Chapter 4

Earth

Chapter Objectives:

1. Describe the interior of Earth. Discuss each layer in terms of composition, density, and dynamics.

2. Discuss seismology. Differentiate between P-waves, S-waves, and surface waves.

3. Describe the theory of continental drift. Describe the theory of plate tectonics. Differentiate between the two.

4. List and describe the three types of plate margins.

5. List and define the 3 types of rocks. Describe the rock cycle.

6. Define mineral and rock. Discuss the connection between the two.

7. Differentiate between weathering and erosion. Define chemical and physical weathering.

8. Define soil and discuss its constituent components.

9. List and describe the 6 soil forming factors.

Introduction

The planet Earth is a fairly special place. In all of our studies of the Universe, we have found no planet or moon that is quite like it. It has everything that we and all other creatures need for life. It has a solid surface, is not so hot that most elements vaporize and not so cold that all are frozen solid, has a reasonably stable atmosphere, has puddles of liquid water that are quite large, and has access to a steady energy source, the Sun. In our own Solar System, there are planets and several moons with one or two of these features, but none that comes close to all of them. We live in an age when we have begun to discover planets outside of our Solar System, and so far, we still have not found anything like our own. In the vastness of the Universe, it is almost certain that there is one like it, but the variety does not seem common.

Given this uniqueness and our utter dependence upon it having these attributes, you would think that we would give more thought and study to it. However, if you are like most people that have been through the American educational system, the last and only time that you studied Earth was some time during the Eighth Grade. Before that time, you did not receive much instruction in science other than some physical and life science. After that time, you were probably exposed to the classic physical science/biology/chemistry sequence most people get in high school. This common practice drives geologist crazy, as many of them cannot understand how so many people can live on a planet without knowing too much about it.

Over the next several chapters, we are going to be studying energy sources that come from the solid portion of Earth. In order to make sure that everyone has the same understanding of this, we will quickly review this information that you received in Eighth Grade, and quite possibly, build upon it in this chapter.

Interior

The best place to start with our study of Earth is on the inside. For most of human existence, the Earth's interior has remained an utter mystery, as our ability to penetrate it physically is very limited. Many ancient cultures, especially those near active volcanoes, believed that the interior of Earth was extremely hot with various caverns. This image of the interior persisted until quite recently. Sir Edmund Halley (of Halley's Comet fame) proposed a hollow section of Earth as recently as 1692 in order to account for the low density that he believed the planet had. His idea maintained some support among scientists well into the 1800's. Today, we know that these ideas are incorrect, as the interior of Earth is solid. Later, we will discuss how we know this, but for now, let us state what our current understanding of the interior of Earth is and how it operates.

Fig. 1: Sir Edmund Halley

During the first billion years of the existence of our Solar System, our Earth was able to accumulate mass from accreting smaller planetismals, which were asteroid-like objects of rock, metal, and ice. The combination of the mass, the force of gravity, radioactive decay, and the igniting Sun caused this early Earth to pass into a molten state. This fluid nature allowed heavier elements to be drawn more toward the center of the planet,

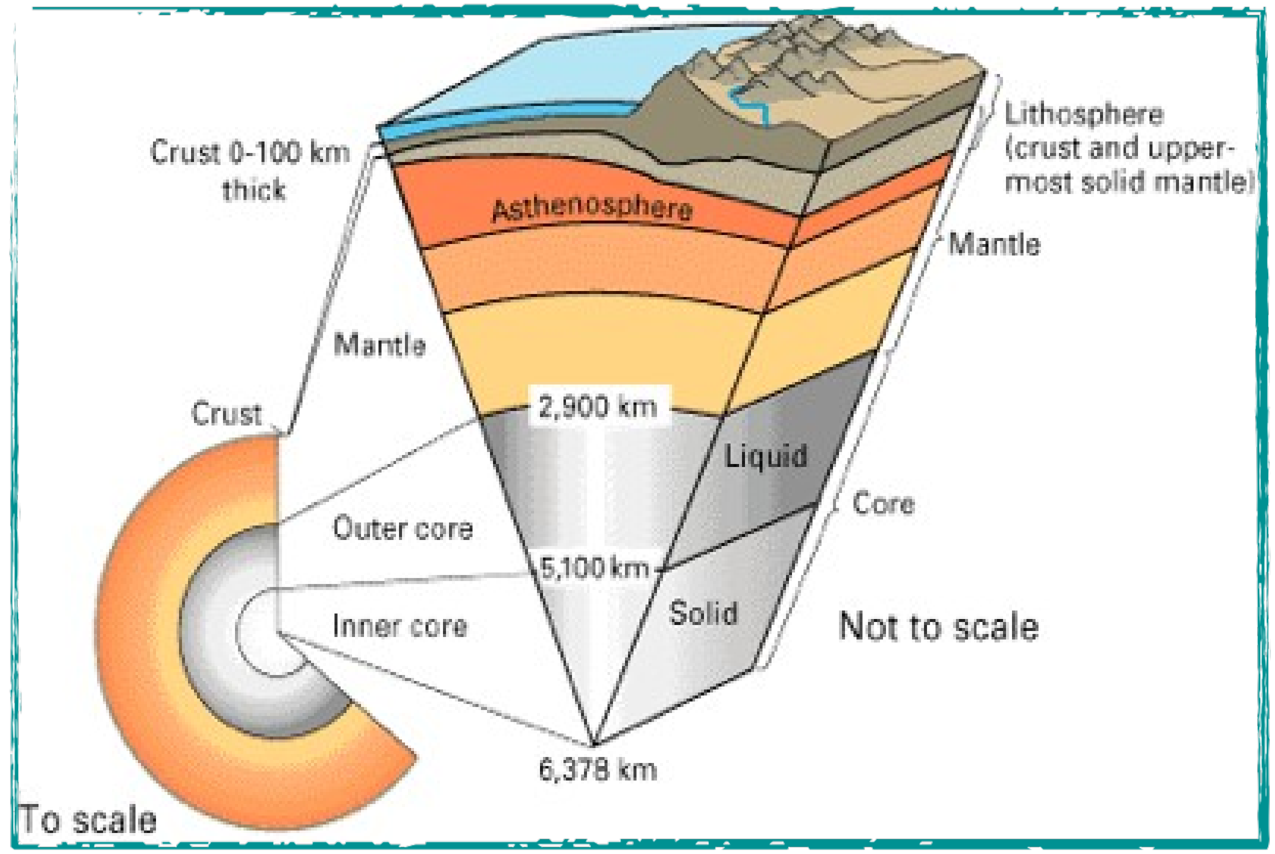
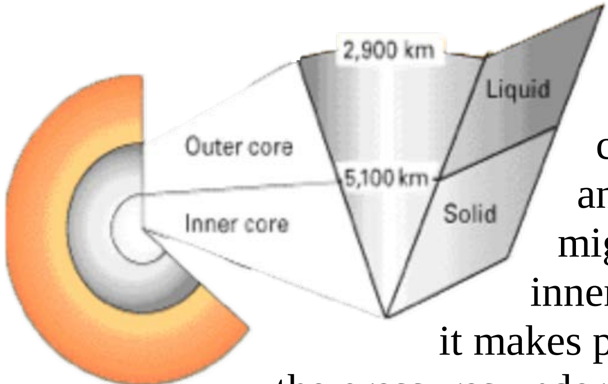


Fig. 2: cross sections of Earth

displacing lighter elements to the regions furthest from the center. As Earth cooled and began to solidify, this segregation led to three distinct layers within Earth based upon the density and composition of the materials. While there is not much exchange of materials between these layers today, there is an interplay of energy that connects all three and creates a very dynamic planet.

Core

The innermost section of Earth, the core, is comprised mostly of nickel and iron. This section extends from the center of the planet out to about 3,500 kilometers from the center. While this is over half of the radius of Earth, it is only about 15% of the Earth's volume; however, the density of nickel and iron (10-12 grams/ cubic centimeter) is high enough to significantly increase the overall density of the planet such that Earth has the greatest density of any body in the Solar System (5.5 grams/ cubic centimeter).



The core is not really a single layer, though, as it is comprised of an inner solid core and an outer liquid core. While this might seem counter intuitive, as the inner core is hotter than the outer core, it makes perfect sense when you consider the pressures under which the core finds itself. The inner core has the weight of the outer core and the rest of the planet pressing down on it, which is great enough that the inner core cannot remain as a liquid even at those temperatures. As you move to the outer core, the pressure is reduced, as there is less weight above it. This reduced weight is enough that the iron/nickel material is able to turn to a liquid even though the material is cooler than the inner core.

The presence of an outer liquid core made of iron and nickel is incredibly important to life on our planet. Since the inner and outer core are not rigidly connected to the rest of the planet, they are free to rotate independently of the rest of the planet. This differential rotation rate gives a net circulation of metal with respect to the rest of the planet for long periods of time. As we saw in the last chapter on electricity, this net current of hot metal, which contains free electrons, corresponds to a large electrical current. The large current generates a magnetic field, which we are able to measure very strongly at the surface. This magnetic field extends into outer space, where it shields the planet's surface from incoming high energy ionizing particles. If these particles were allowed to get to the surface, they would cause genetic damage on a grand scale and greatly damage living tissue on this planet. As it is, the magnetic field curves the path of these particles into circles which give off high-energy radiation (creates the Van Allen radiation belts) while they slowly drift toward the North and South magnetic poles of Earth. As they encounter Earth's atmosphere, they ionize the gases to create the aurora borealis and aurora australis.

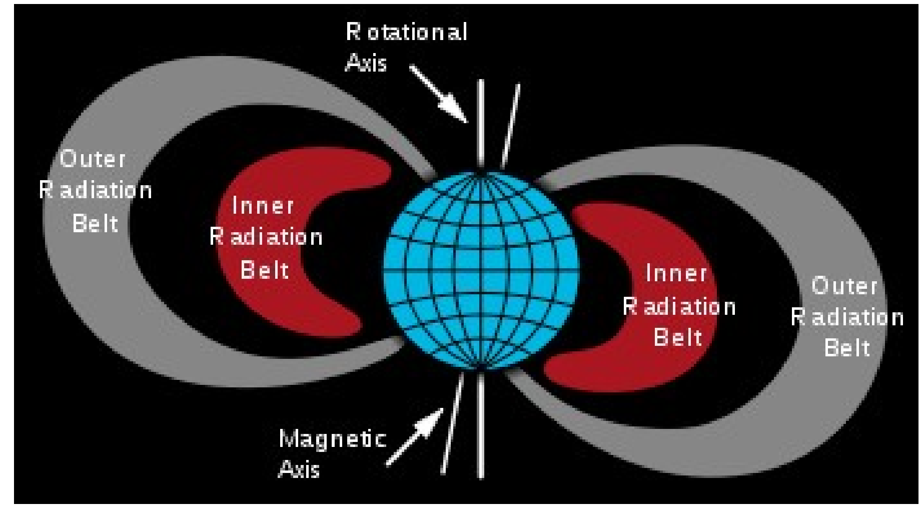
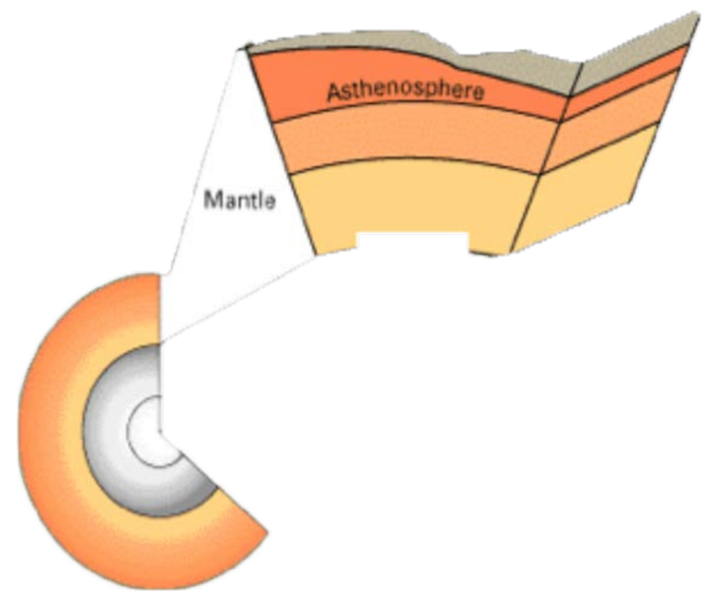


