





Volcanic eruptions, such as this one at Italy's Mount Etna, provide data about the nature of Earth's interior (Photo by Marco Fulle).

*This chapter was prepared by Professor Michael Wyession, Washington University.

If you could slice any planet in half, the first thing you would notice is that it would be divided into distinct layers. The heaviest materials (metals) would be at the bottom. Lighter solids (rocks) would be in the middle. Liquids and gases would be at the top. Within Earth, we know these layers as the iron core, the rocky mantle and crust, the liquid ocean, and the gaseous atmosphere. More than 95 percent of the variations in composition and temperature within Earth are due to layering. However, this is not the end of the story. If it were, Earth would be a dead, lifeless cinder floating in space.

There are also small horizontal variations in composition and temperature at depth that indicate the interior of our planet is very active. The rocks of the mantle and crust are in constant motion, not only moving about through plate tectonics, but also continuously recycling between the surface and the deep interior. It is also from Earth's deep interior that the water and air of our oceans and atmosphere are replenished, allowing life to exist at the surface.

Discovering and identifying the patterns of Earth's deep motions have not been easy. Light does not travel through rock, so we must find other ways to "see" into our planet. The seismic waves associated with earthquakes are one means used to investigate Earth's interior. Other techniques include mineral physics experiments that can recreate the conditions of extreme temperature and pressure inside planets and gravity measurements that show where there are internal variations in the distribution of mass. Examining Earth's magnetic field gives clues to the patterns of flow of liquid iron in the core. Taken together, all these different fields of study give us a picture of Earth as a churning, varied, complex planet that continues to change and evolve over time.

Gravity and Layered Planets

If a bottle were filled with clay, iron filings, water, and air and then shaken, it would appear to have a single, muddy composition. If that bottle were allowed to sit, however, the different materials would settle out into layers. The iron filings, which are the densest, would sink to the bottom. Above the iron would be the clay, then water, then air. This is what happens inside planets. At their birth, planets form from an accumulation of nebular debris but quickly begin to form layers. The iron sinks to form the core, rock forms the mantle and crust, and gases form the atmosphere. All large bodies in the solar system have iron cores and rocky mantles, even Jupiter, Saturn, and the Sun. A profile of Earth's layered structure is shown in Figure 12.1.

For both the bottle of mud and Earth it is the force of gravity that is responsible for the layering. Figure 12.1 shows another interesting effect of gravity. Not only does the density change between layers, but it changes within layers. This is because materials compress when you squeeze them. Rock having the composition of the upper mantle has a density of about 3.3 g/cm^3 at Earth's surface. But, take that rock to the base of the mantle and its density increases to 5.6 g/cm^3 , nearly twice as much. The intense pressure of the overlying rock causes rock at the base of the mantle to be compressed into nearly half its volume!

This increase in density occurs partly because the intense pressure causes atoms to shrink in size. However, atoms do not all compress at the same rate. It is easier to compress negative ions than positive ions. Negative ions have more electrons than protons, and tend to be "fluffier" than positive ions. When rocks are squeezed, the negative ions (such as O^{2-}) compress more easily than the positive ions (such as Si^{+4} and Mg^{+2}), so the ratios of the ionic sizes change. When these ratios change enough, the structure of a mineral is no longer stable, and the atoms rearrange into a more stable and denser structure. This is called a *mineral phase change*, as discussed in Chapter 3. The increase in density of mantle rocks is due both to the compression of existing minerals and to the transition to new "high-pressure" minerals.

Probing Earth's Interior: "Seeing" Seismic Waves

The best way to learn about Earth's interior is to dig or drill a hole and examine it directly. Unfortunately, this is only possible at shallow depths. The deepest a drill has ever penetrated is only 12.3 kilometers, which is about 1/500 of the way to Earth's center! Even this was an extraordinary accomplishment because temperature and pressure increase so rapidly with depth.

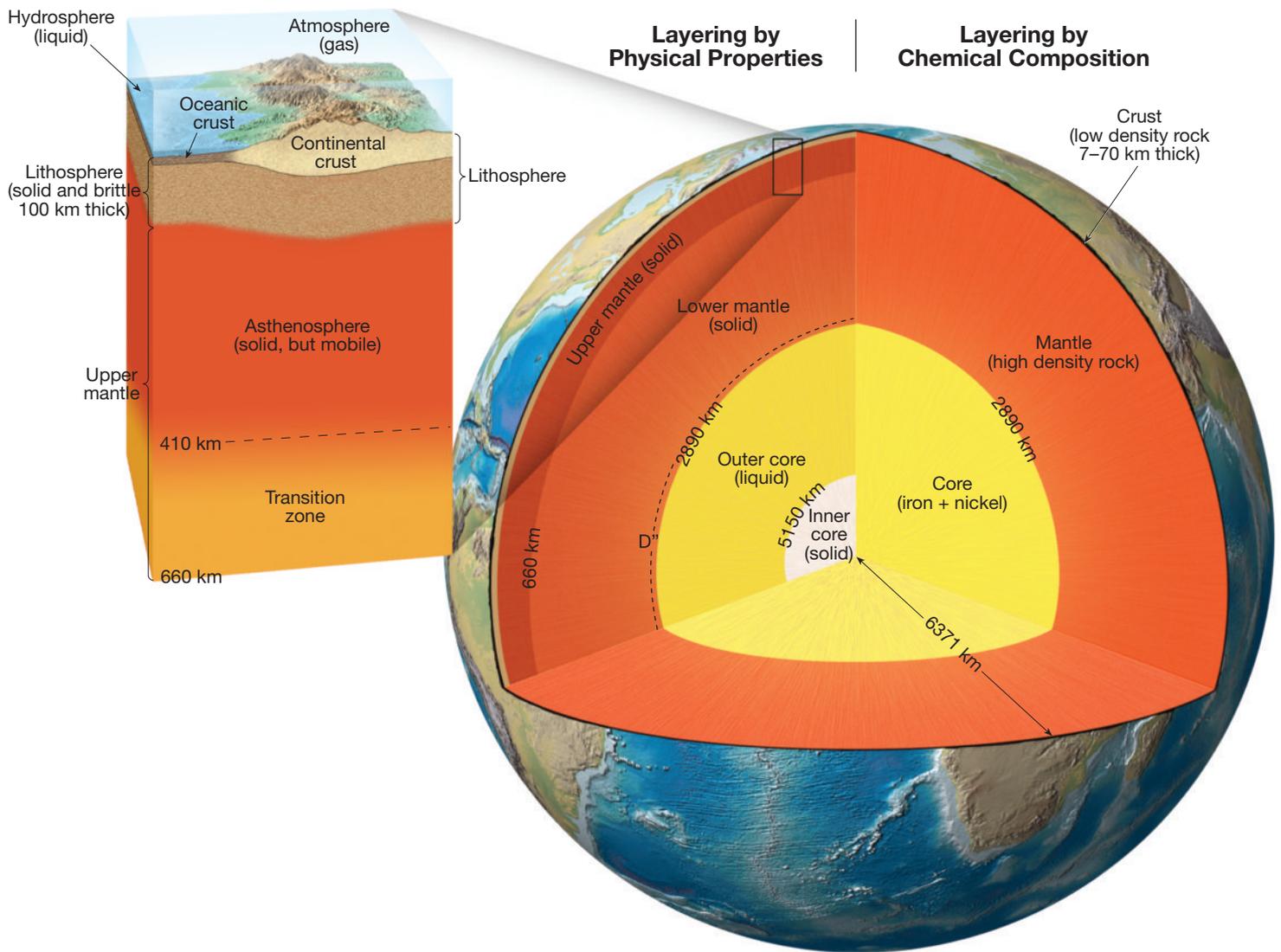


FIGURE 12.1 The layers of Earth, shown both in terms of physical properties and chemical composition. The physical properties of Earth's layers (left) include the physical state of the material (solid, liquid, or gas) as well as how stiff the material is (for example, the distinction between the lithosphere and asthenosphere). The chemical layers are mainly determined by density, with the heaviest materials at the center and the lightest ones at the surface.

Fortunately, many earthquakes are large enough that their seismic waves travel all the way through Earth and can be recorded on the other side (Figure 12.2). This means that the seismic waves act like medical X rays used to take images of a person's insides. There are about 100 to 200 earthquakes each year that are large enough (about $M_w > 6$) to be well-recorded by seismographs all around the globe. These large earthquakes provide the means to "see" into our planet. Consequently, they have been the source of most of the data that allowed us to figure out the nature of Earth's interior.

Interpreting the waves recorded on seismograms in order to identify Earth structure is difficult. This is because seismic waves usually do not travel along straight paths. Instead, seismic waves are reflected, refracted, and diffracted as they pass through our planet. They reflect off of

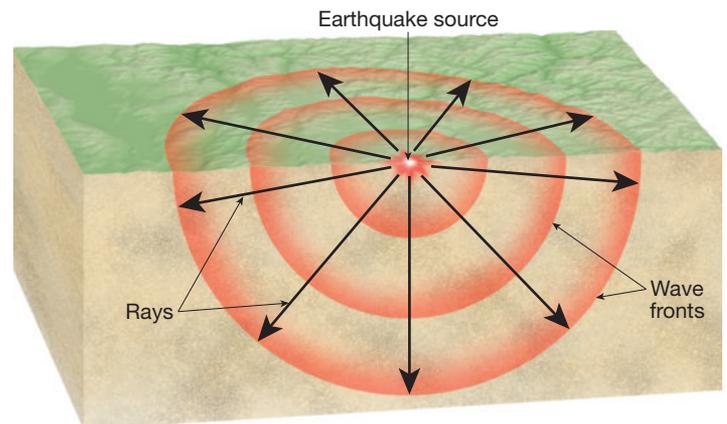


FIGURE 12.2 When traveling through a medium with uniform properties, seismic waves spread out from an earthquake source (focus) as spherically shaped structures called *wave fronts*. It is common practice however, to consider the paths taken by these waves as *seismic rays*, lines drawn perpendicular to the wave front as shown in this diagram.

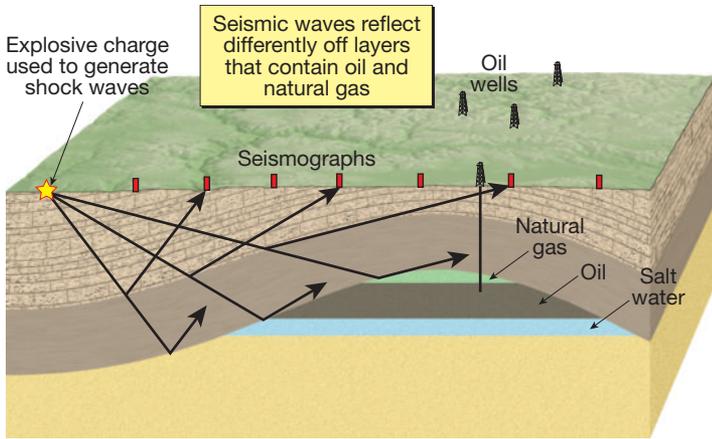


FIGURE 12.3 Reflected seismic waves are used to search for oil and natural gas underground. The seismic waves from explosions reflect differently from layers of rock that contain liquid oil and natural gas, and thus are used to map petroleum reservoirs in Earth's crust.

boundaries between different layers, they refract (or bend) when passing from one layer to another layer, and they diffract around any obstacles they encounter. These different wave behaviors have been used to identify the boundaries that exist within Earth.

As Figure 12.3 shows, changes in the composition or structure of rock cause seismic waves to reflect off of boundaries between different materials. This is especially

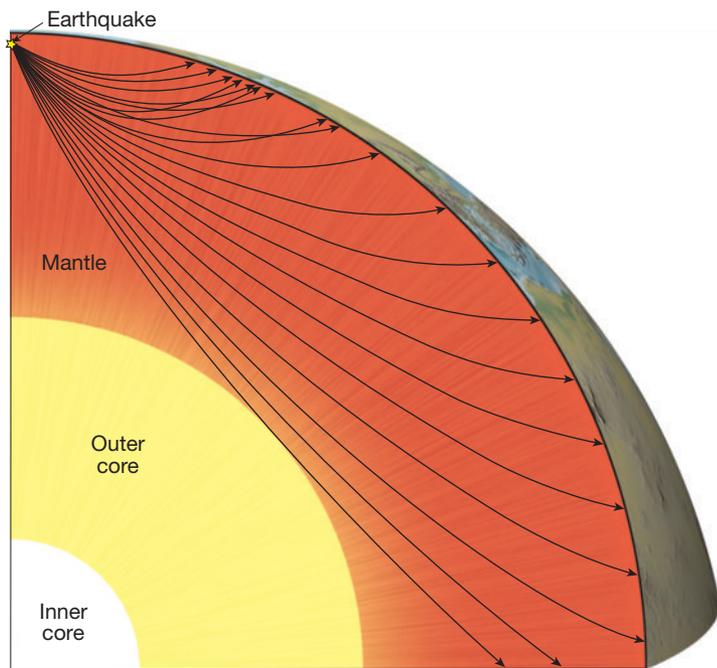


FIGURE 12.4 Slice through Earth's mantle showing some of the ray paths that seismic waves from an earthquake would take. The rays follow curved (refracting) paths rather than straight paths because the seismic velocity of rocks increases with depth in the mantle, a result of increasing pressure with depth. Notice the complicated ray paths in the upper mantle, with some even crossing each other. This is due to the sudden seismic velocity increases that result from mineral phase changes at increasing pressures.

