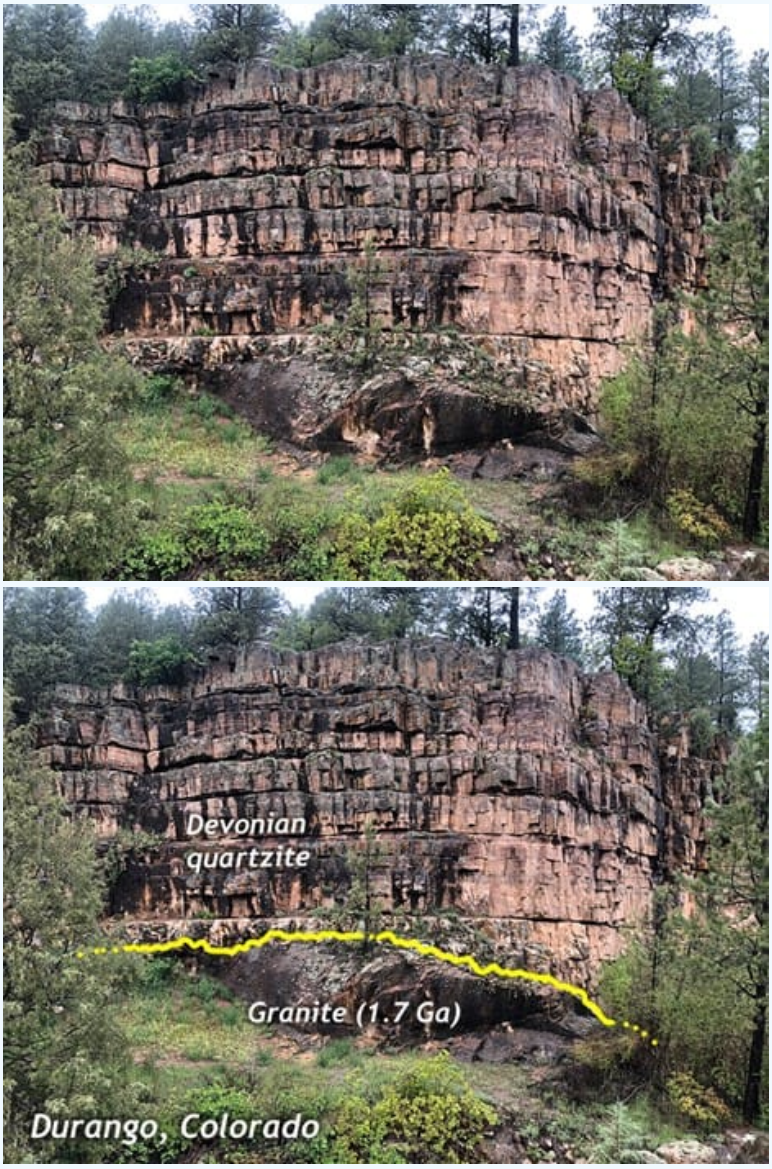


Figure 8.3.12: The “Great Unconformity” (really a nonconformity) north of Durango Colorado. 1.7 Ga Baker’s Bridge granite overlain by late Devonian Ignacio quartzite. (Photograph by Cody Mason; reproduced with permission.)



Figure 8.3.13: Nonconformity in Gallinas Canyon, New Mexico. (Photo by Howell Bosbyshell; reproduced with permission.)

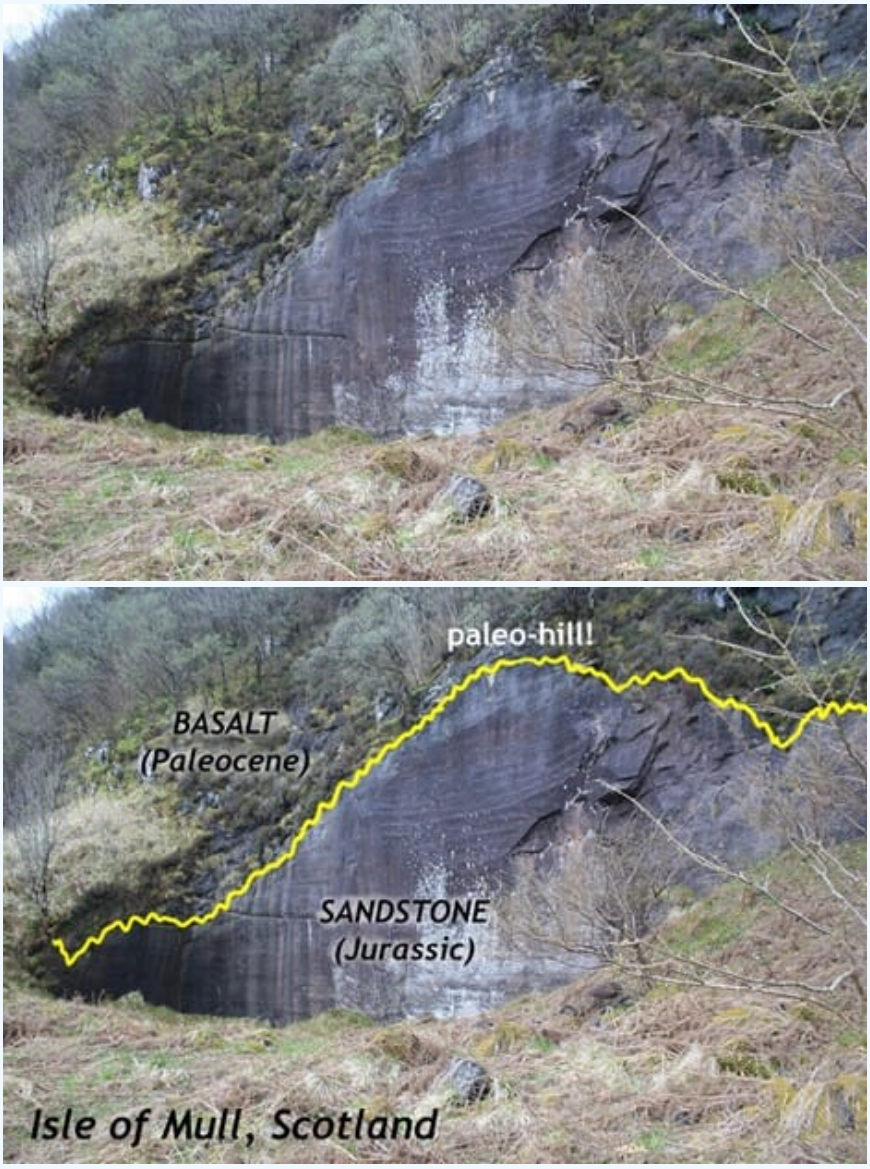


Figure 8.3.14: A paleo-hill of Jurassic sandstone, overlain by Paleocene volcanics mainly basalts. It was quarried for its rock used to make the approach to a local lighthouse on the Isle of Mull. (Photo by James Westland; reproduced with permission.)

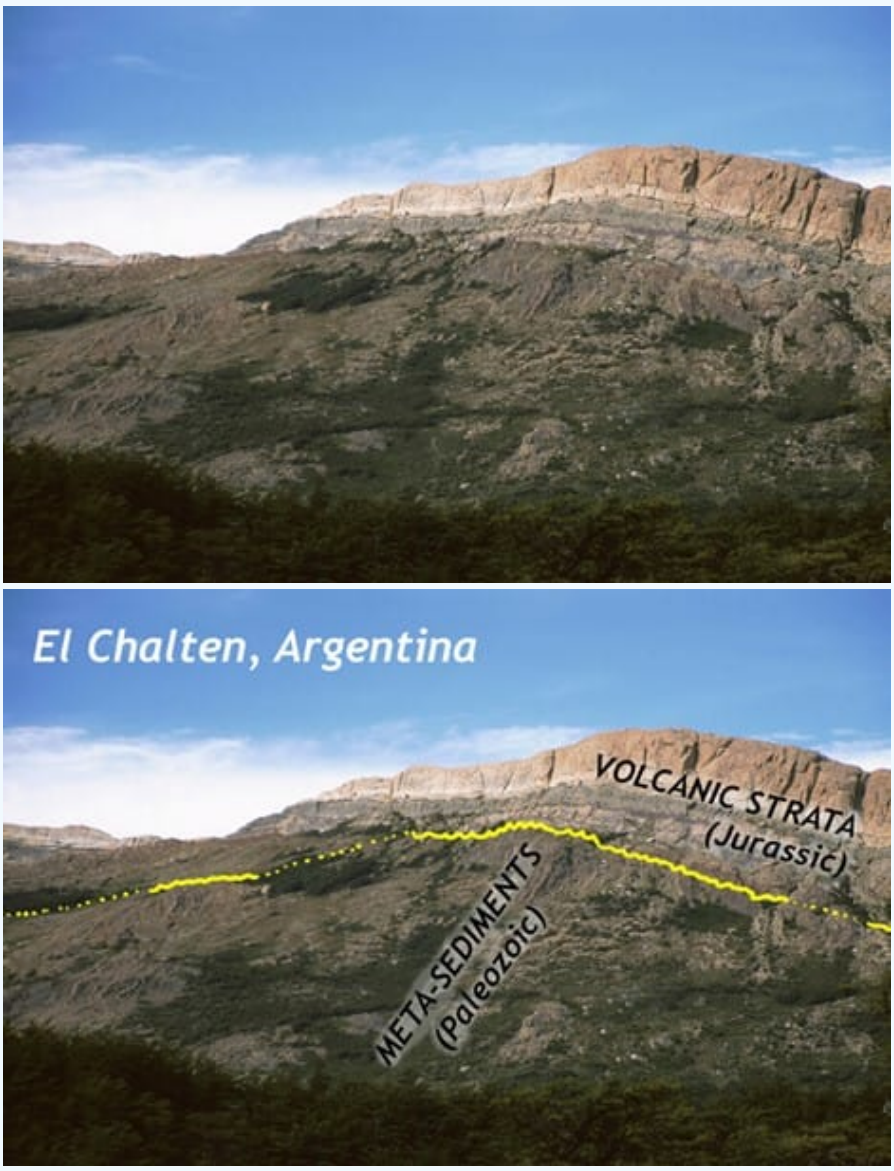


Figure 8.3.15: Angular unconformity at El Chalten, Argentina. Jurassic volcanic strata overlie Paleozoic meta-sedimentary strata. (Photo by Donna Anderson; reproduced with permission.)



Figure 8.3.16: Angular unconformity marked by dipping Paleocene interbedded sands and muds with overlying horizontal Eocene conglomerates. Bighorn Basin, Wyoming. (Photo by Amanda Owen; reproduced with permission.)

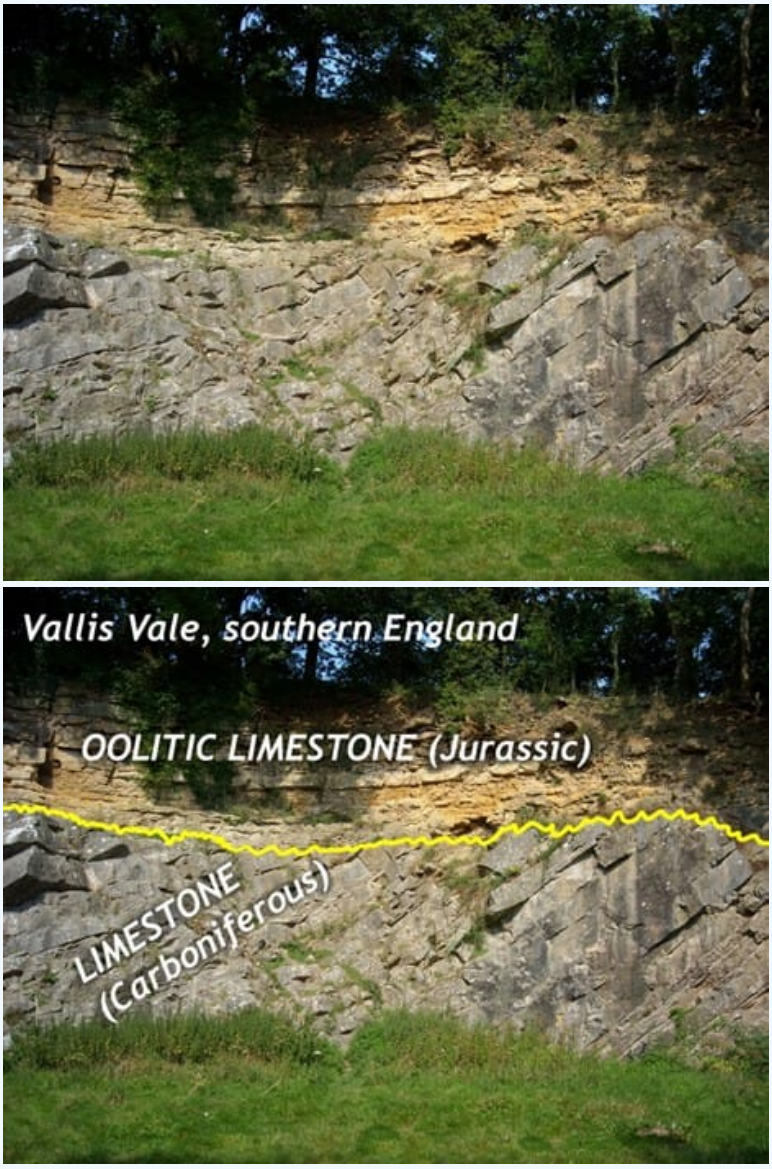


Figure 8.3.17: Probably one of the best examples of an angular unconformity in Southern England. An unconformity is a buried erosional surface separating strata of different ages; it indicates that sediment deposition was not continuous. In this unconformity at the very eastern end of the Mendips, younger 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 8.3.18: 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 8.3.19: Tilted Pleistocene terrace deposits over tilted Miocene Astoria Formation creating an angular unconformity. 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 8.3.20: This is an upside-down unconformity (specifically a disconformity)! 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.)

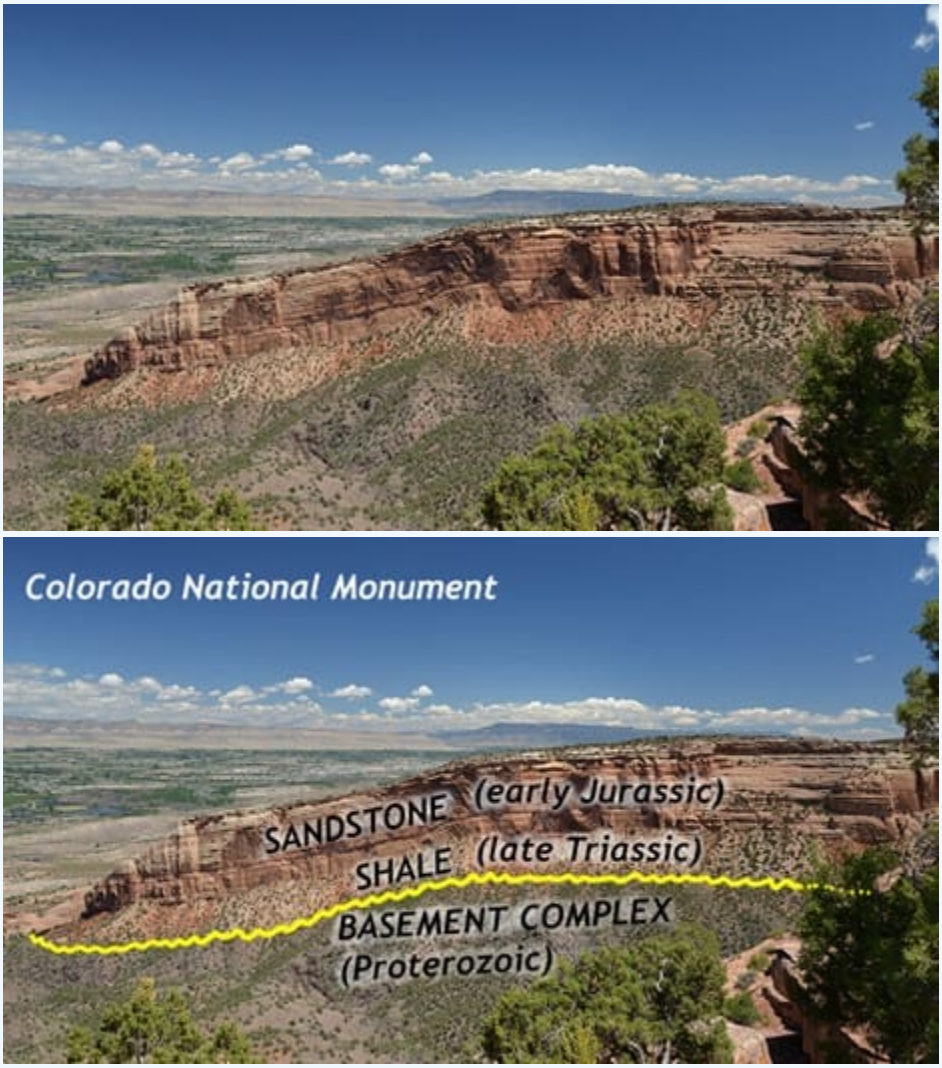


Figure 8.3.21: A monocline overprinting an unconformity. Proterozoic basement rocks (purple, low in canyon) are overlain by the Chinle Formation, with the Wingate Sandstone as the cliff-forming unit. Colorado National Monument. (Photo by Robert Mahon; reproduced with permission.

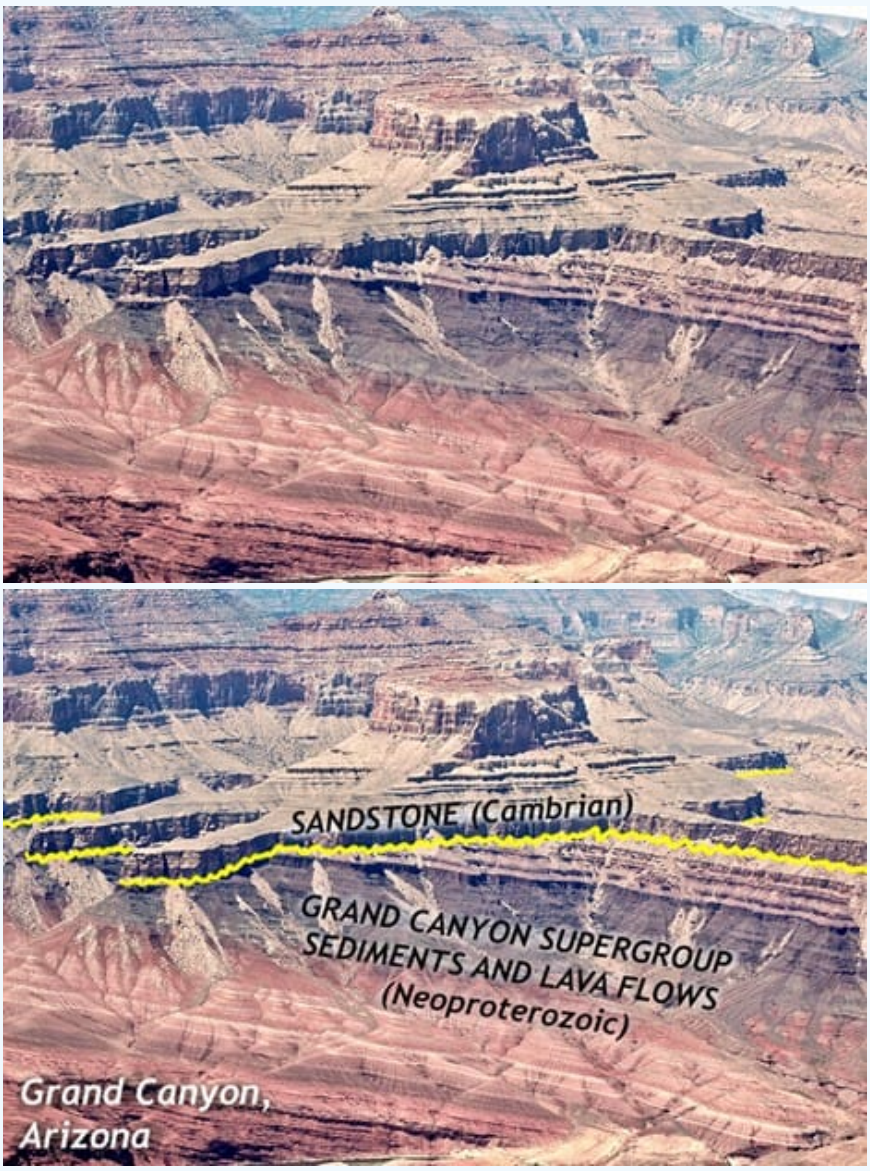


Figure 8.3.22: Angular unconformity in the Grand Canyon. Lower tilted layers are Neoproterozoic strata of the Grand Canyon supergroup (various sedimentary rock and some lava flows). The cliff-forming unit extending horizontally across the field of view is Cambrian-aged Tapeats Sandstone. (Photo by Zoltán Sylvester; reproduced with permission.)



Figure 8.3.23: An angular unconformity created as Pleistocene terrace deposits formed over tilted deltaic deposits of the Eocene Coaledo Formation, Shore Acres State Park, Oregon. (Photo by Lockwood DeWitt; reproduced with permission.)



Figure 8.3.24: 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 8.3.25: 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.)

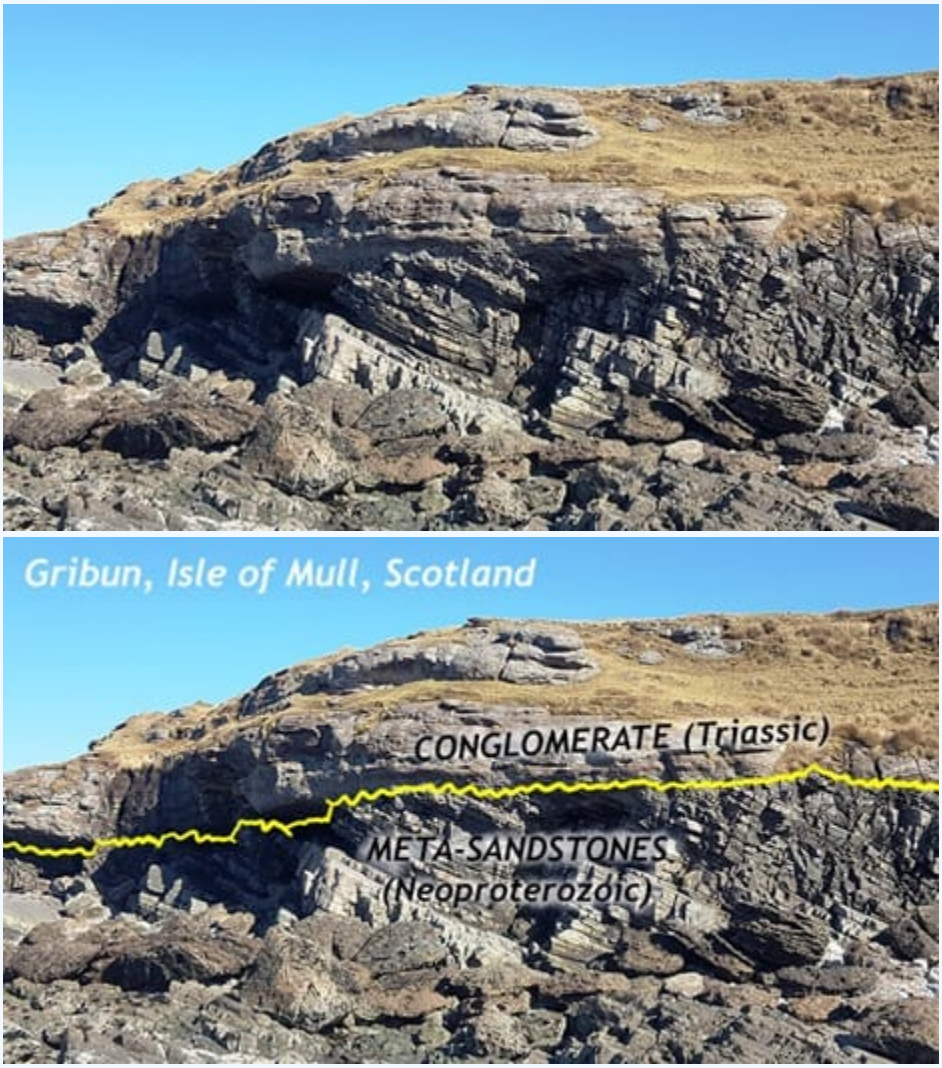


Figure 8.3.26: At Gribun on the Isle of Mull is an angular unconformity. The underlying dipping rocks are Neoproterozoic psammites (meta-sandstones). There are horizontal Triassic conglomerates above. Most of the boulders on the shore are the conglomerate. Higher up, but not seen in this photo, there are also Paleocene basalts. (Photo by James Westland; reproduced with permission.)

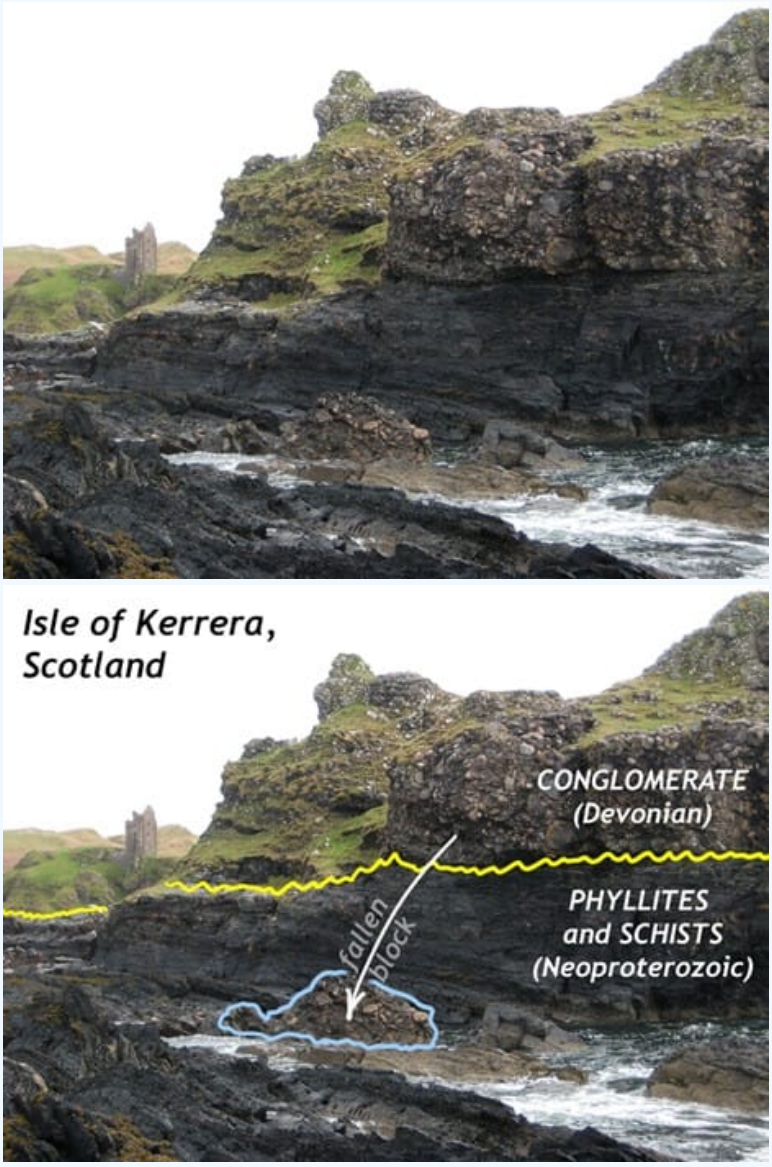


Figure 8.3.27: The classic Scottish angular unconformity between Dalradian metasedimentary slates/phyllites/metalimestones and the overlying basal conglomerate of Late Devonian Old Red Sandstone. This exposure is on the Isle of Kerrera. Gylen Castle ruin for scale. (Photo by Roderick Brown; reproduced with permission.)



Figure 8.3.28: Basal Cambrian arenites (sandstones) on Ercall Granophyre (an igneous rock, like a granite) creating a nonconformity. Ercall quarries, Shropshire. (Photo by Jon Radley, University of Birmingham; reproduced with permission.)

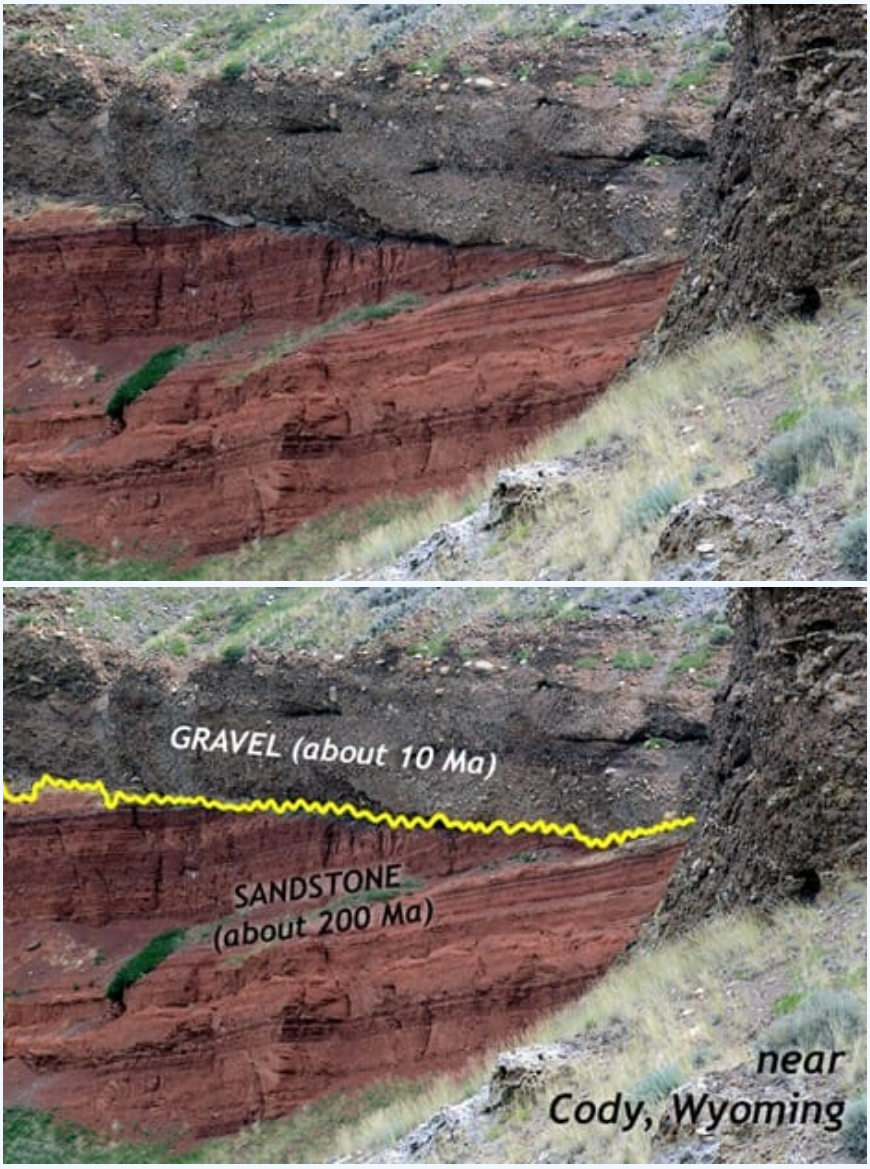


Figure 8.3.29: Angular unconformity near Cody, Wyoming. The lower unit is Triassic Chugwater Formation; the upper unit is Quaternary terrace gravels. (Photo by Ben Edwards; reproduced with permission.)

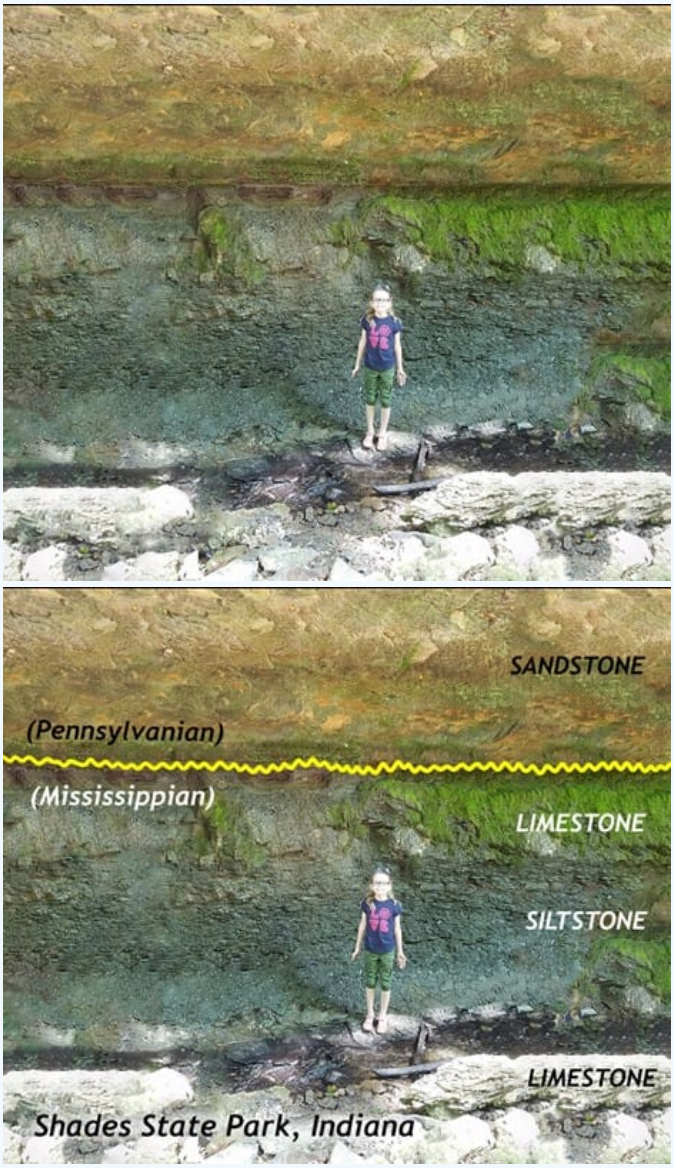


Figure 8.3.30: A gorge in Shades State Park, Indiana shows the Mansfield Sandstone (Pennsylvanian) lying unconformly over limestones and siltstones of the Mississippian aged Borden Group (creating a disconformity). Glacial waters cut into the sandstone, creating these gorges, exposing the strata. (Photo by Troy Simpson; reproduced here with permission)

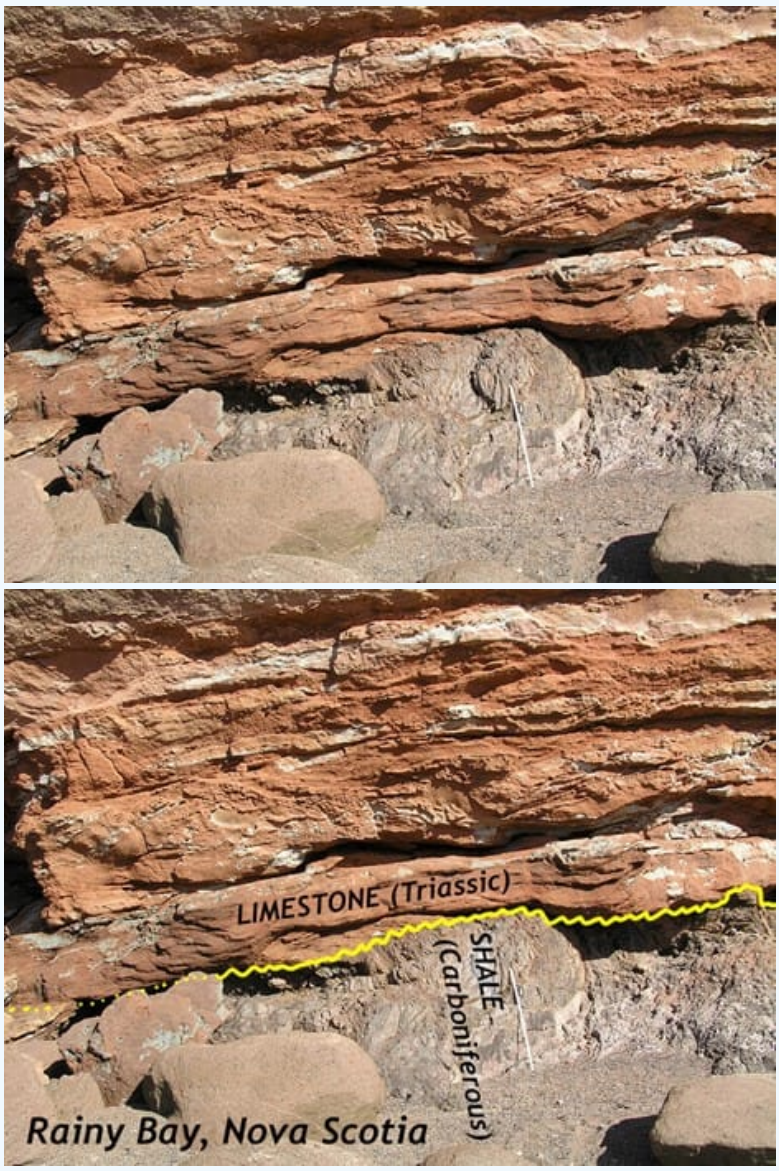


Figure 8.3.31: Gently dipping Triassic strata over vertical Carboniferous layers, Rainy Bay, Nova Scotia, Canada creating an angular unconformity. (Photo by Jon Noad; reproduced with permission.)

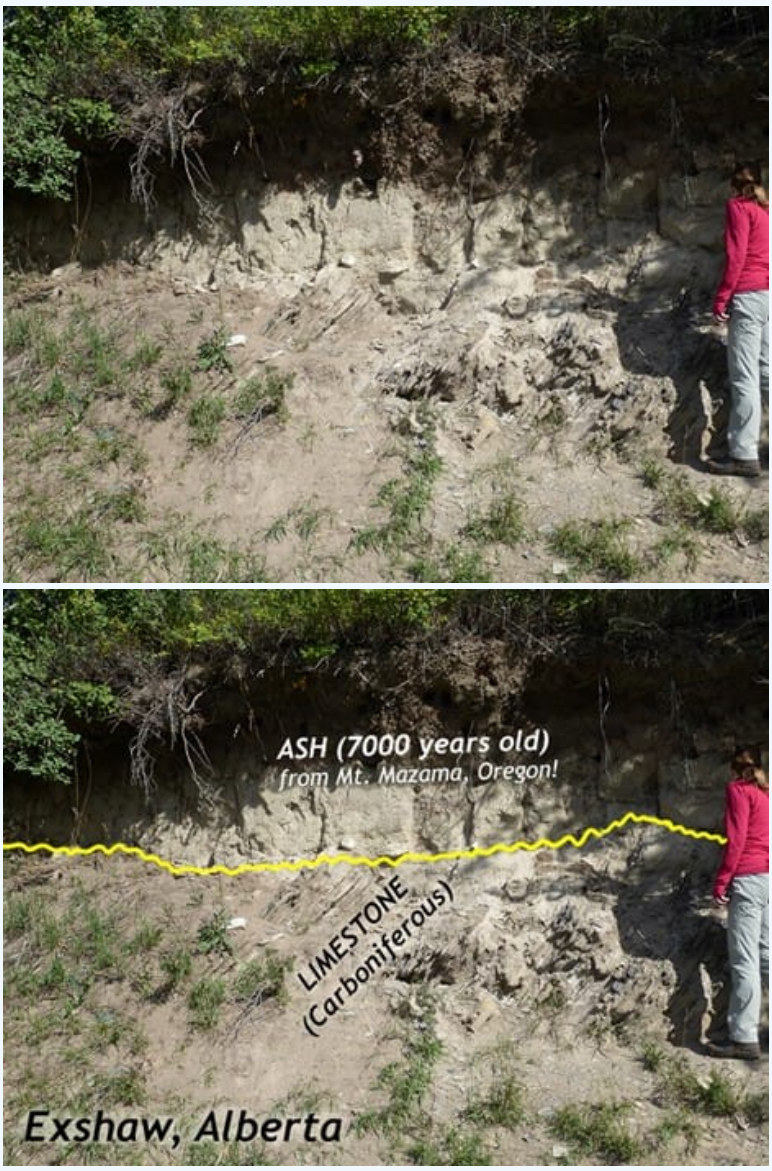


Figure 8.3.32: 7,000-year-old Mazama Ash Bed overlying steeply dipping Carboniferous limestone creating an angular unconformity. Exshaw, Alberta, Canada. (Photo by Jon Noad; reproduced with permission.)



Figure 8.3.33: 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 8.3.34: 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 creating an angular unconformity. (Photo by Troy Simpson; reproduced here with permission.)



Figure 8.3.35: 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.)

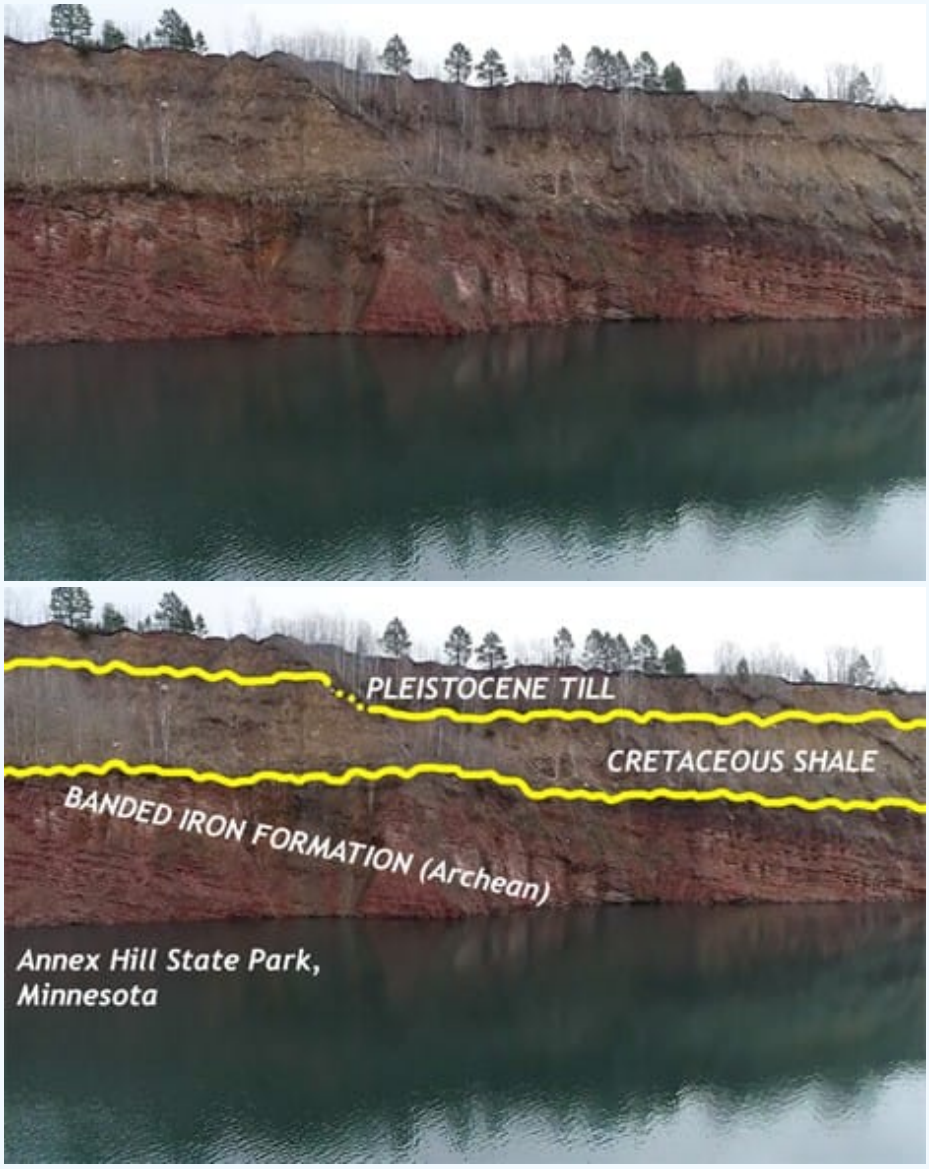


Figure 8.3.36: Annex Hill State Park in Minnesota shows two unconformities: an angular unconformity between Archean banded iron formation and overlaying Cretaceous shale, and a disconformity between the shale and overlying Pleistocene till.

Key Terms

- **angular unconformity** - a geological feature where younger, horizontal sedimentary rock layers are deposited on top of older, tilted or folded sedimentary rock layers, creating an angular discordance between the two sets of strata
- **disconformity** - a gap in the geological record that occurs between parallel layers of sedimentary rock
- **nonconformity** - a gap in the geologic record where sedimentary rock overlies igneous or metamorphic rock
- **unconformity** - a gap in the geologic record that represents an interruption in the deposition of sedimentary layers

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8.4: Fossils and Age Dating

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; 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 (explored elsewhere in this chapter) 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.

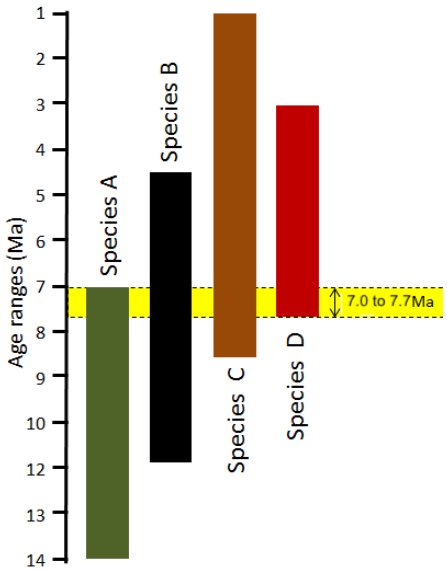


Figure 8.4.1: **Geologic Range Chart**. The application of bracketing to constrain the age of a rock based on several fossils. In this diagram, the colored bars represent the time range during which each of the four species (A, B, C, D) existed on Earth. Although each species lived for several million 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 Figure 8.4.1–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 in Figure 8.4.2 is Triassic in age, specifically living in the range between 251-237 Ma. The fossil ammonoid in Figure 8.4.3 is Cretaceous in age, specifically living in the range between 113-100 Ma. Select the play button to explore these fossils in 3D.

Interactive Content

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Figure 8.4.2: 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.)

Interactive Content

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Figure 8.4.3: 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.

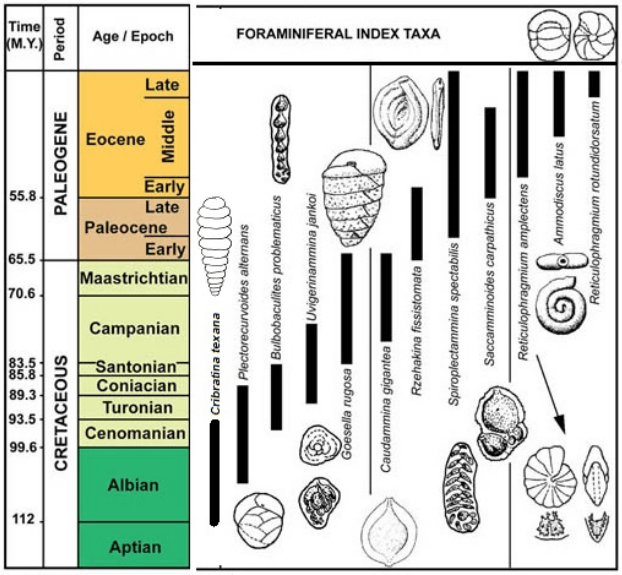


Figure 8.4.4: Select index foraminifera from the Cretaceous and Paleogene. (For educational purposes from the Geologic Timescale Foundation, https://timescalefoundation.org/. Modified by Shelley Jaye.)

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Figure 8.4.5: 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.

Interactive Content

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Key Terms

- **index fossil** - organisms that lived for relatively short time periods and distributed over a wide geographic area
- **paleontology** - the study of the history of life on Earth as based on fossils

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8.5: Absolute Age Dating

Numerical (or Absolute) Dating

Numerical (also known as absolute) **dating** deals with assigning actual dates (in years before the present) to geological events. This contrasts 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 this science is radiometric dating (also known as isotopic dating).

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.



Figure 8.5.1: Archbishop James Ussher (1581-1656) (public domain; WikiMedia Commons).

In the 1800s, practitioners of the young science of geology applied the uniformitarian views of Hutton and Lyell (discussed elsewhere 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 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 years. 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 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.

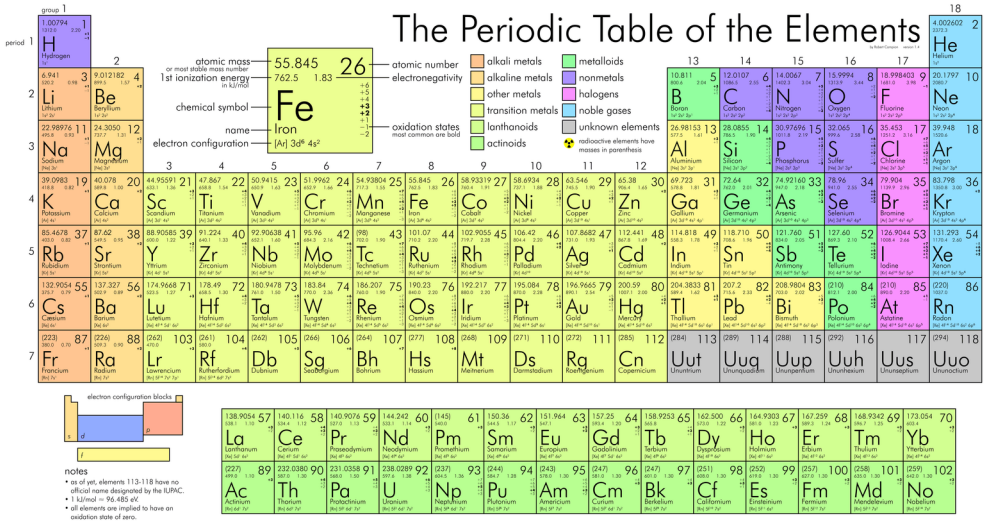


Figure 8.5.2: The periodic table of the elements of the world, showing element names, oxidation states, electron negativities, first ionization energies, and electron configurations. (CC0 1.0 Universal; Public domain from Wikimedia)

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 the number of protons plus the number of neutrons.

Consider, for example, the three different isotopes of Carbon each with 6 protons (which is what makes it 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

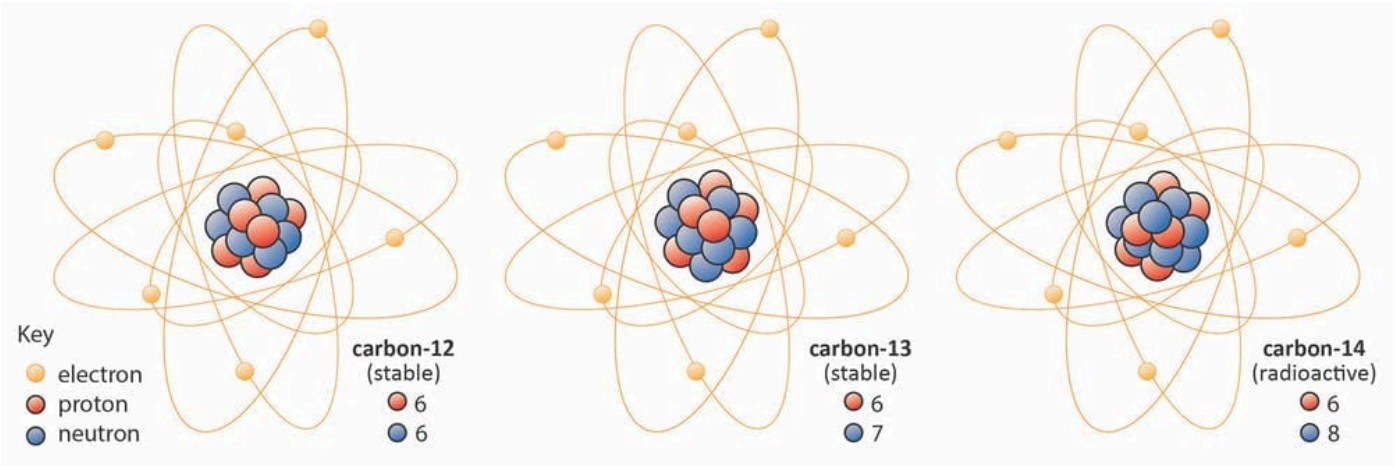


Figure 8.5.3: The three isotopes of carbon. (CC-ND 2.0; Tessa Koumoundouros from: https://theconversation.com/explainer-what-is-an-isotope-10688.)

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.

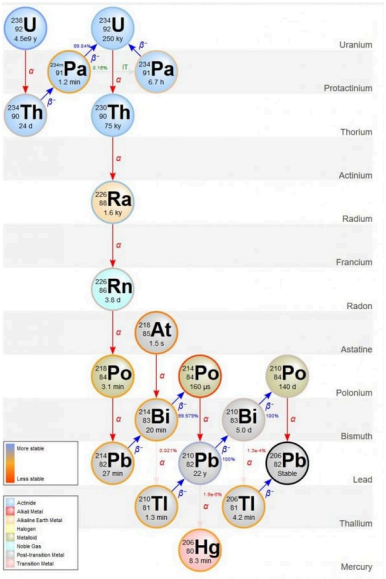


Figure 8.5.4: Decay chain of U-238 to stable Pb-206. (Creative Commons Attribution-NonCommercial-ShareAlike 4.0 International License; Salt Lake Community College, An Introduction to Geology, from: http://opengeology.org/textbook/7-geologic-time/#721_Radioactive_Decay/.)

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 ($ZrSiO_4$). Stable atoms of Zr dominate in the crystal lattice however radioactive ^{238}U atoms can substitute for Zr. The daughter product of the decay of ^{238}U is ^{206}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 $^{238}U \rightarrow {}^{206}Pb$ 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 8.5.1: 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.

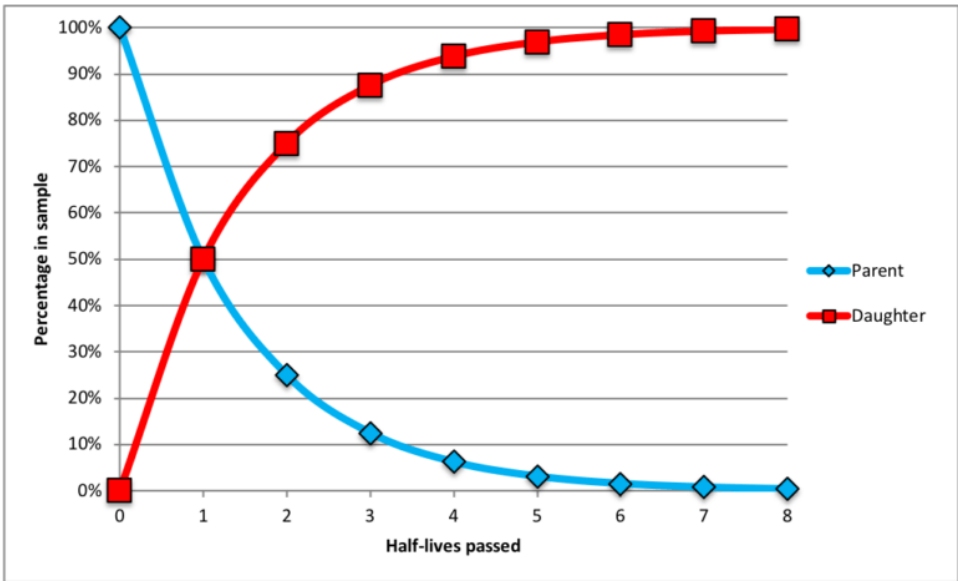


Figure 8.5.5: Radioactive Decay Curve diagram. Relationship between the amount of radioactive parent atoms in a sample relative to the number of daughter atoms over the passage of time, measured in half-lives. (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; 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.

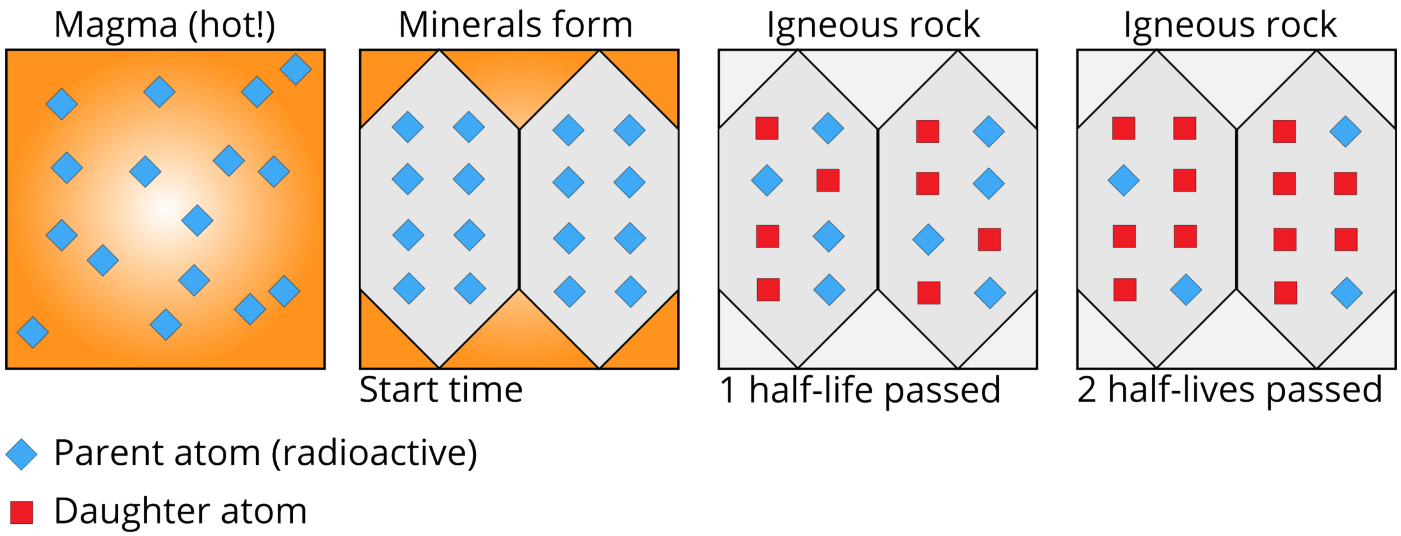


Figure 8.5.6: An example of how the initial number of radioactive parent atoms (blue diamonds) in two mineral grains (gray hexagons) changes over time (measured in half-lives) relative to the number of daughter products (red squares). (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

The left-most box in Figure 8.5.6 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? (Answer: only one as half of the two blue dots in each of the minerals would convert to daughter; so there would be one blue dot per crystal)

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 $^{40}\mathrm{K} \rightarrow {}^{40}\mathrm{Ar}$. 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:

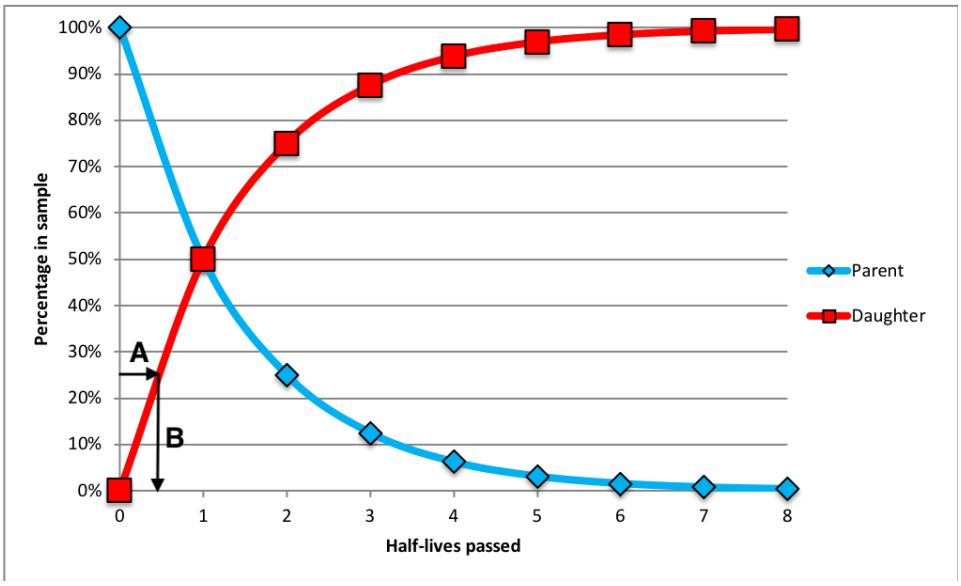


Figure 8.5.7: Radioactive Decay Curve Half Life Plot – example problem.

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):

$$1.25 \text{ By X } 0.4 = 500 \text{ My}$$

Note

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. 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 8.5.8: 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.

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.

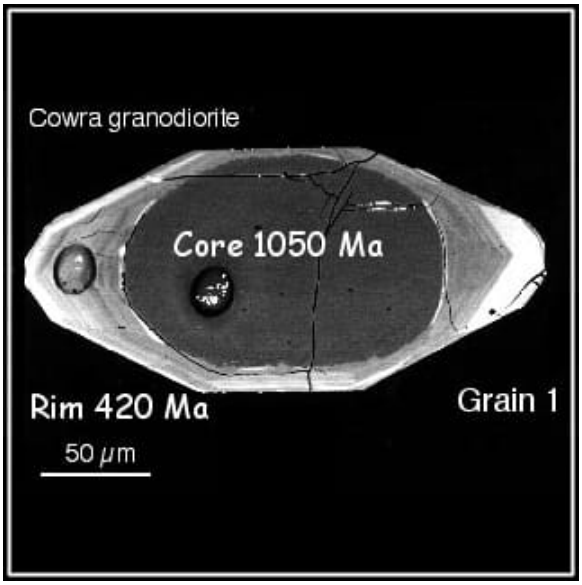


Figure 8.5.9: SHRIMP allowed in situ dating of discrete zones in zircon. This revealed for the first time the true complexities of the geochronology of this mineral as well as its ability to record individual geological events. (With permission from the Research School of Earth Sciences, Australian National 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.

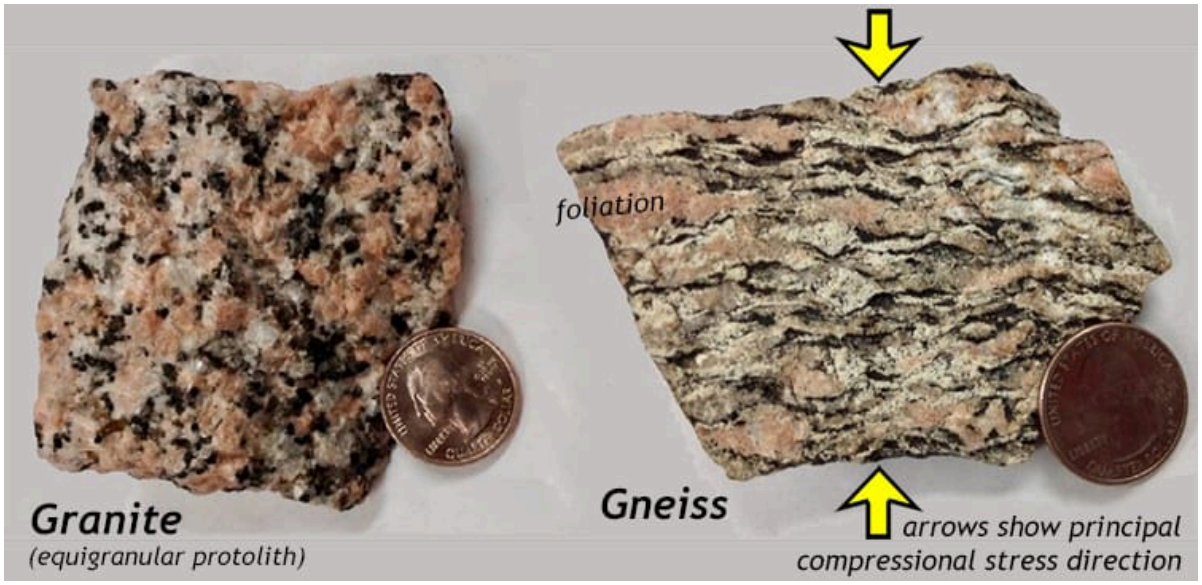


Figure 8.5.10: A comparison of granite and gneiss. The texture of the protolith granite is the same in every direction (equigranular) but the gneiss has a pronounced foliation, which we infer to have developed from squeezing (compressional stress) in a direction that was perpendicular to the resulting foliation.