Fig. 3: Van Allen radiation belts

Mantle

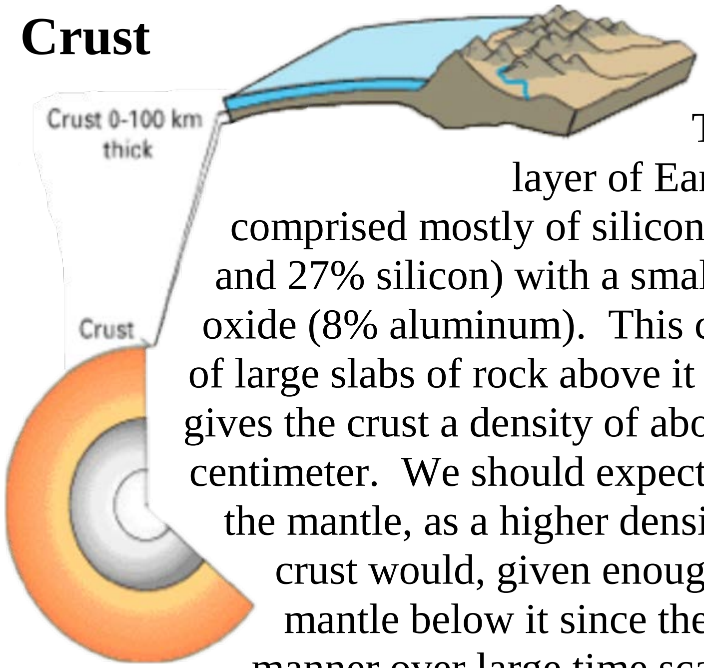


Outside of the core is the mantle, a layer that extends from about 2,900 kilometers to almost 100 kilometers below the surface. This means the mantle comprises about 85%

of the volume of Earth. The composition of the mantle is mostly silicon dioxide and magnesium oxide, with peridotite being the predominant mineral formed. This composition gives the mantle a much lower density than the core, with values close to 3 grams/cubic centimeter near the crust and 5 grams/ cubic centimeter as it nears the core.

The mantle differs from the core in other ways. It exhibits a plastic behavior physically. This does not mean that it is plastic (we just stated that it was mostly oxygen, silicon, and magnesium, not a petroleum product); it means that it behaves like a solid over short time scales, but moves like a liquid over long time scales. If you were able to travel to the mantle, it would seem like solid rock as you touched it. However, if you were to mark it in some way and be able to observe the mark over times scales of millions of years, you would notice that it appears to flow. The best analogy is to candle wax, which is hard enough to be able to mold into shapes, but which will slowly flow over time scales of years and decades such that its bottom starts to bulge outward. This plastic behavior means that the mantle is able to convect heat from the core up to the crust, as it moves over massive time scales. This is important to our discussion of the layer above the mantle: the crust.

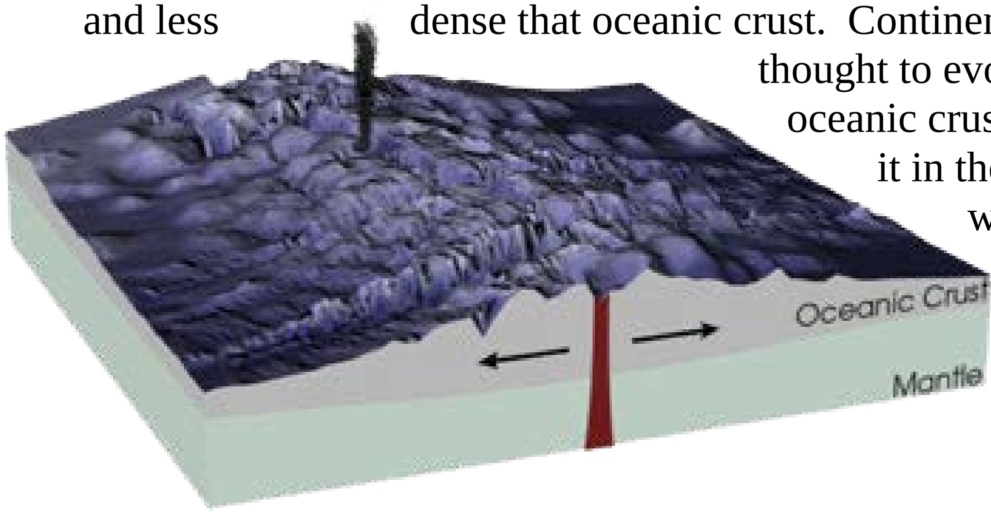
Crust



The outermost and thinnest layer of Earth is the crust, which is comprised mostly of silicon dioxide (over 47% oxygen and 27% silicon) with a small amount of aluminum oxide (8% aluminum). This composition, plus the lack of large slabs of rock above it putting it under pressure, gives the crust a density of about 2.5 - 3.0 grams/cubic centimeter. We should expect a density less than that of the mantle, as a higher density would mean that the crust would, given enough time, sink into the mantle below it since the mantle has a plastic manner over large time scales. This combination of lower density and plastic behavior of the mantle results in the crust literally floating on top of the mantle, which makes it susceptible to the principles that all floating objects obey.

While it is hard to imagine if you have lived in Jonesboro for some time, there is topography to the crust, with mountain ranges that jut almost 10 kilometers above sea level and ocean trenches that sink an equal distance below sea level. Just like with an iceberg or a boat, the only way to have a floating object like a mountain range sticking up so high is if it has a broad base that sinks very far into the mantle. Similarly, the only way to have a trench that is so low is if it has almost no base that sticks into the mantle. Thus, the thickness of the crust is highly variable, with thicknesses on the order of 50-100 kilometers in the continental regions and on the order of 5-10 kilometers under the oceans.

These two different areas have much different types of crust. The oceanic crust is composed mostly of basalt, gabbro, and diabase. It is very poor in silicon dioxide (only about 50% of the rock) and high in chemicals like iron and magnesium. From our discussion of the mantle, you will note that this composition looks similar to that region. Oceanic crust is denser than continental crust, and where the two types of crust are forced together, the oceanic crust will sink below. This chemical composition also means that oceanic crust is dark in color. Continental crust, on the other hand, is comprised more of rocks like andesite, rhyolite, and granite. These rocks are higher in silicon dioxide (over 63%) as well as aluminum, sodium, and potassium, and are lighter in color and less dense that oceanic crust. Continental crust is thought to evolve from oceanic crust by melting it in the presence of water and/or gases like carbon dioxide and then cooling.



Earthquakes

Earth is about 4,000 miles in radius. We have drilled only about 8 miles into it, having not even reached the mantle at this time in history. Therefore, having only physically probed Earth to such a shallow depth, how can we make the claim that this is what the interior of the planet looks like? The answer is seismology, which is the study of how sound waves travel. By listening to echoes and refractions that have passed through Earth, we can determine what the interior looks like, similar to what hospitals do when they perform an ultrasound on a patient. Of course, a planet is much larger than a patient, which means that we are going to need a very loud source of sound in order to penetrate all the way through Earth. Fortunately for the study of the interior, the planet itself supplies such a loud sound in the form of an earthquake.

TABLE 1: HISTORICAL EARTHQUAKES

YEAR	LOCATION	DEAD
1290	Hopeh Province, China	100,000
1556	Shensi Province, China	830,000
1667	Shemaka, Russia	80,000
1737	Calcutta, India	300,000
1908	Messina, Italy	75,000
1920	Kansu Province, China	180,000
1923	Tokyo, Japan	142,807
1932	Kansu Province, China	70,000
1976	Tangshan, China	255,000
2010	Haiti	230,000

Of course, people who are caught in an earthquake do not consider it good fortune. The awesome power unleashed can cause large losses in life and property. Even the planet, which we often think of as solid and unchanging, can be irrevocably scarred by one of these events. In ancient days, the occurrence of an earthquake was often attributed to the anger of the gods who were displeased with some action of mankind. Today, we know that they occur when stresses that have built up in the crust are relieved by large blocks of rock moving very rapidly. However, for all of our knowledge of these disasters, we have yet to be able to accurately predict them.

Of the different types of natural disasters, the one that has the greatest reputation for loss of life and societal damage is an earthquake. History is replete with stories of massive damage from these events. Until very recently, the recording of earthquakes required that a civilization be near the epicenter of the earthquake and that this society have some way of recording it. It is, therefore, not surprising that the earliest recordings of earthquakes that we have today come from China in 1177 B.C. China, which has had a literate, recording society for many millennia, is situated in a location that has many large and lethal earthquakes. Throughout time, they have recorded a large amount of death and destruction, from the 1556 Shensi earthquake that killed 830,000 to the 2008 Sichuan earthquake that killed over 68,000 people. Other countries have also recorded vast amounts of death and damage, as seen in Table 1.

Fig. 4: 1964 Alaskan earthquake

The history of recorded earthquakes in the U.S. is filled with many large earthquakes, but few that rival the loss of life as those already mentioned. In 1811-1812, a series of large earthquakes that have been estimated above an 8 on the Richter scale hit near the area of New Madrid, Missouri. The vibrations from these earthquakes were felt as far away as the New York and Boston. In 1886, a devastating earthquake hit Charleston, South Carolina that killed over 60 people and destroyed almost all buildings in the city. Possibly the most destructive earthquake in the U.S. occurred in San Francisco during 1906. This earthquake caused a series of fires that burned large sections of the city and killed over 700 people.

RICHTER MAGNITUDES	
MAGNITUDE	**DESCRIPTION**
less than 2.0	micro
2.0-3.9	minor
4.0-4.9	light
5.0-5.9	moderate
6.0-6.9	strong
7.0-7.9	major
8.0-9.9	great
10.0+	massive/epic

Most of these large earthquakes occurred before we were able to accurately measure their magnitudes. One of the largest that has occurred since we have been able to measure magnitudes happened in Alaska on March 27, 1964. It had a magnitude of 9.2 and created vertical displacement of over 11 meters high. The focus was situated under the Prince William Sound, and resulted in a tidal wave that killed 110 people in Alaska, Canada, Hawaii, and the continental U.S. It was so large that it caused the water in pools in Texas and Louisiana to slosh.3

Plate Tectonics

In order to understand in better detail the reasons for earthquakes, we must come to a greater understanding of how our crust operates. In 1912, Alfred Wegener, a German meteorologist, proposed what seemed to be a rather preposterous scientific theory. He posited that the crust of the Earth was split up into giant plates of rock that were free to move about on the top of the mantle. This theory, known as continental drift, had several key pieces of evidence to support it. The first of these was that easternmost margins of the continental shelves of North and South America seemed to fit into the westernmost margins of those of Europe and Africa. Furthermore, the geographic features, such as rock layers, mountain ranges, and glacial scoring, lined up if one were to place the continents together as such. The fossil record also showed some anomalies that were easily explained if these continents had been placed together some time in the past. Although there was much to support this theory, it suffered from one great problem: there was no known mechanism at the time that could have driven the rock layers to their locations. It also did not help that Wegener was a meteorologist and not a geologist, as geologist considered him an outsider who was not to be taken seriously.