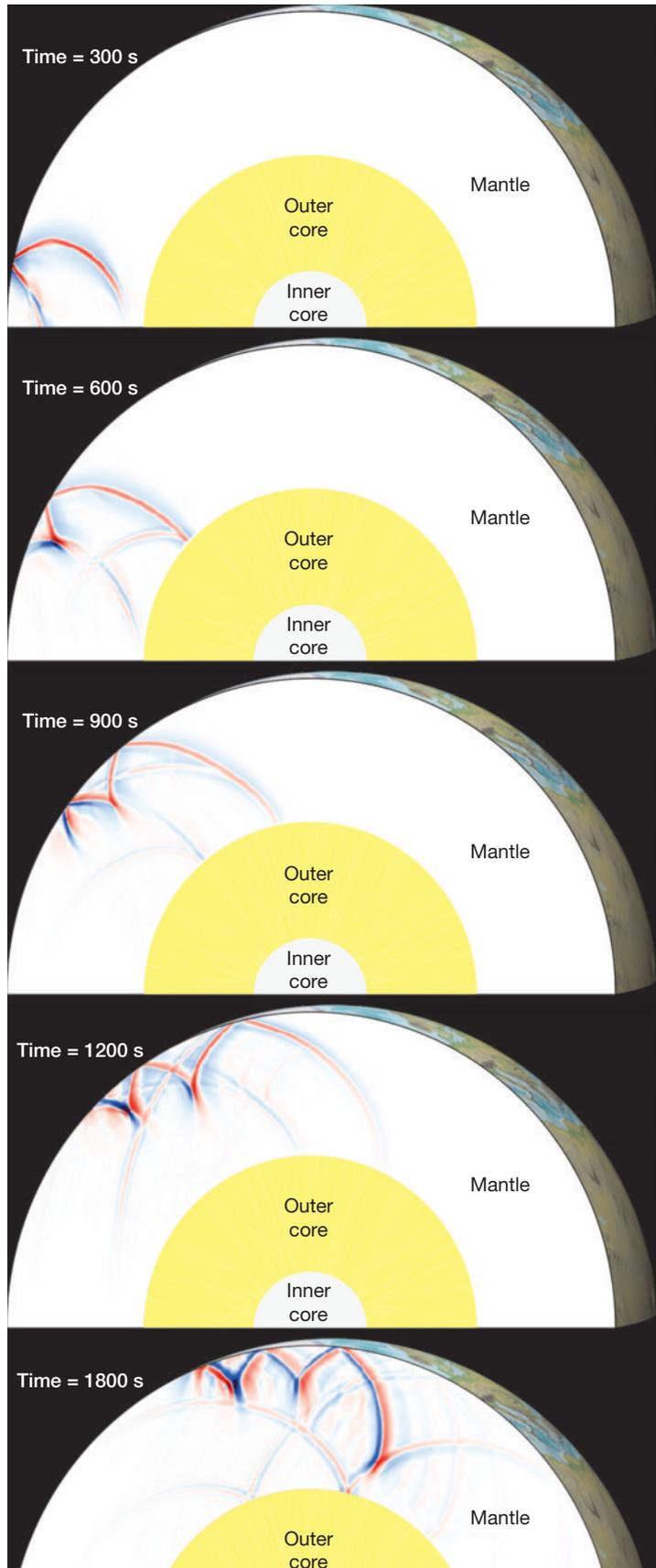


FIGURE 12.5 Five snapshots in time showing the locations of S waves within Earth's mantle following an earthquake. In addition to refracting and diffracting, S waves reflect from boundaries such as the core-mantle boundary. Note that S waves do not penetrate the outer core because S waves do not travel through liquids.

BOX 12.1 ▶ UNDERSTANDING EARTH

Re-creating the Deep Earth

Seismology alone cannot determine what Earth is made of. Additional information must be obtained by some other means so that the seismic velocities can be interpreted in terms of rock type. This is done using *mineral physics* experiments performed in laboratories. By squeezing and heating minerals and rocks, physical properties like stiffness, compressibility, and density (and therefore seismic velocities) can be directly measured. This means that the conditions of the mantle and core can be simulated, and the results compared to seismic modeling.

Most mineral physics experiments are done using giant presses involving very hard carbonized steel. The highest pressures, however, are obtained using diamond-anvil presses like the one shown in Figure 12.A. These take advantage of two important characteristics of diamonds—their hardness and transparency. The tips of two diamonds are cut off, and a small sample of mineral or rock is placed in between. Pressures as great as those in the interior of Jupiter have been obtained by squeezing the two diamonds together.

High temperatures are achieved by firing a laser beam through the diamond and into the mineral sample.

Besides measuring seismic velocities at the conditions of different depths within Earth, there are other important mineral

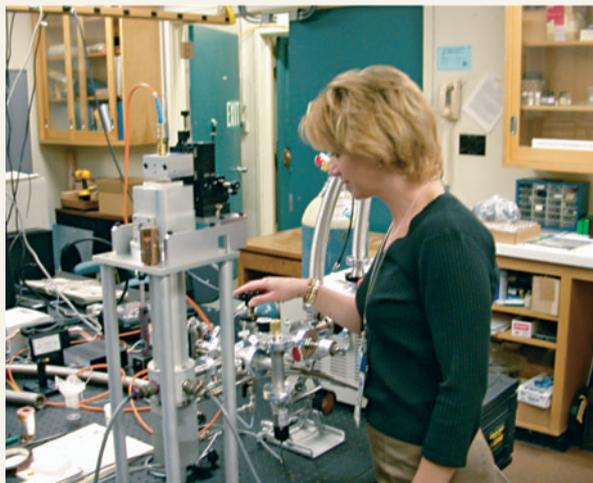


FIGURE 12.A High-pressure experiments inside a diamond anvil cell (left photo) can recreate the conditions at the center of a planet. The whole apparatus is small, and can fit on top of a table. High pressures are generated by cutting the tips off of high-quality diamonds (right photo), putting a small sample of rock between, squeezing the diamonds together and heating the sample with a laser. (Left photo courtesy of Lawrence Livermore National Laboratory; right photo by Douglass L. Peck Photography)

physics experiments. One experiment determines the temperature at which minerals will begin to melt under various pressures. Another experiment determines (at different temperatures) the pressures at which one mineral phase will become unstable and convert into a new “high-pres-

sure” phase. Yet another involves making these same tests for slightly different mineral compositions. All of these experiments are needed because, as will be discussed later, there are three-dimensional changes in composition and temperature within Earth.

important in the exploration for oil and natural gas where artificially generated seismic waves are used to probe the crust. Petroleum tends to get trapped in certain kinds of geological structures, and these structures are identified by mapping out the layering of the upper crust. The price of gasoline would be much more expensive without the existence of seismic imaging because a huge number of drill wells would have to be randomly deployed to find oil. Using seismic waves, companies can just drill in the places that are most likely to have petroleum. Seismic waves also reflect off of boundaries between the crust, mantle, outer core, inner core, and off of other boundaries within the mantle as well.

One of the most noticeable behaviors of seismic waves is that they follow strongly curved paths (Figure 12.4). This occurs because the velocity of seismic waves generally increases with depth. In addition, seismic waves travel faster when rock is stiffer or less compressible. These properties of stiffness and compressibility are then used to interpret the composition and temperature of the rock. For instance, when rock is hotter it becomes less stiff (imagine taking a frozen chocolate bar and then heating it up!), and waves

travel more slowly. Waves also travel at different speeds through rocks of different compositions. Thus the speed that seismic waves travel can help determine both the kind of rock that is inside Earth and how hot it is.

Within Earth’s mantle, where there are both sharp boundaries and gradual seismic velocity changes, the pattern of seismic waves becomes quite complex. Figure 12.5 shows what S waves from a deep earthquake look like as they travel through the mantle. Note how the single wave from the shock is soon broken up into many different waves, which appear on the seismograms as separate signals.

Earth's Layers



Earth's Interior

▶ Earth's Layered Structure

Combining the data obtained from seismological studies and mineral physics has given us a layer-by-layer understanding of the composition of Earth (see Box 12.1). Seismic velocities,

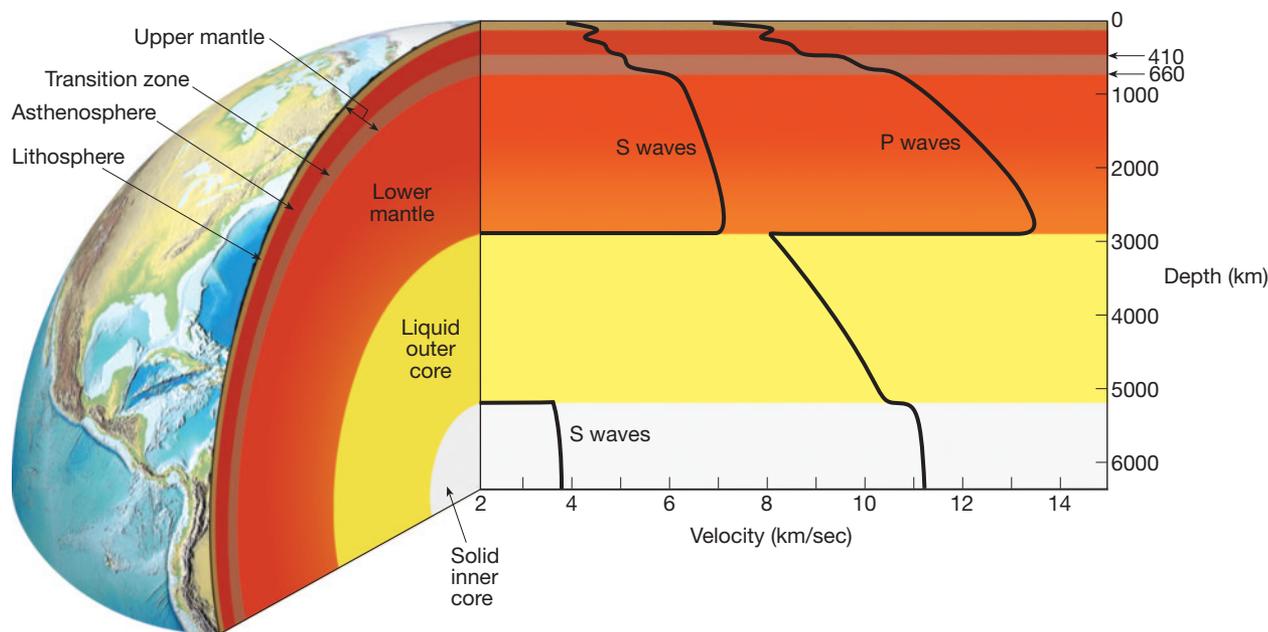


FIGURE 12.6 Cutaway of Earth showing its different layers and the average velocities of P and S waves at each depth. S waves are an indication of how rigid the material is—the inner core is less rigid than the mantle, and the liquid outer core has no rigidity.

as a function of depth, are shown in Figure 12.6. By examining the behavior of a variety of rocks at the pressures corresponding to these depths, geologists have been able to figure out the compositions of Earth's crust, mantle, and core.

Earth's Crust

Earth's **crust** is of two different types—continental crust and oceanic crust. Both share the word “crust,” but the similarity ends there. Continental and oceanic crusts have very different compositions, histories, ages, and styles of formation. In fact, the ocean crust is much more similar to rock of the mantle than to rock of the continental crust.

Oceanic Crust Seismic imaging has shown that the ocean crust is usually about 7 kilometers (4.5 miles) thick. All ocean crust forms at mid-ocean ridges, which separate two diverging tectonic plates. Ocean crust has P wave velocities of about 5–7 km/s and a density of about 3.0 g/cm³, which agrees with experimental values for the rocks basalt and gabbro. The composition and formation of ocean crust is discussed further in Chapter 13.

Continental Crust While oceanic crust is fairly uniform throughout the oceans, no two continental regions have the same structure or composition. Continental crust averages about 40 kilometers (25 miles) in thickness, but can be more than 70 kilometers (45 miles) thick in certain mountainous regions like the Himalayas and Andes. The thinnest crust in North America is beneath the Basin and Range region of the western United States, where the crust can be as thin as 20 kilometers (12 miles). The thickest North American crust, beneath the Rockies, is more than 50 kilometers (30 miles) thick.

Seismic velocities within continents are quite variable, suggesting that the composition of continental crust must also vary greatly. This agrees with what has been learned about the different ways continents can form, discussed in Chapter 22. In general, however, continents have a density of about 2.7 gm/cm³, which is much lower than both oceanic crust and mantle rock. This lower density explains why continents are buoyant—acting like giant rafts, floating atop tectonic plates, and why they cannot be subducted into the mantle.

Discovering Boundaries: The Moho The boundary between the crust and mantle, called the **Moho**, was one of the first features of Earth's interior discovered using seismic waves. Croatian seismologist Andrija Mohorovičić discovered this boundary in 1909, which is named in his honor. At the base of the continents P waves travel about 6 km/s but abruptly increase to 8 km/s at a slightly greater depth.

Andrija Mohorovičić cleverly used this large jump in seismic velocity to discover the Moho. He noticed that there were two different sets of seismic waves that were recorded at seismographs located within a few hundred kilometers of an earthquake. One set of waves moved across the ground at about 6 km/s, and the other set of waves moved across the ground at about 8 km/s. From these two waves, Mohorovičić correctly determined that the different waves were coming from two different layers, as shown in Figure 12.7.