Note

Here is a reminder of a few of the more complex concepts in this section.

- In order to date old objects, you must use an isotope with a large half-life, otherwise all of the parent decays to daughter product too quickly.
- Carbon-14 can not be used to date rocks like granite for two reasons. First, they are too old and Carbon-14's half-life is just under 6,000 years. This means it can't date materials past about 50,000 years old. Second, Carbon-14 is only found in living objects, and granite was never alive.
- Clastic sedimentary rocks can not be dated using isotopic age dating as they are made of pieces of other rocks. When you date a mineral in the clastic rock, it will produce the age of the rock it came from!
- Isotopic age dating of metamorphic rocks is possible but complicated. Since the "clock" is reset when the minerals get to a certain temperature during metamorphism and then decay starts once the temperature is lowered to a certain amount, problems can arise. Sometimes the "clock" is not completely reset and the rock has an age that is too old, reflecting the age of the protolith.

Key Terms

- **daughter product** – the stable atom that forms as a result of the decay of a radioactive parent isotope
- **half-life** – the amount of time it takes for half of the atoms in a sample of a radioactive isotope to decay into daughter products
- **isotopes** – atoms of the same element that have the same number of protons but different numbers of neutrons, giving them different atomic masses
- **mass number** – the total number of protons and neutrons in the nucleus of an atom
- **numerical dating** - the type of geologic dating in which you assign a number age to a geologic event
- **parent** – the original radioactive isotope that undergoes decay to form a daughter product
- **radioactive decay** – the natural process by which an unstable atomic nucleus loses energy by emitting particles or radiation, transforming into a more stable form (often producing a daughter isotope)

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8.6: The Geological Time Scale

Overview

The geological time scale 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. It 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 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.

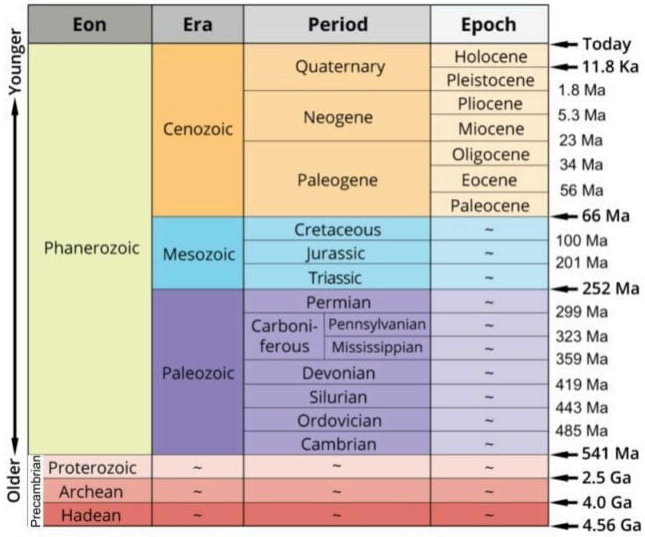


Figure 8.6.1: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

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.

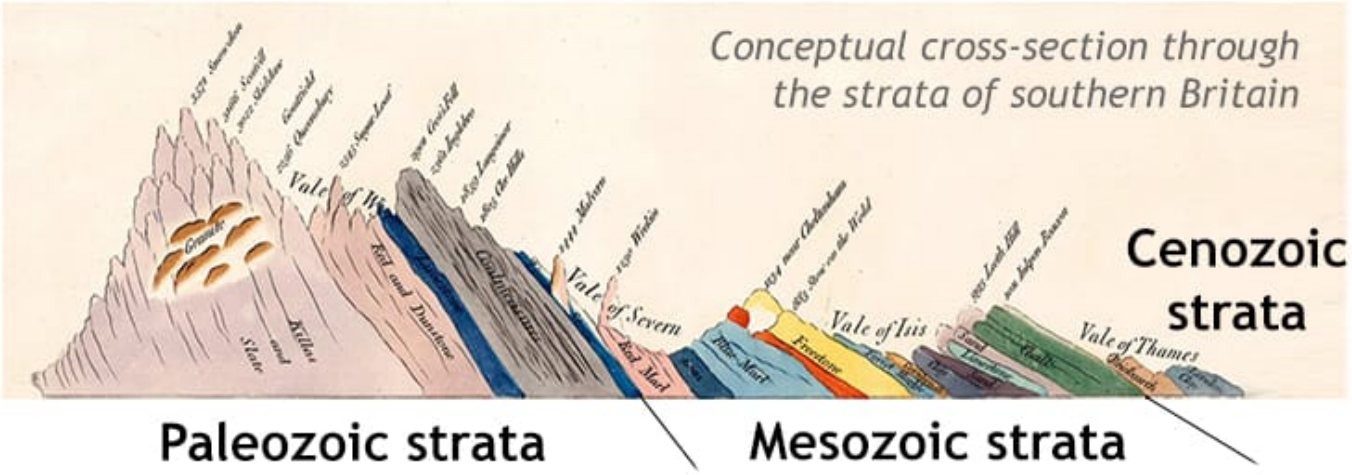


Figure 8.6.2: William Smith's "Sketch of the succession of strata and their relative altitudes," an inset on his geological map of England and Wales (with era names added). (Modified from a version by Steven Earle here: https://opentextbc.ca/physicalgeology2ed/chapter/8-1-the-geological-time-scale/)

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.

Another important principle of relative dating, the principle of faunal succession, allows fossils to act as "timestamps" that reveal the age of the strata in which they are found.

William Smith (1769 – 1839), one of the founding fathers in our early understanding of geologic time in the 1700s and early 1800s conducted work in understanding early relative dating principles of superposition and faunal succession which 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.

The geologic timescale: Where did these names come from?

Phanerozoic eon – Greek for "visible life"

Cenozoic era – Greek for "new life"
- **Quaternary** – Latin for "fourth"
  - **Holocene** – Greek for "entirely new"
  - **Pleistocene** – Greek for "mostly new"
- **Neogene** – Greek for "new birth / generation"
  - **Pliocene** – Greek for "more new"
  - **Miocene** – Greek for "less new"
- **Paleogene** – Greek for "ancient birth / generation"
  - **Oligocene** – Greek for "few new"
  - **Eocene** – Greek for "dawn of the new"
  - **Paleocene** – Greek for "ancient new"
- * **Tertiary** (Latin for "third") is an archaic term left over from the Wernerian chronology. It is no longer officially used as part of the geologic timescale.

Mesozoic era – Greek for "middle life"
- **Cretaceous** -- Latin word for chalk ("creta") and was first applied to extensive deposits of this age that form white cliffs along the English Channel between Great Britain and France.
- **Jurassic** – named after the Jura Mountains between France and Switzerland, where rocks of this age were first studied.
- **Triassic** – refers to the threefold division of rocks of this age in Germany.

Paleozoic era – Greek for "ancient life"
- **Permian** – named after the province of Perm, Russia, where rocks of this age were first studied
- **Pennsylvanian** – named after the state of Pennsylvania (USA), where rocks of this age are widespread
- **Mississippian** – named after the Mississippi River valley (USA), which contains good exposures of rocks of this age
- **Devonian** – named after Devonshire, England (UK), where rocks of this age were first studied
- **Silurian** – named after a Celtic tribe called the Silures (UK).
- **Ordovician** – named after a Celtic tribe called the Ordovices (UK).
- **Cambrian** – Roman name for Wales (UK), where rocks of this age were first studied

Precambrian* – all the time prior to the start of the Cambrian (not a formal era/eon name)
* synonymous with "**Cryptozoic**" – Greek for "hidden life"

Proterozoic eon – Greek for "former life"
Neoproterozoic – Greek for the "new" Proterozoic
Mesoproterozoic – Greek for the "middle" Proterozoic
Paleoproterozoic – Greek for the "ancient" Proterozoic

Archean eon – Greek for "ancient"

Hadean eon – Greek for "hellish" or "unseen" (the Netherworld)

Most recent

Most ancient

Figure 8.6.3: 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 understand the geological time scale. This is most easily done by first breaking the time scale into its component parts: eons, eras, periods, and epochs. This is similar to how we break time up into millennia, centuries, decades, and years.

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.

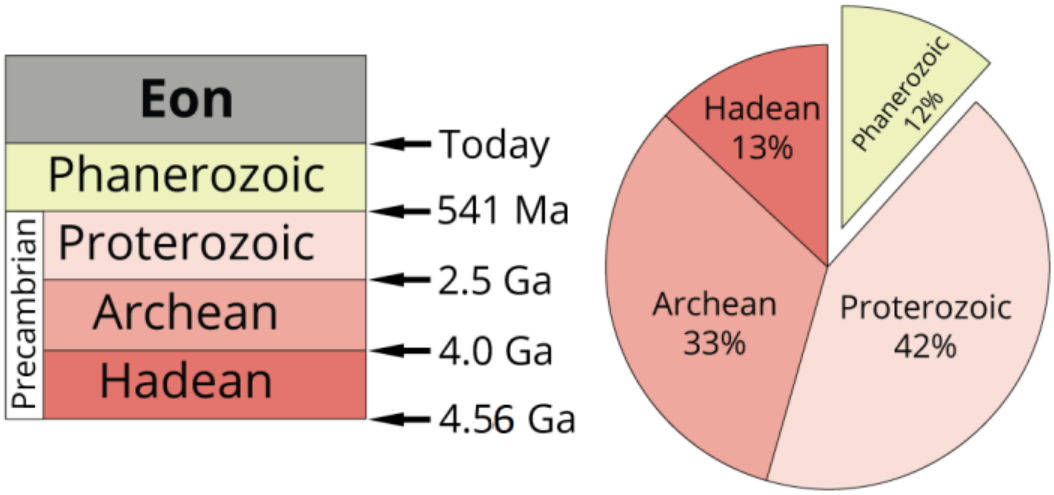


Figure 8.6.4: Left: the four eons of geological time. *Right:* the "Precambrian" eons (Hadean, Archean, and Proterozoic) represent 88% of geological time. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License.; Image by Jonathan R. 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.

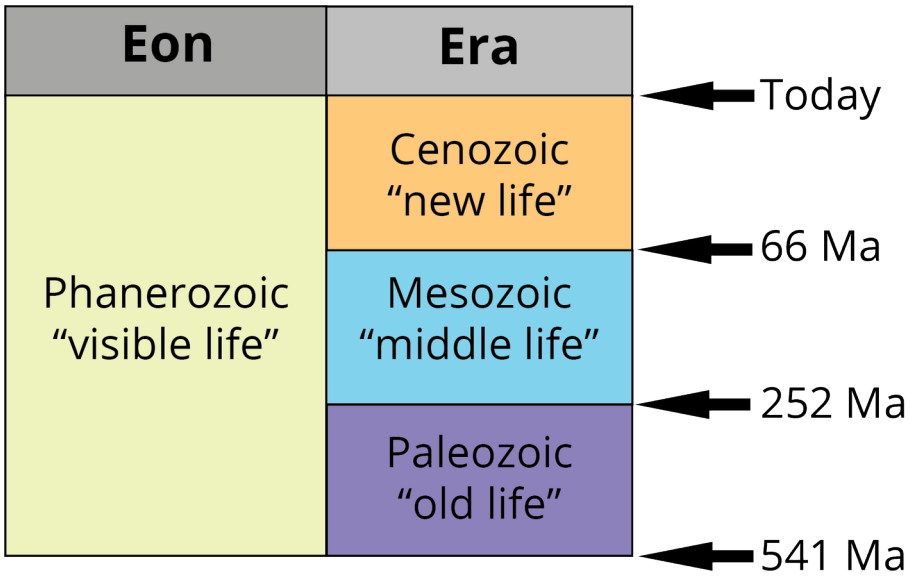


Figure 8.6.5: The three eras of the Phanerozoic eon. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

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.

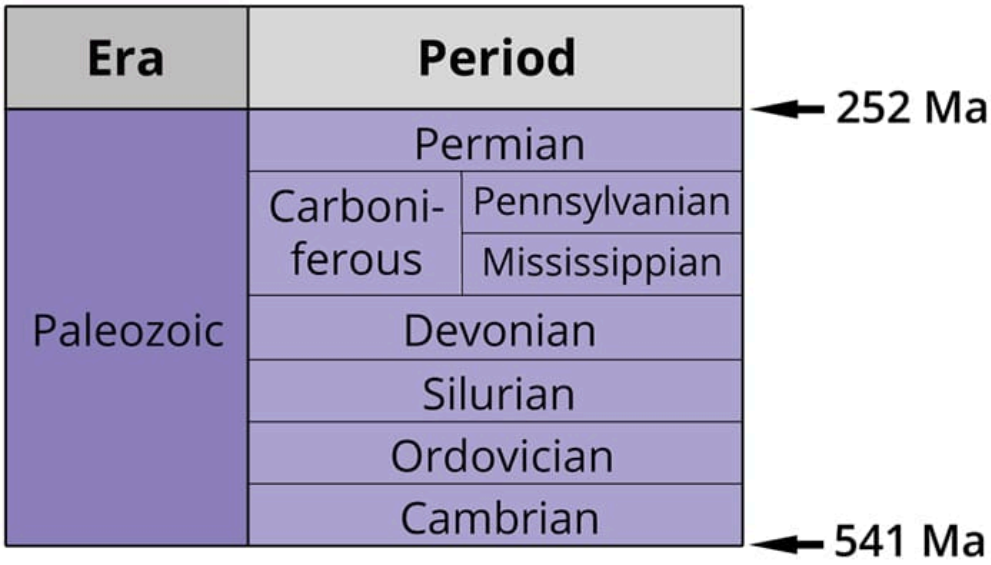


Figure 8.6.6: The seven periods of the Paleozoic era. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

The Mesozoic era is divided into the Triassic, Jurassic, and Cretaceous periods.

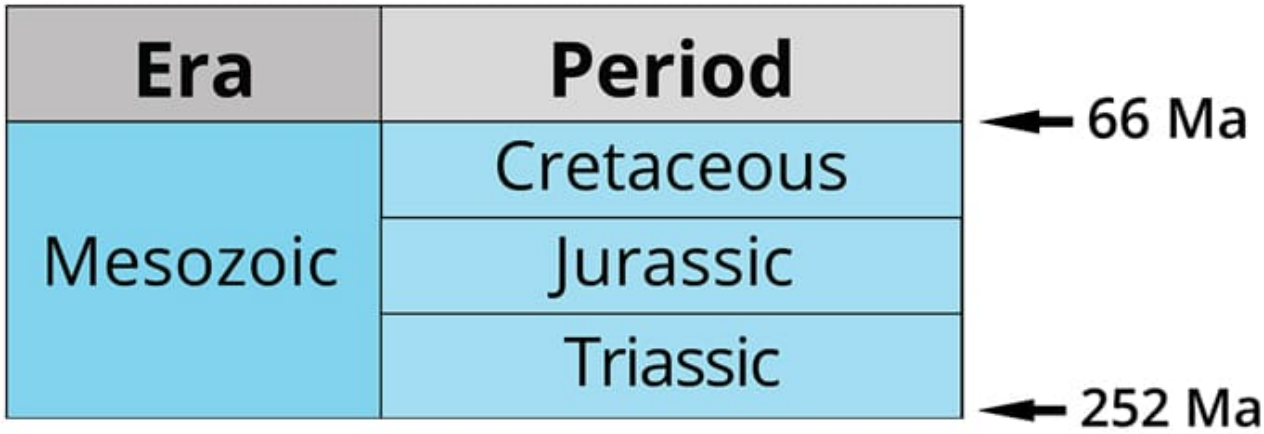


Figure 8.6.7: The three periods of the Mesozoic era. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

Finally, the Cenozoic era is divided into three periods: the Paleogene, Neogene, and Quaternary.

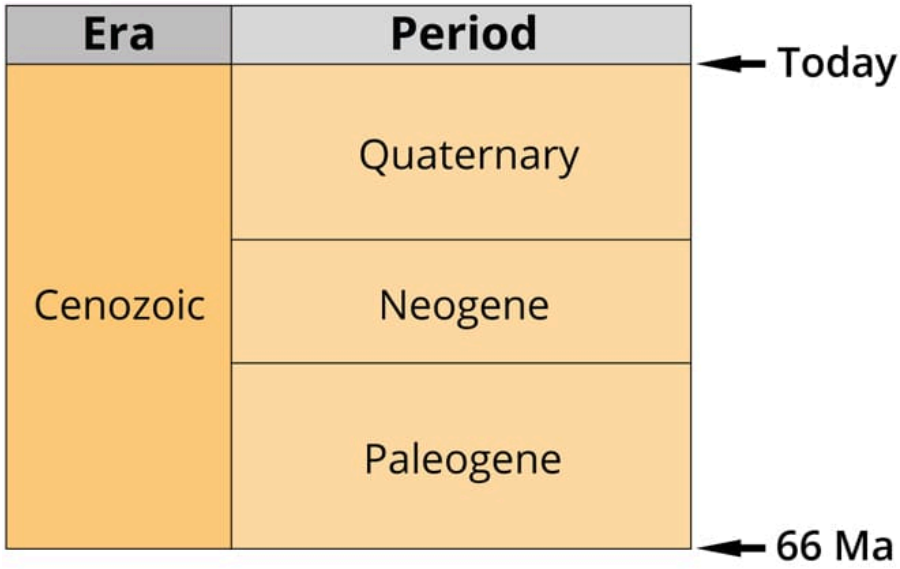


Figure 8.6.8: The three periods of the Cenozoic era. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

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.

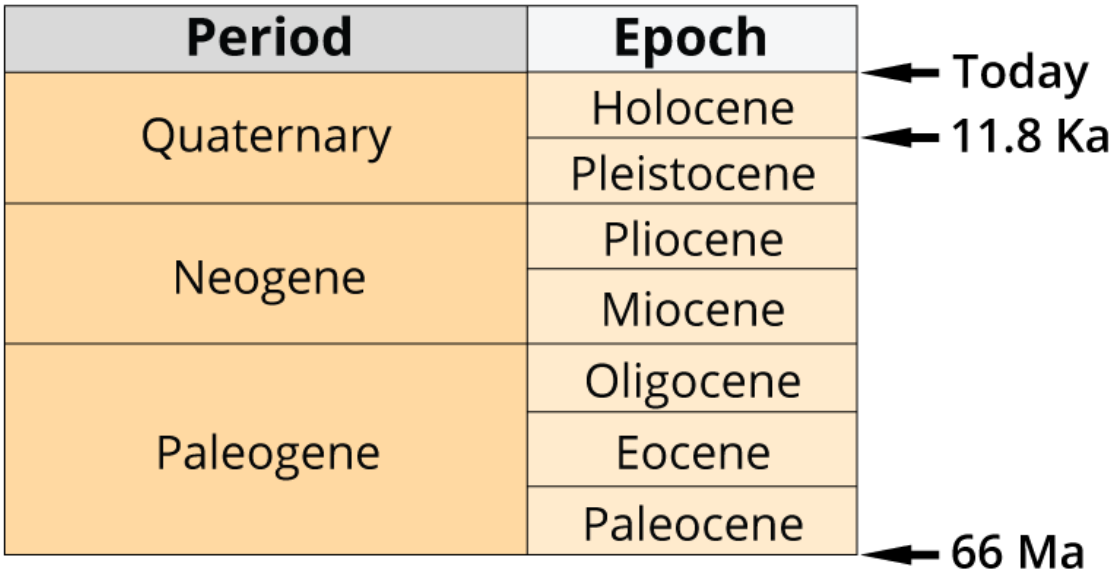


Figure 8.6.9: Epochs of the Paleogene, Neogene, and Quaternary periods. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

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.

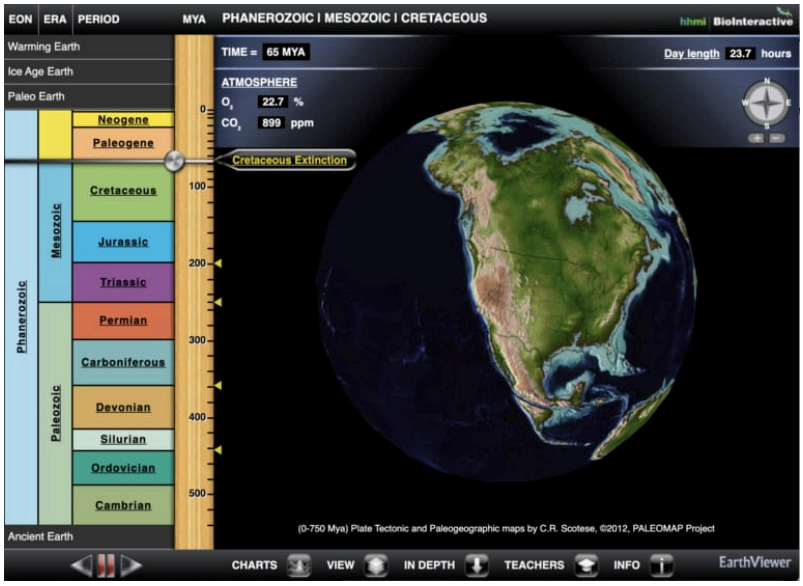


Figure 8.6.10: Static image of the Earth Viewer from Howard Hughes Medical Institute Biointeractive. (From: https://www.biointeractive.org/classroom-resources/earthviewer. Use for educational purposes.)

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 that you may find helpful as you memorize the time scale.

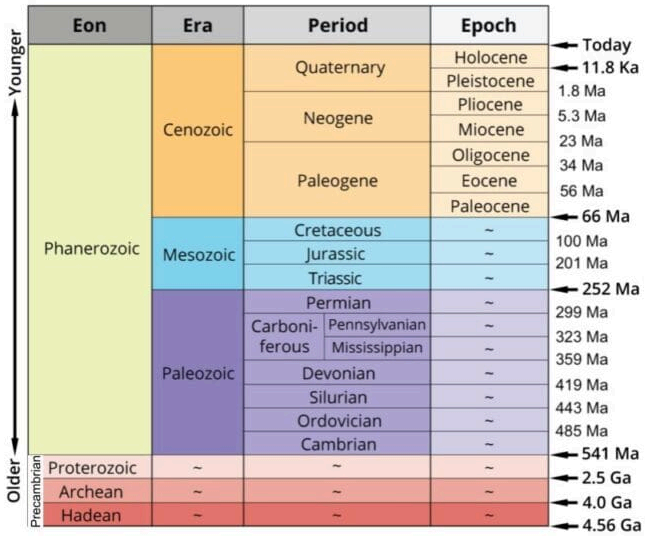


Figure 8.6.11: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

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.

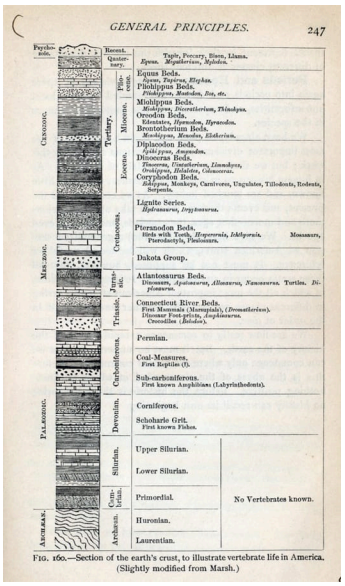


Figure 8.6.12: Geological time scale from Le Conte (1885). (public domain)

Key Terms

- **stratigraphic column** - a diagram that shows a vertical sequence of rock layers in the order they were deposited representing the rock units, their thicknesses, lithologies (rock types), and sometimes fossils or other features to illustrate the geologic history of an area

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8.7: Key Points

1. Earth is 4.6 billion years old.
2. Uniformitarianism suggests the chemical and physical laws of nature have not changed over the course of Earth's long history which allows geologists to study present day processes and assume things happened at similar rates in the past. This allows them to build a timeline.
3. Relative dating allows geologists to put geologic events in order of occurrence but does not give timeframes to any of the events.
4. The principle of superposition tells geologists sedimentary beds are generally older at the bottom and younger at the top of a stacked sequence.
5. The principle of faunal succession allows geologists to determine two different sedimentary beds containing similar fossils in two different areas are the same age.
6. The principle of original horizontality tells geologists that tilting events occur after the deposition of sedimentary beds.
7. The principle of lateral continuity allows geologists to connect sedimentary beds across valleys.
8. The principle of cross-cutting relationships tells geologists that faults and intrusions are younger than the rocks they cut through.
9. The principle of inclusions tells geologists a piece of rock included in another rock is older than the rock that surrounds it.
10. Unconformities represent time that is missing from the rock record due to a pause in deposition or the occurrence of erosion.
11. Fossils are remnants of past animal or plant life in rocks. If we know when that species lived, we can narrow down the time period in which a particular rock formed.
12. An index fossil represents a species that lived for a short time but was very widespread. It is useful because it lived widely but for a short period of time so it really narrows down the age of the rock.
13. Numerical dating allows geologists to know the exact age of a rock or geologic event in years.
14. Radiometric dating is the most common form of numerical dating which uses the decay of isotopes in the minerals of a rock to calculate the age of the rock.
15. One of the major limitations of radiometric dating is it can not be used on clastic sedimentary rocks since they are made of pieces of pre-existing rocks that each have their own age.
16. The geological time scale organizes Earth's 4.6-billion-year history into a series of nested time units dividing time into eons, eras, periods, epochs, and ages, based on major events such as mountain building, climate shifts, and changes in life forms preserved in the fossil record.

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9: EVOLUTION

Introduction

The biosphere is filled with patterns. This is true not only in the ecosystems around us today, but also in our planet's fossil past. These patterns force scientists to ask questions:

- 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?
- Why are ecosystems arranged in a hierarchy with primary producers on the bottom and omnivorous organisms on top?

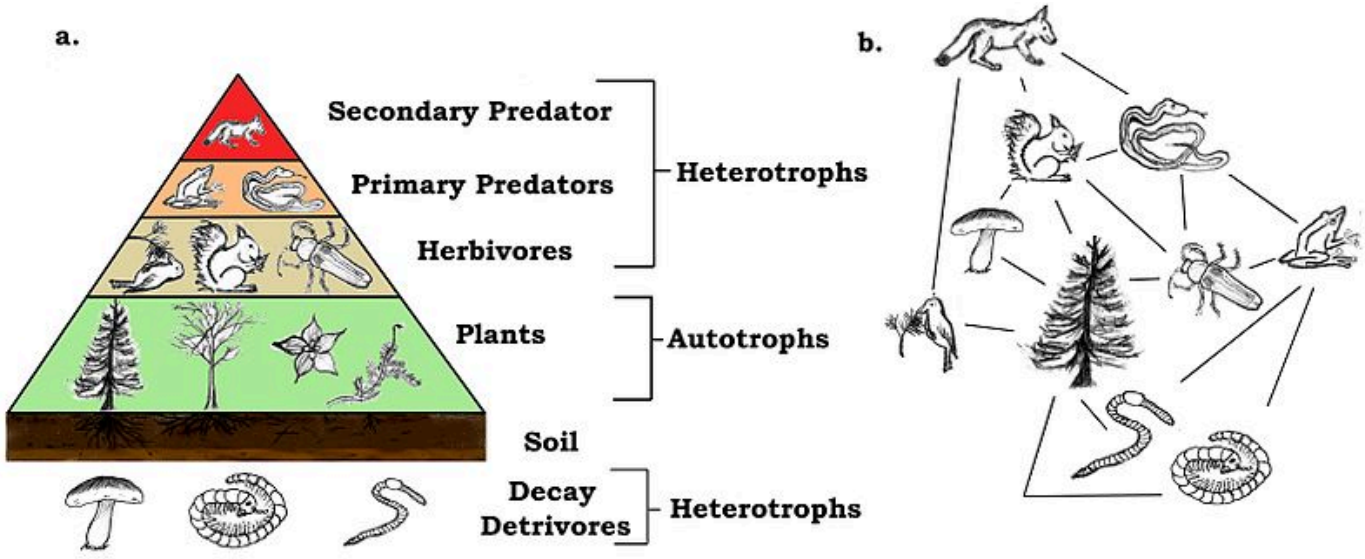


Figure 9.1: The trophic pyramid is based upon observations of food web, or energy transfer, as observed in living ecosystems. As a model, it explains the patterns of food production and predation we see in living systems. It can also be applied to fossil organisms because such energy transfer processes, from decomposing heterotrophs to omnivorous heterotrophs also apply to past ecosystems (CC BY 3.0; Thompsma, Wikimedia Commons).

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.

9.1: What is Evolution?

9.2: Development of the Theory of Evolution

9.3: A Modern View of Evolution

9.4: Evolutionary Patterns in Fossils

9.5: Key Points

9.6: Further Reading

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9.1: What is Evolution?

Descent with Inherited Modification

"Descent with inherited modification" is the essential idea of **evolution**. Biological evolution results from the mutation of genetic material through generations of offspring. Such evolution from one generation to another is the result of the accumulation of changes in DNA. These 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 (or **genotype**) are physically expressed through its appearance (or **phenotype**). Such changes in genes can lead to longer horns, more toes, improved leg bone structure for walking on dry land, or bipedalism in primates.

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 the questions of our origin. You are reading this chapter, and that is a result of evolutionary change!

Evolution tells us that we are all related. Of course 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, which speciated into the many branches of the Tree of Life. Some modern life 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 (these are considered "highly derived").

Evolutionary change is observable everywhere, like in the development of antibiotic resistance among bacteria. A stroll through a wooded area will reveal an ecosystem of food webs where an intricate dance is performed daily between organisms that have evolved together. A decomposing organism is broken down by other organisms specially adapted for that purpose.



Figure 9.1.1: Decomposition of a log in the woods, aided by fungi, microorganisms, and insects. (CC BY 2.0; Hauke Musicaloris, Flickr) .

As a theory that has been itself evolving for more than a century, evolution 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)

Key Terms

- **evolution** - the process of gradual change in the characteristics of a population leading to the development of new species over many generations
- **genotype** - a unique sequence of DNA
- **phenotype** - the set of observable characteristics of an organism, resulting from the interaction between its genetic makeup (genotype) and environmental factors

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9.2: Development of the Theory of Evolution

Organizing the Tree of Life: Linnaean Classification

The beginnings of recorded classification go back to Aristotle in the 4th century BC, with his observations of invertebrates while living on the islands of Greece. His method of classifying organisms began by separating animals from plants, then separating 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 through the middle ages and right up to the time of Linnaeus in the 18th century.

Classification of organisms in the years just prior to Linnaeaus' work 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. His system relies on shared characteristics - what organisms have in common - to create descriptive names and organize them into groups or "bins."

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 most inclusive category. Kingdoms get divided up into "Phyla," then "Classes," "Orders," "Families," "Genera," and finally "Species."

In 1990, Carl Woese added a higher taxonomic rank, the Domain, because of the discovery of three distinct forms of RNA that separate the Archaea, Eukarya, and Bacteria from one another. 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 later insights 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!

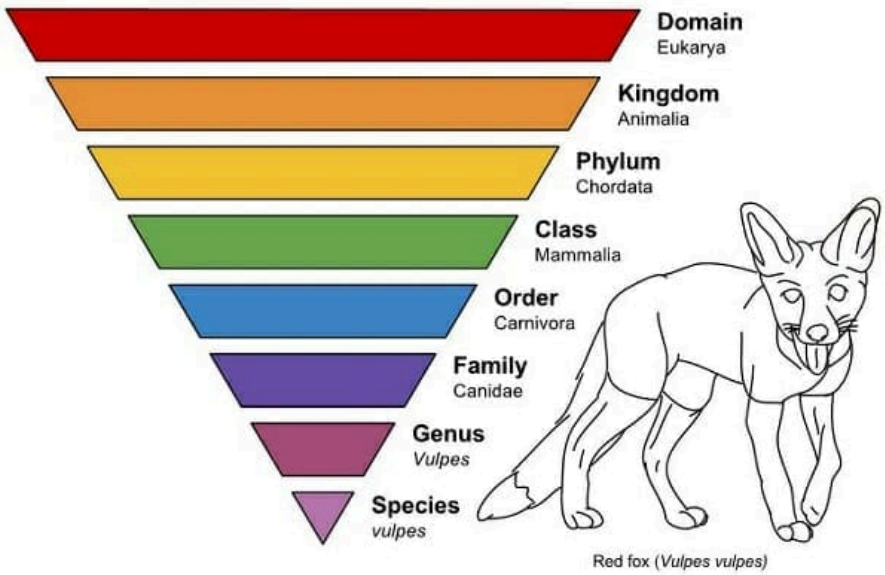


Figure 9.2.1: Red Fox classification according to the Linnaean System. Including the later addition of the "Domain", wolves belong to the Kingdom Animalia and, as the system progresses, end with the genus and species names Vulpes vulpes.

Evolutionary Foundations: Lamarckian Inheritance

Determining a mechanism for 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.

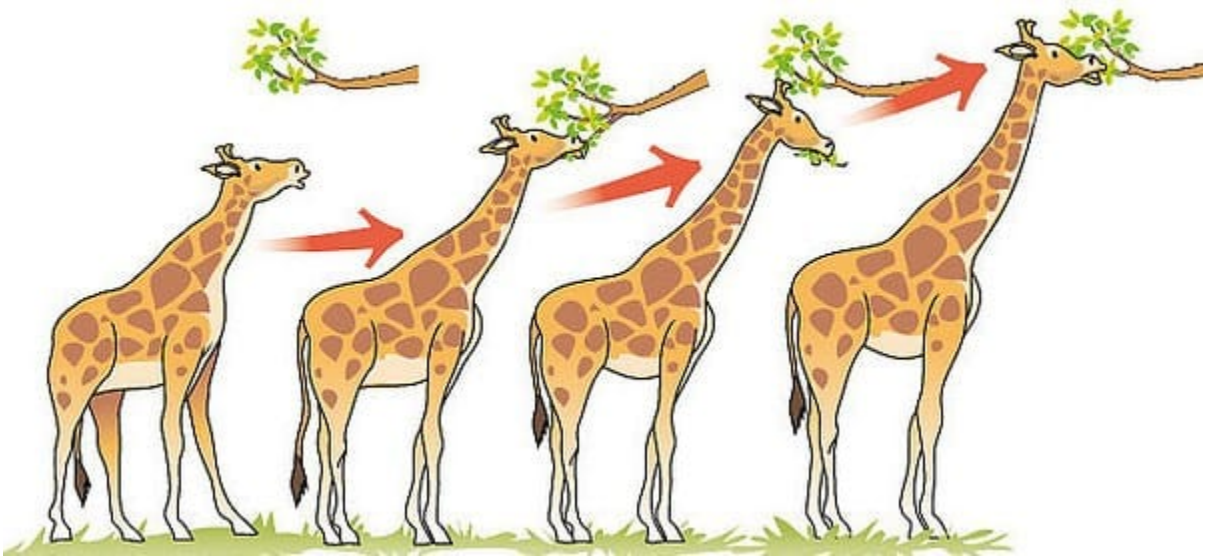


Figure 9.2.2: Lamarckian inheritance of acquired characteristics says that these giraffe, through their own efforts, lengthened their necks and passed the trait of a slightly longer neck to their offspring. (CC BY-SA 4.0, Sandritaverooka, Wikimedia Commons.)

A modern example of **Lamarck's inheritance of acquired traits** is a parent that undergoes plastic surgery to change their body. A mother who alters the shape of her nose through surgery will not pass the altered shape down to her children, because that new shape is not coded in her genetics. We know now that Lamarckian inheritance is a flawed understanding of inherited characteristics, but Lamarck 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 was a major step forward.

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.

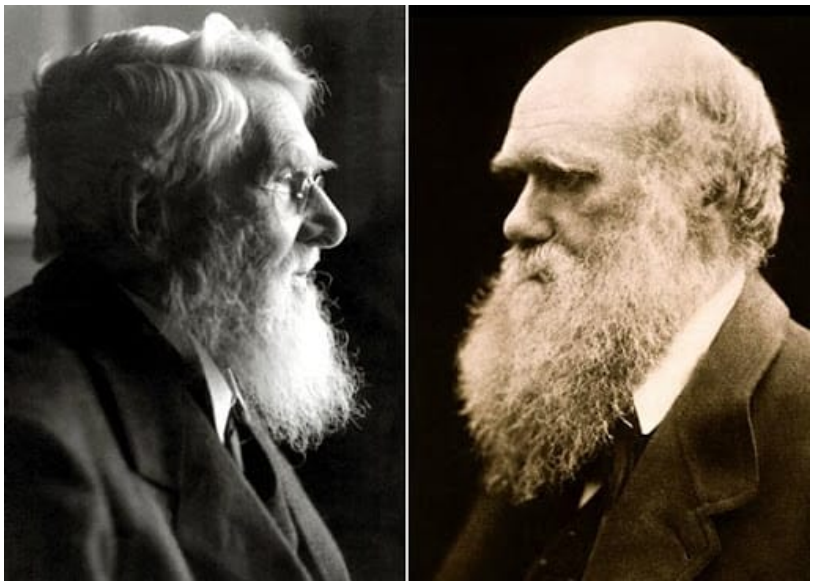


Figure 9.2.3: Photo of Alfred Russel Wallace (left) and Charles Darwin (right), co-discoverers of natural selection.

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.

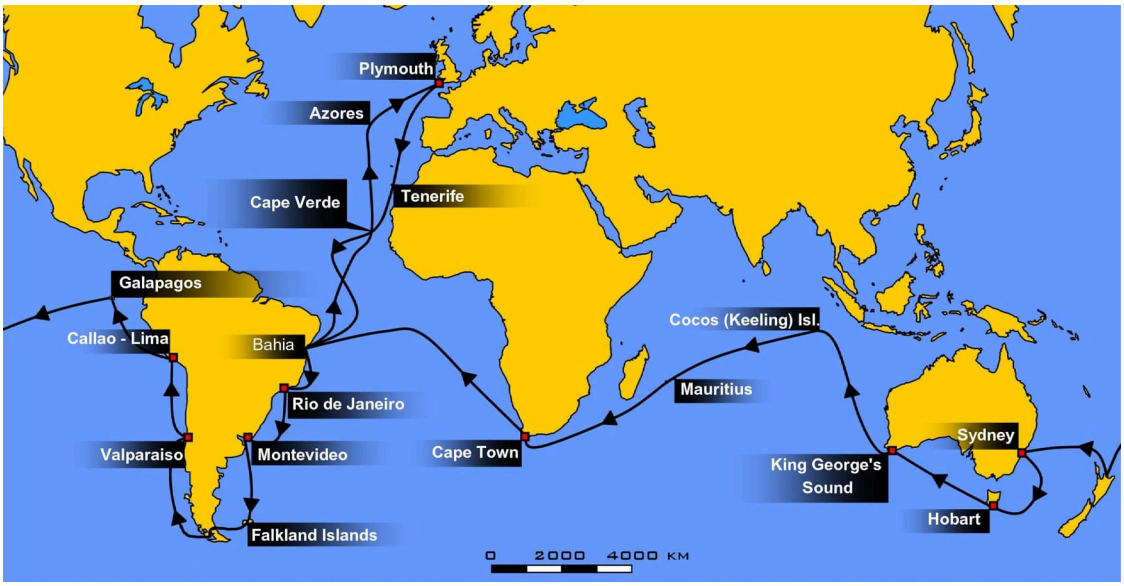


Figure 9.2.4: Stops along Charles Darwin's second voyage of "The Beagle" (1831-1836). These includes locations from Australia to Cape Town, South Africa, to South America and, eventually, back to England.

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.

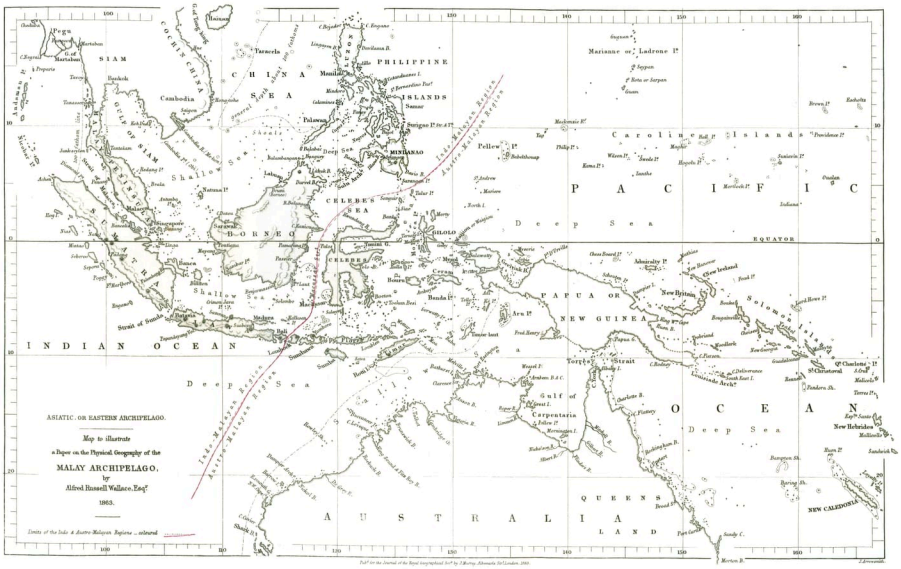


Figure 9.2.5: Map of Indonesia, Australia, and environs depicting "Wallace's Line". This line between landmasses separates Asian from Australasian faunas and was described by Wallace. The differences in species across this line provided Wallace with the patterns he used to describe his views on natural 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. Given these straits were about 35 km in width, the mobility of an organism was key. Plant pollen and seeds, blown by the wind or transported by water, allowed growth on both sides of the strait. Large mammals and most birds, by contrast, could not traverse it. This separation, over time, led 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 write letters to each other. 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.

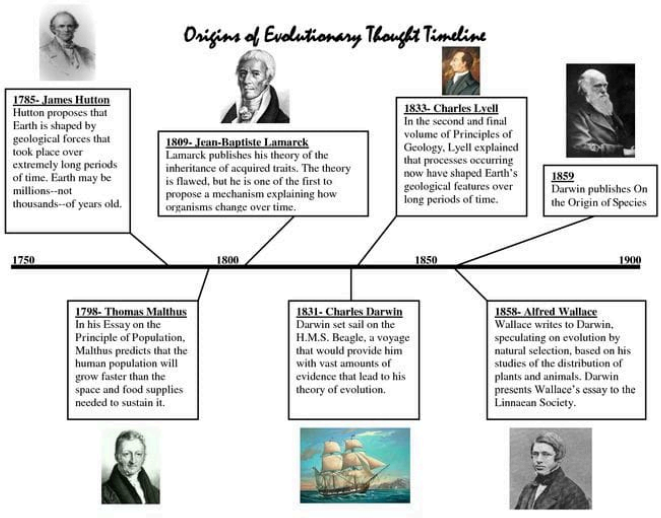


Figure 9.2.6: A brief timeline of evolutionary thought, from James Hutton to Charles Darwin and Alfred Russell Wallace.

Meanwhile, Back in Europe: Mendelian Genetics

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.

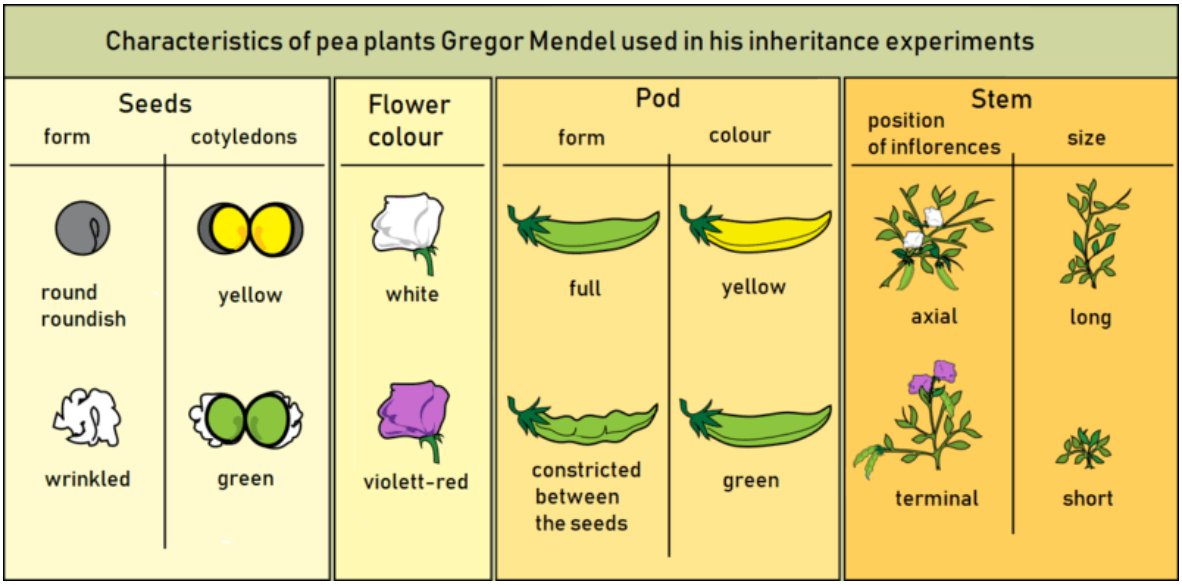


Figure 9.2.7: Characteristics of pea plants Mendel used in his inheritance experiments. These include seed form, cotyledon color, flower color, pod form and color, flower position on stem, and stem size.

There are three principles in Mendelian genetics. (It is worthwhile to note that, though the term “laws” here are accepted within the scientific community, they may be more accurately referred to as “principles.”)

1. Law of Segregation: Genes come in two forms called alleles.
2. Law of Dominance: Alleles can be dominant (expressed in an organism’s appearance) and recessive (not expressed). 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.
3. Law of Independent Assortment: The selection of one allele of a gene over another, however, occurs independently of selection occurring among other genes.

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 “polygenic”, such as the color of fruit fly eyes, where several genes contribute to a physically expressed variation.



Figure 9.2.8: Variations in fruit fly eye color (from the top, clockwise): brown, cinnabar, sepia, vermilion, white, wild.

YouTube Video

This interactive element is not available in this format. Visit the link or scan the QR code to access it online.

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Key Terms

- **binomial nomenclature** - a system of naming plants and animals by assigning a two-part, Latin-based name to each, consisting of the genus and the species
- **invertebrate** - an animal that does not have a backbone or spinal column
- **Lamarck's inheritance of acquired traits** - the theory that an organism's offspring inherit characteristics that the parent acquired during its lifetime through the use or disuse of organs
- **natural selection** - the method by which organisms with traits better suited to their environment are more likely to survive and reproduce, passing those advantageous traits to their offspring

- **vertebrate** - an animal with a backbone, or spinal column, and an internal skeleton

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9.3: A Modern View of Evolution

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. 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**.

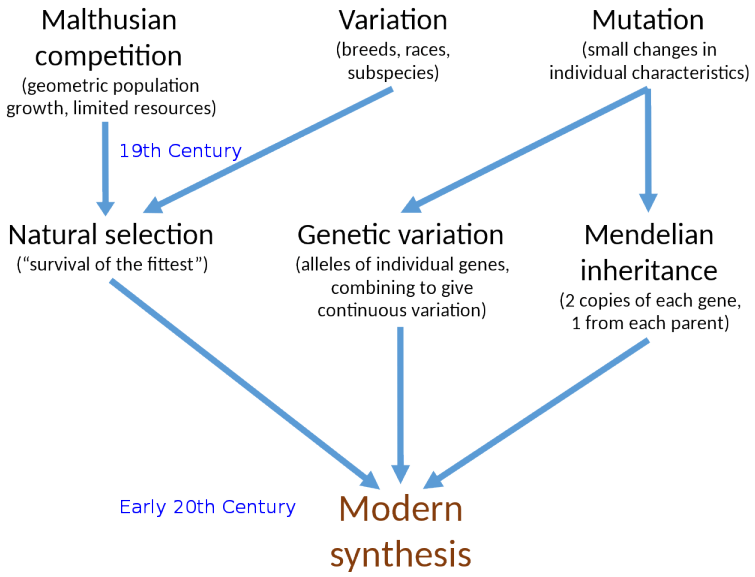


Figure 9.3.1: Diagram of the Modern Synthesis of evolutionary theory. The combination of principles undergirding population dynamics with understandings from genetics is synthesized into a larger and more coherent model of how organisms and populations evolve.

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.

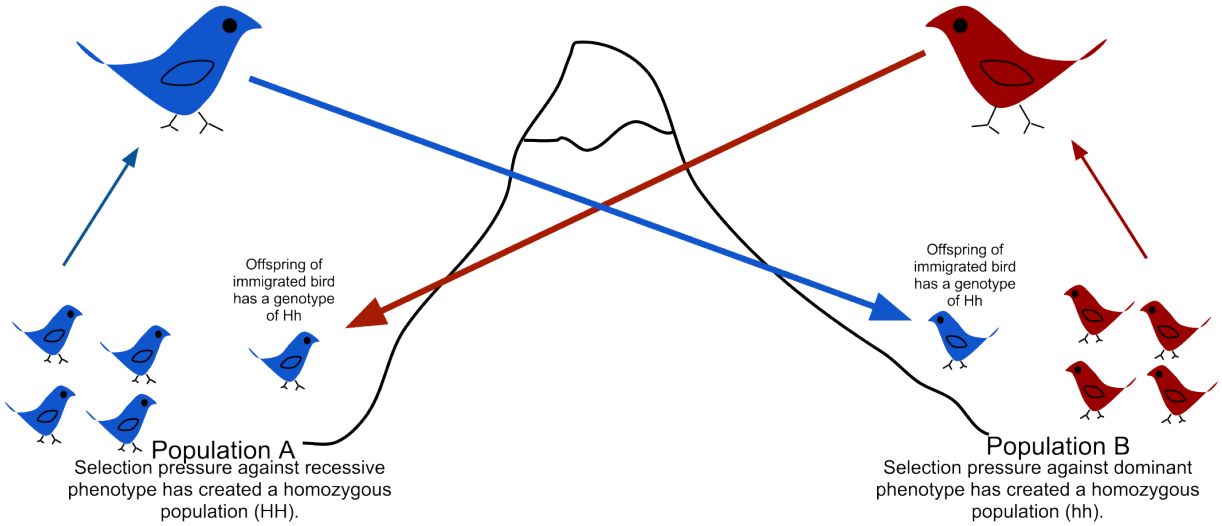


Figure 9.3.2: Gene flow, the transfer of alleles across populations, is depicted here showing birds from two populations living on either side of a mountain. Species A, the blue birds, contains the dominant alleles while species B, the red birds, contain the recessive alleles. A member of species A is 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.

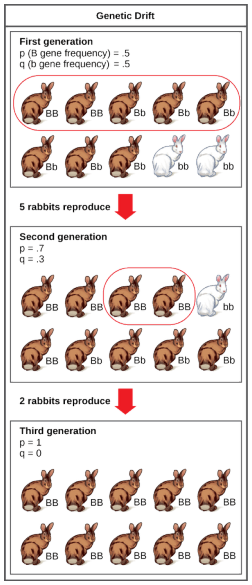


Figure 9.3.3: An example of genetic drift using rabbits. Genetic drift allows for the removal of an allele from a population due to chance. Even through the allele frequency for brown and white fur coats is equal in the first generation, the dominant allele (for a brown coat) quickly displays with more 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 migration occurs sideways, 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 so inclined, 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.

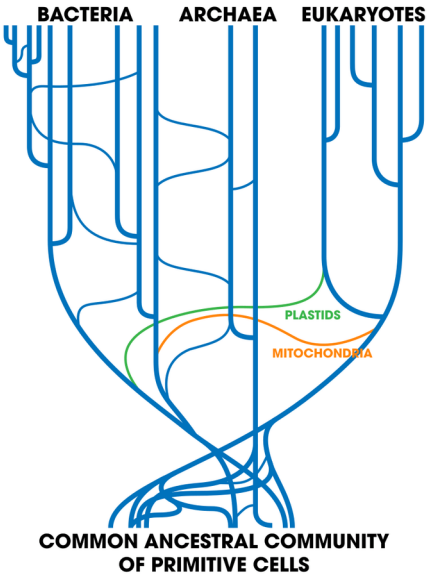


Figure 9.3.4: Horizontal gene transfer as a form of inheritance, in this case through the transfer of genetic material across Kingdoms through plastids and mitochondrial materials (Smets, 2005).

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 9.3.5: 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.

Who are the "Fittest"?

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).

YouTube Video

This interactive element is not available in this format. Visit the link or scan the QR code to access it online.

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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.

Key Terms

- **artificial selection** - the process by which humans deliberately breed organisms for specific and preferred traits
- **gene migration** - the movement of genetic material from one population to another through the migration of individuals into a new area
- **genetic drift** - a change in the frequency of an allele over time caused by some individuals reproducing more successfully than others
- **genetic mutations** - changes in the sequence of nucleic bases in the coding portion of a DNA molecule
- **genetic symbiosis** - the transfer of genetic material from one species to another due to close physical interactions
- **horizontal gene migration** - the movement of genetic material between organisms that are not parent and offspring
- **neo-Darwinian evolutionary theory** - the theory that combines Charles Darwin's concept of natural selection with modern genetics, including Mendelian inheritance and the understanding of mutations

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9.4: Evolutionary Patterns in Fossils

Figured Stones to Fossils: Nicolas Steno

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.

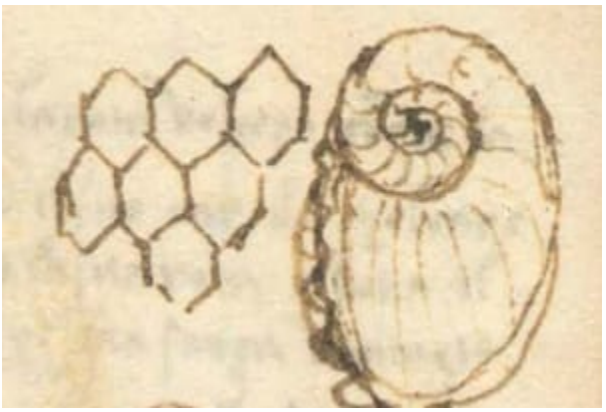


Figure 9.4.1: Drawing of Palaeodytion specimen, Leonardo da Vinci, Paris Manuscript I

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 Nicolas 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 9.4.2). 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.

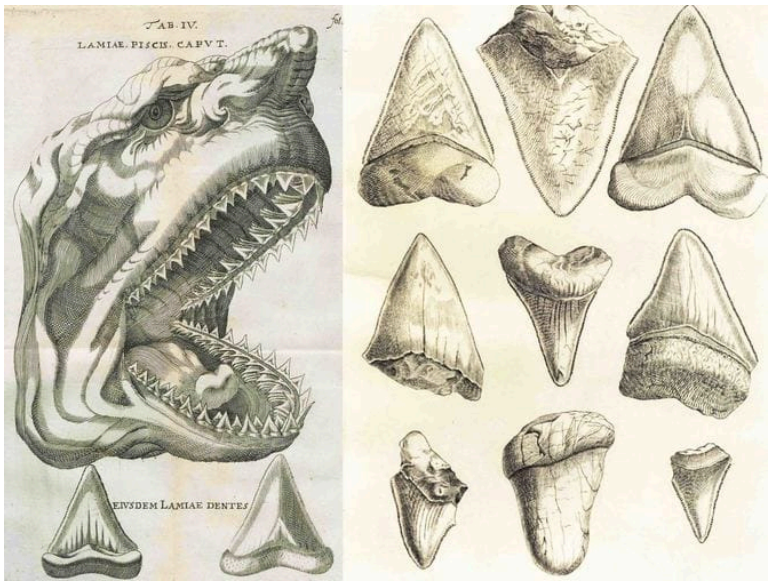


Figure 9.4.2: Woodcut of figured stones from Nicholas Steno's 1667 monograph on the dissection of a Great White Shark. Similar stones, found embedded in rocks in the northern Italian Alps, Steno identified as sharks' teeth. From "Canis carchariae dissectum caput" (Steno, 1669)

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 Georges 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.

The Principle of Faunal 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.

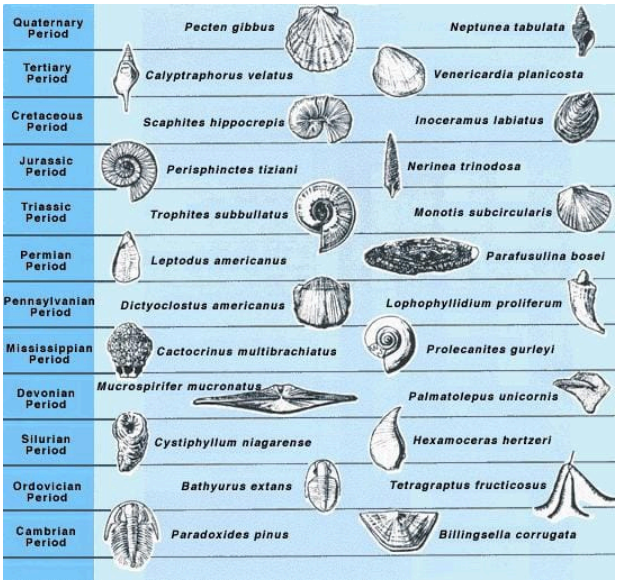


Figure 9.4.3: Faunal Succession through geologic time. Certain fossils are widespread in certain time intervals and can come to define them. Over time, they represent evolutionary change. (USGS)

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.

An Inherently Biased Record: Signor-Lipps Effect

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.

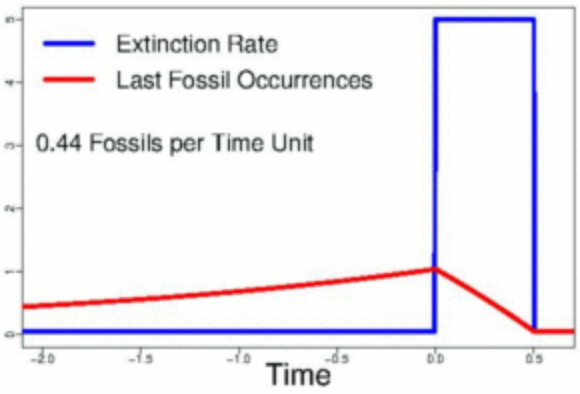


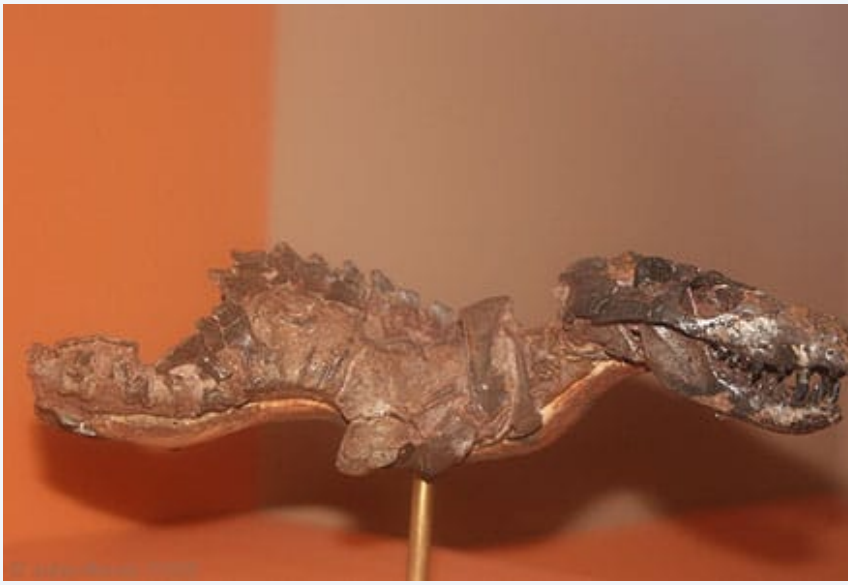
Figure 9.4.4: An animation of the Signor-Lipps effect, demonstrating how the rate of last fossil occurrences approximate the extinction rate as the number of fossils per time unit increases (Wikipedia: NikelsenK)

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
Avialae	*Archaeopteryx lithographica*	Jaws with sharp teeth, Long bony tail, Feathers	

Fossil Group or Clade	Organism	Features	Image
Stegocephalia	*Tiktaalik roseae*	Scales, Gills, Flattened head, Sturdy interior fin bones, Ears capable of hearing in both air and water	
Hominin	*Ardipithecus ramidus*	Grasping big toe, Reduced canine tooth size, bipedal features in the pelvis and base of skull	

Fossil Group or Clade	Organism	Features	Image
Epicynodontia	*Thrinaxodon liorhinus*	Possibly had hair, secondary palate, Nasals pitted with foramina, inner ear bone	
Cetacea	*Basilosaurus cetoides*	First fully aquatic mammal, nearly complete loss of hindlimbs	

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 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 9.4.5: Glossopteris sp. leaves from the Permian of Australia.



Figure 9.4.6: 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 (sudden appearances) or outages (sudden 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.

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.

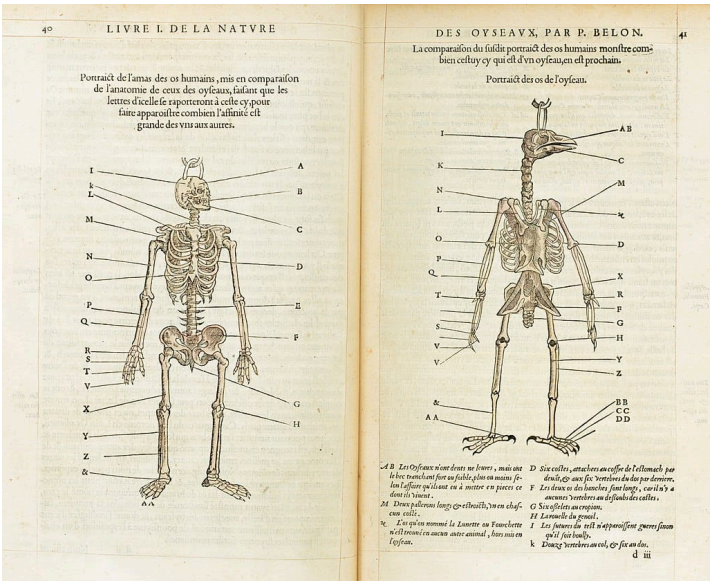


Figure 9.4.7: Pierre Belon's early attempt at an analysis of homologous traits in his book, Natural History of Birds, 1555.

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 wings 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.

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.

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

Pan troglodytes right forelimb by Blackburn Lab on Sketchfab

Snowy egret right forelimb

Snowy Egret Wing Skeleton by Blackburn Lab on Sketchfab

Figure 9.4.8: 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.

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.