Fig. 5: Alfred Wegener

After World War II, more evidence was discovered that helped support the idea that the crust was made of moving plates. This evidence came about as the military tried to map out the bottom of the ocean so that submarines could travel safely at depths which are not illuminated. The discovery of the Mid-Atlantic Ridge running down the middle of the Atlantic was a very startling find. This volcanic ridge was almost perfectly in the center of the Atlantic and stood up over 10,000 feet above the sea floor. As scientists probed it, they found an unusual pattern to the magnetic field in the rock on both sides of the Ridge.

Fig. 6: tourists visit the Mid-Atlantic Ridge where it extends above sea level in Iceland

Measurements showed that the magnetic field in the rock shifted from north to south and back to north over and over as one moved away from the Ridge. This variation was symmetric across the Ridge, with slabs of rock that were equidistant on either side have the same orientation.

By this time, scientist knew that the magnetic field of the planet switches from north to south and back over long time scales, as the velocity of the liquid outer core changes over time. The fact that the pattern repeated itself on both sides means that the rock that was equidistant from the Ridge was of the same age. Radiometric dating of rock samples from the region showed that rocks closest to the Ridge are relatively young compared to rocks further away. The only logical explanation for this was that new rock was being created by the volcanoes along the Ridge, and that this rock moved further away from the Ridge as time passed. This means that the crust is coming apart at this region, which gave evidence in support of Wegener's theory.

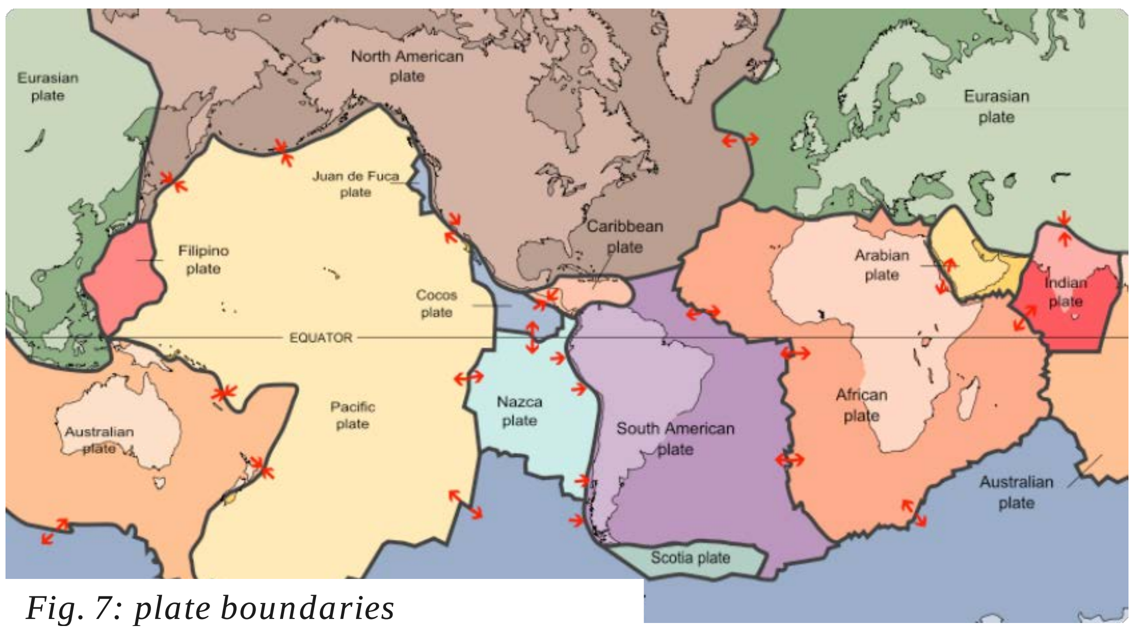


Fig. 7: plate boundaries

Today, we have further evidence to support this theory that has come to be known as plate tectonics. We also have evidence that mantle convection is what is driving this motion. Since the crust is floating on top of the mantle, it makes sense that movements in the mantle would cause the crust to also move. The theory of plate tectonics is now widely accepted and used as a basis for all work in geology and paleontology.

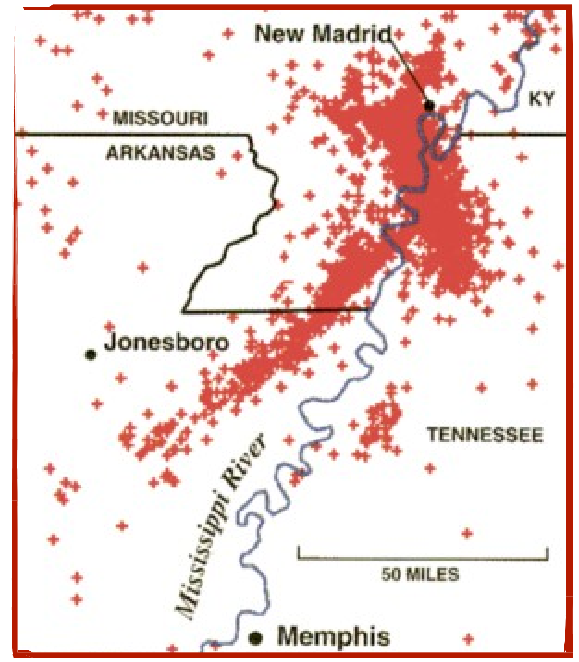


Fig. 8: earthquakes from 1974-2005 along the New Madrid Fault Line

It is this fact that the plates of the crust are moving that accounts for almost all earthquakes. The plates are not greasy pucks sliding around on frictionless ice. They are giant slabs of solid material that are moving past other solid pieces of rock with great deals of friction, stress, and strain. As they move, the plates deform, warp, and stretch, sometimes buckling in two or cracking in place. This means that earthquakes can occur in any location on the Earth. However, they are much more likely to occur at the boundaries between plates, as this is where the most amount of stress and buckling occur. To understand this, let us look at the types of boundaries that there are.

Plate Boundaries

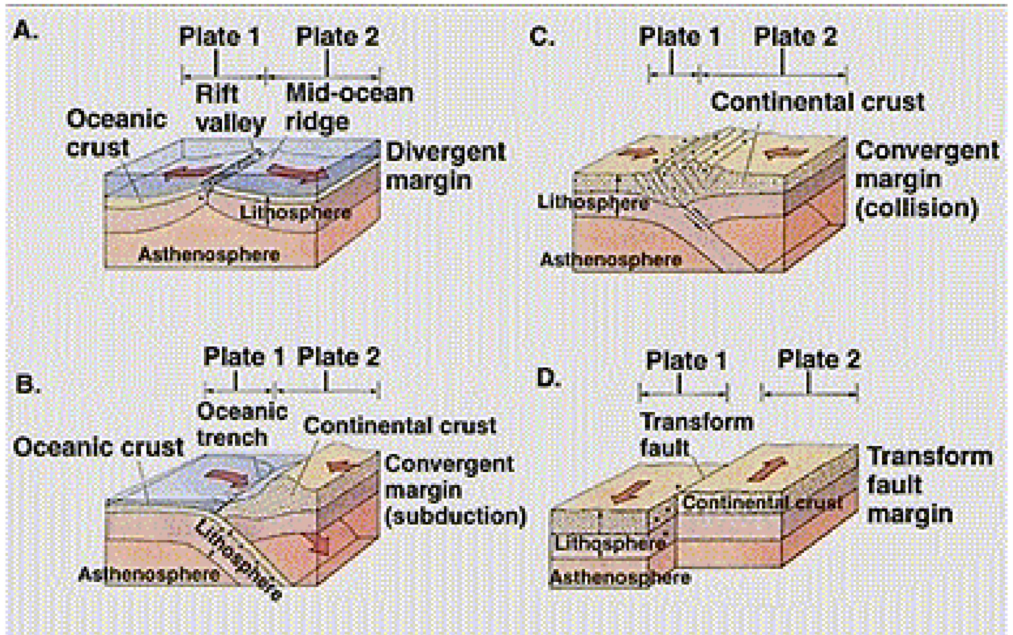


Fig. 9: types of plate boundaries

Since the Earth is not changing in size, the fact that there are moving plates means that there will be three types of boundaries between the plates. There will be boundaries where plates are moving apart, where plates are moving together, and where plates are moving past each other. Divergent zones (Figure 9A) are characterized by large-scale volcanoes and shallow, small magnitude earthquakes. As the plates come apart, molten magma from near the crust-mantle interface is allowed to seep to the surface and fill in the cracks. This magma pushing up causes some stress and cracking in the surrounding rock, thus creating small earthquakes. The Mid-Atlantic Ridge is a perfect example of this type of zone.

Convergent zones are separated into two categories depending upon the types of rocks on either side of the boundary. When continental crust (thicker and less dense) crashes into oceanic crust (thinner and denser), the oceanic crust goes under the continental crust to form a subduction zone (Figure 9B). The oceanic crust that is driven down toward the mantle melts as it encounters higher temperatures and pressure. As this magma is driven up through cracks created in the continental crust, dissolved gases within it begin to expand out of solution due to the reduced pressure. This forces the magma up through the cracks faster, which often results in explosive vulcanism. These zones are characterized by deep-seated earthquakes and stratovolcanoes that form island arc near the continental shelf. Examples of these types of volcanoes exist throughout the entire rim of the Pacific Ocean (New Zealand, Philippines, Japan, Aleutian Islands, Mt. St. Helens). When these deep-seated earthquakes occur with epicenters that are underneath the ocean, they can result in tsunamis, which can travel very far and cause giant tidal waves halfway around the world, reeking damage and killing people.