When a shallow earthquake occurs, there is a *direct wave* that moves straight through the crust and is recorded at nearby seismographs. (In Figure 12.7, the slope of the line gives the velocity of direct waves through the crust.) Seismic waves will also follow a path down through the crust and along the top of the mantle. These are called *refracted waves*

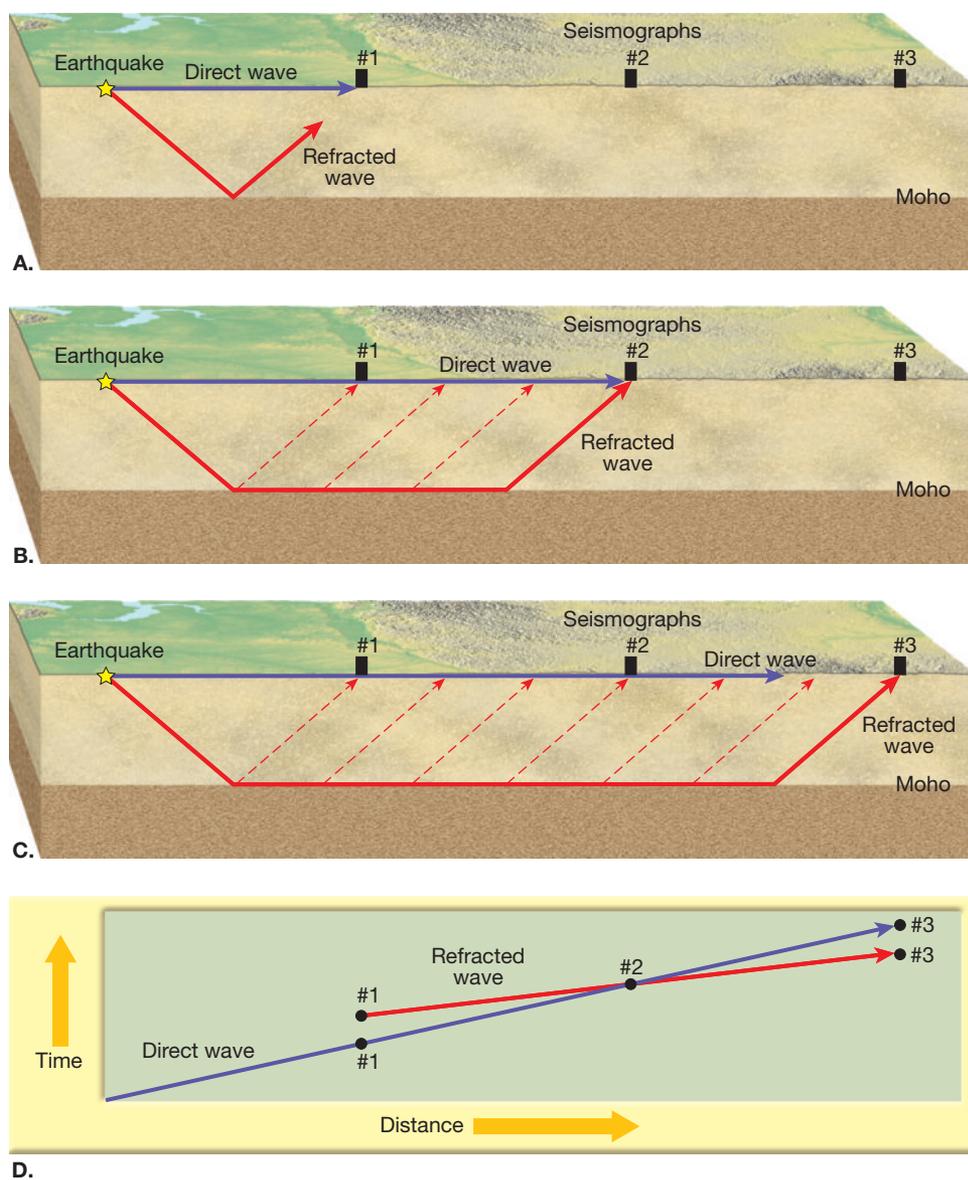


FIGURE 12.7 Diagram showing seismic waves from an earthquake arriving at three different seismographs. Over a short distance, such as at seismograph #1, the direct wave arrives first. For greater distances, such as at seismograph #3, the refracted wave arrives first. At the cross-over point, which in this diagram occurs at seismograph #2, both waves arrive at the same time. The distance to the cross-over point increases with the depth of the Moho, and therefore can be used to determine the thickness of the crust.

because they are bent, or refracted, as they enter the mantle. These refracted P waves will travel across the ground at the speed of the waves in the mantle (8 km/s). At nearby distances the direct wave arrives first. However, at greater distances the refracted wave is the first to arrive. The point at which both waves arrive at the same time, called the *cross-over*, can be used to determine the depth of the Moho. Thus, using just these two waves and an array of seismographs, you can determine the thickness of the crust for any location.

The difference between direct and refracted waves is analogous to driving along local roads or taking the interstate highway. For short distances you will arrive faster if you just take the local roads. For greater distances, the short-

er time occurs if you first drive to the interstate and travel along it. The cross-over point, where both routes take the same amount of time, is directly related to how far you are from a major highway. (Or, when determining the depth of the Moho, how far the mantle [fast layer] is from the surface.)

Earth's Mantle

More than 82 percent of Earth's volume is contained within the **mantle**, a nearly 2900-kilometer thick shell extending from the base of the crust (Moho) to the liquid outer core. Because S waves readily travel through the mantle, we know that it is a solid rocky layer composed of silicate minerals that are enriched in iron and magnesium. However, despite its solid nature rock in the mantle is quite hot and capable of flow, albeit at very slow velocities.

The Upper Mantle The upper mantle extends from the Moho down to a depth of about 660 km. The upper mantle can be divided into three different parts. The top layer of the upper mantle is part of the stiff **lithosphere**, and beneath that is the weaker **asthenosphere**. These layers are a result of the temperature structure of Earth, and so are discussed later in this chapter. The bottom part of the upper mantle is called the *transition zone*.

We have a good sense of what the upper mantle is made of because mantle rocks are brought to the surface by several different geological processes. The seismic velocities we observe for the mantle are consistent with a rock called peridotite. Mantle *peridotite* is an ultramafic rock mostly composed of the

minerals olivine and pyroxene. It is richer in the metals magnesium and iron than the minerals found in either the continental or oceanic crust.

The olivine crystals in peridotite display a very important property called **seismic anisotropy**, which means that seismic waves travel at different speeds along different paths through the crystals. With olivine, the crystals tend to line up with their fast directions pointing in the same direction that the rock is flowing. This is very fortunate for geologists, because if the fastest seismic wave direction through regions of the upper mantle can be found, then the direction the olivine is moving is also found. Seismology therefore not only provides a snapshot of the structure of Earth's

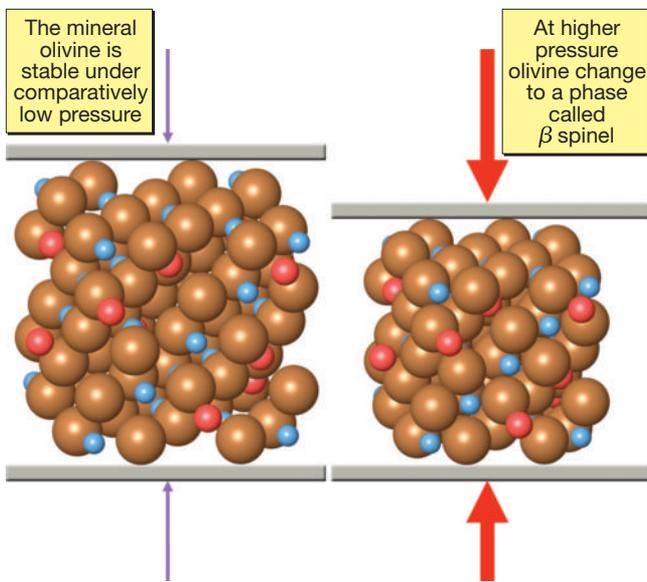


FIGURE 12.8 Demonstration of the effect of pressure on the structure of minerals. The mineral olivine, which is stable in the upper mantle, is no longer stable at the pressures in the transition zone, and converts to denser phases. In the top of the transition zone olivine converts to a mineral phase called β -spinel. The atoms are the same, but they are compressed into a more compact crystalline structure.

interior at this very moment but also a sense of where the rock within Earth will move to in the future.

Transition Zone From about 410 km to about 660 km in depth is the part of the upper mantle called the **transition zone**. The top of the transition zone is identified by a sudden increase in density from about 3.5 to 3.7 g/cm³. Like the Moho, this boundary reflects seismic waves. Unlike the Moho, this boundary is not due to a change in chemical composition but to a change in mineral phase. The chemical composition above and below the 410-kilometer discontinuity is the same. However, the mineral olivine, which is stable in the upper mantle, is no longer stable at the pressures in the transition zone and converts to denser phases (Figure 12.8). In the top half of the transition zone olivine converts to a phase called β -spinel, and in the bottom half β -spinel converts into a true spinel structure called ringwoodite.

The most unusual thing about the transition zone is that it is capable of holding a great deal of water, up to 2 percent by weight. This is much more than for the peridotite of the upper mantle, which can only hold about

0.1 percent of its weight as water. Because the transition zone is 10 percent of the volume of Earth, it could potentially hold up to five times the volume of Earth's oceans. Water cycles slowly through the planet, brought down into the mantle with subducting oceanic lithosphere and carried upward by rising plumes of mantle rock. How much water is actually contained within the transition zone is not known.

The Lower Mantle From 660 kilometers deep to the top of the core, at a depth of 2891 kilometers, is the **lower mantle**. Beneath the 660-kilometer discontinuity, both olivine and pyroxene take the form of the mineral *perovskite* (Fe, Mg) SiO₃. The lower mantle is by far the largest layer of Earth, occupying 56 percent of the volume of the planet. This means that perovskite is the single most abundant material within Earth.

The D'' Layer In the bottom few hundred kilometers of the mantle, just above the core, is a highly variable and unusual layer called the **D'' layer** (pronounced "dee double-prime"). This is a boundary layer between the rocky mantle and the liquid iron outer core (Figure 12.9). The D'' layer is a

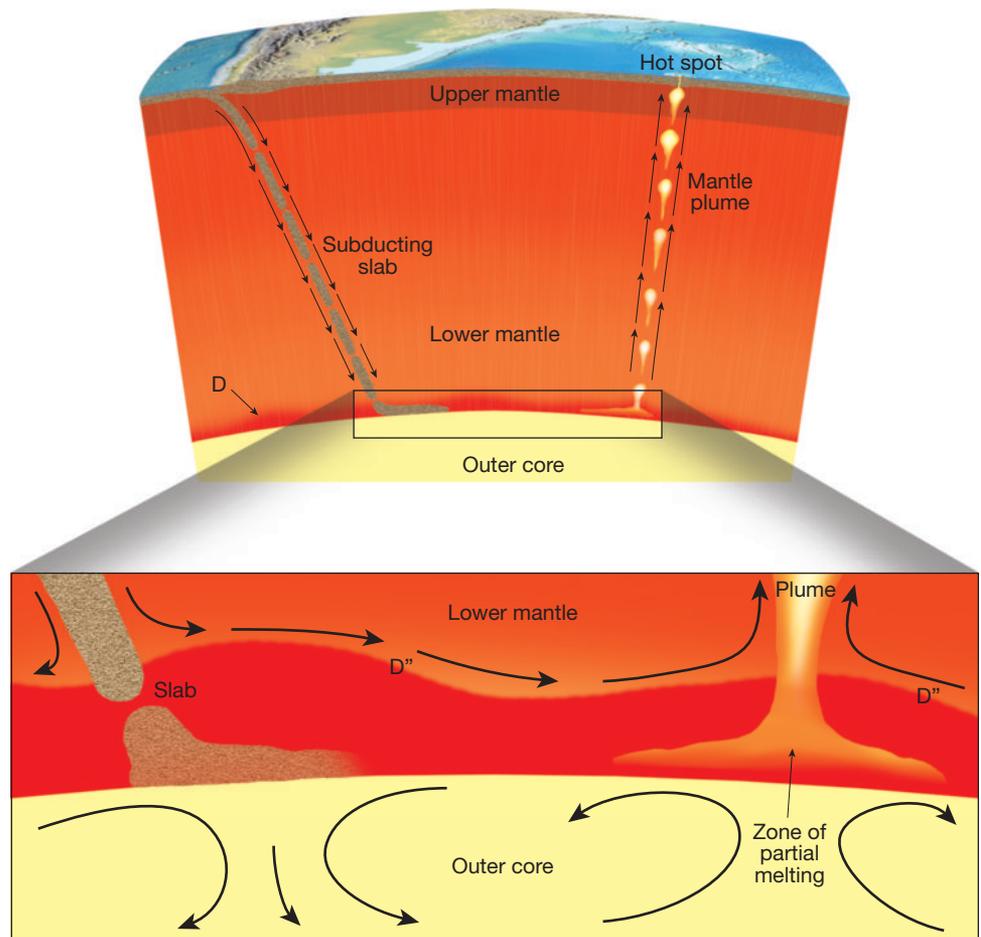


FIGURE 12.9 Schematic of the variable and unusual D'' layer at the base of the mantle. Like the lithosphere at the top of the mantle, the D'' layer contains large horizontal variations in both temperature and composition. Many scientists believe that D'' is the graveyard of some subducted ocean lithosphere and the birthplace of some mantle plumes.

lot like the lithosphere, which is the boundary layer between the mantle and the ocean/atmosphere layer. Both the lithosphere and D" layer have large variations in composition and temperature. The difference in lithospheric temperature between hot mid-ocean ridges and cold abyssal seafloor is more than 1000°C. The horizontal changes in temperature within the D" layer are similar. The composition of the lithosphere varies greatly, with either continental or ocean crust embedded in it. There also seem to be large slabs of differing rock types embedded within D".

The very base of D", the part of the mantle directly in contact with the hot liquid iron core, is like Earth's surface in that there are "upside-down mountains" of rock that protrude into the core. Furthermore, in some regions of the core-mantle boundary, the base of D" seems to be hot enough to be partially molten. This may be the cause of narrow zones at the very base of the mantle where P-wave velocities decrease by 10 percent and S-wave velocities decrease by 30 percent.

Discovering Boundaries: The Core-Mantle Boundary Evidence that Earth has a distinct central core was uncovered in 1906 by a British geologist, Richard Dixon Oldham. (In 1914 Beno Gutenberg calculated the depth to the core boundary

as 2900 kilometers, a value that has stood the test of time.) Oldham observed that at distances of more than about 100° from a large earthquake P and S waves were absent, or very weak. In other words, the central core produced a "shadow zone" for seismic waves as shown in Figure 12.10.

As Oldham predicted, Earth's core exhibits markedly different elastic properties from the mantle above, which causes considerable refraction of P waves—similar to how light is refracted (bent) as it passes from air to water. In addition, because the outer core is liquid iron, it blocks the transmission of S waves (recall S waves do not travel through liquids).

Figure 12.10 shows the locations of the P and S wave shadow zones and how the paths of the waves are affected by the core. While there are still P and S waves that arrive in the shadow zone, they are all very different than would be expected for a planet without a core.

