Chapter 9 Earth's Interior

Learning Objectives

After carefully reading this chapter, completing the exercises within it, and answering the questions at the end, you should be able to:

- Explain the variations in the composition and characteristics of Earth's different layers.
- Compare the characteristics and behaviour of the two types of seismic body waves.
- Summarize the variations in seismic-wave velocity as a function of rock type and temperature and pressure conditions.
- Explain some of the ways that seismic data can be used to understand planetary interiors.
- 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.

In order to understand how Earth works, and especially the mechanisms of plate tectonics (covered in Chapter 10), we need to know something about the inside of our planet — what it's made of, and what goes on in there. We have a variety of ways of knowing, and these will be discussed in this chapter, but the one thing we can't do is go down and look! Fortunately there are a few places where mantle rock is exposed on Earth's surface, and we have some samples of material from the insides of other planetary bodies, in the form of meteorites that have landed on Earth (Figure 9.0.1). We also have a great deal of seismic information that can help us understand the nature of Earth's interior.

Figure 9.0.1 Left: a fragment of the Tagish Lake meteorite, discovered in 2000 on the ice of Tagish Lake, B.C. It is a "stony" meteorite that is dominated by ferromagnesian silicate minerals, and is similar in composition to Earth's mantle. Right: part of the Elbogen meteorite discovered in Germany around 1400. It is an iron meteorite, similar in composition to Earth's core. Both samples are a few centimetres across.

Earth's interior is broadly divided by composition and depth into crust, mantle, and core (Figure 9.0.2). The crust is primarily (roughly 95%) made up of igneous rock and metamorphic rock with an overall composition between intermediate and felsic. The remaining 5% is made up of sedimentary rock, which is dominated by mudstone.

The mantle includes several layers, all with the same overall ultramafic composition. The upper mantle is typically composed of peridotite, a rock dominated by olivine and pyroxene. The lower mantle has a similar chemical composition, but because of the extreme pressures, different minerals are present, including spinels and garnets. The properties of the mantle also vary with depth, as follows:

- **Lithosphere**: solid
- **Asthenosphere**: partially liquid
- Upper and lower mantle: solid but plastic (the difference between the two is in the type of minerals)
- **"D" layer** (the part of the mantle within 200 kilometres of the core): partially liquid
- The **core-mantle boundary** (CMB) is at a depth of 2,900 kilometres.

The core is primarily composed of iron, with lesser amounts of nickel (about 5%) and several percent oxygen. It is extremely hot (roughly 3500° to 5000°C). The outer core is liquid while the inner core is solid—even though it is hotter—because the pressure is so much greater at that depth.

Although the CMB is just about half of the way to Earth's centre, the mantle, being on the outside, is by far the major component of Earth. The mantle makes up 82.5% of the volume, the core 16.1%, and the crust only 1.4%.

In the remainder of this chapter, we'll look first at how we know about Earth's interior structure, and then at the properties of the different layers and the processes that take place within them.

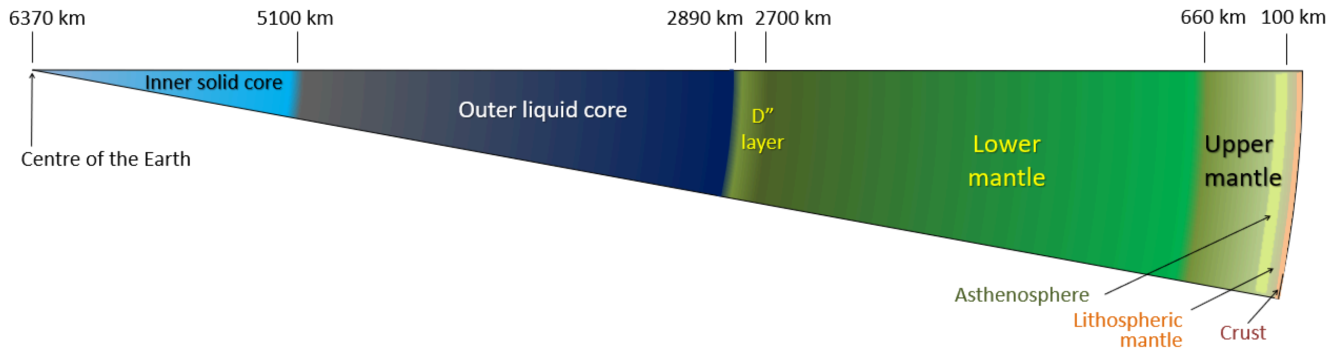

Figure 9.0.2 Earth's layers: crust is pink, mantle is green, core is blue. [Image Description]

Image Descriptions

Figure 9.0.2 image description: Layers of the earth

Layer	**Kilometres below the Earth's surface**	**Thickness (kilometres)**
Crust and Lithospheric part of the mantle	0 to 100	100
Asthenosphere and Upper mantle	100 to 660	560
Lower mantle	660 to 2,700	2,040
D" layer	2,700 to 2,890	190
Outer liquid core	2,890 to 5,100	2,210
Inner solid core	5,100 to 6,370	1,270

[Return to Figure 9.0.2]

Media Attributions

9.1 Understanding Earth Through Seismology

Seismology is the study of vibrations within Earth. These vibrations are caused by various events: earthquakes, extraterrestrial impacts, explosions, storm waves hitting the shore, and tidal effects. Of course, seismic techniques have been most widely applied to the detection and study of earthquakes, but there are many other applications, and arguably seismic waves provide the most important information that we have concerning Earth's interior. Before going any deeper into Earth, however, we need to take a look at the properties of seismic waves. The types of waves that are useful for understanding Earth's interior are called **body waves**, meaning that, unlike the surface waves on the ocean, they are transmitted through Earth materials.

Imagine hitting a large block of strong rock (e.g., granite) with a heavy sledgehammer (Figure 9.1.1). At the point where the hammer strikes it, a small part of the rock will be compressed by a fraction of a millimetre. That compression will transfer to the neighbouring part of the rock, and so on through to the far side of the rock—all in a fraction of a second. This is known as a compression wave, and it can be illustrated by holding a loose spring (like a Slinky) that is attached to something (or someone) at the other end. If you give it a sharp push so the coils are compressed, the compression propagates (travels) along the length of the spring and back (Figure 9.1.2). You can think of a compression wave as a "push" wave—it's called a **P wave** (although the "P" stands for "primary" because P waves arrive first at seismic stations).

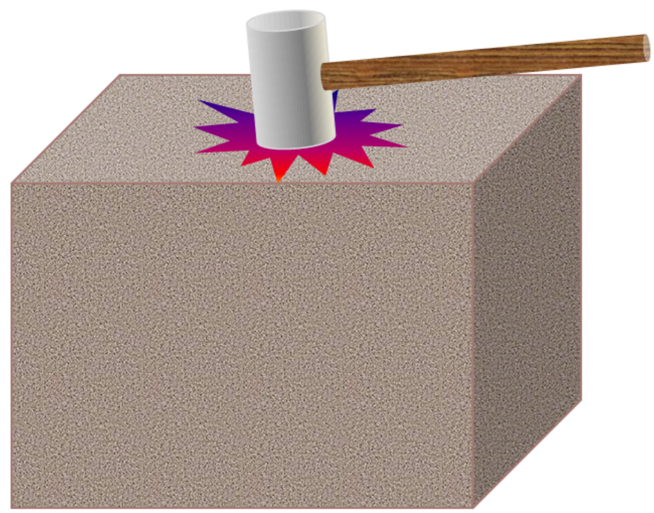

Figure 9.1.1 Hitting a large block of rock with a heavy hammer will create seismic waves within the rock. Please don't try this at home!

When we hit a rock with a hammer, we also create a different type of body wave, one that is characterized by back-and-forth vibrations (as opposed to compressions). This is known as a shear wave (**S wave**, where the "S" stands for "secondary"), and an analogy would be what happens when you flick a length of rope with an up-and-down motion. As shown in Figure 9.1.2, a wave will form in the rope, which will travel to the end of the rope and back.