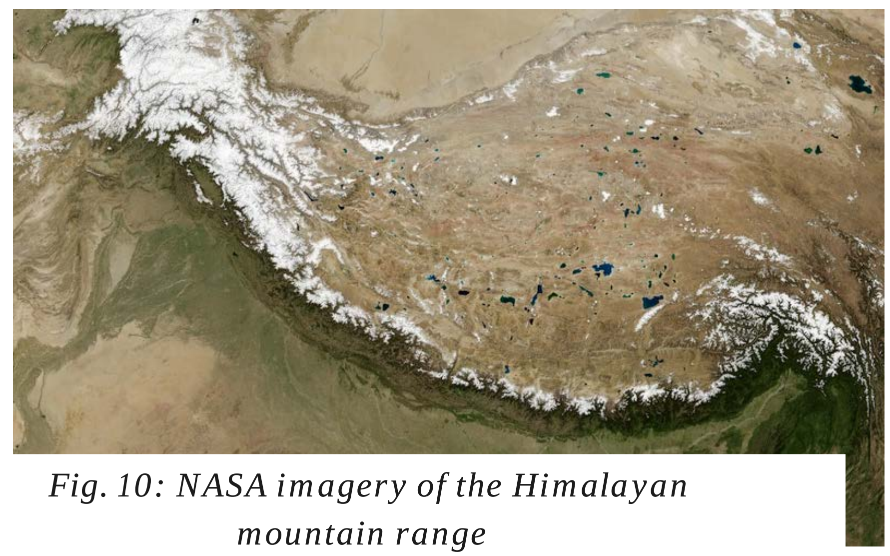


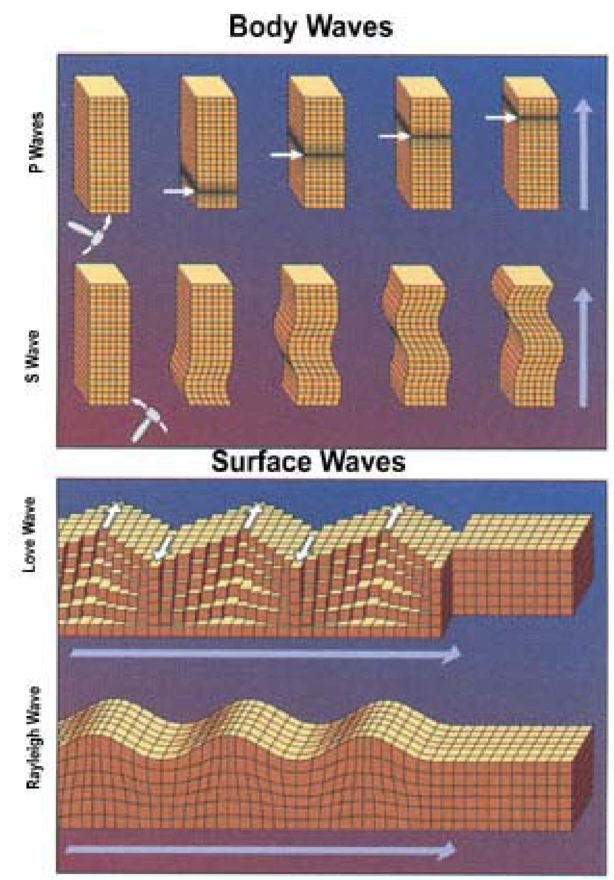
Fig. 10: NASA imagery of the Himalayan mountain range

The other type of convergent zone occurs when continental crust crashes into continental crust (Figure 9C). When this occurs, the two plates fold and crumple at the boundary, uplifting material into mountain chains. This results in few, if any, volcanoes and numerous shallow earthquakes. The best example of this is the Himalaya Mountains that form the boundary between the Indian Plate and Asia. India is very quickly (in geological time scales) burrowing under Asia. As it disappears, more material is piling up and causing the Himalayans to get even taller. Similar processes formed the Appalachian Mountains about 300 million years ago. The difference in the sizes of these two mountain chains is a result of 300 million years of erosion in the Appalachians and the fact that the Himalayans are still being formed.

The last type of boundary is a transform fault (Figure 9D) that will form between two plates that are sliding past one another. This type of boundary has never resulted in volcanoes in recorded history. The most likely explanation of this is that there is no crustal plate melting, nor any exposure to the mantle, with this type of boundary. However, earthquakes are very prevalent in this type of zone, as there is a great deal of friction between the plates that will cause no movement for long periods of time while stress builds up. Earthquakes in this zone are usually shallow. In the U.S., our favorite example of this type of boundary is the San Andreas Fault, which runs down the western coastline of California from San Francisco to Los Angeles. As the North American Plate lurches northwesterly past the Pacific Plate, earthquakes are spawned that cause death and destruction along this boundary.

As previously stated, earthquakes do occur in other locations than plate boundaries. They will occur as stress is relieved slowly over time from the entire plate moving in the direction it is going. Oftentimes, these earthquakes occur at former plate boundaries between the “welded” sections of the plate. For instance, the Ozark and Appalachian Mountains are the site of a former boundary between proto North America, Africa, and Europe over 300 million years ago4. These mountains were formed when the proto-continents crashed into one another and welded together. Because of their history, they are a weak area of the plate. Possibly the most famous intracontinental earthquake zone in the U.S. exists near New Madrid, Missouri, where the famous 1811-1812 earthquakes occurred.

Earthquake Waves



When a violent release of energy occurs because of rock movement, vibrations are generated as the rock moves past each other. These vibrations create waves, which travel away from the focus of the earthquake. Some of these waves only travel along the surface of the earth and are, therefore, called surface waves. Those that travel through the interior of the earth are called body waves, and fall into two different categories: compressional and shear waves. The frequencies of some of these waves are high enough to be heard, while others fall below our hearing level of about 10 hertz. As they travel, they obey all of the principles of other types of wave motion, meaning that they will reflect and refract at boundaries between parts of the Earth that have different wave speeds.

The surface waves are designated as either Rayleigh or Love waves, depending upon whether there motion is vertical or horizontal (Figure 10). These types of waves have the slowest wave speed, but are usually responsible for most of the damage that occurs during an earthquake, as their motion drives the ground up and down, and side to side. Since these waves spread in all surface directions, their amplitudes decrease with distance away from the epicenter, making them truly dangerous only within a hundred miles or less of the epicenter.

Of the two types of body waves, shear or S-waves have the slowest speed. These waves travel in a transverse fashion, as the material through which it travels is moved at right angles to the direction that the wave is traveling. The analog to this type of wave is the type of wave that travels on a plucked, taut string. As the wave travels down the string, the string vibrates back and forth. This is different from the compressional or P-waves, which travel by material movement in the same direction as the wave is moving. To visualize how this wave travels, imagine a Slinky that is placed on a floor and pulled into a straight line. If the Slinky is moved back and forth in the direction of the line that the Slinky makes, a compressional wave will travel down its length.

This difference between P and S-waves in how they travel is very important. In order for S-waves to propagate, they require a fairly stiff material matrix, i.e. they need for the molecules in the material through which they are traveling to be tightly bound to their nearest neighbors. This is so that there will be internal forces that will pull material back to its original location as it is perturbed away from it. Because of this, S-waves cannot travel through a liquid or a gas. P-waves, on the other hand, have no such restriction, and can travel through all types of media (sound waves in air are an example of a compressional wave).

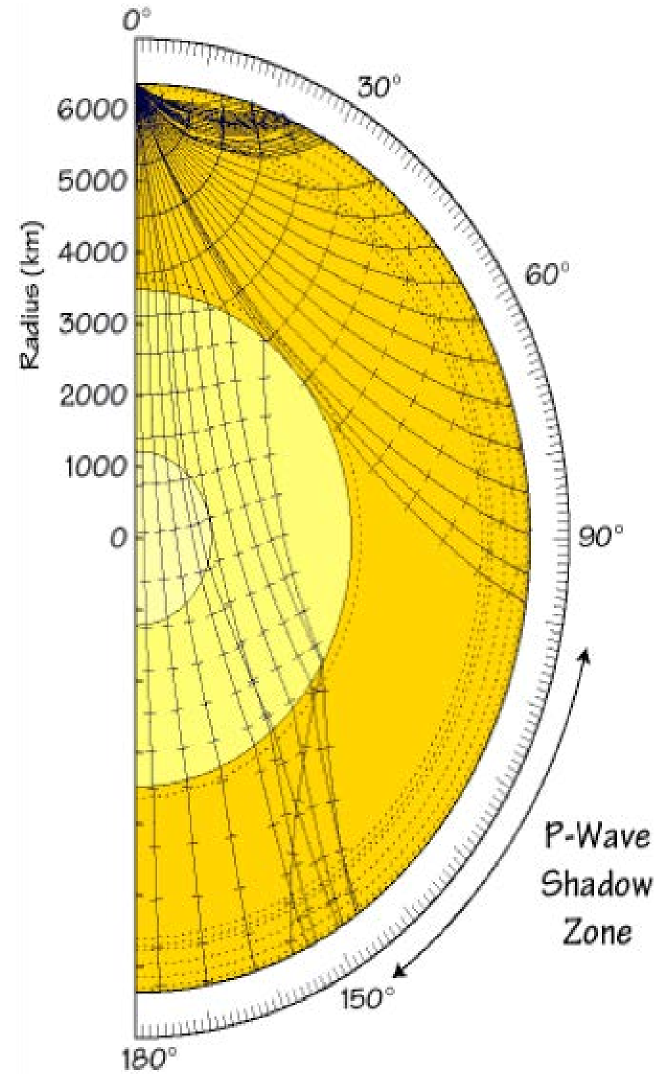
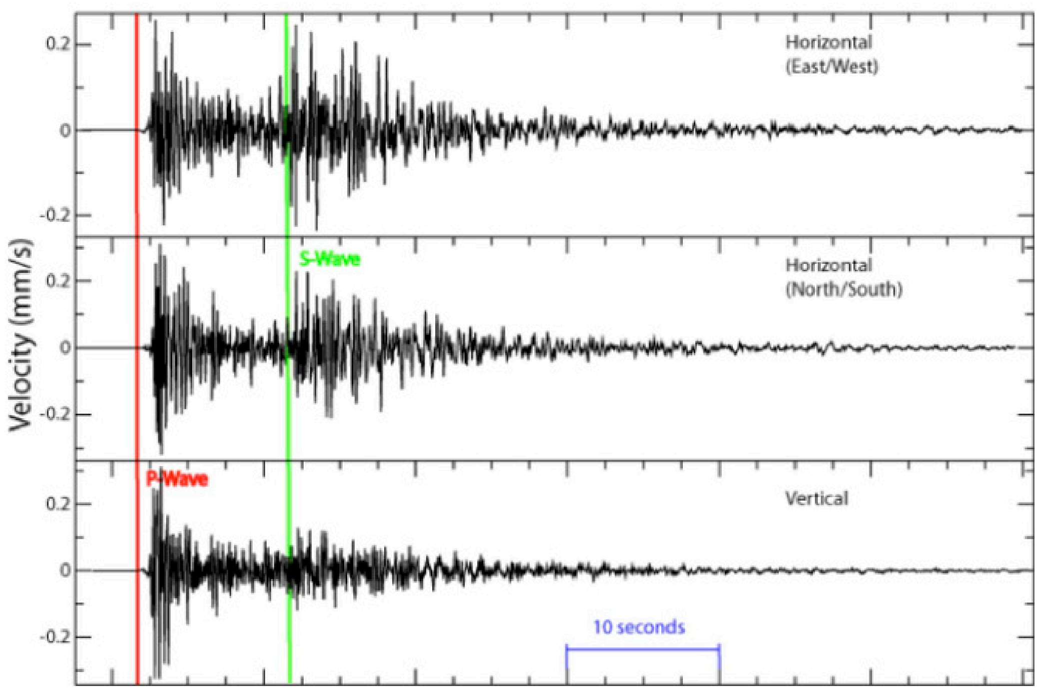


Fig. 11: earthquake shadow zone

It is this difference that allows us to determine that the outer core is liquid, while the inner core is solid. As S-waves impinge upon the outer core, they are not able to travel through it, and only reflect from the interface. P-waves do travel through it, but the differences in velocity between the solid mantle and the liquid outer core cause the P-waves to refract significantly. Once the P-waves get to the inner core, the encounter another change in speed, and reflect and refract from that interface. The P-waves that do pass all of the way through the core and strike the core-mantle interface are refracted and reflected significantly once more. The result of all of this is that S-waves will not be detected on the opposite of Earth from an earthquake, while P-waves will exhibit a shadow zone centered at about 120 degrees from the earthquake because of the big angles of refraction that are encountered at the core-mantle interface.