Earth's Core

The Outer Core The boundary between the mantle and the outer core, called the *core-mantle boundary*, is the most significant within Earth in terms of changes in material properties. P waves drop from 13.7 to 8.1 km/s at the core-mantle

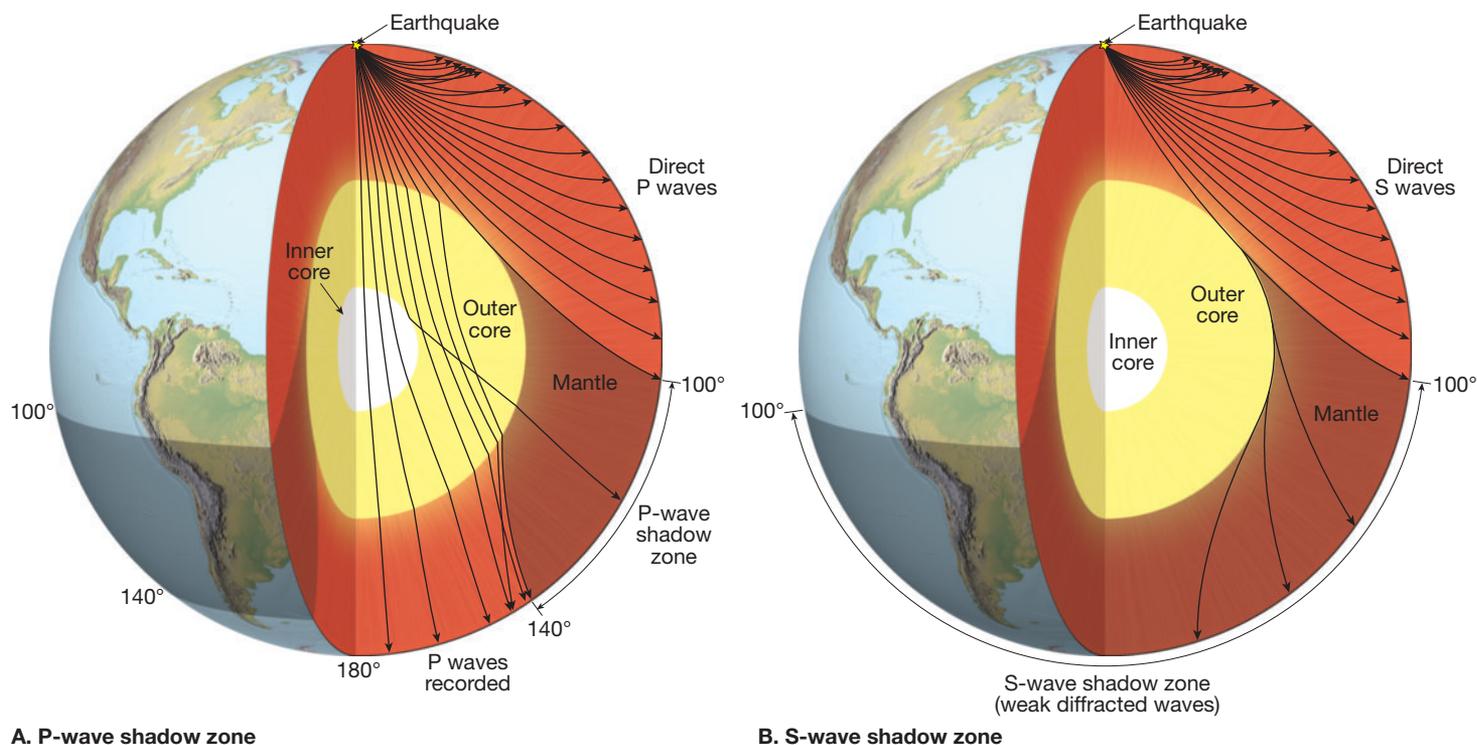


FIGURE 12.10 Two views of Earth's interior showing the effects of the outer and inner cores on the ray paths of P and S waves. **A.** When P waves interact with the slow-velocity liquid iron of the outer core, their rays are refracted downward. This creates a shadow zone where no direct P waves are recorded (although diffracted P waves travel there). The P waves that travel through the core are called PKP waves. The "K" represents the path through the core, and comes from the German word for core, which is *kern*. The increase in seismic velocity at the top of the inner core can refract waves sharply so that some arrive within the shadow zone, which is shown here as a single ray. **B.** The core is an obstacle to S waves, because they cannot pass through liquids. Therefore, a large shadow zone exists for S waves. However, some S waves diffract around the core and are recorded on the other side of the planet.

boundary, and S waves drop dramatically from 7.3 km/s to zero. Because S waves do not pass through liquids, the lack of any S waves in the **outer core** means that it is liquid. The change in density, from 5.6 to 9.9 g/cm³, is even larger than the rock–air difference observed at Earth's surface.

Based on our knowledge of the composition of meteorites and the Sun, geologists expect Earth to contain a great deal of iron. However, this iron is mostly missing from the crust and mantle. This fact, plus the great density of the core, tells us that it is mostly made of iron and some nickel, which has a similar density as iron.

The **core** is only about 1/6 of Earth's volume, but because iron is so dense, the core accounts for 1/3 of Earth's mass, and iron is Earth's most abundant element when measured by mass. The outer core is not pure iron, however. Its density and seismic velocities suggest that the outer core contains about 15 percent of other, lighter elements. These are likely to include sulfur, oxygen, silicon, and hydrogen. Based on mineral physics experiments, this is not surprising. For instance, pure iron melts at a very high temperature, but an iron-sulfur mixture melts at a much lower temperature. When Earth was forming and heating up, the iron that sank to form the core melted more easily in the presence of the sulfur, pulling it down into the core as well.

The Inner Core At the center of the core is a solid sphere of iron with lesser amounts of nickel called the **inner core**. In drawings like Figure 12.1 the inner core looks much larger than it really is. The inner core is actually very small, only 1/142 (less than one percent) of the volume of Earth. The inner core did not exist early in Earth's history, when the planet was hotter. However, as the planet cooled, iron began to crystallize at the center to form the solid inner core. Even today, the inner core continues to grow in size as the planet cools. The inner core does not contain the quantity of light elements found in the outer core.

The inner core is separated from the mantle by the liquid outer core, and is therefore free to move independently. Recent studies suggest that the inner core is actually rotating faster than the crust and mantle, lapping them every few hundred years (Figure 12.11). The inner core's small size and great distance from the surface make it the most difficult region within Earth to examine.

Discovering Boundaries: The Inner Core–Outer Core Boundary The boundary between the solid inner core and liquid outer core was discovered in 1936 by Danish seismologist Inge Lehman. She could not tell whether the inner core was actually solid or not, but using basic trigonometry reasoned that some P waves were being strongly refracted by a sudden increase in seismic velocities at the inner core–outer core boundary. This is the opposite situation as what occurs to produce the P-wave shadow zone. When seismic velocities suddenly decrease, such as at the mantle–outer core boundary, waves get bent so that there is a shadow zone where no direct waves arrive. When seismic waves suddenly increase, as they do at the outer core–inner core boundary, waves get bent so that several P waves can sometimes arrive

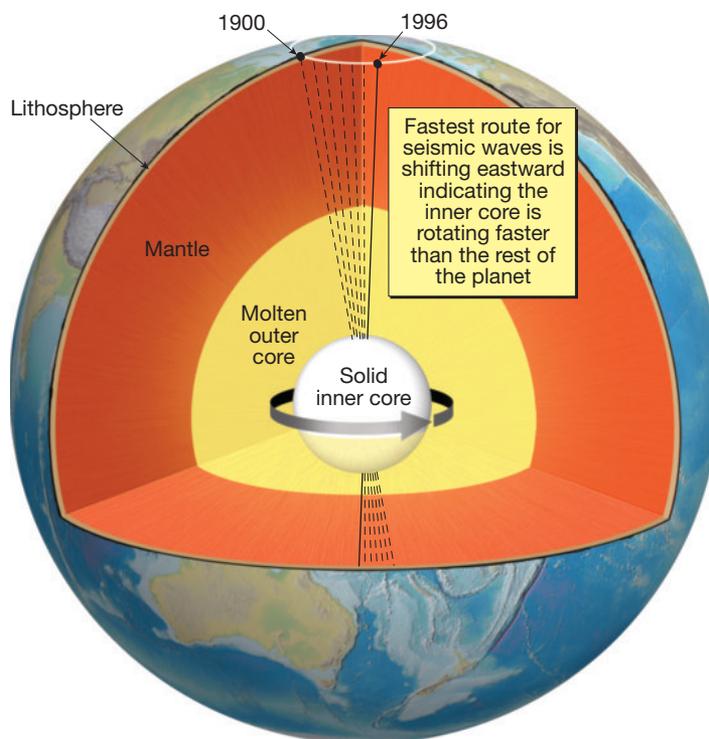


FIGURE 12.11 The solid inner core is separated from the mantle by the liquid outer core, and moves independently. Slight variations in the travel times of seismic waves through the core, measured over many decades, suggest that the inner core actually rotates faster than the mantle. The reason for this is not yet understood.

at a single location. In the case of the inner core, these waves can even be refracted enough to arrive within the P-wave shadow zone. Both of these occurrences, shown in Figure 12.10A, are proof of a distinct inner core.

Earth's Temperature

One way to describe a planet is by the composition of its layers, as was done in the preceding discussions. Another way is to examine the change in temperature with depth. This is very important for understanding the movements of rock within a planet. As you are probably aware, heat flows from hotter regions toward colder regions. Earth is about 5500°C at its center and 0°C at its surface, so heat is continually flowing toward the surface. It is this flow of heat that produces the convective flow of rock and metal in the mantle and core, and, in the process, plate tectonics.

We can measure the rate at which Earth is cooling by measuring the rate at which heat is escaping at Earth's surface. Measurements around the planet have shown that the average flow of heat at the surface is 87 milliwatts per square meter. This is not a lot, as it would take the energy emitted from about 690 square meters, roughly the size of a baseball diamond, to power one 60-watt light bulb. However, because Earth's surface is so large, heat leaves at a rate of 44 terrawatts per year, which is about three times the total world rate of energy consumption.

As Figure 12.12 shows, heat does not leave Earth's surface at the same rate in all locations. The rate of heat flow is highest near mid-ocean ridges, where hot magma is consistently rising toward the surface. Energy flow is also high in many continental regions because of particularly high levels of radioactive isotopes there. Heat flow is lowest in the areas of old, cold, ocean seafloor abyssal plains.

How Did Earth Get So Hot?

Earth, like all planets in our solar system, has had two thermal stages of existence. The first stage occurred during Earth's formation and involved a very rapid increase in internal temperature. The second stage has been the very slow process of cooling down. The first stage was very brief, taking only about 50 million years. The second stage has taken the remaining 4.5 billion years of Earth history and will continue for about another 4.5 billion years, until the Sun becomes a red giant star and Earth is destroyed.

As discussed in Chapter 1, Earth formed through a very violent process involving the collisions of countless planetesimals ("baby planets") during the birth of our solar system. With each collision, the kinetic energy of motion was converted into heat. As the early Earth grew in size, it rapidly got hotter. Several factors contributed to the early increase in temperature. The planet contained many relatively short-lived radioactive isotopes, such as Aluminum-26 and Calcium-41. As these isotopes decayed to stable isotopes, they released a great deal of energy. Further, as Earth's mass increased, so did its gravitational force of attraction, and this force caused the entire Earth to be compressed. This compression led to an increase in Earth's temperature, much the

same way that compressing air in a bicycle pump causes the whole pump to get hot.

Two other events caused Earth's temperature to rise suddenly. The first was the collapse of the iron core. At some point during Earth's growth the temperature got high enough that iron started to melt. Droplets of liquid iron began to form and sink toward Earth's center to form the core. The sinking of these iron droplets released additional heat, which caused more iron to melt and sink, which released more heat and so on. The core likely formed quickly by this runaway process.

The second significant event that heated our planet was the collision of a Mars-sized object with Earth that led to the formation of the Moon. At this time the entire core was molten, and most, if not all, of the mantle as well. From that point, about 4.5 billion years ago, to the present, Earth has slowly and steadily cooled down.

If Earth's only source of heat were from its early formation, our planet would have cooled to a frozen cinder billions of years ago. However, Earth's mantle and crust contain enough long-lived radioactive isotopes to keep Earth warm. There are four main radioactive isotopes that keep our planet cooking as if on a slow burner: uranium-235, uranium-238, thorium-232 and potassium-40. As was shown in Table 9.1 (page 262), the half-lives of these four isotopes are on the order of billions of years. As a result, large quantities of these isotopes remain. Radioactivity therefore plays two vital roles in geology. It provides the means for determining the ages of rocks, as discussed in Chapter 9. Even more importantly, however, it has kept mantle convection and plate tectonics active for billions of years.

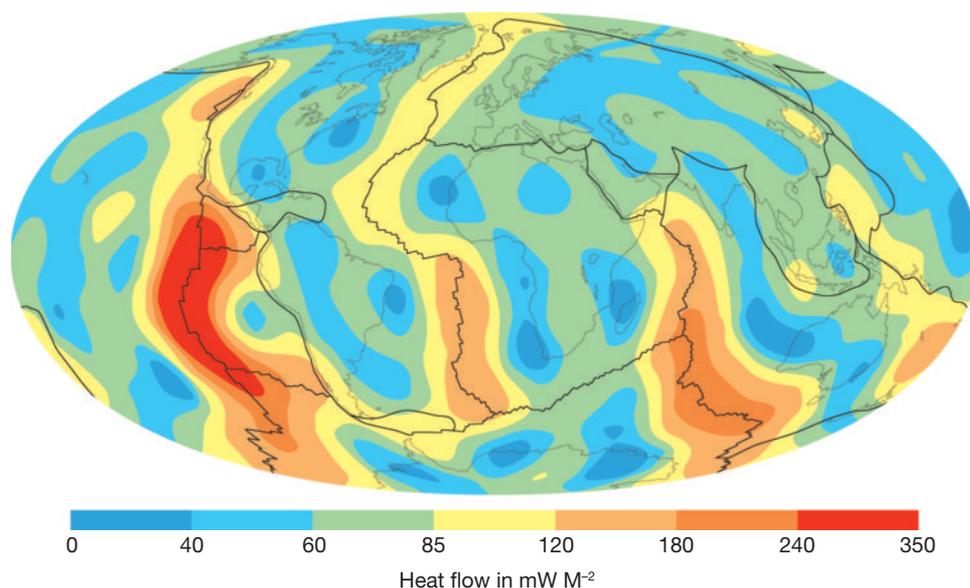


FIGURE 12.12 A map of the rate of heat flow out of Earth as it gradually cools over time, measured in milliwatts per square meter. Earth loses most of its heat near mid-ocean ridges, where magma rises toward the surface to fill the cracks formed when tectonic plates pull apart. Continents lose heat faster than old ocean seafloor because they contain higher amounts of heat-producing radioactive isotopes.

