Chapter 3. Earth's Interior

Adapted by Karla Panchuk, University of Saskatchewan from *Physical Geology* by Steven Earle

Figure 3.1 The red rocks of the Tablelands in Gros Morne National Park are a sample of Earth's mantle. Top: The red rocks of the Tablelands are on the right, and stand in contrast with the green surroundings. Bottom: A closer view of Tablelands terrain, showing rocks weathered red, and a near absence of plant life. *Source: Top photograph- Leos Kral (2008) CC BY-NC-SA 2.0; Bottom photograph: Tara Joyce (2013) CC BY-SA 2.0. See Appendix C for more attributions.*

Learning Objectives

After reading this chapter and answering the review questions at the end, you should be able to:

- Explain the variations in the composition and characteristics of Earth's different layers.
- Explain how seismic data can be used to understand the structure of Earth's interior.
- Describe the temperature variations within Earth and their implications for internal processes such as mantle convection.
- Explain the origins of Earth's magnetic field and the timing of magnetic field reversals.
- Describe the isostatic relationship between the crust and the mantle, and the implications of that relationship for geological processes on Earth

The barren red rocks of the Tablelands stand in stark contrast to their lush green surroundings in Gros Morne National Park (Figure 3.1, top). If the Tablelands appear out of place, it's because they are. The Tablelands are one of few places on Earth where you can walk directly on the rocks of Earth's mantle, thanks to an accident of plate tectonics that happened hundreds of millions of years ago. The red colour of the Tablelands rocks comes from iron-bearing minerals reacting with oxygen. Unaltered, the rocks are dark green (Figure 3.2). The rocks lack vegetation because the chemical composition of the rocks does not provide adequate nutrients for plants.

Figure 3.2 Tablelands mantle rock with reddish weathering rind, and dark green fresh surface. Scale in cm. *Source: Karla Panchuk (2017) CC BY 4.0*

Locations like the Tablelands are one way we can learn about Earth's interior. Meteorites derived from smashed differentiated bodies (asteroids that separated into mantle and core) are another. Asteroids that formed at a similar distance from the sun as Earth had a mineral composition akin to Earth's. When these objects were shattered in giant collisions, the result was stony meteorites from fragmented mantle rock, and iron meteorites from fragmented core. Some fragments sampled the result of violent encounters that mixed the two (Figure 3.3).

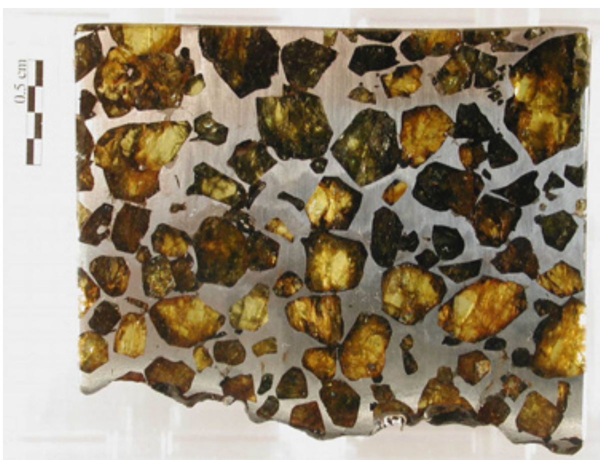

Figure 3.3 Cut and polished slab of a stony-iron meteorite called a pallasite, thought to have formed in a collision that smashed mantle rocks against the metal core of an asteroid early in the solar system's history. Green and brown crystals are the mineral olivine. The metal between the olivine crystals is an iron-nickel mineral. *Source: Muséum de Toulouse (2012) CC BY-NC 2.0*

We also get information about the structure of Earth's interior by analyzing the speeds and paths of earthquake vibrations, called **seismic waves**.

We need to know something about the inside of our planet— what it’s made of, and what happens within it— in order to understand how Earth works, especially the mechanisms of plate tectonics. It is fortunate that there are many ways for geologists gather information about Earth's interior, because one thing they can't do is go down and look at it.

3.1 Earth's Layers: Crust, Mantle, & Core

Earth consists of three main layers: the **crust**, the **mantle**, and the **core** (Figure 3.4). The core accounts for almost half of Earth's radius, but it amounts to only 16.1% of Earth's volume. Most of Earth's volume (82.5%) is its mantle, and only a small fraction (1.4%) is its crust.

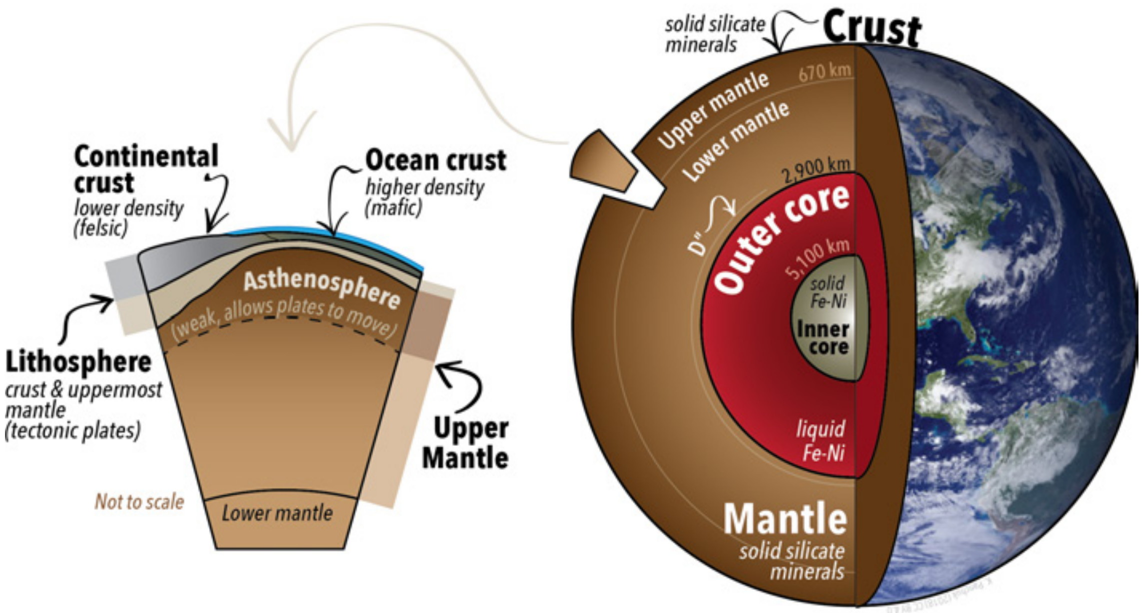

Figure 3.4 Earth's interior. Right- crust, mantle, and outer and inner core to scale. Left- Cutaway showing continental and ocean crust, and upper mantle layers. The lithosphere is the crust plus the uppermost layer of the mantle. *Source: Karla Panchuk (2018) CC BY 4.0. Earth photo by NASA (n.d.) Public Domain*

Crust

The Earth's outermost layer, its **crust**, is rocky and rigid. There are two kinds of crust: **continental crust**, and **ocean crust**. Continental crust is thicker, and predominantly **felsic** in composition, meaning that it contains minerals that are richer in silica. The composition is important because it makes continental crust less dense than ocean crust.