The detection of these waves is done through the use of a seismograph. As the waves pass by the seismograph, they will jiggle the base of device, which causes some form of detection (pen on paper, piezoelectric pads, etc.). The amplitude of the wave will depend upon the strength of the earthquake and the distance it is from the seismograph, as the wave energy is spread over a greater region as it travels and loses intensity due to reflections. The time that it takes for the waves to reach the seismograph also depends on this distance, as well as the type of wave. The differences in their wave speeds means that P-waves hit a spot first, followed by S-waves and then surface waves.

Earthquake Magnitude

The earliest attempts to quantify the magnitude of an earthquake relied upon assessing the damage near its epicenter. Different numbers were assigned to an earthquake depending upon whether buildings were knocked down, trees toppled over, or things fell off of shelves. The scale that was developed for doing this was known as the Mercalli scale, after its inventor Giuseppe Mercalli. The problem with using this scale to quantify earthquakes was that it was too subjective. Depending upon many factors such as the quality of the buildings, the population density of the region, and the investigator doing the survey, different numbers could be assigned to the same earthquake.

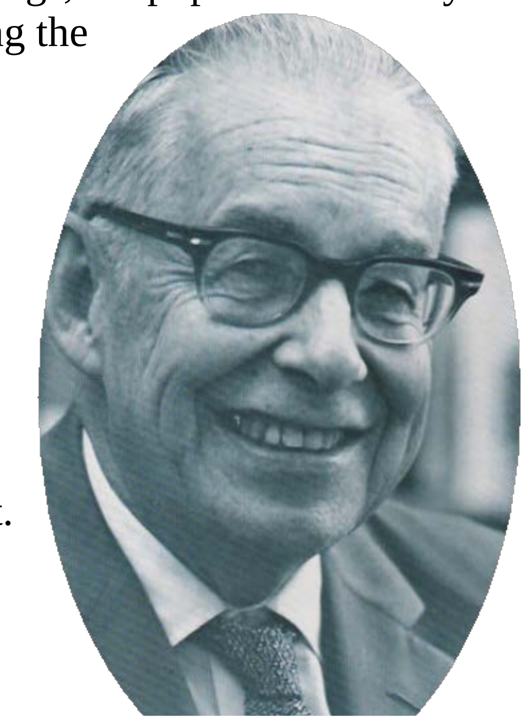


Fig. 12: Charles Richter

In order to get a more objective measure of the magnitude of an earthquake, Dr. Charles Richter developed the Richter scale in 1935. This system assigned each earthquake a number dependent upon the amplitude of the waves generated by it. Since there was such a broad range of values of amplitudes for earthquakes measured, Richter made the scale logarithmic. This means that each increase of one on the Richter scale corresponds to an increase of 10 in the amplitude of the waves. Thus, on the Richter scale, a magnitude-8 earthquake is 10 times greater in amplitude than a magnitude-7 and 100 times greater than a magnitude-6.

While the Richter scale does allow one to objectively measure an earthquake, it still does not give one a sense for the damage that an earthquake can render. For instance, the 1964 Alaska earthquake that we mentioned before was 9.2 on the Richter scale, the second highest value every recorded. While Figure 4 attests to the damage to buildings on the fault line being severe, only 110 people died in this earthquake. This is far less than the numbers of people killed in the earthquakes listed in Table 1, even though each of these earthquakes was probably less than a magnitude 9 on the Richter scale. In order to give one a sense of how much damage is done by an earthquake, scientists have developed the Modified Mercalli scale. Earthquakes are given a letter from Roman numeral I up to Roman numeral XII based upon eyewitness accounts and observations of the epicenter after the earthquake. While not as precise as the Richter scale, its narrowed classification schemes over the old Mercalli scale do allow for a reasonable estimation as to the damage to humans and their buildings.

Rock Types

In our discussion of the crust above, we talked about the forms of rock found in the layer. Mostly, we talked about the chemical constituents of the rock and the subsequent density and color this gave. While the chemical make-up of the rock is very important, it is also important to know how the rock came to be. The chemistry of the rock refers to the minerals in the rock, while the origin refers to the rock type.

In our everyday lives, we often find confusion between the terms rock and mineral. People will sometimes use the terms interchangeably since they are both found in the ground. However, they are distinctly different things. Minerals are solids with a

definite chemical composition and crystalline structure. While rocks can be made of minerals, and therefore have some of these same properties, they can also be made of materials such as volcanic glass that do not contain a single crystal.

The problem is that we do not have a really clear definition of a rock. Different sources will define the term rock in a variety of fashions, with each definition leaving out some outlying versions of rock while including others. For the purposes of this lab, we are going to define rock as any coherent, naturally occurring substance that is usually composed of minerals.

What does this definition leave in, and what does it exclude? Granite and limestone are definitely included, as they are naturally occurring solids that are composed of minerals. Some substances that do not have minerals are also included, such as volcanic glass and anthracitic and bituminous coal. Manmade substances that look like rocks, such as industrial diamonds and concrete, are not included, as they are not natural. Water is not a rock since it is not coherent, but glacial ice would be included since it is. Other people's definition of rock would exclude some of these and include others. Therefore, before you start discussing rocks, it is best to come to some understanding as to definition.

Identifying rocks is sometimes very easy when you have a good sample, but can be quite hard when you have an anomalous rock. Identifying large differences, such as those between granite and limestone, can be quite easy. However, small differences within type will complicate a lot of the process. For example, granite contains a familiar mineral known as quartz. Most people have seen quartz and can identify it quite well when it is in a pure sample: clear to white, six-sided crystal shape, hard. As you start adding impurities into the quartz, it becomes harder to identify it, which makes identifying the granite harder. Like most things that are classified, it is easy to identify those that fit the standard, but it is extremely hard to identify the outliers.

Igneous

Rocks are classified into three large families based upon the rock genesis: igneous, metamorphic, and sedimentary. Igneous rocks are the simplest type to explain. They are created when molten lava cools and hardens. This means that they were the first type of rock to appear on this planet after it finished with its molten stage. They are also the most abundant type of rock on the surface of the Earth, comprising over 80% of the crustal material.

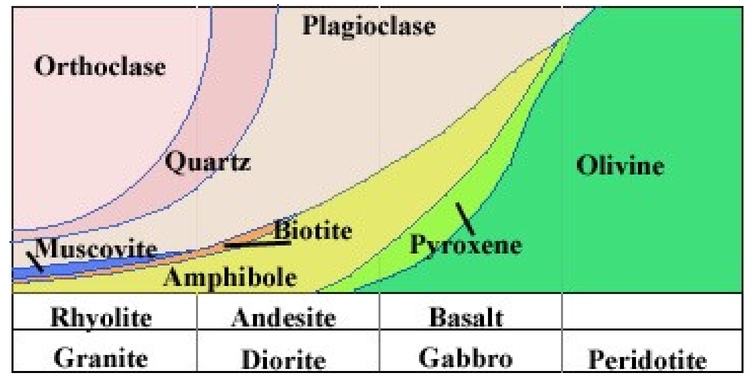


Fig. 13: igneous rock types

Igneous rocks are further subdivided into smaller categories based upon their chemical make up and the location in which they cooled. The chemical backbone of all igneous rocks is silicon dioxide (SiO_2). The different chemical compositions of igneous rock have to do with different metals that will combine with silicon dioxide. There is a broad spectrum of chemicals that will do this. On one end of the spectrum, we have igneous rocks that form from sialic magma that is rich in aluminum and potassium (ex. continental crust). This magma is generally light in color and thick. On the other end of the spectrum are rocks that form from simatic magma that is rich in magnesium and iron (oceanic crust). This magma is generally dark in color and rather thin. In between these two extremes are a variety of rocks that form from magma of varying chemical compounds.

The location of cooling matters to the type of igneous rock since it determines crystal size. If magma cools at the surface of Earth, it will harden very quickly before crystals have a lot of time to form. This type of rock will develop either no crystals (glass) or very small crystals that will require the aid of a lens to see. If the magma cools deep below the surface, it will take a tremendous amount of time to harden, as the surrounding rock will insulate it.

In this scenario, crystals will have a long time to form, which will result in very large crystals in the rock. We distinguish between these two different types by saying that a rock is volcanic (cooled at the surface) or plutonic (cooled below the surface).

Neither of these two factors is really quantized, i.e. both of them can change very slightly over a broad range. However, for classification reasons, we generally place them into a narrow set of groups, as in Figure 13. The three most commonly seen types of magma are rhyolitic, andesitic, and basaltic. Rhyolitic magma is derived from the melting of continental crust. Because it is very viscous, it does not flow well through crack in the crust, and so it rarely makes it to the surface. Thus, rhyolitic magma usually results in granite, which is the plutonic, large crystal version. Andesitic magma comes from the melting of continental and oceanic crust, like that found at a subduction zone. The lower viscosity of this magma means that it reaches the surface more readily than rhyolitic magma. However, it is still thick enough to form steep stratovolcanoes when it does reach the surface. Basaltic magma comes from the melting of oceanic crust and is the thinnest of all of the magmas. Its viscosity means that it usually does make it to the surface, resulting in a small crystal igneous rock called basalt.

Metamorphic

Metamorphic rocks result when pressure, temperature, and chemical conditions produce a change in the crystalline shape or composition of a rock without actually melting a rock. Pressure can squeeze crystals along a plane, causing them to re-form so that the rock looks foliated in that direction. This can cause problems with identification, as this foliation can look like sediment layers that are prevalent in other types of rocks. Increased temperature can give the atoms and molecules in the rock greater kinetic energy, allowing the crystals to grow bigger or change shape without going through a melting phase. Thus, a rock that started out as looking closer to volcanic can look more like plutonic. Changes in the chemical conditions can allow new elements to enter the rock and react, changing the crystals atom by atom. This can even occur when the initial substance was not a rock, such as in petrified wood where the carbon atoms in the wood are slowly replaced with silicon atoms.

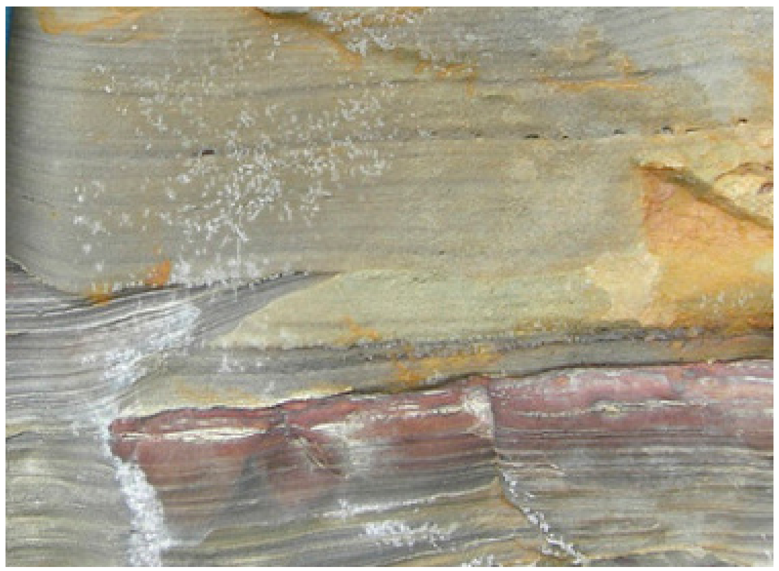
Of course, the amount of change can vary from a small amount to a large amount. If very little change is made, then it is considered low-grade metamorphism. Greater changes result in high-grade metamorphic rock. Besides the amount of metamorphism, the type of rock depends upon what the original rock was and whether the result is foliated or not. The following tables list some of the more common types of metamorphic rock.