Heat Flow

Heat travels by three different mechanisms: *radiation*, *conduction*, and *convection*. Within a planet all three are active, but are more or less significant within different layers. The motions of rock and metal in the interior of a planet are entirely dependent upon how heat is able to move from one layer to the next. The regions where radiation, convection, and conduction are important in controlling the flow of heat out of Earth are shown in Figure 12.13. As you can see, only two of these processes (*convection* and *conduction*) operate within Earth's interior, and these will be considered next.

Convection The transfer of heat by moving material in a fluid-like manner and carrying the heat with it is called **convection**. It is the primary means by which heat is transferred within Earth. You are familiar with convection if you

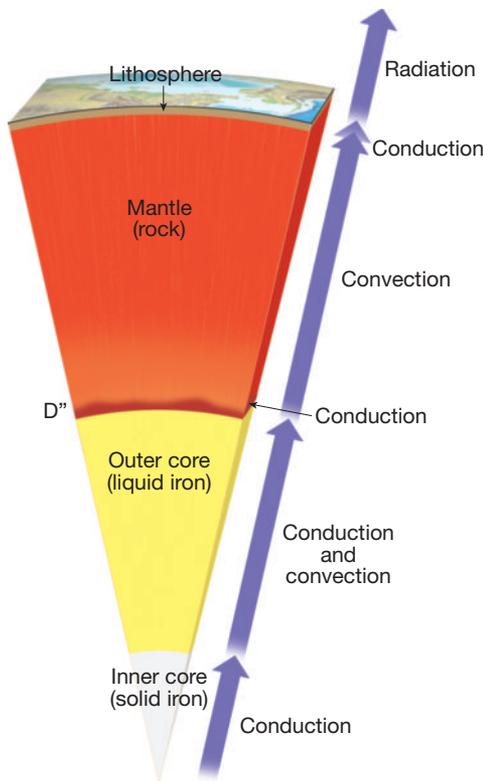


FIGURE 12.13 Diagram showing the dominant style of heat transfer at different depths within Earth as the planet cools down. Earth ultimately loses its heat to space through radiation. However, heat travels from Earth's interior to the surface, through the processes of heat convection and conduction.

have ever watched a pot of boiling water. The water seems to be rolling—rising up in the middle of the pot, and then down along the sides (Figure 12.14). This pattern is called a *convection cycle*, and occurs within Earth's mantle and outer core, and possibly within the inner core as well.

Convection occurs because of several factors—thermal expansion, gravity, and fluidity. When the water at the bottom of the pot is heated, it expands. The colder and heavier water at the top of the pot sinks and replaces the hot water at the bottom, which then rises to the top. The driving force for convection is the force of gravity, pulling down on the water. If you tried to boil water while floating in outer space, with no strong gravity present, you would find that your pot of boiling water would not convect.

Last, the material has to be fluid enough to be able to flow. Scientists usually measure a material's fluidity in terms of its *resistance* to flow, called its **viscosity**. Water flows easily and has a low viscosity. The liquid iron of Earth's outer core likely has a viscosity close to water, and it also convects very easily. Materials that are very viscous do not flow easily but can still convect. Catsup is 50,000 times more viscous than water, but it still flows. Rock in the lower mantle is 10 trillion trillion (10^{25}) times more viscous than water, but it too flows.

The temperatures at the top and bottom of a convection cycle determine how vigorous the convection is. Earth's surface is very cold, compared to the interior, so newly formed

ocean lithosphere cools rapidly. This causes the ocean lithosphere to contract and become denser and heavier, and in time it sinks back into the mantle at subduction zones. These cold sinking slabs eventually descend to the base of the mantle absorbing heat along the way. When rock in the lower mantle becomes warm enough, it rises back toward the surface, some of it eventually making its way to mid-ocean ridges to become new ocean lithosphere (Figure 12.15). Oceanic lithosphere can therefore be thought of as the top part of the mantle convection cycle. In a similar manner, plate tectonics can be viewed as the surface expression of mantle convection, which is Earth's primary mechanism for cooling down.

Convection can sometimes occur in a way that is not driven by heat flow. This is called *chemical convection*, and it occurs when changes in density result through chemical and not thermal means. Chemical convection is an important mechanism within the outer core. As iron crystallizes and sinks to form the solid inner core, it leaves behind a melt that contains a higher percentage of lighter elements. Because this liquid is more buoyant than the surrounding material it rises upward, creating convection.

Conduction The flow of heat through a material is called **conduction**. Heat conducts in two ways: (1) through the collisions of atoms and (2) through the flow of electrons. In rocks, atoms are locked in place but are constantly oscillating. If one side of a rock is heated up, its atoms will oscillate more energetically. This will increase the intensity of the collisions with their neighboring atoms, and like a domino ef-



FIGURE 12.14 A simple example of convection, which is heat transfer that involves the actual movement of a substance. Here the flame warms the water in the bottom of the beaker. This heated water expands, becomes less dense (more buoyant), and rises. Simultaneously, the cooler, denser water near the top sinks.

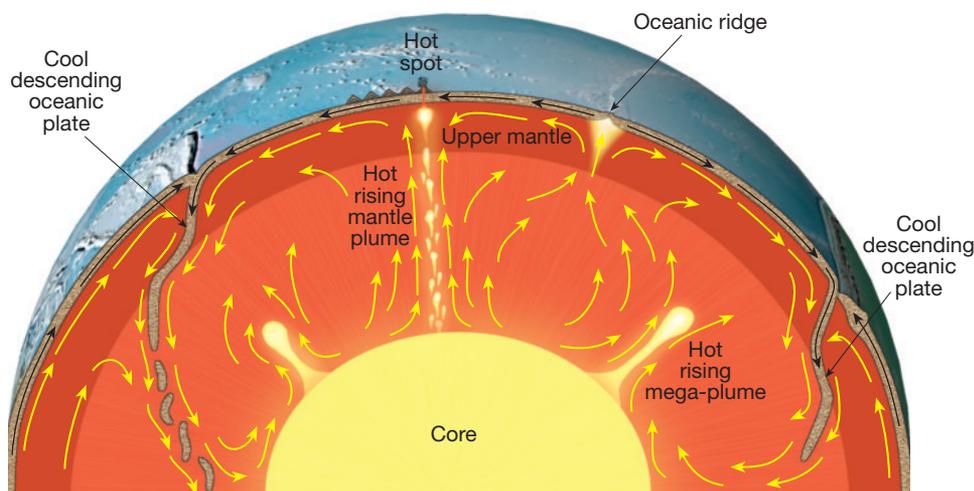


FIGURE 12.15 Diagram showing convection within Earth's mantle. The entire mantle is in motion, driven by the sinking of cold oceanic lithosphere back into the deep mantle. This is like stirring a pot of stew with downward strokes of a spoon. The upward flow of rock likely occurs through a combination of mantle plumes and a broad return flow of rock to replace the ocean lithosphere that leaves the surface at subduction zones.

fect, the energy will slowly propagate all the way through the rock. Conduction occurs much more quickly in metals. Though the atoms of metals are also locked in place, some of their electrons are free to move through the material, and these electrons can carry heat quickly from one side of a metal object to another.

There is a huge variation in the rate of conduction for different materials. For example, heat conducts about 40,000 times more easily through a diamond than through air. Most rocks are poor conductors of heat. Conduction is therefore not an efficient way to move heat through most of Earth. However, there are places where conduction is important; these include the lithosphere, the D'' layer, and the core.

Conduction, through the flow of electrons, is likely very important in both the solid iron inner core and liquid iron outer core. Once heat conducts from the inner core into the outer core, convection may play a significant role in carrying heat to the top of the core. However, heat can only pass from the core to the mantle through conduction, not convection. This is because iron is much too dense to intrude into the lighter mantle floating on top. For heat to leave the core, it must conduct across the core–mantle boundary and thus through the D'' layer. Once the heat reaches the lower mantle it is carried toward the surface through mantle convection.

Next, heat from the mantle moves to Earth's surface through the rocky lithosphere. There are a few places where convection carries heat directly to the surface, in the form of volcanic lava. Everywhere else the heat must make its final journey to the surface by conducting slowly across the stiff, rigid lithosphere.

Earth's Temperature Profile

The profile of Earth's average temperature at each depth is called the **geothermal gradient** or **geotherm**, (Figure 12.16A). Earth's temperature increases from about 0°C at the

surface to more than 5000°C at Earth's center. Within Earth's crust, temperature increases rapidly—as great as 30°C per kilometer of depth. You can experience this in deep mines. The deepest diamond mines in South Africa go to depths of more than 3 kilometers, where the temperature is more than 50°C (120°F). The temperature doesn't continue to increase at such a rapid rate, however, or the whole planet would be molten below a depth of 100 kilometers.

At the base of the lithosphere, about 100 kilometers down, the temperature is roughly 1400°C. However, you would need to go to almost the bottom of the mantle before the temperature doubled to 2800°C. For most of the mantle, the temperature increases very slowly—about 0.3°C per kilometer. However, the D'' layer acts as a thermal

boundary layer, and the temperature there increases by more than 1000°C from the top to the bottom. Finally, temperatures increase only gradually across the outer and inner cores.

Determining temperatures inside Earth is difficult, and there are large uncertainties. In fact, the temperature at Earth's center may be as high as 8000°C. You may be wondering how geoscientists measure Earth's deep temperatures. The best way is to use mineral physics experiments that measure the temperatures and pressures at which materials change. For example, the basis for the geotherm for the upper mantle (shown in Figure 12.16A) comes from experiments that establish the temperatures at which the mineral olivine makes the phase changes that cause the 410- and 660-kilometer discontinuities. Similar experiments are used to determine the temperature at which liquid and solid iron would coexist at the boundary between the inner and outer core.

Also plotted in Figure 12.16A is the curve for the average melting point of material at each depth. How close the geotherm is to the melting point of a material not only determines whether a material is molten or not, but how stiff it is. Figure 12.16B shows the viscosity of material in the crust and mantle. High-viscosity regions, like the lithosphere, are very stiff. Low-viscosity regions, like the asthenosphere or D'', are much softer. Notice how viscosity is directly related to how close the geotherm and melting point curves are in Figure 12.16A. When rock gets close to its melting point, it begins to weaken and get soft.

Both the geotherm and melting point curves generally increase gradually with depth. This is a result of the continual increase in pressure. Squeezing a material raises its temperature by causing the atoms to collide with each other more often, so the geotherm increases. This is the reason for the gradual increase in temperature seen across the middle of the mantle and across the core. However, squeezing a material

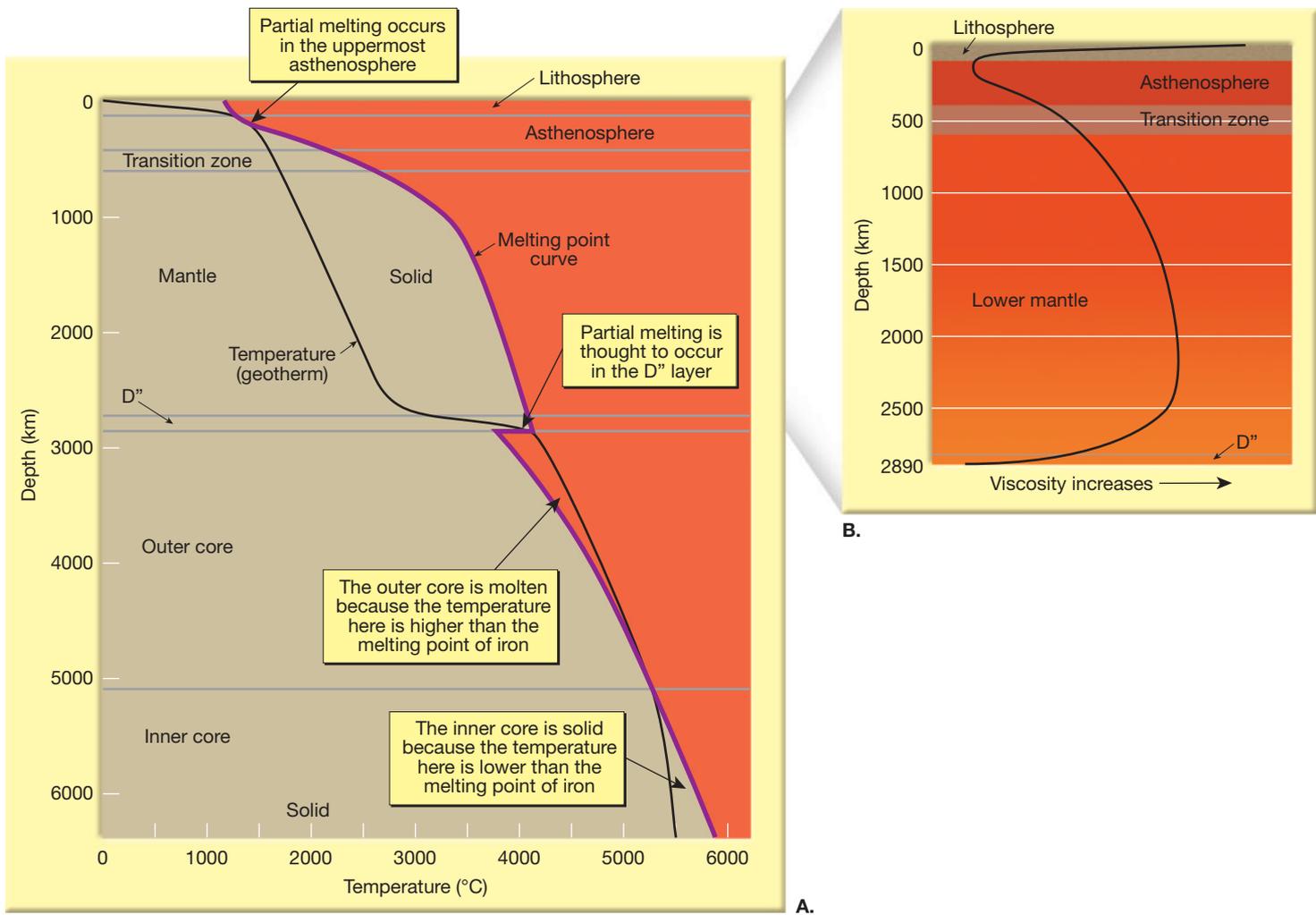


FIGURE 12.16 These graphs show how the viscosity of Earth materials at various depths is related to Earth's geotherm and the melting point of these materials. **A.** Earth's temperature profile with depth, or *geotherm*. Note that Earth's temperature increases gradually in most places. Within Earth's two major thermal boundary layers, the lithosphere and the D'' layer at the core-mantle boundary, the temperature increases rapidly over short distances. Also shown is the melting point curve for the materials (rock or metal) found at various depths. Where the geotherm crosses above (to the right of) the melting point curve, as in the outer core, the material is molten. **B.** This graph shows how viscosity (resistance to flow) changes with depth from Earth's surface to the bottom of the mantle. High viscosities, as in the crust and lithosphere, show rock that is stiffer and flows less easily. If you compare these two figures, you can see that rocks are weakest and flow more easily at depths where the temperature of rocks are close to melting (the asthenosphere and D'' layer).

also makes it harder to melt because liquids usually take up more volume than solids. Higher pressures result in less room for rock to expand into, so materials under pressure tend towards being solid. This causes the melting point curve to also increase with depth. Generally the melting point curve increases more rapidly with depth than the geotherm. However, in two layers, the uppermost asthenosphere and D'' layer, Earth's temperature is high enough that some rock begins to melt.