Ocean crust is thinner, and predominantly **mafic** in composition. Mafic rocks contain minerals with less silica, but more iron and magnesium. Mafic rocks (and therefore ocean crust) are denser than the felsic rocks of continental crust.

The crust floats on the mantle. Continental crust floats higher in the mantle than ocean crust because of the lower density of continental crust. An important consequence of the difference in density is that if tectonic plates happen to bring ocean crust and continental crust into collision, the plate with ocean crust will be forced down into the mantle beneath the plate with continental crust.

Mantle

The mantle is almost entirely solid rock, but it is in constant motion, flowing very slowly. It is **ultramafic** in composition, meaning it has even more iron and magnesium than mafic rocks, and even less silica. Although the mantle has a similar chemical composition throughout, it has layers with different mineral compositions and different physical properties. It can have different mineral compositions and still be the same in chemical composition because the increasing pressure deeper in the mantle causes mineral structures to be reconfigured.

Rocks higher in the mantle are typically composed of **peridotite**, a rock dominated by the minerals olivine and pyroxene. The Tablelands rock in Figure 3.2 is a type of peridotite. Lower in the mantle, extreme pressures transform minerals and create rocks like **eclogite** (Figure 3.5), which contains garnets.

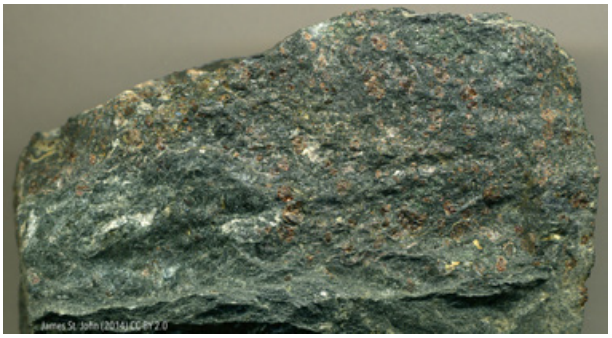

Figure 3.5 Eclogite from the Swiss-Italian Alps. Reddish brown spots are garnets. *Source: James St. John (2014) CC BY 2.0*

Lithosphere

The lithosphere can't be classified neatly as either crust or mantle because it consists of both. It is formed from the crust as well as the uppermost layer of the mantle which is stuck to the underside of the crust. Tectonic plates are fragments of lithosphere.

Asthenosphere

Beneath the lithosphere is the asthenosphere.Tiny amounts of melted rock dispersed through the otherwise solid asthenosphere make the asthenosphere weak compared to the lithosphere. The weakness of the asthenosphere is important for plate tectonics because it deforms as fragments of lithosphere move around upon and through it. Without a weak asthenosphere, plates would be locked in place, unable to move as they do now.

D"

The D" (dee double prime) layer is a mysterious layer beginning approximately 200 km above the boundary between the core and mantle. (This boundary is referred to as the core-mantle boundary.) We know it exists because of how seismic waves change speed as they move through it, but it isn't clear why it's different from the rest of the mantle. One idea is that it is minerals are undergoing another transition in this region because of pressure and temperature conditions, similar to the transition between the upper and lower mantle. Other ideas are that small pools of melt are present, or that the differences in seismic properties are due to subducted slabs of lithosphere resting on the core-mantle boundary.

Core

The core is primarily composed of iron, with lesser amounts of nickel. Lighter elements such as sulfur, oxygen, or silicon may also be present. The core is extremely hot (~3500° to more than 6000°C). But despite the fact that the boundary between the inner and outer core is approximately as hot as the surface of the sun, only the outer core is liquid. The inner core is solid because the pressure at that depth is so high that it keeps the core from melting.

3.2 Imaging Earth’s Interior

Seismology is the study of vibrations within Earth. These vibrations are caused by events such as earthquakes, extraterrestrial impacts, explosions, storm waves hitting the shore, and tides. Seismology is applied to the detection and study of earthquakes, but seismic waves also provide important information about Earth’s interior.

Seismic waves travel through different materials at different speeds, and we can apply knowledge of how they interact with different materials to understand Earth's layers and internal structures. Similar to the way that ultrasound is used to image the human body, we can measure how long it takes for seismic waves to travel from their source to a recording station.

Another feature of seismic waves is that some, called **P-waves**, can travel rapidly though both liquids and solids, but others, called **S-waves**, can only travel though solids, and are slower than P-waves. Observing where P-waves travel, and S-waves do not, allows us to identify regions within Earth that are melted.

Seismic Wave Paths

Seismic waves travel in all directions from their source, but it is convenient to imagine the path traced by one point on the wave front, and represent that path as a **seismic ray** (arrows, Figure 3.6).

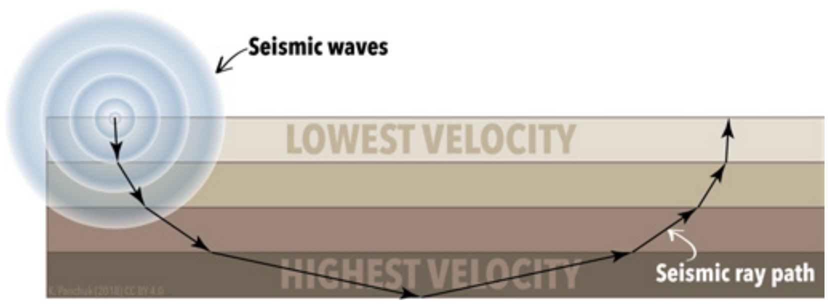

Figure 3.6 Seismic waves and seismic rays. The paths of seismic waves can be represented as rays. Seismic ray paths are bent when they enter a rock layer with a different seismic velocity. *Source: Karla Panchuk (2018) CC BY 4.0*

When seismic waves encounter a different rock layer, some might bounce off the layer, or **reflect**, as in the bottom layer of Figure 3.6. But some waves will travel through the layer. If the wave travels at a different speed in the new layer, its path will be bent, or **refracted**, as it crosses into the new layer. If the wave can travel faster in the new layer, it will be bent slightly toward the contact between the two layers. In Figure 3.6, the ray can travel progressively faster in each layer as it goes down through the layers, and it is bent slightly upward each time it crosses into the next layer. The reverse happens if the wave slows down. On the right side of the diagram, the wave is moving upward through slower and slower layers. It is bent away from the faster layer each time, causing it to take a more direct path to the surface.