FOLIATED METAMORPHIC ROCK		
ROCK	PARENT ROCK	DESCRIPTION
slate	shale	very fine-grained, dense; splits into thin plates with smooth surfaces; harder than shale; dark gray
phyllite	shale, slate	satiny sheen on foliation surfaces due to microscopic mica flakes; often gray or gray-green
schist	shale, slate, phyllite, lava, coal	foliation due to abundance and alignment of elongated or platy minerals, such as micas; quartz, feldspar, garnet, and corundum often present; naed for mineral content
amphibolite	basalts, carbonate-rich muds	black schist composed primarily of aligned amphibole crystals; often shiny on split surfaces
gneiss	any rock or mixture of rock	coarse-grained; characterized by hands of quartz and feldspar separated by schistose layers of micas

NON-FOLIATED METAMORPHIC ROCK		
ROCK	PARENT ROCK	DESCRIPTION
meta-conglomerate	conglomerate	cobbles and pebbles flattened in a parallel arrangement; often high in silica and surrounded by micaceous matric
quartzite	sandstone	composed mainly of fused quartz grains; extremely durable and dense; fractures across original sand grain boundaries; surface less abrasive than sandstone
marble	limestone	recrystallized calcite; reacts with hydrochloric acid; usually massive in structure; white or pastel in color; reflects light on freshly broken surfaces
serpentinite	peridotite, basalt	lime green to dark green; dense; massive; formed by addition of water to simatic rocks; commonly displays slickensides; closely related to asbestos
eclogite	simatic rock	commonly a granular mass of green pyroxene crystals containing scattered reddish garnets; some are foliated; some may be igneous in origin; all products of high pressure
hornfels	any pre-existing rock	dense; rough; fine-grained rock with conchoidal fracture; formed by heat alteration without the addition of new chemicals; when scattered visible crystals are present, name includes these crystals
anthracite	bituminous	black, shiny, hard; low density; will not rub off on fingers

Sedimentary

Sedimentary rock is, quite possibly, the easiest type of rock to understand. It is formed from sediments that are cemented together in order to form a coherent solid. The different types of sedimentary rocks will depend upon what sediments are involved and how they are cemented together. Sediments can come together one of three ways. The first of these is to have smaller pieces of rock come together in some type of depositional environment, such as a lake, a beach, or an alluvial fan. When they do, the type of rock that is formed is known as a clastic sedimentary rock. Sediments can also come together from a dissolved state (ex. chlorine ions in water) when saturation limits are reached and



precipitation occurs. This is known as a chemical sedimentary rock. Finally, plants or animals can pull the dissolved sediments out of solution and later precipitate them as sediment. This will form a biogenic sedimentary rock.

The cement that is used to hold these sediments together is normally provided by some type of dissolution process. For chemical sedimentary rocks, this is quite obvious. For clastics, the cement comes from the local environment. Most depositional environments are local low spots, and thus, are places in which water will collect. As sediment begins to pile up, pressure causes the sediment grains to be pushed very close together. Water, which is one of the greatest solvents known, will pick up some dissolved ions as it sits and moves in between the sediment grains. Some of the dissolved ions will precipitate out and glue the individual grains together.

There are many different features to sedimentary rock. Different sized sediments will form different types of rocks. For instance, sand grains that cement together will form sandstone, while small flat clay grains will form shale. Because organisms that have died can also be moved to the same spot as sediments, one often finds fossils within clastic and biogenic sedimentary rocks. Yearly or seasonal changes in the types of sediments that reach a region will show layering within a clastic rock. These and many more factors give the same type of sedimentary rock vastly different appearances, making rock identification a challenge.

The following tables contains most of the common types of sedimentary rock.

SEDIMENTARY ROCK

ROCK	ORIGIN	DESCRIPTION
conglomerate	stream deposition	rounded particles greater than 4mm, cemented by finer material
breccia	landslides, grinding along faults	distinguished from conglomerate by angularity of the particles
tillite	glacial deposition	frequently a mixture of conglomerate and breccia clasts in a siltstone matrix; no layering
quartz sandstone	dunes, beaches, sand bars	particles 1/16mm to 2mm; cemented together; may be angular or rounded; small grains of clear quartz; gritty feel
arkose	weathered granite	a sandstone containing considerable feldspar particles which appear white or pink, usually angular and may be as coarse as 4mm.
graywacke	weathered basalt	a sandstone containing rock fragments, glass shards, and altered plagioclase embedded in clay matrix
siltstone	stream deposition, nearshore marine	a fine-grained, compact rock intermediate between sandstone and shale
shale	quiet marine, lakes, swamps	particles smaller than 1/256mm; may show closely spaced bedding planes; feels smooth
rock salt	desert dry lakes, marginal marine	crystalline or massive halite; hardness of 2 1/2; commonly white or pale orange

SEDIMENTARY ROCK

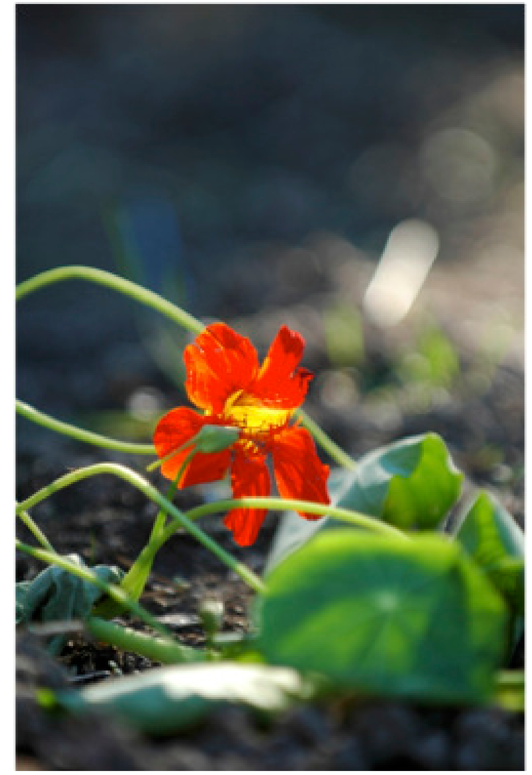
ROCK	ORIGIN	DESCRIPTION
gypsum rock	same as salt	massive or granular gypsum; hardness of 2; white, gray, or pastel
chert	chemical precipitate	massive silica; conchoidal fracture; hardness of 7; dark colored variety is flint; commonly in limestones
limestone	chemical precipitate, marine, hot springs	composed of calcite; hardness of 3; reacts with hydrochloric acid; usually dense, fine-grained, white or gray, some biogenic
dolomite	marine mud flats, deep ocean muds	a form of limestone containing considerable magnesium; reacts sluggishly with cold hydrochloric acid
coquina	beaches, reefs	visible calcite shell fragments loosely cemented together
chalk	quiet marine	white, soft; form of limestone; reacts vigorously with cold hydrochloric acid
diatomite	marine	formed by the deposition of microscopic diatom shells made of silica; white, low density
bituminous coal	swamps	dark brown or black; low density; partially decomposed remains of land plants; often banded

Soil

Having a solid surface is very important for life on Earth. However, if rock were the only form of solid available, we would be in trouble, as plants would have a very tough time growing. For this, we need soil, which makes understanding soil very important to our studies. First, though, we need to define soil to differentiate it from another popular term, dirt.

Soil or dirt: what's the difference? Depending upon whom you ask, you might get a radically different answer. Some sources state that the only difference between them has to do with their location: soil is the unconsolidated material on the ground, dirt is that same matter on your hands or clothes, and sediment is the same material on the bottom of a river or lake. Others define the differences based upon the size and shape of the material grains. For the purposes of this book, we are going to define things the following ways. Soil is a complex, unconsolidated mixture of inorganic, organic, and living material that is found on the immediate surface of the ground that supports plant life. Dirt is any fine-grained, unconsolidated mixture that comes from the ground. Sediment is granular material that has been eroded by the forces of nature. Thus, soil can be considered dirt, and it can consist of sediments, but dirt and sediments are not necessarily soil.

Fig. 14: soil runoff near a fresh clearcut



It is this last part of the definition of soil that is so important to us. Without soil, there would be no plant life on the surface of the land. Without plant life, we would not exist. We need it for food. We need it for oxygen. We need it for clothing and shelter. We need it for energy. There is a vital interplay between soil and plants, which makes it vitally important that we understand the fragile nature of soils.

Given this important role, it is amazing that we often treat soil like dirt. We strip away the overlying vegetation to plant crops or build houses, exposing it to erosional forces that wash and blow it away. We pour toxic herbicides and insecticides on it, removing or changing the important organisms that make the soil what it is. We irrigate it with water bearing minute traces of salt, slowly killing the soil as the salt concentrations build up to lethal quantities.

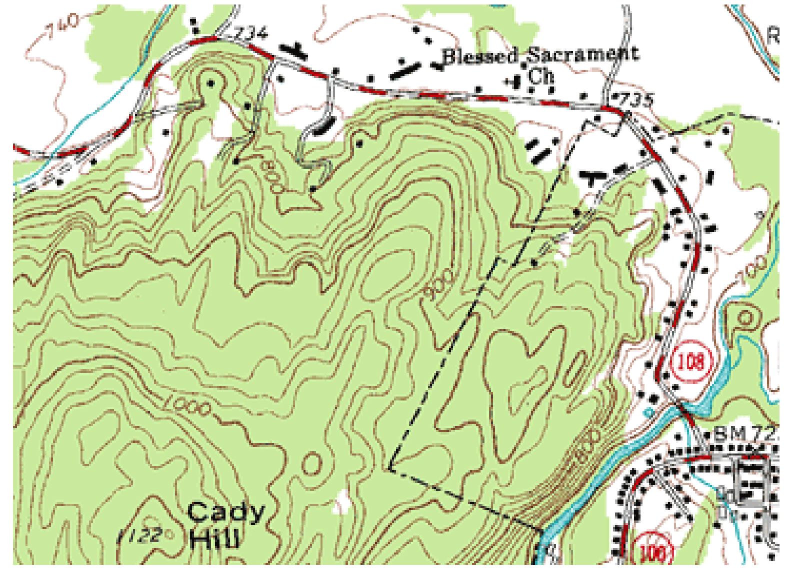
Soil Forming Factors

The complex nature of soil means that the type of soil that a region has depends on many factors. To see this, just look at the soils you find in different locations. Is the soil in the desert like that in a rainforest? Is the soil in Arctic regions like that near the equator? What about the soil on a mountain compared to that in a valley? These locations all have different soils because of a myriad of factors. There is a significant amount of interplay between these factors, though, that complicates the process of soil creation. To see this, let's look at what some of the more important factors are.