We can now understand why different layers of Earth behave the way they do. The lithosphere is stiff because its temperature is much colder than its melting temperature. The asthenosphere is weaker and softer because it is very close to its melting temperature, and partial melting likely occurs in some places. The existence of the weak asthenosphere

is critical to the existence of plate tectonics on Earth—it allows the stiff sheets of lithosphere to slide across it. Without an asthenosphere, Earth's mantle would still convect, but it would not have tectonic plates.

Most of the lower mantle is very stiff, and rock moves more sluggishly there. It is thought that convective flow occurs several times slower in the lower mantle than in the upper mantle. However, at the very base of the mantle, where Earth's temperature again approaches the melting point of rock, the D'' layer is relatively weak, and rock there flows more easily.

In the core, the temperature increases much more slowly than does the pressure. From the core-mantle boundary to Earth's center, the temperature may only increase by about 40 percent, or from 4000° to 5500°C. However, over the same

depth the pressure nearly triples, going from 1.36 to 3.64 megabars. Even though iron in the outer core is cooler than iron in the inner core it is under much less pressure and remains a liquid. Stated another way, iron in the inner core, although very hot, is under such great pressure that the inner core is solid.

Earth's Three-Dimensional Structure

As you have seen, Earth is not perfectly layered. At the surface there are large horizontal differences: oceans, continents, mountains, valleys, trenches, mid-ocean ridges, and so on. Geophysical observations show that horizontal variations are not limited to the surface—they also occur within Earth, and are directly related to the process of mantle convection and plate tectonics. Three-dimensional structure within Earth is identified primarily with a kind of seismic imaging called seismic tomography. It is also studied by examining variations in Earth's gravitational and magnetic fields.

Earth's Gravity

The most significant cause for changes in the force of gravity, at the surface, is due to Earth's rotation. Because Earth rotates around its axis, once every day, the acceleration due to gravity* is less at the equator (9.78 m/s^2) than at the poles (9.83 m/s^2). This happens for two reasons. Earth's rotation causes a centrifugal force that is in proportion to the distance away from the axis of rotation. (This is similar to the force that appears to try to throw you off of a moving merry-go-round.) Centrifugal force acts to throw objects upward at the equator, where the force is greatest.

In addition, Earth's rotation has caused Earth's shape to be slightly flattened, with the equator further from Earth's center (6378 km) than the poles (6357 km). This weakens gravity at the equator because gravitational force is smaller when objects are further apart. Your weight will actually be less by 0.5% at the equator than at the poles (Figure 12.17). Earth is therefore not a perfect sphere, but has the shape of an *oblate ellipsoid*. The amount of flattening is 1 part in 298. This shape was one of the first clues geologists had that Earth's mantle, although essentially solid, behaves like a fluid and is able to flow over very long time scales.

Gravity measurements show that there are more variations in Earth's shape than just its slightly elliptical nature. The density of rock within Earth is different in different locations. This is obvious for the crust, where geologists find rocks with different compositions and densities at different places at the surface. For instance, the igneous rocks that make up extensive lava flows in the northwestern United States are denser than the sedimentary rocks that outcrop in the Midwest. Density differences for rocks of different compositions extend throughout the crust and also into the

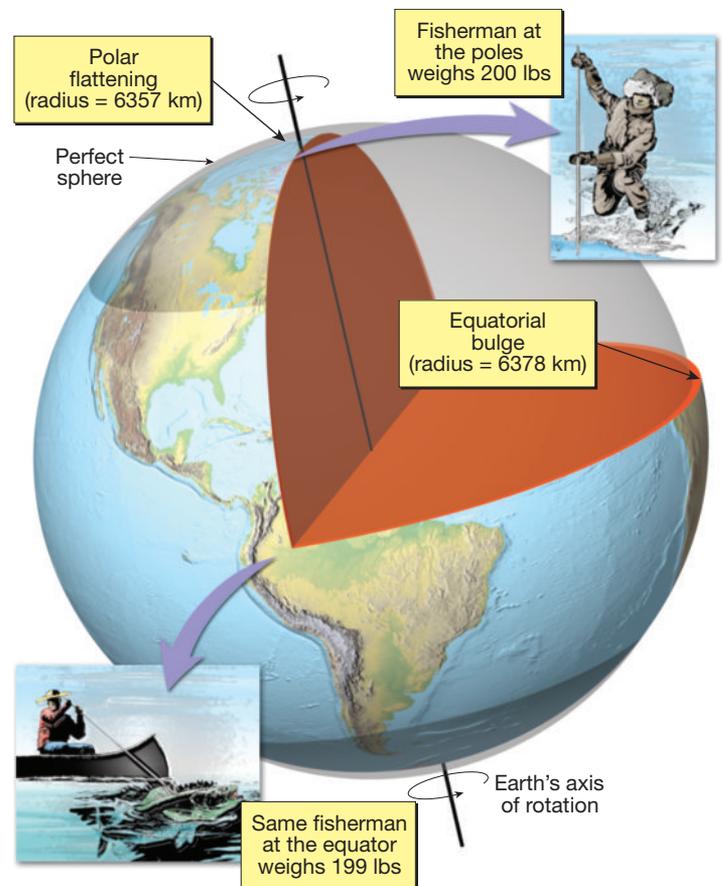


FIGURE 12.17 Drawing of Earth demonstrating the bulge at the equator and flattening of the poles that occur because of Earth's rotation. The combination of Earth's elliptical shape and its daily rotation actually cause the force of gravity to be weaker at the equator than at the poles. This difference is large enough to be measured on a bathroom scale. Imagine two fishermen of equal mass both standing at sea level. If the one at the North Pole weighs 200 pounds, the one at the equator would weigh only 199 pounds.

mantle. If there is denser rock underground, the resulting increased mass will cause a greater gravitational force. Because metals and metal ores tend to be much denser than silicate rocks, gravity anomalies (differences from the expected) have long been used to help prospect for mineral deposits. A map of gravity anomalies for the United States is shown in Figure 12.18. The majority of the variations shown are caused by density differences that result from changes in composition. Another significant cause of gravity variations at the surface is due to topography on land and bathymetry (the topography of the ocean floor). For example, over the oceans, where there are seamounts, water is being replaced by rock, and this increases the gravitational pull in that area. These gravitational anomalies change the level of the sea surface. Contrary to what you might think, the sea surface is elevated above seamounts, ridges, and underwater plateaus, not depressed. Though the downward force of gravity acting on the overlying water is increased, this increase in gravity also causes the surrounding water to be pulled toward these elevated features, causing the sea surface to rise. In fact, the topography of the seafloor is actually

*The force of gravity causes objects, such as an apple to accelerate as it falls to the ground, hence the expression "acceleration due to gravity."

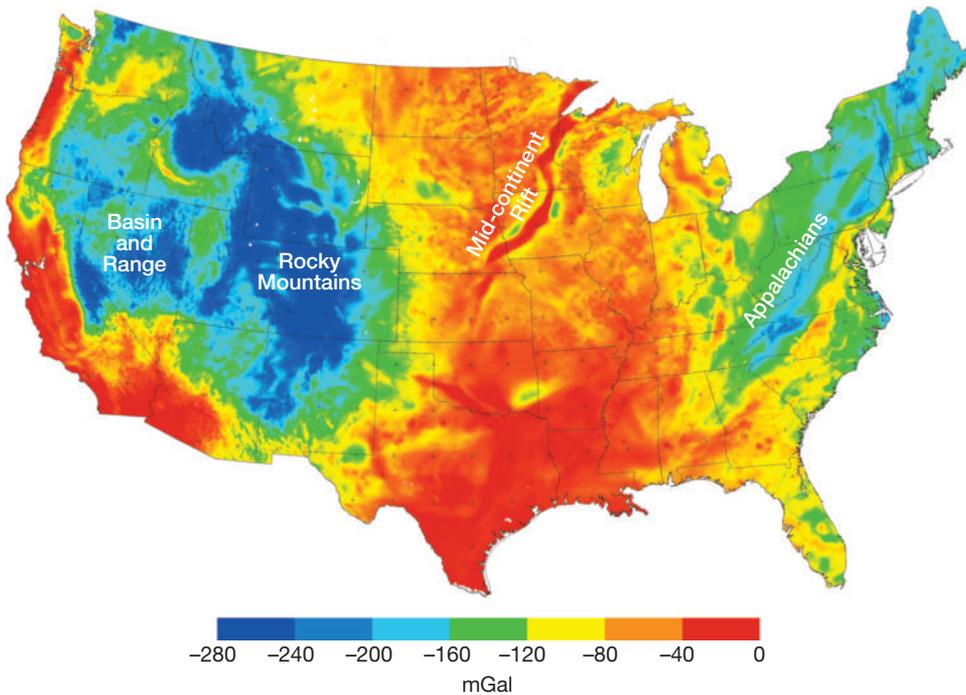


FIGURE 12.18 A map of gravity anomalies beneath the continental United States. Changing elevation changes the strength of Earth's gravity, so values are calculated for what would be measured if you were at sea level at each location. This allows the gravity anomalies to be compared across the map. The negative anomalies (blue) beneath the Rockies and Appalachians show us that the crust has deep roots beneath the mountains there. The negative anomaly (blue) in the Basin and Range Province is the result of hotter, tectonically active crust (rifting and volcanoes). The narrow positive anomaly (red) that runs in a line down the middle of the country is the mid-continent rift, where denser volcanic rocks entered the crust more than a billion years ago.

measured globally with satellites that use radar to measure the elevation of the sea surface (see Figure 13.5).

Changes in density deep beneath the surface also cause variations in the shape of Earth's surface. The height of the surface of the oceans actually changes vertically by about 200 meters due to very large-scale density variations within the mantle. The shape of this surface, measured from the perfect ellipsoid due to rotation, is called the **geoid**. A map of global geoid variations is shown in Figure 12.19. The width of the geoid anomalies can be an indication of the depth of the density anomalies that cause them. When underground density anomalies are near the surface, the geoid variations are narrow. When density anomalies are very deep, the geoid anomalies are very broad, sometimes thousands of kilometers across. These large-scale geoid anomalies are a result of the large upwellings and downwellings of mantle convection.

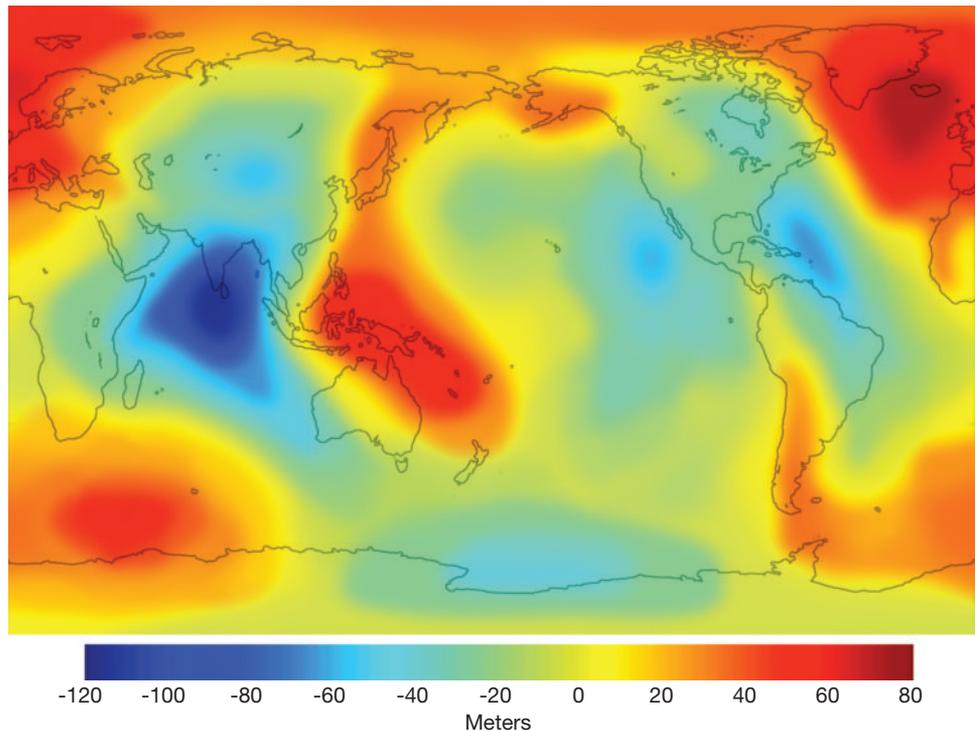


FIGURE 12.19 Map of the large-scale geoid, made from satellite measurements. The geoid is the shape of Earth that differs from what is expected from just Earth's rotation. The geoid is a result of broad differences in density within Earth's interior. The effect is that the sea surface varies in height by more than 200 meters due to movements within Earth's deep interior.