Seismic velocities are higher in more rigid layers, so broadly speaking, they get faster deeper within Earth, because higher pressures make layers more rigid. They tend to take curved paths through the Earth because refraction bends their path until they are reflected and directed upward again, as in Figure 3.6.

Discoveries with Seismic Waves

The Moho: Where Crust Meets Mantle

One of the first discoveries about Earth's interior made through seismology was in the early 1900s by Croatian seismologist Andrija Mohorovičić (pronounced *Moho-ro-vi-chich*). He noticed that sometimes, seismic waves arrived at seismic stations (measuring locations) farther from an earthquake *before* they arrived at closer ones. He reasoned that the waves that traveled farther were faster because they bent down and traveled faster through different rocks (those of the mantle) before being bent upward back into the crust (Figure 3.7).

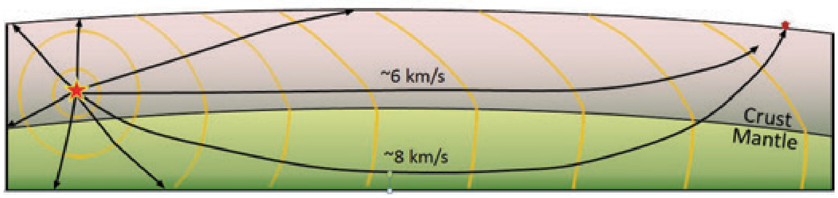

Figure 3.7 Depiction of seismic waves emanating from an earthquake (red star). Some waves travel through the crust to the seismic station (at ~6 km/s), while others go down into the mantle (where they travel at ~8 km/s) and are bent upward toward the surface, reaching the station before the ones that travelled only through the crust. *Source: Steven Earle (2016) CC BY 4.0*

The boundary between the crust and the mantle is now known as the **Mohorovičić discontinuity** (or Moho). Its depth is between 60 – 80 km beneath major mountain ranges, 30 – 50 km beneath most of the continental crust, and 5 – 10 km beneath ocean crust.

The Core-Mantle Boundary

Arguments for a liquid outer core were supported by a distinctive signature in the global distribution of seismic waves from earthquakes. When an earthquake occurs, there is a zone on the opposite side of Earth where S-waves are not measured. This **S-wave shadow zone** begins 103° on either side of the earthquake, for a total angular distance of 154° (Figure 3.8, left). There is also a **P-wave shadow zone** on either side of the earthquake, from 103° to 150° (Figure 3.8, right).

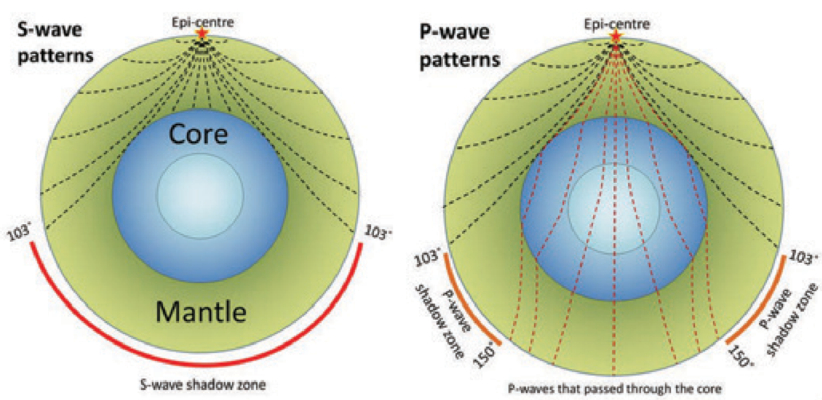

Figure 3.8 Patterns of seismic wave propagation through Earth's mantle and core. S-waves do not travel through the liquid outer core, so they leave a shadow on Earth's far side. P-waves do travel through the core, but because the waves that enter the core are refracted, there are also P-wave shadow zones. *Source: Steven Earle (2016) CC BY 4.0*

The S-wave shadow zone occurs because S-waves cannot travel through the liquid outer core. The P-wave shadow zone occurs because seismic velocities are much lower in the liquid outer core than in the overlying mantle, and the P-waves are refracted in a way that leaves a gap. Not only do the shadow zones tell us that the outer core is liquid, the size of the shadow zones allows us to calculate the size of the core, and the location of the core-mantle boundary.

Seismic Portrait of Earth's Layers

The change seismic wave velocity with depth in Earth (Figure 3.9) has been determined over the past several decades by analyzing seismic signals from large earthquakes all around the world. Earth's layers are detectable as changes in velocity with depth. The asthenosphere is visible as a low velocity zone within the upper mantle (Figure 3.9, left). There is an abrupt increase in P-wave velocity at 420 km, showing the depth at which minerals transform into structures that are more stable at higher pressures and temperatures.

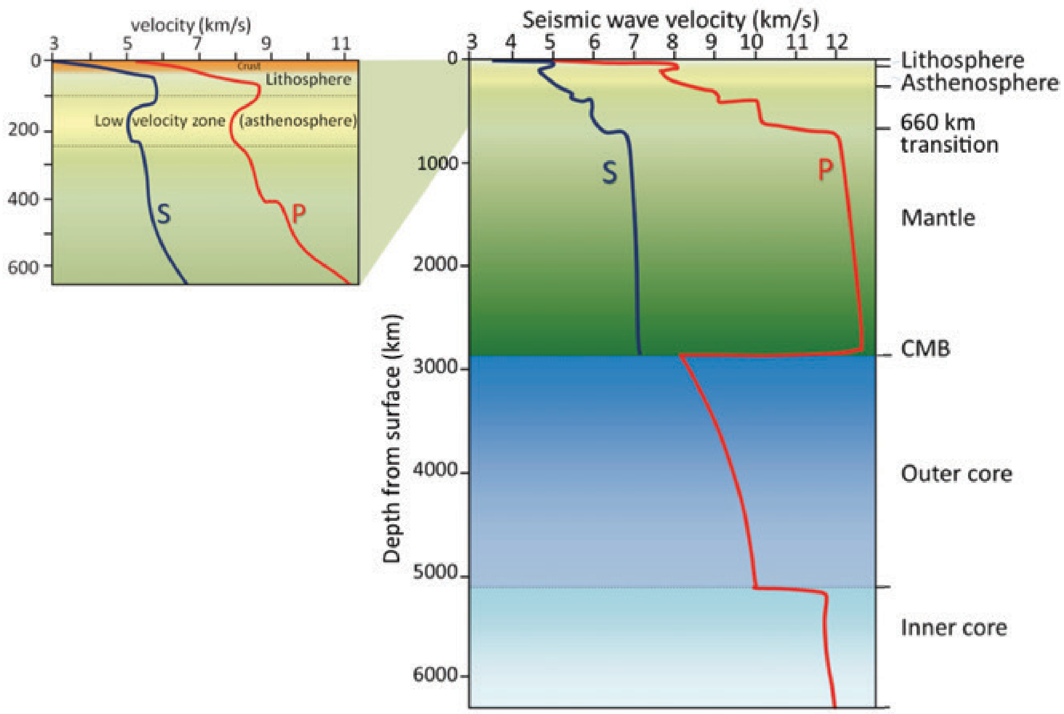

Figure 3.9 P-wave and S-wave velocity variations with depth from the crust through the upper mantle (left) and from the crust through to the core (right). *Source: Karla Panchuk (2018) CC BY 4.0, modified after Steven Earle (2016) CC BY 4.0*