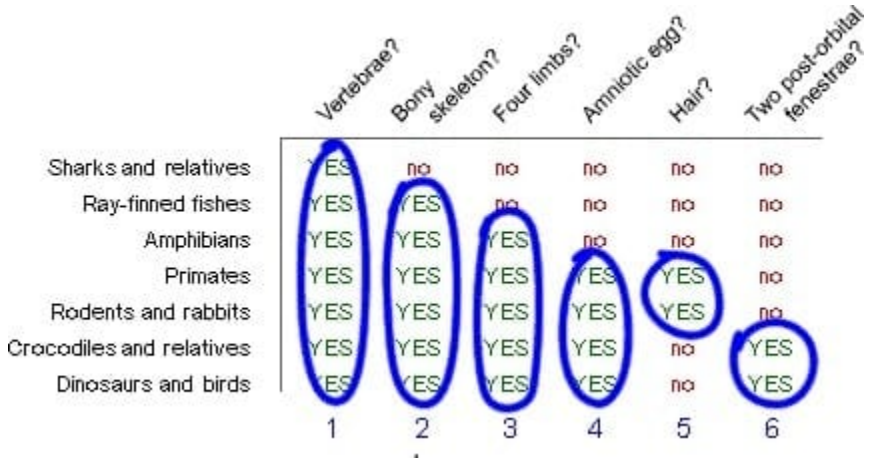


Figure 9.4.9: A character matrix can be used to analyze evolutionary relationships among organisms. Shared derived characters, rather than shared ancestral traits, are used to parsimoniously differentiate between a group of organisms. (Image: University of California Museum of Palentology, https://evolution.berkeley.edu/admin/media)

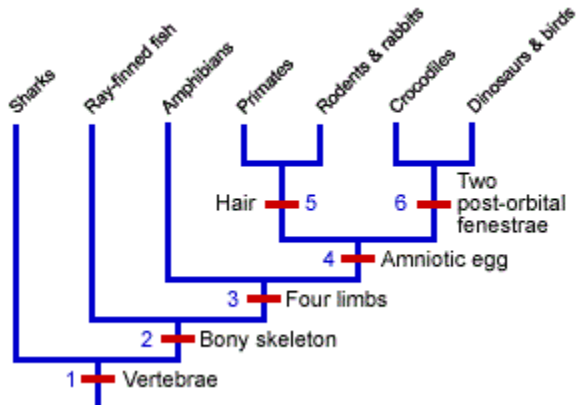


Figure 9.4.10: From a character table, a cladogram can be created to visually explore shared derived characters, their development through speciation (origination), and their development over time. In this case, connections between sharks, primates and other mammals, and dinosaurs are explored in a larger 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 9.4.10) 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.

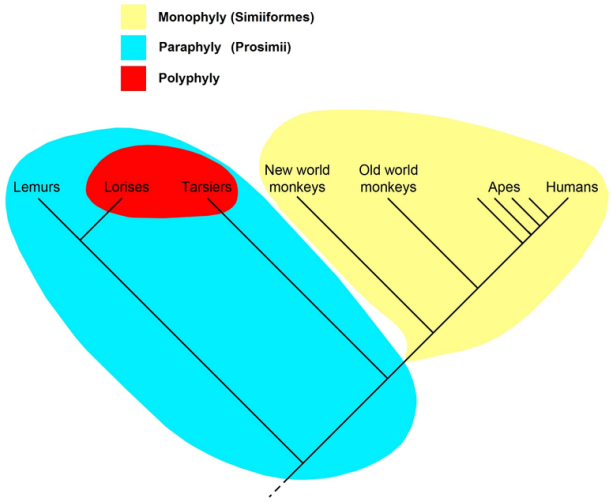


Figure 9.4.11: A primate cladogram showing the differences, or groups, of different kinds of taxon. Monophyletic taxon are in yellow. Paraphyletic taxon are outlined in blue. Polyphyletic taxon are outlined in red.

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.

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!

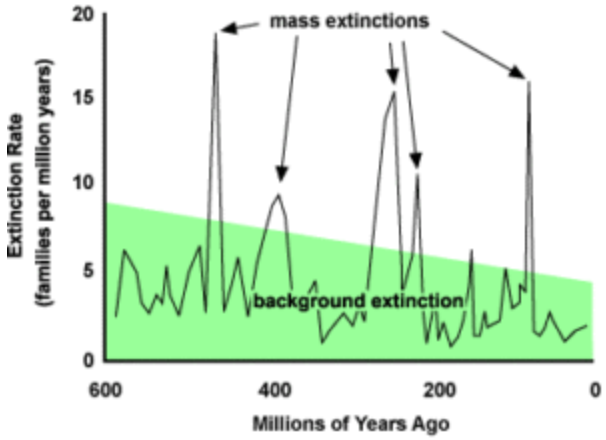


Figure 9.4.12: Mass extinctions versus background extinctions. The green shaded area represents extinction that is occurring all the time. Species turnover at low rates, or even when rates are a bit elevated above background levels, are not unusual (Permission: University of California Museum of Paleontology's 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.

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".

Key Terms

- **background extinction** - the normal, ongoing rate of species loss that occurs as a result of natural ecological and evolutionary processes
- **homology** - a similarity in structure, genetics, or developmental pathways among different organisms that results from shared ancestry
- **mass extinction** - a short time interval during which an exceptionally high proportion of Earth's species becomes extinct across multiple ecosystems and taxonomic groups
- **vestigial structures** - anatomical features or behaviors that persist in an organism but have lost most or all of their original function through evolution

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9.5: Key Points

1. Evolution results from the mutation of genetic material and the subsequent accumulation of changes in DNA through many generations of offspring.
2. Evolution represents a coherent view of life and is built upon a foundation of evidence from the fossil record, biochemistry, observations of naturalists, and observations still being made in labs, in nature and on other planets.
3. Carl von Linne introduced binomial nomenclature to create a simple taxonomy in which organisms were classified in a nested hierarchy with Kingdoms at the highest rank and then divided into “Phyla,” then “Classes,” “Orders,” “Families,” “Genera,” and finally “Species.”
4. Early evolutionary theory introduced natural selection, the process by which species change over time as organisms with enough favorable characteristics survive long enough to pass on their genes.
5. Understanding how natural selection worked relied on the understanding of how genes were passed down from parents to offspring and how certain traits are passed on or expressed.
6. Several key processes drive evolutionary change: genetic mutation, gene migration, genetic drift, horizontal gene transfer, and natural selection.
7. Sexual selection involves species preferentially having their genetic material passed on due to having more desirable physical features.
8. Artificial selection is controlled by humans as we manipulate what genes are passed down from one generation to another such that the species evolves to have favorable traits.
9. Early on, fossils were not though of as evidence of past life but eventually they were accepted as evidence of ancient life.
10. By studying fossils early on, the concept of extinction was introduced.
11. Once the ages of various fossils were established, scientists concluded the age of the rock containing the fossil had to be the same age as the fossil itself.
12. Transitional fossils show the link between two major groups as they have traits of the ancestral and descendant groups.
13. Extinction of species is important to evolutionary theory as it makes space for other species to evolve into.
14. Mass extinctions occur as a large number of the species on Earth die off. There are five widely accepted mass extinction events: 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

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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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10: HADEAN AND ARCHEAN

Introduction

Earth formed about 4.6 billion years ago from a solar nebula, along with the rest of our solar system. It grew through the accretion of nebular material that ranged in size from space dust to planetesimals (minute planets). In its earliest stages, our planet looked and behaved much differently than the majestic blue and green planet we know today. Since the Hadean, Earth has experienced many events (Figure 10.1) that have allowed life as we know it to evolve.

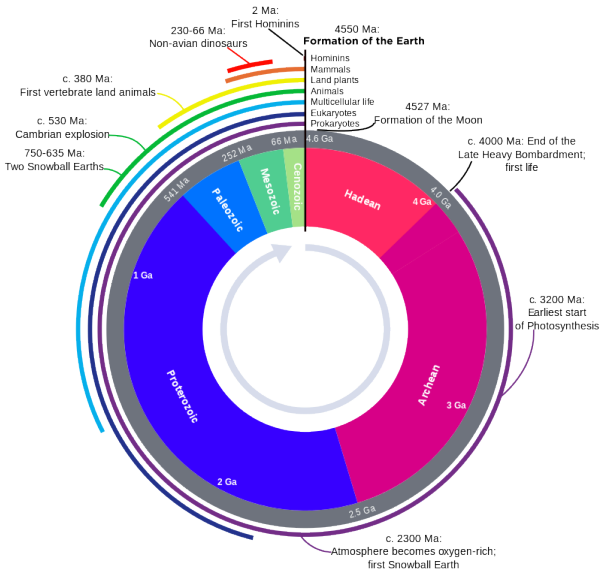


Figure 10.1: Geologic time on Earth, represented circularly, to show the individual time divisions and important events. Ga=billion years ago, Ma=million years ago. From An Introduction to Geology, Chapter 8: Earth History.

Here is the the simplified geologic time scale used in the Geologic Time chapter of this book. The Hadean and Archean eons are highlighted at the bottom.

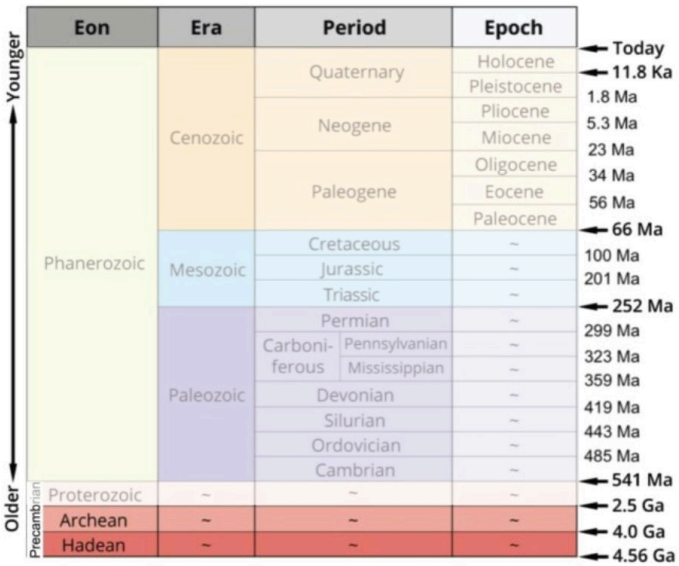


Figure 10.2: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

10.1: Hadean Eon

10.2: Earth's Oldest Rocks

10.3: Archean Eon

10.4: When Did Life Begin?

10.5: Key Points

10.6: Further Reading

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10.1: Hadean Eon

The Geologic Time Scale

The first two "chapters" in the story of Earth are the Hadean and Archean Eons. During these times, Earth transformed from an uninhabitable hellscape into a place that started to resemble the planet we know today. The names Hadean (derived from Hades, the Greek God of the underworld) and Archean (from the Greek word for the beginning) come from the **Geologic Time Scale**, the table of contents for the "Book of Earth." It chronologically arranges and names chapters in Earth's history according to major geologic and biologic events in Earth's history, recorded as significant changes in rocks and fossils.

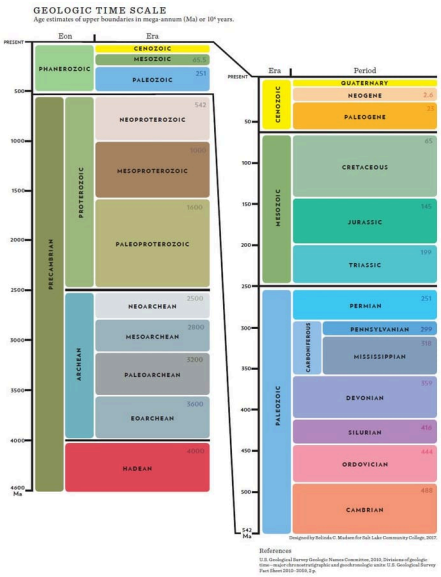


Figure 10.1.1: Geologic Time Scale with ages shown

Hadean Eon

Earth's oldest Eon of geologic time is appropriately named the "Hadean", after the Greek god of the underworld, Hades, and dates from 4.6–4.0 billion years ago. The **Hadean** can be described as "Hell on Earth," the time when an extremely hostile environment existed with magma oceans boiling on the surface, widespread volcanic activity, a superheated atmosphere of noxious gases, and frequent bolide (meteorites, comments, and astroid) impacts.



Figure 10.1.2: Artist's impression of the Earth in the Hadean.

The new Earth was incredibly hot due to gravitational compression, radioactive decay, and bolide impacts. Most of this initial heat still exists inside the Earth, which is crucial for driving geologic processes today, like plate tectonics. As Earth cooled through this Eon, the planet began to take on its current structure.

Differentiation of Earth's Interior

Planetary geologists believe that toward the end of the **accretionary** stage in Earth's formation, 4.5 – 4.6 Ga (billions of years ago) when the planet was being assembled, Earth was pummeled by larger, planetesimal-sized space debris. 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 inside the early Earth. This double whammy of interior and exterior heat 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, iron-associated elements sank toward the core and lighter elements rose toward the surface.

Earth's Original Crust

As differentiation proceeded, the surface magma ocean cooled, and a thin, early crust began to form from high-melting-point silicate minerals (see Bowen's Reaction Series), such as olivine and calcium-rich plagioclase, that could crystallize at high temperatures. Meteors, asteroids, and comets continued their impacts, puncturing the earliest crust so magma continually welled up, flowing over Earth's surface and crystallizing into komatiite, a volcanic rock composed of the mineral olivine. Rising heat from the core prevented large crustal blocks from remaining intact. Denser mafic and ultramafic minerals (minerals rich in iron and magnesium) sank to form the mantle, while the highest-density materials, like iron and nickel, sank into the core. Differentiation transformed Earth from a homogeneous sphere of molten material into a heterogeneous planet with distinct layers: a felsic and mafic crust, an ultramafic mantle, and an iron and nickel core.

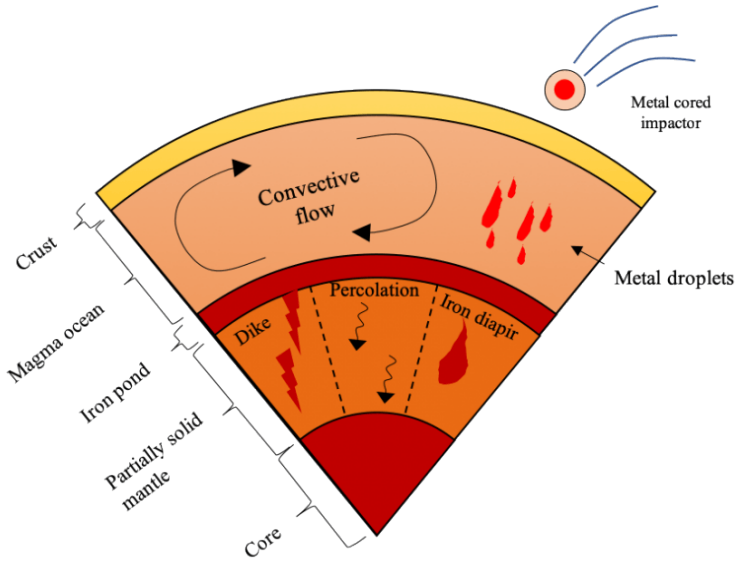


Figure 10.1.3: Hypothetical core-mantle differentiation processes: Percolation, diking, and diapirism. After Rubie et al. (2015). (CC BY-SA 4.0; By AlexInMetal – Own work, on Wikimedia.)

Magma Composition

Differentiation produced our iron-rich inner and outer cores, the silicate-dominated mantle, and, ultimately, the crust. The relatively low-density silicates are minerals composed chiefly of silicon and oxygen, bound with other oxygen-seeking elements such as aluminum, calcium, potassium, sodium, iron, and magnesium. In the early Hadean, Earth was continually bombarded by space debris containing all Earth's natural elements in varying proportions.

Heat powered a convecting ocean of silicate mush, which brought diapirs (teardrop-shaped blobs of deep magma) toward the surface. There, it pooled and the exterior cooled in thick komatiite masses, forming the first crustal pieces. The dense komatiite masses were partially melted by heat from below, causing them to drip back into the magma ocean for recycling. This remelting process further differentiated the early crust, as the "partial melt" was more silica-enriched: the most silica-rich minerals in the komatiite would be the first to melt. This "evolved" silica-rich magma was also less dense and more buoyant, allowing it to rise and float on Earth's surface. This process of magma evolution leads to the varying compositions we recognize today when classifying igneous rocks. Notice where komatiite appears on the diagram below, all the way to the right, with the least amount of silica. As magma evolution proceeds, the composition becomes more silica-enriched, creating magma that becomes increasingly more "felsic" in composition.

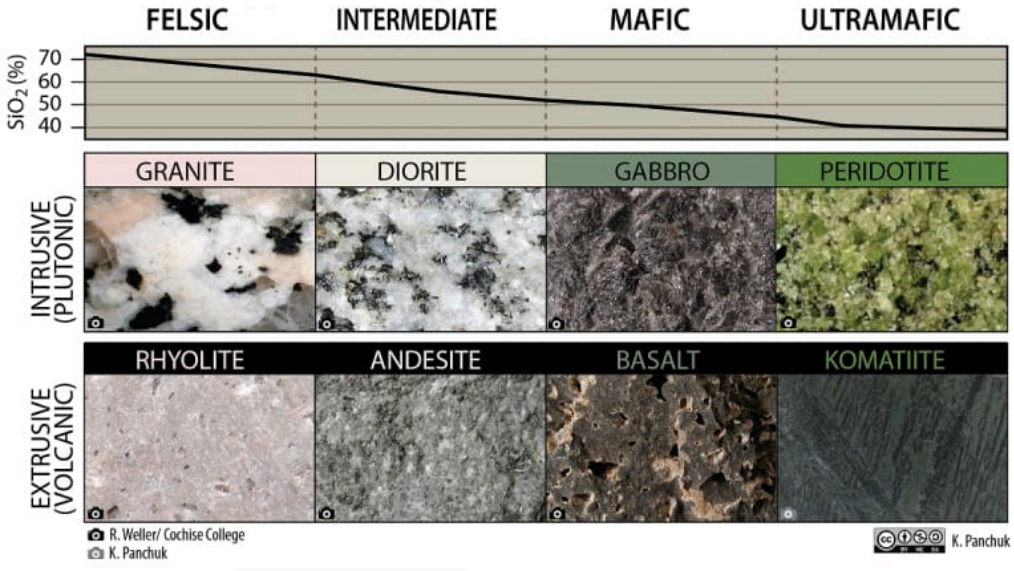


Figure 10.1.4: Igneous rock classification diagram.

As Earth cooled and the mantle solidified, komatiite became increasingly rare because there wasn't enough heat to melt the minerals needed to produce the highly ultramafic lavas required to form it. [3] The exact formation of the felsic crust of our continents is debated because making granite requires plate tectonic processes, and the timing of the start of plate tectonics remains unknown. Fortunately, recent discoveries in the Jack Hills of western Australia may hold the evidence needed to end this debate.

The Jack Hills Zircon

What followed the formation of thin, perturbed komatiite crust is highly debated and an area of intense 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 (map below). It is a common mineral in felsic rocks, with a composition similar to that of continental crust, such as granite. It is uncommon in oceanic crustal rocks, such as basalt, and typically does not occur in ultramafic rocks, such as komatiite.

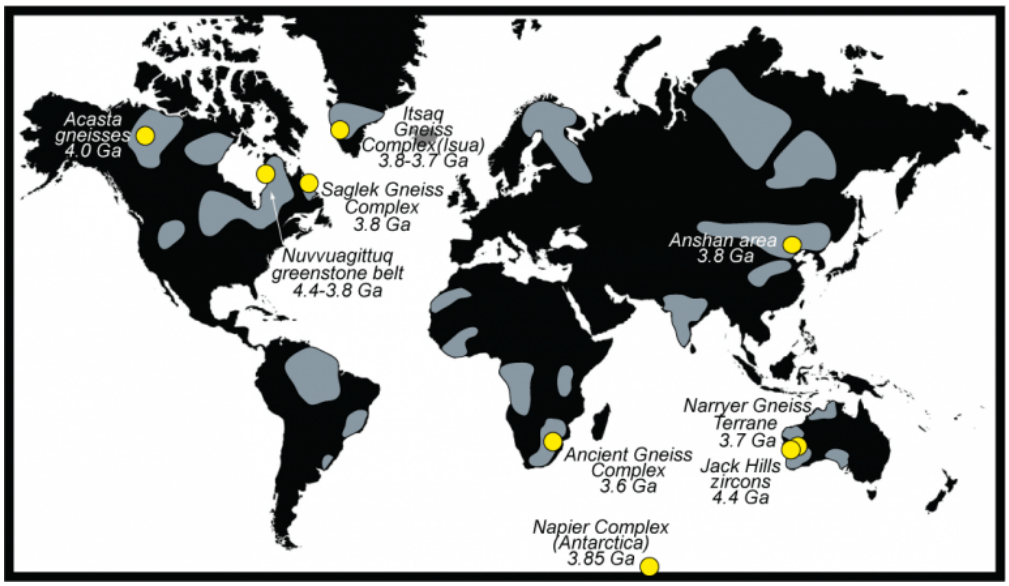


Figure 10.1.5: Archean Cratons (light grey) and locations of Earth's oldest rocks on a world map. By Jonathan O'Neil. (With permission from: UOttawa Early Earth Blog)



Figure 10.1.6: 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, and the sediment contained in the metaconglomerate must be older than the rock itself, according to the principle of inclusions (particles contained within a rock are older than the rock itself). The original conglomerate's depositional environment may have been an alluvial fan, a fan-shaped pile of sediment formed where a stream flows out of a mountain. Zircon crystals were extracted from the metaconglomerate for analysis.

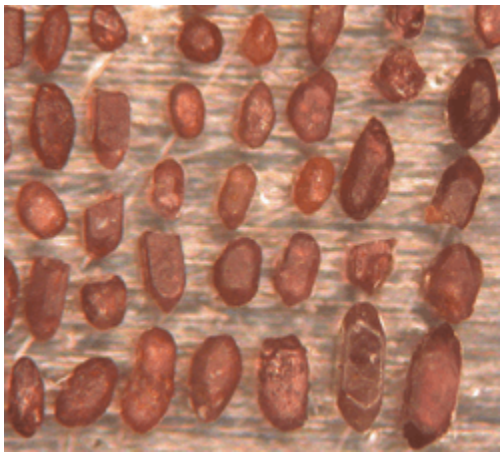


Figure 10.1.7: This image shows an array of Jack Hill zircons arranged on a piece of tape. Largest zircons ~ 1mm. (CC BY-NC-SA 3.0; Science Education Resource Center)

Zircon is a tiny, very durable mineral and a natural timekeeper. Zircon typically forms during magma crystallization, where radioactive uranium can substitute for zirconium in the mineral lattice. Following crystallization and the formation of an igneous rock, 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 results in a decrease in the number of protons, ultimately changing uranium into lead. The rate of this decay is well known, allowing scientists to accurately date zircon. Radiometric dating analysis of the Jack Hills detrital zircon grains yields 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!

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, a crust of varying composition existed, some of which was more felsic in composition, more like the continental crust that exists today, and that mountains had been uplifted from which sediment was transported by water. So, did plate tectonics start 4.4 billion years ago and, with it, the formation of our modern continents? Considering other evidence, this seems unlikely; however, this

discovery implies that modern geologic processes began shortly after Earth's structure formed and that the earliest Earth more closely resembled today's Earth than previously believed.

Analysis of oxygen-isotope ratios (atoms with varying numbers of neutrons) in zircons revealed even more incredible evidence about the environment of the early Hadean. Analysis of the relative amounts of different isotopes of oxygen $^{16}\mathrm{O}$ and $^{18}\mathrm{O}$ (ratio denoted with the lowercase Greek delta $\delta^{18}\mathrm{O}$) in the Jack Hills zircon are skewed toward “heavy” $^{18}\mathrm{O}$, as opposed to the more common “light” $^{16}\mathrm{O}$, indicating it formed by cool, wet, sedimentary processes at the Earth’s surface. This could mean that the magma that formed the zircons came from melted seafloor sediment. So, not only was the very young Earth capable of making rock more felsic than its original ultramafic komatiite crust, but it was also cool enough to have liquid water in oceans. These are surprisingly familiar conclusions about a “hellish” young planet.

Since the discovery of Hadean-age zircon in the Jack Hills area of Australia, other detrital zircons have been found in Archean-age rocks in other parts of the world.



Figure 10.1.8: Hadean age zircon discovery locations on a world map. (CC BY 4.0 International; By karaclc on Wikimedia.)

Origin of Earth’s Water

Explanations for the origin of Earth’s water include volcanic outgassing, comets, and meteorites. The volcanic outgassing hypothesis for the origin of Earth’s water holds that it originated within the planet and emerged through 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 composed of dust and ice, with some or all 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. However, the chemistry of Earth’s water isotopically matches water found in meteorites better than that of comets. In particular, chondrite meteorites contain water in the crystal structure of some of their minerals. This water could have been released from minerals when these meteorites bombarded the very young, hot Earth. Oxygen would have been released upon impact and melting, allowing it to combine with atmospheric hydrogen to make water.

YouTube Video

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In the end, all three sources likely contributed to the origin of Earth's water.

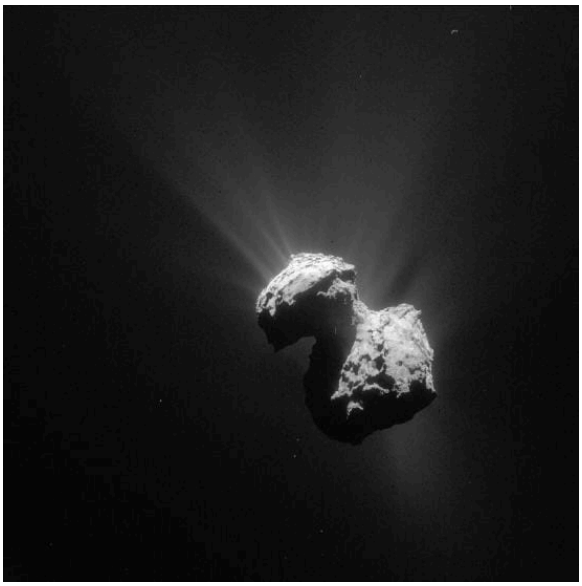


Figure 10.1.9: Water vapor leaves comet 67P/Churyumov–Gerasimenko.

Key Terms

- **Geologic Time Scale** - a graphical representation of Earth's history based on the rock record that shows the timing and relationships of geologic and biologic events
- **Hadean** - the earliest part of the Precambrian Eon starting at the conception of Earth up to 4 Ga
- **accretion(ary)** - the process of small particles and gases within the solar nebula combining together to form larger particles and, eventually, planets
- **differentiation** - the process of a planet developing a layered structure according to density and chemical composition

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10.2: Earth's Oldest Rocks

The Acasta Gneiss Complex

The Jack Hills detrital zircons indicate that rocks that were eroded into sediment were present around 4.404 Ga. Ancient rocks exceeding 3.5 billion years old are found on all continents, but 4.4 Ga rocks have not been found.

The oldest intact rock on Earth is the Acasta Gneiss Complex, part of the North American craton in Northwest Canada. This complex includes a variety of highly deformed and metamorphosed tonalite-trondhjemite-granodiorite (TTG) rock. TTGs are similar to granite in that they are silica-rich and felsic, but they differ in their mineralogy. They are typical of igneous rocks produced from magma along active continental margins today.



Figure 10.2.1: 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 radiometrically dated using U/Pb isotopic dating techniques on zircon formed during the crystallization of these rocks. The presence of zircon in an indicates that the original TTG was derived from magma that had “evolved” (it was formed through remelting a pre-existing rock) and that solidified between 3.92 and 4.02 Ga. Its composition, similar to that of continental crust rocks today, indicates that continental crust existed in the Hadean. Extensive geochemical analysis has led to the interpretation that the oldest rock of the Acasta Gneiss Complex was derived from the partial melting of 4.3 Ga mafic Hadean crust [1] [2].



Figure 10.2.2: Acasta gneiss in exposure. Notice the tight folds from tectonic deformation. (CC BY NC ND 4.0 International; By: The Traveling Geologist)

The Nuvvuagittuq Greenstone Belt

The dating of the rock from the Nuvvuagittuq Greenstone Belt (NGB) has been controversial, but if accepted, it may actually provide the best representation of Earth's earliest crust. The NGB is located in northern Quebec, on the eastern shore of Hudson Bay. It's mostly metamorphosed mafic and ultramafic volcanic rock that initially erupted in an ancient ocean, as evidenced by the presence of pillow lava: A lava structure that forms during underwater eruptions. Lava that erupts underwater forms conspicuous pillow shapes as a crust solidifies instantly around the oozing lava as it spills out underwater.

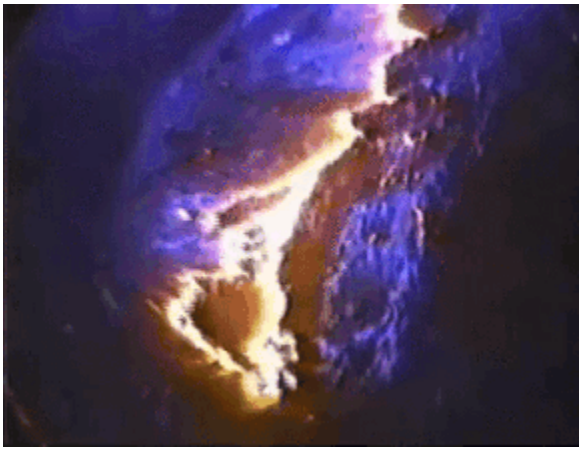


Figure 10.2.3: Modern pillow lava forming underwater.

Another rock type in the NGB is banded iron formation (BIF). BIFs are sedimentary rocks that formed in the ocean as sedimentary iron minerals formed and settled out of the water column.



Figure 10.2.4: 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.)

Around the NGB is a felsic intrusive rock called tonalite. The NGB and tonalite are cross-cut by a pegmatite, another type of intrusive igneous rock. The tonalite and pegmatite intrusive bodies contain zircons, which have been accurately dated at 3.77 Ga. This means that the NGB is at least 3.77 Ga old, since, according to the principle of cross-cutting relationships, it must have existed to be cut by the pegmatite.



Figure 10.2.5: 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 at McGill University in Quebec, Canada, challenged the assumed date for NGB by investigating an odd-looking amphibolite patch within NGB. An amphibolite is a metamorphosed mafic rock and does not contain zircon. O'Neil used a dating technique and isotopic-ratio comparisons of the rare but ubiquitous elements samarium (Sm) and neodymium (Nd). Using this technique, O'Neil determined the actual age of the NGB to be 4.28 Ga [3]. Further analysis of additional samples since then has pushed the date back even further to 4.31 Ga [4]. If true, the NGB rock would represent the oldest preserved Hadean rock on the planet, providing a snapshot of Earth's earliest crust.

Not all geologists investigating these rocks concur with O'Neil's findings, and this controversy will likely continue until zircon is found within the NGB.

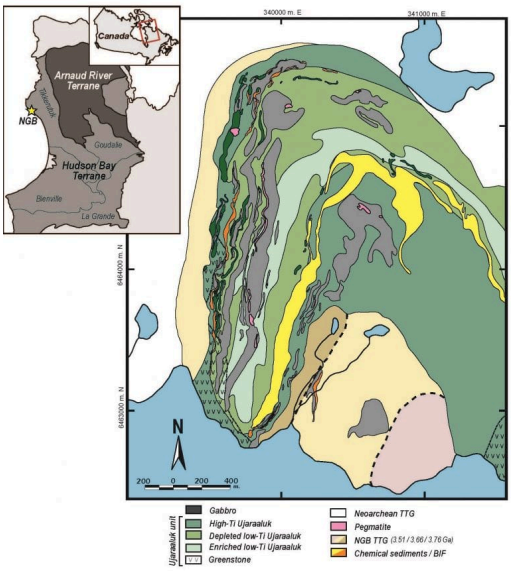


Figure 10.2.6: Geologic map of the Nuvvuagittuq Greenstone Belt (with permission by Jonathan O'Neil.)

Key Terms

- **Acasta Gneiss Complex** - located in Canada, it preserves the oldest rocks on Earth some of which where derived from the partial melting of 4.3 billion year old Hadean crust
- **Nuvvuagittuq Greenstone Belt (NGB)** - located in Canada, rock from this has been dated back to 4.31 Ga making it the oldest preserved rock on earth (though the dating is controversial)
- **tonalite-trondhjemite-granodiorite (TTG)** - rocks similar to granite that have been metamorphosed and are part of the Acasta Gneiss Complex

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10.3: Archean Eon

Overview

The Archean Eon, lasting from 4.0 to 2.5 billion years ago, is named after the Greek word for beginning, which seems appropriate, as this time period marks the beginning of the observable rock record: assuming that the 4 billion years old Acosta Gneiss is the oldest intact rock (see previous section), and the start of Earth operating with modern geological processes, such as plate tectonics. As we have already learned, rocks and minerals existed during the Hadean Eon, but the rock record from the Archean is considerably more complete.

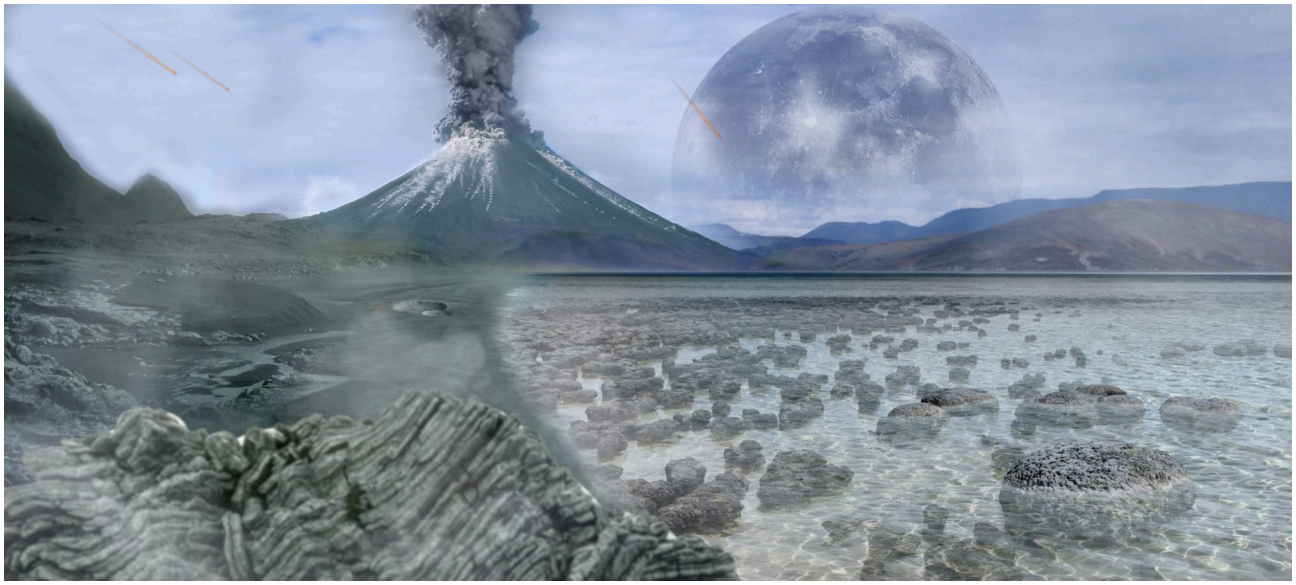


Figure 10.3.1: Artist's impression of the Archean. Note the meteorite, volcanic activity, and stromatolite deposits in the shallow sea.

Late Heavy Bombardment

Objects were chaotically flying around at the start of the solar system, building the planets and moons. Around the start of the Archean, about 4.1–3.8 billion years ago, a second large spike in asteroid and comet impacts occurred on Earth and the Moon. This event is called the **Late Heavy Bombardment**. Meteorites and comets in stable or semi-stable orbits became unstable and impacted objects throughout the solar system. During the late heavy bombardment, the Earth, Moon, and all the planets in the solar system were pummeled by material from the asteroid and **Kuiper belts**. Most of the Moon's craters are from this event, where samples have been collected to support the timing.

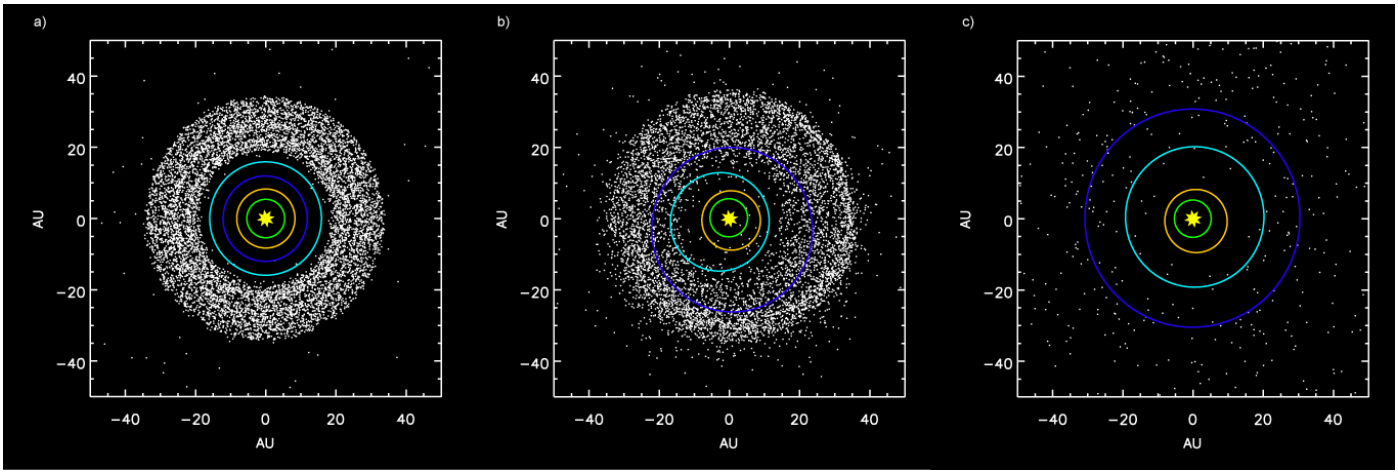


Figure 10.3.2: Simulation of before, during, and after the late heavy bombardment.

It is widely accepted that the planets in our solar system experienced extensive asteroid and comet bombardment during their formation; however, another process must have caused the second increase in impacts hundreds of millions of years later, once they were fully formed. A leading theory blames gravitational resonance between Jupiter and Saturn, in which the gravity of these two giants disturbed orbits within the asteroid belts and the Kuiper Belt, based on a similar process observed in another solar system, Eta Corvi.

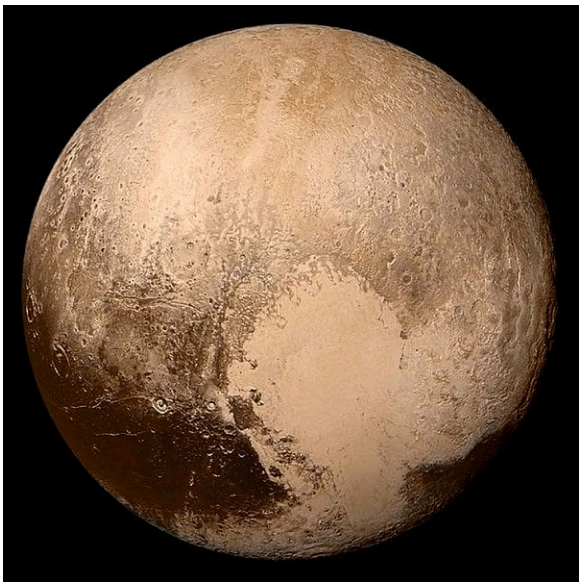


Figure 10.3.3: 2015 image from NASA's New Horizons probe of Pluto. The lack of impacts found on the Tombaugh Regio (the heart-shaped plain, lower right) has been inferred as being younger than the Late Heavy Bombardment and the surrounding surface due to its lack of impacts.

The First Continents and Plate Tectonics

The continents were formed incrementally throughout Earth's early history. After all, only about 8% of continental crust is of Archean age (*Goodwin A.M. Principles of Precambrian Geology. Academic Press; London, UK: 1996.* [Google Scholar]). Initially, when Earth was hotter, lighter, more felsic materials floated upward, while heavier ultramafic materials and metallic iron sank. This differentiation process led to the layering of the Earth, with lighter materials forming the lithosphere and continental crust, and heavier materials forming the interior layers. This internal layering is revealed through seismic data. While the timing and specific processes remain debated, volcanic activity must have brought the first continental material to Earth's surface during the Hadean, and the continents formed through some incremental process during the Archean. Forming continents was initially challenging because a significantly hotter core drove vigorous mantle convection that rifted apart young, thin, and weak crustal pieces.

For plate tectonics to work as it does currently, there must be continents. As the Earth cooled, permanent continental crust could form, and plate tectonic-like activity could begin. The video below, "The World Before Plate Tectonics," provides a useful discussion. It will start at 1:49 and should be viewed until at least 4:22.

YouTube Video

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It seems that early plate tectonic processes started at least at the end of the Archean, suggesting that a stable continental lithosphere was already in place. An earlier beginning is possible, but evidence of earlier tectonic activity has been erased by subduction and the rock cycle. In any event, once stable pieces of continental crust were established, plate tectonics moved them, and they started to grow through accretion (smashing together) and continental volcanic activity.

YouTube Video

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The cores of the continents were made in the Archean. These tectonically stable interiors are called **cratons**. Cratons have two main parts: the shield, composed of crystalline basement rocks such as gneiss and greenstone, and the platform, which is covered by sedimentary rock. Most cratons have remained relatively unchanged, with most tectonic activity, such as subduction and accretion, occurring around cratons instead of within them. Shields were the building blocks for the Proterozoic continents that now define Earth's land surface, and contain the oldest rocks.

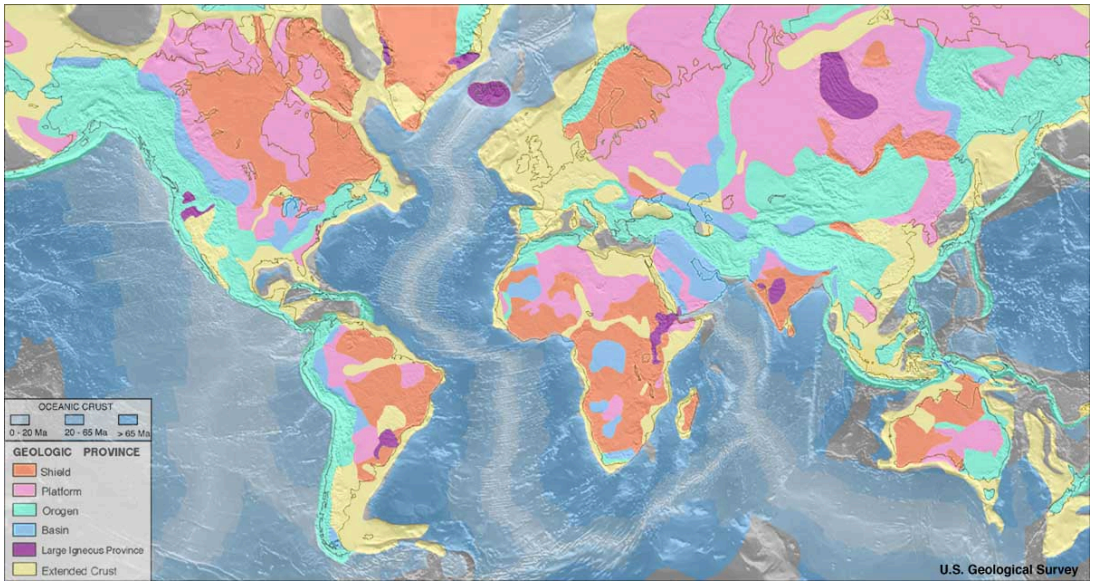


Figure 10.3.5: Geologic provinces of Earth. Cratons are pink and orange.

Sedimentary Rocks

Interestingly, the oldest Archean sedimentary rocks are entirely marine chemical sedimentary rocks. There are no siliciclastic (detrital) sedimentary rocks, which are made from sediments derived from the physical weathering of crustal rocks. There are no sandstones or conglomerates, for example. This implies that no land was exposed above sea level, and no rock exposures were available for weathering and erosion.

Key Terms

- **accretion** - the addition of crustal material through collisions
- **chemical sedimentary rocks** - rocks that form through mineral precipitation in water
- **craton** - tectonically stable interiors of continents composed of the shield, which is crystalline basement rock like gneiss and greenstone, and the platform of sedimentary rocks covering the shield
- **Kuiper Belt** - a doughnut-shaped region of icy objects beyond the orbit of Neptune; it is home to Pluto and most of the known dwarf planets and some comets
- **Late Heavy Bombardment** - a period of extraordinary meteorite impacts at the start of the Archean, thought to have been triggered by the gravitational interaction of Jupiter and Saturn

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10.4: When Did Life Begin?

First Life on Earth

Life most likely started during the late Hadean or early Archean eon. While the timing and location of the origin of life on Earth remain debated, it is widely recognized as the most critical, planet-changing event, and various theories have been developed to explain the conditions that enabled its emergence.



Figure 10.4.1: Fossils of microbial mats from Sweden

Some of the most concrete evidence of early life comes from **stromatolites**, fossilized microbial mats dating to 3.5 billion years ago, found in Australia. These photosynthetic microorganisms, like cyanobacteria, lived in the shallow oceans with abundant sunlight. They trapped sediment as they grew, resulting in a distinctive layered, mound-like structure (see below).



Stromatolite mounds. Image from Wilson44691, CC0, via Wikimedia Commons.

It's likely, however, that life on Earth started earlier, perhaps even in the Hadean. For example, microscopic filaments from a hydrothermal vent deposit in Quebec, Canada, prove that life could have started as early as 3.8–4.3 billion years ago.

Conditions for First Life

Conventional scientific thought suggests that the extreme heat from the early meteorite bombardment of Earth during the Hadean would have made conditions at the surface too harsh for life to begin, with portions of the early crust melted from these collisions (e.g., Marchi et al., 2014). Despite the relentless impacts of the Late Heavy Bombardment at the start of the Archean, though, some studies 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 about 4.3 billion years ago, relatively early in the Hadean. While it's possible life arose during this earliest chapter of Earth's history, finding evidence is difficult because the Hadean is poorly represented in the rock record.

Despite the extreme conditions, signs of life are preserved in very old rocks. Some of the oldest, the 3.8-4.3 billion-year-old Nuvvuagittuq Greenstone Belt (NGB), contains **banded iron formations** (BIF) with microscopic tubes of iron mineral hematite. These tubes match the shape and size of those made by bacteria in modern hydrothermal vent environments, where they consume 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. Lifeforms prefer the lighter isotope of carbon, carbon-12. If this discovery holds true, this will be the oldest evidence of fossil life yet discovered. The NGB banded iron formation can be interpreted as having been deposited on the ancient seafloor, as represented by greenstone. This NGB has been dated by cross-cutting igneous intrusions to have formed at least 3.77 Ga. This would predate any previously discovered fossil evidence by at least several hundred million years.

Video: World's Oldest Fossils (UCL)

YouTube Video

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Hydrothermal Vents

Today, hydrothermal vents are found in the deep ocean along mid-oceanic ridges at divergent plate boundaries and 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 and mixed with fluids released from the rising magma. Superheated water can transport numerous dissolved minerals and mobile ions, which are transported to the seafloor and then precipitate 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 present is the key to the past", the first organisms at Hadean hydrothermal vents were therefore also likely sulfur-reducing bacteria. These bacteria were chemoautotrophs, meaning they extracted the energy necessary for cellular functions from inorganic compounds. In the case of sulfur-reducing bacteria, they “breathe” sulfate (), and reduce sulfur to hydrogen sulfide (). Sulfur-reducing bacteria are found today at the Mid-Atlantic Ridge and the East Pacific Rise, and in pools associated with land-based hydrothermal activity, such as the hot springs and geysers of Yellowstone National Park in Wyoming. These prokaryotes belong to the Domain Archaea, which encompasses a diverse array of microbes that can survive in chemically harsh conditions across a wide range of temperatures.



Figure 10.4.2: Apex Chert of western Australia contains some of the oldest known fossils

How Did Life Begin?

The origin of life on Earth is not yet known. 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 Earth's early atmosphere and oceans, which was very different from today's. The oxygen-free atmosphere produced a reducing environment with abundant methane, carbon dioxide, sulfur, and nitrogen compounds. This is what the atmosphere is like on other bodies in the solar system. Researchers in the famous Miller-Urey experiment simulated early Earth's atmosphere and lightning within a sealed vessel. After igniting sparks within the vessel, they discovered the formation of amino acids, the fundamental building blocks of proteins.

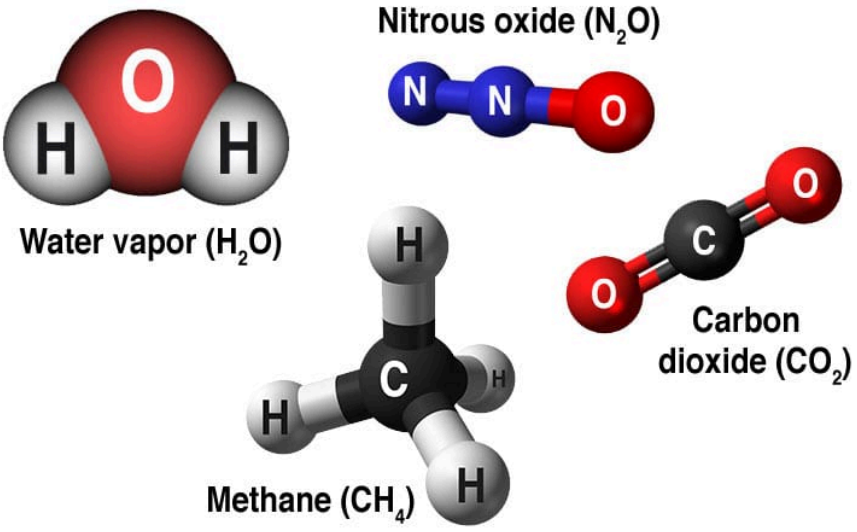


Figure 10.4.3: Greenhouse gases were more common in Earth's early atmosphere.

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

Another explanation is that life began around hydrothermal vents on a deep-sea mid-ocean ridge. Life was discovered along modern mid-ocean ridge systems in 1977, sparking this idea. Using uniformitarianism, we can reason that hydrothermal vents

also provided the nutrients for life in the ancient past.

Another possibility is that life or its building blocks came to Earth from space, carried aboard comets or other objects.

Video: Did Asteroids Spark Life?

YouTube Video

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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.

The Prokaryotic Pioneers

The first life forms on Earth were likely sulfur-reducing prokaryotes. **Prokaryotes** are single-celled bacteria, small in size (0.1 – 10.0 μm), with no nucleus or other organelles. Nevertheless, these simple organisms have changed the trajectory of Earth's history.

Two of the three Domains of life depicted in the Phylogenetic Tree of Live below, Archaea and Bacteria, are prokaryotes. Eukarya, to which you belong, is the third. 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.

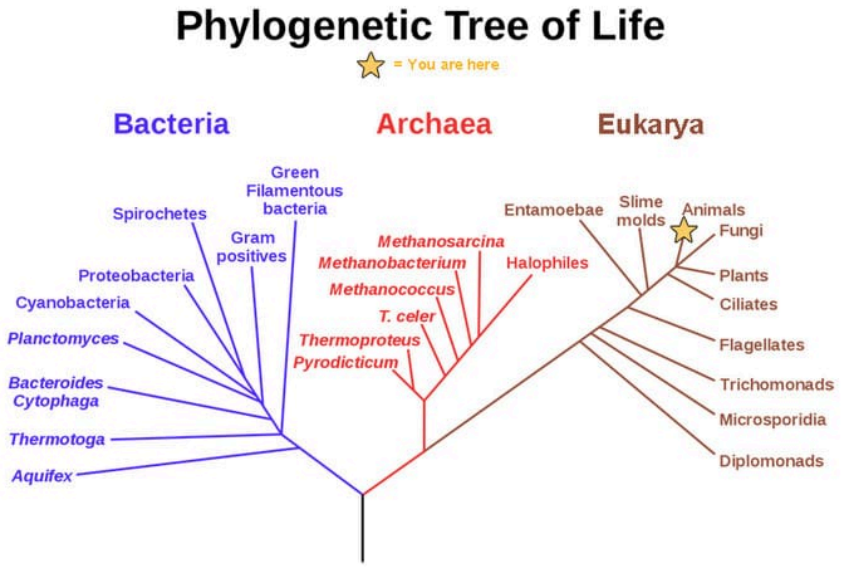


Figure 10.4.4: Phylogenetiac tree of life showing the three major Domains. (Public Domain image.)

In many rocks of similar age as those preserving the Archean sulfur-reducing bacteria, 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 Australia's 3.5 billion year old (Archean) Dresser Formation.



Figure 10.4.5: Stromatolite from the Archean Dresser Formation, Western Australia. (Creative Commons Attribution 2.0 Generic license.)



Figure 10.4.6: Modern unicellular bacteria from a microbial mat in Guerrero Negro, Baja California, Mexico. (Public Domain via NASA.)

Cyanobacteria are photoautotrophs that require sunlight to produce 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 most 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 slowest-building and life-on-Earth-changing events of all time, the Great Oxidation Event (GOE). BIFs help tell this story.

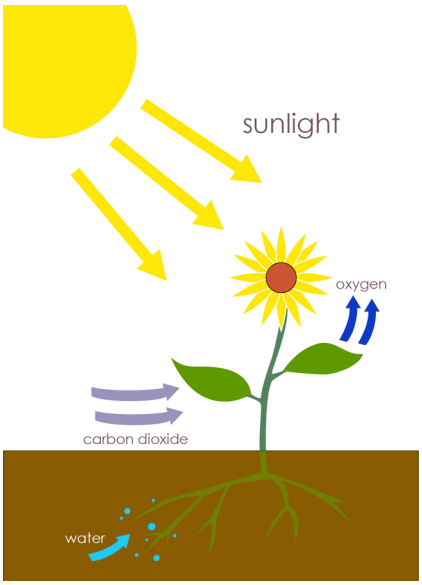


Figure 10.4.1: Diagram showing the main products and reactants in photosynthesis. The one product that is not shown is sugar, which is the chemical energy that goes into constructing the plant, and the energy that is stored in the plant which is used later by the plant or by animals that consume the plant.

An oxygenated world also significantly changed the planet's chemistry. For example, iron remained in solution in the non-oxygenated environment of the earlier Archean Eon. However, once the environment was oxygenated, iron combined with free oxygen to form solid precipitates of iron oxide, such as hematite or magnetite. These precipitates accumulated into expansive Archean and Proterozoic mineral deposits with red chert known as **banded-iron formations.**

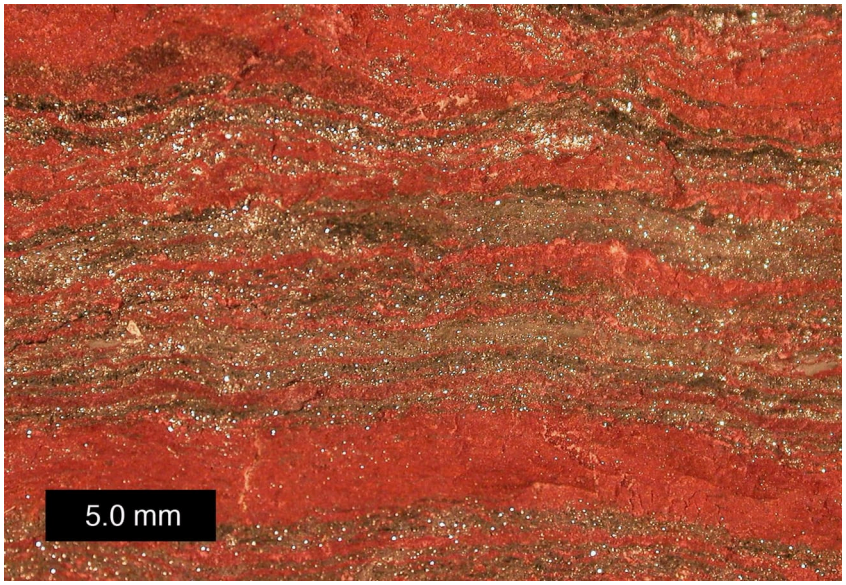


Figure 10.4.2: Alternating bands of iron-rich and silica-rich mud, formed as oxygen combined with dissolved iron.

The formation of iron oxide minerals and red chert (see figure) in the oceans persisted for a long time; it prevented oxygen levels from increasing significantly, since precipitation removed oxygen from 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 responsible for triggering Earth's first ice age around 2.1 billion years ago, the Huron Glaciation. Free oxygen reacted with methane (CH_4) in the atmosphere to produce carbon dioxide (CO_2). Carbon dioxide and CH_4 are called greenhouse gases because they trap heat within the Earth's atmosphere, like the insulated glass of a greenhouse. Methane more effectively insulates than CO_2, so as CH_4 was converted to CO_2 and the concentration of CH_4 decreased, the greenhouse effect also decreased, and the planet cooled.

The interpretation of BIFs is that excess oxygen produced by photosynthesizing cyanobacteria in shallow oceans was available to combine with iron ions in seawater, forming the dark, oxidized iron layers (see rock below). The source of iron is likely a combination of both ions released from hydrothermal vents and 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. The decline of BIFs in the rock record continued as atmospheric oxygen levels rose. With more atmospheric oxygen available, iron ions were oxidized on land before being transported to the oceans, consuming the iron that had previously been oxidized there. Consequently, BIFs stopped forming.



Figure 10.4.7: Banded iron formation, Barberton Greenstone Belt, South Africa. (Creative Commons Attribution-Share Alike 1.0 Generic license.)

Key Terms

- **Banded Iron Formations (BIF)** - sedimentary rock units with distinctive alternating layers of dark gray to black iron oxides and red to yellow chert that formed in seawater from oxygen produced by photosynthetic cyanobacteria
- **prokaryotes** - organisms whose cells lack a nucleus and other organelles
- **stromatolites** - layered sedimentary structures formed by the growth of bacteria or algae

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10.5: Key Points

- Earth formed about 4.6 billion years ago from a solar nebula, along with the rest of our solar system. It grew through the accretion of nebular material that ranged in size from space dust to planetesimals (minute planets).
- The Hadean was a time when magma oceans boiled on the surface, there was widespread volcanic activity and a superheated atmosphere of noxious gases, and frequent impacts by extraterrestrial objects.
- Earth's water may have originated from volcanic outgassing, comets, and/or meteorites.
- Earth's first crust was ultramafic to mafic in composition and was frequently rifted apart by heat rising from Earth's hotter core.
- Based on lab analysis of zircon grains found in metamorphosed sedimentary rock, a felsic continental crust may have started forming in the Hadean, perhaps as early as 4.4 Ga.
- The oldest intact rock on Earth is the 4.02 Ga Acasta Gneiss Complex of northwest Canada. Its composition, similar to that of continental crust rocks today, indicates that continental crust existed in the Hadean.
- The Nuvvuagittuq Greenstone Belt (NGB) may be as old as 4.3 Ga, and may be the best representation of Earth's earliest crust. It's metamorphosed mafic and ultramafic volcanic rock that initially erupted in an ancient ocean.
- The cratons or cores of the continents formed during the Archean as differentiation occurred and lighter materials formed continental crust.
- Plate tectonic processes started at the end of the Archean and added material to the cratons through accretion and continental volcanic activity.
- First life that is definitely preserved in the fossil record are 3.5 billion year old stromatolites beds in Australia. Still, microscopic hematite tubes imply that life could have started even earlier along hydrothermal vent systems like those found today along mid-ocean ridges.

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11: PROTEROZOIC EON

Introduction

The Proterozoic Eon, meaning “earlier life,” follows the Archean Eon and ranges from 2.5 billion to 541 million years old. During this time, most of the continental cores (cratons) had formed, and plate tectonic processes started. The formation and breakup of supercontinents dramatically disturbed the carbon cycle, leading to global glaciations known as "Snowball Earths".

Photosynthesis by microbial organisms, like cyanobacteria, completely transformed the oceans and later the atmosphere by adding massive amounts of free oxygen gas (O_2), initiating what is called the Great Oxygenation Event (GOE). Cyanobacteria were incorporated into larger cells through endosymbiosis, making the first eukaryotes, then multicellular organisms, then the first animals near the end of the Proterozoic Eon.

Here is the the simplified geologic time scale used in the Geologic Time chapter of this book. The Proterozoic Eon is highlighted at the bottom.

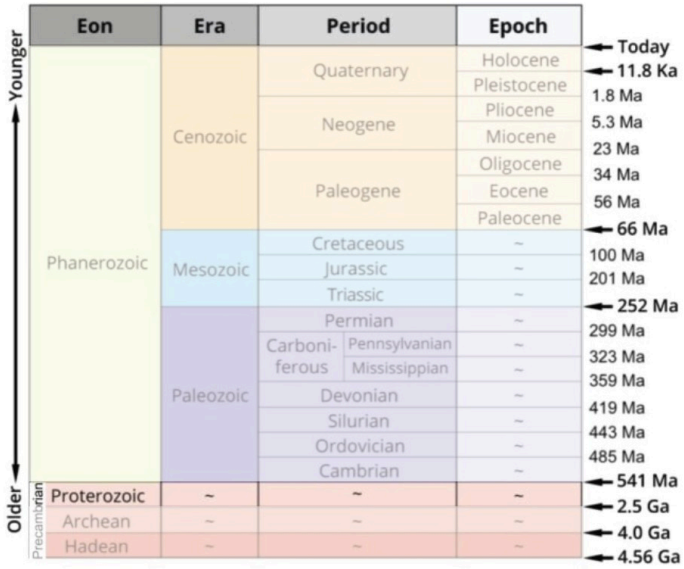


Figure 11.1: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

11.1: Proterozoic Geography

11.2: Proterozoic Glaciations

11.3: Proterozoic Atmosphere

11.4: Proterozoic Life

11.5: Key Points

11.6: Further Reading

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11.1: Proterozoic Geography

Rodinia

At the start of the Proterozoic Eon, most of the cratons had formed, and Earth's core had cooled sufficiently that crustal pieces weren't constantly being ripped apart by heat rising from the mantle. These Archean continents were moving, though it's unclear how closely their mechanism of movement resembled modern plate tectonics. Collisions of these ancient plates formed the **supercontinent Rodinia** about 1.2 billion years ago, causing the first mountain-building episode of the Appalachian Mountains. Geologists have reconstructed Rodinia by matching and aligning ancient mountain chains, assembling the continental crust pieces like a jigsaw puzzle, and using the **paleomagnetic** markers in rocks to orient and position the ancient continental pieces.

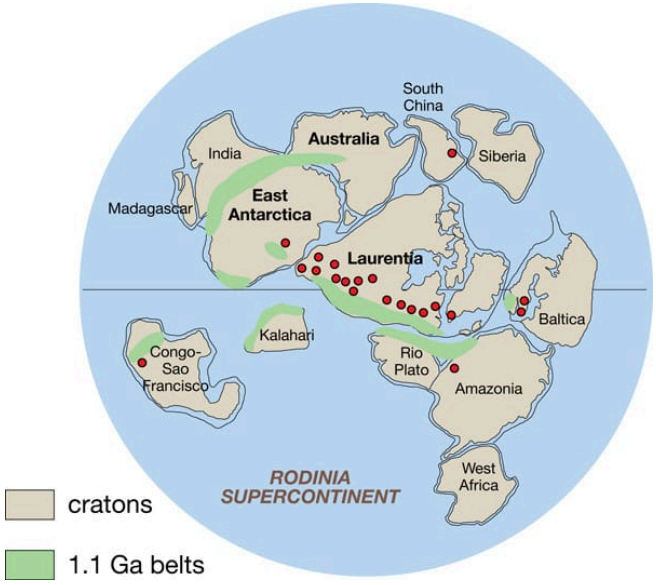


Figure 11.1.1: One possible reconstruction of Rodinia 1.1 billion years ago. (John Goodge, modified from Dalziel (1997).)