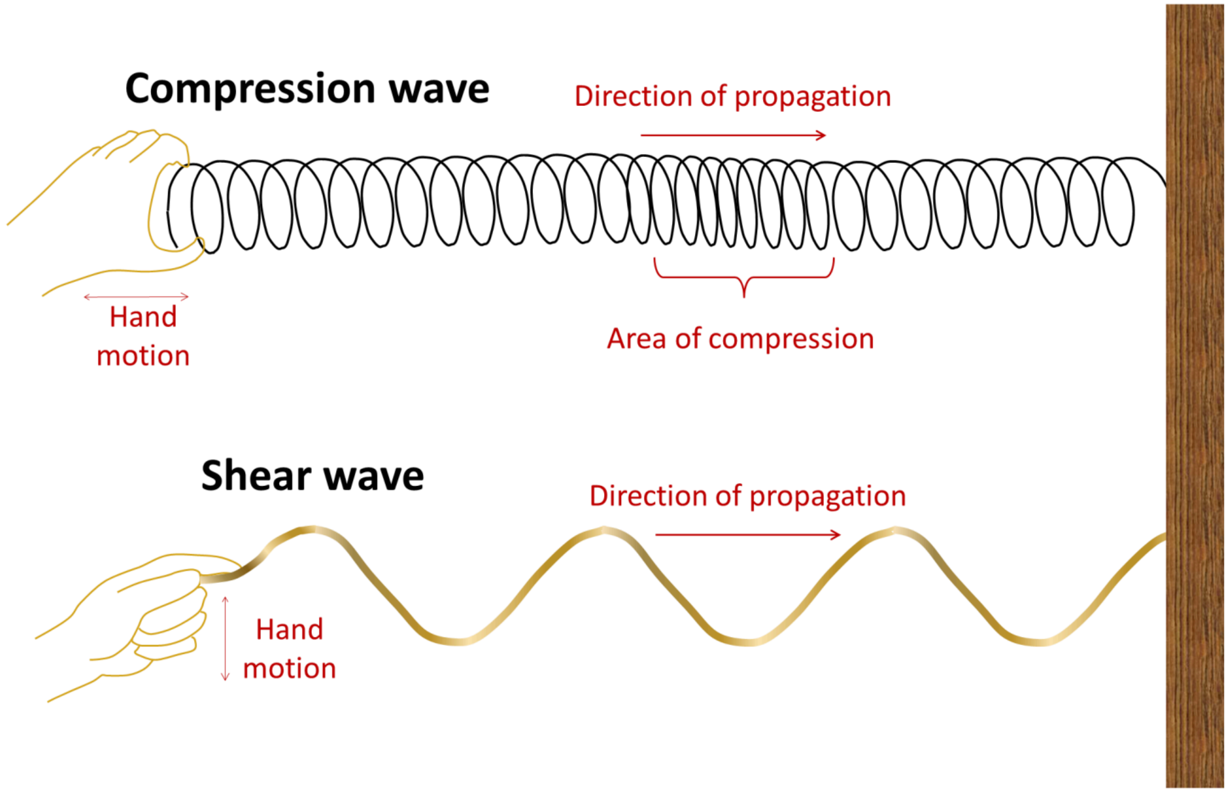

Figure 9.1.2 A compression wave can be illustrated by a spring (like a Slinky) that is given a sharp push at one end. A shear wave can be illustrated by a rope that is given a quick flick.

Compression waves and shear waves travel very quickly through geological materials. As shown in Figure 9.1.3, typical P wave velocities are between 0.5 kilometres per second (km/s) and 2.5 km/s in unconsolidated sediments, and between 3.0 km/s and 6.5 km/s in solid crustal rocks. Of the common rocks of the crust, velocities are greatest in basalt and granite. S waves are slower than P waves, with velocities between 0.1 km/s and 0.8 km/s in soft sediments, and between 1.5 km/s and 3.8 km/s in solid rocks.

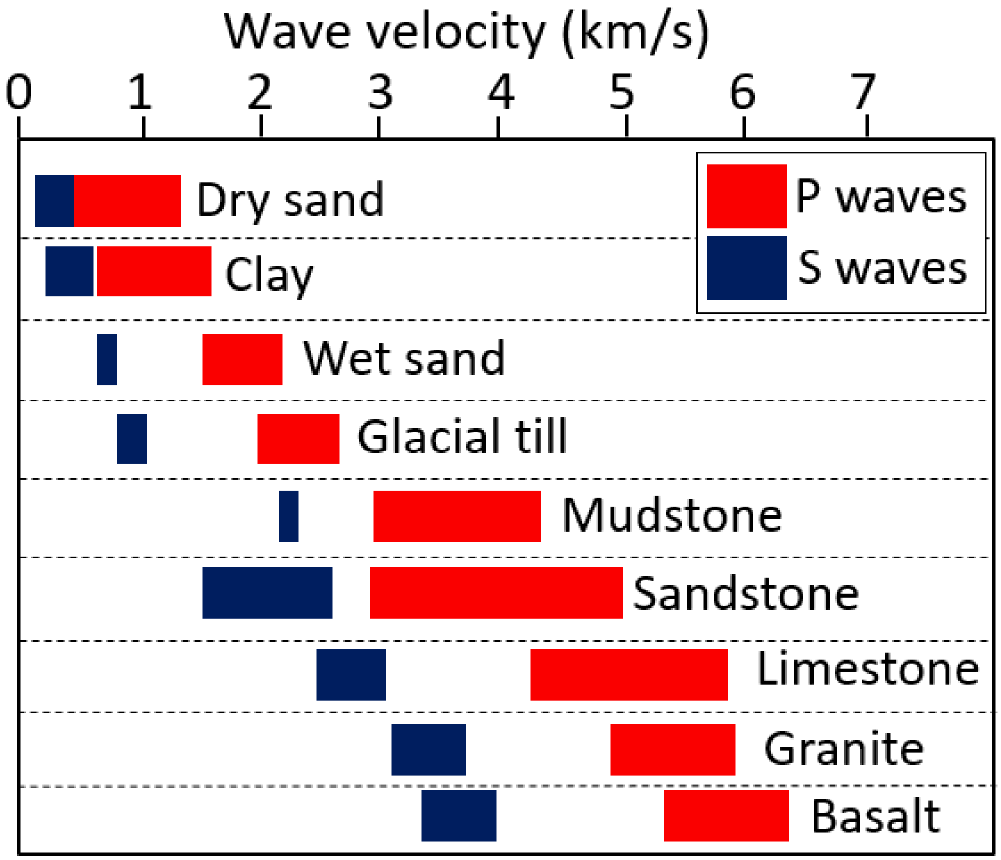

Figure 9.1.3 Typical velocities of P-waves (red) and S-waves (blue) in sediments and in solid crustal rocks. [Image Description]

Exercise 9.1 How soon will seismic waves get here?

Imagine that a strong earthquake takes place on Vancouver Island within Strathcona Park (west of Courtenay). Assuming that the crustal average P wave velocity is 5 km per second, how long will it take (in seconds) for the first seismic waves (P waves) to reach you in the following places (distances from the epicentre are shown)?

1. Nanaimo (120 km away)
2. Surrey (200 km away)
3. Kamloops (390 km away)

See Appendix 3 for Exercise 9.1 answers.

Mantle rock is generally denser and stronger than crustal rock and both P- and S-waves travel faster through the mantle than they do through the crust. Moreover, seismic-wave velocities are related to how tightly compressed a rock is, and the level of compression increases dramatically with depth. Finally, seismic waves are affected by the phase state of rock. They are slowed if there is any degree of melting in the rock. If the material is completely liquid, P waves are slowed dramatically and S waves are stopped altogether.

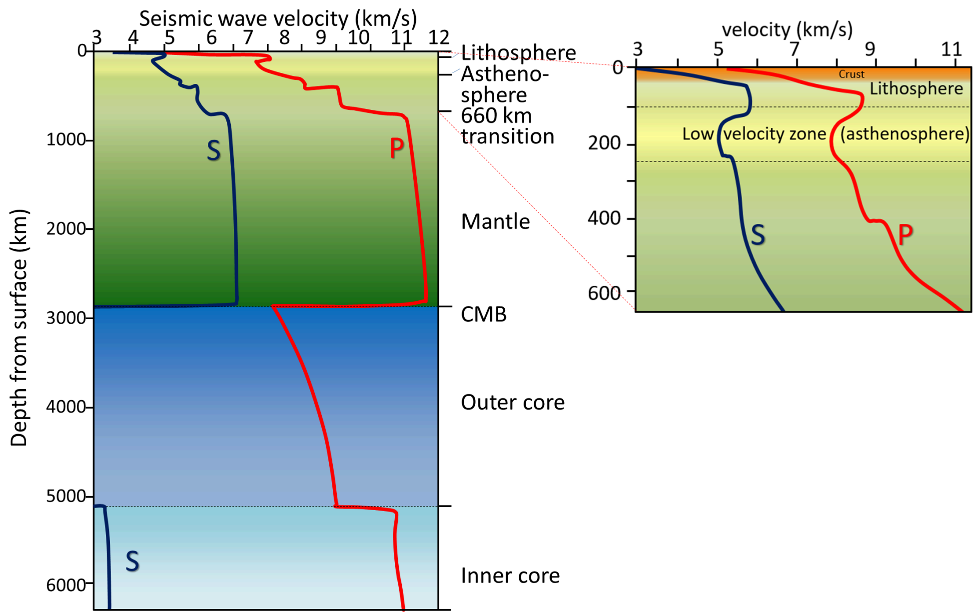

Figure 9.1.4 P wave (red) and S wave (blue) velocity variations with depth in Earth. The diagram on the right shows an expanded view of the upper 660 kilometres of the curves in the diagram on the left. [Image Description]

As shown on the right-hand part of Figure 9.1.4, the upper approximately 100 km of the Earth is known as the lithosphere. This includes the rigid upper part of the mantle (or lithospheric mantle) and the crust. The next 150 km is the asthenosphere or low velocity zone (because seismic waves are slowed as they pass through that material). As we'll see below, that part of the mantle is close to it's melting point and in some regions may be partially molten.