The boundary between the upper and lower mantle is visible at 660 km as a sudden change from rapidly increasing P- and S-wave velocities to slow or no change in P-wave and S-wave velocities (Figure 3.9, right). The core-mantle boundary (CMB in Figure 3.9) is apparent as a sudden drop in P-wave velocities, where seismic waves move from solid mantle to liquid outer core. The boundary between the outer core and inner core is marked by a sudden increase in P-wave velocity after 5000 km, where seismic waves move from a liquid back into a solid again.

Seismic Images of Plate Tectonic Structures

Using data from many seismometers and hundreds of earthquakes, it is possible to create images from the seismic properties of the mantle. This technique is known as **seismic tomography**. Tomography can be used to map out slabs of lithosphere that are entering the mantle, or have disappeared within it. Those slabs are cooler, and therefore more rigid than surrounding mantle rocks, so seismic waves travel through them faster. In Figure 3.10, higher-than-average seismic velocities in cool slabs are indicated in dark blue.

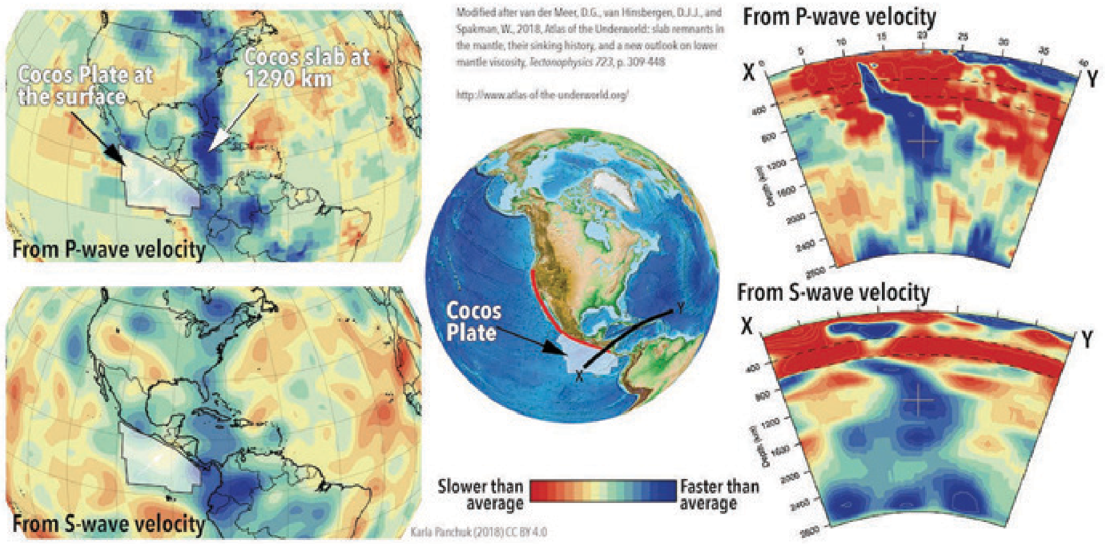

Figure 3.10 P-waves and S-waves used to map out the location of the Cocos slab of lithosphere. The slab appears in dark blue, indicating higher than average seismic wave velocities. Left- Tomograms showing seismic wave anomalies for a 1290 km surface. Right- Cross-sections along the transect marked X-Y on the globe. *Source: Karla Panchuk (2018) CC BY 4.0, modified after van der Meer et al. (2018) CC BY 4.0*

Thanks to the tomograms, we can see that the Cocos plate, which is colliding with Central America, is part of a much larger slab of lithosphere that has already settled onto the mantle. Tomograms representing a surface at 1290 km depth (Figure 3.10, left) show that at that level, the Cocos slab is beneath the Caribbean Sea. The tomograms on the right show a vertical view along the line X-Y marked on the globe. The vertical tomograms show us that the Cocos slab extends all the way down to the core-mantle boundary.

Visit the Underworld

What is the Atlas of the Underworld?

The *Atlas of the Underworld* is a catalog of more than 90 slabs of lithosphere that have been imaged within the mantle using seismic tomography. The *Atlas* includes tomographic images, locator maps, and geological histories for each slab. The catalog can be searched online at http://www.atlas-of-the-underworld.org/ or viewed in the original publication by van der Meer et al. (2018). The *Atlas of the Underworld* is an open-access resource.

The HADES Underworld Explorer

Create your own tomographic cross-sections for locations anywhere in the world by using this intuitive drag-and-drop tool. Visit the HADES Underworld Explorer at http://www.atlas-of-the-underworld.org/hades-underworld-explorer/

3.3 Earth's Interior Heat

Earth Gets Hotter the Deeper You Go

Earth's temperature increases with depth, but not at a uniform rate (Figure 3.11). Earth's **geothermal gradient** is 15° to 30°C/km within the crust. It then drops off dramatically through the mantle, increases more quickly at the base of the mantle, and then increases slowly through the core. The temperature is approximately 1000°C at the base of the crust, around 3500°C at the base of the mantle, and approximately 6,000°C at Earth's centre.