Laurentia and Modern Plate Tectonics

Rodinia broke up about 750 to 600 million years ago. One of the fragments was **Laurentia**, the continental precursor to North America. The break-up may also signal the start of modern plate tectonics, with the rigid continental crust of Rodinia breaking along rift zones and subduction forming between oceanic and continental crustal slabs. It's breakup also created many shallow-water, biologically favorable environments that fostered the evolutionary breakthroughs marking the start of the next eon, the Phanerozoic.

Key Terms

- **Laurentia** - a Proterozoic craton that forms the core of the North American continent
- **paleomagnetic** - ancient magnetic orientation
- **Rodinia** - a supercontinent that formed and broke up in the late Proterozoic
- **supercontinent** - a continent composed of a collection of smaller continental masses

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11.2: Proterozoic Glaciations

The Snowball Earth

The “**Snowball Earth**” glaciations were a series of ice ages during the Neoproterozoic (late Proterozoic) Era of geologic time, roughly coinciding with the breakup of the supercontinent Rodinia. These ice ages were thought to have been so intense that the planet's entire surface may have frozen over - from the poles to the equator. The name of this extraordinary chapter in Earth's history comes from Caltech geophysicist Joe Kirschvink, who said in a 1992 scientific paper that Earth would have looked like a giant snowball when viewed from outer space.



Figure 11.2.1: Artistic interpretation of the Snowball Earth. By Oleg Kuznetsov - 3depix - http://3depix.com/, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons

These Snowball Earth episodes are the Sturtian glaciation (about 720 to 660 Ma) and the Marinoan glaciation (about 645 to 640 Ma) in the Cryogenian Period ("cryo" from the Greek for cold), with the Gaskiers glaciation (in the Ediacaran Period at about 580 Ma) being the possible third event in the sequence. There is 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 been less extensive.

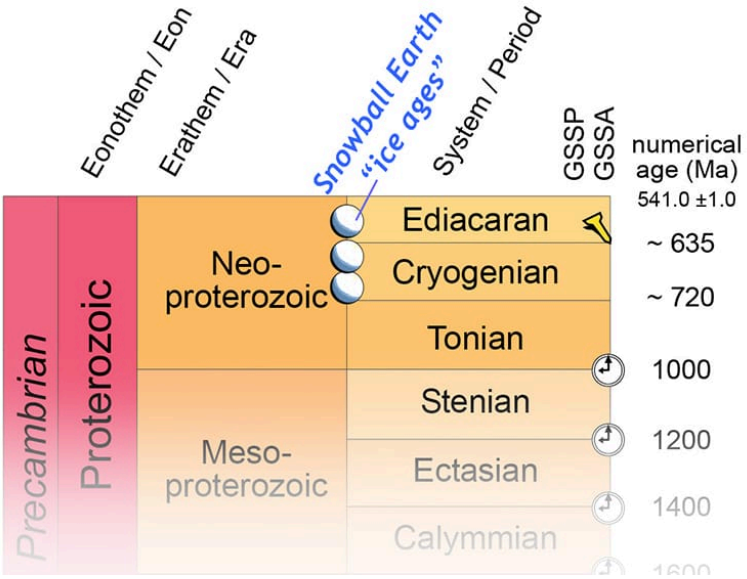


Figure 11.2.2: A close-up of the Neoproterozoic Era as depicted in the ICS timescale, showing the Cryogenian and Ediacaran periods.

Evidence for Neoproterozoic Glaciations

Why do geologists think there were ice ages during the Neoproterozoic? Ice ages are times in Earth's history when there are glaciers that are expanding, flowing, and moving sediment around. The ice eventually melts, but its impacts on the sedimentary record are distinctive and long-lasting. 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 glaciation, 39 localities were documented on six continents, and for the Marinoan, there are 48 places where glacial 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 subtle features are best seen on the ancient bedrock surfaces where the glacier flowed. However, this is rarely exposed, so this is the least common evidence for Snowball Earth.

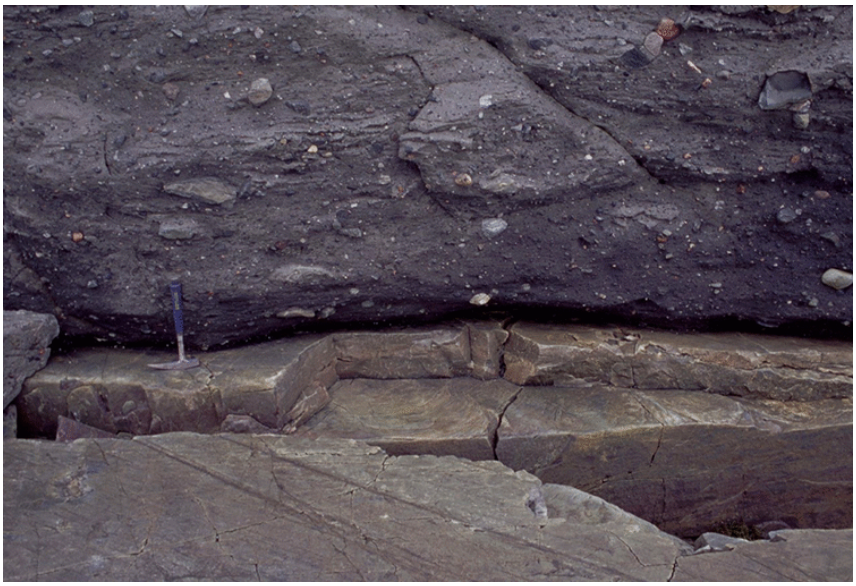


Figure 11.2.3: Striated glacial pavement (unconformity surface) below Neoproterozoic Smalfjord tillite, western Norway. (Photo by Galen Halverson; reproduced with permission.)

Tillites & Diamictites

Glaciers produce deposits of very poorly sorted sediment: giant 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. The photo below shows **tillite**, till that has been lithified into rock. Note the large faceted clasts “floating” in all orientations within the finer-grained matrix. Applying Uniformitarianism, we can conclude that the Neoproterozoic tillites are evidence of Snowball Earth glaciations, as they have the same texture as deposits produced by melting glaciers today.



Figure 11.2.4: Gaskiers tillite, Newfoundland. (Callan Bentley photo)

Another rock name frequently used to describe poorly sorted sedimentary deposits is diamictite. **Diamictites** are poorly sorted but aren't necessarily from glacial deposits. For instance, a landslide could produce a very poorly sorted mix of sedimentary grains, which could be lithified to make a diamictite. All tillites are diamictites, but not all diamictites are tillites.



Figure 11.2.5: A large (cobble-sized) faceted clast within diamictite of the Gaskiers Formation, western Avalon Peninsula, Newfoundland.

One way to distinguish whether a diamictite has a glacial origin (that is, whether it is a tillite) is based on the character of the larger clasts. Some of the boulders and cobbles in a true tillite will show faceted shapes, where they have flat faces ground down by being dragged along over resistant surfaces. These facets will frequently show randomly oriented scratches, too. Together,

these features provide compelling evidence of the past grinding action of glacial ice.



Figure 11.2.6: A faceted shape bearing intersecting scratches is strong evidence of a clast having been ground down through glacial action. (Callan Bentley photo)

Dropstones

The third signature of glaciation in the Neoproterozoic era is the presence of **dropstones** in marine sedimentary rocks. These fine-grained, thinly-bedded and laminated deepwater deposits of clay and silt, contain random clasts, the dropstones, that are significantly larger, such as pebbles, cobbles, or boulders. The dropstones pierced into the soft, pre-existing sedimentary layers, disrupting their continuity - the layers are truncated, with post-dropstone layers of fine sediment draped atop them.

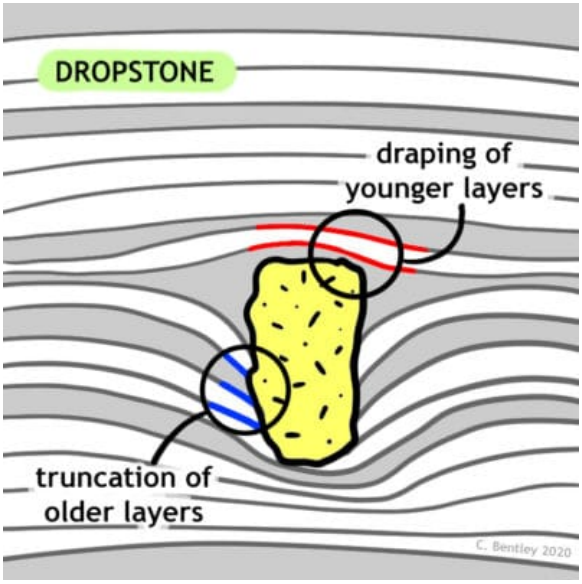


Figure 11.2.7: Cartoon cross-section of a dropstone, in cross-section, showing key relationships to the surrounding sedimentary layers. (Callan Bentley cartoon)

The only plausible candidate for transporting random, large clasts far offshore, where they could be dropped into thin sedimentary layers, is ice. It is likely that when glaciers from the Snowball Earth ice sheets reached the sea, they calved off (broke off) as icebergs full of gravel, sand, cobbles, and silt. We see this process today in Antarctica and Greenland. These icebergs float out to deep water, melt, and let their load of sediment fall through the water and pierce the sediment of the seafloor. The photos below show large clasts included with fine layers.

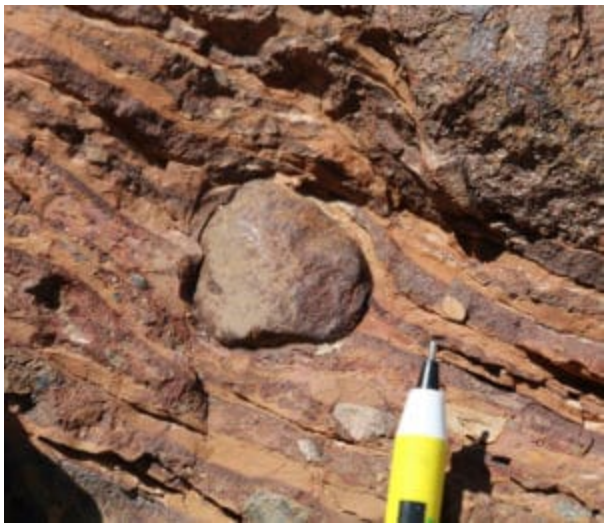


Figure 11.2.8: Dropstone in Kingston Peak diamictite, Sperry Wash, Death Valley region, California. Photo from Callan Bentley.

Evidence for Glaciers at the Equator

We have established that there is ample evidence of glacial activity preserved in the rock record from the Neoproterozoic era. However, it is also true that there is still plenty of glacial activity on modern Earth, and no one would mistake our modern planet for a giant snowball. So, why do we think the Snowball Earth glaciations were global freeze-overs, not glaciations confined to high-latitude continents like Greenland and Antarctica or to tall mountains, as is the case today? The answer comes from examining paleomagnetic signatures within the sedimentary rocks, which enables us to reconstruct the ancient latitude at which these rocks were formed.

During the mid-1980s to late 1990s, the paleomagnetic inclination (and thus paleo-latitude) of sea-level marine strata in central Australia was measured. These strata formed about 635 Ma and include signatures of glacial influence (such as dropstones) and show tidal rhythms preserved as varves (annual layer of terrestrial sediment). These strata are important because (1) we know they are glacial, but (2) they also show fine layers that allow for the precise measurement of their subtle magnetic signatures. Luckily, these samples have also been undisturbed since they formed: their magnetism has not been reset by a metamorphic event, for instance. The quality of the paleomagnetic measurements is robust, meaning they are trustworthy and have passed several tests validating the results. The results of the studies showed the sediments formed within 10° of the equator.



Figure 11.2.9: 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.)

Evidence of glaciers within 10 degrees of the equator is significant; the equator is generally warm year-round and not where we would expect continental ice sheets to flow 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 close to the equator during the Neoproterozoic. Therefore, if glaciers were flowing into the sea at such equatorial latitudes, it implies an exceptionally frigid climate for the planet.

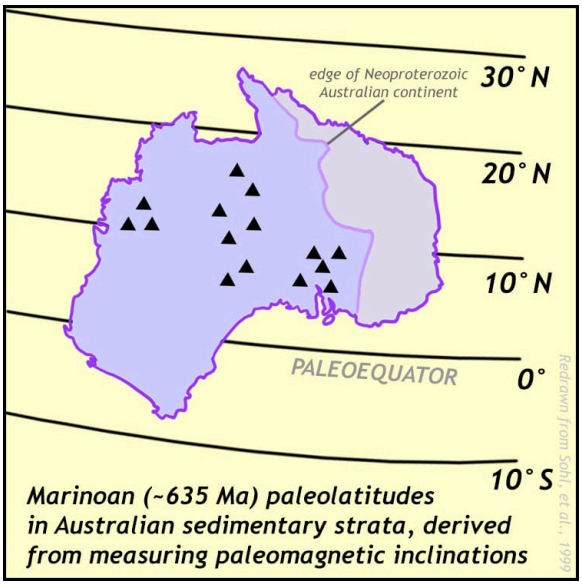


Figure 11.2.10: Marinoan-aged glaciogenic sedimentary rocks in central Australia show near-equatorial paleolatitudes. (Callan Bentley redrawing of a figure from Sohl, et al. (1999))

Furthermore, because snow and ice are highly reflective, a global ice cover extending all the way down to low latitudes would result in a highly reflective planetary surface. Overall, Earth would have had a very high **albedo**. Consequently, sunlight would have been reflected away instead of absorbed by the planet, further deepening the cold spell of this time.

Rapid Melting of Snowballs: Cap Carbonates

One intriguing feature of Snowball Earth glacial deposits is that many of these distinctive sedimentary rocks, with their characteristic signatures of glaciation, are conformably overlain by deposits of limestone and/or dolostone. Because these **carbonate** strata always overlie the glacial strata, they are called “cap carbonates.”



Figure 11.2.11: 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 carbonates tell us is that temperature conditions abruptly switched from the deep chill of an ice age to the balmy tropical warmth of a greenhouse period. Shallow water limestones form best today in places like the Bahamas or offshore of Florida. The cap carbonates deposited atop the glacial sediments and are much more extensive, though, covering significantly more area, suggesting they were deposited during a major transgressive event. Several lines of evidence point to

a rapid rate of cap carbonate deposition. The conformable 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 sequences of strata challenge our understanding of climate change on Earth.

Overall, the cap carbonates show a transgression (glacial ice melted and was added as water to the ocean, causing sea levels to rise) and a massive chemical precipitation of carbon-containing rock, indicating the ocean was in a state of carbonate oversaturation. The cap carbonate transgressive sequences for Sturtian and Marinoan glaciations are distinctive and consistent in deposits of that age worldwide.

Return of the Banded Iron Formation

Another chemical sedimentary rock in some Snowball Earth sedimentary sequences is the banded iron formation (BIF). This is cited as evidence of radically altered oceanic chemistry during the Snowball glaciation and, therefore, indirect evidence of extensive sea ice cover.



Figure 11.2.12: A dropstone (of quartzite) in banded iron formation strata of the Rapitan Group, Northwestern Canada. (Photo by Galen Halverson; reproduced with permission.)

In the Archean, the chemical precipitation of banded iron formation was a consequence of anoxic ocean waters. Generally, BIFs are considered an “extinct rock” that stop forming around 1.6 billion years ago with the build-up of free oxygen in the planet’s atmosphere and oceans. But, there is more to the story...

During the Snowball Earth glaciations, BIFs reappeared in the geological record because the conditions that allowed their formation, anoxic oceans, occurred again, but this time due to the Snowball Earth event. Their reappearance after a long absence is cited as strong evidence that sea ice covered the Earth’s oceans: 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. Consequently, deep-sea black smokers once again "polluted" ocean water with dissolved iron, and BIFs were deposited.

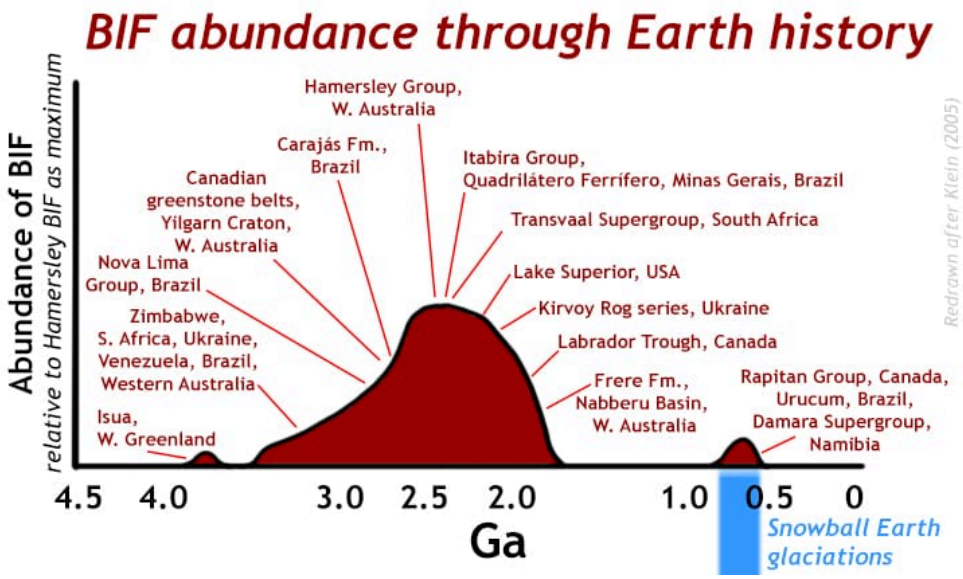


Figure 11.2.13: A plot showing the relative abundance of banded iron formation (BIF) through geologic time. Note the small "BIF blip" that coincides with the timing of the Neoproterozoic Snowball Earth glaciations. (Callan Bentley figure, redrawn from Klein (2005))

Starting and Stopping a Snowball

What caused the Snowball Earth events? The most accepted model for initiating the Snowball Earth glaciations assumes that CO_2 levels were dropping. This can happen when the amount of CO_2 consumed through chemical weathering exceeds the amount emitted by volcanic activity. Once these conditions were in place, ice would have begun to form at the poles, creating floating sea ice that expanded over time, gradually reaching lower latitudes. Sea ice has a high albedo: about 92% of incoming sunlight is reflected away, so little energy is absorbed from the incoming sunlight. In contrast, ocean water has a low albedo, reflecting only about 8% of incoming sunlight, with the rest being absorbed as heat energy.

As sea ice built up in the polar regions, its high albedo reflected away an increasing amount of the energy the planet received from the Sun. This constantly increasing albedo was a positive feedback system that enhanced the CO_2-induced cooling. Eventually, when the ice reached about 33° of latitude, the albedo-induced cooling effect was unstoppable, and the remainder of the planet's surface 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.

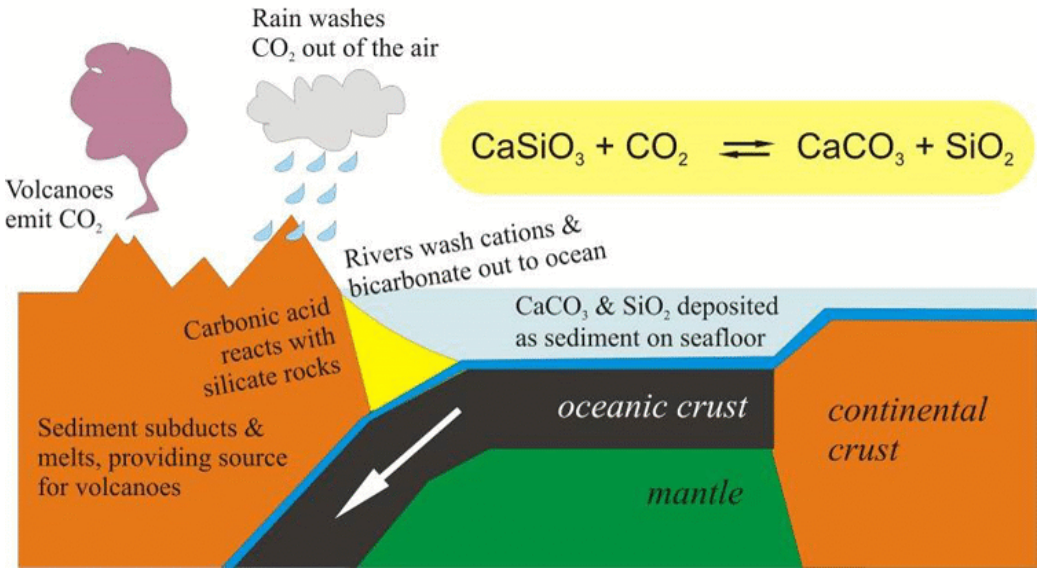


Figure 11.2.14: How the Neoproterozoic carbon cycle is thought to have changed to initiate a Snowball Earth glaciation (and sow the seeds for its eventual melting) <b.gif">original by Paul Hoffman (2006)).</b.gif">

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. Above sea level, much of the carbon cycle would no longer be functional, with essentially no chemical weathering happening. So, how was it possible that the planet could warm up again?

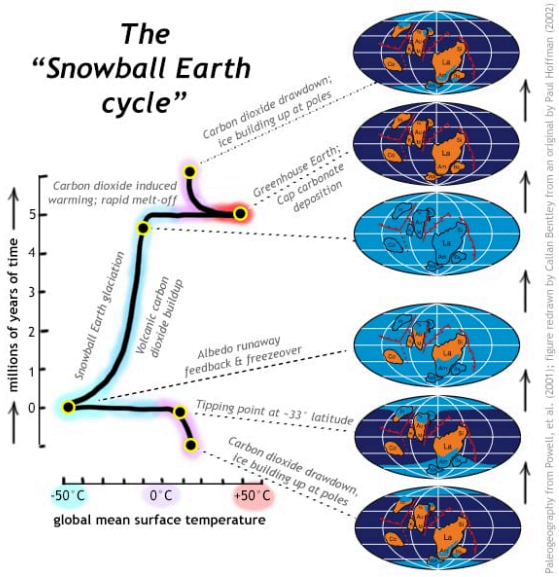


Figure 11.2.15: The theoretical cycle of events that turned on and off the two "hard" Snowball Earth glaciations. (Callan Bentley figure, redrawn from an original by Paul Hoffman (2002))

You may recall that plate tectonics is not driven by energy from the Sun; instead, it is powered by Earth's internal heat. So, despite the frigid conditions, subduction and seafloor spreading continued, driving volcanic activity, emitting gases including water vapor and carbon dioxide. 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 concentration of this greenhouse gas would have rapidly increased. Eventually, after about 5 million years or so, enough CO_2 accumulated that the resulting greenhouse heat retention would have overcome the albedo's cooling effects. Sea ice would have started melting, likely first near 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.

Eventually, Snowball Earth came to an end, and the post-glacial greenhouse Earth had begun. With an atmosphere 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 rapidly accumulated, and, gradually, CO_2 was drawn down again.

Key Terms

- **albedo** - the ability of a surface to reflect sunlight
- **carbonates** - sedimentary rocks primarily composed of carbonate minerals like limestone (calcite or aragonite) and dolomite -
- **diamictite** - generic term for poorly sorted sediment; it could be from glacial activity but could also be from landslide or other sedimentary activities
- **dropstones** - random large clasts found within fine-grained, thinly-bedded and laminated deepwater deposits of clay and silt that were transported out into the ocean by an iceberg that then melted and dropped it
- **paleomagnetism** - the study of the Earth's ancient magnetic field, preserved in rocks and sediments
- **Snowball Earth** - periods of time when glaciers were so extensive they covered most of Earth
- **striations** - scratches or grooves that result when a glacier takes a larger particle of rock and drags it against the bedrock, gouging out a scratch mark that shows the direction the glacier was flowing
- **tillite** - glacial deposits formed from very poorly sorted sediment; glacial diamictite

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11.3: Proterozoic Atmosphere

The Great Oxidation "Event"

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.

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 11.3.1: Deformed banded iron formation, Soudan, Minnesota. Chemical sedimentary rocks like these imply an oxygen-free atmosphere. (Callan Bentley photo)

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, because they were getting oxidized by atmospheric oxygen, 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 11.3.2: Detrital pyrite in conglomerate of the Steyn Reef Placer, Witwatersrand Supergroup of South Africa. (Photo by Frances Deegan; reproduced with permission.)

Origin of Oxygen

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).



Figure 11.3.3: Modern stromatolites, Shark Bay, Australia. (Photo by Alicejmichel via Wikimedia)

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 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.

Definition: Term

- **banded iron formation** - unusual chemical sedimentary rocks that could only form in an ocean that is largely oxygen-free
- **cyanobacteria** - single or multicellular bacteria that use chlorophyll to absorb energy from light and were most responsible for constructing stromatolites
- **detrital pyrite** - rounded, transported pyrite found in sedimentary rocks, serving as an indicator of an oxygen-free Earth atmosphere
- **stramatolite** - layered, sedimentary rock structures formed by photosynthetic cyanobacteria trapping sediment in shallow water

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11.4: Proterozoic Life

Eukaryotes: Evolution through Endosymbiosis

A large evolutionary step occurred during the Proterozoic Eon, with the appearance of **eukaryotes** around 1.8 billion years ago in the Paleoproterozoic. Eukaryotic cells are more complex, having nuclei and organelles. The nuclear DNA is capable of more complex replication and regulation than that of **prokaryotes**. The organelles include mitochondria for producing energy and chloroplasts for **photosynthesis** which is the processes where plants take in carbon dioxide to convert sunlight into chemical energy which fuels their metabolism and releases oxygen to their surroundings. The eukaryote branch in the tree of life gave rise to fungi, plants, and animals.

The accumulation of oxygen in the atmosphere due to prokaryotic photosynthesis created a new resource that life could exploit as a source of energy. The new complex eukaryotic cells were equipped 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.

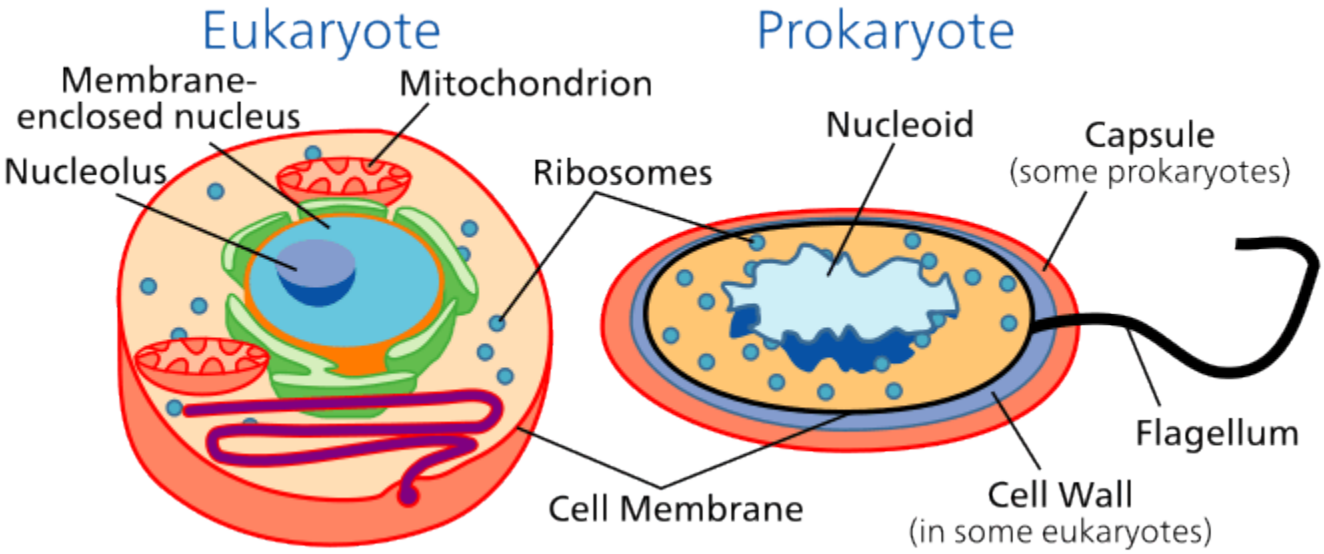


Figure 11.4.1: Prokaryotes versus eukaryotes. (Public Domain image; Science Primer (National Center for Biotechnology Information))

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).

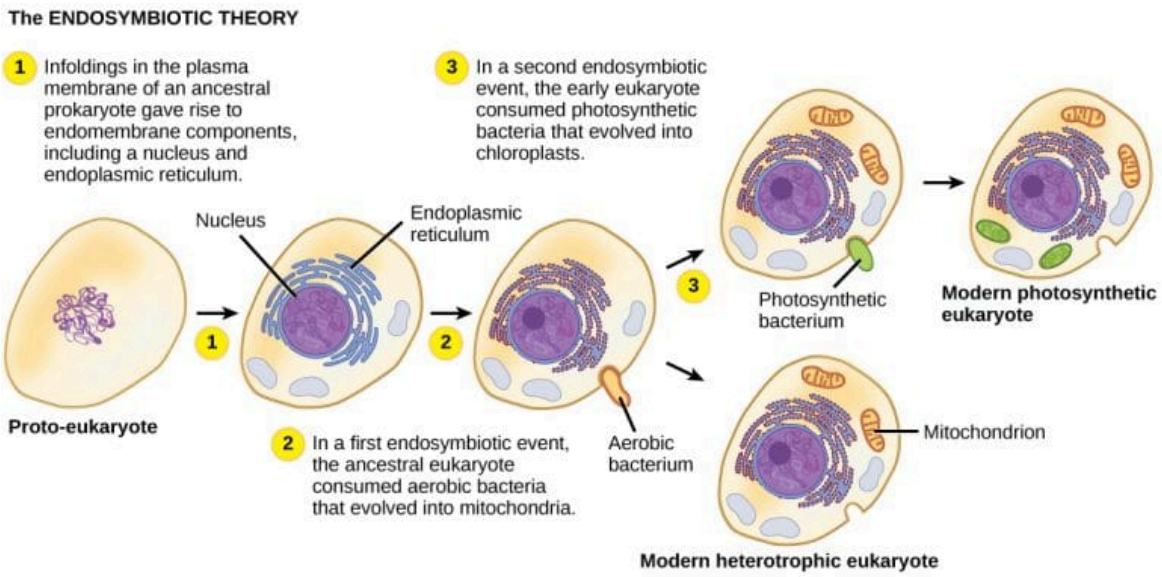


Figure 11.4.2: Schematic of endosymbiotic theory showing the evolution of eukaryotes from prokaryotes. (Image via Creative Commons Attribution-Share Alike 4.0 International license)

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.

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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.

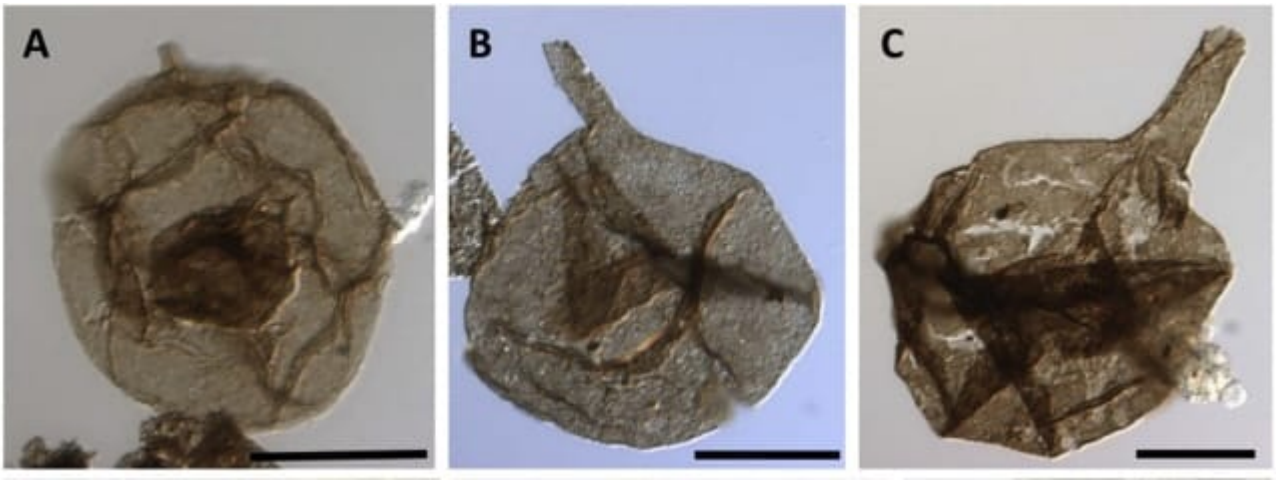


Figure 11.4.3: Selected Neoproterozoic acritarchs (*Germinosphaera cf. G. bispinosa)* from the Mingolsheim core (Germany). All scale bars 20 µm. (B. van de Schootbrugge via Wikimedia, CC BY 4.0)

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.

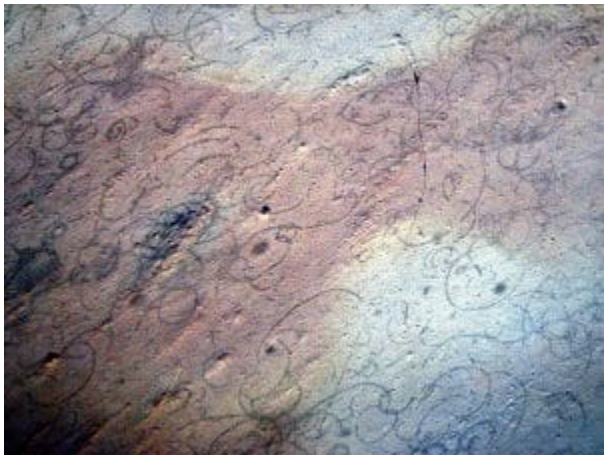


Figure 11.4.4: Grypania spiralis fossils. (Creative Commons Attribution-Share Alike 3.0 Unported license.)

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.

Multicellular Life: The Ediacaran Biota

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** which are bacteria that use chlorophyll to absorb energy from light. 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.

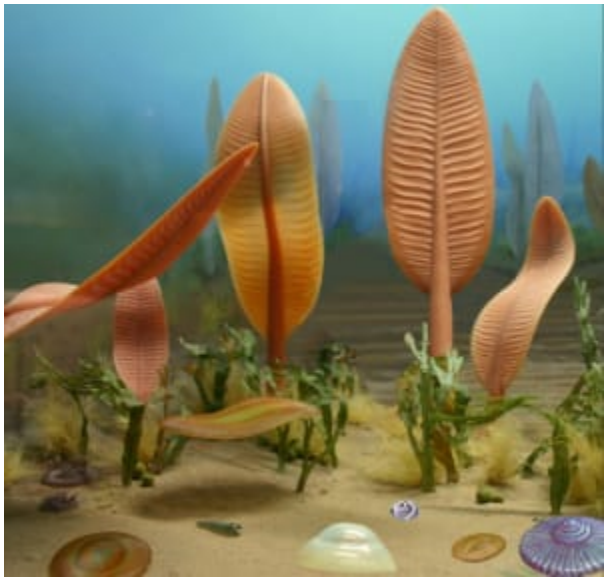


Figure 11.4.5: Artistic rendering of Ediacaran shallow seas. (Creative Commons Attribution-Share Alike 4.0 International license.)

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.

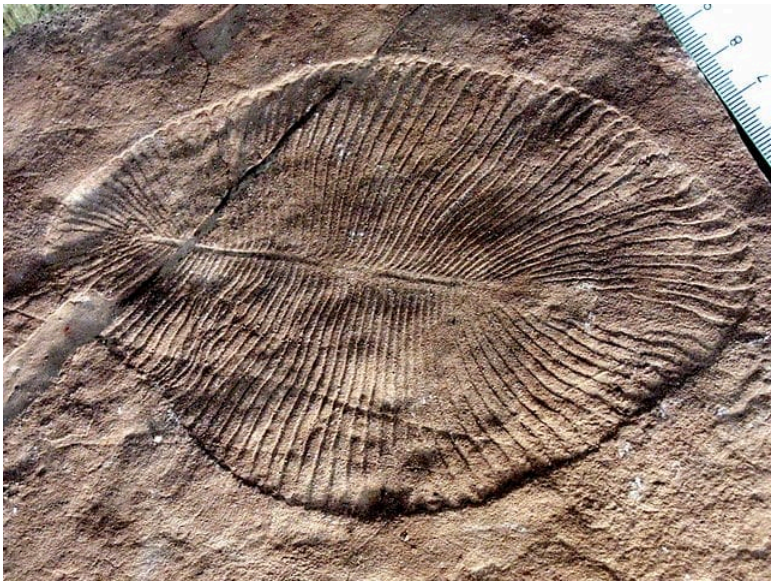


Figure 11.4.6: Dickinsonia tenuis fossil. (Creative Commons Attribution-Share Alike 4.0 International license.)

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.

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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.

Key Terms

- **aerobic** - involving requiring oxygen
- **Ediacaran biota** - plants and animals that evolved in the late Proterozoic and represent the earliest known complex multicellular organisms
- **endosymbiosis** - the process of two organisms merging into one to form a mutually beneficial relationship
- **eukaryotes** - organisms with membrane-bound nuclei and include all animals, plants, fungi, seaweeds, and many unicellular organisms
- **photosynthesis** - a system of biological processes where plants take in CO_2 to convert sunlight into chemical energy which fuels their metabolism and releases oxygen to their surroundings
- **prokaryotes** - single-celled organisms that lack nuclei and predate eukaryotes

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11.5: Key Points

- Widespread Neoproterozoic glaciation is well-documented, with Cryogenian and Ediacaran glacial deposits found even at paleoequatorial latitudes.
- Glacial sediments are overlain by “cap carbonates,” warm-water limestones interpreted as forming during rapid post-glacial sea-level rise.
- The Snowball Earth hypothesis explains these deposits, proposing that Earth was nearly or entirely ice-covered during major Cryogenian glaciations.
- Glaciation was likely initiated by CO_2 drawdown from intense tropical weathering, which reduced greenhouse warming.
- A runaway ice-albedo feedback amplified cooling, allowing ice to expand toward the equator and further reduce absorbed solar radiation.
- Deglaciation occurred when volcanic outgassing rebuilt atmospheric CO_2, strengthening greenhouse warming enough to melt global ice cover.
- Melting released large amounts of dissolved ions, enabling widespread precipitation of thick carbonate sequences during marine transgression.
- Sturtian and Marinoan glaciations show associated banded iron formations, indicating ocean anoxia consistent with extensive sea-ice cover.
- The Gaskiers glaciation was shorter and less extensive, and its end closely precedes the appearance of the earliest Ediacaran fossil animals.

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11.6: Further Reading

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12: PHANEROZOIC EON - PALEOZOIC ERA

Introduction

The Phanerozoic Eon is the most recent, from 541 million years ago to today; it means “visible life” because an abundance of fossils marks the Phanerozoic rock record. Phanerozoic organisms had evolved hard body parts like claws, scales, shells, and bones that made it more likely they would be preserved as fossils. Rocks older than the first period of this eon, the Cambrian, are called Precambrian and rarely contain fossils because these organisms had soft body parts. Rocks of the Phanerozoic are younger and more common, and contain the majority of visible fossils; the abundance of preserved fossils in rocks from this Eon makes them the most informative storytellers of Earth’s history. The Phanerozoic is subdivided into three Eras, the Paleozoic (“ancient life”), Mesozoic (“middle life”), and Cenozoic (“recent life”). This chapter focuses on the Paleozoic Era.

Here is the the simplified geologic time scale used in the Geologic Time chapter of this book. The Paleozoic Era is highlighted toward the bottom.

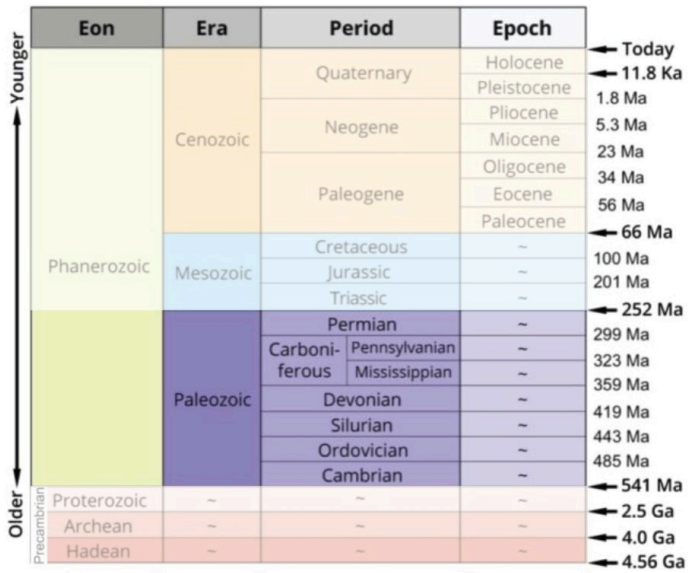


Figure 12.1: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

Marine organisms dominated life in the early Paleozoic Era, 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.



Figure 12.2: The trilobites had a hard exoskeleton, and were an early arthropod, the same group that includes modern insects, crustaceans, and arachnids.

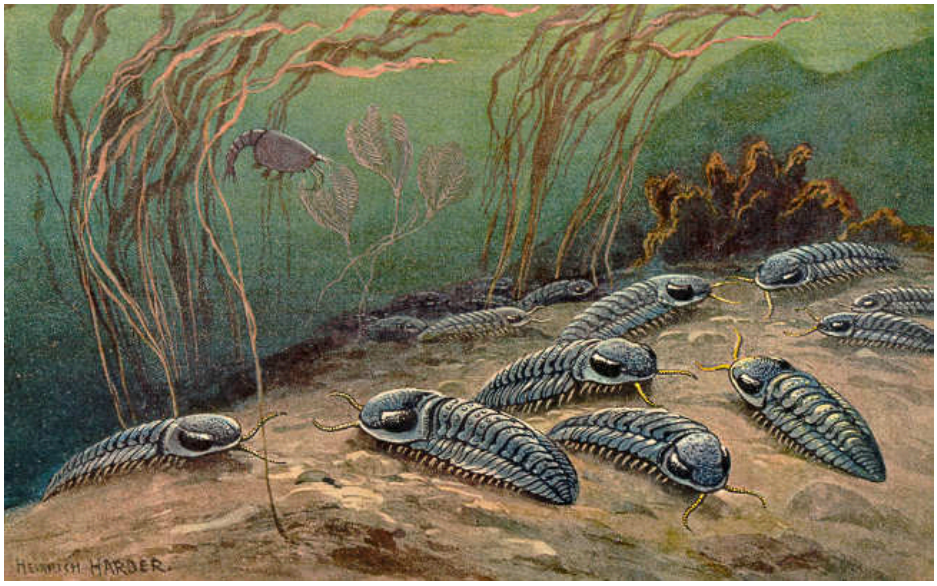


Figure 12.3: Trilobites, by Heinrich Harder, 1916.

During the Paleozoic Era, sea levels rose and fell during four transgressive-regressive sequences. Most of North America was a shallow, tropical ocean, as evidenced by the abundant marine sedimentary rocks such as limestones containing corals, ooids, and other fossils covering much of the continent. Unconformities document falling sea levels when rock sequences were lost to erosion.



Figure 12.4: Laurentia, which makes up the North American craton.

By the late Paleozoic, the supercontinent Pangea was close to fully assembled. This resulted from the collisions and accretions of island arcs and continents, as well as the closures of ocean basins. In North America, these tectonic events happened along 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. The Panthalassa ocean basin surrounded Pangea. Subsequent plate movement formed the huge Tethys Sea bay, dividing Pangea into smaller supercontinents, Laurasia and Gondwana. Laurasia comprised the North American craton, Laurentia, and Eurasia. Gondwana comprised the remaining continents of South America, Africa, India, Australia, and Antarctica. The east coast of North America had a very high mountain range from continental crust collisions and the creation of Pangea. The west coast of North America had smaller and isolated volcanic highlands from subduction and the associated 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 12.5: A reconstruction of Pangaea, showing approximate positions of modern continents.

The video below, "From the Cambrian Explosion to the Great Dying", offers a concise and thorough summary of the Era.

YouTube Video

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12.1: Geography and Tectonics of the Early Paleozoic

12.2: Evidence for the Taconian Orogeny

12.3: Geography and Tectonics of the Middle Paleozoic

12.4: Geography and Tectonics of the Late Paleozoic

12.5: Early Paleozoic Life - The Cambrian and Ordovician

12.6: Middle Paleozoic Life - The Silurian and Devonian

12.7: Late Paleozoic Life - The Carboniferous and Permian

12.8: Key Points

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12.1: Geography and Tectonics of the Early Paleozoic

Overview

At the start of the Paleozoic, in the Cambrian Period, the continents were near or below the equator, making their configuration essentially opposite of the world today. Laurentia (proto North America) had recently separated from the supercontinent Gondwanaland, and was situated along the equator and mostly above sea level. As the Paleozoic progressed, the global climate warmed, melting glaciers and sea levels rose. The lithology of sedimentary rocks deposited during this time tells the story of sea level transgressions during the early part of this Era. Near the end of the Ordovician, sea levels regressed as Gondwanaland moved across the South Pole and an ice sheet grew across the cratons. This relatively short glaciation period caused one of the greatest extinctions of marine organisms (Stanley and Luczaj, 2015).

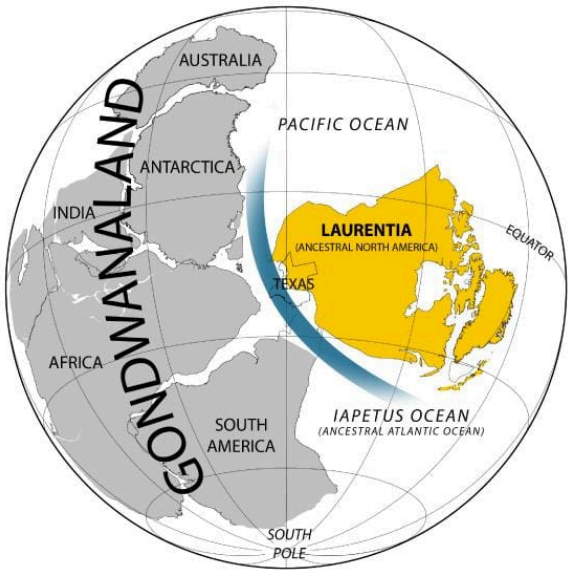


Figure 12.1.1: Laurentia (ancestral North America) during the Cambrian with the state of Texas for reference. Credit: Ian Dalziel, University of Texas at Austin.

North American Continent

The eastern margin of the ancestral North American continent was active, with subduction driving mountain building. The first of three mountain-building events that constructed the Appalachian Mountains, the Taconian Orogeny, happened as Laurentia collided with several islands situated between the continent and Gondwanaland.

In contrast to the East Coast, the West Coast was a tectonically inactive passive margin during the early Paleozoic. The western edge of the North American continent was near the present-day Nevada-Utah border and formed an expansive shallow continental shelf near the paleoequator. Starting in the Devonian, the west coast transformed into an active continental margin as an island arc collision drove the Antler Orogeny. This marked the start of subduction and accretion of continental material onto the western margin of North America, which would eventually become portions of the states of California, Oregon, and Washington. The Sonoma orogeny began in the late Paleozoic when subduction caused North America to collide with another island arc. Eventually, a subduction zone was established along the entire western margin that persisted through the Mesozoic Era.

The video below summarizes the tectonic and environmental changes taking place in North America during the Paleozoic. On the left side, sea level transgressions and regressions are indicated by blue and brown, respectively. The transgressive sequences are named after the locations where their deposits were originally studied: **Sauk, Tippecanoe, Kaskaskia, and Absaroka**. To the right is a geologic timeline and the geographic evolution of North America.

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The Wilson Cycle

Canadian geophysicist J. Tuzo Wilson developed a model that explains how an ocean basin opens and closes, beginning with a divergent boundary and ending with collisional subduction between two continents. This model was developed using observations of fossil similarities on either side of the Atlantic and structural evidence that the Atlantic Ocean opened roughly where an ancient ocean had closed. This led Wilson to propose that a “proto-Atlantic Ocean” existed between the continents during the Paleozoic Era. This ocean closed in stages, bringing together dissimilar rocks as the supercontinent Pangaea formed.

The Sauk, Tippecanoe, Kaskaskia, and Absaroka sequences were deposited during one very long period of episodic deposition, marking the opening and closing of the Iapetus Ocean as illustrated by the Wilson Cycle. This simple model can also be applied to other contexts, such as the closing of the Tethys Sea as India moved northward toward Asia. Watch this simple schematic video of the Wilson Cycle below.

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Stages of the Wilson Cycle

A Wilson Cycle has nine stages, 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.

The closing of the rift 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. The Sauk, Tippecanoe, Kaskaskia, and Absaroka sequences fit in the model from stage C to before stage H.

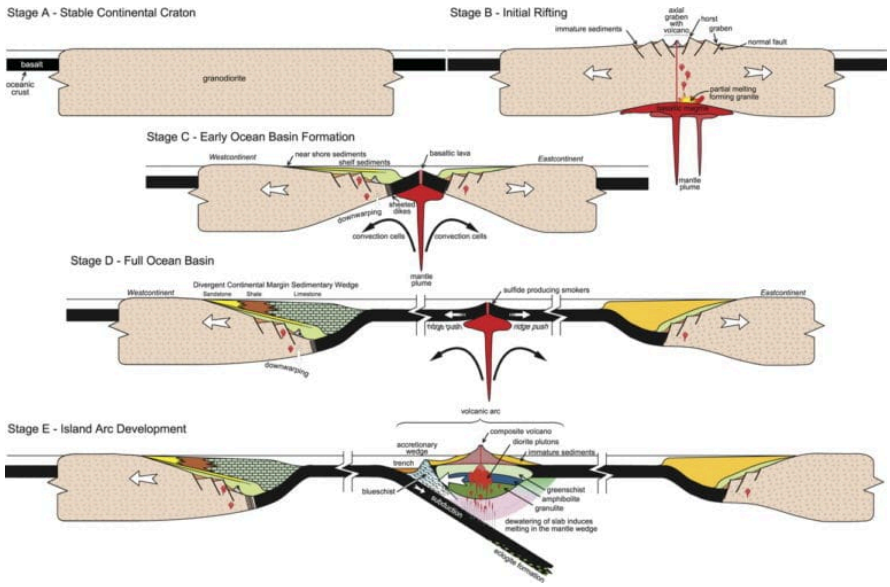


Figure 12.1.2: Stages A through E of a Wilson Cycle (Whitmeyer et al., 2007)

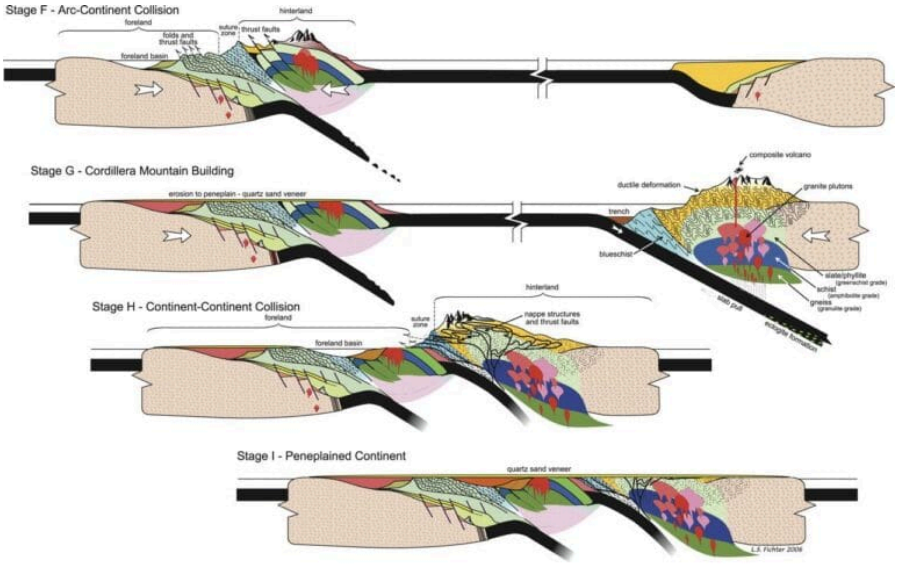


Figure 12.1.3: Stages F through I of a Wilson Cycle (Whitmeyer et al., 2007).

The Sauk and Tippecanoe Transgressive Sequences

The Sauk, Tippecanoe, Kaskaskia, and Absaroka marine transgressions deposited sediment across most of North America from the Cambrian through the Permian Periods. In the late Proterozoic, sea levels rose as the supercontinent Rodinia broke up. Seawater flowed across Laurentia, forming the Sauk Sea, and sediment was deposited atop the long-eroded surface of the craton. This resulted in the Great Unconformity, representing a gap of geologic time ranging from 500 million to 1.2 billion years. Carbonates, shale, sandstone, and volcanic ash were deposited with each pulse of sea level rise, recording rising or falling sea levels, volcanism, and mountain building.

Here, we'll examine the formations produced during the Sauk and Tippecanoe transgressions in greater detail to see what story their rocks can tell.

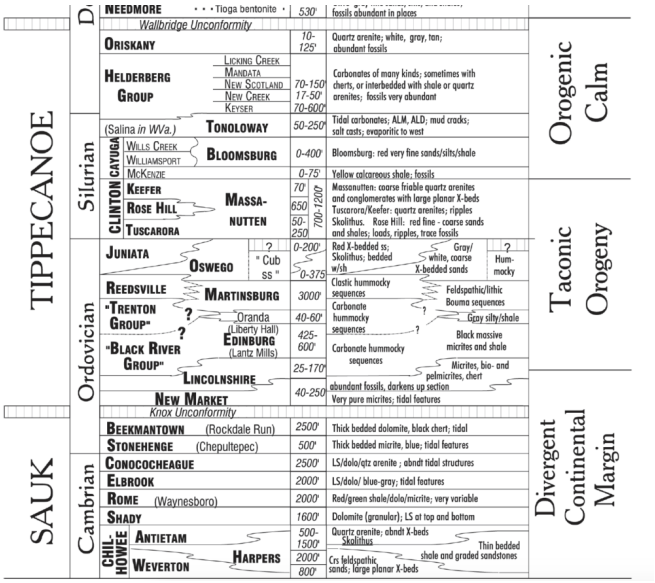


Figure 12.1.4: A stratigraphic column summarizing the lithology and related tectonic activity of the Sauk and Tippecanoe transgressions.

At the start of the Paleozoic, the Laurentian craton was mostly above sea level as reflected by thick accumulations of siliclastic sediment (terrestrial) in early Cambrian formations, like the Chil-Howee. Cross-bedding is common in these rocks, attesting that wind played an important role in transporting and depositing sediment: a consequence of a continent devoid of vegetation. The early Ordovician New Market Limestone, as indicated by its lithology, fossils, and structure, preserves a record of a shallow sea. Examine the carbonate rocks of the New Market and Lincolnshire formations in the GIGAmacro image below.

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Figure 12.1.5: Carbonate rocks of the New Market and Lincolnshire formations.

Later in the Ordovician, deposits from the Tippecanoe transgression reached the seafloor as a type of landslide called **turbidity currents**. Evidence of this is the mix of terrestrial sediment, marine carbonates, and finely preserved fossils. The GIGAmacro image below shows well-preserved graptolite fossils (darker, caterpillar-like markings) in a muddy limestone. Such preservation can result from rapid burial, as would be the case with submarine landslides.

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Figure 12.1.6: Graptolite fossils preserved in a muddy limestone.

The Tippacanoe Sequence also reflects mountain building from the Taconian Orogeny. The Silurian Massanutten Sandstone formed from the siliclastic sediment eroded from the highlands produced by the Taconian Orogeny. This formation contains sedimentary structures like ripple marks and cross-bedding, reflecting deposition in shoreline and dune environments. See the hand sample below.

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Figure 12.1.7: Massanutten Sandstone produced by Taconian Orogeny.

For reference, Laurentia may have looked like this during the Silurian.



Figure 12.1.8: Late Silurian (Taconian Orogeny) Laurentia.

With time, the Taconian mountains were eroded down to a lowland. This and renewed transgression are reflected by the transition from the coarse sands of the Massanutten to the fine-grained intertidal Bloomsburg and Tonoloway Formations, then the fossil-rich limestone of the Helderberg Formation. The latter formed as a carbonate bank of limestone and dolomite in a warm, shallow sea during the early Devonian.

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Figure 12.1.9: Bloomsburg Formation, Fort Valley, VA.

The Tippacanoe Sequence ends with a significant sea level regression and prolonged erosion. This is marked by the Wallbridge Unconformity, upon which the Kaskaskia, and Absaroka marine transgressions deposited sediment, further adding strata atop the Laurentian craton.

The interpretation of stratigraphy and unconformities, as exemplified by examination of the Sauk and Tippecanoe sequences, can provide a model for interpreting the changes we see in **facies**. These changes, resulting from the complex interplay among tectonics, sedimentation, and relative sea level, constitute the story of the strata. Consequently, we know the geologic history of a basin, like the Iapetus Sea, and the paleography of a continent, like Laurentia. Through interpreting stratigraphy, rocks can tell their story!

Key Terms

- **Absaroka Sequence** - A group of North American sedimentary formations formed during marine transgressions in the Mississippian, Permian, Triassic, and Jurassic periods
- **facies** - Characteristics of a rock unit that reflect the origin of the rock and distinguish it from adjacent rock units. These characteristics include mineralogy, sedimentary structures, fossil content, and texture. These provide clues about the depositional environment in which a rock was formed
- **Kaskaskia Sequence** - A group of North American sedimentary formations formed during marine transgressions from the Devonian through the Pennsylvanian periods
- **Sauk Sequence** - A group of North American sedimentary formations formed during marine transgressions in the late Proterozoic and Ordovician periods
- **Tippecanoe Sequence** - A group of North American sedimentary formations formed during marine transgressions in the Ordovician, Silurian, and Devonian periods

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12.2: Evidence for the Taconian Orogeny

Introduction

During the late Ordovician, subduction pulled a volcanic island arc into Laurentia (ancestral North America), causing mountain-building in the region that is today New England. The Taconian, or **Taconic Orogeny**, produced an extensive mountain belt, with peaks rivaling those found in the Swiss Alps and vast volumes of sediments from their erosion. Geoscientists today use both the resulting sedimentary rocks and the metamorphic and igneous rocks that formed at the roots of the mountain belt as evidence for this important event.

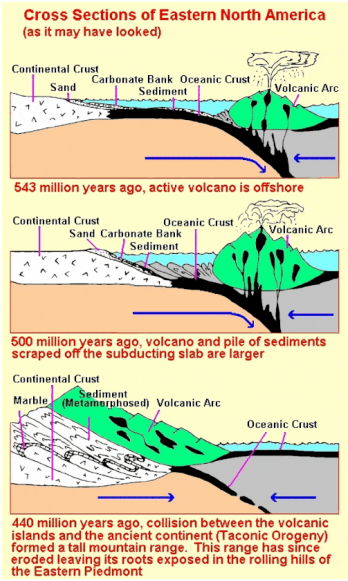


Figure 12.2.1: The sequence of events that led to the Taconic (Taconian) Orogeny. Image from Lyn Topinka, United States Geological Survey. Modified from Plank and Schenck, 1998, Delaware Piedmont Geology, Delaware Geological Society., Public domain, via Wikimedia Commons.L

Before the orogeny, the edge of Laurentia was a passive margin: through the Cambrian and well into the Ordovician, there was no tectonic activity anywhere nearby. Submerged under a shallow 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, and abundant fossils of filter feeders like bryozoans and brachiopods, and low quantities of clastic sediments, indicate clear water and no highlands.



Figure 12.2.2: Ooids from the pre-Taconian Cambrian-aged Conococheague Formation, Shenandoah County, Virginia.

Evidence from Igneous and Metamorphic Rocks

The cause of the Taconian Orogeny was a collision between two tectonic plates: the oceanic segment of the ancestral North American plate and the plate including the floor of the Iapetus Ocean. Subduction resulted in a volcanic island arc forming in the middle of the Iapetus Ocean.

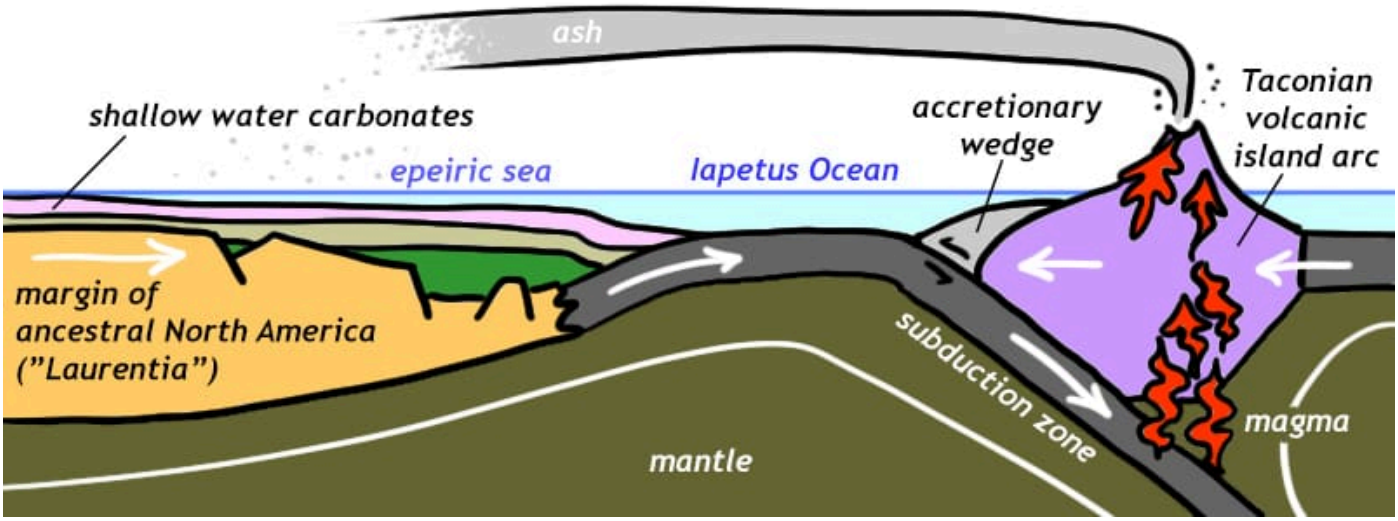


Figure 12.2.3: The tectonic situation that would lead to the Taconian Orogeny: subduction of the oceanic margin of the ancestral North American plate resulted in a volcanic island arc that drew ever closer, building up an accretionary wedge of Iapetan ocean floor and deepwater sediments.

Isotopic ages of rocks from this time tell us two things: that subduction was happening and, later, the volcanic arc accreted onto Laurentia. The Port Deposit Tonalite, a metamorphosed granitoid, is a nice example. Prior to the Taconian Orogeny, it was an unmetamorphosed body of intrusive igneous rock, the product of subduction-generated magma. Zircons from the tonalite reveal a magmatic crystallization age of 515 Ma. At the same time, biotites indicate the igneous rock was metamorphosed 490-480 Ma when the Taconian volcanic island arc collided with the ancient continent.