Accurate seismometers have been used for earthquake studies since the late 1800s, and systematic use of seismic data to understand Earth's interior started in the early 1900s. The rate of change of seismic waves with depth in Earth (as shown in Figure 9.1.4) has been determined over the past several decades by analyzing seismic signals from large earthquakes at seismic stations around the world. Small differences in arrival time of signals at different locations have been interpreted to show that:

- Velocities are greater in mantle rock than in the crust.
- Velocities generally increase with pressure, and therefore with depth.
- Velocities slow in the area between a 100 and 250 kilometre depth (called the "low-velocity zone"; equivalent to the asthenosphere).
- Velocities increase dramatically at 660 kilometre depth (because of a mineralogical transition).
- Velocities slow in the region just above the core-mantle boundary (the D" (d-double-prime) layer or "ultra-low-velocity zone").
- S waves do not pass through the outer liquid part of the core, but S waves can be created by P waves at the surface of the inner core and their inner core velocity is 3.6 km/s.

- P wave velocities increase dramatically at the boundary between the liquid outer core and the solid inner core.

One of the first discoveries about Earth's interior made through seismology was in 1909 when Croatian seismologist Andrija Mohorovičić (pronounced *Moho-ro-vi-chich*) realized that at certain distances from an earthquake, two separate sets of seismic waves arrived at a seismic station within a few seconds of each other. He reasoned that the waves that went down into the mantle, traveled through the mantle, and then were bent upward back into the crust, reached the seismic station first because although they had farther to go, they traveled faster through mantle rock (as shown in Figure 9.1.5). The boundary between the crust and the mantle is known as the **Mohorovičić discontinuity** (or Moho). Its depth is between 30 and 40 kilometres beneath most of the continental crust, and between 5 and 10 kilometres beneath the oceanic crust.

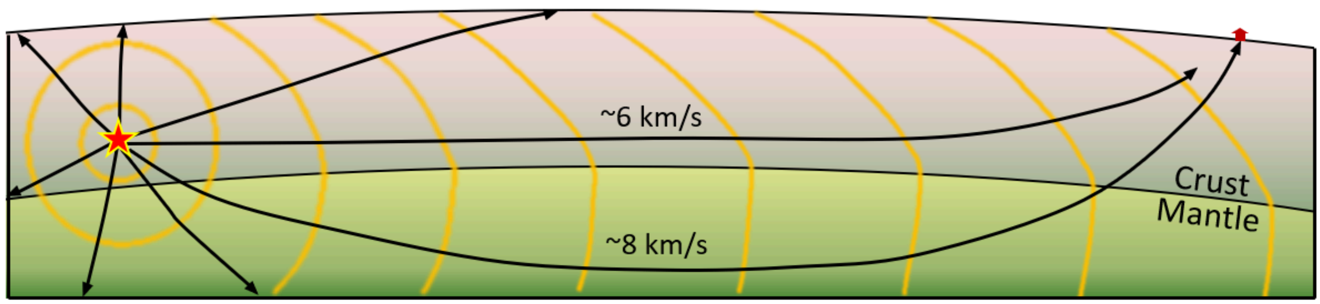

Figure 9.1.5 Depiction of seismic waves emanating from an earthquake (red star). Some waves travel through the crust to the seismic station (at about 6 km/s), while others go down into the mantle (where they travel at around 8 km/s) and are bent upward toward the surface, reaching the station before the ones that traveled only through the crust.

Our current understanding of the patterns of seismic wave transmission through Earth is summarized in Figure 9.1.6. Because of the gradual increase in density (and therefore rock strength) with depth, all waves are refracted (toward the lower density material) as they travel through homogenous parts of Earth and thus tend to curve outward toward the surface. Waves are also refracted at boundaries within Earth, such as at the Moho, at the core-mantle boundary (CMB), and at the outer-core/inner-core boundary.

S waves do not travel through liquids—they are stopped at the CMB—and there is an S wave shadow on the side of Earth opposite a seismic source. The angular distance from the seismic source to the shadow zone is 103° on either side, so the total angular distance of the shadow zone is 154°. We can use this information to infer the depth to the CMB.

P waves do travel through liquids, so they can make it through the liquid part of the core. Because of the refraction that takes place at the CMB, waves that travel through the core are bent away from the surface, and this creates a P wave shadow zone on either side, from 103° to 150°. This information can be used to discover the differences between the inner and outer parts of the core.

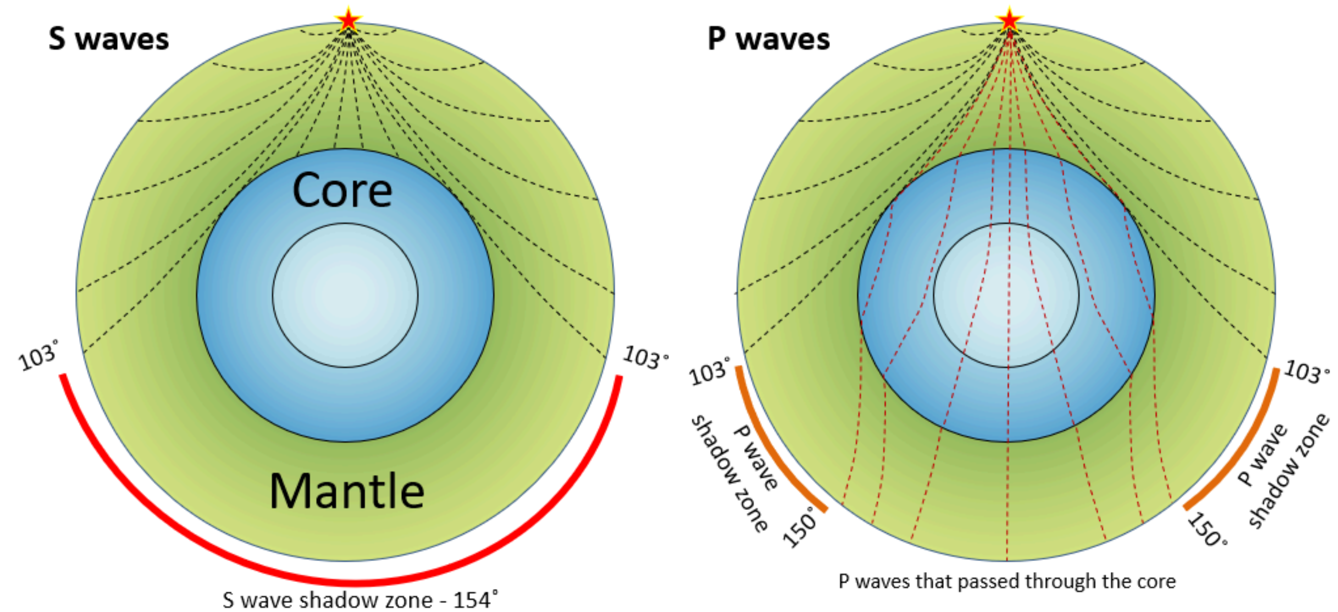

Figure 9.1.6 Patterns of seismic wave propagation through Earth's mantle and core. S waves do not travel through the liquid outer core, so they leave a shadow on Earth's far side where they cannot get to. P waves do travel through the core, but because the waves that enter the core are refracted, there are also P wave shadow zones.

Exercise 9.2 Liquid Cores in Other Planets

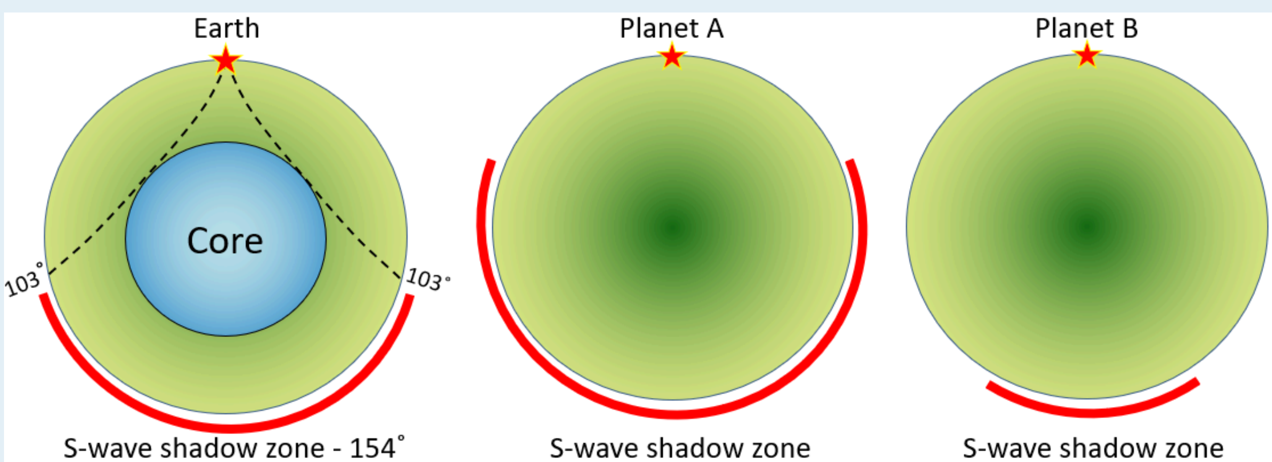

Figure 9.1.7

We know that other planets must have (or at least did have) liquid cores like ours, and we could use seismic data to find out how big they are. The S wave shadow zones on planets A and B are shown. Using the same method used for Earth (on the left), sketch in the outlines of the cores for these two other planets.