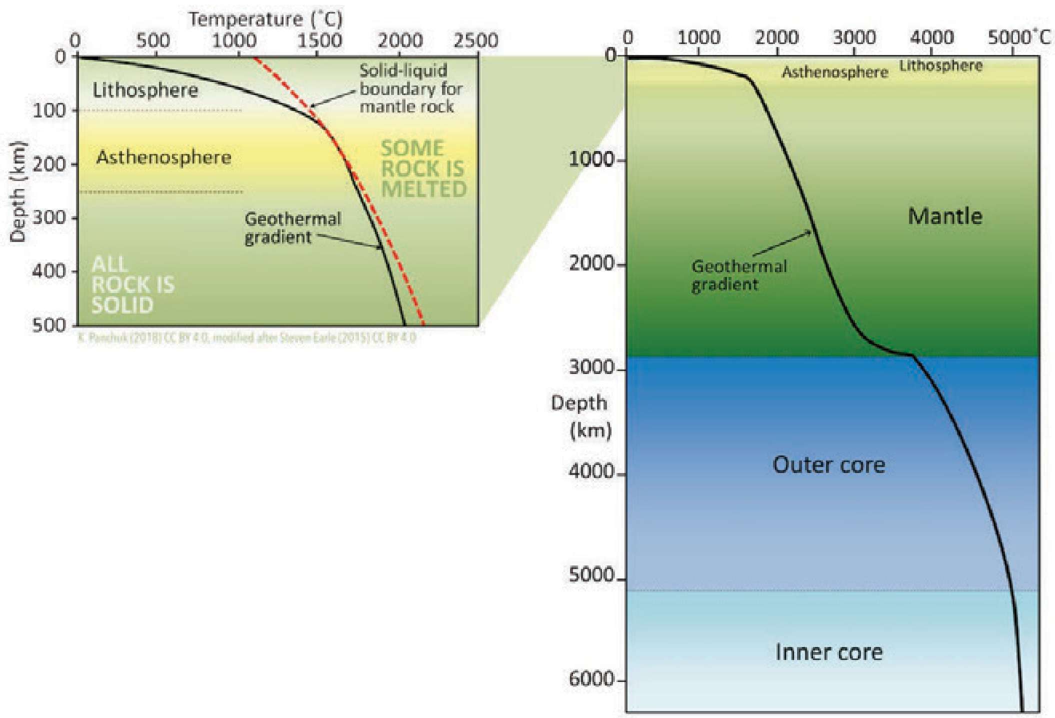

Figure 3.11 Geothermal gradient (change in temperature with depth). Left- Geothermal gradient in the crust and upper mantle. The geothermal gradient remains below the melting temperature of rock, except in the asthenosphere. There, temperatures are high enough to melt some of the minerals. Right- Geothermal gradient throughout Earth. Rapid changes occur in the uppermost mantle, and at the core-mantle boundary. *Source: Karla Panchuk (2018) CC BY 4.0, modified after Steven Earle (2016) CC BY 4.0*

In spite of high temperatures within Earth, mantle rocks are almost entirely solid. High pressures keep them from melting. The red dashed line in Figure 3.11 (right) shows the minimum temperature at which dry mantle rocks will melt. Rocks at temperatures to the left of the line will remain solid. In rocks at temperatures to the right of the line, *some* minerals will begin to melt. Notice that the red dashed line goes further to the right for greater depths, and therefore greater pressures. Now compare the geothermal gradient with the red dashed line. The geothermal gradient is to the left of the red line, except in the asthenosphere, where small amounts of melt are present.

Convection Helps to Move Heat Within Earth

The fact that the temperature gradient is much lower in the main part of the mantle than in the lithosphere has been interpreted as evidence of **convection** in the mantle. When the mantle convects, heat is transferred through the mantle by physically moving hot rocks. Mantle convection is the result of heat transfer from the core to the base of the lower mantle. As with a pot of soup on a hot stove (Figure 3.12), the material near the heat source (the soup at the bottom of the pot) becomes hot and expands, making it less dense than the material above. Buoyancy causes it to rise, and cooler material flows in from the sides. Of course, convection in the soup pot is much faster than convection in the mantle. Mantle convection occurs at rates of centimetres per year.

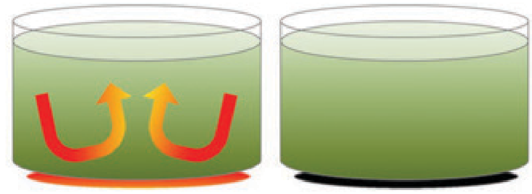

Figure 3.12 Convection in a pot of soup on a hot stove (left). As long as heat is being transferred from below, the liquid will convect. If the heat is turned off (right), the liquid remains hot for a while, but convection will cease. *Source: Steven Earle (2015) CC-BY 4.0*

Convection carries heat to the surface of the mantle much faster than heating by **conduction**. Conduction is heat transfer by collisions between molecules, and is how heat is transferred from the stove to the soup pot. A convecting mantle is an essential feature of plate tectonics, because the higher rate of heat transfer is necessary to keep the asthenosphere weak. Earth's mantle will stop convecting once the core has cooled to the point where there is not enough heat transfer to overcome the strength of the rock. This has already happened on smaller planets like Mercury and Mars, as well as on Earth's moon. When mantle convection stops, the end of plate tectonics will follow.

Models of Mantle Convection

In the soup pot example, convection moves hot soup from the bottom of the pot to the top. Some geologists think that Earth's convection works the same way— hot rock from the base of the mantle moves all the way to the top of the mantle before cooling and sinking back down again. This view is referred to as **whole-mantle convection** (Figure 3.8, left). Other geologists think that the upper and lower mantle are too different to convect as one. They point to slabs of lithosphere that are sinking back into the mantle, some of which seem to perch on the boundary between the upper and lower mantle, rather than sinking straight through. They also note chemical differences in magma originating in different parts of the mantle—differences that are not consistent with the entire mantle being well stirred. They argue that **double-layered convection** is a better fit with the observations (Figure 3.13, right). Still others argue that there may be some locations where convection goes from the bottom of the mantle to the top, and some where it doesn't (Figure 3.13, middle).

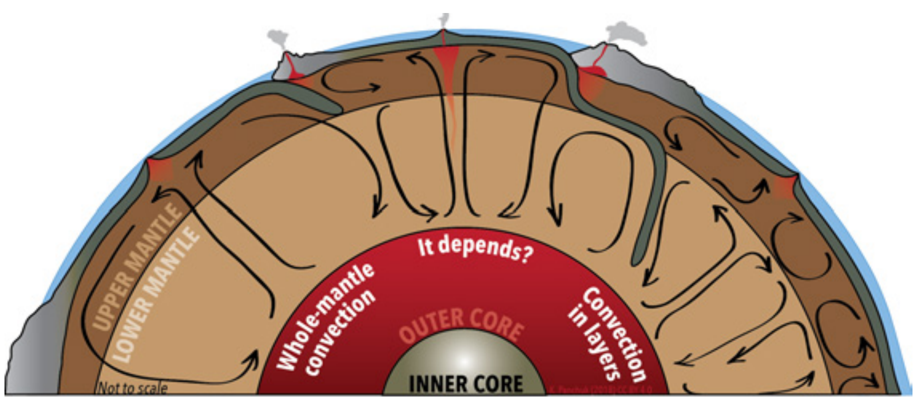

Figure 3.13 Models of mantle convection. Left- whole mantle convection. Rocks rise from the core-mantle boundary to the top of the mantle, then sink to the bottom again. Right- Two-layer convection, in which upper and lower mantle convect at different rates. Middle- Convection paths vary depending on the circumstances. *Source: Karla Panchuk (2018) CC BY 4.0*

Why Is Earth Hot Inside?

The heat of Earth's interior comes from a variety of sources. These include the heat contained in the objects that accreted to form Earth, and the heat produced when they collided. As Earth grew larger, the increased pressure on Earth's interior caused it to compress and heat up. Heat also came from friction when melted material was redistributed within Earth, forming the core and mantle.