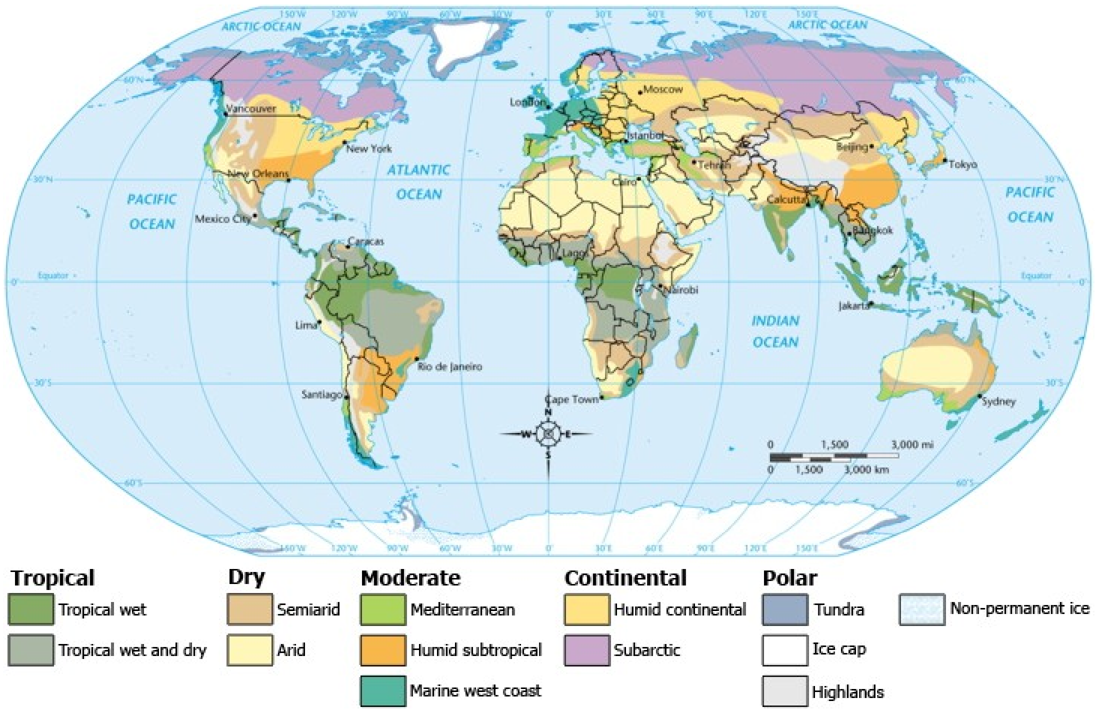
One of the most important factors is the type of rock from which the inorganic material in the soil originated. The parent rock provides the soil with a great deal of the chemical backbone for the soil. For instance, a soil that contains sediment from limestone will be high in calcium, and will also have a basic pH. This will be much different than one that derives from granite, which will have a higher sodium or aluminum content and a pH that tends more toward neutral or acidic.

The type of parent rock will also affect the soil in other ways. The grain size of the inorganic materials is also affected by the parent rock's hardness, fracture characteristics, and the crystalline structure of any minerals in the rock. Rocks containing mica will produce small, flat grains like that found in clay, while rocks with a lot of quartz might produce rounder grains like that found in sand.

Of course, the parent rock for the soil might not be a local rock, though. This is especially true in our region, as there are no local rocks. The reason for this is because of another important factor, topography. Areas that are steeper will be more susceptible to mass wasting and erosional forces. This will remove much of the smaller grains of inorganic material, leaving behind only the larger variety. The smaller grains will be moved to places where the land is not as steep, such as a valley or an alluvial plain. This means that the parent rock for a local soil can actually be hundreds or even thousands of miles away.



Topography does not work alone in this regard. The climate of the region will also greatly affect this sedimentation process in several ways. Before the parent rock can become part of the soil, it must be broken down either physically or chemically. Rain,



especially exceptionally acidic rain, will leach elements out of a rock, which moves them into the local soil as well as weakening the chemical bonds within the rock, making it more likely to fracture. Water and wind flowing across the rock can also physically breakdown the rock, as can ice that expands in cracks and pore spaces. The rain and wind also operate as erosional forces to move the weathered sediments to new locations.

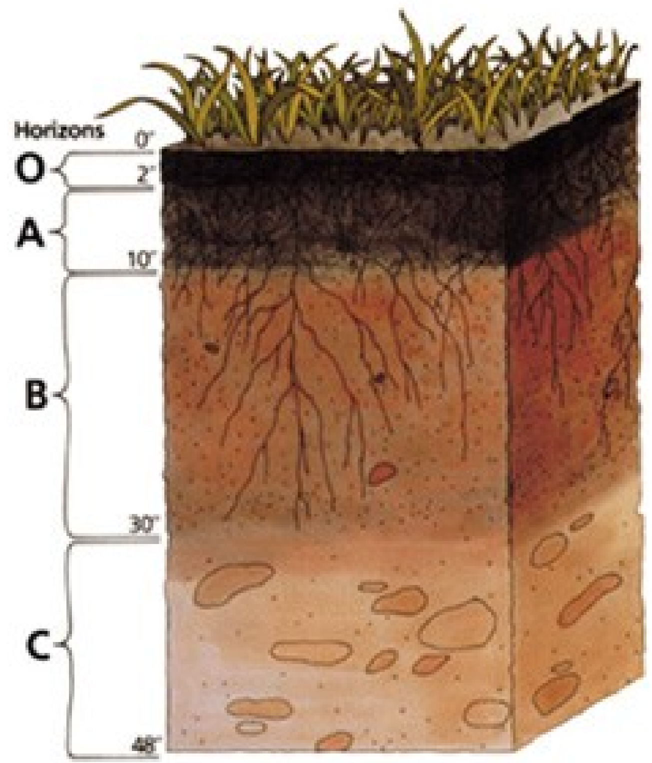
But climate does not only affect the soil by weathering and erosion. The types of organisms that live on and in the soil also depend on the climate. Cacti do not grow well in cold locations that get a lot of rain, and worms have a tough time eating their way through frozen tundra. These biotic factors are very important to the soil, which, after all, is a mixture of inorganic, organic, and living matter. The waste products and dead remains of these organisms provide the organic material for the soil. These organisms also hold the soil in place, covering and protecting it from erosional forces that would strip material away. They might also help further break down the inorganic material in the surroundings, either chemically (acids from the organisms, such as pine sap) or physically (roots from a tree fracturing rock).

All of these factors depend on an even greater one: time. A soil does not just spring up overnight. Nor does it remain a constant. It takes time for all of these factors to play out and bring a soil to maturity. Imagine a region of freshly exposed rock. Depending upon the type of rock, it can take anywhere from a couple of years to thousands of years before any reasonable concentration of sediments begins to pile up. The organisms that can live on this small amount of fresh sediment might do well initially, but as a soil begins to take shape, it might become more advantageous for other forms of life to inhabit it. As the weathering and erosion proceed, the topography of the area will change, which might even change the climate as weather patterns might shift.

Soils are constantly changing. However, before mankind became ubiquitous, this change was often gradual and slow. Today, human activity has become one of the greatest factors affecting a soil. As stated before, we move large sections of sediments, we irrigate soils that never had water, and we change the organisms that live on and in the soil. We make molehills out of mountains, and vice versa.

Soil Horizons

If you were to begin digging into a mature soil, you would notice that the color, texture, and other properties of the soil changed as you went deeper. If you were to dig deep enough, you would see that the soil appeared to be in very distinct layers. These layers, known as soil horizons, occur because of the different chemical and biological processes that take place in these zones.



Depending upon the type of soil, there can be up to 5 different horizons. These are denoted by the letters O, A, B, C, and E. Not all soils will have these horizons, with some immature soils having none. Most soils have at least three of these (A, B, and C).

If it is present, the top horizon will by the O layer, which is comprised of organic matter. This layer is normally found in forest soils, where dead leaves and other detritus can build up on a yearly basis. Below the O layer will be the A horizon, which is where the organic material is mixed in with the inorganic material. This layer is usually darker in color, and if present, means that the soil will generally be fertile for plant life. In a forest environment, there will sometimes by an E horizon below the A that is a result of water becoming acidic as it passes through the O and A horizons and then leaching minerals out of the soil. Below this horizon if it is present, or below the A if it is not, is the B layer, which is where the minerals and clay grains accumulate. In some regions, this layer can be very thick and tightly pored, resulting in hardpan that can very effectively impede the flow of water through it. Below the B horizon is the C, which contains the parent inorganic material for the soil. It is little affected from the original soil before it matured.

Types of Soil

In 1975, the U.S. Department of Agriculture created a taxonomic scheme that grouped all soils into one of 12 major groups known as orders. Underneath these orders are smaller groups and sub-groups. In all, there are over 100 different types of soil that have been classified by scientists. We do not have the time or space to list them all here. For our purposes, we will stick to the 12 orders of soil[3].

Alfisols are a well-developed, highly fertile soil that forms in forests. These soils have undergone some leaching (water stripping some chemicals from the soil as it percolates through it), leaving them with a subsurface layer of clay. This clay allows these soils to remain moist, which helps to keep them fertile. They are usually found in temperate zones, which makes them ideal for farming. In the U.S., a wide swath of land from south Texas up through Michigan is composed primarily of this soil type.

Andisols are formed from the ash and ejecta from volcanoes. These soils are very high in glass grains and materials with pores. This latter property means that these soils have a high ability to hold water. In the U.S., these soils are found mostly in the Hawaiian Islands and the Northwest near the active and recently active Cascade Mountains.

Aridisols, as the name implies, are soils that form in regions that are dry for long periods of the year. These soils have a high calcium carbonate concentration, with layers of clay, silica, salt, and gypsum in the subsurface regions. They make up about 8% of the land in the U.S., existing mostly in the desert Southwest.

Entisols are characterized by being fairly new soils, which means that they have not had much time to develop any horizons or substrata. They are usually found on rocky hillsides, but they can also be found on river deltas and shorelines. Essentially, all soils that do not fit into one of the other 11 orders are classified as an entisol. In the U.S., the greatest concentration of these soils is found in the Rocky Mountain and High Plains regions.

Gelisols form in locations that have permafrost within 2 meters of the surface. This means that they are limited to regions near the Poles, or in high, mountainous zones. Because organic matter does not decompose at a fast rate in such areas, gelisols contain a lot of carbon material. In the U.S., these soils are found mostly in Alaska.

Histosols are soils that contain at least 20-30% organic material and are more than 40 centimeters thick. They are found in locations where the presence of water prevents organic matter from decomposing quickly. This means that histosols are commonly found in low-lying swampy regions. The Gulf Coast and Great Lakes regions contain most of the histosol soils in the U.S.

Inceptisols are slightly more mature versions of entisols. They have begun to develop soil horizons, but have none of the features found in the 10 other orders. As with entisols, they are found mostly in mountainous or hilly regions, such as in the Rocky and Appalachian Mountains.

Mollisols are found in grassland areas and have a relatively rich, dark-colored surface zone as a result of the organic matter from the being added from the grass. The fertile nature of these soils makes them excellent media for growing grain crops. The Great Plains region of the U.S., the Breadbasket to the World, is an example of this type of soil.

Oxisols are the heavily oxidized soils found in tropical and subtropical rainforest. These soils have undergone heavy amounts of weathering and are very low in fertility outside of the very thin layer of organic material on the surface. Because water has leached most of the other minerals out of the soil, oxisols are very high in concentrations of aluminum and iron and are mined extensively in countries where rainforest are being chopped down. In the U.S., oxisols are limited to regions of Hawaii.