See Appendix 3 for Exercise 9.2 answers.

Using data from many seismometers and hundreds of earthquakes, it is possible to create a two- or three-dimensional image of the seismic properties of part of the mantle. This technique is known as seismic tomography, and an example of the result is shown in Figure 9.1.8.

The Pacific Plate subducts beneath Tonga and appears in Figure 9.1.8 as a 100 kilometre thick slab of cold (blue-coloured) oceanic crust that has pushed down into the surrounding hot mantle. The cold rock is more rigid than the surrounding hot mantle rock, so it is characterized by slightly faster seismic velocities. There is volcanism in the Lau spreading centre and also in the Fiji area, and the warm rock in these areas has slower seismic velocities (yellow and red colours).

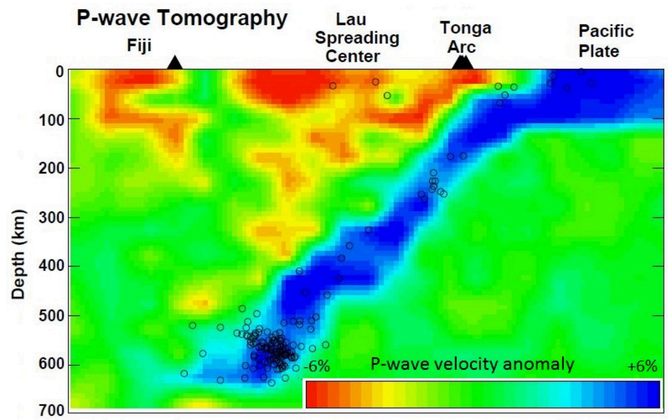

Figure 9.1.8 P-wave tomographic profile of area in the southern Pacific Ocean from southeast of Tonga to Fiji. Blue represents rock that has relatively high seismic velocities, while yellow and red represent rock with low velocities. Open circles are earthquakes used in the study.

Image descriptions

Figure 9.1.3 image description: Wave velocity in different materials in kilometres per second.

Material	S Wave (kilometres per second)	P Wave (kilometres per second)
Dry sand	0.1 to 0.4	0.4 to 1.3
Clay	0.2 to 0.6	0.6 to 1.6
Wet sand	0.7 to 0.8	1.5 to 2.2
Till	0.8 to 1.0	1.9 to 2.6
Mudstone	2.1 to 2.3	3.0 to 4.3
Sandstone	1.4 to 2.5	3.0 to 5.0
Limestone	2.4 to 3.1	4.2 to 5.8
Granite	3.0 to 3.7	4.9 to 5.9
Basalt	3.3 to 4.0	5.2 to 6.2

[Return to Figure 9.1.3]

Figure 9.1.4 image description: P-wave and S-wave velocity variations with depth in Earth.

Layer	Depth from surface (km)	S-Wave velocity (kilometres per second)	P-Wave velocity (kilometres per second)
Crust	0 to 30	3.0 to 4.6	5.3 to 7.0
Lithosphere	30 to 100	4.6 to 5.8	7.0 to 8.7
Asthenosphere	100 to 250	5.0 to 5.9	7.8 to 8.5
Mantle	250 to 2890	5. to 7.0	8.2 to 12.6
Outer core	2890 to 5100	0	8.0 to 10.1
Inner core	5100 to 6370	0	11.8 to 12.0

[Return to Figure 9.1.4]

Media Attributions

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- Figure 9.1.3: "P Wave Velocity, m/s" and "Shear Wave Velocity, m/s" by the US Environment Protection Agency. Edited by Steven Earle. Public domain.
- Figure 9.1.8: "P-wave Tomography" by D. Zhao, Y. Xu, D.A. Wiens, L. Dorman, J. Hildebrand, and S. Webb. (Science, p. 278, 254-257, 1997). Used with permission.

9.2 The Temperature of Earth's Interior

As we've discussed in the context of metamorphism, Earth's internal temperature increases with depth. However, as shown in Figure 9.2.1 (right), that rate of increase is not linear. The temperature gradient is around 15° to 30°C per kilometre within the upper 100 kilometres; 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 around 1000°C at the base of the crust, around 3500°C at the base of the mantle, and around 5,000°C at Earth's centre. The temperature gradient within the lithosphere (upper 100 kilometres) is quite variable depending on the tectonic setting. Gradients are lowest in the central parts of continents, higher in the vicinity of subduction zones, and higher still at divergent boundaries.

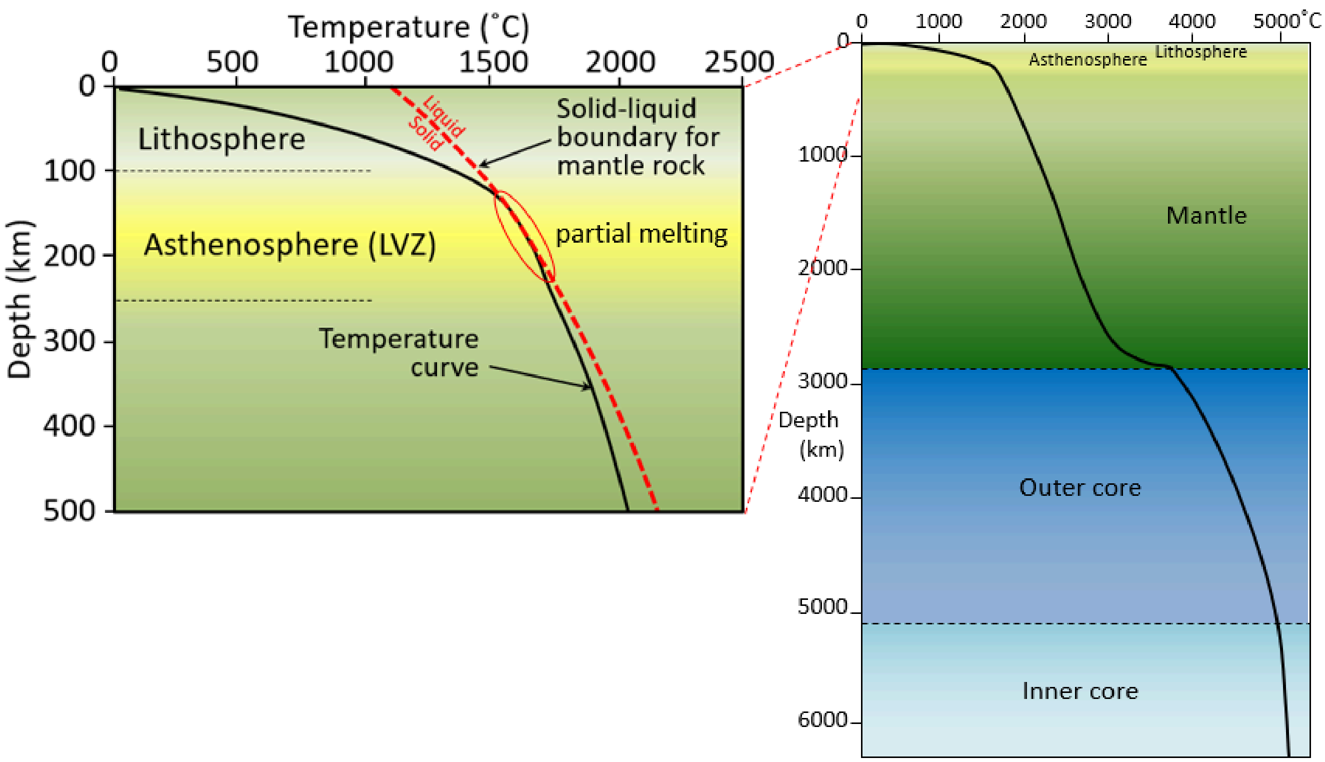

Figure 9.2.1 Right: generalized rate of temperature increase with depth within Earth. Temperature increases to the right, so the flatter the line, the steeper the temperature gradient. Our understanding of the temperature gradient comes from seismic wave information and knowledge of the melting points of Earth's materials. Left: Rate of temperature increase with depth in Earth's upper 500 kilometres, compared with the dry mantle rock melting curve (red dashed line). LVZ= low-velocity zone. [Image Description]

Figure 9.2.1 (left) shows a typical temperature curve for the upper 500 kilometres of the mantle in more detail, along with the melting curve for dry mantle rock. (Mantle rock will melt under conditions to the right of the dashed red line.) In general the mantle is not molten because the temperature lies to the left of the melting curve, but within the depth interval between 100 and 250 kilometres the temperature curve comes very close to the melting boundary for dry mantle rock. At these depths, therefore, mantle rock is either very nearly melted or partially melted. In some situations, where extra heat is present and the temperature line crosses over the melting line, or where water is present, it may be completely molten.

This region of the mantle—the asthenosphere—is also known as the low-velocity zone because seismic waves are slowed within rock that is near its melting point. Below 250 kilometres the temperature stays on the left side of the melting line; in other words, the mantle is solid from here all the way down to the D" layer near the core-mantle-boundary.