A major source of Earth's heat is **radioactivity**, the energy released when the unstable atoms decay. The radioactive isotopes uranium-235 (^{235}U), uranium-238 (^{238}U), potassium-40 (^{40}K), and thorium-232 (^{232}Th) in Earth's mantle are the primary source. Radioactive decay produced more heat early in Earth's history than it does today, because fewer atoms of those isotopes are left today (Figure 3.14). Heat contributed by radioactivity is now roughly a quarter what it was when Earth formed.

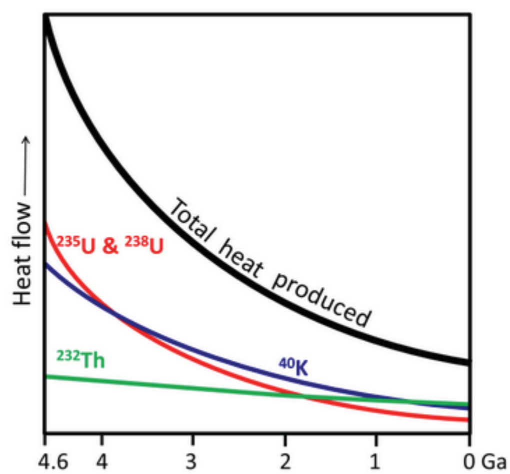

Figure 3.14 Production of heat within the Earth over time by radioactive decay of uranium, thorium, and potassium. Heat production has decreased over time as the abundance of radioactive atoms has decreased. *Source: Steven Earle (2015) CC BY 4.0, modified after Arevalo et al. (2009)*

3.4 Earth's Magnetic Field

Earth's liquid iron core convects because it is heated from beneath by the inner core. Because iron is a metal and conducts electricity (even when molten), its motion generates a magnetic field.

Earth's magnetic field is defined by north and south poles representing lines of magnetic force flowing into Earth in the northern hemisphere and out of Earth in the southern hemisphere (Figure 3.15). Because of the shape of the field lines, the magnetic force is oriented at different angles to the surface in different locations. The tilt, or **inclination** of magnetic field lines is represented by the tilt of compass needles in Figure 3.15. At the north and south poles, the force is vertical. The force is horizontal at the equator. Everywhere in between, the magnetic force is at an intermediate angle to the surface.

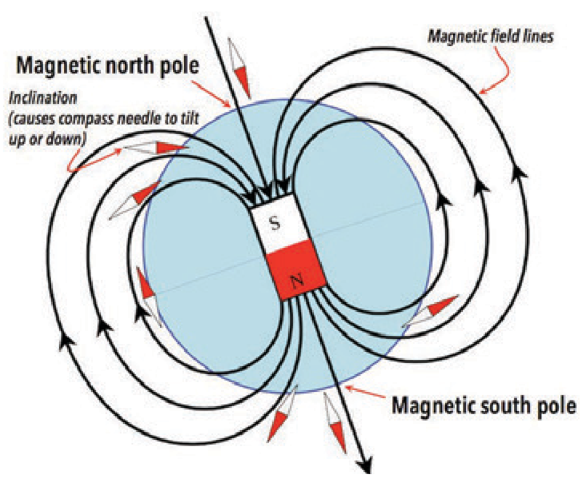

Figure 3.15 Earth's magnetic field depicted as the field of a bar magnet coinciding with the core. The south pole of the magnet points to Earth's magnetic north pole. The red and white compass needles represent the orientation of the magnetic field at various locations on Earth's surface. *Source: Karla Panchuk (2018) CC BY-SA 4.0, modified after Steven Earle (2015) CC BY-SA 4.0, and T. Stein (2008) CC BY-SA 3.0 view source*

Exercise: Magnetic Inclination

Regular compasses point only to the north magnetic pole, but if you had a magnetic dip meter (or a smartphone with the appropriate app), you could also measure the angle of the magnetic field at your location in the up-and-down sense. However, you don't need a dip meter or app to do this exercise! Using Figure 3.15 as a guide, describe the general location on Earth where the vertical angles would be as follows:

1. Straight down
2. Down at a steep angle
3. Up at a steep angle
4. Parallel to flat ground

Earth's magnetic field is generated within the outer core by the convective movement of liquid iron, but although convection is continuous, the magnetic field is not stable. Periodically, the magnetic field decays and then becomes re-established. When it does re-establish, the polarity may have reversed (i.e., your compass would point south rather than north). Over the past 250 Ma, there have been hundreds of magnetic field reversals, and their timing has been anything but regular. The shortest ones that geologists have been able to identify lasted only a few thousand years, and the longest one was more than 30 million years, during the Cretaceous Period (Figure 3.16).

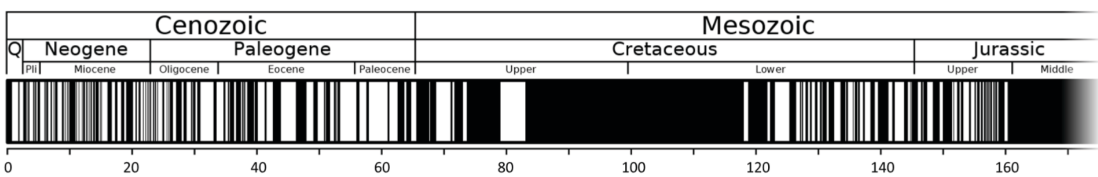

Figure 3.16 Magnetic field reversal chronology for the past 170 Ma. Black stripes mark times when the magnetic field was oriented the same as today. *Source: Steven Earle (2015) CC BY 4.0, modified after AnomieX (2010) Public Domain*

Changes in Earth's magnetic field have been studied using mathematical models that simulate convection in the outer core (Figure 3.17). Reversals happened spontaneously when the model was run to simulate a period of several hundred thousand years. Spontaneous reversals can happen because convection does not occur in an orderly way, in spite of what the bar magnet analogy may suggest. Many small-scale variations occur in convection patterns within the inner core, and Earth's magnetic field over all is the sum of those variations. Magnetic reversals do not happen as frequently as they might, if not for the solid inner core. Magnetic field changes take much longer within the inner core, so reversals in the outer core do not always coincide with reversals in the inner core. Both are required in order for Earth's magnetic field to flip.