Seismic Tomography

The three-dimensional changes in composition and density that are detected with gravity measurements can actually be viewed using seismology. In a technique called **seismic tomography**, a very large number of seismic observations are combined to make three-dimensional models of Earth's interior. These models typically involve collecting signals from many different earthquakes recorded at many seismograph stations, in order to "see" all parts of Earth. Seismic tomography is very similar to medical tomography, in which doctors use techniques like CT scans to make three-dimensional images of a person's body.

Seismic tomography usually involves identifying regions where P or S waves travel faster or slower than average for that depth. These seismic velocity "anomalies" are then interpreted as variations in material properties such as temperature, composition, mineral phase, or water content. For instance, increasing the temperature of rock about 100°C can decrease S velocities

by about 1 percent, so images from seismic tomography are often interpreted in terms of temperature variations.

Figure 12.20 shows an example of S-wave velocity tomography for the mantle beneath North America. *Red colors* show regions where waves travel slower than average and *blue colors* show regions where waves travel faster than average. This colored diagram shows some very significant patterns. Continental lithosphere has fast seismic velocities, when compared to oceanic lithosphere, because it is older and thus has been cooling at the surface for a long time. Seismic imaging also shows that continental lithosphere (deep blue areas) can be very deep, extending more than 300 kilometers into the mantle. This deep continental lithosphere is called the **tectosphere**. Beneath some of the oldest portions of continents the tectosphere connects continuously with the lower mantle, without a strong presence of an asthenosphere. The opposite situation occurs at oceanic ridges, which exhibit slow seismic velocities (bright red areas) because they are very hot (Figure 12.20).

In the mid-mantle beneath North America, you can see a tongue of fast seismic velocities (light blue) that represents a sheet of ancient Pacific Ocean lithosphere known as the Farallon plate. This plate used to subduct beneath North America all along its western edge. Remnants of the plate still subduct beneath Oregon and Washington in the form of the Juan de Fuca plate and beneath Mexico as the Cocos plate. The segment of this former ocean sea floor seen in Figure 12.20 used to descend beneath California, but is now completely subducted. The Farallon slab is now sinking down through the lower mantle toward the core–mantle bound-

ary, where it is slowly heating up and will eventually become mixed back into the mantle. Over time, this slab will become hot enough to begin to rise back to the surface. This kind of upward return flow may be what is observed as the reddish orange areas at the right and left sides of the figure.

The large region of slow seismic velocities at the base of the mantle beneath Africa (the large reddish orange region at the lower right of Figure 12.20) is called the African superplume—a region of upward flow in the mantle. These slow velocities are likely due to both unusually high temperatures and rock that is highly enriched in iron. The rising rock cannot easily break through the African tectosphere, so it seems to be deflected to both sides of Africa, perhaps supplying new magma to both the Mid-Atlantic and Indian Ocean spreading centers.

Images from seismic tomography, like the one shown in Figure 12.20, reveal the whole-mantle cycle of convection. Sheets of cold ancient ocean sea floor sink to the base of the mantle, where they warm, expand, and rise back toward the surface again.

Earth's Magnetic Field

Convection of liquid iron in the outer core is vigorous and gives rise to Earth's magnetic field. Because material in the outer core flows so easily, horizontal temperature variations there are very small—likely less than 1°C. Such small temperature differences create indistinguishable differences in seismic velocities, so the outer core appears uniform at each depth when viewed with seismic waves. However, the patterns of flow in the outer core create variations in Earth's magnetic field, and these can be observed at Earth's surface.

Flow in the outer core is thought to occur for three main reasons:

1. As heat conducts out of the core into the surrounding mantle, the outermost core fluid cools, becomes denser, and sinks. This is a form of thermally driven convection.
2. Crystallization of solid iron at the bottom of the outer core, to form the inner core, releases fluid that is depleted in iron and therefore relatively buoyant. As this fluid rises up and away from the inner core boundary it helps drive convection. This is a form of chemically driven convection.
3. There may be radioactive isotopes like potassium-40 within the core that could provide additional heat to drive thermal convection.

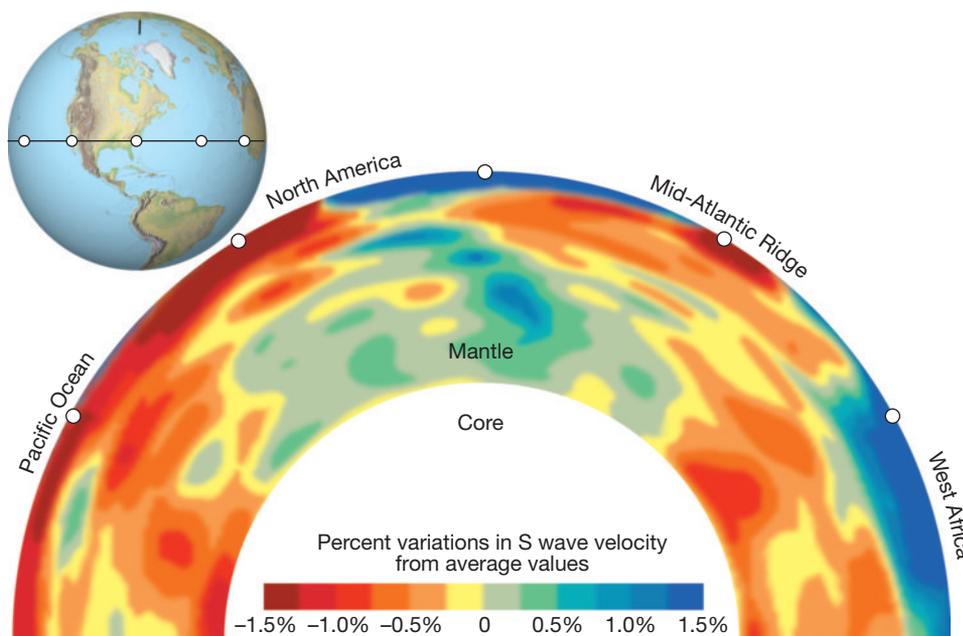


FIGURE 12.20 A seismic tomographic slice through Earth showing mantle structure. Colors show variations in the speed of S waves from their average values. Older portions of continents such as eastern North America and Africa are cold and stiff, so their blue colors show fast S wave speeds. The western United States is hotter and tectonically active, making that portion of the continent warmer and weaker, which slows S waves. The large blue structure extending far below North America is a sheet of cold, dense ancient Pacific seafloor that is sinking toward the base of the mantle. The large orange structures beneath western Africa and the Pacific Ocean are thought to be megaplumes of warm material that are rising toward the surface.

The relative importance of these three mechanisms is still uncertain.

The Geodynamo As the core fluid rises, its path becomes twisted through a phenomenon called the *Coriolis effect*, which is a result of Earth's rotation. The fluid ends up moving in spiraling columns as shown in Figure 12.21. Because the fluid is electrically charged, it generates a magnetic field through a process called a geodynamo that is similar to an electromagnet. If a wire is wrapped around an iron nail and an electric current passed through it, the nail will generate a magnetic field that looks a lot like the field from a bar magnet (Figure 12.22A, B). This is called a dipolar field—a type of magnetic field that has two poles (a north and south magnetic pole). As Figure 12.22C shows, the magnetic field that emanates from Earth's outer core has the same dipolar form.

However, the convection in the outer core is not quite so simple. More than 90% of Earth's magnetic field takes the form of a dipolar field, but the remainder of the field is the result of other more complicated patterns of convection in the core. In addition, some of the features of Earth's magnetic field change over time. For centuries sailors have used compasses to determine direction. Consequently, a great deal of attention has been paid to keeping track of the direction that compass needles point. One observed change in the magnetic field is a gradual "westward drift" of the non-dipole part of the magnetic field. In order to explain this, we first need to look at how the magnetic field is measured.

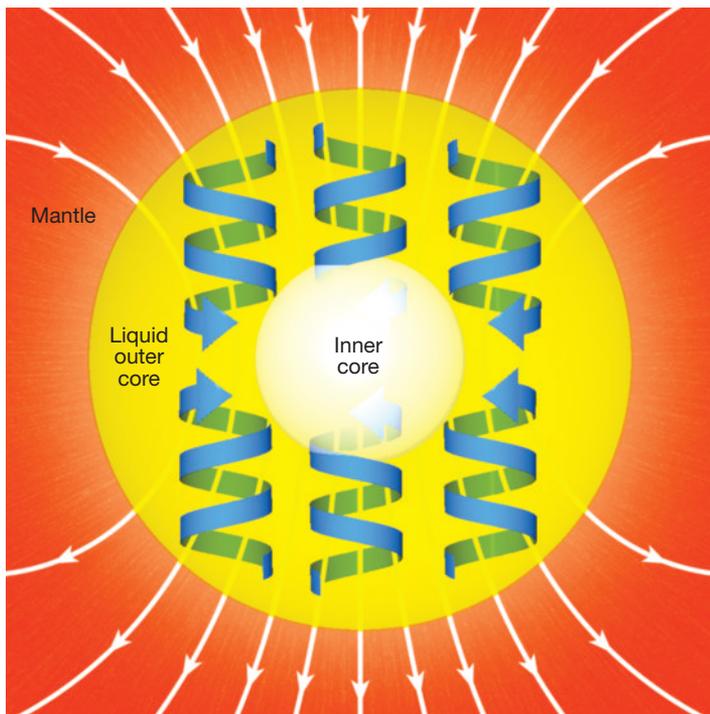
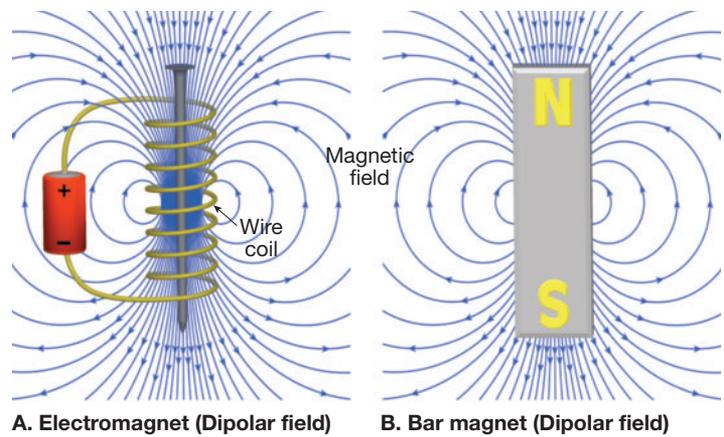
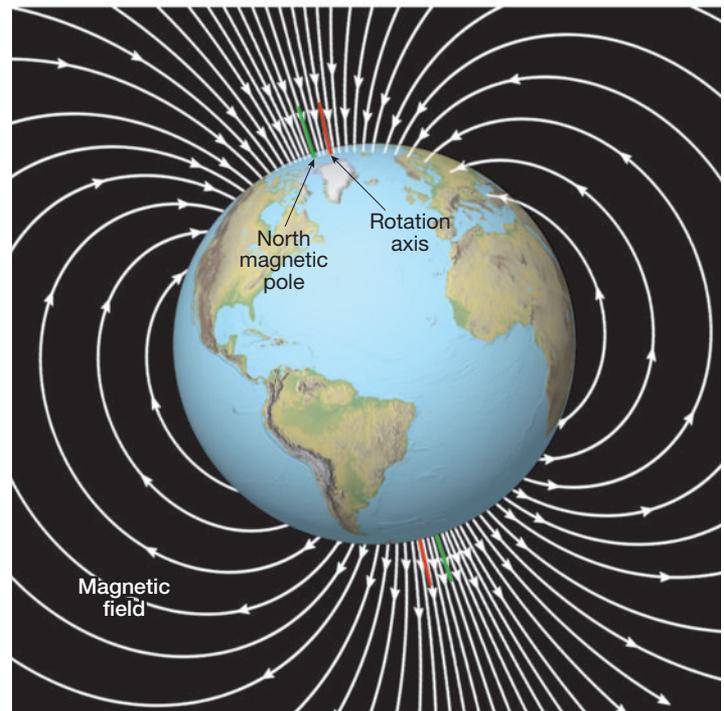


FIGURE 12.21 Illustration of the kind of convection patterns within Earth's liquid iron outer core that could give rise to the magnetic field we measure at the surface. It is thought that convection takes the form of cylindrical gyres of rotating molten iron that are aligned in the direction of Earth's axis of rotation.



A. Electromagnet (Dipolar field) **B. Bar magnet (Dipolar field)**



C. Earth's magnetic field (Dipolar field)

FIGURE 12.22 Demonstration of the similarity of Earth's magnetic field to that of an electromagnet (A), which consists of an electrical current passed through a coil of wire, or bar magnet (B). While it was once thought that Earth's core acts like a large bar magnet, scientists now think that Earth's magnetic field (C) is more like an electromagnet, and that the cylinders of spiraling liquid iron shown in Figure 12.21 behave like the coil of current passing through the wires of an electromagnet.

At any point on Earth's surface, the direction that the magnetic field is pointing is measured with two angles, called *declination* and *inclination*. The declination measures the direction to the magnetic north pole with respect to the direction to the geographic North Pole (Earth's axis of rotation). The inclination measures the downward tilt of the magnetic lines of force at any location. It is what your compass would read if you could tilt it on its side. At the magnetic north pole the field points directly downward. At the equator it is horizontal (Figure 12.23). In North America it tilts downward at an intermediate angle.