The fact that the temperature gradient is much less in the main part of the mantle than in the lithosphere has been interpreted to indicate that the mantle is convecting, and therefore that heat from depth is being brought toward the surface faster than it would be with only heat conduction. As we'll see in Chapter 10, a convecting mantle is an key feature of plate tectonics.

The convection of the mantle is a product of the upward transfer of heat from the core to the lower mantle. As in a pot of soup on a hot stove (Figure 9.2.2), the material near the heat source becomes hot and expands, making it lighter than the material above. The force of buoyancy causes it to rise, and cooler material flows in from the sides. The mantle convects in this way because the heat transfer from below is not perfectly even, and also because, even though mantle material is solid rock, it is sufficiently plastic to slowly flow (at rates of centimetres per year) as long as a steady force is applied to it.

As in the soup pot example, Earth's mantle will no longer convect 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.

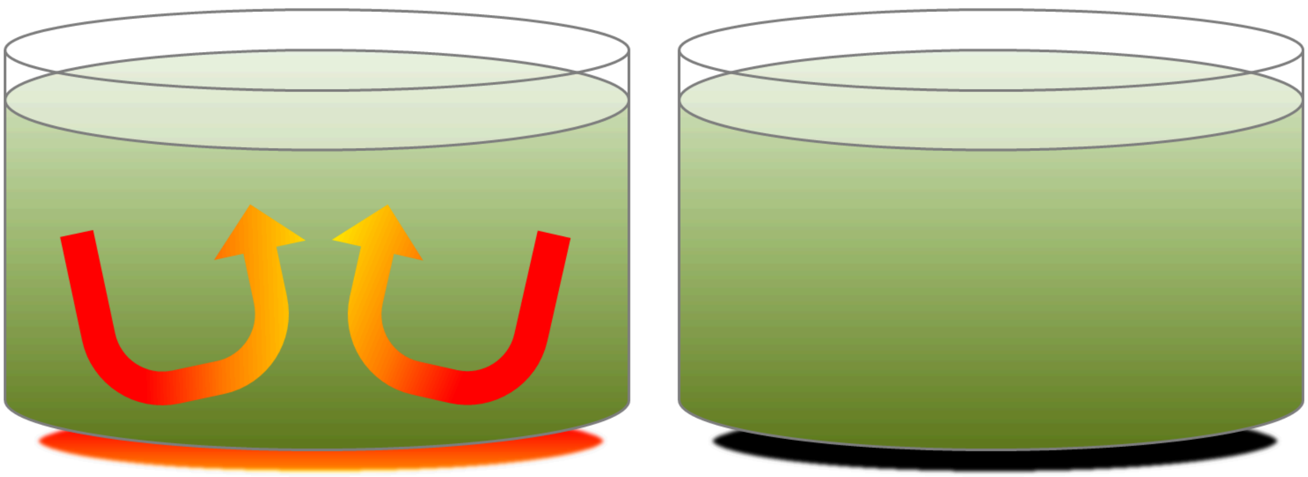

Figure 9.2.2 Convection in a pot of soup on a hot stove (left). As long as heat is being transferred from below, the liquid will convect. If the heat is turned off (right), the liquid remains hot for a while, but convection will cease.

Why is the inside of the Earth hot?

The heat of Earth's interior comes from two main sources, each contributing about 50% of the heat. One of those is the frictional heat left over from the collisions of large and small particles that created Earth in the first place, plus the subsequent frictional heat of redistribution of material within Earth by gravitational forces (e.g., sinking of iron to form the core).

The other source is **radioactivity**, specifically the spontaneous radioactive decay of the isotopes ^{235}U, ^{238}U, ^{40}K, and ^{232}Th, which are primarily present in the mantle. As shown on Figure 9.2.3, the total heat produced that way has been decreasing over time (because these isotopes are getting used up), and is now roughly 25% of what it was when Earth formed. This means that Earth's interior is slowly becoming cooler.

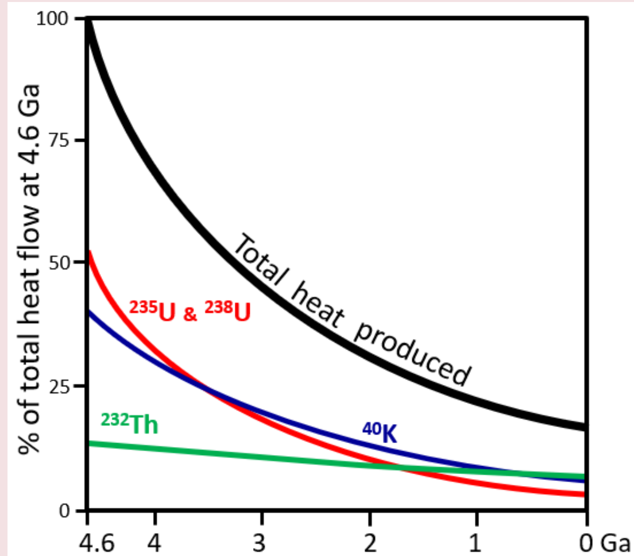

Figure 9.2.3 Heat flow on Earth from radioactive decay

Image Descriptions

Figure 9.2.1 image description: Temperature increase within the Earth.

Layer	Depth (kilometres)	Temperature increase (Celsius)	Temperature increase rate (Degrees per kilometre)
Lithosphere	0 to 100	0 to 1400	14
Asthenosphere	100 to 250	1400 to 1700	2
Mantle	250 to 1000	1700 to 2100	0.53
	1000 to 2000	2100 to 2600	0.5
	2000 to 2890	2600 to 3800	1.35
Outer Core	2890 to 4000	3800 to 4600	0.72
	4000 to 5100	4600 to 5000	0.36
Inner core	5100 to 6370	5000 to 5100	0.079

[Return to Figure 9.2.1]

Media Attributions

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- Figure 9.1.3: © Steven Earle. CC BY. Based on R. Arevalo, W. McDonough, M. Luong, , The K/U ratio of Earth: insights into mantle composition, structure and thermal evolution, Earth and Planetary Science Letters, V 278, p. 361-369, 2009.

9.3 Earth's Magnetic Field

Heat is also being transferred from the solid inner core to the liquid outer core, and this leads to convection of the liquid iron of the outer 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 the North and South Poles that align generally with the axis of rotation (Figure 9.3.1). The lines of magnetic force flow into Earth in the northern hemisphere and out of Earth in the southern hemisphere. Because of the shape of the field lines, the magnetic force trends at different angles to the surface in different locations (red arrows of Figure 9.3.1). At the North and South magnetic poles, the force is vertical. Anywhere near to the equator the force is horizontal, and everywhere in between, the magnetic force is at some intermediate angle to the surface. As we'll see in Chapter 10, the variations in these orientations provide a critical piece of evidence to the understanding of continental drift as an aspect of plate tectonics.

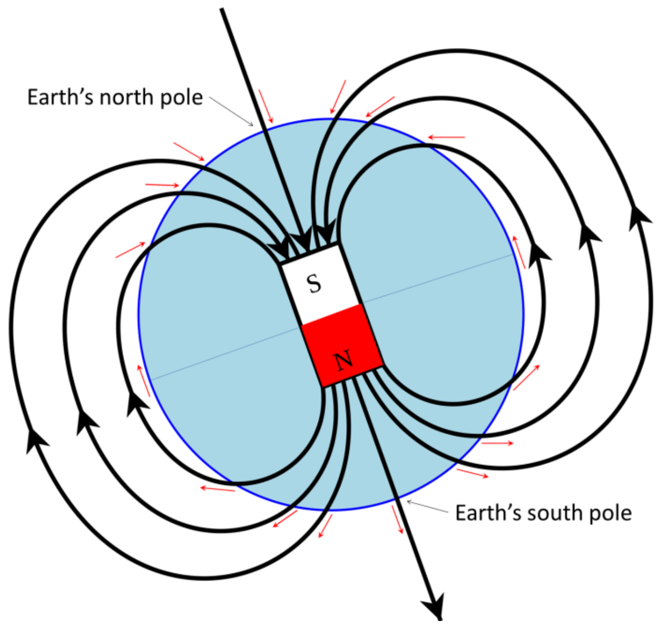

Figure 9.3.1 Depiction of Earth's magnetic field as a bar magnet coinciding with the core. The south pole of such a magnet points to Earth's North Pole. The red arrows represent the orientation of the magnetic field at various locations on Earth's surface.

Earth's magnetic field is generated within the outer core by the convective movement of liquid iron, but as we discovered in Chapter 8, the magnetic field is not stable over geological time. For reasons that are not completely understood, the magnetic field decays periodically and then becomes re-established. When it does re-establish, it may be oriented the way it was before the decay, or it may be oriented with the reversed polarity. Over the past 250 Ma, there have been a few hundred magnetic field reversals, and their timing has been anything but regular. The shortest ones that geologists have been able to define lasted only a few thousand years, and the longest one was more than 30 million years, during the Cretaceous (Figure 9.3.2).

Exercise 9.3 What would a magnetic dip meter tell you?

Regular compasses point only to the north magnetic pole, but if you have a magnetic dip meter you could also measure the angle of the magnetic field at your location in the up-and-down sense.

Using Figure 9.3.1 as a guide, describe where you'd be on Earth if the vertical angles are as follows:

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

See Appendix 3 for Exercise 9.3 answers.