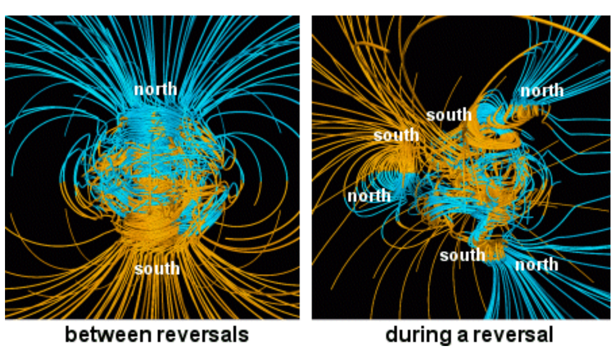

Figure 3.17 Earth's magnetic field between reversals (left) and during a reversal (right). The lines represent magnetic field lines: blue where the field points toward Earth's centre and yellow where it points away. The rotation axis of Earth is vertical, and the outline of the core is shown as a dashed white circle. *Source: NASA (2007) Public Domain*

3.5 Isostasy

Lithospheric Plates Float on the Mantle

The mantle is able to convect because it can deform by flowing over very long timescales. This means that tectonic plates are *floating* on the mantle, like a raft floating in the water, rather than *resting* on the mantle like a raft sitting on the ground. How high the lithosphere floats will depend on the balance between gravity pulling the lithosphere down, and the force of buoyancy as the mantle resists the downward motion of the lithosphere. **Isostasy** is the state in which the force of gravity pulling the plate toward Earth's centre is balanced by the resistance of the mantle to letting the plate sink.

To see how isostasy works, consider the rafts in Figure 3.18. The raft on the right is sitting on solid concrete. The raft will remain at the same elevation whether there are two people on it, or four, because the concrete is too strong to deform. In contrast, isostasy is in play for the rafts on the left, which are floating in a swimming pool full of peanut butter. With only one person on board, the raft floats high in the peanut butter, but with three people, it sinks dangerously low. Peanut butter, rather than water, is used in this example because the viscosity of peanut butter (its stiffness or resistance to flowing) more closely represents the relationship between the tectonic plates and the mantle. Although peanut butter has a similar density to water, it's higher viscosity means that if a person is added to a raft, it will take longer for the raft to settle lower into the peanut butter that it would take the raft to sink into water.

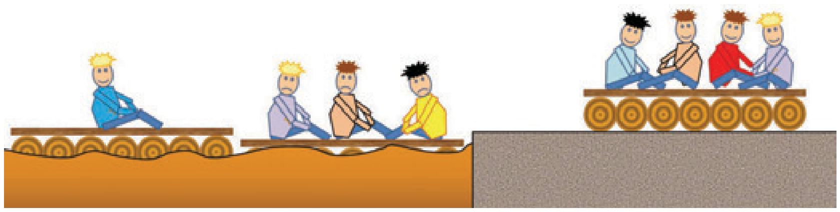

Figure 3.18 Illustration of isostatic relationships between rafts and peanut butter (left), and a non-isostatic relationship between a raft and solid ground (right). *Source: Steven Earle (2015) CC BY 4.0*

The relationship of Earth's crust to the mantle is similar to the relationship of the rafts to the peanut butter. The raft with one person on it floats comfortably high. Even with three people on it the raft is less dense than the peanut butter, so it floats, but it floats uncomfortably low for those three people. The crust, with an average density of around 2.6 g/cm^3, is less dense than the mantle (average density of ~3.4 g/cm^3 near the surface, but more at depth), and so it is floating on the mantle. When weight is added to the crust through the process of mountain building, the crust slowly sinks deeper into the mantle, and the mantle material that was there is pushed aside (Figure 3.19, left). When erosion removes material from the mountains over tens of millions of years, decreasing the weight, the crust rebounds and the mantle rock flows back (Figure 3.19, right).

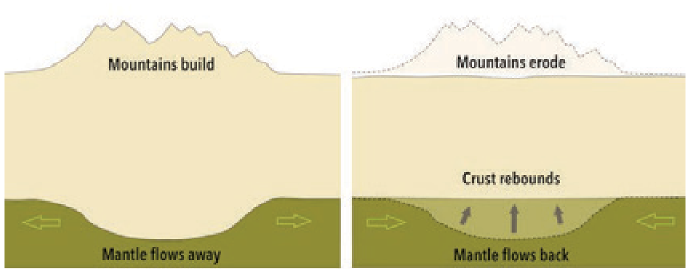

Figure 3.19 Illustration of isostatic relationships between rafts and peanut butter (left), and a non-isostatic relationship between a raft and solid ground (right). *Source: Steven Earle (2015) CC BY 4.0*

Isostasy and Glacial Rebound

The crust and mantle respond in a similar way to glaciation. Thick accumulations of glacial ice add weight to the crust, and the crust subsides, pushing the mantle out of the way. The Greenland ice sheet, at over 2,500 m thick, has depressed the crust below sea level (Figure 3.20a). When the ice eventually melts, the crust and mantle will slowly rebound (Figure 3.20b), but full rebound will likely take more than 10,000 years (3.20c).

Large parts of Canada are still rebounding as a result of the loss of glacial ice over the past 12,000 years, as are other parts of the world (Figure 3.21). The highest rate of uplift is in a large area west of Hudson Bay, where the Laurentide Ice Sheet was the thickest, at over 3,000 m. Ice finally left this region around 8,000 years ago, and the crust is currently rebounding at nearly 2 cm/year. Strong isostatic rebound is also occurring in northern Europe where the Fenno-Scandian Ice Sheet was thickest, and in the eastern part of Antarctica, which also experienced significant ice loss during the Holocene.

Glacial rebound in one location means subsidence in surrounding areas (Figure 3.21, yellow through red regions). Regions surrounding the former Laurentide and Fenno-Scandian Ice Sheets that were lifted up when mantle rock was forced aside and beneath them are now subsiding as the mantle rock flows back.

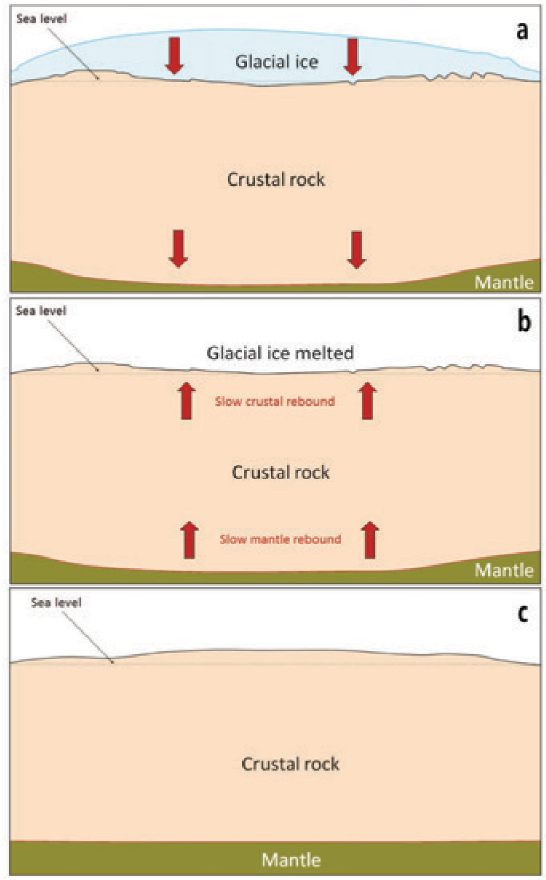

Figure 3.20 Cross-section through the crust in the northern part of Greenland. a) Up to 2,500 m of ice depresses the crust downward (red arrows) and below sea level. b) After complete melting. Isostatic rebound would be slower than the rate of melting, leaving central Greenland at or below sea level for thousands of years. c) Complete rebound after ~10,000 years raises central Greenland above sea level again. *Source: Steven Earle (2015) CC BY 4.0*