Figure 12.2.4: The Port Deposit Tonalite of Cecil County, Maryland.

Additional evidence for the accretion of the Taconic Island Arc is contained in the remains of the ancient mountains of this region, the Piedmont geologic province of Virginia, Maryland, Washington, D.C., Pennsylvania, New Jersey, and New York, and parts of New England. The "story-tellers" are the crushed and cooked remains of the Iapetus Ocean and the Taconian volcanic island arc. These rocks have been metamorphosed to various degrees, from greenschist facies 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 (intrusive and extrusive, mafic and felsic). In some cases, the subsequent metamorphic recrystallization was slight enough to preserve primary volcanic and sedimentary structures, such as the graded beds in the meta-graywacke of the Mather Gorge Formation, which reflect oceanic processes. Specifically, these graded beds formed from deep submarine deposition of clastic sediment by turbidity currents in the Iapetus Ocean.



Figure 12.2.5: Relict graded bed in Mather Gorge Formation metagraywacke, near Potomac, Maryland.

These rocks also help tell a broader story of the timing of the Taconian Orogeny. Radiometric dating shows metamorphism happening around 460 Ma, as reflected by the crystallization ages for **migmatites**. The Piedmont is also comprised of felsic igneous rocks that crystallized out of the subduction-made magma. These include formations like the Occoquan Granite, the Georgetown Intrusive Suite, and the Kensington Tonalite, all of which have isotopic ages in the 474 to 450 Ma range, which are the product of later-stage subduction. The gigapixel image below is of a sample of migmatite from Orange County, Virginia. It contains “leucosomes” that represent the formerly molten portion of this rock, which is otherwise a schist. Similar partial melting occurs today beneath active modern mountain belts such as the Himalayas.

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Figure 12.2.6: Migmatite from Orange County, Virginia representing partial melting that occurs beneath active mountain belts.

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 smashed into Laurentia, compressing and metamorphosing the Iapetan ocean sediments caught in between. Today, traces of ancient Taconian Orogeny faults are found throughout New England and New York City.



Figure 12.2.7: 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.

Evidence from Sedimentary Rocks

As mountains grow, they produce large amounts of sediment through erosion. This debris accumulates in the adjacent basins, sometimes terrestrial and sometimes marine. Before plate tectonics was understood to be the driving force of mountain building, geologists used accumulations of clastic sediment as evidence of orogenies because gravel, sand, and mud must come from somewhere. Sequences of clastic sediment can be organized into two groups: deep marine turbidity current deposits called “flysch,” and a terrestrial “red-bed” sequence, termed “molasse.” These European terms are useful for summarizing the sedimentary signature of the Taconian Orogeny.

Taconian Flysch

As the Ordovician approaches, the stratigraphic record shows limestones get “dirtier” as clay and silt deposits mix with the carbonate deposits. The first indication of tectonic uplift. Over time, going up and younger in the stratigraphic sequence, the passive margin carbonates give way to shales with limestone, 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 as it grew taller. This flysch is the marine sedimentary record of Taconian mountain-building and its erosion.

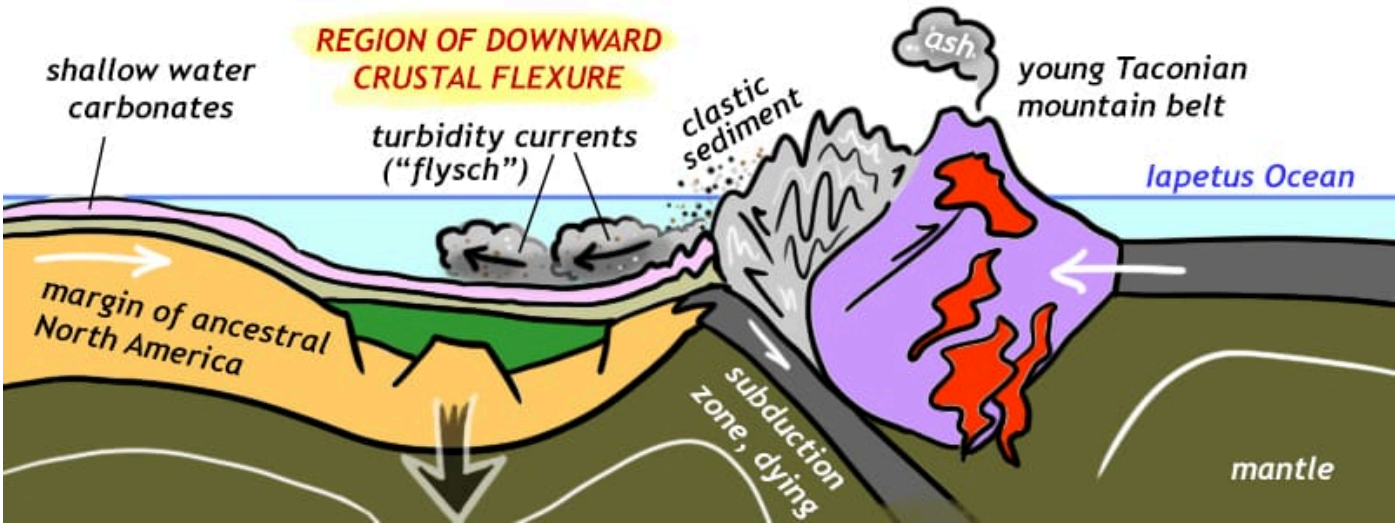


Figure 12.2.8: The deepening of the sedimentary basin adjacent to the young Taconian mountain belt was accomplished as the edge of ancestral North American continent flexed downward. Turbidity currents flowed into this deepened basin, depositing shale and graywacke: the Taconian "flysch."

A record of these turbidity currents is preserved as graded beds in graywacke, separated by shale layers. These distinctive deep-sea sedimentary sequences resulted from repeated submarine avalanches that delivered vast quantities of sand and mud into the deep ocean.

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Figure 12.2.9: Evidence of turbidity currents in the Devonian Trimmers Rock in Newport, Pennsylvania.

The transition from pre-Taconian shallow-water carbonates to Taconian Orogeny deepwater turbidites suggests that the water got deeper. There are likely two causes of this: (1) compression along the convergent boundary, which caused the leading edge of the ancestral North America plate to fold downward, and (2) the increasing weight of the accretionary wedge of sediment caused the oceanic crust of the passive margin to sag downward under this extra weight. Fossils further confirm this scenario, as deeper water faunas like brachiopods, crinoids, and a nautiloid are preserved in some of these deposits. Layers of ash are preserved, too, presumably sourced from the approaching volcanic island arc. These ash layers weather today to a yellowish, crumbly clay material called bentonite, including zircons that can be dated, which helps constrain the age of the sedimentary strata above and below the bentonites.

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Figure 12.2.10: Gigapixel image of a rock containing brachiopods and crinoids.

Taconian Molasse

Eventually, the basin collecting the flysch deposits filled. Rivers flowing out of the Taconian mountains deposited their load of sediment in alluvial fans, stream channels, flood plains, and deltas. Using the volume of flysch and molasse deposits, one can estimate the total volume of the mountains to be 600,000 cubic km of rock. Considering the width of the metamorphic belt that was previously discussed and the metamorphic grade of the rock within the belt (which is a reflection of the overlying pressure acting on those rocks), we can use the volume estimate to estimate the height of the Taconian peaks to be around 4000 meters or 13,000 feet.

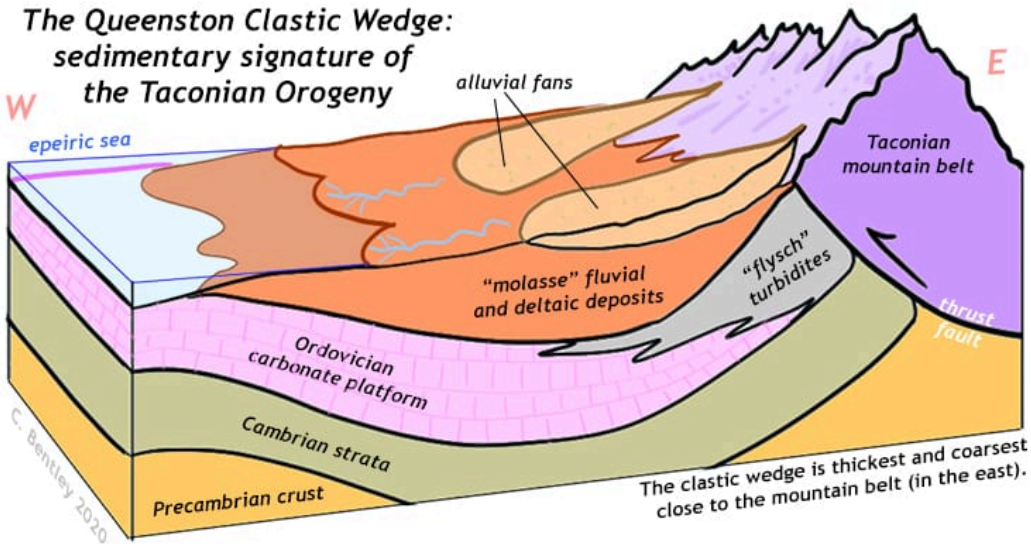


Figure 12.2.11: The Queenston Clastic Wedge was deposited west of the Taconian mountain belt. The molasse is thickest and coarsest close to the mountain belt to the east, and thins and fines to the west.

After the Taconian Mountains had been worn down, conditions returned to passive margin sedimentation in the Silurian, and new carbonates were deposited into the Devonian. This reprieve from active margin conditions would resume with the Acadian Orogeny in the middle to late Devonian.

Key Terms

- **migmatites** - a rock having both metamorphic parts and igneous parts resulting from partial melting
- **protolith** - the original, unmetamorphosed rock
- **Taconic Orogeny** - an extensive mountain building event 440-470 million years ago caused when subduction pulled a volcanic island arc into Laurentia

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12.3: Geography and Tectonics of the Middle Paleozoic

Overview

The Silurian and Devonian periods comprise the middle Paleozoic. Sea levels rose during this time as the Ordovician glaciers melted, and sediment was deposited across most continents which helped record the history we are about to discuss.

Mountain Building

Eastern North America

In the late Silurian through the Devonian, Laurentia experienced two mountain-building events, the **Caledonian** and **Acadian orogenies**. The Caledonian effected the region that is today northern Laurentia, Greenland, the British Isle, and Norway, while the Acadian Orogeny happened further south along the eastern margin of current North America resulting in another wave of Appalachian mountain-building. One of the most recognizable unconformities on Earth was formed from the crustal deformation accompanying these events.

We know it was relatively warm in the Silurian through the middle Devonian, with extensive shallow seas as evidenced by the widespread coral reef deposits. Evaporite deposits reflect arid conditions and an episode of sea level regression near the end of the Silurian.

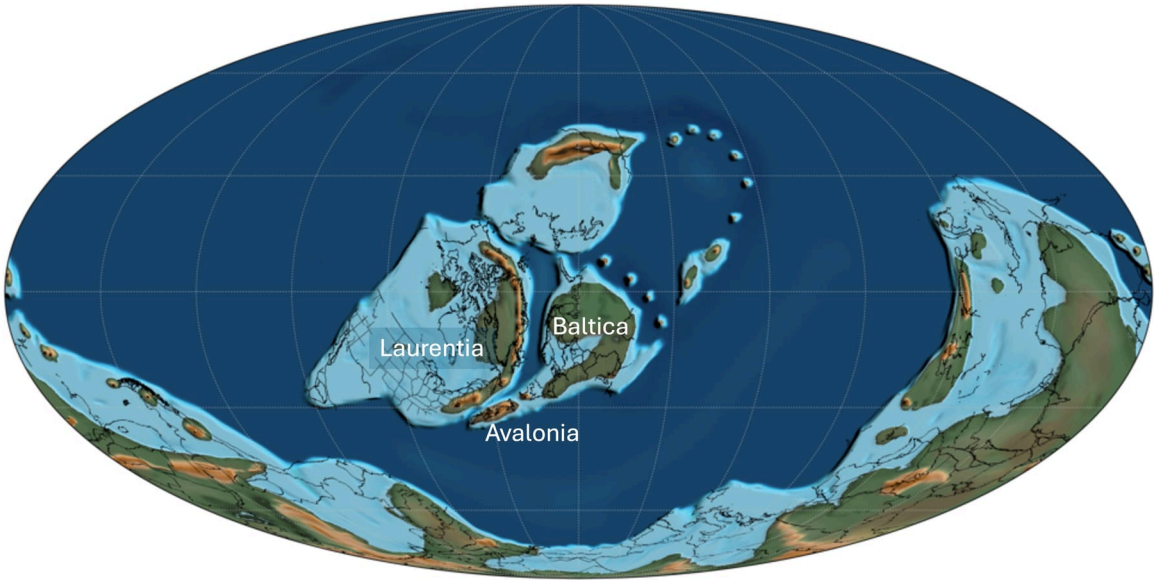


Figure 12.3.1: World map during the Silurian modified from Scotese, Christopher R.; Vérard, Christian; Burgener, Landon; Elling, Reece P.; Kocsis, Ádám T., CC BY 4.0 <https://creativecommons.org/licenses/by/4.0>, via Wikimedia Commons.



Figure 12.3.2: Angular unconformity with nearly verical Silurian strata overlain by gently dipping Devonian strata. Image from dave souza, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons.

In addition to unconformities and deformed crust, sedimentary sequences record the mountain-building of the Middle Paleozoic. They tell a story of prolonged erosion during the first part of the Silurian that established a temporary passive margin along which coral-stromatolite reef communities flourished. These deposits are abruptly overlain by deep-water flysch deposits, signaling the subsidence of the shallow sea, as Avalonia approached and caused the downward deformation of the foreland basin containing the Iapetus Sea. As mountain-building progressed, erosion of the highlands filled this basin, and the flysch deposits were replaced with coarser-grained terrestrial and shallow sea molasse sediments. This sequence of events was similar to what happened during the Taconian Orogeny and produced a depositional sequence that can be seen throughout the eastern US today.

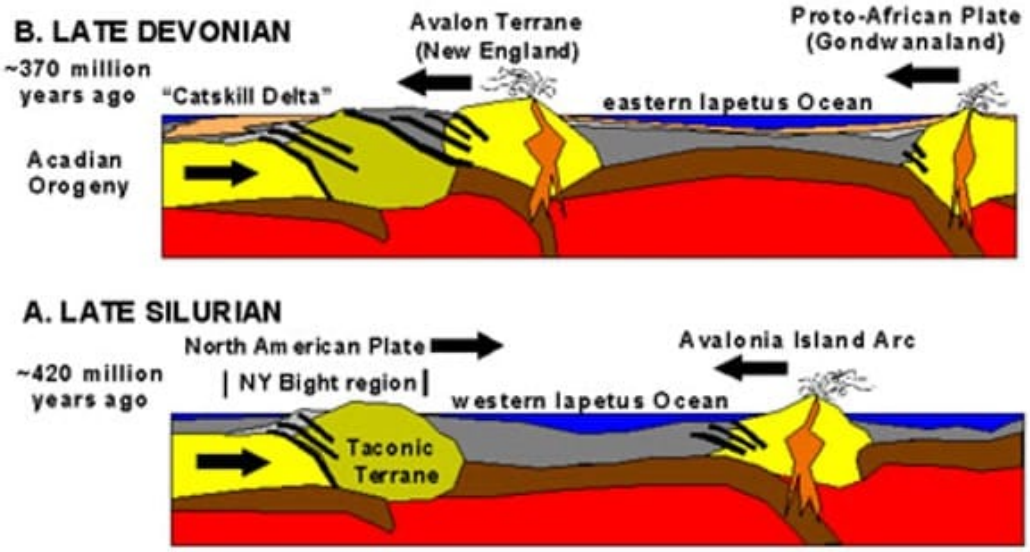


Figure 12.3.3: Image modified from USGS, Public domain, via Wikimedia Commons.

Western North America

Near the end of the Devonian, the western margin of Laurentia became active due to the collision with an island arc. This caused the **Antler Orogeny**, the first mountain-building event of the Cordilleran.

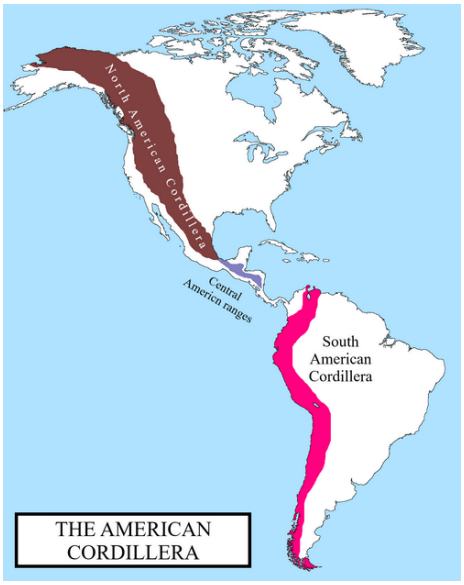


Figure 12.3.4: Map of the American Cordillera. Image from Knightoftheswords281, CC BY 4.0 <https://creativecommons.org/licenses/by/4.0>, via Wikimedia Commons.

Mass Extinction Event

The **late Devonian extinction event** demonstrates how biology, geology, and the atmosphere are interconnected. The spread of plants across the continents removed CO2 from the atmosphere, reducing the greenhouse effect and leading to global cooling. Plants also grew taller during the Devonian, and consequently, their root grew deeper. This caused the weathering of rock, which also used atmospheric CO2 and released elements, like Phosphorus, that made their way into the oceans. Phosphorus stimulated the growth of marine algae, thereby increasing the activity of bacteria that break down dead algae. The decay process required oxygen, and the oceans eventually became too deficient to support life, leading to widespread extinctions.

Key Terms

- **Acadian Orogeny** - a mountain-building event affecting a more southern part of the east coast of Laurentia occurring during the late Silurian through the Devonian resulting in the second wave of Appalachian mountain-building
- **Antler Orogeny** - first mountain-building event of the Cordilleran occurring at the end of the Devonian
- **Caledonian Orogeny** - a mountain-building event affecting the northern part of the east coast of Laurentia occurring during the late Silurian through the Devonian
- **late Devonian extinction event** - an extinction event at the end of the Devonian occurring as the oceans became too deficient in oxygen to support life

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12.4: Geography and Tectonics of the Late Paleozoic

Overview

The Carboniferous and Permian Periods comprise the late Paleozoic, a chapter in Earth's history that experienced the most intense ice age of the Paleozoic Era, followed by intense global warming and the greatest extinction of life our planet has experienced. High sea levels at the start of the Carboniferous created widespread shallow seas and extensive limestone deposits, and warm conditions in the equatorial regions allowed for coal forests to thrive. In these freshwater swamps, plants rapidly grew, died, and were buried, resulting in thick accumulations of peat. Over time, after burial, carbon-rich peat is transformed into coal. By the end of the Carboniferous, the coal forests died out in North America, and Pangaea was nearly fully assembled.

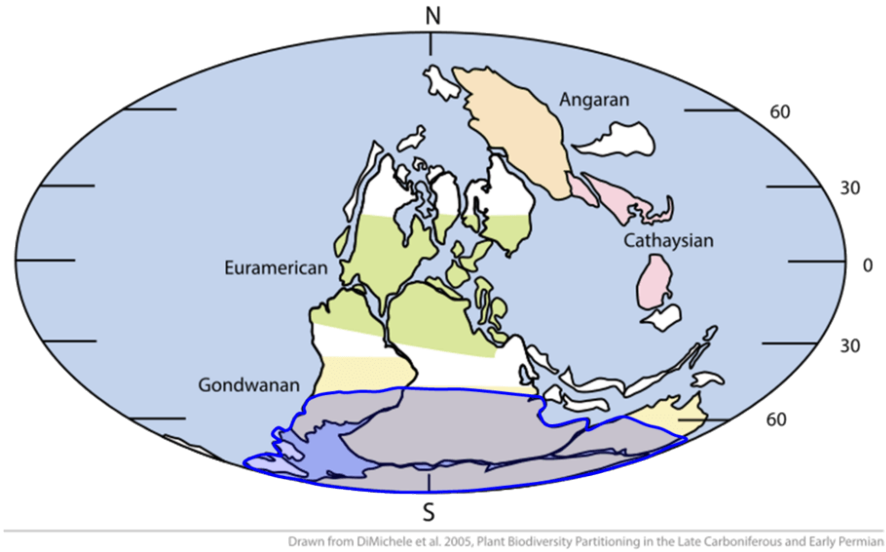


Figure. Configuration of the continents during the Carboniferous. Image from GeoPotinga, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons

The Carboniferous Period was named in Britain for the coal deposits that define rocks from this time. However, in North America, rock in the early half of this period is limestone, as typified by strata in Mississippi; in Pennsylvania, coal is the dominant lithology in late Carboniferous deposits. Consequently, this period is subdivided into the Mississippian and Pennsylvanian Periods in North America.

Mountain Building

The third and final episode of Appalachian mountain building happened during this time, the Alleghenian Orogeny. This folded strata into mountains in the southern part of the range during the collision of Euramerica and Gondwanaland. This convergence deformed the crust along the southern margin of North America, where today is Texas and Oklahoma, building the Ouachita Mountains.

Along North America's western margin, the Sonoma Orogeny picked up where the Antler Orogeny left off, building up mountains through the accretion of island arcs and other fragments of continental crust. This event also thrust seafloor sediments onto the continent and added considerable volume to its western margin.

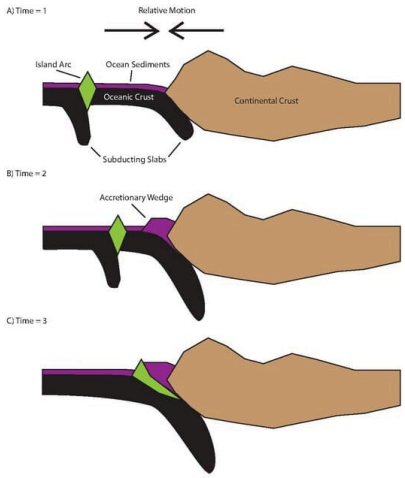
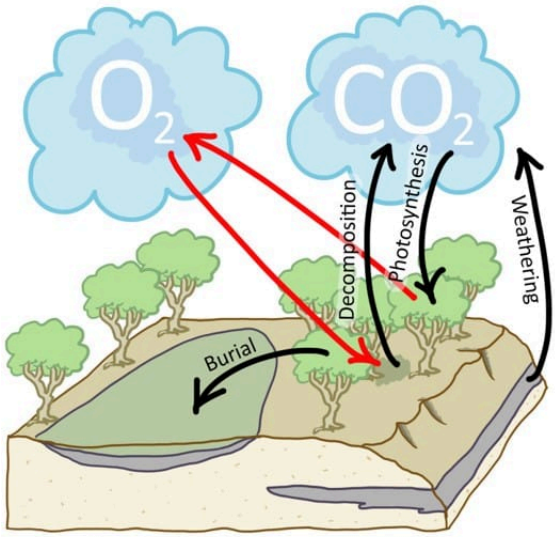


Figure. Image from BZenith, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons.

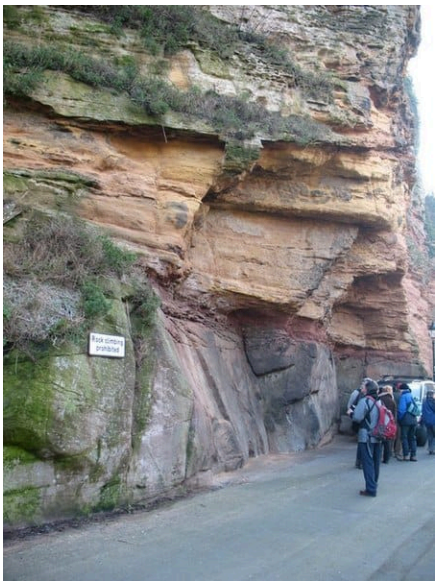
Ice Age

The most significant ice age of the Paleozoic Era started in the mid-Carboniferous. It lasted until the early Permian, when glaciers at low elevations were found even in equatorial regions, like during Snowball Earth. This happened because vegetation was rapidly buried in coal swamps, where bacterial decomposition couldn't keep pace with burial. Consequently, carbon was being buried instead of returning to the atmosphere through decomposition processes (see Carbon cycle). CO2 levels fell, and oxygen levels rose to as high as 35% (today, the atmosphere has 21% oxygen) as global warming decreased. This resulted in global cooling by lowering the concentration of the greenhouse gas CO2. It also increased the size of insects, which took advantage of the higher oxygen levels in the air (Stanley and Luczaj, 2015).



Carbon and oxygen fluxes between plants, the atmosphere, and the geosphere. (Callan Bentley cartoon.)

Evidence for this ice includes a global unconformity that formed as the ice age locked up water in glaciers, lowering sea levels, and exposing continental rock to erosion. The unconformity is exposed in outcrops around the world. More evidence comes in the form of tree rings, which have seasonal growth patterns indicative of the significant temperature swings associated with ice ages.



The Carboniferous unconformity. Image from Gordon Hatton.

The ice age was followed by global warming and drying that persisted through the end of the Paleozoic and were the driving factors behind the greatest extinction of life our planet has experienced, the Permian-Triassic (P-T) extinction. The drying was caused by the assembly of Pangea, which cut off inland portions from ocean-derived moisture due to its immense size and mountain ranges.

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12.5: Early Paleozoic Life - The Cambrian and Ordovician

The Cambrian Explosion: Hard Parts and Burrowing

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 12.5.1: Anomalocaris reconstruction by the MUSE science museum in Italy.

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 offered several evolutionary advantages: 1) it offered protection from physical and chemical variations in the environment, such as changes in water temperature and salinity; 2) it offered protection from biological predators; and, 3) it allowed muscles additional mechanical leverage during locomotion, making it easier to burrow.

Burrowing by a variety of new invertebrates churned up the sediment, disturbing the continuity of the bedding. This **bioturbation** is a “trace fossil” signal of the Cambrian Explosion. Indeed, the boundary between the Ediacaran and Cambrian Periods is defined by the appearance of a trace fossil, *Treptichnus pedum*, which indicates an expansion of animal mobility and more complex interactions in the ecosystem. The development of burrowing also increased organisms' ability to withstand seasonal shifts in currents and exploit food sources below the seafloor, which would further increase their chances of survival.

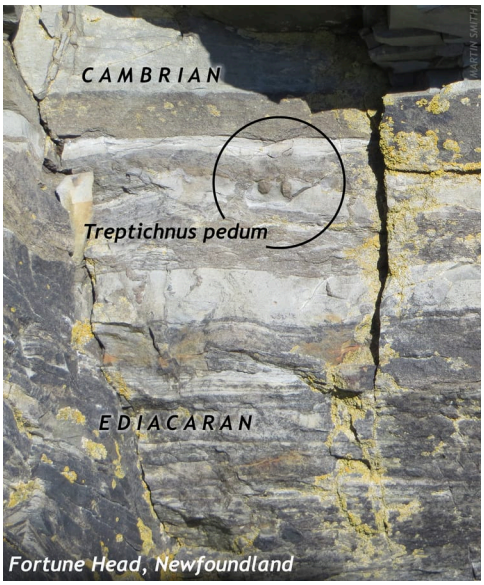


Figure 12.5.2: Treptichnus pedum defines the Ediacaran / Cambrian transition at the GSSP at Fortune Head, Newfoundland. (C. Bentley annotation of Martin Smith photo via Wikimedia.)

The Cambrian Fauna

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.

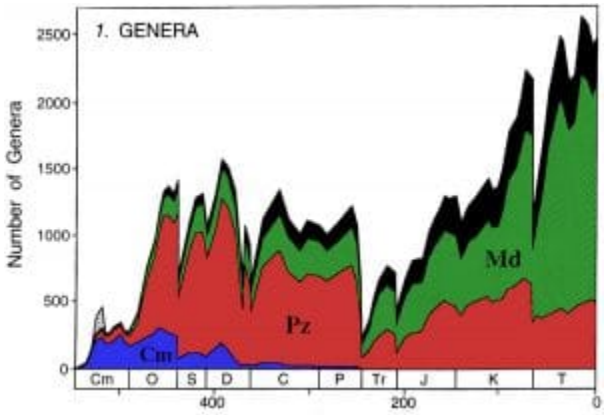


Figure 12.5.3: Sepkoski's (2002) Phanerozoic marine diversity curve. Total height of graph is total number of genera. Blue shading represents the Cambrian fauna, red shading is the Palozoic fauna, green shading is the Modern fauna, and black represents microfossils.

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.

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 is responsible for the incredible increase in diversity 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 12.5.4: Asaphiscus wheeleri, Cambrian trilobite from Utah. (Public domain image.)

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 around today, inarticulate brachiopods are much less common than they were in the early Paleozoic.

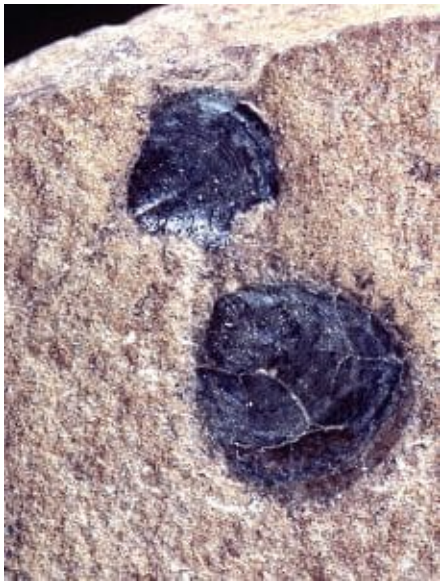


Figure 12.5.5: Inarticulate brachiopod *Lingula* sp. from the Cambrian Conasauga Formation of Alabama. (Photo courtesy of Encyclopedia of Alabama)

Exceptional Preservation: Burgess Shale Fauna

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.

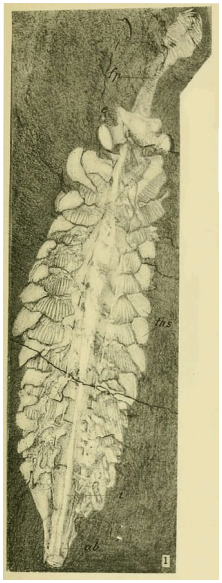


Figure 12.5.6: Original plate from Walcott's 1912 description of Opabinia, with labels: fp = frontal appendage, e = eye, ths = thoracic somites, i = intestine, ab = abdominal segment.

The Great Ordovician Biodiversification Event: Reefs Expand

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 (GOBE)**, 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 12.5.7: A modern coral reef.

One important evolutionary advancement during the Ordovician Period was reef-building organisms, mostly colonial coral. Corals took advantage of the ocean chemistry, using calcite to build large structures 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.



Figure 12.5.8: Guadalupe Mountains National Park in Texas is made of a giant fossil reef complex.

The Paleozoic Fauna

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 Sepkoski's diversity curve 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.



Figure 12.5.9: Diorama of a Late Ordovician seafloor with many representatives of the Paleozoic fauna. (Creative Commons Attribution 2.0 Generic license)

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.

End-Ordovician Mass 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 mass extinction was entirely limited to marine organisms, since land had yet to be colonized at that time. After the mass extinction was over, it took 50 million years for Earth's oceans to recover their former levels of diversity.

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 °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 did well in the aftermath of the end-Ordovician extinction.



Figure 12.5.10: Faceted and striated clast extracted from Ordovician strata in Arabia. (Modified from Figure 3 of Masri (2017).)

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. 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.



Figure 12.5.11: Map showing the inferred extent of end-Ordovician glaciation in Gondwana. (Map by Callan Bentley, redrawn after an original by Hoffmann and Linnemann, 2013.)

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. So, the earliest of five **mass extinction** events documented in the fossil record was initiated by cooling climate.

YouTube Video

This interactive element is not available in this format. Visit the link or scan the QR code to access it online.

View this YouTube Video online

4-minute video describing mass extinctions.

Key Terms

- **bioturbation** - fossil-based evidence of burrowing found as the continuity of the bedding is disrupted
- **Burgess Shale** - a Cambrian lagerstätte that allowed scientists to study Cambrian animals in immense detail
- **Cambrian Explosion** - the sudden appearance of biological diversity at the start of the Paleozoic Era
- **chordata** - a major animal phylum containing all vertebrates and some specialized invertebrates all sharing specific characteristics such as having a hollow dorsal nerve cord

- **evolutionary fauna** - a group of marine animal taxa with hard skeletons that appear, evolve, and go extinct together as a distinct, dominant community over a particular geological time
- **Great Ordovician Biodiversification Event (GOBE)** - an event during the late Ordovician in which vertebrates and invertebrates became more diverse and complex at family, genus, and species level
- **Lagerstätte** - a site of exceptional fossil preservation that includes impressions of soft body parts
- **mass extinction** - a widespread, rapid decrease in the Earth's biodiversity, where at least 75% of all species perish within a geologically short interval

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12.6: Middle Paleozoic Life - The Silurian and Devonian

Evolutionary Innovations: Vertebrate Diversification and Life on Land

After the End-Ordovician Mass Extinction, 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. 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 12.6.1: The armor-plated fish (placoderm) Bothriolepis panderi from the Devonian of Russia.

The Age of Fishes

Throughout the Cambrian and into the Silurian, jawless fish (agnathans) and armored jawless fish (ostracoderms) diversified. Jaws in vertebrates likely evolved from the gills in ostracodermi, and the spiny sharks (acanthodians) displayed the first jaws in the early Silurian. Other jawed fish, including armored fish (placoderms) and bony fish (osteichthyes) appeared during the Silurian and radiated during the Devonian. The oldest cartilaginous fish (chondrichthyes), i.e. sharks and rays, radiated in the earliest Devonian.

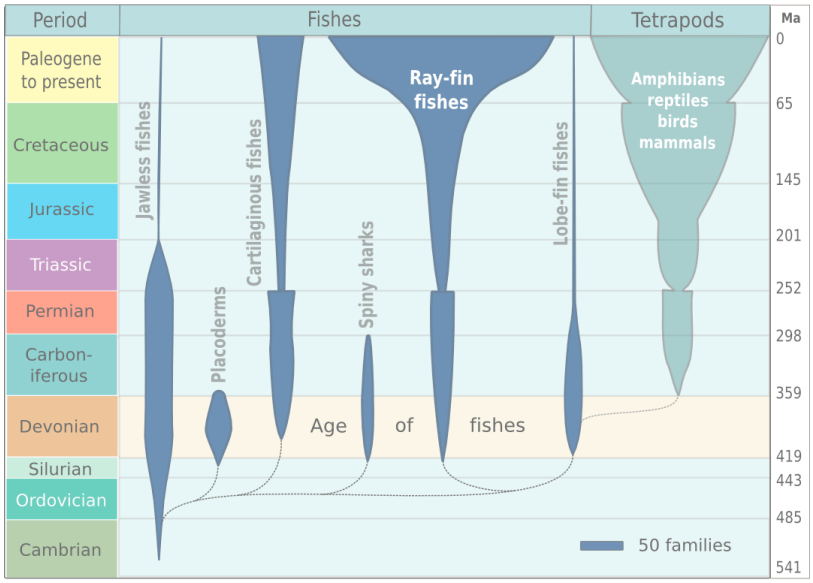


Figure 12.6.2: Spindle diagram of fish diversity through time. Vertical length of shapes indicate geologic range of species while the width represents diversity. (Creative Commons Attribution-Share Alike 3.0 Unported license.)

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 also 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. Several other fossils from the Devonian are more tetrapod-like than fish-like but were not fully terrestrial, as the first fully terrestrial tetrapods would not evolve until the early Carboniferous period.

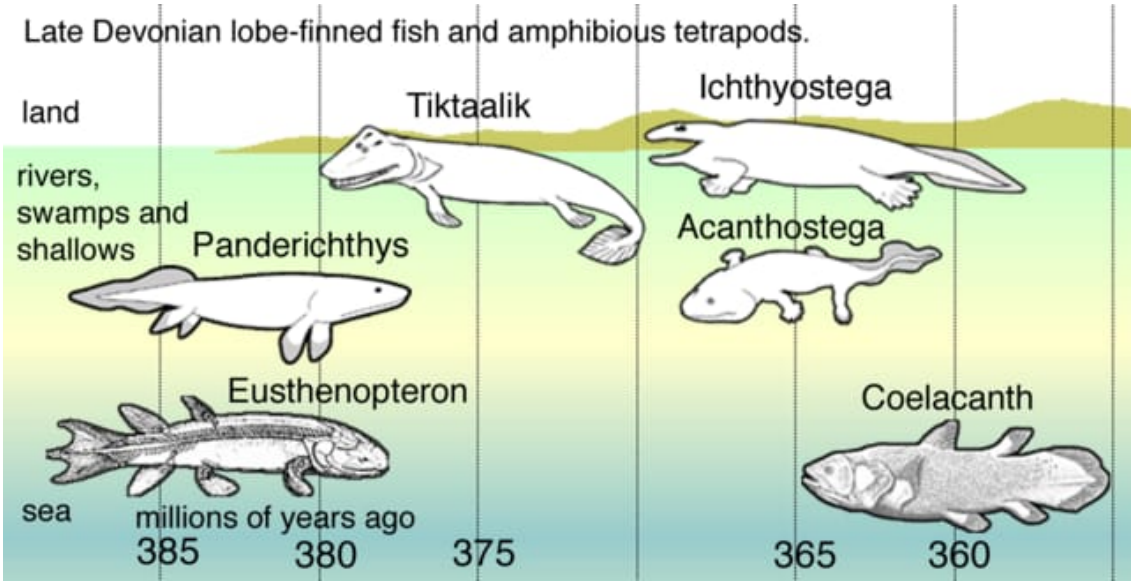


Figure 12.6.3: Several different types of fish and amphibians that led to walking on land.

Invertebrate Animals on Land

Invertebrate groups diversified significantly throughout the Paleozoic oceans, 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 by invertebrates is thought to have occurred during the early Paleozoic, but it was not instantaneous. Trace fossils from millipedes have been found from the Late Ordovician, and by the Silurian and Devonian, spiders, scorpions, centipedes, and silverfish all found their way onto land. Aquatic invertebrates would have had an advantage due to their exoskeletons, which allowed for protection and moisture retention as these organisms adapted first to freshwater ecosystems and then to dry 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.



Figure 12.6.4: Modern horseshoe crab showing specialized appendages and book gills. (Creative Commons CC0 1.0 Universal Public Domain Dedication)

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 appeared, often found in association with plant fossils, scorpions, and myriapods (i.e. millipedes, centipedes, and eventually insects). 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.

Plants on Land

Like the first animals, the first multicellular plants evolved in the ocean: marine green and red algae 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, which 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 they reproduce via spores. Later in the Ordovician, the first true vascular plants (trachaeophytes) evolved, and by the Devonian, this group looked like modern plants: 1) green with waxy leaves (containing cutin), 2) possessing a lignin-reinforced stem (for structural support), 3) with roots that conducted water and nutrients through their tissues, and 4) using seeds for reproduction.



Figure 12.6.5: Moss (a bryophyte) with spore releasing structures. (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.

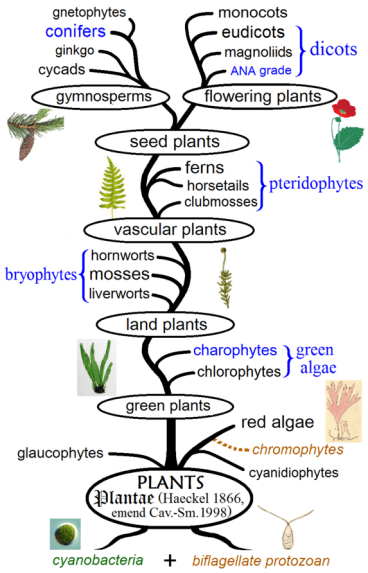


Figure 12.6.6: Diagram showing the phylogeny of major plant groups. (Creative Commons Attribution-Share Alike 4.0 International license.)

Gymnosperms were the dominant group of vascular, seeded plants that developed during the Devonian; this group includes seed ferns and eventually conifers, cycads, and ginkgos and are characterized by the ability to reproduce via wind-blown pollen released from cones. The most common type of plant on Earth today - the flowering plants (angiosperms) - would not evolve until the Mesozoic.

Late Devonian Mass Extinction

Toward the end of the Devonian, things started getting rough for organisms on Earth, but unlike other mass extinctions, this one took a while. The Late Devonian Mass Extinction was spread out over more than 20 million years, maybe as many as 25 million. It's not 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 (the Frasnian-Famennian stage boundary, or Kellwasser Event) and 359 Ma (the end-Devonian, or Hangenberg Event).

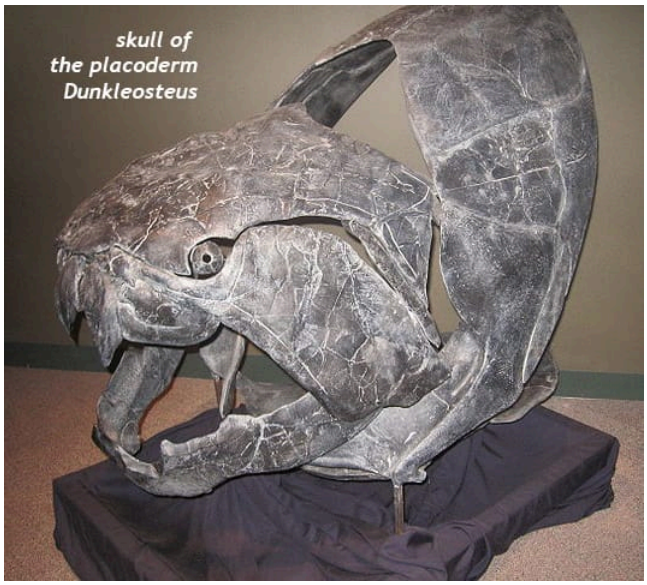


Figure 12.6.7: Dunkleosteus, a predatory placoderm from the Devonian that disappeared from the fossil record after the Hangenberg Event. (Public domain.)

By the end of this extinction event, 99% of reefs had been destroyed, the diverse fish world had been severely cut back, and many marine groups had been severely pruned 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. Taking over their shallow-sea habitat after the extinction were silica-skeleton sponges, which otherwise prefer cooler, deeper water.

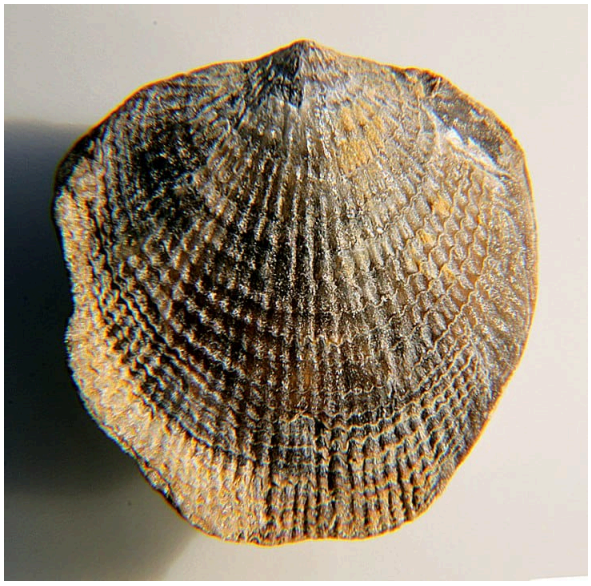


Figure 12.6.8: Atrypa reticularis, from Devonian strata in Germany. (CC-BY; Dwergenpaartje photo, via Wikimedia.)

One of the first clues that things went awry in the Late Devonian is the proliferation of black shale in the rock record. Black shale is mudrock that is rich in organic carbon; if muddy sediment is deposited under low-oxygen or **anoxic** (no oxygen) conditions, any organic carbon will be buried in the sedimentary record, giving its host rock a dark gray or black color. This is good evidence that the shallow ocean experienced prolonged and repeated periods of anoxic conditions throughout the Late Devonian.

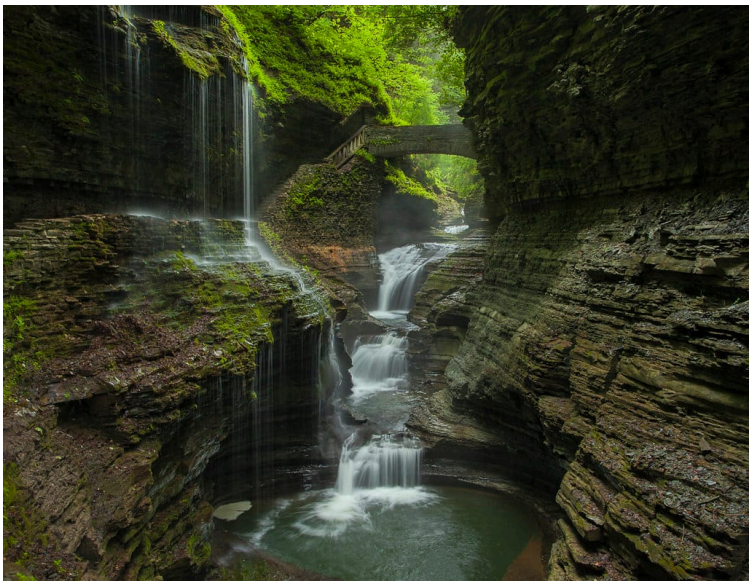


Figure 12.6.9: Watkins Glen, near Ithaca, New York, is a gorge cut into Devonian black shale of the Genesee Group. (CC-BY; Photograph by Barbara Friedman, via Flickr.)

Why was so much of the ocean anoxic in the late Devonian? The primary suspect in the case of Devonian anoxia and extinction is the proliferation of trees. Yes, you read that right: *trees*. 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.

Some of these liberated nutrients 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, which, when it reaches the ocean, accelerates the rate of reproduction of algae in a process called **eutrophication**. When these supercharged populations of algae die, they die in large numbers; the rotting of billions of dead algae cells consumes all the available oxygen in the water, creating anoxic "dead zones". Any fish swimming into such a dead zone must turn around immediately, or suffocate to death. (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.)



Figure 12.6.10: Eutrophication of the Potomac River adjacent to a wastewater input. (CC-BY; Photo by Alexandr Trubetskoy.)

Once the supply of oxygen is exhausted, remaining unrotted dead algae rain down on the seafloor, resulting in 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 enticing targets for natural gas exploration because of events that occurred during the Late Devonian Mass Extinction.

Key Terms

- **anoxic** - having no oxygen
- **arthropods** - a phylum of invertebrate animals characterized by a segmented body, a hard exoskeleton, and jointed appendages
- **eutrophication** - the over-enrichment of water by nutrients, primarily phosphorus and nitrogen, causing excessive algae growth and oxygen depletion
- **tetrapods** - a vertebrate animal with four limbs

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12.7: Late Paleozoic Life - The Carboniferous and Permian

The Carboniferous and Permian: Tetrapods and Plants Take Over the Land

The next time period, called the Carboniferous (North American geologists have subdivided this into the Mississippian and Pennsylvanian periods), saw the highest levels of oxygen ever known, with forests (e.g., ferns, club mosses) and swamps dominating the landscape. 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 12.7.1: A reconstruction of the giant arthropod (insects and their relatives) Arthropleura.

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.

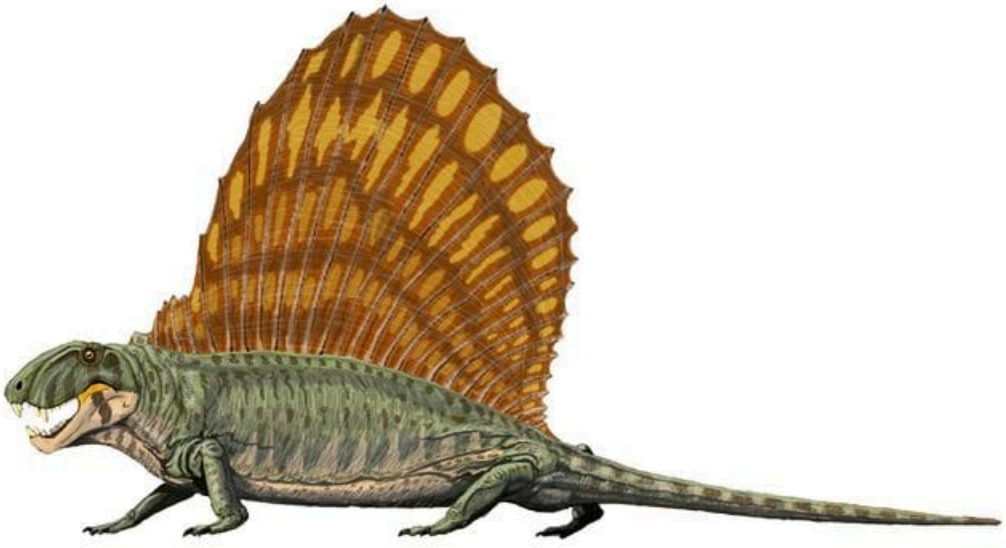


Figure 12.7.2: Reconstruction of Dimetrodon.

Invertebrates Learn to Fly

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.



Figure 12.7.3: Large fossilized Meganeura dragonfly. (Creative Commons Attribution-Share Alike 3.0 Unported, 2.5 Generic, 2.0 Generic and 1.0 Generic license.)

Tetrapod Evolution: Introducing Amphibians

The first fully terrestrial tetrapods transitioned to land in the Mississippian (early Carboniferous) period and were the ancestors of today's amphibians (although modern amphibian groups generally evolved during the Triassic). The amphibians were able to breathe air and live on land but still needed water to nurture their soft eggs. 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 12.7.4: Skeleton of Eryops, an early large carnivorous amphibian, from the Smithsonian's National Natural History Museum. (Image by Daderot via Wikimedia.)

Other structural changes to the vertebrate body plan were also required for success on land. The term "tetrapod" refers to animals able to walk 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 musculoskeletal template in their ancestors would have allowed for the evolution of muscular arms and legs. Specifically, sarcopterygian 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.

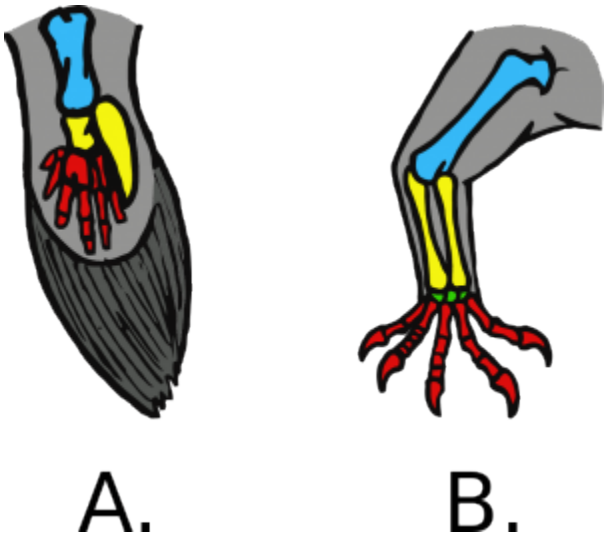


Figure 12.7.5: Comparison between the fins of A., Crossopterygiian fishes and the legs of B., tetrapods. Bones which correspond to each other have the same color. (Public domain image.)

Other changes are not particularly well-fossilized but also would have been key to success on land: 1) development of a three-chambered heart and more complex lungs to extract oxygen from air, 2) continual modification to auditory systems to allow for better hearing in air, and 3) development of more waterproof skin to retain tissue moisture. By the Mississippian (early Carboniferous) period, tetrapods had evolved into two main groups, amphibians and amniotes, from a common tetrapod ancestor.

Rise of the Amniotes: Origination of Reptiles

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 was the **amniotes**: a clade that includes reptiles (including dinosaurs and birds) and mammals. Exactly which animal qualifies as the first amniote is a subject of debate; *Casineria*, first described in 1999 from fossils from Scotland is a good candidate (although its exact classification is controversial). If this animal scurried in front of you right now, you would probably identify it as a reptile, even though it would not be considered a true reptile.

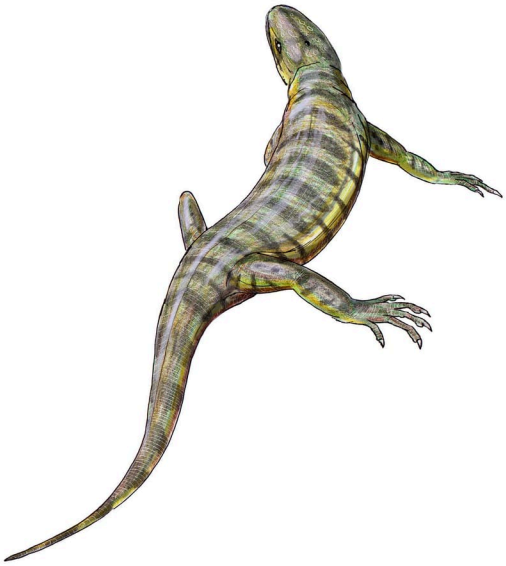


Figure 12.7.6: Casineria reconstruction (by ДиБгд via Wikimedia.)

The amniotic egg was an important evolutionary innovation that allowed tetrapods to expand out of terrestrial aquatic environments. 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.

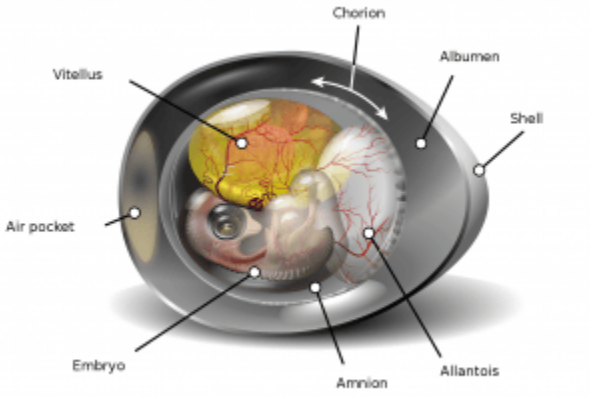


Figure 12.7.7: Modern chicken egg showing structures of amniotic egg. (Creative Commons Attribution-Share Alike 4.0 International license.)

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 Mass 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.

Amniotes Branch Off: Synapsids and Mammals

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).



Figure 12.7.8: Cynognathus, a cynodont from the Triassic of South Africa, pencil drawing by N. Tamura. (Creative Commons Attribution-Share Alike 3.0 Unported license.)

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 12.7.9: 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 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 12.7.10: Dimetrodon reconstruction. The posture is controversial, but here shown upright, based on possible Dimetrodon trackways. (Reconstruction by Max Bellomio 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.

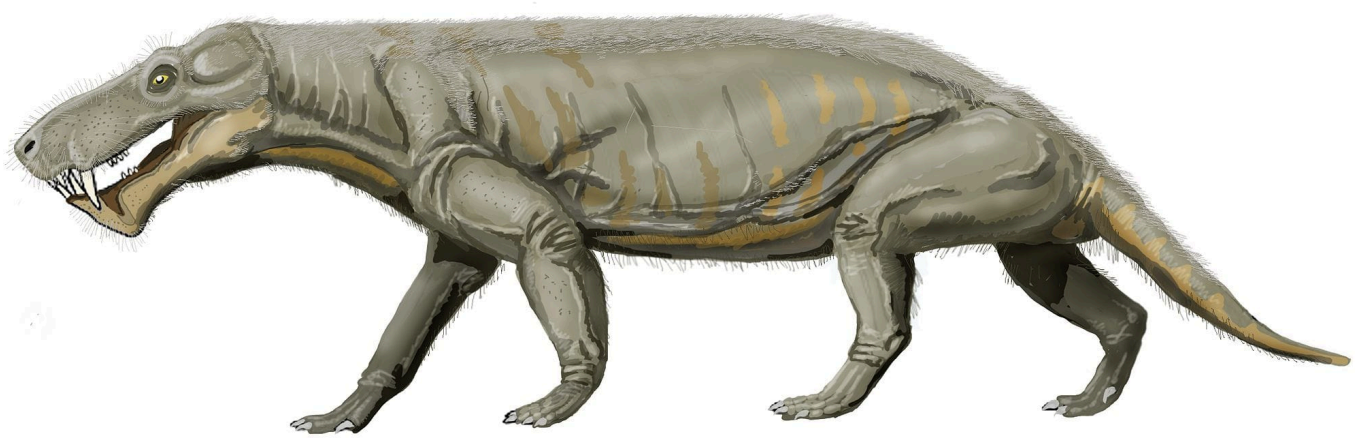


Figure 12.7.11: Gorgonops, a large Permian predatory therapsid. (Reconstruction by Dmitry Bogdanov via Wikimedia.)

End-Permian Mass Extinction

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 of marine invertebrates. Giant single-celled fusilinid foraminiferids were rolling around the seafloor like fat grains of rice.



Figure 12.7.12: Fusulinid foram fossil from Permian limestones in west Texas. (Callan Bentley photo.)

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.



Figure 12.7.13: Glossopteris sp. Late Permian, Australia. At the Royal Tyrrell Museum of Palaeontology. (Bloobityboop via Wikimedia, licensed under CC BY-SA 4.0.)

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.

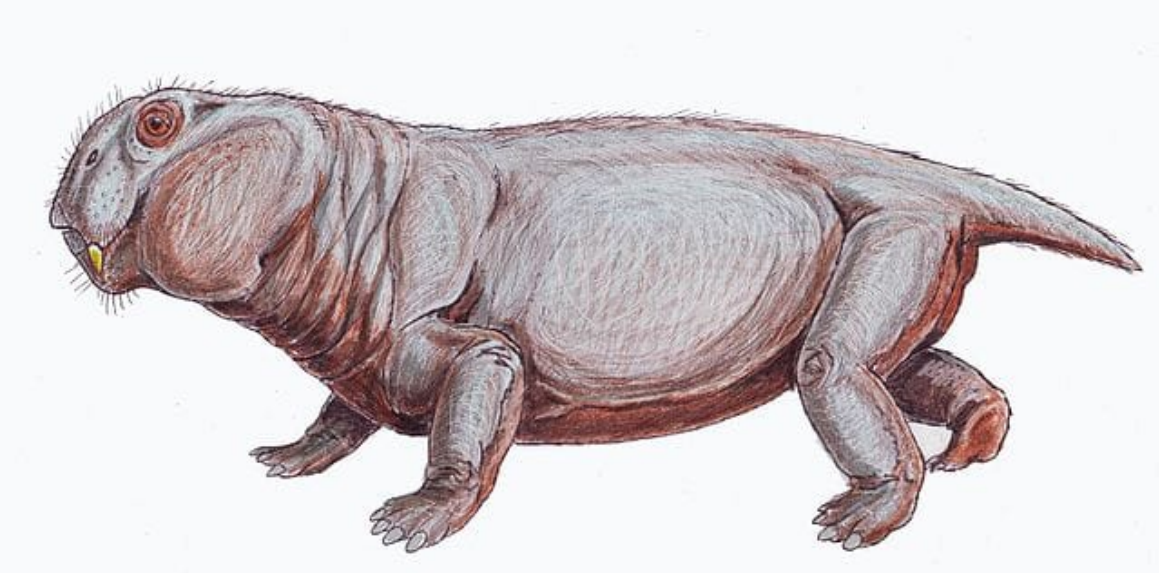


Figure 12.7.14: Lystrosaurus murrayi, about the size of a small dog. (CC-BY; Sketch by Dmitry Bogdanov, via Wikimedia.)

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.



Figure 12.7.15: 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.)

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 ^{18}O-based water molecules as likely to evaporate as ^{16}O-based water molecules. Using the ratio of ^{16}O to ^{18}O, it has been calculated that the global ocean warmed by about 6 °C. As with the discussion of 'cooling by 5 °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 CO_2 in the late Permian atmosphere.

Now let's consider where that extra CO_2 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.

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 12.7.16: Putorana Plateau, Russia, part of the Siberian Traps flood basalt province. (CC-BY; Photo by Александр Лещёнок, via Wikimedia.)

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 CO_2, which is a potent greenhouse gas.

In addition, 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 CO_2 to the total that was released! Estimates are that the level of CO_2 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 2026, the world was at about 430 ppm, up from ~280 ppm in pre-Industrial times.) So the end of the Permian had almost twenty times as much CO_2 in its atmosphere as we do today, an insane level that would have warmed the planet to an astonishing degree.

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 — stopped 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.

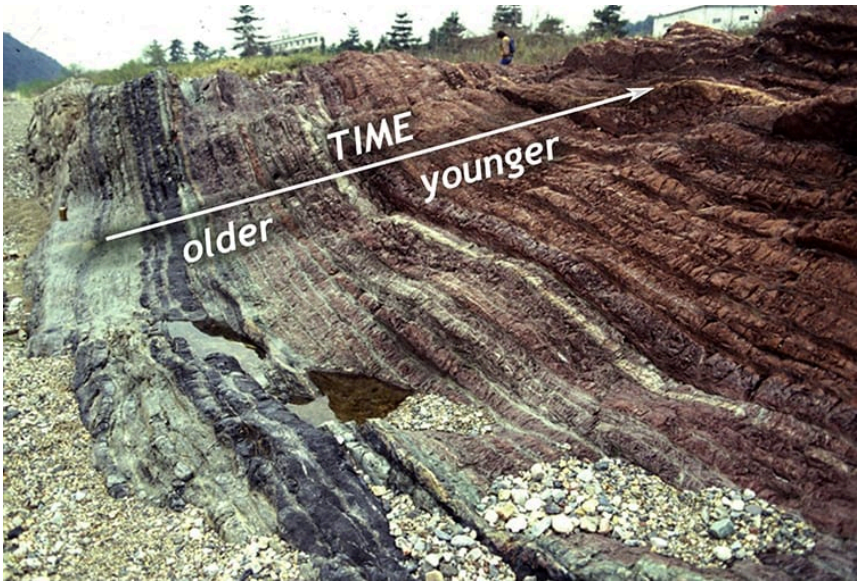


Figure 12.7.17: Outcrops in Inuyama, Japan, show a Triassic sequence of sedimentary strata that record changes in deep ocean oxygen levels after the end-Permian mass extinction event. The strata have been tectonically rotated due to accretion with the Japanese mainland. (Photo by Yukio Isozaki; reproduced with permission.)

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 (H_2S) as a waste product. You know H_2S: 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 (FeS_2) 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 12.7.18: 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 H_2S. In 2005, Lee Kump and colleagues suggested something bold – that the oceans were so euxinic during the end-Permian that H_2S 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.

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, *Claraia* is so omnipresent and numerous in early Triassic strata that it makes an excellent index fossil for the first part of the Triassic period: a bizarro "Clamworld."

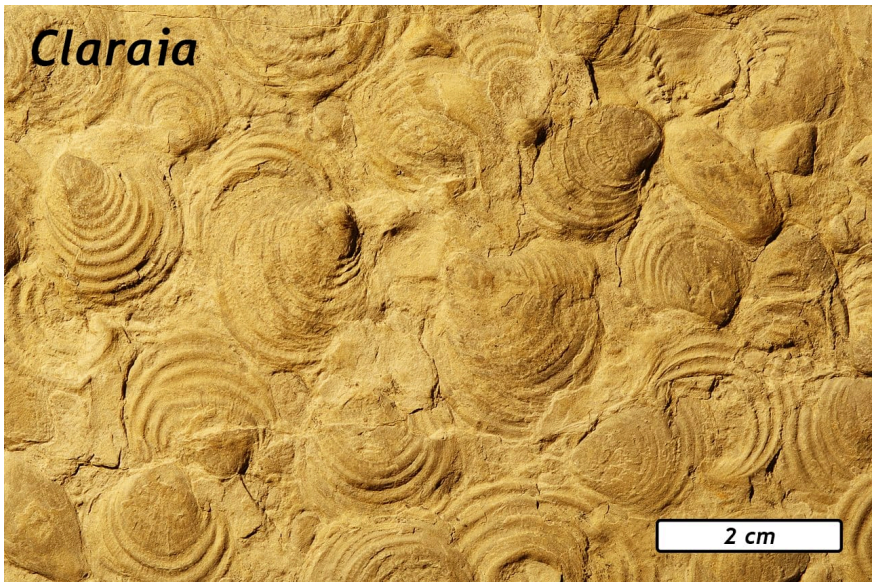


Figure 12.7.19: A layer of limestone made of nothing but shells of the clam *Claraia*. (Modified from a photo by Toedi3614, via Wikimedia.)