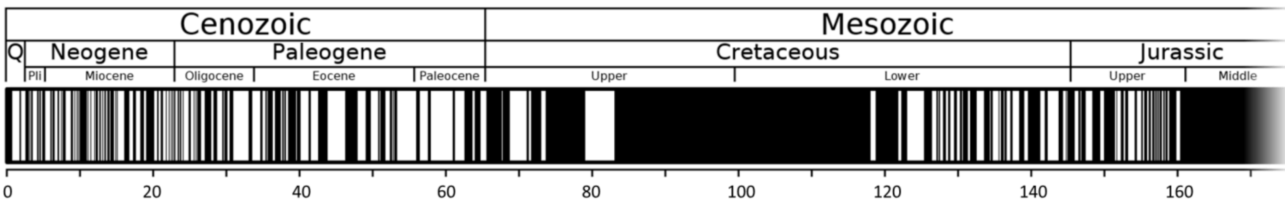

Figure 9.3.2 Magnetic field reversal chronology for the past 170 Ma. The first 5 Ma of the magnetic chronology are shown in more detail in Figure 8.5.3, although the time scale is in the opposite direction in that figure.

Changes in Earth's magnetic field have been studied using a mathematical model, and reversals have been shown to take place when the model was run to simulate a period of several hundred thousand years. The fact that field reversals took place shows that the model is a reasonably accurate representation of the Earth. According to the lead author of the study, Gary Glatzmaier, of University of California at Santa Cruz: "Our solution shows how convection in the fluid outer core is continually trying to reverse the field but that the solid inner core inhibits magnetic reversals because the field in the inner core can only change on the much longer time scale of diffusion. Only once in many attempts is a reversal successful, which is probably the reason why the times between reversals of the Earth's field are long and randomly distributed." A depiction of Earth's magnetic field lines during a stable period and during a reversal is shown in Figure 9.3.3. To read more about these phenomena see Glatzmaier's Geodynamo website.

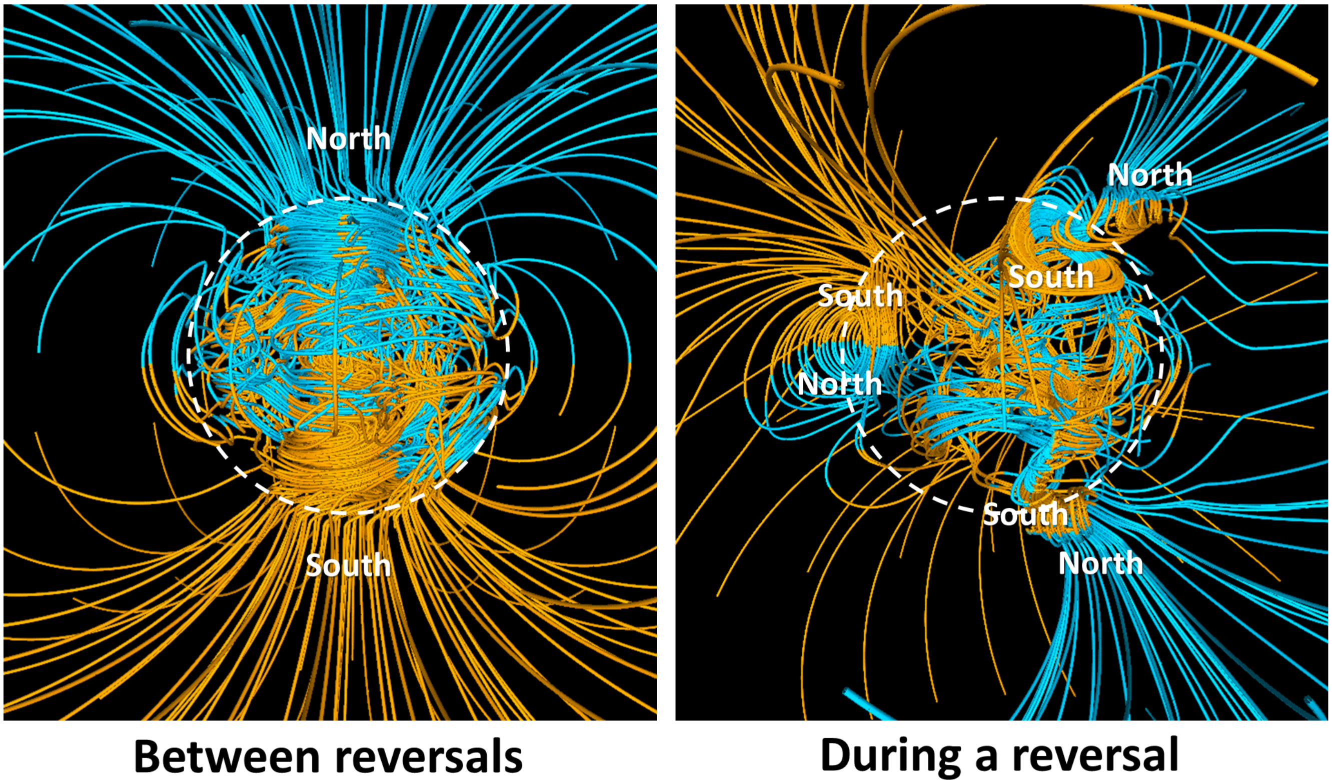

Figure 9.3.3 Depiction of Earth's magnetic field between reversals (left) and during a reversal (right). The lines represent magnetic field lines: blue where the field points toward Earth's centre and yellow where it points away. The rotation axis of Earth is vertical, and the outline of the core is shown as a dashed white circle. [Image Description]

Image Descriptions

Figure 9.3.3 image description: The earth's magnetic fields is normally very uniform with the magnetic field pointing towards the earth in the north and away from the earth in the south. During a reversal, the Earth's magnetic field becomes very convoluted. [Return to Figure 9.3.3]

Media Attributions

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- Figure 9.3.3: "Supercomputer models of Earth's magnetic field" by NASA. Public domain.

9.4 Isostasy

Theory holds that the mantle is able to convect because of its plasticity, and this plasticity also allows for another very important Earth process known as **isostasy**. The literal meaning of the word isostasy is "equal standstill," but the importance behind it is the principle that Earth's crust is *floating* on the mantle, like a raft floating in the water, rather than *resting* on the mantle like a raft sitting on the ground.

The relationship between the crust and the mantle is illustrated in Figure 9.4.1 On the right is an example of a non-isostatic relationship between a raft and solid concrete. It's possible to load the raft up with lots of people, and it still won't sink into the concrete. On the left, the relationship is an isostatic one between two different rafts and 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. We're using peanut butter here, rather than water, because its viscosity more closely represents the relationship between the crust and the mantle. Although it has about the same density as water, peanut butter is much more viscous (stiff), and so although the three-person raft will sink into the peanut butter, it will do so quite slowly.

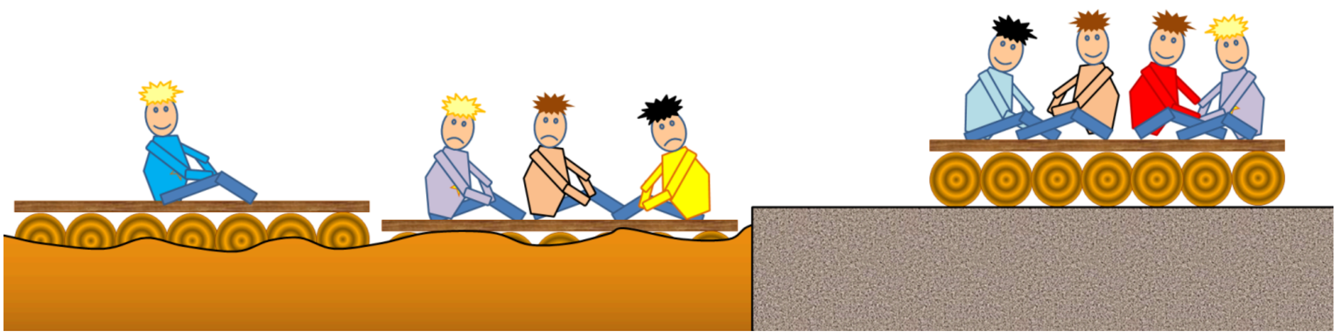


Figure 9.4.1 Illustration of a non-isostatic relationship between a raft and solid ground (right) and of isostatic relationships between rafts and peanut butter (left).

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 grams per cubic centimetre (g/cm^3), is less dense than the mantle (average density of approximately 3.4 g/cm^3 near the surface, but more than that at depth), and so it is floating on the "plastic" mantle. When more weight is added to the crust, through the process of mountain building, it slowly sinks deeper into the mantle and the mantle material that was there is pushed aside (Figure 9.4.2, left). When that weight is removed by erosion over tens of millions of years, the crust rebounds and the mantle rock flows back (Figure 9.4.2, right).