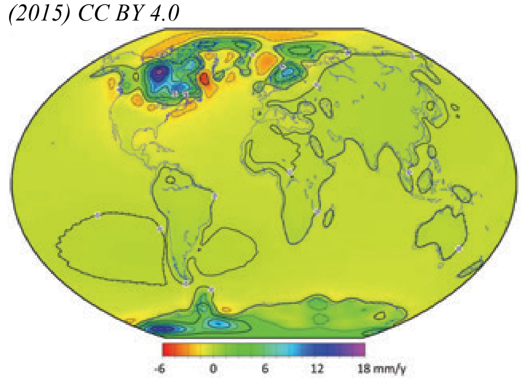

Figure 3.21 Current rates of post-glacial isostatic uplift (green, blue, and purple shades) and subsidence (yellow and orange). Subsidence is taking place where the mantle is slowly flowing back toward areas that are experiencing post-glacial uplift. *Source: Steven Earle (2015) CC BY 4.0, modified after Erik Ivins, JPL (2010) Public Domain*

How Can the Mantle Be Both Solid and Plastic?

You might be wondering how it is possible that Earth's mantle is solid, rigid rock, and yet it convects and flows like a very viscous liquid. The explanation is that the mantle behaves as a non-Newtonian fluid, meaning that it responds differently to stresses depending on how quickly the stress is applied.

A good example of non-Newtonian behaviour is the deformation of Silly Putty, which can bounce when it is compressed rapidly when dropped, and will break if you pull on it sharply. But will deform in a liquid manner if stress is applied slowly. The force of gravity applied over a period of hours can cause it to deform like a liquid, dripping through a hole in a glass tabletop (Figure 3.22). Similarly, the mantle will flow when placed under the slow but steady stress of a growing (or melting) ice sheet.

Chapter 3 Summary

Figure 3.22 Silly Putty exhibiting plastic behaviour when acted upon by gravity over several hours. *Source: Erik Skiff (2006) CC BY-SA 2006*

The topics covered in this chapter can be summarized as follows:

3.1 Earth's Layers

Earth is divided into a rocky crust and mantle, and a core consisting largely of iron. The crust and the uppermost mantle form the lithosphere, which is broken into tectonic plates. The next layer, the asthenosphere, allows the plates to move because it deforms by flowing.

3.2 Imaging Earth's Interior

Seismic waves that travel through Earth are either P-waves or S-waves. P-waves are faster than S-waves, and can pass through fluids. Earth's layers can be identified by looking at changes in the velocity of seismic waves. Seismic wave shadow zones contributed to knowledge of the depth of the core-mantle boundary, and the knowledge that the outer core is liquid. Plate tectonic structures within Earth can also be mapped using the seismic waves generated by earthquakes.

3.3 Earth's Interior Heat

Earth’s temperature increases with depth (to around 6000°C at the centre), but the rate of increase is not the same everywhere. In the lithosphere, thickness and plate tectonic setting are are factors. Deeper within the mantle, convection currents are more important.

3.4 Earth's Magnetic Field

Earth's magnetic field is generated by convection of the liquid outer core. The magnetic field is similar to that of a bar magnet, and has force directions that vary with latitude. The polarity of the field is not constant, meaning that the positions of the north and south magnetic poles have flipped from “normal” (as it is now) to reversed and back many times in Earth's history.

3.5 Isostasy

The plastic nature of the mantle, which allows for mantle convection, also determines the nature of the relationship between the crust and the mantle. The crust floats on the mantle in an isostatic relationship. Where the crust becomes thicker and heavier because of mountain building, it pushes farther down into the mantle. Oceanic crust, being denser than continental crust, floats lower on the mantle than continental crust.

Review Questions

1. What parts of Earth are most closely represented by stony meteorites and iron meteorites?
2. Draw a simple diagram of Earth's layers, and label the approximate locations of the following boundaries: crust/mantle, mantle/core, outer core/inner core.
3. How do P-waves and S-waves differ?

4. Why does P-wave velocity decrease suddenly at the core-mantle boundary?
5. Why do both P-waves and S-waves gradually bend as they move through the mantle?
6. What is the evidence for mantle convection, and what causes mantle convection?
7. How is Earth's magnetic field generated?
8. When were the last two reversals of Earth's magnetic field?
9. What property of the mantle is essential for the isostatic relationship between the crust and the mantle?
10. How would you expect the depth to the crust-mantle boundary in the area of the Rocky Mountains to differ from that in central Saskatchewan?
11. British Columbia is still experiencing weak post-glacial isostatic uplift, especially in the interior, but also along the coast (see Figure 3.21). Meanwhile, offshore areas are experiencing weak isostatic subsidence. Why?

Answers at the end of the chapter

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Answers to Review Questions

1. Stony meteorites are similar in composition to Earth's mantle, while iron meteorites are similar to the core.

2. Compare your answer to Figure 3.4.

3. P-waves can pass through a liquid, and travel approximately twice as fast as S-waves (which cannot pass through a liquid).

4. P-wave velocity decreases at the core-mantle boundary because the outer core is liquid.

5. The mantle gets increasingly dense and strong with depth because of the increasing pressure. This difference affects both P-wave and S-wave velocities, and they are refracted toward the lower density mantle material (meaning they are bent out toward Earth's surface).

6. The key evidence for mantle convection is that the rate of temperature increase with depth within the mantle is less than expected. This can only be explained by a mantle that is mixing by convection. The mechanism for convection is the transfer of heat from the core to the mantle, causing the to mantle flow.

7. Earth's magnetic field is generated within the liquid outer core because liquid metal is convecting.

8. The last two reversals of Earth's magnetic field were at the beginning of the present Brunhes normal chron (0.78 Ma), and at the end of the Jaramillo normal subchron (0.90 Ma).

9. The isostatic relationship between the crust and the mantle is dependent on the fact that over very long timescales, the mantle deforms by flowing.

10. In the area of the Rocky Mountains the crust is thickened and pushed down into the mantle. In Saskatchewan the crust is thinner and does not extend as far into the mantle.

11. During the Pleistocene glaciation, British Columbia was pushed down by glacial ice. Mantle rock flowed slowly out from under the weighted-down crust and toward the ocean floor. Now that the land area is rebounding, that mantle rock is flowing back and the offshore areas are subsiding.