This was the biggest 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).

Key Terms

- **amniotes** - a group of tetrapod vertebrates comprising reptiles, birds, and mammals defined by their ability to lay eggs on land or retain fertilized eggs within the mother
- **synapsids** - a lineage of amniotes that includes mammals and their extinct "mammal-like reptile" ancestors, characterized by a specific skull structure with a single opening behind each eye
- **therapsids** - a major clade of tetrapods that gave rise to mammals

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12.8: Key Points

- The Cambrian Explosion marks the rapid rise of animal diversity and the first widespread appearance of hard body parts.
- Burrowing and bioturbation in the Cambrian transformed seafloor ecosystems and indicate increased animal mobility.
- Cambrian fauna included trilobites, brachiopods, and early echinoderms dominated early Paleozoic oceans.
- The Great Ordovician Biodiversification Event (GOBE) produced major increases in species-level diversity.
- The end-Ordovician mass extinction was driven by global cooling and glaciation on Gondwana. It was the first of Earth's five mass extinctions.
- Early Paleozoic tectonics shifted Laurentia's east coast from a passive margin to an active subduction zone.
- The Taconian Orogeny formed the first Appalachian Mountains during the early Paleozoic. The area went through two more events, the Acadian Orogeny during the middle Paleozoic and the Alleghenian Orogeny during the late Paleozoic.
- Major marine transgressions (Sauk, Tippecanoe, Kaskaskia, Absaroka) repeatedly flooded North America.
- Silurian innovations included the evolution of jaws and the first land plants and arthropods.
- The Devonian is the "Age of Fishes," with major diversification of placoderms, acanthodians, and lobe-finned fish.
- Lobe-finned fish such as Tiktaalik represent transitional forms leading to the first tetrapods.
- Arthropods were the first animals to colonize land, evolving book lungs and tracheae.
- Plants transformed Earth's surface through weathering, soil formation, and the evolution of vascular tissues.
- The Late Devonian mass extinction was driven by forest expansion, nutrient runoff, and widespread ocean anoxia. It was the second of Earth's five mass extinctions.
- Late Paleozoic climate swings included a major ice age followed by warming, alongside mountain building (Alleghenian, Sonoma) and the setup for the Permian–Triassic extinction. It was the third of Earth's five mass extinctions.
- Tetrapods diversified on land during the Carboniferous and Permian, giving rise to amphibians, reptiles, and early synapsids such as Dimetrodon and therapsids.
- Vast Carboniferous forests and high oxygen levels fueled the rise of giant arthropods and the first flying insects, which underwent a major adaptive radiation.
- The end-Permian mass extinction driven by massive CO_2 release, global warming, ocean stagnation, and euxinia was the most severe extinction in Earth's history and reshaped life on land and in the oceans.

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13: PHANEROZOIC EON - MESOZOIC ERA

Introduction

The Mesozoic (“middle life”) Era started 252 million years ago after the Permian-Triassic mass extinction event and ended 66 million years ago when another mass extinction event occurred. It is comprised of three periods: the Triassic Period, the Jurassic Period, and finally the Cretaceous Period.

Here is the the simplified geologic time scale used in the Geologic Time chapter of this book. The Mesozoic Era is highlighted in the middle.

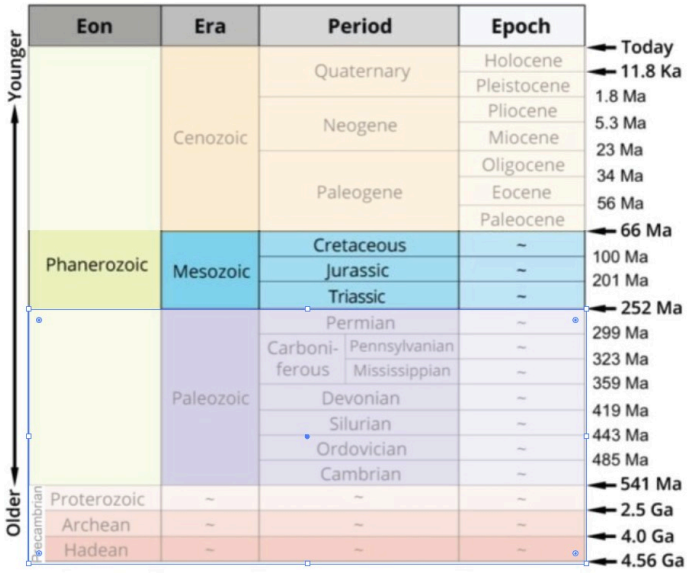


Figure 13.1: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

As Pangea started to break apart during this era, North America's landscape continued to take shape. Mammals, birds, and flowering plants continued to develop however the main characters for this chapter in Earth's history are probably the dinosaurs. For this reason, the Mesozoic is known as the age of reptiles.

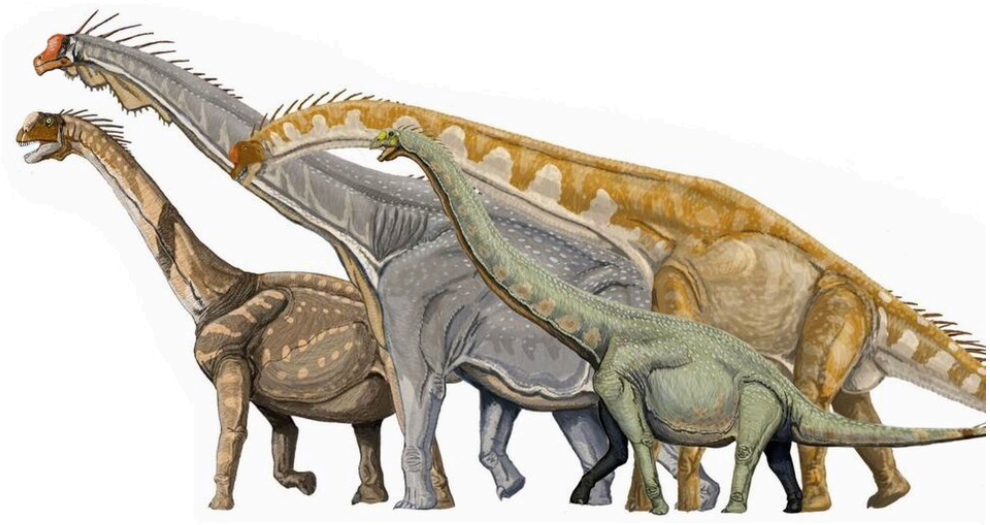


Figure 13.2: Sauropod Dinosaurs. Image from Богданов dmitrchel@mail.ru, CC BY-SA 3.0 <http://creativecommons.org/licenses/by-sa/3.0/>, via Wikimedia Commons.

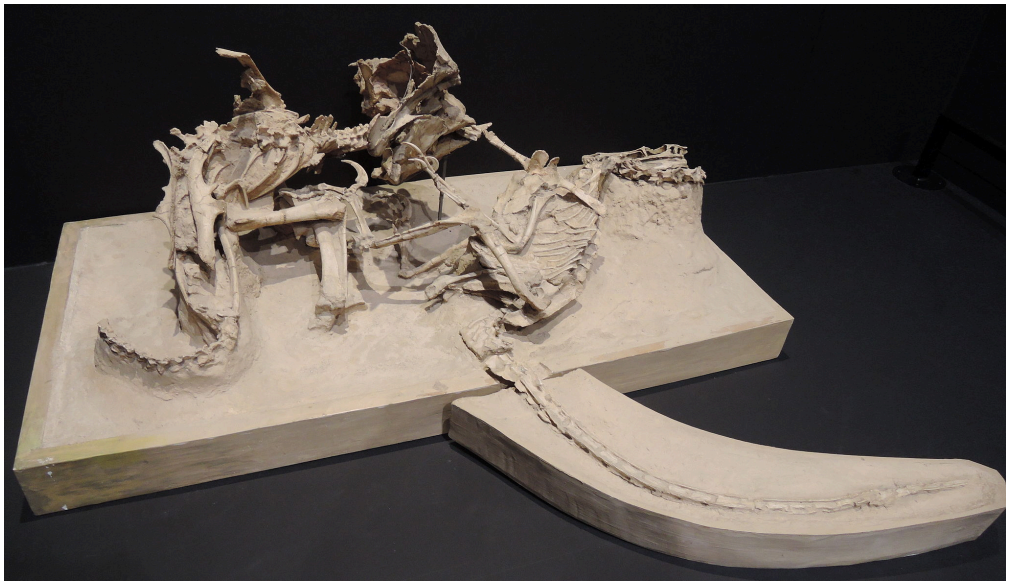


Figure 13.3: Perhaps the greatest fossil ever found, a velociraptor attacked a protoceratops, and both were fossilized mid sequence.

13.1: Paleography and Tectonics of the Mesozoic

13.2: Mesozoic Evolution - Reptiles and Amphibians

13.3: Mesozoic Evolution - Amniotes

13.4: Mesozoic Evolution - Dinosaurs

13.5: Mesozoic Mass Extinctions

13.6: Key Points

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13.1: Paleography and Tectonics of the Mesozoic

Breakup of Pangea

Early in the Mesozoic, Pangaea was fully assembled, but by the late Triassic, around 210 million years ago, Pangaea began to break up. Evidence for this process includes the age of the sediments in the Newark Supergroup, which was deposited in rift valleys formed during continental rifting. Additional evidence is the ages of the Palisades sill and of the eastern part of North America and the Atlantic Ocean floor.

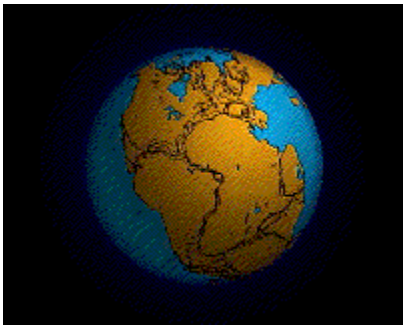


Figure 13.1.1: Animation showing Pangea breaking up

Due to seafloor spreading, the oldest rocks on the Atlantic floor are along the coasts of northern Africa and the east coast of North America, while the youngest are along the mid-ocean ridge. The age of these oldest basalts supports a timing for the breakup around 210 Ma.

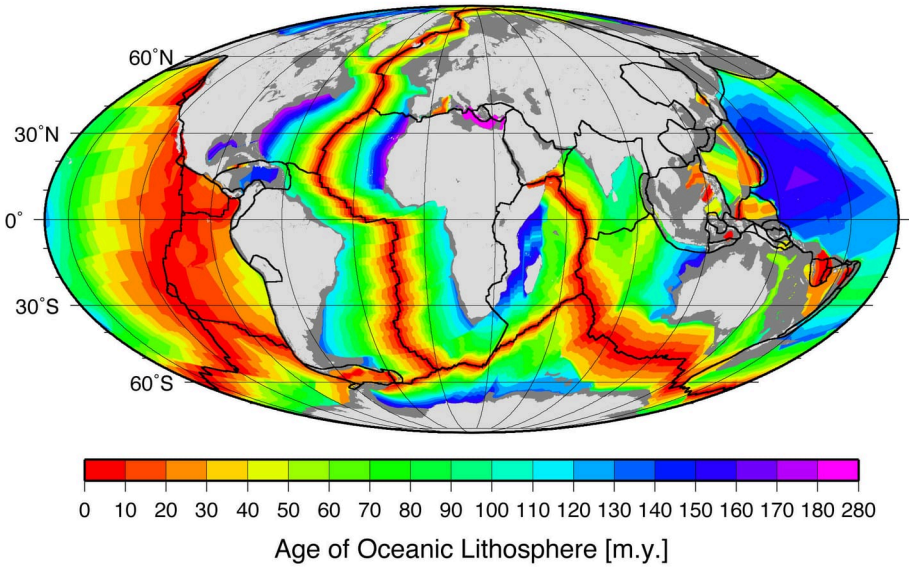


Figure 13.1.2: Age of oceanic lithosphere, in millions of years. Notice the differences in the Atlantic Ocean along the coasts of the continents.

This age pattern shows how the Atlantic Ocean opened as the young Mid-Atlantic Ridge began to create the seafloor. This means that the Atlantic Ocean began to form and open up 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 rifted plate margins eventually became the passive plate boundaries along the east coast of the Americas today. Here is a video by Tanya Atwater of Pangea breaking apart.

During the Mesozoic, the western margin of North America was active as it moved away from Pangaea, and the eastern margin became passive after the rifting that broke the continent free from Eurasia subsided. The convergence of North America with the ancient oceanic Farallon Plate drove subduction and the accretion of terranes in the west during four orogenic episodes:

the Sonoma (continuing from the late Paleozoic), Nevada, Sevier, and Laramide orogenies. These tectonic events built up the mountain ranges that make up the North American Cordilleran and added considerable mass to the continent through the accretion of terranes.



Figure 13.1.3: Map of the American Cordillera. Image from Knightoftheswords281, CC BY 4.0 <https://creativecommons.org/licenses/by/4.0>, via Wikimedia Commons.

In response to the mountain building, basins developed into which sediment was deposited from the adjacent highlands. One was the Cretaceous Western Interior Foreland Basin, which flooded during high sea levels, forming the Cretaceous Interior Seaway. This created a shallow epicontinental seaway that extended from the Gulf of Mexico to the Arctic Ocean, dividing North America into two separate land masses: Laramidia to the west and Appalachia to the east, for approximately 25 million years.



Figure 13.1.4: The Cretaceous Interior Seaway in the mid-Cretaceous.

The Sonoma Orogeny

During the **Sonoma Orogeny**, which spanned from the late Permian to the Triassic, mountains were formed through the accretion

of island arcs and other fragments of continental crust. This event also thrust seafloor sediments onto the continent, adding considerable volume to its western margin.

The Navadan Orogeny

Like the Sonoma Orogeny, the **Nevadan Orogeny** occurred approximately 155 Ma to 145 Ma (Schweikert, R. et al., 1984)* during the Late Jurassic to Early Cretaceous period, when North America collided with an island arc that had formed offshore. This resulted in the accretion of more material onto the continent, creating a large Andes Mountains-style volcanic arc and a belt of thrust faults (thrust belt), and generated magma that crystallized into rock that forms the core of mountain ranges like the Sierra Nevadas, the San Gabriels, and the San Bernardinos in California.

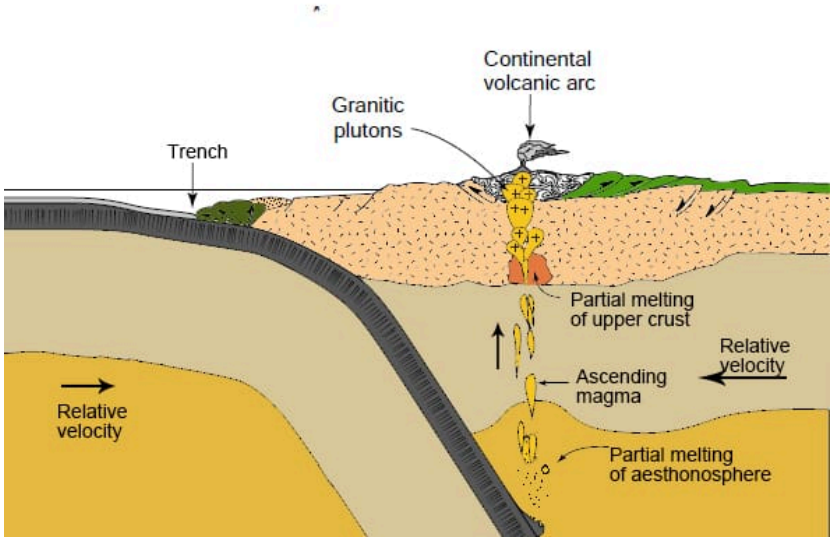


Figure 13.1.5: Andes Mountains-style subduction zone. Image from MichaelZuo, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons.

The deformation of the Nevadan orogeny is most easily observed in the Sierra Nevada Mountains of California and the Klamath Mountains along the coast of California and Oregon.



Figure 13.1.6: A map of California's geomorphic provinces, including the Klamath and Sierra Nevada Mountains. Image from the California Department of Conservation: https://www.conservation.ca.gov/cgs/...fornia-geotour.

This deformation is preserved as folded and faulted rock.

Interactive Content

This interactive element is not available in this format. Visit the link or scan the QR code to access it online.

View this Interactive Content online

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.

The Sevier Orogeny

The next orogeny of western North America was the **Sevier Orogeny** that spanned 160 million years (Ma) ago to around 50 Ma (Giallorenzo, M.A., et al., 2018)**. While terranes still caused crustal deformation, the Sevier orogeny occurred further inland and was more widespread than the Nevadan orogeny.



Figure 13.1.7: The Sevier fold and thrust belt. Image from Quickdraw123, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons.

The earliest events associated with the Sevier orogeny are as early as 160 Ma, which predates the Nevadan Orogeny. The latest deformation occurred approximately 50 million years ago, during the Cenozoic Era, while most deformation events took place between 120 and 80 million years ago, during the Cretaceous Period.

Figure 13.1.8: Paleogeography of Earth around 105 Ma.

The most significant aspect of the tectonic event was that the angle of the subducting plate decreased. This occurred as the

spreading center between the Farallon and Pacific plates approached the continental margin, meaning the ocean floor of the Farallon plate was relatively young and warm, making it more buoyant.

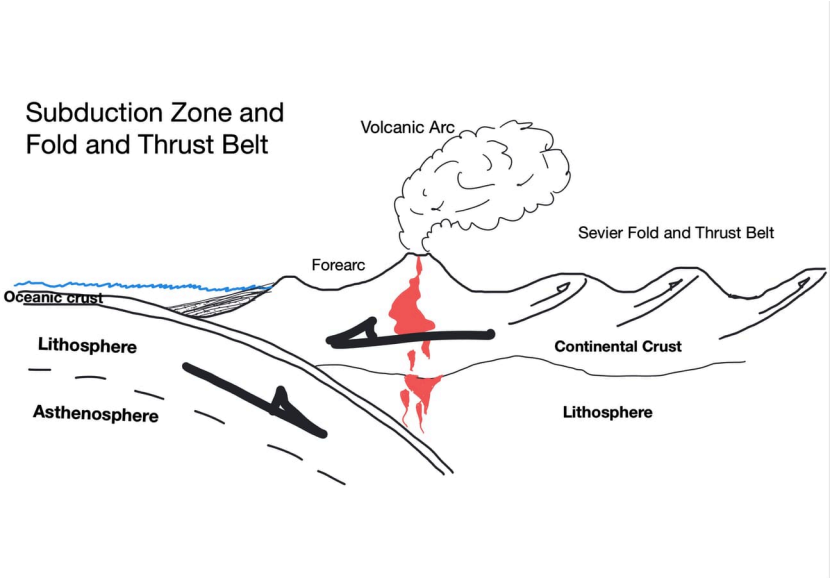


Figure 13.1.9: Sketch of a subduction zone and fold and thrust belt.

This caused more resistance to subduction and additional convergent forces in the overriding North American plate, resulting in a fold-and-thrust belt that extends far 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. Evidence for this includes igneous and sedimentary rocks that were thrust and folded upward from deeper in the crust, along faults, placing older rock atop younger rock. As new terranes accreted to the continent from the west, they pushed the pre-existing sedimentary layers out of the way towards the east. While most of the Farallon Plate was subducted during this event, 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.

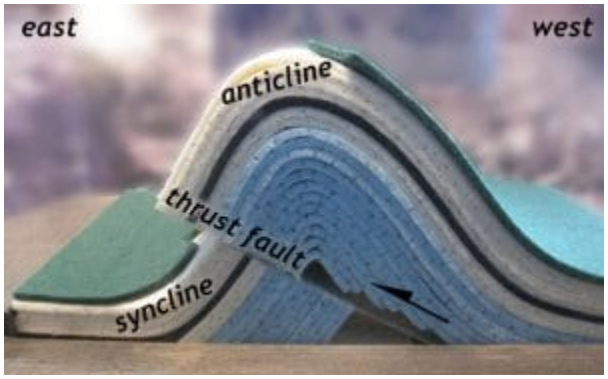


Figure 13.1.10: This felt model of Turtle Mountain, Alberta is on display at the Frank Slide Visitor Center. The structure shown is very similar to the structure of many faults and folds in the Rockies caused by Sevier deformation.

As 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 typical of active and ancient compressional mountain belts worldwide. This style of deformation is known as thin-skinned deformation.

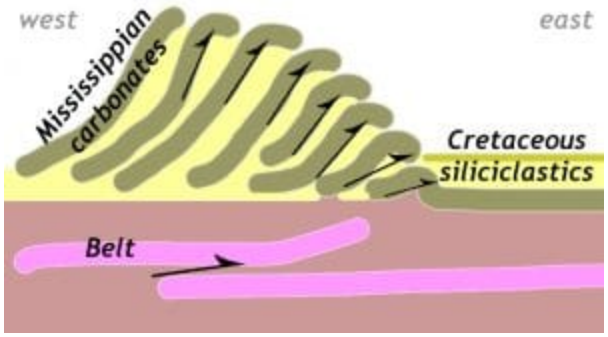


Figure 13.1.11: Cartoon cross-section through an imbricate stack of thrust sheets, as seen at Sun River Canyon, near Glacier National Park.

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 when their orientations are so close to horizontal. The rock body above the thrust fault is sometimes referred to as a thrust sheet or nappe.

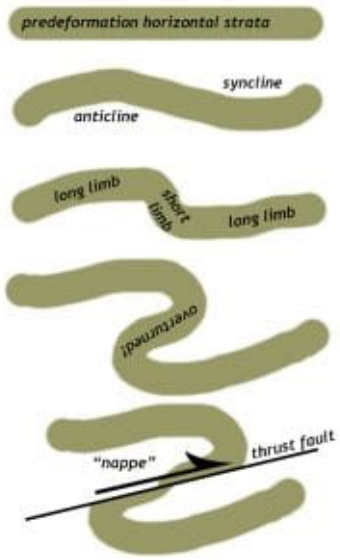


Figure 13.1.12: A cartoon cross-section showing 5 steps in the folding and thrust faulting of an asymmetric anticline to make a "nappe."

Nappes form when horizontal rock layers are compressed, buckling into asymmetric folds and thrust faults. Once a fault forms, 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) where the crust is weaker. 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. As deformation proceeds, the older, more western thrust sheets get rotated to steeper dip angles. This whole "stack" of nappes is said to be "shingled," or "imbricated." The animated GIF below illustrates these processes, although in reality, they occur over tens of thousands of years.

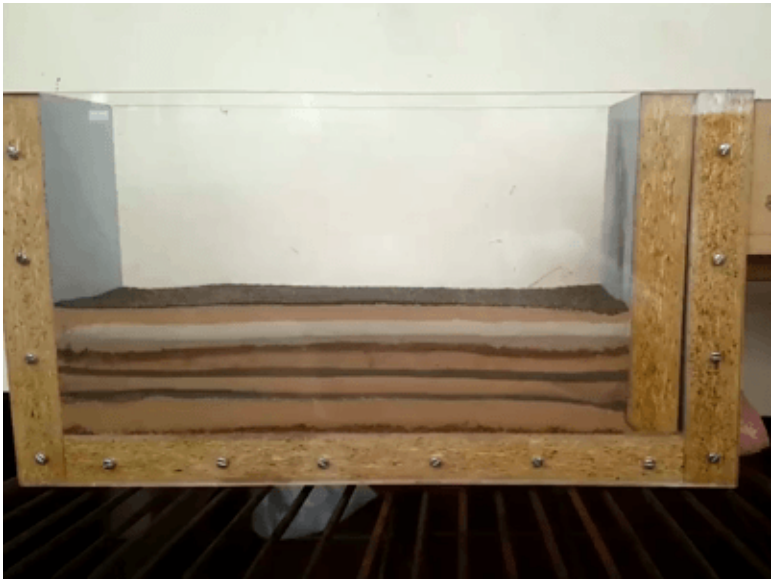


Figure 13.1.13: Sandbox deformation experiment showing how layers of sand fold and fault due to lateral compression. This closely mimics the evolution of fold-and-thrust faults in the Sevier orogeny (except in this model, compression comes in from the east instead of the west). Model and source video by Marco Martins-Ferreira.

Laramide Orogeny

The fourth mountain-building event that occurred in the western US during the late Mesozoic and early Cenozoic, spanning approximately 80 to 35 million years ago, was the **Laramide Orogeny**. Its timing overlaps the Sevier orogeny and was also caused by the subduction of the Farallon Plate. Still, it was distinct from the Sevier, due to several factors: (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.

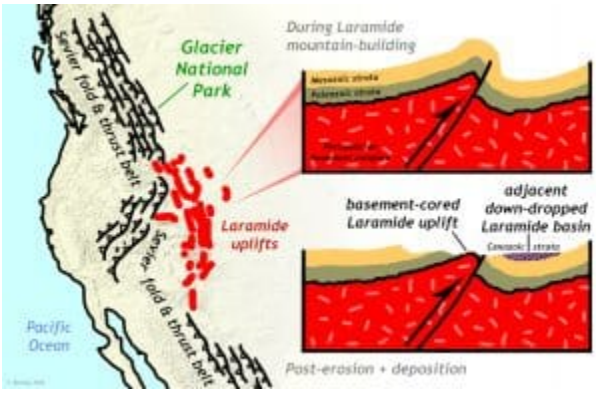


Figure 13.1.14: Map showing the geography of the Sevier fold and thrust belt relative to Laramide uplifts. Cross-sectional views show the structure of the basement-cored Laramide uplifts.

The subducting plate was younger and hotter than during the Sevier orogeny, resulting in the subduction angle becoming increasingly shallow over time. Eventually, it became so small that the slab was dragging along the base of the overriding plate. This is known as flat-slab subduction. This put pressure on the overriding plate, causing warps in the crust to develop from beneath, affecting the overlying continent hundreds of miles east of the continental margin, where it built high mountains. The very low angle also eliminated the lithospheric mantle wedge as the ocean plate was in direct contact with the crust, which prevented magma from forming. Consequently, there was no magmatism during the Laramide orogeny, or at least none related to subduction as it had occurred in the previous orogenies.

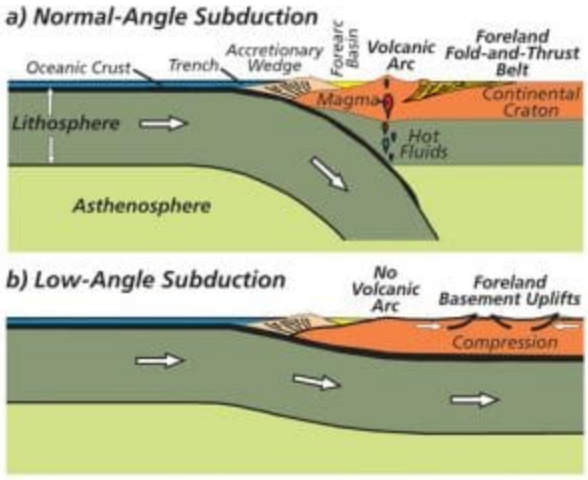


Figure 13.1.15: Normal (a) and flat-slab (b) subduction. Picture by R. J. Lillie and Parks and Plates by the National Park Service.

Deformation style

The Laramide orogeny's style of deformation is thick-skinned, meaning that the crystalline rocks deep in the crust were deformed, resulting in 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 Earth's surface, generating lowlands. These sedimentary basins were filled with large volumes of detrital sediment from nearby eroding mountains, creating new Cenozoic-aged 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.

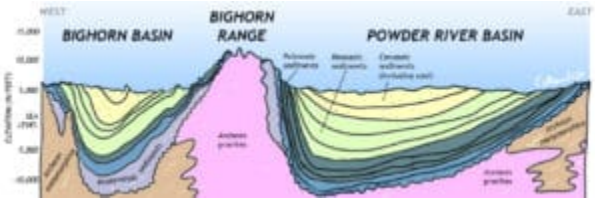


Figure 13.1.16: The Laramide Orogeny generated huge synclinal basins where sediments accumulated through the Cenozoic.

Location

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 (deep crystalline rock) Laramide deformation. In contrast, the Sevier fold and thrust belt is longer and further west, in a sinuous arc that runs from the Alberta/British Columbia border through western Montana, Wyoming, Utah, and Nevada, and to Mexico.



Figure 13.1.17: 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.

Key Terms

- **Laramide Orogeny** - a major mountain-building event that affected western North America from roughly the Late Cretaceous to the early Paleogene (about 80–40 million years ago.
- **Nevada Orogeny** - a Middle to Late Jurassic (roughly 170–140 million years ago) compressional event along the western margin of North Americ
- **Sevier Orogeny** - a major compressional mountain-building episode that occurred from the Late Jurassic through the Late Cretaceous (approximately 140–50 million years ago)
- **Sonoma Orogeny** - an Early Triassic (about 250–230 million years ago) mountain-building event along the western margin of North America

* Schweikert, Richard; Bogan, Nicholas L.; Girty, Gary H.; Hanson, Richard E.; Merguerian, Charles (1984). "Timing and Structural Expression of the Nevadan Orogeny, Sierra Nevada, California". *Geological Society of America Bulletin*. **95** (8): 967–979. doi:10.1130/0016-7606(1984)95<967:taseot>2.0.co;2.

** Giallorenzo, M.A.; Wells, M.L.; Yonkee, W.A.; Stockli, D.F.; Wernicke, B.P. (2018-03-01). "Timing of exhumation, Wheeler Pass thrust sheet, southern Nevada and California: Late Jurassic to middle Cretaceous evolution of the southern Sevier fold-and-thrust belt". *GSA Bulletin*. **130** (3–4): 558–579. doi:10.1130/B31777.1. ISSN 0016-7606.

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13.2: Mesozoic Evolution - Reptiles and Amphibians

Introduction

Since the Mesozoic Era begins with the Permian-Triassic mass extinction, the Triassic saw devastated ecosystems that took over 30 million years to fully re-emerge. The first appearance of many modern animal groups that would later flourish occurred at this time. This includes frogs (amphibians), turtles (reptiles), marine ichthyosaurs and plesiosaurs (marine reptiles), mammals, and the archosaurs.



Figure 13.2.1: A Mesozoic scene from the late Jurassic.

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 which ultimately dominated this era. 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.

Where Did Reptiles Come From?

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 13.2.2: The famous Berlin Specimen of Archaeopteryx. (Picture by H. Raab (User: Vesta) via Wikimedia.)

Amphibians: Precursors to Reptiles

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.



Figure 13.2.3: Skeleton of Eryops, an early large carnivorous amphibian, from the Smithsonian's National Natural History Museum. (Image by Daderot via Wikimedia.)

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*.

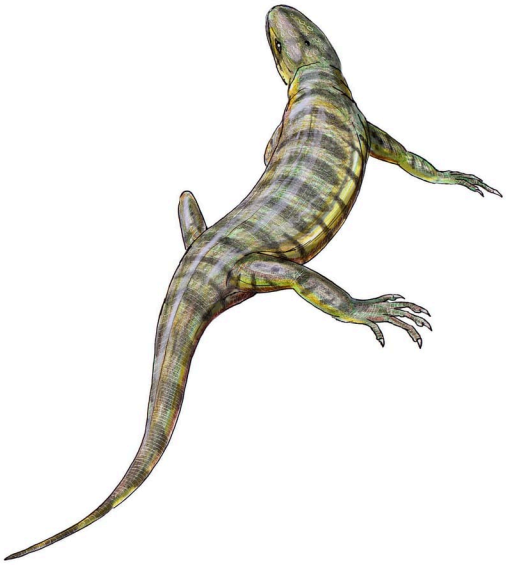


Figure 13.2.4: Casineria reconstruction (by ДиБгд via Wikimedia.)

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13.3: Mesozoic Evolution - Amniotes

Evolution of Amphibians

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.

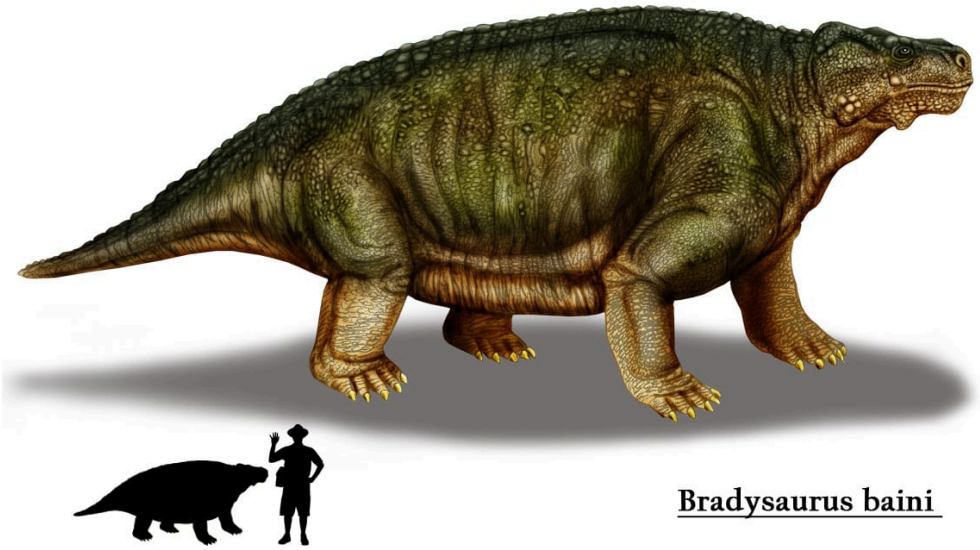


Figure 13.3.1: Bradysaurus, a parareptile (pareiasaur), with a size comparison to an adult human. (Image by Karkemish via Wikimedia.)

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.

Amniote Branch One: Synapsids and Mammals



Figure 13.3.2: 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 13.3.3: 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.

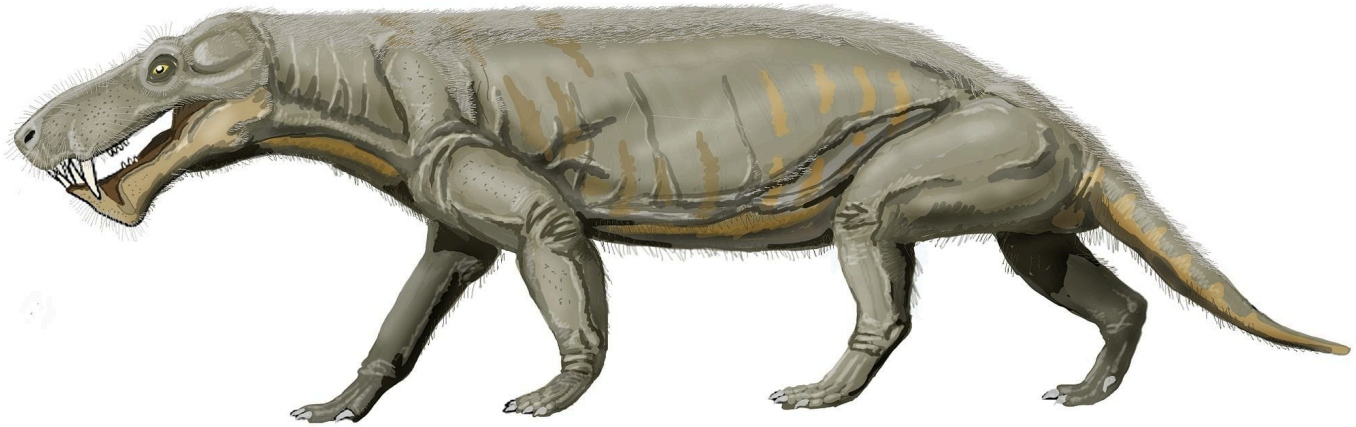


Figure 13.3.4: Gorgonops, a large Permian predatory therapsid. (Reconstruction by Dmitry Bogdanov via Wikimedia.)

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 13.3.5: Skull of Cynognathus. (Photo by Ghedoghedo via Wikimedia.)

Most of the therapsids went extinct at the Permian-Triassic mass extinction, and even more went extinct at the mass extinction occurring at the end of the Triassic Period. One subgroup that survived both were the cynodonts. Meaning 'dog tooth,' cynodonts possessed even more mammalian traits.

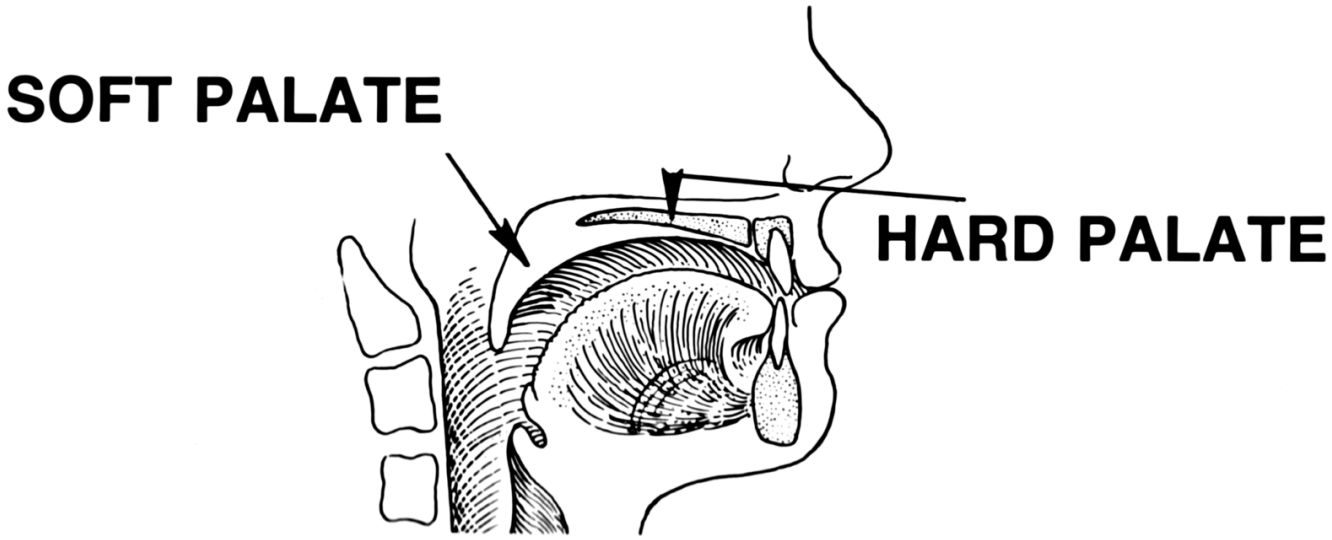


Figure 13.3.6: This cross-section of a human shows the hard palate and soft palate that all cynodonts and their relatives (including mammals and humans) have. The hard palate is partially made of bone, and the soft palate is not. The hard palate can be found in the fossil record because it can be fossilized 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.

True Mammals



Figure 13.3.7: 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 13.3.8: 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.

Amniote Branch Two: Sauropsids (Anapsids and Diapsids)

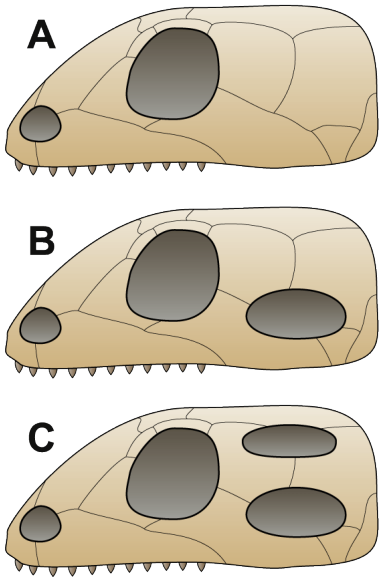


Figure 13.3.9: Major amniote skull types: A = anapsid, B = synapsid, C = diapsid. (Image by Preto(m) via Wikimedia.)

The **sauropsids**, the branch that would eventually evolve into reptiles, dinosaurs, and birds, are broken down into two major groups based on specific features of the skull: 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.

Anapsids

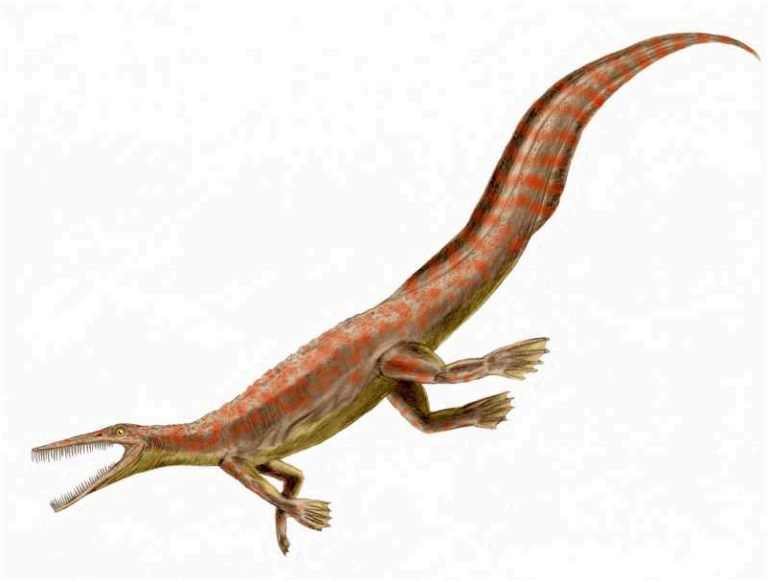


Figure 13.3.10: Mesosaurus reconstruction drawn (by Nobu Tamura via Wikimedia.)

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.

Diapsids: Lepidosaurs

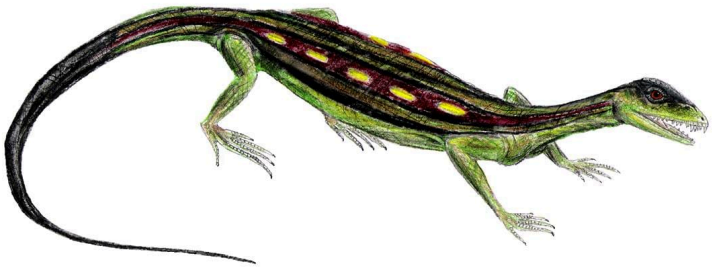


Figure 13.3.11: Araeoscelis reconstruction (by Smokeybjb via Wikimedia.)

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 13.3.12: 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.

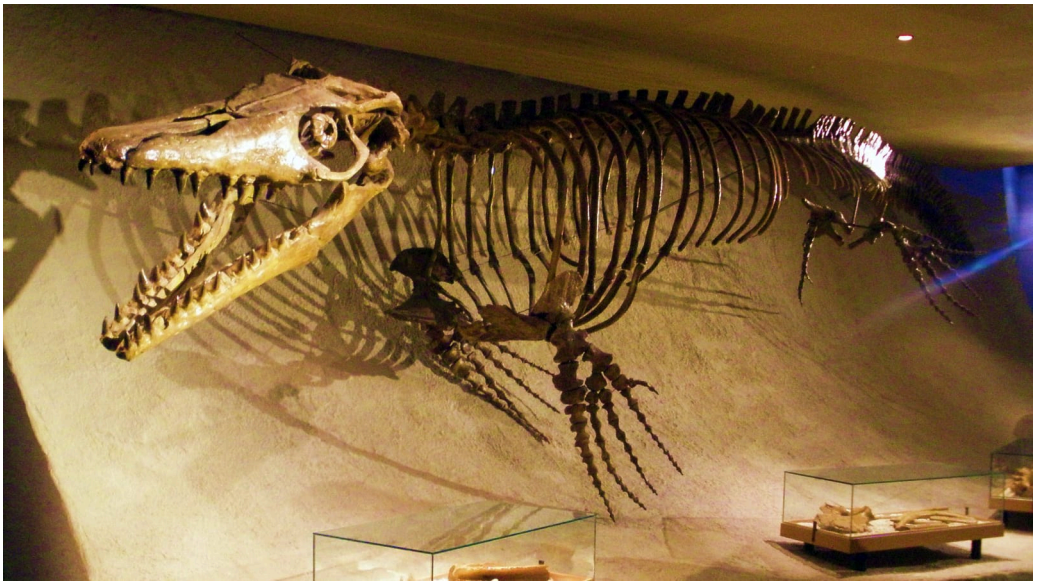


Figure 13.3.13: Mosasaurus skeleton displayed at the Natural History Museum of Maastricht, Netherlands. (Photograph by Ghedoghedo via Wikimedia.)

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.



Figure 13.3.14: Protorosaurus reconstruction (by Nobu Tamura via Wikimedia.)

Diapsids: Archosaurs

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.

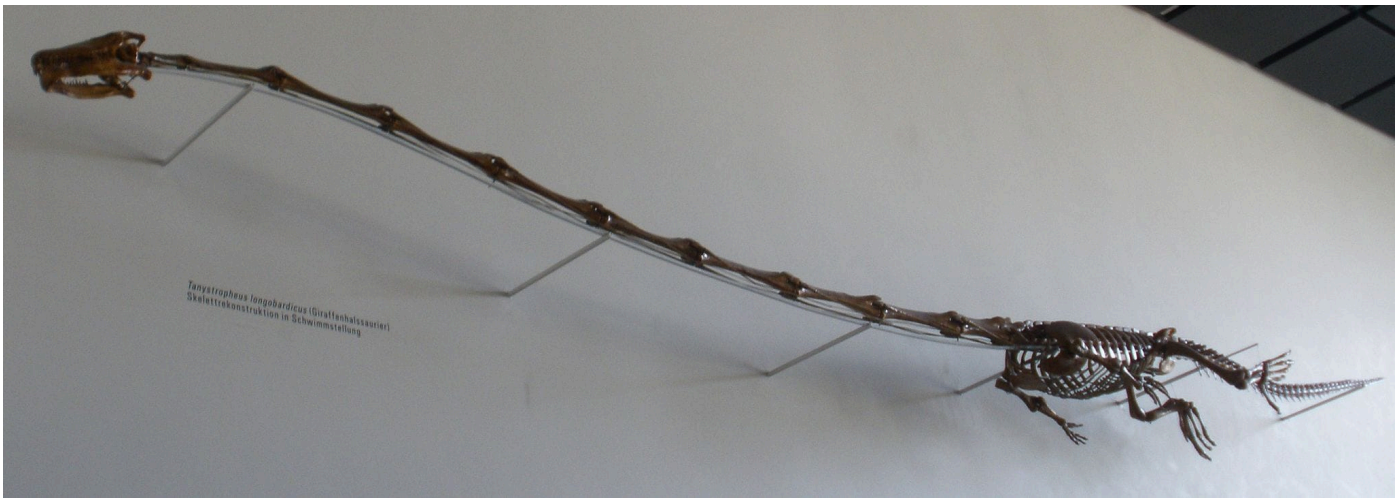


Figure 13.3.15: Tanystropheus, and early archosaur, had a long neck like sauropod dinosaurs which came much later. (Photo taken at the Paleontology Museum of Zurich by Ghedoghedo via Wikimedia.)

In the Triassic, in the open landscape and ecospace following the Great Dying (AKA the end Permian-Triassic Mass Extinction), the archosaurs diversified greatly. Many different lines of archosauromorphs that originated soon went extinct at the next extinction, at the end of the Triassic Period. 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.



Figure 13.3.16: 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.

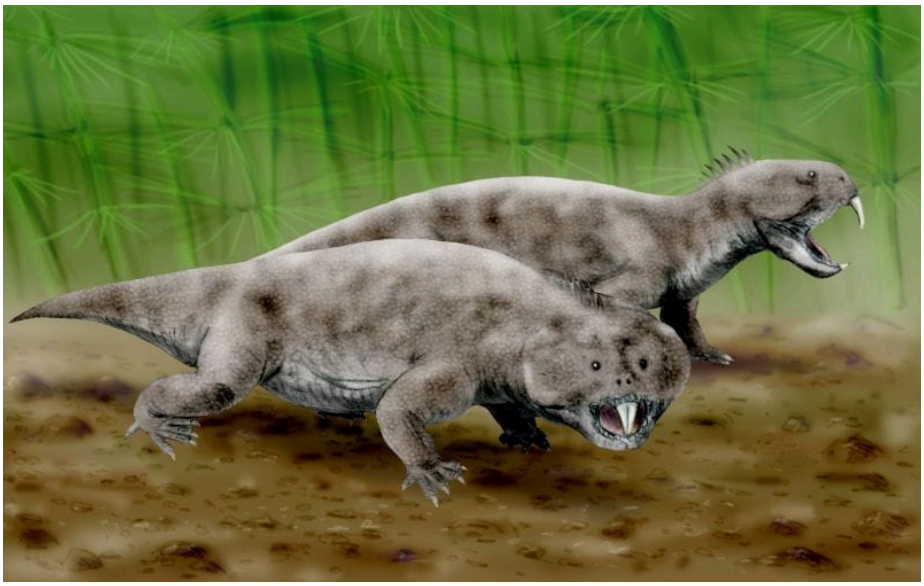


Figure \({13{.}3{.}\eatSpaces{17} \endSpaces}\: Hyperodapedon, a common rhynchosaur. (Drawing by Nobu Tamura via Wikimedia.)

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.

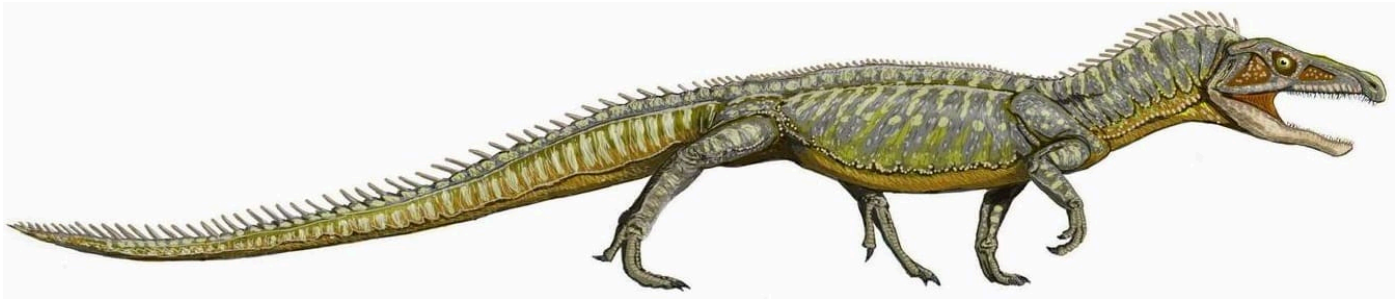


Figure 13.3.18:2Proterosuchus reconstruction (by ДиБгд via Wikimedia.)

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.

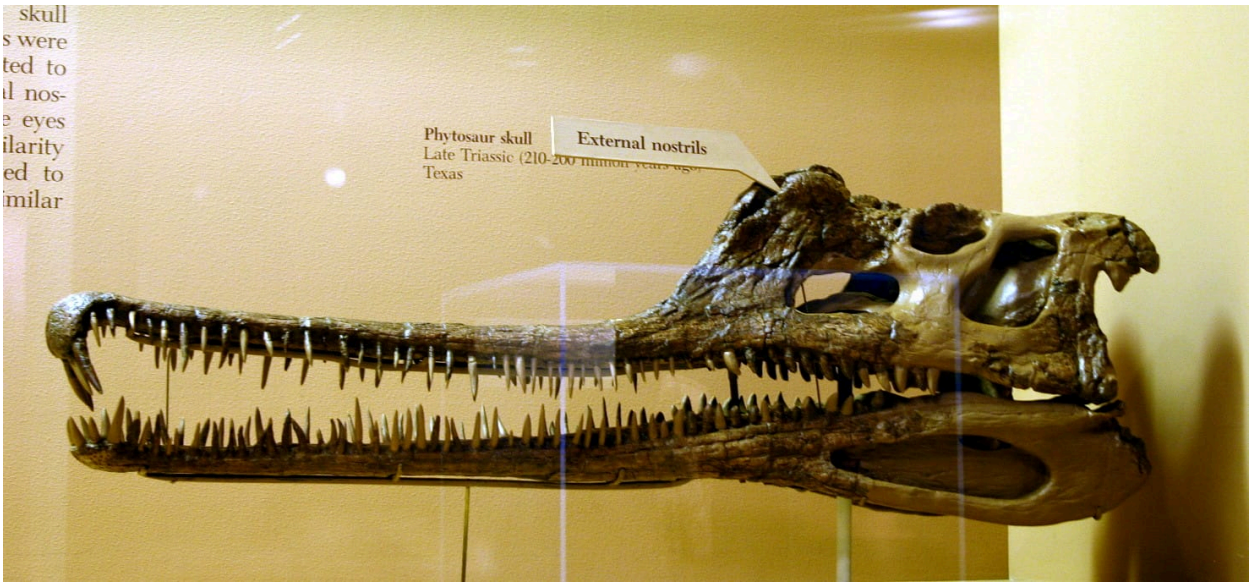


Figure 13.3.19: The skull of phytosaur. Note the nostril above the eye on the top of the skull. (Photo by Ryan Somma.)

Diapsids: Archosaurs - Pseudosuchia

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.

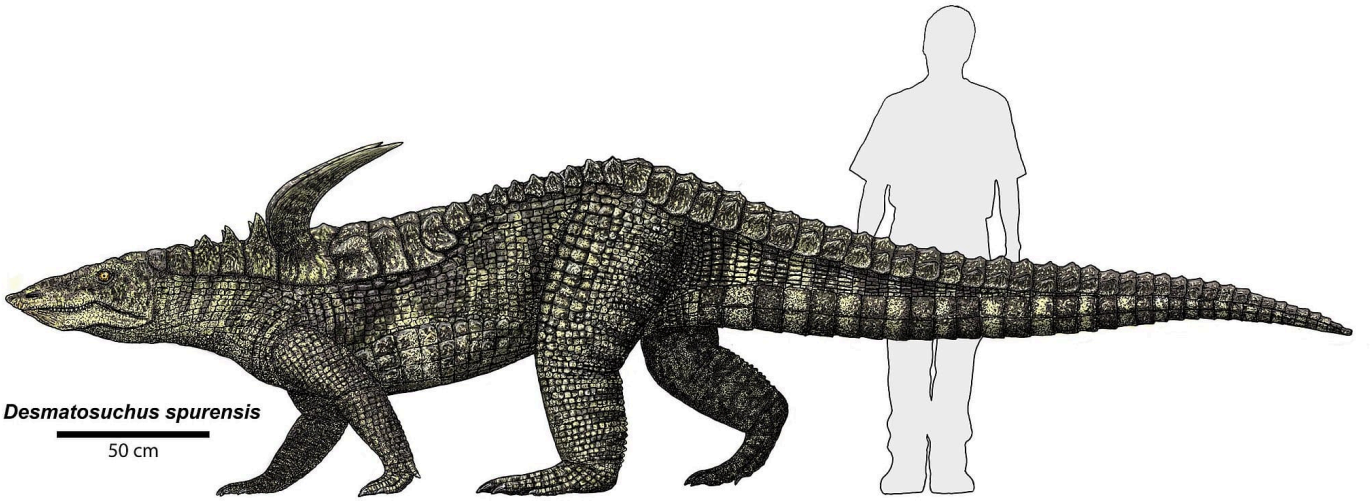


Figure 13.3.20: Desmatosuchus, an aetosaur with thick armor and spikes and a pig-like nose.

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.

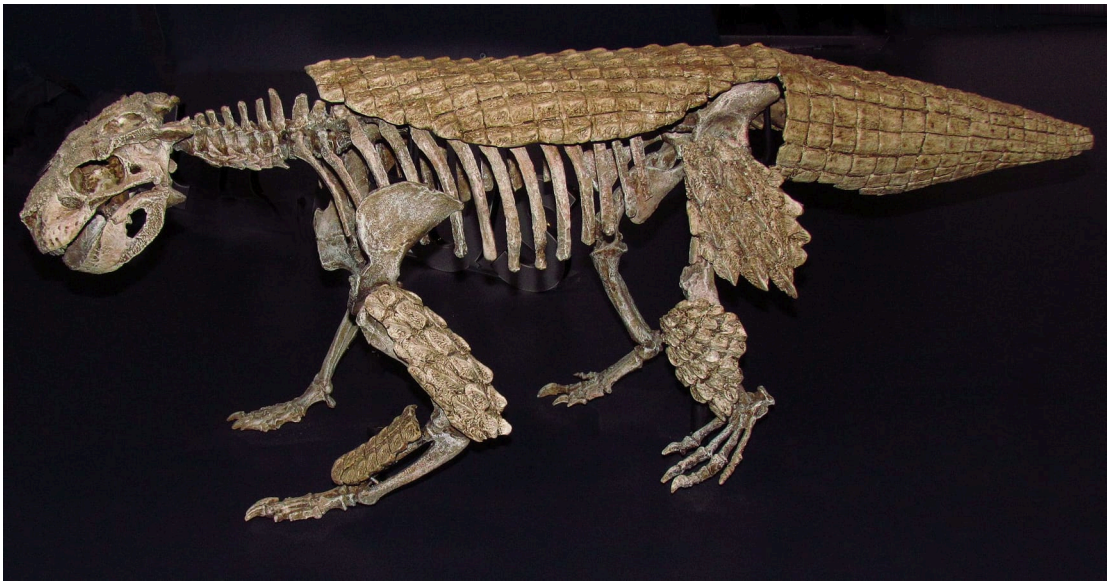


Figure 13.3.21: Simosuchus, an example of a strange early croc. (Photo of a cast found in the Royal Ontario Museum by D. Gordon E. Robertson via Wikimedia.)

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.

Diapsids: Archosaurs - Avemetatarsalia

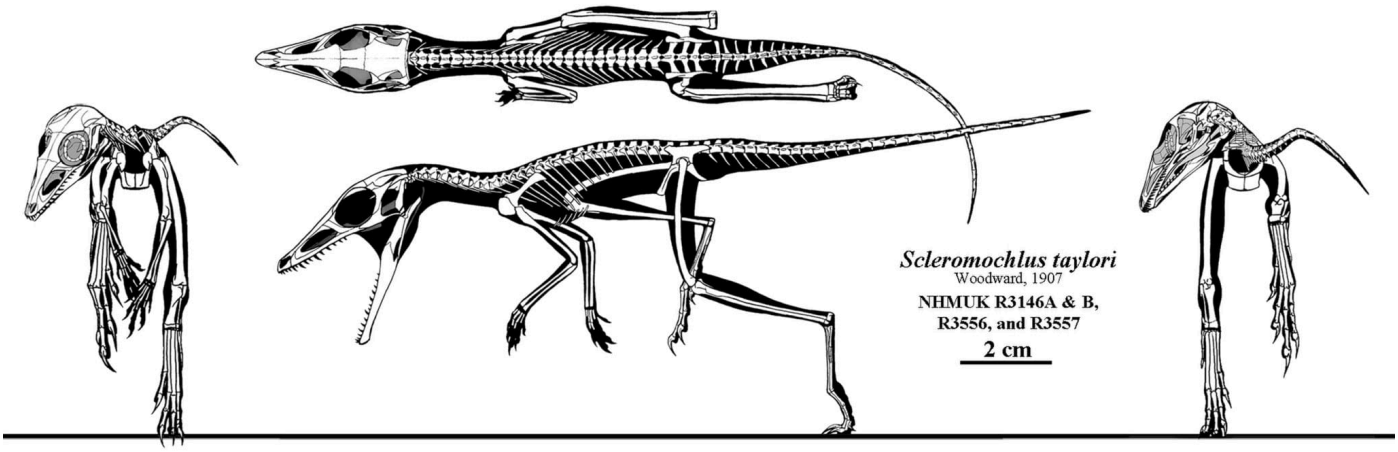


Figure 13.3.22: Scleromochlus, a possible early ancestor of pterosaurs. (Drawing by Jaime A. Headden via Wikimedia.)

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 13.3.23: Photograph of a well-preserved Rhamphorhynchus fossil from the Natural History Museum in Brussels (by MØtty 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.

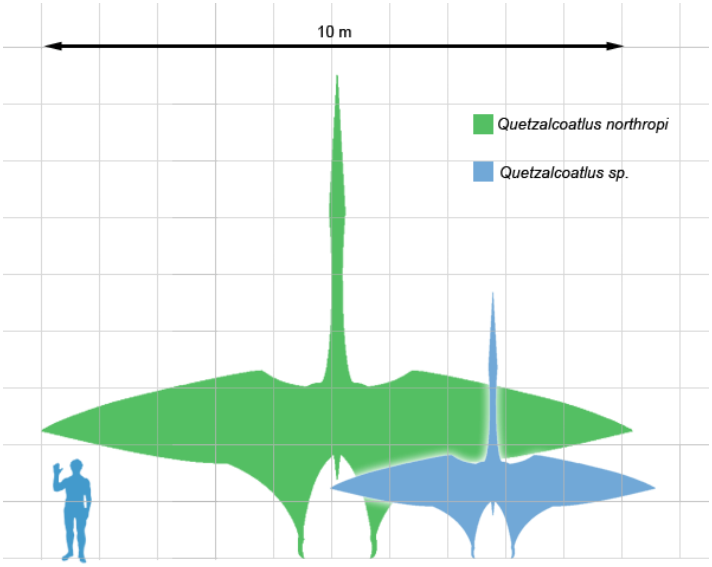


Figure 13.3.24: Size comparison between two Quetzalcoatlus species and a human. (Illustration by Matt Martyniuk (Dinoguy2), Mark Witton and Darren Naish via Wikimedia.)

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!

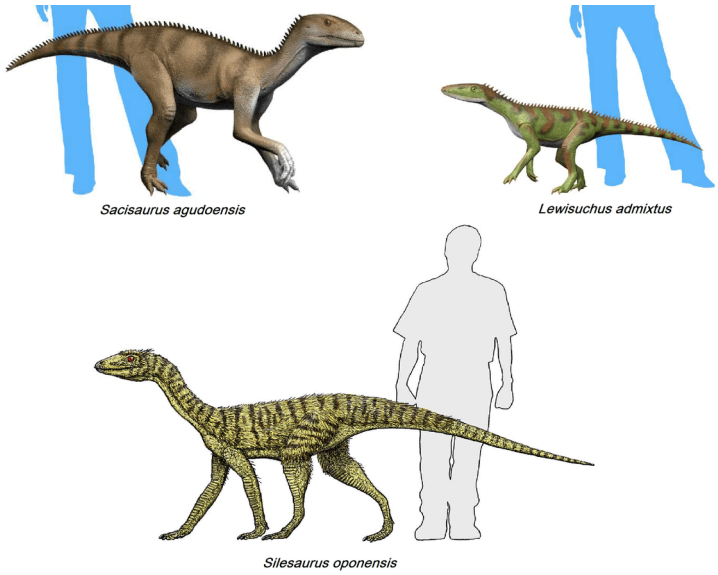


Figure 13.3.25: Different sileasaurs compared to humans. The interpretation of them being quadrupedal verses bipedal is debated. (Drawings by Nobu Tamura and US National Park Service via Wikimedia.)

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.

Unknown Diapsids: Ichthyosaurs



Figure 13.3.26: 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 13.3.27: 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.