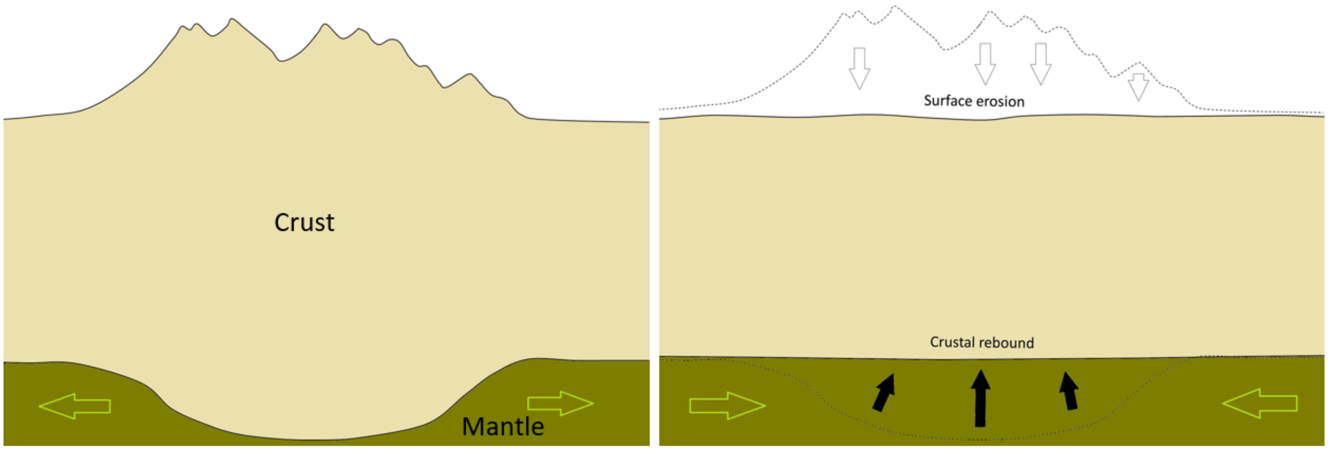


Figure 9.4.2 Illustration of the isostatic relationship between the crust and the mantle. Following a period of mountain building, mass has been added to a part of the crust, and the thickened crust has pushed down into the mantle (left). Over the following tens of millions of years, the mountain chain is eroded and the crust rebounds (right). The green arrows represent slow mantle flow.

The crust and mantle respond in a similar way to glaciation and deglaciation as they do to the growth and erosion of mountain ranges. Thick accumulations of glacial ice add weight to the crust, and as the mantle beneath is squeezed to the sides, the crust subsides. This process is illustrated for the current ice sheet on Greenland in Figure 9.4.3 (a and b). The Greenland Ice Sheet at this location is over 2,500 metres thick, and the crust beneath the thickest part has been depressed to the point where it is below sea level over a wide area. When the ice eventually melts, the crust and mantle will slowly rebound, but full rebound will likely take more than 10,000 years (Figure 9.4.3c).

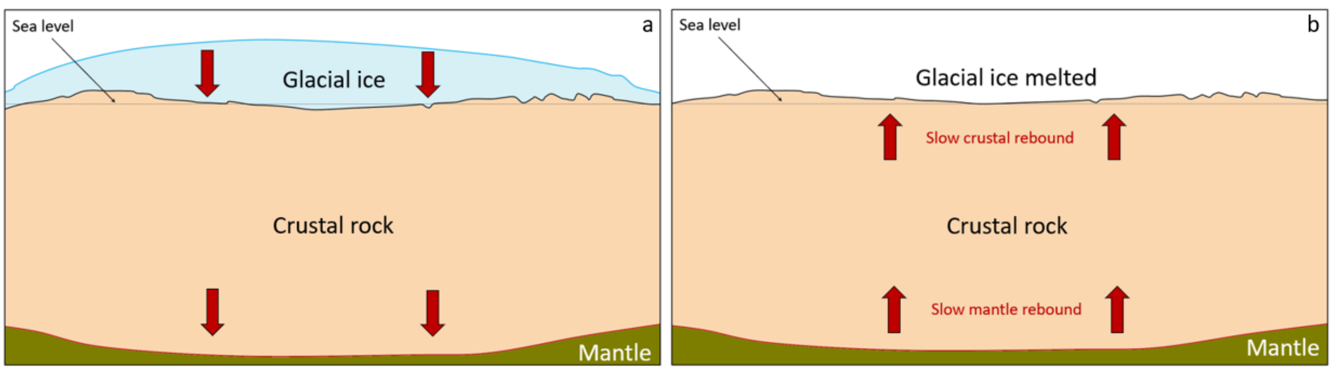


Figure 9.4.3 (a) A cross-section through the crust in the northern part of Greenland (The ice thickness is based on data from NASA and the Center for Remote Sensing of Ice Sheets, but the crust thickness is less than it should be for the sake of illustration.) The maximum ice thickness is over 2,500 m. The red arrows represent downward pressure on the mantle because of the mass of the ice. (b) Depiction of the situation after complete melting of the ice sheet, a process that could happen within 2,000 years if people and their governments continue to ignore climate change. The isostatic rebound of the mantle would not be able to keep up with this rate of melting, so for several thousand years the central part of Greenland would remain close to sea level, in some areas even below sea level.

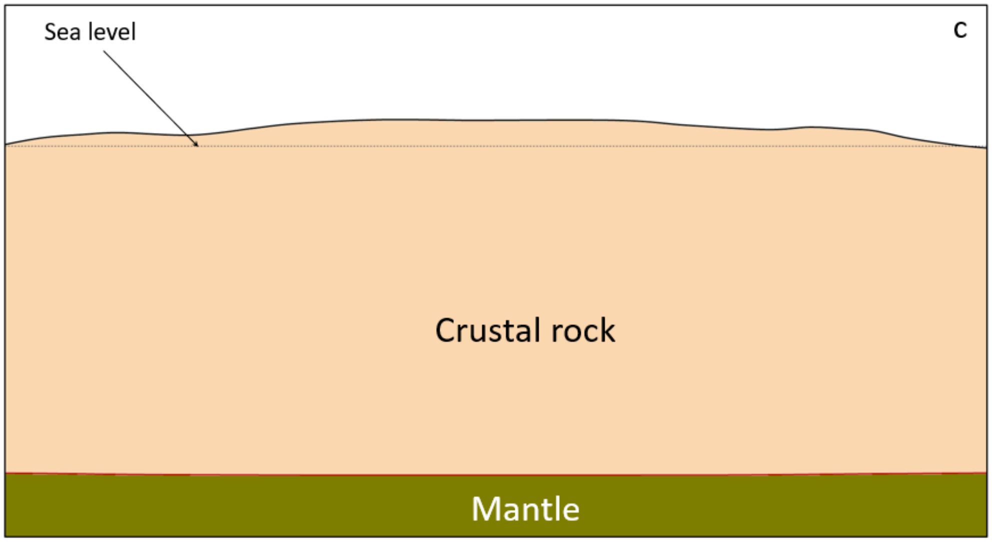


Figure 9.4.3c Depiction of the crust beneath a post-glacial Greenland once isostatic equilibrium is achieved. It is likely that complete rebound of the mantle would take more than 10,000 years.

How can the mantle be both solid and plastic?

You might be wondering how it is possible that Earth's mantle is rigid enough to break during an earthquake, 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 this is the behaviour of the material known as Silly Putty, which can bounce and will break if you pull on it sharply, but will deform like a liquid if stress is applied slowly. In this photo, Silly Putty was placed over a hole in a glass tabletop, and in response to gravity, it slowly flowed into the hole. The mantle will flow when placed under the slow but steady stress of a growing (or melting) ice sheet.