Spodosols are highly acidic soils that form in coniferous forests. Water leaching through these soils becomes acidic, which causes heavy weathering of the lower horizons in the soil. Much like the oxisols, this leads to high concentrations of aluminum and iron in the subsurface horizons. These soils have a very low fertility for any kind of crop other than trees that favor acidity like pines. The Northeastern U.S. and the Upper Peninsula region of Michigan are where these soils are found in the U.S.

Ultisols are very similar to spodosols in that they are highly leached, acidic soils in forest environments. They are characterized by a subsurface clay layer that is high in iron, and are found in temperate and subtropical zones that receive a fair amount of rain. The southeastern region is the primary location for these soils in the U.S.

Vertisols are clay-rich soils that experience swelling and shrinking depending upon the water content. These soils are an engineering nightmare, as the total volume of the soil changes depending water. They are limited in the U.S. to sections of Texas and the Delta region of Mississippi.

Problems

1. From what location within the interior of Earth does the lava that comes out of volcanoes originate?

2. What is the difference between continental drift and plate tectonics?

3. Why does the topsoil in the Jonesboro region have such a thick layer of clay several inches below the surface?

4. What type of rock constitutes the majority of Ozark Mountains? What kind of processes created it? What does this say about the geological history of the region?

5. Are erosion and weathering the same thing?

6. Anthracite coal is considered a rock. Given that coal starts out as plant material, what type of rock is it?

7. Why are the oldest rocks on the surface of Earth only about 3.3 billion years old if the planet is thought to be over 4 billion years old? Why are such rocks hard to find?

8. The world's largest earthquakes, as determined by the Richter Scale, rarely produce the largest amount of death and destruction. Why?

9. The rainforests of Brazil are an excellent spot to mine for aluminum (unfortunately). Why?

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2. The World Book Encyclopedia, Volume 6, p. 19, Field Enterprises Educational Corporation, Chicago, 1974.

http://neic.usgs.gov/neis/eqlists/USA/1964_03_28.html, September 13, 2003.

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Appendix A

The following tables are from General Geology of the Western United States - A Laboratory Manual by Bassett and O-Dunn, pp. 6-18, Peek Publications, Palo Alto, CA, 1980.

MINERALS WITH METALLIC LUSTER				
Name & Composition	Hardness	Color	Streak	Features
Graphite C	1	silver-gray	black	marks paper, greasy feel, light in weight, one perfect cleavage
Molybdenite MoS_2	1 - 1 1/2	bluish-gray	greenish-gray	soft, flexible, shiny plates (one perfect cleavage); often with hexagonal outline, marks paper
Galena PbS	2 1/2	silver-gray	black	cube or octahedron cystals, cubic cleavage, bright luster, heavy
Native Copper Cu	2 1/2 - 3	copper-rose	copper-rose	greenish-gray surface film where altered; heavy and malleable; rare crystals, usually in compact masses, often has a pale green surface coating of malachite
Native Gold Au	2 1/2 - 3	gold, white-gold, rose	gold, white-gold, rose	color varies with impurities; extremely heavy; may be gouged or sliced with a knife; dissolves in aqua regia; rare small crystals and dendrites, nuggets in sedimentary deposits
Native Silver Ag	2 1/2 - 3	silver-white	silver-white	tarnishes dark gray or black; irregular fracture; very heavy; sectile; may occur as dendrites (see Gold) and wires in calcite and other minerals
Bornite Cu_5FeS	3	rose to brown	gray-black	iridescent alteration coating common; brittle conchoidal fracture; “peacock ore”
Chalcopyrite $CuFeS_2$	3 1/2	brass-yellow	greenish-black	often tarnished iridescent or chalky greenish-blue; brittle, fairly soft; usually massive; conchoidal fracture
Pyrite FeS_2	6	light brass-yellow	black	occurs in cubes with grooved faces, and pyritohedrons with 5-sided faces; called “fool’s gold”; much lighter than true gold; poor cleavage; fragile
Magnetite Fe_3O_4	6	black	black	magnetic, granular or octahedral crystals common; no cleavage
Specular Hematite Fe_2O_3	6	shiny steel-gray	dark red	glittering flakes or wavy sheets; streak is distinctive; tendency to flake obscures true hardness

MATERIALS WITH NON-METALLIC LUSTER				
Name & Composition	Hardness	Col	Streak	Description
Talc Mg3Si4O10(OH)2	1	white, pale	pearly	extremely soft, soapy feel; impurities may increase hardness; one perfect cleavage; often in scaly masses
Kaolinit e	1 - 2 1/2	white, cream	earthy, dull	soft, powdery texture; smells earthy when damp; usually in clay-like masses with dull appearance
Native Sulfur S	1 1/2 - 2 1/2	yellow	resinous, greasy	low hardness; light in weight; detectable sulfur odor; often in well-developed blocky crystals, or as fine coating on volcanic rock
Gypsum CaSO4 ~2H2O	2	colorless, white, sometimes	vitreous to pearly	soft; one perfect cleavage; selenite is clear and may occur in large (1 m) sword-like crystals, or in bladed groups incorporating sand and known as “desert roses”, satin spar is fibrous, alabaster is massive
Borax Na2B4O7 • 1OH2O	2	white	vitreous	short, stubby crystals; conchoidal fracture; brittle, soft; also in earthy, massive forms
Chlorite	2	light to dark green	vitreous to earthy	micaceous habit; one good cleavage; flakes are not elastic like mica
Carnotite K2(UO2)2(VO4) 3H2O[2]	2	canary yellow	dull, earthy	usually a coating or powder in sandstone or other rock; imparts a strong yellow color; vary radioactive; hardness indeterminate
Cinnabar HgS	2 - 2 1/2	cinnamon red	adamantine to dull	color diagnostic; may appear almost metallic or in earthy, pinkish-red masses; scarlet streak; toxic
Biotite Mica K(Mg,Fe)3 AlSi3O10 (OH)2	2 1/2	dark brown, black	vitreous	occurs in six-sided mica “books” and as scattered flakes; peels into flexible greenish-brown sheets along one perfect cleavage; black mica
Muscovite Mica KAl2(AlSi3 O10) (OH)2	2 1/2	colorless, pale	vitreous to pearly	occurs in six-sided mica “books” and as scattered flakes; peels into thin transparent sheets along one perfect cleavage; white mica
Lepidolite Mica KLi2(AlSi4O10) (OH)2	2 1/2 - 4	colorless, lilac, yellow	vitreous to pearly	lilac color is diagnostic; often in granular masses of small mica “books”; lavender mica
Halite NaCl	2 1/2	colorless, salmon,	vitreous to greasy	easily dissolves in water; often has stepped-down “hopper” faces; cubic cleavage; crystal masses or coating on other material

MATERIALS WITH NON-METALLIC LUSTER				
Name & Composition	Hardne	Color	Streak	Descriptio
Asbestos $Mg6Si4O10(OH)8$	2 1/2 - 3	light green, light brown	silky	long, thread-like fibers with silky sheen; the commercial variety is fibrous serpentine
Calcite CaCO4	3	colorless, white (rarely pastel)	vitreous	effervesces freely in cold dilute hydrochloric acid; perfect rhombohedral cleavage; doubly refracting; frequently in rhombohedral crystals; hundreds of other forms known, may be fluorescent
Barite BaSO4	3	colorless, white, blue	vitreous	heavy for a non-metal; often occurs as tabular crystals; such crystals in circular arrangement form “barite roses”; perfect cleavage
Bauxite	3 - 3 1/2	white, usually stained with goethite	earthy	pea-sized round concretionary grains show color banding in cream, yellow, and brown; actually a rock made up of various hydrous aluminum oxides
Sphalerite ZnS	3 1/2	usually yellow-brown; also black, green, red	adamantine to metallic	light yellow streak in most colors varieties; heavy; perfect dodecahedral cleavage; cleavage chunks often triangular in shape; occurs as crystals, compact masses, and coatings
Azurite Cu3(CO3)2(OH)2 **Malachite** Cu2CO3(OH)2	3 1/2 - 4	azure blue bright green	dull velvety	colors and association distinctive; both effervesce in hydrochloric acid; azurite often in radiating masses; malachite frequently in curved masses exhibiting color banding in shades of greens
Dolomite CaMg (CO3)2	3 1/2 - 4	white, yellow, pink	vitreous to pearly	slowly effervesces in cold dilute acid when powdered; pale pink color is indicative; often associated with calcite; usually in rhombohedral crystals; perfect rhombohedral cleavage
Fluorite CaF2	4	colorless, all pastels; deep purple	vitreous	crystals often cubic or octahedral; color banding common; octahedral cleavage; usually fluorescent in ultraviolet light
Colemanite Ca2B6O11 ~5H2O	4 1/2	colorless, white	vitreous	may be in stubby glassy crystals, or in compact granular masses; perfect cleavage
Apatite Ca5(PO4)3F	5	white, blue, brown	vitreous	will not scratch glass; commonly in 6-sided prisms; green, blue, yellow; one poor cleavage

MATERIALS WITH NON-METALLIC LUSTER				
Name & Composition	Hardness	Color	Streak	Description
Scheelite $CaWO_4$	5	white, yellow, brown	vitreous	will not scratch glass; heavy; fluoresces; good cleavage; crystal faces may be grooved
Goethite $HFeO_2$	5 - 5 1/2	dark rusty brown, ochre yellow	dull, earthy	streak distinctive yellow-brown; often spongy, porous or earthy; bladed, fibrous; often occurs in cubes and pyritohedrons as an alteration or pyrite; also called limonite
Hematite (earthy) Fe_2O_3	5	dull brownish red to bright red	sub- metallic to earthy	characteristic red-brown streak; often earthy and too powdery for accurate hardness test; may be granular or oolitic; crystals rare; no cleavage
Rhondonite $MnSiO_3$	6	pink to deep rose	vitreous	massive, dense, or granular aggregates often have black veins; color and hardness diagnostic; blocky crystals; nearly 90° cleavage
Hornblende	5 1/2 - 6	greenish-black	vitreous	barely scratches glass; shiny on cleavage faces; opaque; often splintery at edges; usually massive, occasionally in chunky crystals; two directions of cleavage at 124° and 56°
Augite	6	dark green	vitreous to dull	stubby prismatic crystals; usually duller and greener than closely related horn blend; two cleavages at 87° and 93° and uneven fracture
Orthoclase Feldspar $KAlSi_3O_8$	6	white, pink	vitreous	two good cleavages; will scratch glass; wavy internal pattern and pink color distinguishes it from plagioclase when present; may be massive or in large, well-developed, coffin-shaped crystals
Plagloclase Feldspar $NaAlSi_3O_8$ $CaAl_2Si_2O_8$	6	white, gray	vitreous	two good cleavages; will scratch glass; “record grooves”; rectangular cleavage faces often seen in igneous rock
Spodumene $LiAlSi_2O_6$	6 1/2	colorless, white, lavender	vitreous	elongated prismatic crystals; associated with lepidolite, tourmaline, beryl; deep grooves often parallel long crystal faces; perfect prismatic cleavage
Olivine $(Mg,Fe)_2SiO_4$	6 1/2 - 7	olive green	vitreous	crystals often appear as glassy green beads, isolated or in masses; color distinctive; conchoidal fracture
Epidote $Ca_2(Al,Fe)_3\ SiO_{12}\ (OH)$	6 1/2 - 7	light to dark green	vitreous	usually a dull avocado massive; crystals are dark green with striations and well developed cleavage