Unknown Sauropsids - Turtles and Sauropterygia

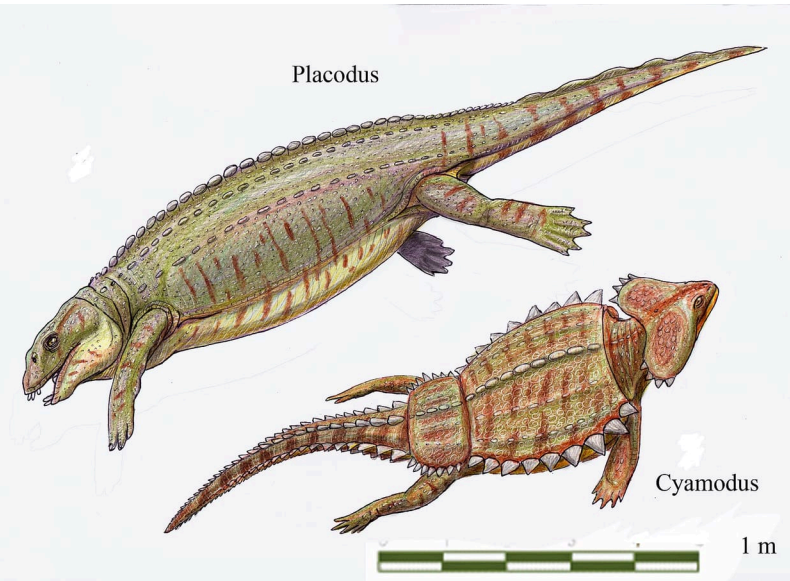


Figure 13.3.28: Examples of placodonts, drawn (by DiBgd via Wikimedia.)

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.

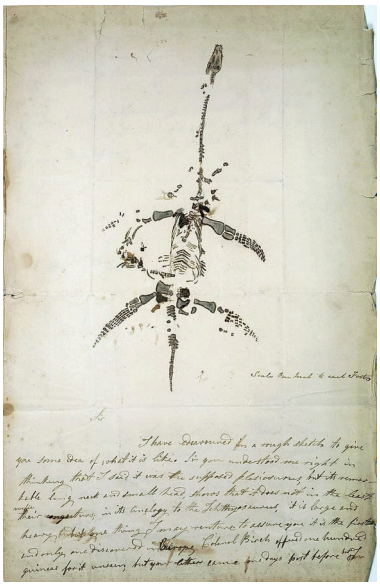


Figure 13.3.29: Drawing by Mary Anning of Plesiosaurus. Anning did early work on plesiosaurs, ichthyosaurs, marine reptiles, invertebrates, pterosaurs, and was the first person in the world to recognize coprolites. However, due to her gender, she was not taken seriously as a scientist. She not only was 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.

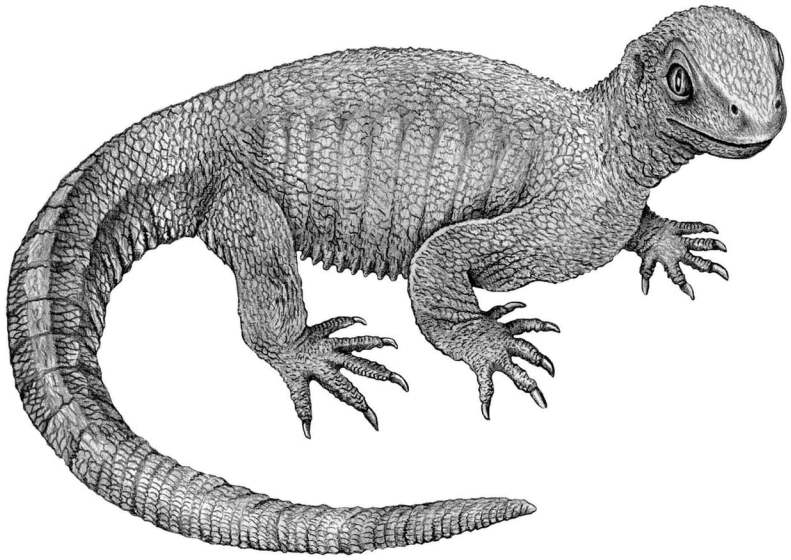


Figure 13.3.30: Pappochelys, possibly the first turtle. (Drawing by Rainer Schoch via Wikimedia.)

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.

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.”

Key Terms

- **amniotes** - a group of animals that evolved from amphibians that includes all reptiles (broadly defined), birds, and mammals
- **anapsids** - the group of sauropsids with reptile skulls with no openings in the rear of the skullavemetatarsalia - bird-line archosaurs
- **archosaurs** - a branch of diapsids that includes the last common ancestor of birds and crocodilians
- **diapsids** - the group of sauropsids with two openings in the rear of the skull

- **lepidosaurs** - a branch of diapsids known for overlapping scales
- **pseudosuchian** - crocodile-line archosaurs
- **sauropsids** - the branch of amniotes that would eventually evolve into reptiles, dinosaurs, and birds
- **synapsids** - the branch of amniotes that would eventually lead to mammals and humans, but started out with a very reptilian-like body plan

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13.4: Mesozoic Evolution - Dinosaurs

Introduction

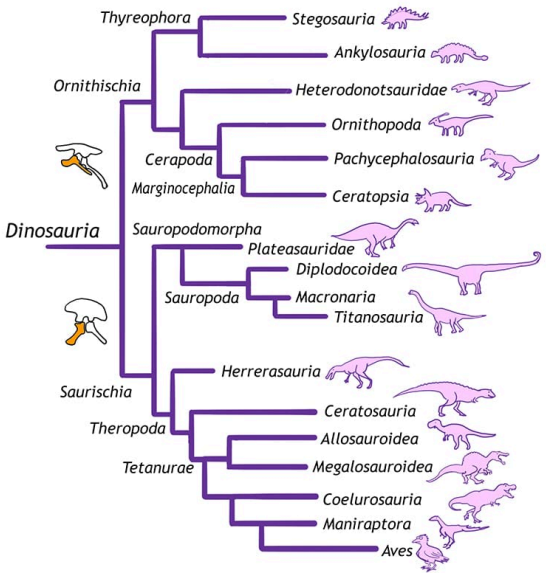


Figure 13.4.1: The major branches of the dinosaur family tree. (Image by Callan Bentley.)

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 section, we will talk about the seven major groups that include the vast majority of known dinosaurs and how those groups relate to one another.

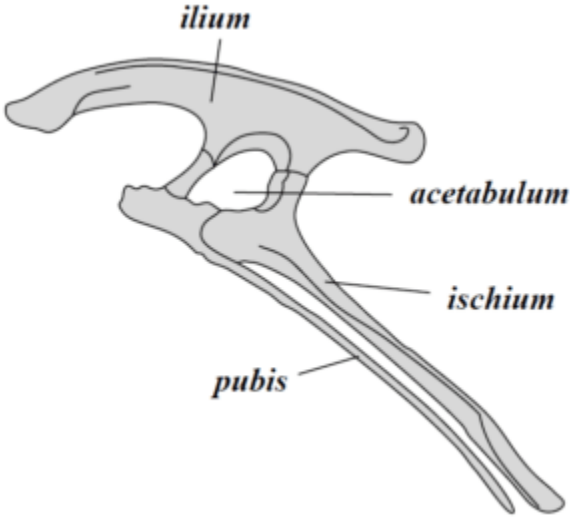


Figure 13.4.2: Ornithischian pelvis with a closed structure. (Image by AdmiralHood on Wikimedia.)

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.

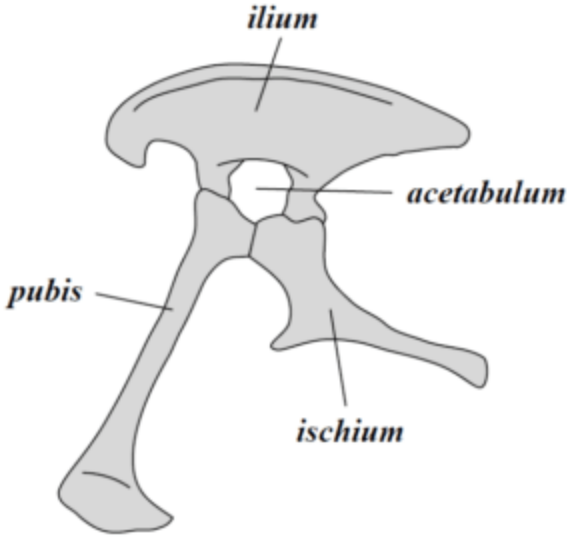


Figure 13.4.3: Saurischian pelvis with an open structure. (Image by AdmiralHood on Wikimedia.)

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.

Saurischian dinosaurs have two major subdivisions: theropods and sauropodomorphs.

Saurischian Dinosaurs: Theropods

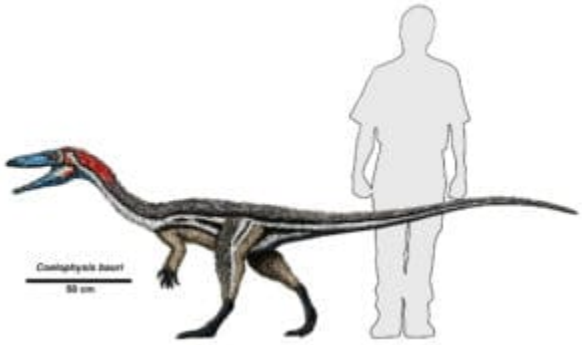


Figure 13.4.4: Drawing of Coelophysis compared to an adult human. (Image from Petrified Forest National Park.)

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 13.4.5: 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 13.4.6: 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 13.4.7: 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.

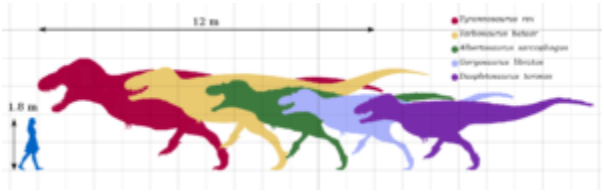


Figure 13.4.8: Largest Tyrannosaurs compared to a human. (Drawn by PaleoGeekSquared on Wikimedia.)

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 13.4.9: 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 13.4.10: 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 13.4.11: 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.

YouTube Video

This interactive element is not available in this format. Visit the link or scan the QR code to access it online.

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Video by PBS that breaks down how dinosaurs are related to birds.

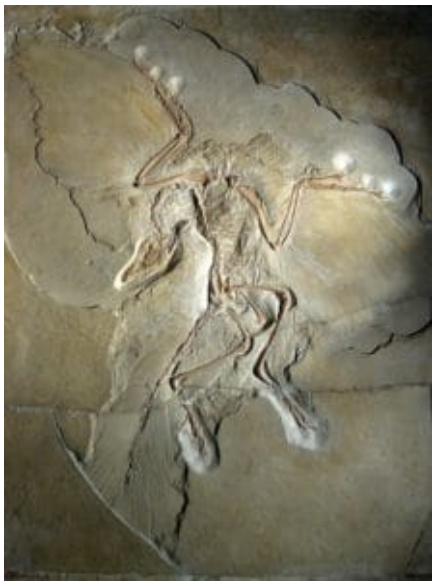


Figure 13.4.12: The famous Berlin Specimen of Archaeopteryx. (Picture by H. Raab (User: Vesta) via Wikimedia.)

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.

Saurischian Dinosaurs: Sauropods



Figure 13.4.13: 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 13.4.14: 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 13.4.15: 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 13.4.16: 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.

Ornithischian Dinosaurs



Figure 13.4.17: 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.

There are two major subgroups within ornithischia: thyreophora and neornithischia.

Ornithischian Dinosaurs: Thyreophora - Stegosaurs



Figure 13.4.18: 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)

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.

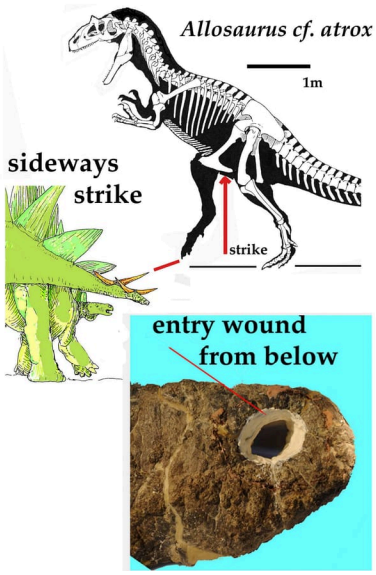


Figure 13.4.19: Cartoon and picture of a Stegosaurus/Allosaurus interaction. (Image care of Robert Bakker.)

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.

Ornithischian Dinosaurs: Thyreophora - Ankylosaurs



Figure 13.4.20: 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.

Ornithischian Dinosaurs: Neornithischia - Ornithopods

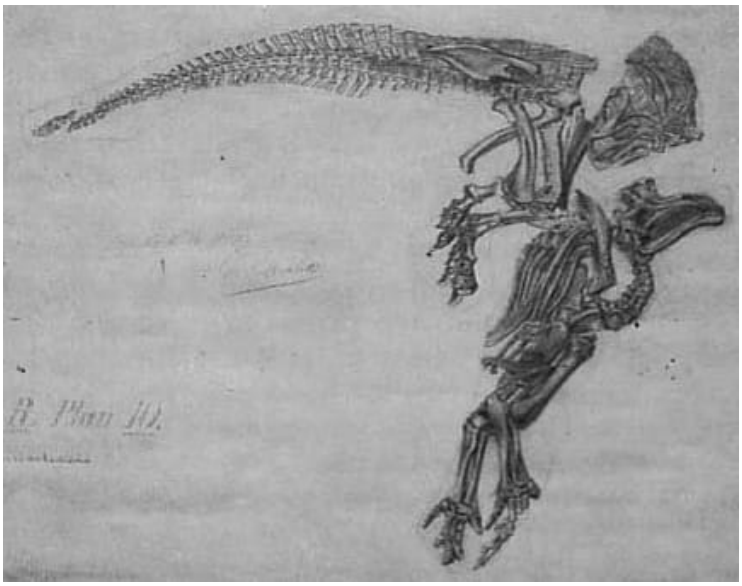


Figure 13.4.21: Drawing of an Iguanodon skeleton by G. Lavette as it was found in 1882, nearly articulated.

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.

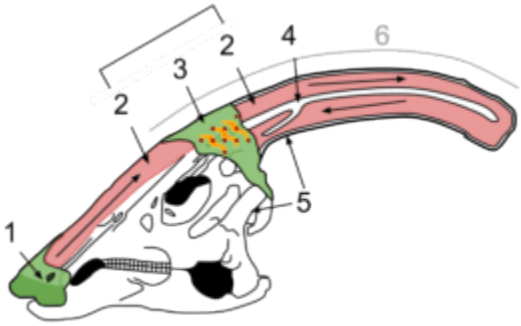


Figure 13.4.22: Parasaurolophus skull with a view into nasal cavity of the cranial crest. It is hypothesized that the Parasaurolophus pushed air through its long cranial crest to make low register sounds that could be heard for miles. 1. Nostril 2. Sinus cavity 3. Skin 4. Intra Cranial bone 5. Skull 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.

Ornithischian Dinosaurs: Neornithischia - Ceratopsians



Figure 13.4.23: 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 13.4.24: 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.

Ornithischian Dinosaurs: Neornithischia - Pachycephalosaurs

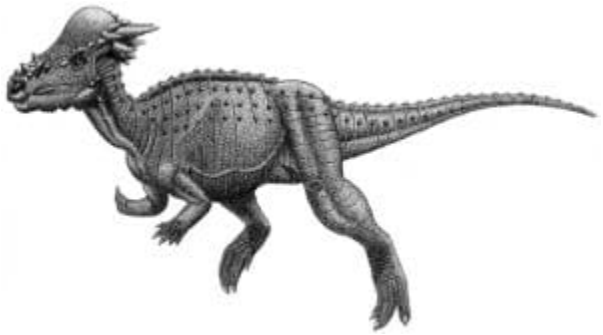


Figure 13.4.25: Sketch of a Pachycepholosaur from the Hell Creek Formation of South Dakota (by Jordan Mallon via Wikimedia.)

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.

Key Terms

- **ornithischians** - herbivorous dinosaurs characterized by a "bird-hipped" pelvic structure in which the pubis is rotated backward
- **sauropods** - large, long-necked, quadrupedal, "lizard-hipped" herbivorous dinosaurs known for their columnar limbs, elongated cervical vertebrae, and massive body size
- **theropods** - bipedal "lizard-hipped" dinosaurs characterized by hollow bones, three-toed limbs

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13.5: Mesozoic Mass Extinctions

Introduction

There are three mass extinctions associated with the Mesozoic Era. The first occurred 252 million years ago. It marked the start of the Mesozoic Era and was addressed in the previous chapter. The second occurred at the end of the Triassic Period, and the third occurred at the end of the Mesozoic Era. These two are explained below.

End of Triassic Extinction

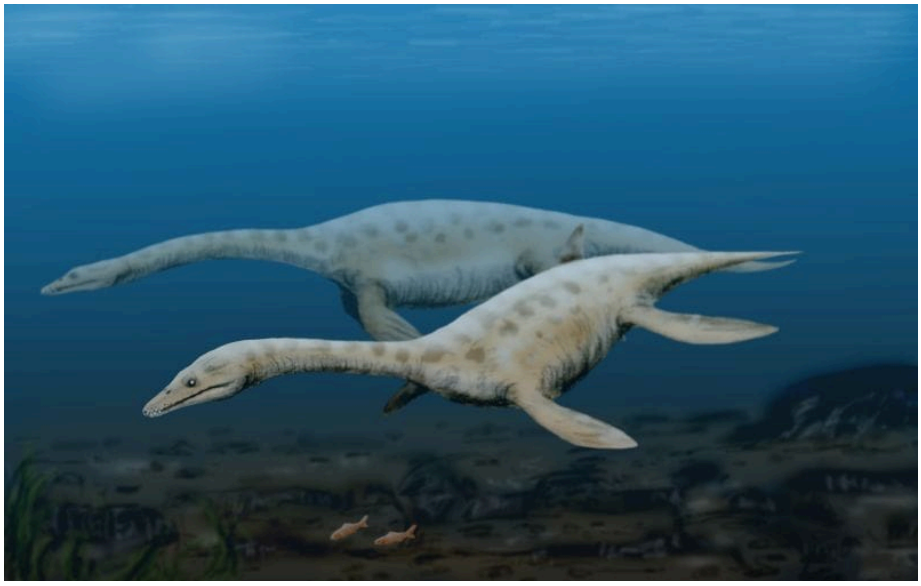


Figure 13.5.1: A drawing of the early plesiosaur Agustasaurus from the Triassic of Nevada.

At the end of the Triassic Period, 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 13.5.2: Reconstruction of the small (<5″) Megazostrodon, one of the first animals considered to be a true mammal.

Effects on Mammals

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.

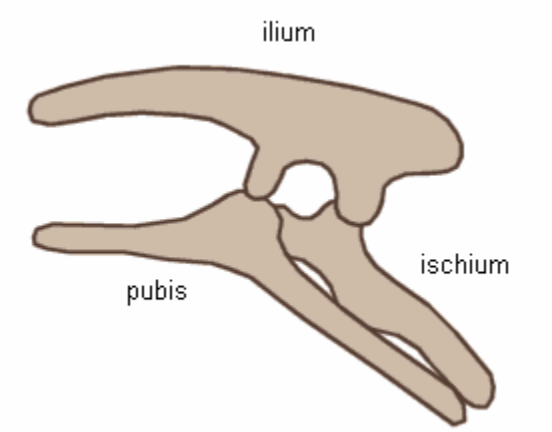


Figure 13.5.3: Closed structure of a ornithischian hip, which is similar to a birds.

Effects on Dinosaurs

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.

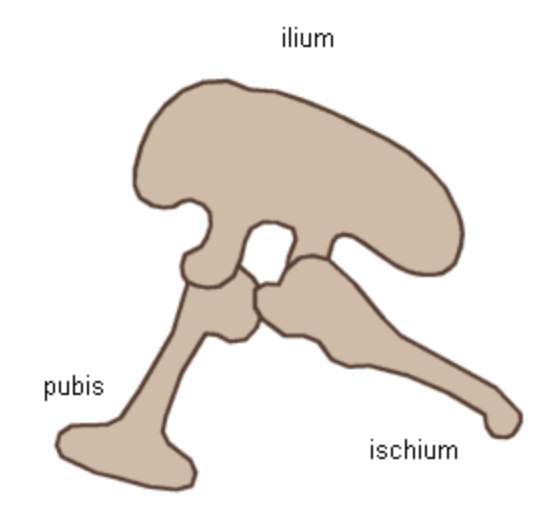


Figure 13.5.4: Open structure of a saurischian hip, which is similar to a lizards.

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 13.5.5: 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 13.5.6: Iconic “Berlin specimen” Archaeopteryx lithographica fossil from Germany.

Effects on Birds

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.



Figure 13.5.7: Reconstructed skeleton of Argentinosaurus, from Naturmuseum Senckenberg in Germany.

Effects on Plants

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-Pg- End of Mesozoic Mass Extinction

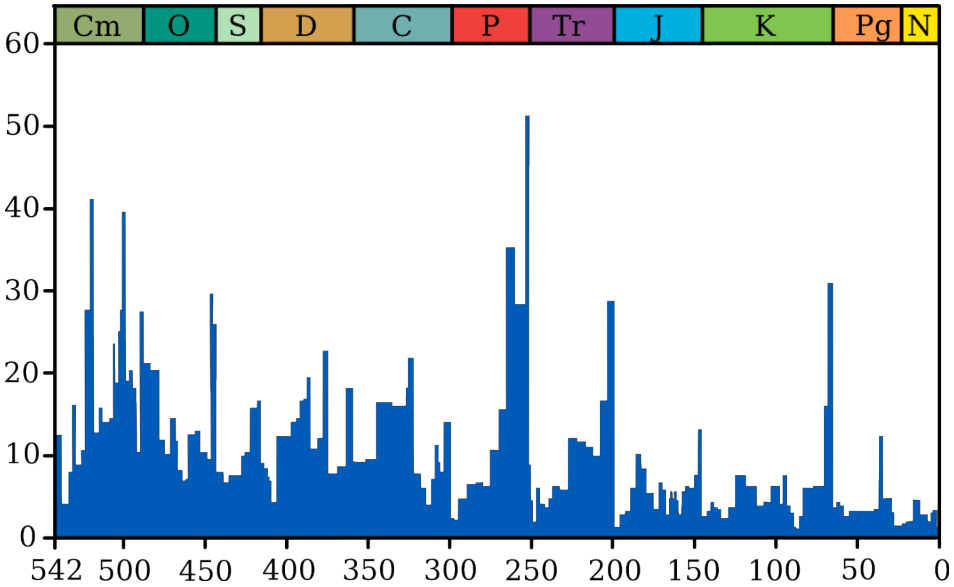


Figure 13.5.8: Graph of the rate of extinctions. Note the large spike at the end of the Cretaceous (labeled as K).

Similar to the end of the Paleozoic era, the Cretaceous Period and Mesozoic Era ended 66 million years ago with the **K–Pg mass extinction** event, also known as the K–T extinction. This event marks the end of the Cretaceous Period (K) and the beginning of the first period of the Cenozoic Period, the Paleogene (Pg).

Cause of Mass Extinction

Extraterrestrial Impact

This extinction event was likely caused by a large 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 13.5.9: 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.

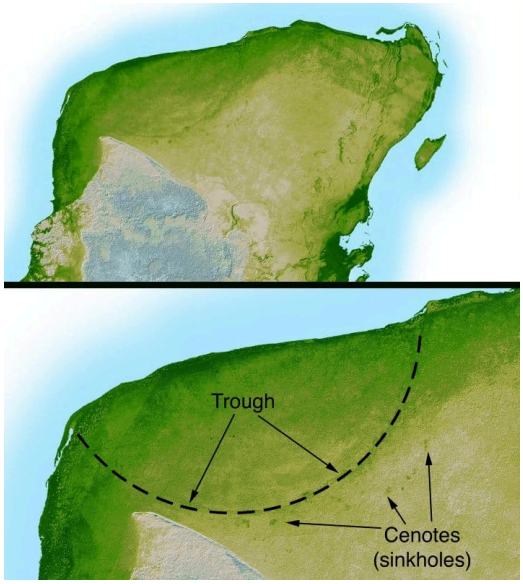
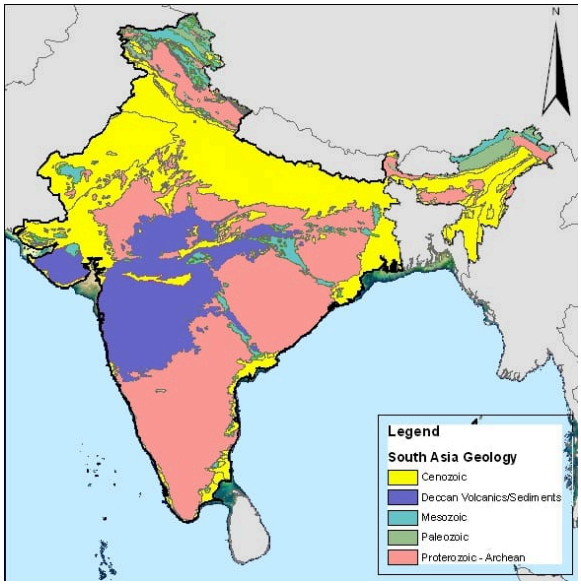


Figure 13.5.10: The land expression of the Chicxulub crater. The other side of the crater is within the Gulf of México.

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.

Alternate Explanations

With all of this information, it seems like the case would be closed. However, there are other events at this time which could have partially aided the demise of so many organisms. For example, sea levels are known to be slowly decreasing at the time of the K-Pg event, which is tied to marine extinctions, 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.



Figure

:

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

It seems that by the end of the Mesozoic, life on Earth was in a fragile state, and at the same time, a meteorite larger than Mount Everest was hurtling towards Earth.

Video: The Day the Dinosaurs Died – Minute by Minute

The events that led to the extinction of the dinosaurs are dramatically summarized in the video "The Day the Dinosaurs Died – Minute by Minute", by Kurzgesagt – In a Nutshell.

YouTube Video

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Video: How the Dinosaurs Actually Died

The argument for volcanic activity causing the extinction of the dinosaurs is well explained in the video "How the Dinosaurs Actually Died", by Kurzgesagt – In a Nutshell.

YouTube Video

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Key Terms

- **Chicxulub Crater** - a ~180-kilometer-wide impact structure buried beneath the Yucatán Peninsula in Mexico formed about 66 million years ago by the collision of a large asteroid, meteoroid or comet
- **Deccan Traps** - one of Earth's largest continental flood basalt provinces, located primarily in western India and erupted in multiple phases around 66 million years ago
- **K–Pg mass extinction** - also referred to as the K-T Extinction, this extinction occurred at 66 million years and eliminated roughly 75% of Earth's species

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13.6: Key Points

1. The Mesozoic Era began 252 million years ago after the Permian-Triassic mass extinction event.
2. Pangaea came together and broke up during the Triassic.
3. For most of the Mesozoic, the western margin of North America was tectonically active, while the eastern margin remained passive. Mountain building happened during four orgonenic events, building up the North American Cordilleran, which includes mountain ranges like the Rockies.
4. In the Cretaceous, sea levels rose, and North America was separated into two landmasses by a shallow seaway, where sediment accumulated.
5. As Pangea assembled and the climate dried out, amphibians evolved into a group called amniotes, some of which were able to lay waterproof eggs on land.
6. Amniotes were broken into the synapsids which would eventually become mammals and the sauropsids containing modern reptiles, birds, and dinosaurs.
7. Dinosaurs first appeared and became extinct in the Mesozoic Era. They are divided into groups based on their hips as either "bird-hipped" in which the pubis and the ischium are close together or "lizard-hipped" in which the pubis and the ischium are far apart.
8. The end of the Triassic Period was marked by a mass extinction perhaps caused by the Central Atlantic Magmatic Province flood basalt that helped spur the evolution of survivors like mammals, pterosaurs (flying reptiles), ichthyosaurs/plesiosaurs/mosasaurs (marine reptiles), and dinosaurs
9. The Mesozoic Era ended 66 million years ago when 75% of species died out during the K-Pg mass extinction event caused, at least in part, by the impact of a large asteroid, meteoroid, or comet and the prolonged and vigorous volcanic activity of the Deccan Traps.

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14: PHANEROZOIC EON - CENOZOIC ERA

Introduction

Here is the the simplified geologic time scale used in the Geologic Time chapter of this book. The Cenozoic Era is highlighted at the top.

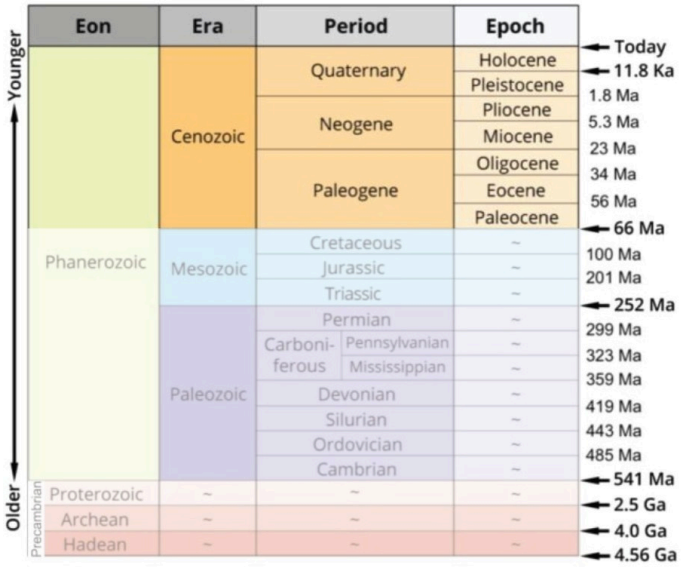


Figure 14.1: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

The Cenozoic, meaning “new life,” is known as the Age of Mammals because it was during this era that mammals became a dominant group, including human ancestors. Birds, too, flourished through adaptive radiation after the demise of the dinosaurs. Most of the Cenozoic Era has been relatively warm, with the main exception being the ice age that began approximately 2.558 million years ago and continues to this day. The East Coast of North America remained a passive continental margin, whereas in the West, tectonic activity led to volcanism and, eventually, the long-standing subduction zone evolved into a transform boundary.

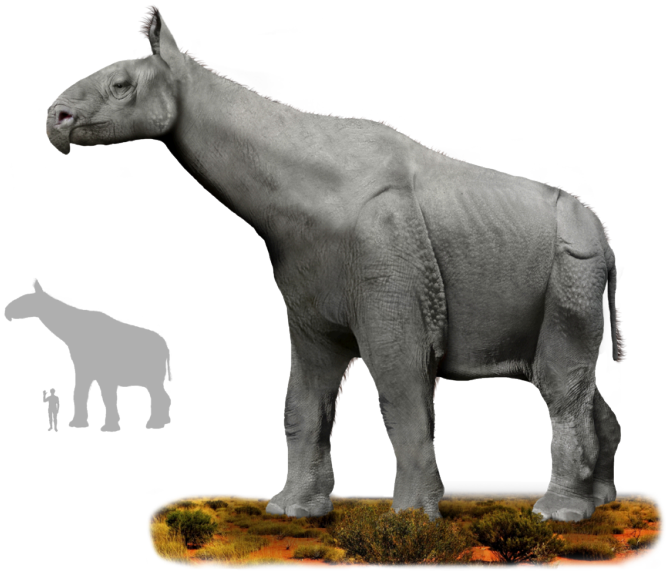


Figure 14.2: Paraceratherium, seen in this reconstruction, was a massive (15-20 ton, 15 foot tall) ancestor of rhinos.

14.1: Paleogeography and Tectonics of the Cenozoic

14.2: An In-Depth Look at the San Andreas Fault and the Basin and Range

14.3: Cenozoic Biology

14.4: Key Points

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14.1: Paleogeography and Tectonics of the Cenozoic

During the Cenozoic, the plates moved the continents into more familiar positions, with the most significant change being the closure of the Tethys Sea through continental collisions that built mountains, such as the Alps, the Zagros, and the Himalayas, the latter of which began about 57 million years ago and continues to this day. This closed off the Atlantic from the Indian Ocean, causing the rearrangement of global ocean circulation patterns.

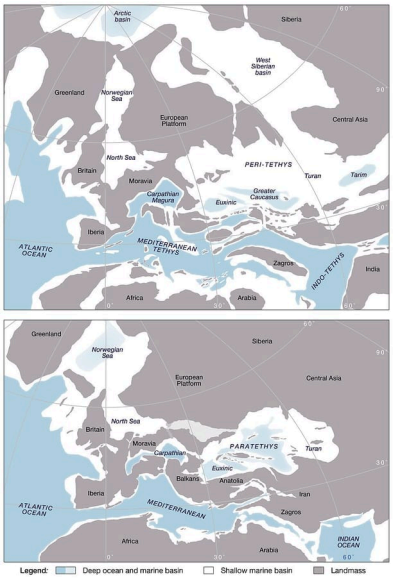


Figure 14.1.1: The closure of the Tethys Sea during the Cenozoic. Dan.palcu, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons.

Perhaps the most significant tectonic event to occur in the Cenozoic regarding North America was the transformation of the west coast of California from a convergent boundary subduction zone to a transform boundary. Subduction along the western margin of the United States had occurred throughout the Mesozoic and into the Cenozoic. The video below is an animation of the last 38 million years of movement in western North America. Note that after the mid-ocean ridge is subducted, the convergent boundary becomes a transform boundary, and continental rifting begins.

YouTube Video

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Near the end of the Mesozoic and into the Cenozoic, the **Laramide Orogeny** built mountains hundreds of miles east of the active western margin of the North American continent. This is attributed to "flat-slab" subduction and the oceanic Farallon Plate being young and close to its spreading center.

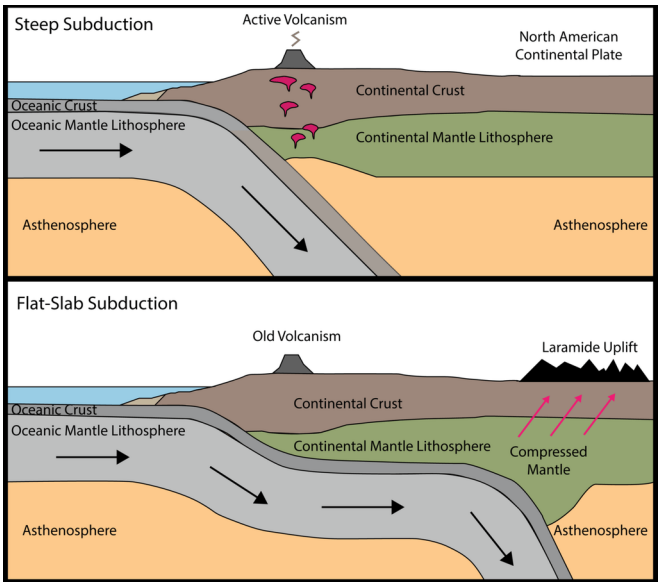


Figure 14.1.2: Shallow subduction during the Laramide Orogeny; public domain (USGS)

Flat-slab subduction happened when the rate of subduction was faster than the spreading of the seafloor between the Farallon and the Pacific plates. Eventually, because the rate of seafloor spreading couldn't keep up with the rate of subduction, the mid-ocean ridge itself was subducted, causing the relative motion along the plate boundary to change from convergent to transform. The transform boundary became 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 has shifted position over the last 18 million years, twisting the mountains around Los Angeles; new faults in the southeastern California deserts may become future San Andreas-style faults.

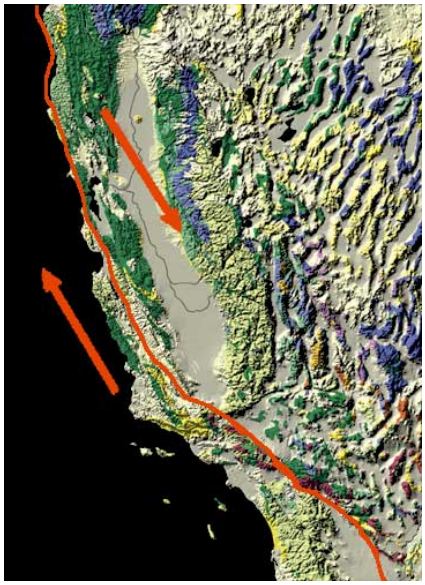


Figure 14.1.3: Map of the San Andreas fault, showing relative motion.

As subduction ceased, the nearly horizontal Farallon slab of oceanic crust began to sink into the mantle, allowing asthenosphere material to rise around it and magma to rise to Earth's surface. This volcanic event is known as the Oligocene ignimbrite flare-up, one of the most significant periods of volcanism in history, including the largest single confirmed eruption, the 5,000 cubic-kilometer Fish Canyon Tuff.

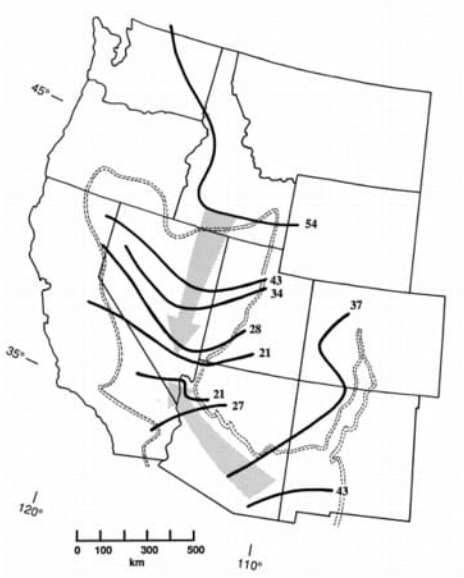


Figure 14.1.4: From Humphreys, 1995. Volcanic activity happened with the double-dashed line. The numbers associated with the solid black lines indicate the timing and location of volcanic activity.

Key Terms

- **Laramide Orogeny** - a major mountain-building event that affected western North America from roughly 80 to 40 million years ago
- **San Andreas Fault System** - a series of faults that mark the transform plate boundary that accommodates horizontal motion between the Pacific Plate and the North American Plate

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14.2: An In-Depth Look at the San Andreas Fault and the Basin and Range

Introduction

Two essential principles of physical geology are that ocean basins grow through seafloor spreading at mid-ocean ridges and close through subduction at some convergent boundaries. However, when these two processes combine, the situation becomes complex, as illustrated by the Cenozoic tectonic history of western North America. This is a rare scenario, and only a few examples of it can be clearly described in the geological record.

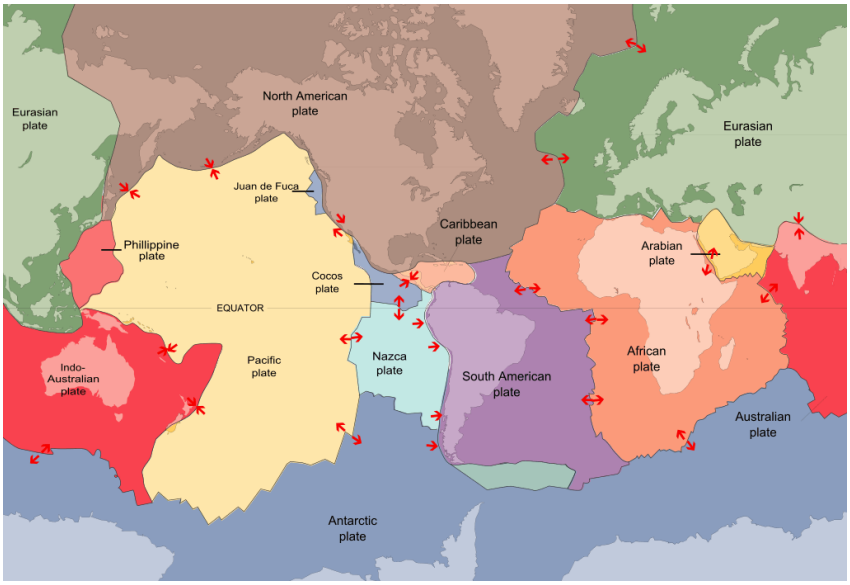


Figure 14.2.1: Plates and plate boundaries with a general sense of motion identified.

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. This happened because the rate of subduction between the North American and Farallon plates was faster than the spreading happening at the mid-ocean ridge between the Pacific and Farallon plates. In other words, seafloor spreading couldn't keep up with subduction, and the mid-ocean ridge was drawn closer and closer to the trench. This meant that the Farallon Plate was still young and hot when it reached the trench, making it buoyant and resistant to sinking. Think of it being more like a balloon filled with air than one filled with sand.

Eventually, the mid-ocean ridge itself was subducted, beginning around 30 million years ago (Ma). This event produced two major tectonic features: the San Andreas Fault System and the **Basin and Range Geologic Province**.

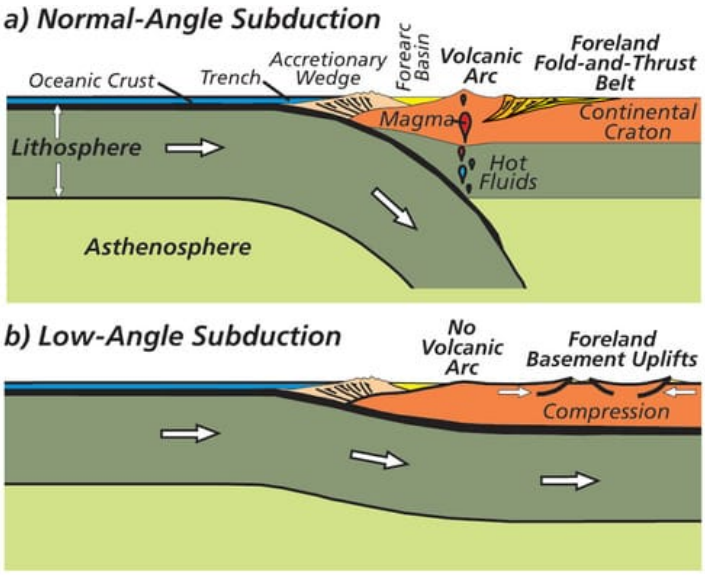


Figure 14.2.2: Normal (a) and flat-slab (b) subduction. (Picture by R. J. Lillie and Parks and Plates by the National Park Service.)

Basin and Range

The Basin and Range is a geomorphic province that extends across much of the southwestern United States and western Mexico. From the eastern margin of the Sierra Nevada Mountains in California, the Basin and Range extends across Nevada to the Wasatch Mountains in western Utah, characterized by a topography featuring a repeating sequence of roughly north-south-oriented, broad valleys (basins) separated by elongated mountain ranges (ranges). So persistent are the orientation and geometry of the basins and ranges that this region was described by Charles Dutton, an important 19th-century geologist, as an “army of caterpillars marching toward Mexico”.



Figure 14.2.3: Sketch of the Basin and Range Geologic Province (by Thatmaceguy via Wikimedia.)

It is probably the best modern example of a wide continental rift zone, a type of divergent boundary within a continent. Unlike typical (narrow) rifts, such as the East African Rift, deformation is distributed over a wide area, and normal faulting distributes tensional stresses over a broad region, creating a series of horsts and grabens.

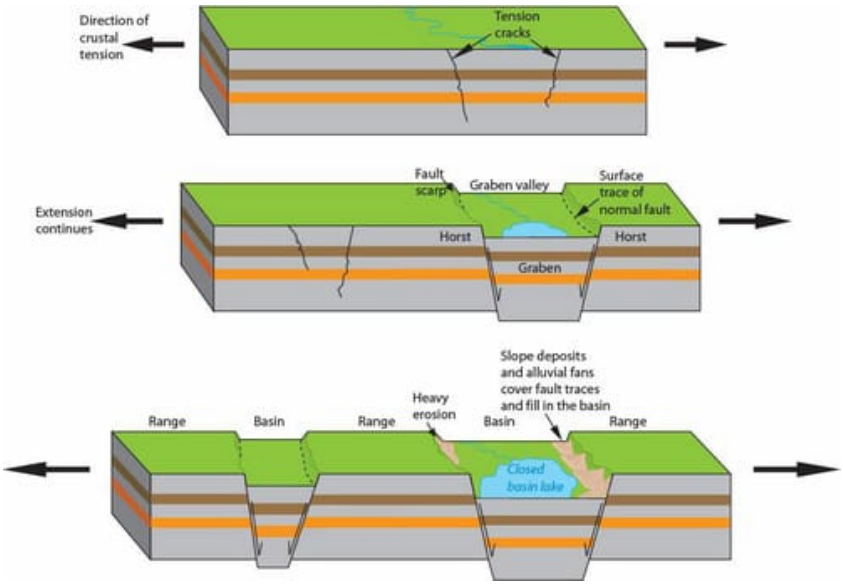


Figure 14.2.4: Generalized rift diagram. A wide rift has many valleys, a narrow rift focuses the extension in one main area. (Sketch by Trista L. Thornberry-Ehrlich, Colorado State University and the National Park Service.)

Regional continental rifting is the cause of the famous Basin and Range physiography: multiple north-south trending mountain ranges separated 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 which is bound by normal faults.



Figure 14.2.5. Photo of the central Basin in Range in Nevada. The dark, linear patches are mountain ranges (horsts) and the lighter areas in between are basins (grabens). (By NASA.)

This faulting has doubled the width of the state of Nevada over the last 50 million years. Most faults in the Basin and Range are either inactive today or exhibit low 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) margins of the Basin and Range are highly active.

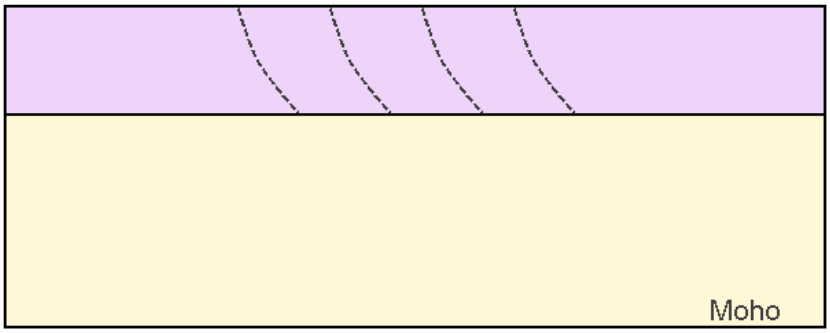


Figure 14.2.6: Animation of tilted-block faulting, one of the ways extension occurs in the Basin and Range (by Aturn4000 via Wikipedia.)

The faulting that started the Basin and Range is complex and not very uniform. In some places, extension dates back as far as 55 million years ago, whereas in others it remains active. The majority of the extension occurred around 20 to 10 million years ago. In general, extensional provinces in the northern Basin and Range began earlier, whereas rifting in the southern portion began later, around 30 million years ago. No single fault system has been active for more than 10 million years (Axen et al., 1993).

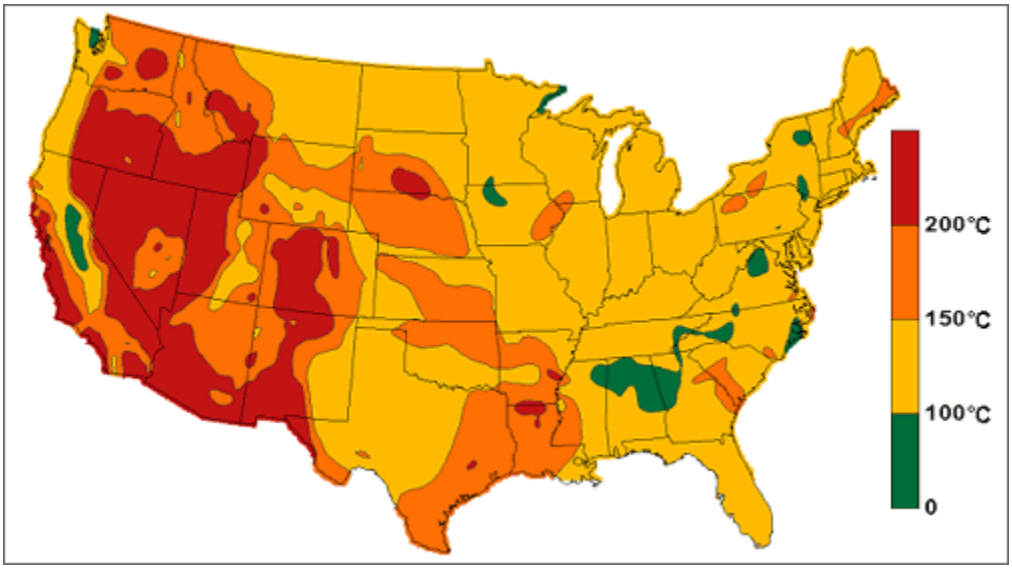


Figure 14.2.7: Geothermal resource map of the US (by the US Department of Energy.)

A significant side effect of Basin and Range extension is heat. Stretching thins the plate, relieving overburden pressure and triggering decompression melting. The stretched and thinned crust makes it easier for magma to reach the surface. This is evident in the extremely 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 14.2.8: 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.)

The upwelling asthenosphere accounts for the history of extensive volcanism across the Basin and Range. The origin of this remains debated, particularly given the absence of volcanism associated with the Laramide orogeny. However, once subduction stopped and a transform boundary was established, it seems that a 'hole' developed in the nearly flat-subducting slab, allowing material to rise from below.

This event, known as the Mid-Tertiary Ignimbrite Flare-up, happened in two waves: one in Washington and Idaho at about 54 Ma, and another in Arizona/New Mexico near the US border with Mexico at 43 Ma. Over time, the northern branch of volcanism shifted south, and the southern branch moved north, converging near Las Vegas at approximately 21 Million Years Ago (Humphries, 1995). This period is characterized by gigantic explosive eruptions, including some of the largest ever known on Earth, and is documented by extensive tuff deposits. Some of these eruptions are more than 5000 times larger than the May 1980 Mt. St. Helens Eruption.

Significance of the Basin and Range

The importance of the Basin and Range to geology is significant. Not only did the extension expose many sedimentary formations, fossils, and resources at or near Earth's surface, but it also contributed to the understanding of normal faulting. Two important and related features were first recognized through studies of Basin and Range extension: the detachment fault and the metamorphic core complex. Detachment faults, also known as low-angle normal faults, were highly controversial when they were first described. Most normal faults are steeply tilted, typically at an angle of around 60 degrees. However, the detachment faults of the Basin and Range are closer to horizontal than vertical. It's thought that these faults either started at high angles and rotated to lower angles with extension or formed at low angles.



Figure 14.2.9: 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.)

Movement and detachment faults also cause the formation and exposure of metamorphic core complexes. These are places where large amounts of extension (dozens of miles in some areas) have allowed deep crustal rock to bulge ductilely toward the surface. The detachment in these cases marks a boundary between highly metamorphosed basement rocks in the footwall, separated by more deformed tilted rocks of the hanging wall.

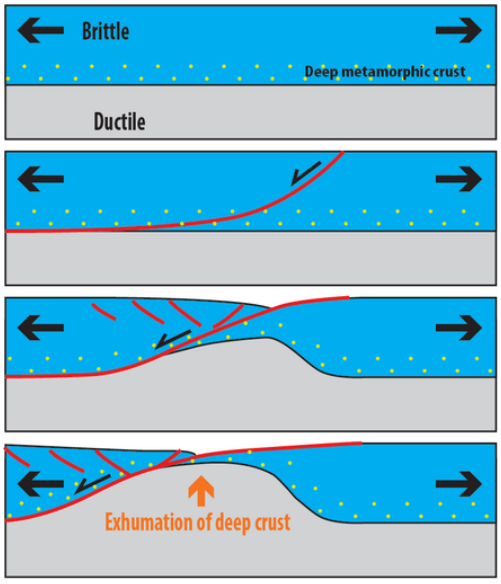


Figure 14.2.10: Sketch of metamorphic core complex that facilitates the upwelling of the mantle (by ManuRoquette via Wikipedia.).

Through progressive extension, the deep rocks bulge upward, and metamorphic rocks that form at depth, such as mylonite, exhibit signs of deformation in the shallow crust, including brecciation. One of the best-studied and understood detachment faults and metamorphic core complexes is in the Whipple Mountains, a mountain range of the Basin and Range, near the border of California and Nevada. Another example includes the Black Mountains near Death Valley.



Figure 14.2.11: Mylonite with garnet. Image from Creative Commons.

Cause of the Basin and Range Formation

The extension in the Basin and Range has two commonly cited causes that could be related or could have occurred simultaneously. One cause is extensional collapse. Beginning in the middle Mesozoic, the western edge of North America experienced prolonged compressional forces due to subduction and terrane accretion, driving the Nevadan, Sevier, and Laramide orogenies. Approximately 60 million years ago, the compression began to decrease, with a complete reduction in this force by approximately 30 million years ago. With the reduction of the forces that had squeezed and overthickened the crust, it began to "relax" and undergo gravitational collapse. As a result, this region of the crust spread out and thinned, contributing to at least some of the Basin and Range extension. One crucial piece of evidence supporting this model is the reactivation of dormant faults. Found all over the west as former thrust faults that turned into normal faults upon the transition to extensional tectonics.

The other cause of extension in the Basin and Range is due to the San Andreas Fault System, a complex yet vital component in understanding the modern plate boundary between the North American and Pacific plates.

San Andreas Fault System

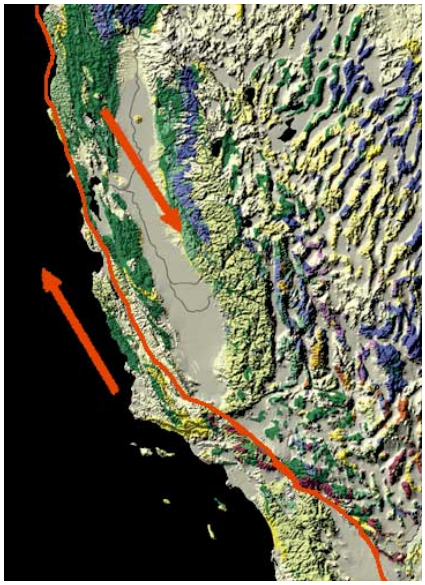


Figure 14.2.12: Map of the San Andreas Fault in California, showing plate motion. (Image by Kate Barton, David Howell, and Joe Vigil of the USGS.)

In 1970, still in the early stages 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 the northern California coast. She used the offshore striping patterns in ocean crust, which form from the freezing of magnetic minerals in basalt, to reconstruct an 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. Once the mid-ocean ridge started subducting, the relative motion of the North American and Pacific plates changed from convergent to transform. This change in transform motion initiated the formation of a strike-slip fault at the plate boundary, which would eventually develop into the San Andreas fault system.

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The current motion, and most likely the past motion, is not exactly parallel to the fault system itself. There is some extensional 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, which rifted the Baja California peninsula from the mainland of Mexico, and the more than 90 degrees of rotation of the Transverse Ranges in Southern California (Ingersoll and Rumelhart, 1999). The video above, illustrated by Tanya Atwater, depicts both events and the widespread extension of the Basin and Range, particularly in Nevada.

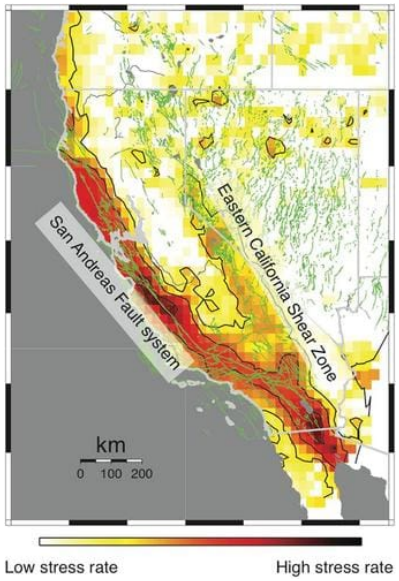


Figure 14.2.13: Stress map of California and nearby regions, showing the Eastern California Shear Zone. (Map by the USGS.)

In the southern section, the San Andreas has occupied at least two other locations (Powell and Weldon, 1992); a new position for the plate boundary appears to be forming in eastern California and western Nevada, within the Eastern California Shear Zone and Walker Lane Fault Zone. This prediction is based on earthquake data indicating that this linear trough accommodates 15-25% of the plate motion between the Pacific and North American plates (Guest, B., 2007).* 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 and 7.1 Ridgecrest Earthquakes. If the plate boundary shifts eastward, California may peel away from the mainland United States, much like Baja California did from Mexico.

* Guest, Bernard; Niemi, Nathan; Wernicke, Brian (1 November 2007). "Stateline fault system: A new component of the Miocene-Quaternary Eastern California shear zone". *Geological Society of America Bulletin*. **119** (11–12): 1337–1347. Bibcode:2007GSAB..119.1337G. doi:10.1130/0016-7606(2007)119[1337:SFSANC]2.0.CO;2. ISSN 0016-7606. Wikidata Q70050019.

Final Thoughts

A 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. However, it can be argued that the plate extends east to Salt Lake City, Utah, as part of the Basin and Range. To the south, the San Andreas merges into a complex series of faults, such as the San Jacinto Fault, and eventually transitions 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, the Queen Charlotte Fault, before curving around to form a subduction zone again near Juneau, Alaska. The difficulty in answering the plate boundary question becomes even more complex when travelling back in time: a mid-ocean ridge that was subducted around 30 million years ago initiated the formation of a transform that would eventually produce the San Andreas Fault. It's an interesting problem that will keep geologists occupied for years to come.

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Key Terms

- **Basin and Range Geologic Province** - a province in the western US containing multiple north-south trending mountain ranges separated by north-south trending valleys

Further reading

https://www.nps.gov/articles/basinrange.htm NPS site on the Basin and Range

Atwater, Tanya. "Implications of plate tectonics for the Cenozoic tectonic evolution of western North America." *Geological Society of America Bulletin* 81, no. 12 (1970): 3513-3536.

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Humphreys, E. D., Post-Laramide removal of the Farallon slab, western United States, *Geology, 23,* 987-990, 1995.

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Miller, Marli B., and Terry L. Pavlis. "The Black Mountains turtlebacks: rosetta stones of Death Valley tectonics." *Earth-Science Reviews* 73, no. 1-4 (2005): 115-138.

Powell, Robert E., R. J. Weldon, and Jonathan C. Matti, eds. *The San Andreas fault system: Displacement, palinspastic reconstruction, and geologic evolution*. Vol. 178. Geological Society of America, 1993.

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14.3: Cenozoic Biology

Introduction

The extinction of all non-avian dinosaurs at the end of the Mesozoic created numerous ecological niches at the start of the Cenozoic. This, along with the absence of predators, enabled the rapid diversification of mammals through **adaptive radiation**. Before long, however, an early order of carnivorous species evolved: the creodonts, whose anatomy was similar to that of today's predators. These types of "hypercarnivores" are described in detail in the video, "When Giant Hypercarnivores Prowled Africa". At the same time, small primate-like mammals evolved to live in trees: the plesiadapiforms (pic).

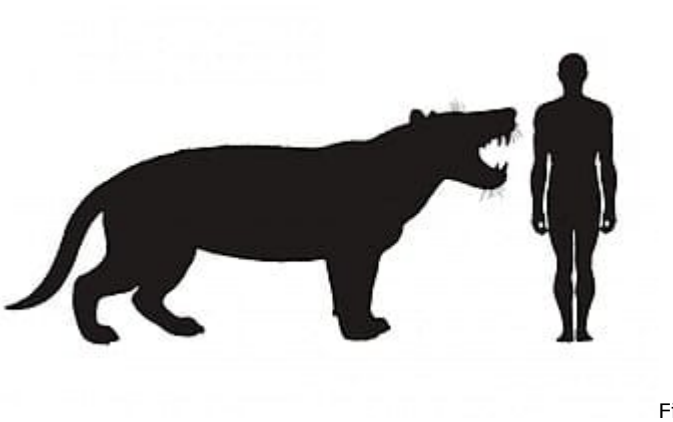


Figure 14.3.1: Simbakubwa, a type of creodont, is compared to a human. Image from: Mauricio Anton, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons.



Figure 14.3.2: Patriofelis, a type of creodont. Image from: Dmitry Bogdanov, CC BY 3.0 <https://creativecommons.org/licenses/by/3.0>, via Wikimedia Commons.

At the start of the Cenozoic Era, the climate was relatively warm, and by the end of the Paleogene Epoch, Earth was experiencing greenhouse conditions, with no continental glaciers anywhere on the planet. This rapid global warming, known as the **Paleocene-Eocene Thermal Maximum**, may have been driven by greenhouse gases released into the atmosphere through increased volcanic activity or the melting of methane ice. The heat was great for reptiles, allowing them to diversify and grow larger. For example, this time saw the world's largest snake, Titanoboa, which grew to 43 feet long, nearly the length of a school bus!

Around 49 million years ago, in the mid-Eocene, the planet began to cool. This gradual shift from greenhouse conditions to eventually icehouse conditions may have been triggered by the widespread growth of the CO2-capturing "micro-fern" in Antarctica, azolla. Through photosynthesis, the azolla may have captured as much 50% of the atmosphere's CO2. The changing

climate conditions led to the extinction of many ancient mammal groups, which were eventually replaced by the ancestors of today's rodents, rhinos, horses, and primates. As the climate cooled and dried, a new biome, the grassland, developed. This significant development drove evolution as animals needed to adapt to eat the new, fibrous grass. In particular, mammals evolved a second stomach, which enabled fermentation of grass and subsequent rechewing to facilitate proper digestion.



Figure 14.3.3: Photograph of azolla from Kurt Stüber [1], CC BY-SA 3.0 <http://creativecommons.org/licenses/by-sa/3.0/>, via Wikimedia Commons.

New World monkeys first appear in the fossil record approximately 26 million years ago. The New World monkeys flourished in the trees, whereas the Old World monkeys adapted to life on the ground. The first apes evolved from Old World monkeys around the beginning of the Miocene Epoch, characterized by the absence of a tail, more flexible shoulders, and shorter spines: traits that eventually helped them to walk upright on two legs. As the climate cooled, grasslands expanded, as did grazing mammals. By the end of the Miocene, approximately 5 million years ago, the lineage of primates that would become humans had diverged from species such as orangutans and gorillas. Our earliest ancestors, *Australopithecus*, inhabited expanding grasslands and may have been able to digest grass. By 2.8 million years ago, a new genus, Homo (as in Homo sapiens), appeared. More detail is given in the following video: From the Fall of Dinos to the Rise of Humans.

YouTube Video

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The Ice Age and Mega Fauna

Earth has experienced at least five ages: periods during which continental ice sheets covered continents, like Greenland and Antarctica today. The ice age we're currently in began 2.6 million years ago, marking the start of the Pleistocene Epoch. During ice ages, Earth naturally cools and warms: **glacial** and **interglacial** periods. Currently, we're experiencing an interglacial period that has been exacerbated by human-caused global warming. The glacial and interglacial periods are the result of changes to Earth's axial tilt (22-24.5 degrees) called obliquity, changes to the direction the axis is pointing (think of a wobbly spinning top) known as precession, and changes in the eccentricity (how much the orbital path varies from a perfect circle) of Earth's orbital path. Combined, these factors can cause more extreme or milder seasons - glacial and interglacial periods.