Figure 9.4.4

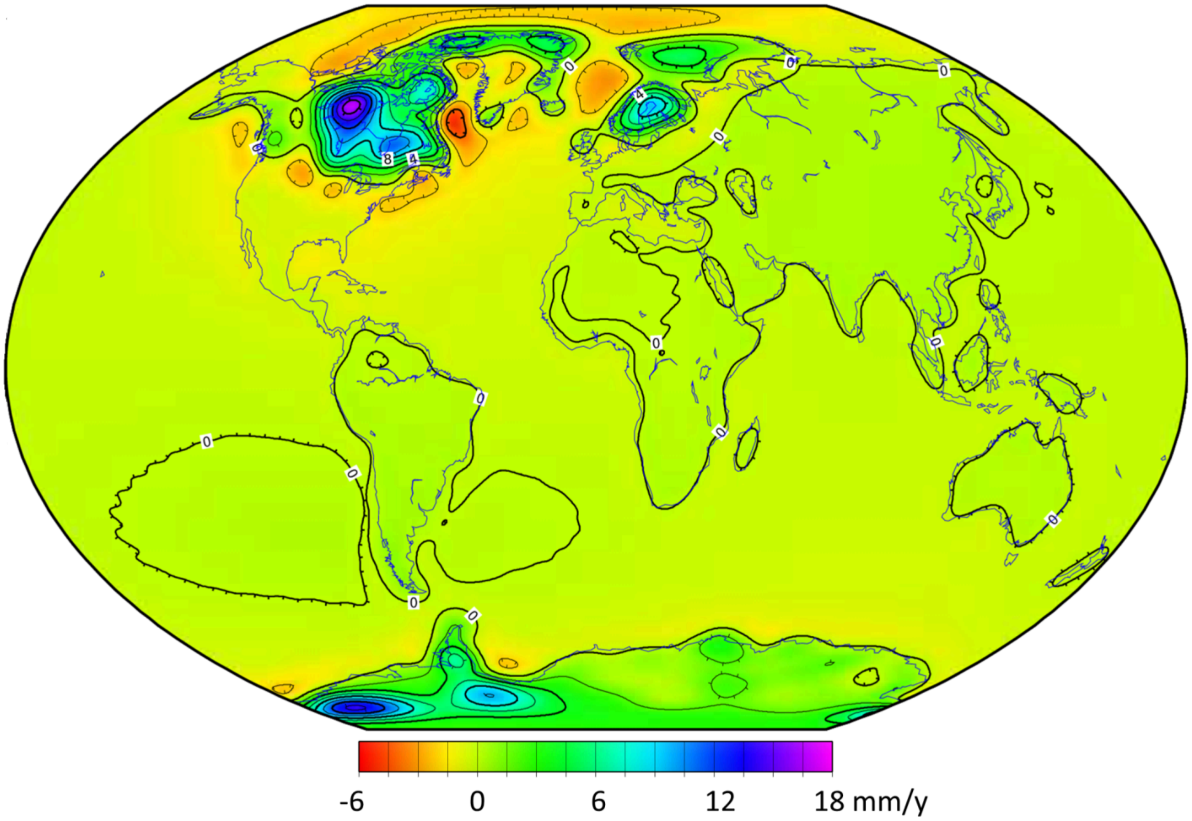


Figure 9.4.5 The current rates of post-glacial isostatic uplift (green, blue, and purple shades) and subsidence (yellow and orange). Subsidence is taking place where the mantle is slowly flowing back toward areas that are experiencing post-glacial uplift.

Large parts of Canada are still rebounding as a result of the loss of glacial ice over the past 12 ka, and as shown in Figure 9.4.5, other parts of the world are also experiencing isostatic rebound. The highest rate of uplift is in within a large area to the west of Hudson Bay, which is where the Laurentide Ice Sheet was the thickest (over 3,000 m). Ice finally left this region around 8,000 years ago, and the crust is currently rebounding at a rate of nearly 2 centimetres per 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.

There are also extensive areas of subsidence surrounding the former Laurentide and Fenno-Scandian Ice Sheets. During glaciation, mantle rock flowed away from the areas beneath the main ice sheets, and this material is now slowly flowing back, as illustrated in Figure 9.4.3b.

Exercise 9.4 Rock density and isostasy

The densities (also known as "specific gravity") of a number of common minerals are given in Table 9.1.

Table 9.1 Densities of common minerals.

Mineral	**Density (grams per cubic centimetre, g/cm^3)**
Quartz	2.65
Feldspar	2.63
Amphibole	3.25
Pyroxene	3.4
Olivine	3.3

The following table provides the approximate proportions of these minerals in the continental crust (typified by granite), oceanic crust (mostly basalt), and mantle (mainly the rock known as peridotite). Assuming that you have 1,000 cm^3 of each rock type, estimate the respective rock-type densities. For each rock type, you will need to multiply the volume of the different minerals in the rock by their density, and then add those numbers to get the total weight for 1,000 cm^3 of that rock. The density is that number divided by 1,000. The continental crust is done for you.

Table 9.2 Determine the density of different kinds of crusts

Rock Type	**Volumes of individual minerals in 1000 cm^3.**	**Grams of individual minerals in 1000 cm^3**	**Total Weight (grams)**	**Density (grams per cubic centimetre, g/cm^3)**
Continental Crust (Granite)	Quartz – 180 cm^3 Feldspar – 760 cm^3 Amphibole – 70 cm^3	Quartz – 477 g Feldspar – 1999 g Amphibole – 277 g	2703 g	2.70
Oceanic Crust (Basalt)	Feldspar – 450 cm^3 Amphibole – 50 cm^3 Pyroxene – 500 cm^3	Feldspar – Amphibole – Pyroxene –		
Mantle (Peridotite)	Pyroxene – 450 cm^3 Olivine – 550 cm^3	Pyroxene – Olivine –		

If continental crust (represented by granite) and oceanic crust (represented by basalt) are like rafts floating on the mantle, what does this tell you about how high or low they should float?

This concept is illustrated in Figure 9.4.6. The dashed line is for reference, showing points at equal distance from Earth's centre.

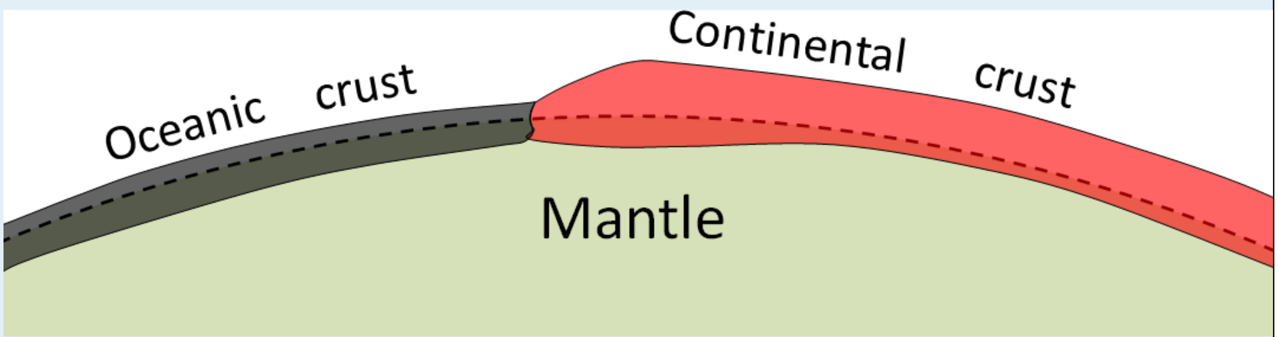


Figure 9.4.6

See Appendix 3 for Exercise 9.4 answers.

Media Attributions

- Figures 9.4.1, 9.4.2, 9.4.3abc, 9.4.6: © Steven Earle. CC BY.
- Figure 9.4.4: "Silly putty dripping" © Eric Skiff. CC BY-SA.
- Figure 9.4.5: "PGR Paulson2007 Rate of Lithospheric Uplift due to PGR" by NASA. Public domain.

Summary

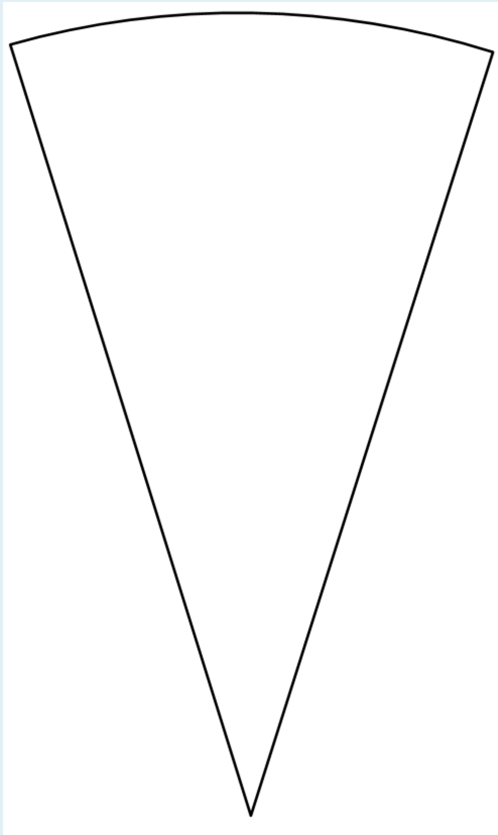
The topics covered in this chapter can be summarized as follows:

Section	Summary
9.1 Understanding Earth Through Seismology	Seismic waves that travel through Earth are either P-waves (compression, or "push" waves) or S-waves (shear waves). P-waves are faster than S-waves, and can pass through fluids. By studying seismic waves, we can discover the nature and temperature characteristics of the various parts of Earth's interior.
9.2 The Temperature of Earth's Interior	Earth's temperature increases with depth (to around 5000°C at the centre), but there are significant variations in the rate of temperature increase. These variations are related to differences in composition and the existence of convection in the mantle and liquid part of the core.
9.3 Earth's Magnetic Field	Because of outer-core convection, Earth has a magnetic field. The magnetic force directions are different at different latitudes. The polarity of the field is not constant, and has flipped from "normal" (as it is now) to reversed and back to normal hundreds of times in the past.
9.4 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 because of mountain building, it pushes farther down into the mantle. Oceanic crust, being heavier than continental crust, floats lower on the mantle.

Questions for Review

Answers to Review Questions can be found in Appendix 2.

1. What parts of Earth are most closely represented by typical stony meteorites and typical iron meteorites?
2. On the below diagram draw (from memory) and label the approximate locations of the following boundaries: crust/mantle, mantle/core, outer core/inner core.



3. Describe the important differences between P-waves and S-waves.
4. Why does P-wave velocity decrease dramatically 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 is the mechanism that causes it?
7. Where and 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. As you can see in Figure 9.22, British Columbia is still experiencing weak post-glacial isostatic uplift, especially in the interior, but also along the coast. Meanwhile offshore areas are experiencing weak isostatic subsidence. Why?