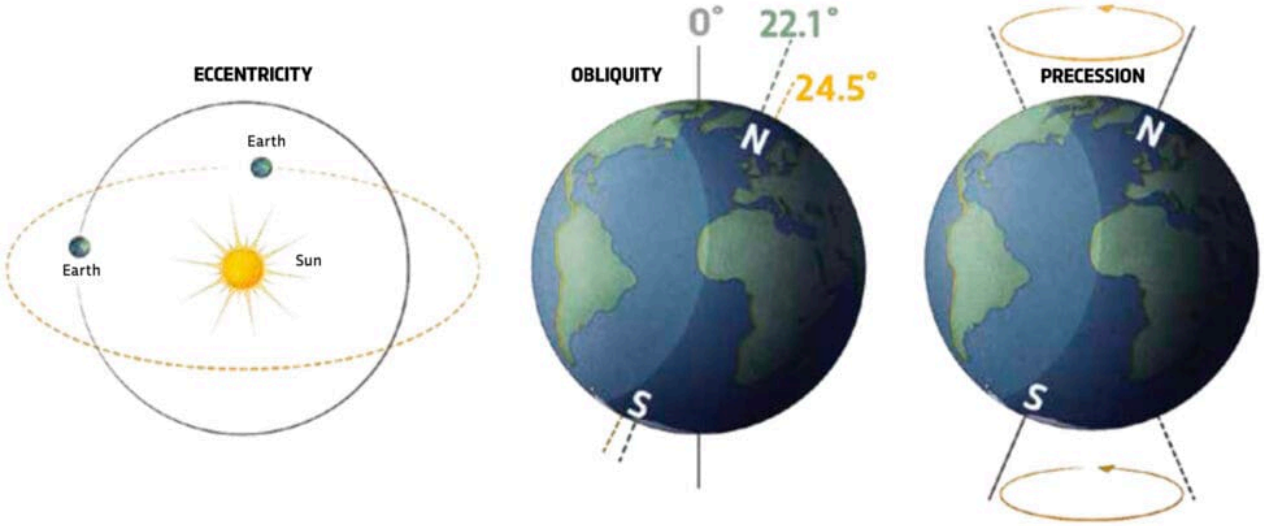


Figure 14.3.4: Eccentricity, Obliquity, and Precession. Image from BBC Science Focus (https://c01.purpledshub.com/bbcscien...ffect-climate/).

The glaciation associated with this ice age significantly shaped Earth's current landscapes. In mountain ranges such as the Sierra Nevada, mountain peaks and intervening valleys are clearly shaped by glacial erosion, whereas in low-lying areas throughout the Midwest and eastern North America, glacial deposits predominate. What brought on this latest ice age is debatable, but it seems that the closure of the Tethys Sea and the equatorial passage between the Atlantic and Pacific oceans were likely contributors, as these events would have disrupted the warm and cold ocean currents, making it more likely that there would be extreme temperature gradients in equatorial vs arctic oceans. In turn, high latitude continents would experience colder winters, and the formation of glaciers would be more likely.

This video discusses the erosional landforms created by glaciers at Long Lake in the Sierra Nevada Mountains in California.

YouTube Video

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During the Pleistocene ice age, some mammal populations grew significantly larger than they do today, including the famous Woolly Mammoths, saber-toothed cats, and the Megalodon. The increase in body size was an adaptation to cool climates. Larger bodies retain heat more effectively, use energy more efficiently, and are more difficult for predators to kill. Additionally, animals continued to fill ecological niches vacated by the dinosaurs, allowing them to reach sizes larger than those of today.

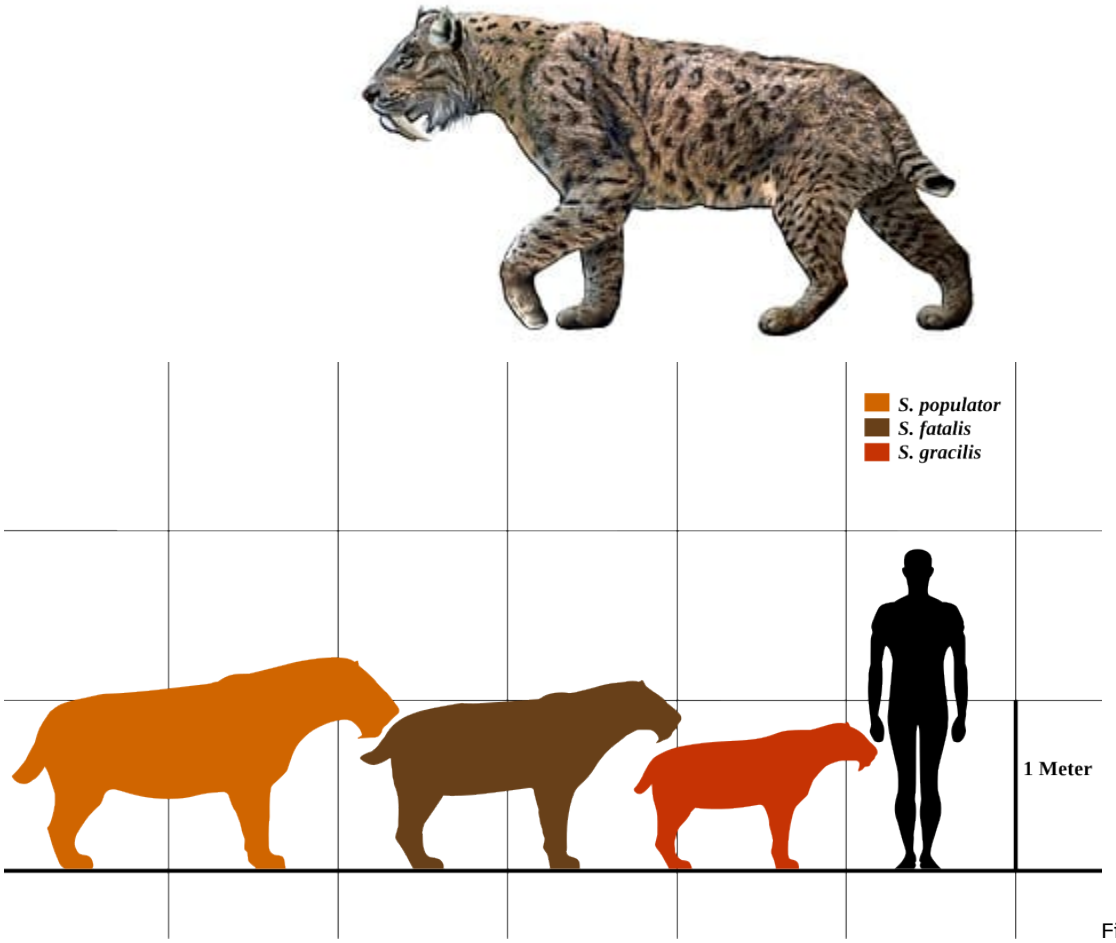


Figure 14.3.5: Smilodon fatalis, a type of saber-toothed cat, and size comparison of commons smildons relative to a human. Images from: Dantheman9758 at English Wikipedia, CC BY 3.0 <https://creativecommons.org/licenses/by/3.0>, via Wikimedia Commons, and Vectorization: Alhadis, CC BY-SA 4.0 <https://creativecommons.org/licenses/by-sa/4.0>, via Wikimedia Commons, respectively.

Cenozoic Evolution

There are five groups of early mammals in the fossil record, based primarily on fossil teeth, the hardest bone in vertebrate skeletons, and therefore most likely to fossilize. For the purposes of this text, the most important group is the Eupantotheres, which diverged into the two main mammalian lineages, marsupials (e.g., *Sinodelphys*) and placentals (eutherians; e.g., *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. Modern-day examples include kangaroos, opossums, and koalas.

Some well-known placental mammal groups have been extensively studied and exhibit interesting evolutionary histories in the Cenozoic. For example, horses began with four toes and evolved into larger forms with a single toe. Cetaceans (marine mammals such as whales and dolphins) originated on land from small bear-like (Mesonychidae) creatures in the early Cenozoic and gradually transitioned to aquatic life. Some of the largest terrestrial animals also evolved, including *Paraceratherium*, a distant relative of rhinoceros.

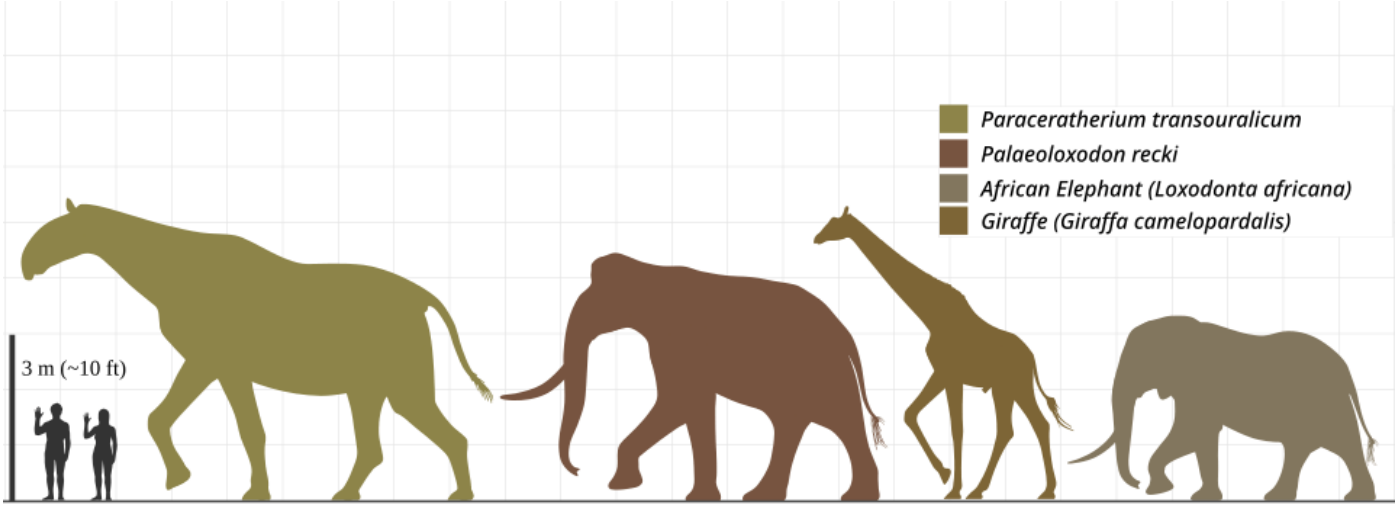


Figure 14.3.6: A comparison of *Paraceratherium* to other mammals. Image from: Steveoc 86 (original silhouettes and diagram) Hemiauchenia (modified version), CC BY 4.0 <https://creativecommons.org/licenses/by/4.0>, via Wikimedia Commons.

However, no evolutionary process has been studied more than human evolution. **Hominids**, the name for human-like primates, started in eastern Africa several million years ago.

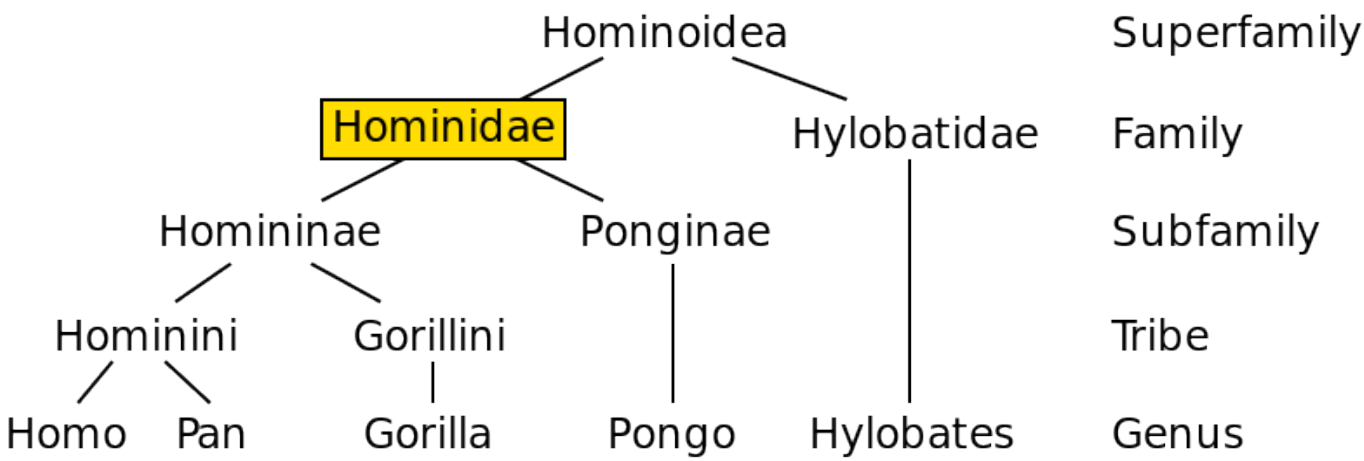


Figure 14.3.7: Family tree of Hominids (Hominadae).

The first critical event in this story is an environmental change from the jungle to the grasslands or savanna biomes. This was probably caused by changes to the Indian Ocean circulation as the Tethys Ocean closed. While bipedalism is known to have evolved before this shift, it is generally believed that our bipedal ancestors, such as *Australopithecus,* had an advantage in covering ground more easily in open environments than their non-bipedal evolutionary cousins. There is also a growing body of evidence, including the famous “Lucy” Australopithecine fossil, that our early ancestors lived in trees. Arboreal animals (tree-living) usually demand a high intelligence to navigate through a three-dimensional world. It is from this lineage that humans evolved, using endurance running to acquire more resources and hunt down prey. 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 14.3.8: Lucy skeleton from the Cleveland Natural History Museum, showing real fossil (brown) and reconstructed skeleton (white).

One advantage that hominids have over other animals is the ability to walk upright, which frees the hands for activities such as tool and weapon use. Another is a large brain. There have been arguments that this increase in brain size can be explained by shifts toward greater meat consumption and the use of fire for cooking, which made meat more nutritious, tool use, and even the emergence of society itself. Regardless of how, this increased cognitive power enabled humans to expand beyond Africa and explore the world, ultimately reaching the Americas via land bridges such as the Bering Land Bridge. Human evolution is a fascinating topic that is covered extensively in anthropology and related textbooks.

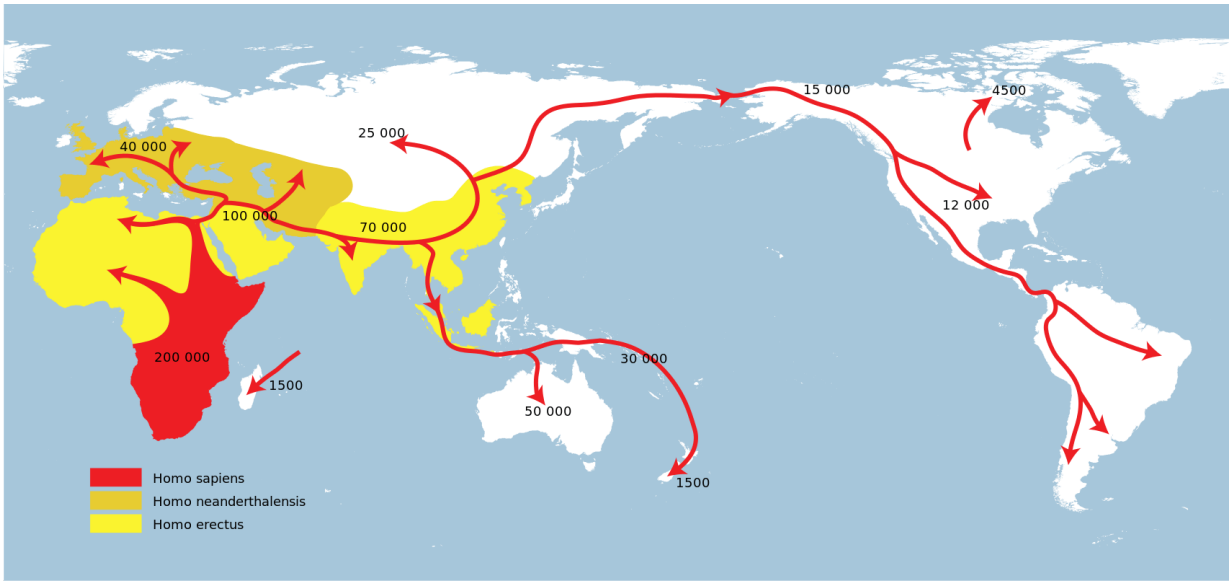


Figure 14.3.9: The hypothesized movement of the homo genus. Years are marked as to the best guess of the timing of movement.

Anthropocene and Extinction

Humans have influenced Earth, its ecosystems, and its climate. Yet human activity cannot explain all of the changes that have occurred in the recent past. The Quaternary Period, the last and current period of the Cenozoic, began 2.58 million years ago with the onset of the current ice age. During this period, ice sheets advanced and retreated, most likely driven by Milankovitch cycles. At this time, various cold-adapted megafauna emerged (e.g., giant sloths, saber-toothed cats, and woolly mammoths), and most of them went extinct as the Earth warmed following the most recent glacial maximum. A long-standing debate concerns the causes of these and other extinctions. Is climate warming to blame, or were they caused by humans? Certainly, we know of recent human-caused extinctions of animals such as the dodo and the passenger pigeon. Can we connect modern extinctions to extinctions

in the recent past? If so, there are several possible explanations for how this happened. Possibly the most widely accepted and oldest is the hunting/overkill hypothesis. This hypothesis is that, as humans hunted large herbivores for food, carnivores were unable to find prey, and human arrival times at locations have been shown to be associated with increased extinction rates in many cases.

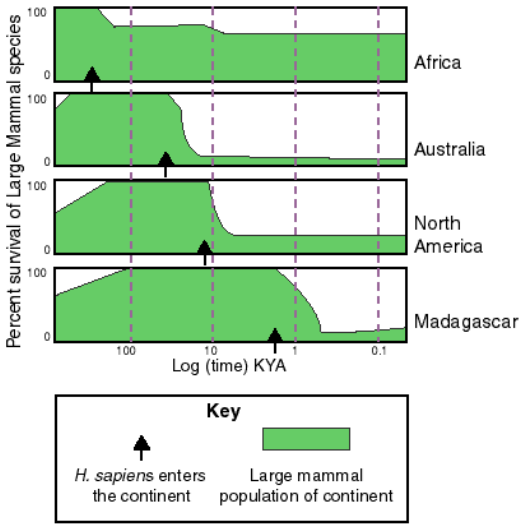


Figure 14.3.10: Graph showing abundance of large mammals and the introduction of humans.

Modern human impact on the environment and the Earth as a whole is unquestioned. In fact, many scientists are starting to suggest that the rise of human civilization ended and/or replaced the Holocene epoch and defines a new geologic time interval: the **Anthropocene**. 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 earth material moved. The biggest scientific debate on this topic is the starting point. Some argue that the human invention of agriculture would be recognized in geologic strata, and that this should be the starting point, around 12,000 years ago. Others link the start of the Industrial Revolution to the subsequent increase in atmospheric carbon dioxide. Either way, the idea is that alien geologists visiting Earth in the distant future would easily recognize the impact of humans on the Earth as the beginning of a new geologic period.



Figure 14.3.11: Bingham Canyon Mine, Utah. This open pit mine is the largest man-made removal of rock in the world.

Key Terms

- **adaptive radiation** - the diversification of a group of organisms into forms filling different ecological niches
- **Anthropocene** - a proposed geological epoch defined by humanity's significant, transformative impact on Earth's climate, ecosystems, and geology
- **glacial** - time interval cold enough to support widespread glaciation
- **hominids** - a group of large-bodied primates that includes modern humans and the great apes
- **interglacial** - time interval between glacials that does not support widespread glaciation
- **Paleocene-Eocene Thermal Maximum** - rapid global warming occurring part way through the Cenozoic Era

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- 2.8: Phanerozoic Eon - Cenozoic Era by Callan Bentley, Karen Layou, Russ Kohrs, Shelley Jaye, Matt Affolter, and Brian Ricketts is licensed CC BY-NC 4.0. Original source: https://opengeology.org/historicalgeology/.

14.4: Key Points

- Plate tectonics moved the continents to their current positions, building mountains across the globe, and closing seaways.
- The convergent boundary established 200 million years ago became a transform boundary when the mid-ocean ridge separating the oceanic Farallon Plate from the Pacific Plate was subducted; the transform boundary became the San Andreas Fault.
- Crustal extension centered in what is today Nevada stretched the crust of North America, creating a regional zone of mountain ranges and linear mountains called the Basin and Range. The stretching and thinning of the crust widened America and triggered voluminous volcanism.
- The current position of the plate boundary may jump eastward, and California could peel away from the rest of America.
- The extinction of dinosaurs enabled mammals to diversify and increase in size.
- Earth cooled during the Cenozoic, allowing the emergence of a new biome, grassland/savannah, and eventually culminating in the ice age.
- Mammals thrived in the habitats vacated by the dinosaurs, with some species becoming significantly larger than modern mammals, such as *Paraceratherium*.
- Hominids adapted to grasslands, eventually learned to use tools and fire, and evolved into the genus *Homo*.

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A

- Absaroka Sequence - A group of North American sedimentary formations formed during marine transgressions in the Mississippian, Permian, Triassic, and Jurassic periods
- abyssal plain – the vast, flat, and deep region of the ocean floor, typically found at depths of 3,000–6,000 meters, covered by fine-grained sediments
- Acadian Orogeny - a mountain-building event affecting a more southern part of the east coast of Laurentia occurring during the late Silurian through the Devonian resulting in the second wave of Appalachian mountain-building
- Acasta Gneiss Complex - located in Canada, it preserves the oldest rocks on Earth some of which where derived from the partial melting of 4.3 billion year old Hadean crust
- accretion - increasing the mass of an object by adding material to it or the addition of crustal material through collisions
- accretion(ary) - the process of small particles and gases within the solar nebula combining together to form larger particles and, eventually, planets
- accretionary wedge - structure near an oceanic trench formed by the accumulation of sediment scraped off the subducting oceanic lithosphere and shoved onto the overriding oceanic or continental plate
- active margin - a continental coastline that is marked by a plate boundary
- adaptive radiation - the diversification of a group of organisms into forms filling different ecological niches
- aeolian – refers to processes, sediments, or landforms produced or transported by wind, such as sand dunes deposits
- aerobic - involving requiring oxygen
- albedo - the ability of a surface to reflect sunlight
- Alleghenian Orogeny - the third and final episode of Appalachian mountain-building causing folding in the southern part of the range and the Ouachita Mountains in Texas and Oklahoma to form occurring during the Carboniferous and Permian periods
- alluvial fan – a fan- or cone-shaped deposit of sediment that forms where a river or stream flows out of a mountainous area onto a flatter plain
- alpine glacier – a glacier that forms in mountainous regions and flows down valleys under the influence of gravity
- amniotes - a group of tetrapod vertebrates comprising reptiles, birds, and mammals defined by their ability to lay eggs on land or retain fertilized eggs within the mother
- amphibolite - non-foliated metamorphic rock made of amphibole (hornblende) formed from a mafic igneous protolith
- anapsids - the group of sauropsids with reptile skulls with no openings in the rear of the skullavemetatarsalia - bird-line archosaurs
- andesite - an aphanitic (intrusive), intermediate igneous rock
- angular unconformity - a geological feature where younger, horizontal sedimentary rock layers are deposited on top of older, tilted or folded sedimentary rock layers, creating an angular discordance between the two sets of strata
- anion - a negatively charged ion formed as the number of electrons exceeds the number of protons
- anoxic - having no oxygen
- Anthropocene - a proposed geological epoch defined by humanity's significant, transformative impact on Earth's climate, ecosystems, and geology
- anticline – a fold in rock layers that is arch-shaped, with the oldest rocks located at the core of the fold and the limbs dipping away from the center
- Antler Orogeny - first mountain-building event of the Cordilleran occurring at the end of the Devonian
- archosaurs - a branch of diapsids that includes the last common ancestor of birds and crocodilians
- arkose sandstone - clastic sedimentary rock composed of sand-sized feldspar particles
- arthropods - a phylum of invertebrate animals characterized by a segmented body, a hard exoskeleton, and jointed appendages
- artificial selection - the process by which humans deliberately breed organisms for specific and preferred traits
- asthenosphere - the layer beneath the lithosphere, which is in a plastic, semi-solid state
- atomic mass - the mass of an atom determined by adding the number of protons and neutrons in the atom
- atomic number - the number of protons in an atom which defines what element the atom is
- augite - a common single-chain silicate mineral rich in iron and magnesium though other elements often substitute in making its formula complex; the most common mineral of the pyroxene family
- axial plane – an imaginary surface that divides a fold into two roughly symmetrical halves, passing through the center of each layer

B

- background extinction - the normal, ongoing rate of species loss that occurs as a result of natural ecological and evolutionary processes
- banded iron formation - unusual chemical sedimentary rocks with distinctive alternating layers of dark gray to black iron oxides and red to yellow chert that could only form in an ocean that is largely oxygen-free
- basalt - an aphanitic (intrusive), mafic igneous rock
- Basin and Range Geologic Province - a province in the western US containing multiple north-south trending mountain ranges separated by north-south trending valleys
- beach – a gently sloping shoreline environment composed of unconsolidated sediment (sand, gravel, or pebbles) shaped by waves, tides, and currents
- bed - a layer of sediment or sedimentary rock that is distinguishable from adjacent layers due to differences in its composition, texture, or other physical properties
- bedding - the layered structure of sedimentary rocks, where layers, are separated by planes
- bedform - a geological feature developed when a moveable bed interfaces with a fluid resulting in the material being shaped by the flow
- binomial nomenclature - a system of naming plants and animals by assigning a two-part, Latin-based name to each, consisting of the genus and the species
- biochemical sedimentary rock - sedimentary rock formed as shells and the body parts of underwater organisms lithify
- biostratigraphy – a branch of stratigraphy that uses fossil content and fossil assemblages to divide and correlate rock layers, helping to establish their relative ages
- biotite - a sheet silicate rich in iron and magnesium making it light in color
- bioturbation - the physical churning and mixing of sediments by the activities of living organisms (such as burrowing)
- bituminous coal - sedimentary rock formed by the breakdown of peat bog material
- body fossil - a preserved remnant or part of an actual organism, such as bones, teeth, shells, leaves, or wood
- Bouma Sequence - a geological model describing the idealized vertical succession of sedimentary structures and lithologies in a turbidite
- Bowen's Reaction Series - a diagram that depicts the order in which minerals crystallize as molten material cools
- braided river – a river system with multiple intertwining channels separated by bars or islands of sediment, usually formed where sediment supply is high and water amount and velocity varies
- breccia - poorly sorted clastic sedimentary rock composed of angular gravel-sized particles
- brittle deformation – the breaking or fracturing of rocks a a result of exposure to stress
- Burgess Shale - a Cambrian lagerstätte that allowed scientists to study Cambrian animals in immense detail

C

- calcite - the only "Big Ten" mineral that is not a silicate; composed of calcium, carbon, and oxygen; typically associated with sedimentary processes
- Caledonian Orogeny - a mountain-building event affecting the northern part of the east coast of Laurentia occurring during the late Silurian through the Devonian
- Cambrian Explosion - the sudden appearance of biological diversity at the start of the Paleozoic Era
- carbonate rocks – sedimentary rocks made mostly of carbonate minerals (like calcite and dolomite), commonly formed in marine environments (e.g., limestone, dolostone)
- carbonates - sedimentary rocks primarily composed of carbonate minerals like limestone (calcite or aragonite) and dolomite -
- carbonization - a specific form of fossilization where organic matter (like plants) is converted into a thin, dark, flattened film of carbon due to heat and pressure, driving off volatile elements and preserving detailed impressions of the original organism
- casts - replicas of organisms, essentially recreating their original shape
- cation - a positively charged ion formed as the number of protons exceeds the number of electrons
- chalk - sedimentary rock made of very small shells made of calcite
- chemical sedimentary rock - sedimentary rock formed as minerals precipitate from ions dissolved in solution
- chert - sedimentary rock made of very small crystals of quartz
- Chicxulub Crater - a ~180-kilometer-wide impact structure buried beneath the Yucatán Peninsula in Mexico formed about 66 million years ago by the collision of a large asteroid, meteoroid or comet
- chondrules - small grains found in certain types of meteorites formed in the early solar nebula as molten droplets quickly

cooled and crystallized
- chordata - a major animal phylum containing all vertebrates and some specialized invertebrates all sharing specific characteristics such as having a hollow dorsal nerve cord
- chronostratigraphy – the study of rock strata in relation to geologic time, focusing on assigning absolute or relative ages to rock units and organizing them into a standardized time framework
- clastic sedimentary rock - sedimentary rock composed of sediment compacted and cemented to form a new rock
- clay - a group of silicate minerals formed through the weathering of pre-existing silicate minerals; also a very small size of material
- claystone - clastic sedimentary rock composed of clay-sized particles
- composition - the materials something is made of
- compressional stress – a type of stress that squeezes rocks together, shortening and thickening the crust
- confining pressure - pressure coming from all directions equally
- conglomerate - poorly sorted clastic sedimentary rock composed of rounded gravel-sized particles
- contact metamorphism - metamorphism that occurs in rock exposed to high temperatures resulting from the proximity to a magma body or lava flow
- continental crust - thicker, less dense crust that when paired with the solid upper mantle creates continental landmasses
- continental volcanic arc - a chain of volcanoes formed on a continental plate as a result of the subduction of oceanic lithosphere
- convection - the movement of the mantle as it transfers heat from the hot core to the brittle lithosphere
- convergent boundary - plate boundary type at which the two plates move toward one another
- correlation – the process of demonstrating equivalence or continuity between rock units or strata at different geographic locations, based on lithology, fossil content, age, or other stratigraphic markers
- covalent bond - a strong type of bond in which atoms share electrons to complete their valence shells rending them stable
- craton - tectonically stable interiors of continents composed of the shield, which is crystalline basement rock like gneiss and greenstone, and the platform of sedimentary rocks covering the shield
- cross section – a vertical slice or profile through the Earth, showing the arrangement and relationships of rock layers, structures, and features beneath the surface
- cross-bed - a sedimentary structure where layers of sediment are deposited at an angle to the main, horizontal bedding plane
- crystalline limestone - a sedimentary rock formed as calcite precipitates
- cyanobacteria - single or multicellular bacteria that use chlorophyll to absorb energy from light and were most responsible for constructing stromatolites

D

- daughter product – the stable atom that forms as a result of the decay of a radioactive parent isotope
- debris flow - mixtures of mud, water, and coarse debris, moving down a slope due to gravity that lack fluid turbulence
- Deccan Traps - one of Earth’s largest continental flood basalt provinces, located primarily in western India and erupted in multiple phases around 66 million years ago
- decompression melting - melting of rock triggered as rock rises and pressure is lowered
- delta – a wedge-shaped deposit of sediment formed at the mouth of a river where it enters a standing body of water, such as a lake or ocean
- desert – a region that receives very little precipitation (typically less than 25 cm per year), characterized by sparse vegetation and dominance of wind and occasional water erosion
- detrital pyrite - rounded, transported pyrite found in sedimentary rocks, serving as an indicator of an oxygen-free Earth atmosphere
- diamictite - generic term for poorly sorted sediment; it could be from glacial activity but could also be from landslide or other sedimentary activities
- diapir - blobs of magma that rise towards Earth's surface
- diapsids - the group of sauropsids with two openings in the rear of the skull
- differential stress - an unequal balance of forces in one or more directions
- differentiation - material settling out according to density or the process of a planet developing a layered structure according to density and chemical composition
- diorite - a phaneritic (intrusive), intermediate igneous rock
- dip – the angle at which a rock layer, fault, or other planar feature is inclined from the horizontal, measured downward in the direction of maximum slope
- disconformity - a gap in the geological record that occurs between parallel layers of sedimentary rock
- discontinuous bed - a sediment ary bed that pinches out in one direction

- divergent boundary - plate boundary type at which the two plates move away from one another
- dolostone - a sedimentary rocks formed as magnesium replaces the calcium in limestone before lithification
- dropstones - random large clasts found within fine-grained, thinly-bedded and laminated deepwater deposits of clay and silt that were transported out into the ocean by an iceberg that then melted and dropped it
- ductile deformation – the bending, flowing, or folding of rocks without breaking when subjected to stress, usually at high temperature and pressure

E

- Ediacaran biota - plants and animals that evolved in the late Proterozoic and represent the earliest known complex multicellular organisms
- elastic deformation – a temporary change in the shape or size of a rock that is reversible; the rock returns to its original form once the stress is removed
- electron - a negatively charged subatomic particle with essentially no mass that resides outside of an atom's nucleus
- endosymbiosis - the process of two organisms merging into one to form a mutually beneficial relationship
- epibole – the sudden, widespread appearance (or disappearance) of a fossil species or assemblage within the stratigraphic record, often used as a marker horizon for correlation
- erosion - the removal of weathered rock material from its original location
- eukaryotes - organisms with membrane-bound nuclei and include all animals, plants, fungi, seaweeds, and many unicellular organisms
- eutrophication - the over-enrichment of water by nutrients, primarily phosphorus and nitrogen, causing excessive algae growth and oxygen depletion
- evolution - the process of gradual change in the characteristics of a population leading to the development of new species over many generations
- evolutionary fauna - a group of marine animal taxa with hard skeletons that appear, evolve, and go extinct together as a distinct, dominant community over a particular geological time
- external mold - an impression of an organism’s outside shape
- extrusive - igneous rock formed as molten material cools on the surface

F

- fabric - the arrangement and organization of a rock's constituent parts and features
- facies - Characteristics of a rock unit that reflect the origin of the rock and distinguish it from adjacent rock units. These characteristics include mineralogy, sedimentary structures, fossil content, and texture. These provide clues about the depositional environment in which a rock was formed
- facies – the set of characteristics (such as rock type, fossils, sedimentary structures, and color) that reflect the conditions in which a sedimentary rock was deposited
- feldspar - a framework silicate mineral with a wide variety of compositions as potassium, sodium, and calcium fit into the structure
- felsic - a composition in which there are high (~70%) concentrations of silica; resultant rocks are typically light in color
- foliated/foliation - the characteristic of a metamorphic rock by which there is a distinct alignment of minerals
- footwall – the block of rock that lies beneath an inclined fault plane
- foreset - an inclined layer of sediment in a bedform that dips in the direction of current flow, forming the steep part of the structure
- formation – a distinct, mappable body of rock with unique physical characteristics, such as color, composition, or particle size, that differentiates it from the surrounding rock layers
- formation – a fundamental lithostratigraphic unit of rock that is mappable, has a distinct set of characteristics (lithology, composition, etc.), and is thick and extensive enough to be recognized at the Earth’s surface or in the subsurface
- fossil – any preserved remains, impression, or trace of an organism that lived in the past; must be older than 11,700 years old
- fossiliferous limestone - sedimentary rock composed of calcite that contains an abundance of fossils

G

- gabbro - a phaneritic (intrusive), mafic igneous rock
- gene migration - the movement of genetic material from one population to another through the migration of individuals into a new area

- genetic drift - a change in the frequency of an allele over time caused by some individuals reproducing more successfully than others
- genetic mutations - changes in the sequence of nucleic bases in the coding portion of a DNA molecule
- genetic symbiosis - the transfer of genetic material from one species to another due to close physical interactions
- genotype - a unique sequence of DNA
- Geologic Time Scale - a graphical representation of Earth's history based on the rock record that shows the timing and relationships of geologic and biologic events
- glacial - time interval cold enough to support widespread glaciation
- gneiss - a foliated metamorphic rock having minerals separated into light and dark bands
- graded bedding - a sedimentary layer characterized by progressively finer-grained particles from the bottom upward
- granite - a phaneritic (intrusive), felsic igneous rock
- Great Ordovician Biodiversification Event (GOBE) - an event during the late Ordovician in which vertebrates and invertebrates became more diverse and complex at family, genus, and species level
- greenstone - non-foliated metamorphic rock made of chlorite formed from a basalt protolith
- greenstone belt - a linear or branching zone of metamorphosed sedimentary and igneous rock within Precambrian cratons; the name comes from the presence of green minerals like chlorite, actinolite, and eclogite
- greywacke - clastic sedimentary rock composed of sand-sized particles that vary in their composition
- group – a lithostratigraphic unit consisting of two or more related formations that share similar geologic features and are grouped together for convenience in mapping and description

H

- Hadean - the earliest part of the Precambrian Eon starting at the conception of Earth up to 4 Ga
- half-life – the amount of time it takes for half of the atoms in a sample of a radioactive isotope to decay into daughter products
- halides - minerals consisting of chlorine bonded with sodium or other cations
- hanging wall – the block of rock that lies above an inclined fault plane
- hominids - a group of large-bodied primates that includes modern humans and the great apes
- homology - a similarity in structure, genetics, or developmental pathways among different organisms that results from shared ancestry
- horizontal gene migration - the movement of genetic material between organisms that are not parent and offspring
- hornblende - a common double-chain silicate mineral; it has a complex formula due to its many possible substitututions

I

- igneous rock - a rock that forms as minerals crystallize when molten material cools
- imbrication - a primary depositional fabric where elongated or flat clasts are arranged in a overlapping pattern indicating flow direction
- index fossil - organisms that lived for relatively short time periods and distributed over a wide geographic area
- interglacial - time interval between glacials that does not support widespread glaciation
- intermediate - a composition in which there are moderate (~60%) concentrations of silica; resultant rocks are typically medium in color
- internal mold – a fossil formed when sediment fills the inside of a shell or other hollow structure; after the shell dissolves, the hardened sediment shows the internal shape
- intrusive - igneous rock formed as molten material cools below the surface
- invertebrate - an animal that does not have a backbone or spinal column
- ionic bond - a type of bond in which atoms lose or gain electrons to become negatively or positively charged and thus attracted to one another
- isotopes – atoms of the same element that have the same number of protons but different numbers of neutrons, giving them different atomic masses

K

- Kaskaskia Sequence - A group of North American sedimentary formations formed during marine transgressions from the Devonian through the Pennsylvanian periods
- komatiite - an aphanitic (intrusive), ultramafic igneous roc

- K–Pg mass extinction - also referred to as the K-T Extinction, this extinction occurred at 66 million years and eliminated roughly 75% of Earth's species
- Kuiper Belt - a doughnut-shaped region of icy objects beyond the orbit of Neptune; it is home to Pluto and most of the known dwarf planets and some comets

L

- lacustrine – pertaining to lakes; describes sediments, deposits, or environments associated with lake basins
- Lagerstätte - a site of exceptional fossil preservation that includes impressions of soft body parts
- Lamarck's inheritance of acquired traits - the theory that an organism's offspring inherit characteristics that the parent acquired during its lifetime through the use or disuse of organs
- Laramide Orogeny - a major mountain-building event that affected western North America from roughly 80 to 40 million years ago
- late Devonian extinction event - an extinction event at the end of the Devonian occurring as the oceans became too deficient in oxygen to support life
- Late Heavy Bombardment - a period of extraordinary meteorite impacts at the start of the Archean, thought to have been triggered by the gravitational interaction of Jupiter and Saturn
- Laurentia - a Proterozoic craton that forms the core of the North American continent
- lava - molten material on Earth's surface
- lenticular bed - a sedimentary bed that pinches out in both directions
- lepidosaurs - a branch of diapsids known for overlapping scales
- limb – the side or arm of a fold, consisting of rock layers that dip away from (or toward) the center of the fold
- lithosphere - the outermost layer of Earth that includes the crust and solid upper mantle; this solid, rock-layer, is broken up into individual pieces called plates
- lithostratigraphy – the branch of stratigraphy that classifies rock units based on their physical and lithologic characteristics, without direct reference to age

M

- mafic - a composition in which there are low (~50%) concentrations of silica; resultant rocks are typically dark in color
- magma - molten material below Earth's surface
- magnetostratigraphy – a stratigraphic method that uses the record of Earth's magnetic field reversals preserved in rocks to establish ages and correlate strata
- mantle plume - long-lived narrow conduits of magma rising from Earth's core to make hot spots
- map – a two-dimensional representation of the Earth's surface used to show the distribution of rock units, structures, and landforms
- marble - non-foliated metamorphic rock made of calcite or dolomite formed as limestone was metamorphosed
- mass extinction - a widespread, rapid decrease in the Earth's biodiversity, where at least 75% of all species perish within a geologically short interval
- mass number – the total number of protons and neutrons in the nucleus of an atom
- meandering river – a river with a single, sinuous channel characterized by large bends (meanders) that migrate laterally across the floodplain
- member – a lithostratigraphic subdivision of a formation, smaller than a formation but still mappable, that has distinctive properties
- metamorphic rock - rock formed when elevated temperatures and pressures cause minerals to change into new minerals without going through the melting phase
- micrite - sedimentary rock made of very fine grains of calcite
- migmatite - a metamorphic rock that is partially melted showing the transition from metamorphic and igneous activity
- mineral - naturally occurring, inorganic solid with an ordered atomic structure and definite chemical composition
- monocline – a simple fold in which otherwise horizontal rock layers bend in one direction, creating a single step-like feature
- mountain belt - a mountain chain formed as a result of the collision of two continental plates
- muscovite - a sheet silicate rich in potassium making it light in color

N

- native elements - minerals that are composed of one element, existing in a pure or nearly pure state

- natural selection - the method by which organisms with traits better suited to their environment are more likely to survive and reproduce, passing those advantageous traits to their offspring
- nebula - a massive cloud of gas and dust in space that is the byproduct of a previous star that suffered a supernova explosion
- neo-Darwinian evolutionary theory - the theory that combines Charles Darwin's concept of natural selection with modern genetics, including Mendelian inheritance and the understanding of mutations
- neutron - a subatomic particle with no charge having a mass of 1 atomic mass unit that resides inside an atom's nucleus
- Nevada Orogeny - a Middle to Late Jurassic (roughly 170–140 million years ago) compressional event along the western margin of North Americ
- nonconformity - a gap in the geologic record where sedimentary rock overlies igneous or metamorphic rock
- non-foliated - the characteristic of a metamorphic rock by which there is no alignment of minerals
- normal fault – a type of dip-slip fault where the hanging wall moves downward relative to the footwall, caused by tensional stress
- nucleus - the central portion of an atom in which protons and neutrons reside
- numerical dating - the type of geologic dating in which you assign a number age to a geologic event
- Nuvvuagittuq Greenstone Belt (NGB) - located in Canada, rock from this has been dated back to 4.31 Ga making it the oldest preserved rock on earth (though the dating is controversial)

O

- oceanic crust - thinner, more dense crust that when paired with solid upper mantle creates the ocean floors
- oceanic ridge - mountain chain at the bottom of the ocean, often in the middle of the ocean, created at divergent boundaries where new crust is created
- olivine - a common silicate mineral rich in iron and magnesium making up much of the ocean floor and the mantle
- oolitic limestone - sedimentary rock formed from small grains of calcite
- organic sedimentary rock - sedimentary rock formed as the remains of organic materials undergo lithification
- ornithischians - herbivorous dinosaurs characterized by a “bird-hipped” pelvic structure in which the pubis is rotated backward
- orogeny - mountain building event occurring due to compression at a convergent plate boundary
- outage – in stratigraphy, more commonly referred to as a hiatus, it is a gap in the geologic record representing a period of non-deposition or erosion between rock layers
- overturned fold – a fold in which one or both limbs are tilted beyond vertical, so that the rock layers on one limb are upside down.
- oxides - minerals consisting of metal ions covalently bonded with oxygen

P

- Paleocene-Eocene Thermal Maximum - rapid global warming occurring part way through the Cenozoic Era
- paleomagnetic - ancient magnetic orientation
- paleomagnetism - the study of the Earth's ancient magnetic field, preserved in rocks and sediments
- paleontology - the study of the history of life on Earth as based on fossils
- parent – the original radioactive isotope that undergoes decay to form a daughter product
- passive margin - a continental coastline that is not marked by a plate boundary
- peridotite - a phaneritic (intrusive), ultramafic igneous rock
- Permian-Triassic (P-T) extinction - mass extinction event at the end of the Paleozoic Era causes by global warming and drying that resulted from the assembly of Pangea
- permineralization – a fossilization process where minerals carried by water fill the pores and empty spaces in bones, wood, or shells, preserving fine details of the original structure
- phenotype - the set of observable characteristics of an organism, resulting from the interaction between its genetic makeup (genotype) and environmental factors
- phosphates - minerals consisting of a PO4 2- cation bonded with various ions
- photosynthesis - a system of biological processes where plants take in CO2 to convert sunlight into chemical energy which fuels their metabolism and releases oxygen to their surroundings
- phyllite - 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
- planetesimals - the nuclei of planets formed in the early solar system from collisions with other objects in the solar system
- plate tectonics - the theory that Earth’s rigid outer shell, the lithosphere, is broken up into individual pieces called plates,

that move and interact with one another along plate boundaries, driving Earth's constructive geologic processes
- plates - pieces of lithosphere composed of oceanic and/or continental crust and the solid upper mantle
- platform – a portion of a craton that is covered by relatively undeformed, horizontally lying or gently tilted sedimentary rocks, overlying older crystalline basement
- plutonic - igneous rock formed as molten material cools below the surface
- porphyroblast - a large crystal of a particular mineral in a metamorphic rock surrounded by small grains
- principle of cross-cutting relationships – a geologic principle stating that a rock or geologic feature (such as a fault or intrusion) that cuts across another rock body must be younger than the rock it cuts through
- principle of faunal succession – the concept that fossils succeed one another in a definite, recognizable order, allowing rock layers to be dated and correlated by the fossils they contain
- principle of inclusions – states that any rock fragments (inclusions) contained within another rock must be older than the rock in which they are enclosed.
- principle of lateral continuity – sedimentary layers originally extend in all directions until they thin out or are interrupted by a barrier, meaning that separated layers on opposite sides of a valley were once continuous
- principle of original horizontality – states that layers of sediment are originally deposited in horizontal or nearly horizontal layers so tilting must have occurred after deposition
- principle of superposition – in an undisturbed sequence of sedimentary rocks, the oldest layers are at the bottom and the youngest layers are at the top
- prokaryotes - single-celled organisms that lack nuclei and predate eukaryotes
- prokaryotes - organisms whose cells lack a nucleus and other organelles
- protolith - the parent rock or the rock from which the metamorphic rock forms
- proton - a positively charged subatomic particle having a mass of 1 atomic mass unit that resides inside an atom's nucleus
- protoplanetary disk - ring of debris around a star
- pseudosuchian - crocodile-line archosaurs

Q

- quartz - a framework silicate mineral composed entirely of silicon and oxygen though small amounts of other elements can cause color changes
- quartz sandstone - clastic sedimentary rock composed of sand-sized quartz particles
- quartzite - non-foliated metamorphic rock made of quartz formed as quartz sandstone was metamorphosed

R

- radioactive decay – the natural process by which an unstable atomic nucleus loses energy by emitting particles or radiation, transforming into a more stable form (often producing a daughter isotope)
- recrystallization – a preservation process where the original minerals of a fossil change into more stable forms without changing the overall shape of the fossil
- recumbent fold – a fold that has been rotated so much that its axial plane is nearly horizontal, causing the limbs to be almost parallel
- regional metamorphism - metamorphism that occurs over a wide area resulting from the collision and compression of tectonic plates
- regression – a geologic process in which the sea level lowers, exposing previously submerged areas and leading to the deposition of progressively shallower-water sediments over deeper-water sediments
- regression - a time period when sea levels drop relative to land masses; caused by sea level drop or land level rise
- relative dating – a method of determining the chronological order of past events by comparing rock layers and fossils, without determining their exact numerical age
- replacement – a fossilization process where the original material of an organism is gradually dissolved and replaced molecule by molecule with minerals, creating a detailed copy
- reverse fault – a type of dip-slip fault where the hanging wall moves upward relative to the footwall, caused by compressional stress
- rhyolite - an aphanitic (intrusive), felsic igneous rock
- ridge push - the elevation of ocean ridges, caused by rising magma, pushes the plates away from the high spot down toward the deeper ocean floor causing the plate to move
- rift valley - elongated low spot created when a divergent boundary develops in continental lithosphere
- rift zones - weak spots in Earth's crust where stretching is occurring creating low spots
- rock - a solid composed of grains of one or more minerals or mineraloids

- rock cycle - a diagram that describes the processes through which the three main types of rocks – igneous, metamorphic, and sedimentary – are created, transformed, and recycled over geological time
- rock gypsum - sedimentary rock made of gypsum
- rock salt - sedimentary rock made of halite
- Rodinia - a supercontinent that formed and broke up in the late Proterozoic

S

- San Andreas Fault System - a series of faults that mark the transform plate boundary that accommodates horizontal motion between the Pacific Plate and the North American Plate
- Sauk Sequence - A group of North American sedimentary formations formed during marine transgressions in the late Proterozoic and Ordovician periods
- sauropods - large, long-necked, quadrupedal, "lizard-hipped" herbivorous dinosaurs known for their columnar limbs, elongated cervical vertebrae, and massive body size
- sauropsids - the branch of amniotes that would eventually evolve into reptiles, dinosaurs, and birds
- schist - a foliated metamorphic rock where minerals are aligned and visible as individual crystals
- seafloor spreading - process where new oceanic crust is formed at oceanic ridges and moves away from the ridge
- sedimentary basin – a low area of the Earth's crust where sediment accumulates over time to form thick, laterally extensive sequences
- sedimentary structures – features that form in sediments at or shortly after deposition, providing clues to the depositional environment (e.g. a current, alternating wet and dry environment, etc.)
- sequence – a relatively conformable package of sedimentary strata bounded above and below by unconformities (erosional or non-depositional surfaces) that record a cycle of sea-level change and sediment deposition
- Sevier Orogeny - a major compressional mountain-building episode that occurred from the Late Jurassic through the Late Cretaceous (approximately 140–50 million years ago)
- shale - clastic sedimentary rock composed of silt- and clay-sized particles that breaks into sheets
- shear stress – a type of stress that causes parts of a rock body to slide past each other in opposite directions
- shield – a large, exposed area of a craton composed of old crystalline igneous and metamorphic rocks, typically tectonically stable and deeply eroded
- silica tetrahedron - a molecule composed of one silicon atom and four oxygen atoms arranged in a tetrahedron with the silicon in the middle
- siliciclastic rocks – sedimentary rocks made mainly from silicate mineral fragments formed from the weathering and erosion of preexisting rocks
- siltstone - clastic sedimentary rock composed of silt-sized particles
- slab pull - the pulling force exerted by an oceanic plate subducting into the mantle due to its own weight
- slate - a fine-grained foliated metamorphic rock which breaks into flat sheets
- Snowball Earth - periods of time when glaciers were so extensive they covered most of Earth
- Sonoma Orogeny - an Early Triassic (about 250–230 million years ago) mountain-building event along the western margin of North America
- Sonoman Orogeny - another episode of mountain-building on the western side of North America caused by the accretion of island arcs and other fragments of continental crust occurring during the Carboniferous and Permian periods
- strain – the measure of change in a rock's shape or volume in response to applied stress
- stramatolite - layered, sedimentary rock structures formed by photosynthetic cyanobacteria trapping sediment in shallow water
- strata – distinct layers of sedimentary rock or volcanic deposits that can be distinguished by composition, texture, or other features
- stratigraphic column - a diagram that shows a vertical sequence of rock layers in the order they were deposited representing the rock units, their thicknesses, lithologies (rock types), and sometimes fossils or other features to illustrate the geologic history of an area
- stratigraphic column - a diagram that shows a vertical sequence of rock layers in the order they were deposited representing the rock units, their thicknesses, lithologies (rock types), and sometimes fossils or other features to illustrate the geologic history of an area
- stratigraphy – the branch of geology concerned with the description, correlation, and interpretation of layered rocks, including their age relationships and depositional environments
- stress – the force applied to a rock per unit area, which can cause deformation

- striations - scratches or grooves that result when a glacier takes a larger particle of rock and drags it against the bedrock, gouging out a scratch mark that shows the direction the glacier was flowing
- strike – the compass direction of the line formed by the intersection of a rock layer or planar feature with a horizontal surface
- strike-slip fault – a fault in which the main movement is horizontal and parallel to the fault's strike, caused by shear stress
- stromatolites - layered sedimentary structures formed by the growth of bacteria or algae
- subduction - process by which dense oceanic lithosphere dives down beneath another tectonic plate
- submarine fan – a large, fan-shaped body of sediment deposited on the deep ocean floor by turbidity currents, usually at the base of continental slopes
- sulfates - minerals consisting of metal ions bonded to a sulfate ion
- sulfides - minerals consisting of metal ions bonded with sulfur
- supercontinent - a continent composed of a collection of smaller continental masses
- supergroup – a lithostratigraphic unit that consists of two or more groups, representing an even larger-scale rock package, useful for describing very thick or regionally extensive stratigraphic successions
- synapsids - a lineage of amniotes that includes mammals and their extinct "mammal-like reptile" ancestors, characterized by a specific skull structure with a single opening behind each eye
- syncline – a fold in rock layers that is trough-shaped, with the youngest rocks located at the core of the fold and the limbs dipping inward toward the hinge

T

- tabular bed - a broad continuous layer of sedimentary material characterized by a consistent thickness and planar bounding surfaces
- Taconic Orogeny - an extensive mountain building event 440-470 million years ago caused when subduction pulled a volcanic island arc into Laurentia
- tensional stress – a type of stress that pulls rocks apart, stretching and thinning the crust
- tetrapods - a vertebrate animal with four limbs
- texture - the size, shape, and arrangement of mineral grains, crystals, or fragments within a rock
- therapsids - a major clade of tetrapods that gave rise to mammals
- theropods - bipedal "lizard-hipped" dinosaurs characterized by hollow bones, three-toed limbs
- thrust fault – a low-angle reverse fault (typically dipping less than 30°) where older rocks are pushed up and over younger rocks due to strong compressional forces
- tidal flat – a low-lying coastal area that is periodically flooded and exposed by tides, often composed of mud and sand
- tillite - glacial deposits formed from very poorly sorted sediment; glacial diamictite
- Tippecanoe Sequence - A group of North American sedimentary formations formed during marine transgressions in the Ordovician, Silurian, and Devonian periods
- tonalite-trondhjemite-granodiorite (TTG) - rocks similar to granite that have been metamorphosed and are part of the Acasta Gneiss Complex
- trace fossil – evidence of an organism's activity such as footprints, burrows, feeding marks, or feces
- transform boundary - plate boundary type at which the two plates slide past one another
- transgression – a geologic process in which the sea level rises or land level drops, resulting in progressively deeper-water sediments being deposited over shallower-water sediments
- travertine - form of limestone deposited by mineral springs
- turbidite – a sedimentary deposit formed by an underwater landslide, typically showing graded bedding with coarser material at the base and finer material at the top

U

- ultramafic - a composition in which there are very low (~40%) concentrations of silica; resultant rocks are typically very dark in color
- unconformity - a gap in the geologic record that represents an interruption in the deposition of sedimentary layers
- uniformitarianism – the principle that the geologic processes we see shaping the Earth today (such as erosion, sedimentation, and volcanism) have operated in the same way throughout Earth's history
- uniformitarianism - the recognition that the processes operating on Earth today must have been operating on Earth in the past as well

V

- vertebrate - an animal with a backbone, or spinal column, and an internal skeleton
- vestigial structures - anatomical features or behaviors that persist in an organism but have lost most or all of their original function through evolution
- volcanic - igneous rock formed as molten material cools on the surface
- volcanic island arc - a chain of volcanic islands formed on an oceanic plate as a result of the subduction of oceanic lithosphere

W

- Walther's Law – a principle stating that the vertical order of sedimentary facies in a stratigraphic sequence reflects the lateral changes that existed at the time of deposition, facies that are found adjacent to one another in space will succeed one another vertically in the rock record
- weathering - breakdown of materials due to exposure to the atmosphere and water

X

- xenolith – a fragment of rock enclosed within another, usually a piece of country rock trapped within an igneous intrusion such that the fragment is older than what it is trapped in

Y

- yield point – the limit beyond which a rock will no longer deform elastically; permanent deformation (ductile or brittle) occurs once stress passes this point

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INDEX

G

Great Oxidation Event
11.3: Proterozoic Atmosphere

S

stromatolites
11.3: Proterozoic Atmosphere

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GLOSSARY

S

Sample Word 1 | Sample Definition 1

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By Page

- Introduction to Historical Geology (Ruppert, Lacy, and Haddad) - *Undeclared*
  - Front Matter - *Undeclared*
    - TitlePage - *Undeclared*
    - InfoPage - *Other*
    - Table of Contents - *Undeclared*
    - Licensing - *Undeclared*
  - 1: What is Historical Geology? - *CC BY-NC 4.0*
  - 2: An Introduction to Earth - *CC BY-NC 4.0*
    - 2.1: The Origin of Earth and Our Solar System - *CC BY-NC 4.0*
    - 2.2: Earth's Structure and the Origin of the Continents and Oceans - *Undeclared*
    - 2.3: Key Points - *CC BY-NC 4.0*
  - 3: Plate Tectonics - *CC BY-NC 4.0*
    - 3.1: Plates Defined - *CC BY-NC 4.0*
    - 3.2: Tectonics Defined - *CC BY-NC 4.0*
    - 3.3: Plate Boundaries and Interiors - *CC BY-NC 4.0*
    - 3.4: Geophysical Phenomena - *CC BY-NC 4.0*
    - 3.5: Driving Forces - *CC BY-NC 4.0*
    - 3.6: The Beginning of Plate Tectonics and the Continents - *CC BY-NC 4.0*
    - 3.7: The Story of Plate Tectonics as Recorded by the Rocks - *CC BY-NC 4.0*
    - 3.8: Key Points - *CC BY-NC 4.0*
  - 4: Earth Materials - Minerals - *CC BY-NC 4.0*
    - 4.1: What is a Mineral? - *CC BY-NC 4.0*
    - 4.2: Building Blocks of Rock - *CC BY-NC 4.0*
    - 4.3: Chemistry of Minerals - *CC BY-NC 4.0*
    - 4.4: Formation of Minerals - *CC BY-NC 4.0*
    - 4.5: The Silicate Minerals - Nine of "The Big Ten" - *CC BY-NC 4.0*
    - 4.6: Other Important Rock Forming Minerals - *CC BY-NC 4.0*
    - 4.7: Key Points - *CC BY-NC 4.0*
  - 5: Earth Materials - Rocks - *CC BY-NC 4.0*
    - 5.1: Igneous Rocks - *CC BY-NC 4.0*
    - 5.2: Metamorphic Rocks - *CC BY-NC 4.0*
    - 5.3: Sediments and Sedimentary Rocks - *CC BY-NC 4.0*
    - 5.4: The Rock Cycle - *CC BY-NC 4.0*
    - 5.5: Key Points - *CC BY-NC 4.0*
  - 6: Sedimentology and Stratigraphy - *CC BY-NC 4.0*
    - 6.1: Sediment Deposition and Bedding - *CC BY-NC 4.0*
    - 6.2: Other Sedimentary Structures - *CC BY-NC 4.0*
    - 6.3: Fossils and Fossilization - *CC BY-NC 4.0*
    - 6.4: Facies - *CC BY-NC 4.0*
    - 6.5: Depositional Environments - *CC BY-NC 4.0*
    - 6.6: Introduction to Strata and Basins - *CC BY-NC 4.0*
    - 6.7: Sequence Stratigraphy and Walther's Law - *CC BY-NC 4.0*
    - 6.8: Stratigraphic Correlation - *CC BY-NC 4.0*
    - 6.9: Key Points - *CC BY-NC 4.0*
    - 6.10: Further Reading - *CC BY-NC 4.0*
  - 7: Geologic Structures - *CC BY-NC-SA 4.0*
    - 7.1: Tectonic Movement - *CC BY-NC 4.0*
    - 7.2: Geological Maps and Cross-Sections - *CC BY-NC-SA 4.0*
    - 7.3: Stress, Strain, and Deformation - *CC BY-NC-SA 4.0*
    - 7.4: Folds - *CC BY-NC 4.0*
    - 7.5: Faults - *CC BY-NC 4.0*
    - 7.6: Key Points - *CC BY-NC 4.0*
  - 8: Geologic Time - *CC BY-NC 4.0*
    - 8.1: Earth's Age and Abbreviating Time Intervals - *CC BY-NC 4.0*
    - 8.2: Uniformitarianism and Relative Age Dating - *CC BY-NC 4.0*

- 8.3: Unconformities - *CC BY-NC 4.0*
- 8.4: Fossils and Age Dating - *CC BY-NC 4.0*
- 8.5: Absolute Age Dating - *CC BY-NC 4.0*
- 8.6: The Geological Time Scale - *CC BY-NC 4.0*
- 8.7: Key Points - *CC BY-NC 4.0*

◦ 9: Evolution - *CC BY-NC 4.0*
- 9.1: What is Evolution? - *Undeclared*
- 9.2: Development of the Theory of Evolution - *Undeclared*
- 9.3: A Modern View of Evolution - *Undeclared*
- 9.4: Evolutionary Patterns in Fossils - *CC BY-NC 4.0*
- 9.5: Key Points - *CC BY-NC 4.0*
- 9.6: Further Reading - *CC BY-NC 4.0*

◦ 10: Hadean and Archean - *CC BY-NC 4.0*
- 10.1: Hadean Eon - *CC BY-NC 4.0*
- 10.2: Earth's Oldest Rocks - *CC BY-NC 4.0*
- 10.3: Archean Eon - *CC BY-NC 4.0*
- 10.4: When Did Life Begin? - *CC BY-NC 4.0*
- 10.5: Key Points - *CC BY-NC 4.0*
- 10.6: Further Reading - *CC BY-NC 4.0*

◦ 11: Proterozoic Eon - *CC BY-NC 4.0*
- 11.1: Proterozoic Geography - *Undeclared*
- 11.2: Proterozoic Glaciations - *CC BY-NC 4.0*
- 11.3: Proterozoic Atmosphere - *CC BY-NC 4.0*
- 11.4: Proterozoic Life - *Undeclared*
- 11.5: Key Points - *CC BY-NC 4.0*
- 11.6: Further Reading - *CC BY-NC 4.0*

◦ 12: Phanerozoic Eon - Paleozoic Era - *CC BY-NC 4.0*
- 12.1: Geography and Tectonics of the Early Paleozoic - *CC BY-NC 4.0*
- 12.2: Evidence for the Taconian Orogeny - *CC BY-NC 4.0*
- 12.3: Geography and Tectonics of the Middle Paleozoic - *CC BY-NC 4.0*
- 12.4: Geography and Tectonics of the Late Paleozoic - *CC BY-NC 4.0*
- 12.5: Early Paleozoic Life - The Cambrian and Ordovician - *CC BY-NC 4.0*
- 12.6: Middle Paleozoic Life - The Silurian and Devonian - *CC BY-NC 4.0*
- 12.7: Late Paleozoic Life - The Carboniferous and Permian - *CC BY-NC 4.0*
- 12.8: Key Points - *CC BY-NC 4.0*

◦ 13: Phanerozoic Eon - Mesozoic Era - *CC BY-NC 4.0*
- 13.1: Paleography and Tectonics of the Mesozoic - *CC BY-NC 4.0*
- 13.2: Mesozoic Evolution - Reptiles and Amphibians - *CC BY-NC 4.0*
- 13.3: Mesozoic Evolution - Amniotes - *CC BY-NC 4.0*
- 13.4: Mesozoic Evolution - Dinosaurs - *CC BY-NC 4.0*
- 13.5: Mesozoic Mass Extinctions - *CC BY-NC 4.0*
- 13.6: Key Points - *CC BY-NC 4.0*

◦ 14: Phanerozoic Eon - Cenozoic Era - *CC BY-NC 4.0*
- 14.1: Paleogeography and Tectonics of the Cenozoic - *CC BY-NC 4.0*
- 14.2: An In-Depth Look at the San Andreas Fault and the Basin and Range - *CC BY-NC 4.0*
- 14.3: Cenozoic Biology - *CC BY-NC 4.0*
- 14.4: Key Points - *CC BY-NC 4.0*

◦ 15: Glossary - *CC BY-NC 4.0*

◦ Back Matter - *Undeclared*
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