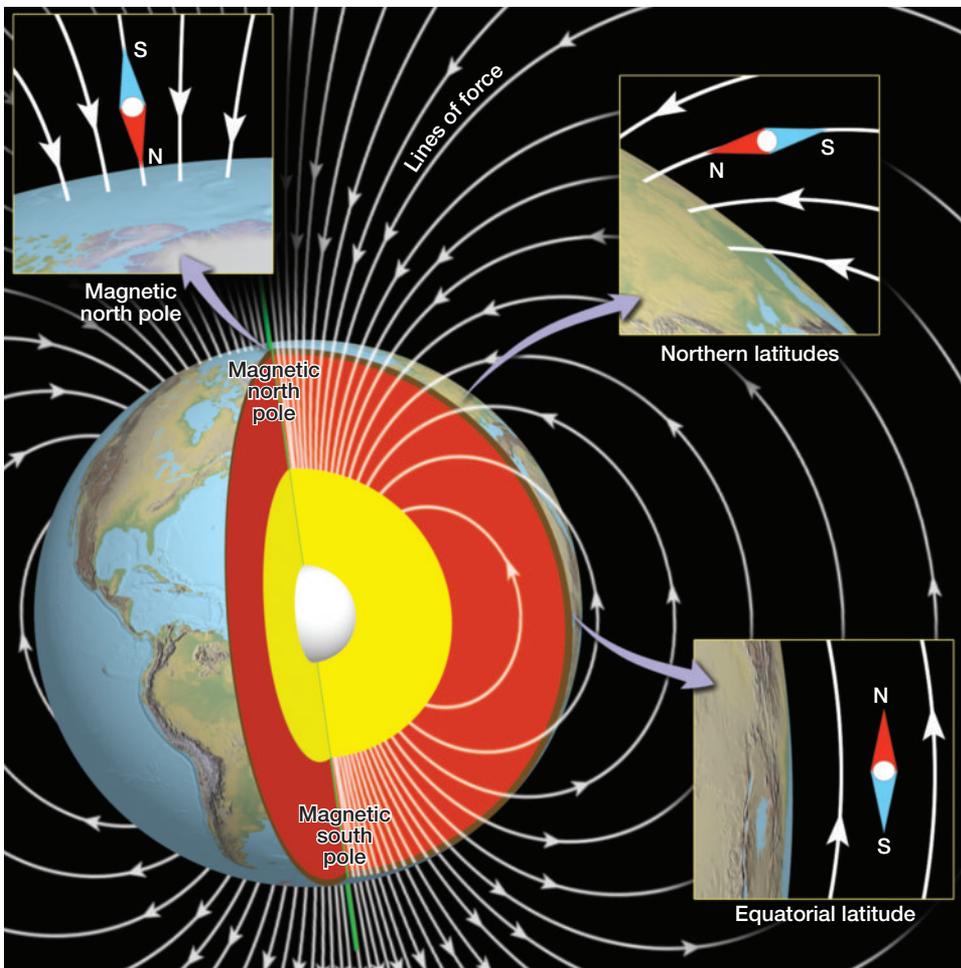


FIGURE 12.23 Drawing that shows the direction of the magnetic field at different locations along Earth's surface. Though a compass measures only the horizontal direction of the magnetic field (the declination), at most locations the field also dips in or out of the surface at a variable angle (inclination).



The location of the magnetic north pole actually moves significant distances during our lifetimes. Earth's magnetic north pole had been located in Canada, but over the past decade moved northward into the Arctic Ocean and is currently moving rapidly toward Siberia at a rate of about 20 kilometers per year (Figure 12.24). The process is not symmetric. Though the magnetic north pole has been moving towards the geographic North Pole, the magnetic south pole has been moving *away* from the South Pole, passing from Antarctica to the Pacific Ocean. This means that core convection changes significantly on a time scale of decades (see Box 12.2).

Magnetic Reversals Although core convection changes over time, causing the magnetic poles to move, the locations of the magnetic poles averaged over thousands of years are

FIGURE 12.24 Map showing the change in measured locations of the north magnetic pole over time. The patterns of convection within the outer core change fast enough that we can see the magnetic field change significantly over our lifetimes.

BOX 12.2 ▶ EARTH AS A SYSTEM

Global Dynamic Connections

The layers of planet Earth are not isolated from one another, but rather their motions are thermally connected. Furthermore, these connected motions don't always occur steadily—sometimes they occur episodically or in pulses. One example shows the possible connection between magnetic reversals, hot-spot volcanoes, and the breakup of the supercontinent of Pangaea.

Pangaea began to break up about 200 million years ago. Plate motions increased and there was a greater amount of subduction of ocean lithosphere. About 80 million

years later the core reversal process shut down and Earth's magnetic field did not reverse for 35 million years. In the tens of millions of years that followed, there were several enormous outpourings of lava that have been linked to the arrival of new hot-spot mantle plumes at the surface.

In one hypothesis, these three events are closely connected. The large amount of subducted lithosphere that followed the breakup of Pangaea could have plunged to the base of the mantle. This would have displaced hot rock at the base

of the mantle, causing much of it to rise to the surface and erupt as flood basalts such as the Deccan Traps in India. The sudden placement of cold ocean lithosphere next to the hot core at the core–mantle boundary would have chilled the uppermost core, causing more vigorous core convection that would have prevented the field from weakening and reversing. This hypothesis, if correct, is an important reminder that Earth is a complex, churning, pulsing planet that is very active in a variety of geological ways.

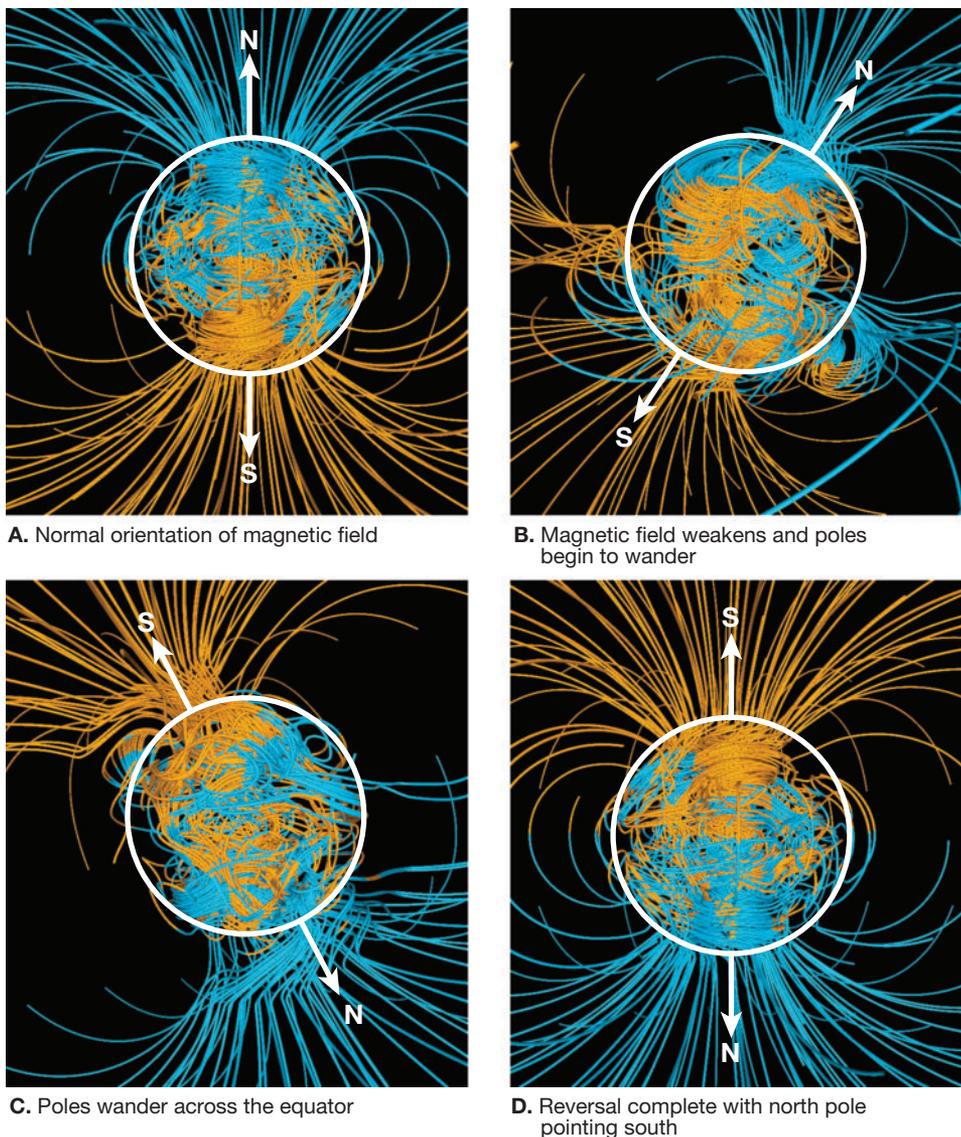


FIGURE 12.25 Computer simulations showing how Earth's magnetic field could reverse direction. The white circle represents the core–mantle boundary and the arrows point to the north (N) and south (S) magnetic poles respectively. During a reversal the strength of the magnetic field weakens and the poles begin to wander greatly, going so far as to cross the equator. When the strength of the field returns to normal levels, the field is regenerated with reverse polarity.

the same as Earth's axis of rotation (geographic poles). There is one major exception to this and that is during periods of magnetic field reversals. At apparently random times, Earth's magnetic field reverses polarity so that the *north* needle on your compass would point to the *south*. (The importance of these reversals in the study of paleomagnetism was described already in Chapter 2). What happens during a reversal is that the strength of the magnetic field decreases to about 10 percent of normal and the locations of the poles begin to wander greatly, going so far as to cross the equator (Figure 12.25). The strength of the magnetic field then returns to its normal levels, and the field is regenerated with reverse polarity. The whole process takes only a few thousand years.

The way that the magnetic field reverses is an indication of the way that the outer core's convection patterns change over relatively short spans. This complex process is now being modeled using high-speed computers as shown in Figure 12.25. The figure also shows how the magnetic field lines become twisted in a complex manner before returning to the more simple dipolar pattern that regularly exists.

The existence of magnetic reversals has been extremely important to geoscientists in providing the foundation for the theory of plate tectonics, but magnetic reversals could have harmful consequences for life on land. Earth's magnetic field creates a large magnetic

layer in space around the planet known as the *magnetosphere*. Along with the atmosphere, the magnetosphere protects Earth's surface from ionized particles emitted by the Sun. These ionized particles form what is called the *solar*

wind. If the strength of the magnetic field decreases greatly during a reversal, the increased amounts of solar wind reaching Earth's surface could cause health hazards for humans and other land-based life forms.

Summary

- Earth is layered with the densest materials at the center and lightest materials forming the outer layer. This layering is a result of gravity, and is similar for all planets. Earth's layers consist of the *inner core* (solid iron), *outer core* (liquid iron), *mantle* (dense rock), *crust* (low-density rock), *ocean* (water), and *atmosphere* (gas). Within layers, the density of materials increases with depth due to compression resulting from the increasing pressure. Within the mantle, increases in density also occur because of mineral phase changes.
- Because it is impossible to drill deep into Earth, *seismic waves* are used to probe Earth's interior. The patterns of seismic waves are complicated because their behavior is influenced by different structures inside the planet before returning to the surface. Seismic waves travel faster through cold rock and slower through hot rock. Seismic waves reflect off of layers composed of different materials. The results of seismic imaging of Earth's interior can be interpreted through comparison with mineral physics experiments. These experiments recreate the temperature and pressure conditions within Earth, and allow scientists to see what rocks and metals are like at various depths.
- *Oceanic crust* and *continental crust* are very different. Oceanic crust is created at mid-ocean ridges, and is fairly similar in composition and thickness everywhere. Continental crust is highly variable, has many different compositions, and is formed in many different ways. Oceanic crust is nowhere older than 200 million years, whereas continental crust can be older than 4 billion years. Oceanic crust is usually about 7 km thick, but continents can be thicker than 70 km. The boundary between the crust and mantle is called the *Moho*.
- The *mantle* comprises most (82%) of Earth's volume. The *upper mantle* extends from the Moho to a depth of 660 km, on average. The upper mantle contains part of the stiff lithosphere, the weak asthenosphere, and the transition zone that may contain significant amounts of water. The lower mantle extends from 660 km down to the core–mantle boundary, 2891 km beneath the surface. At the base of the lower mantle is the variable D'' layer.
- The core is mostly made of iron and nickel, although it contains about 15 percent lighter elements. Because iron is very dense, the core makes up one-third of Earth's mass, and iron is Earth's most abundant element (by mass). The solid inner core grows over time as Earth cools.
- Earth's temperature increases from about 0°C at the surface to about 5500°C at the center of the core (though the exact temperature is very hard to determine). Heat flows unevenly from Earth's interior, with most of the heat loss occurring along the oceanic ridge system. Earth became very hot early in its history (it may have become entirely molten), largely due to the impacts of planetesimals and heat released by radioactive decay (radiogenic heat). Since then, Earth has slowly cooled. Earth is still geologically active because of the radiogenic heat supplied by long-lived radioactive isotopes, including uranium-238, uranium-235, thorium-232, and potassium-40.
- Heat flows from Earth's hot interior to its surface, and does so primarily by convection and conduction. *Convection* transfers heat through the movement of material. *Conduction* transfers heat by collisions between atoms or the motions of electrons. Convection is very important within Earth's mantle and outer core, whereas conduction is most important in the inner core, lithosphere, and D''. Within the asthenosphere and the base of D'' the temperature is close enough to the melting point that some partial melting might occur and the rock is weak enough to flow more easily than elsewhere in the mantle. The weak asthenosphere is very important for plate tectonics because it allows the stiff plates (lithosphere) to move easily across the top of it.
- Rotation causes Earth's shape to take the form of an *oblate ellipsoid*, meaning its equator bulges slightly. The combination of Earth's rotation and its ellipsoidal shape cause gravity to vary significantly—from 9.78 m/s² at the equator to 9.83 m/s² at the poles. Gravity also varies around Earth's surface due to the presence of rocks of different densities. These density differences actually deform Earth's surface, including the ocean surface, by more than 200 meters. This surface is called the *geoid*.
- Three-dimensional images of structure variations within the mantle are made from large numbers of seismic waves using *seismic tomography*. These images show that continental lithosphere can extend several hundred kilometers into the mantle. They also show cold subducted oceanic lithosphere sinking to the base of the mantle and large superplumes of hot rock rising from the core–mantle boundary. This suggests that convection occurs throughout the mantle.
- Convection of the liquid iron in the outer core causes a magnetic *geodynamo* that is responsible for Earth's magnetic field. The convection takes the form of spiraling cylinders that are a result of the Coriolis effect. This field is primarily dipolar; that is, it resembles the field from a bar magnet or electromagnet. The patterns of convection in the outer core change rapidly enough that the magnetic field varies noticeably over our lifetimes.

- The magnetic field randomly reverses, with the north and south poles swapping positions. A reversal takes only a few thousand years, and involves a significant decrease in the strength of the dipolar field. This is important, because

the magnetic field creates a magnetosphere around Earth that protects our planet from much of the Sun's solar wind that would otherwise bombard it. If the magnetosphere weakens, life on land would be adversely affected.

Review Questions

1. What role does gravity play in the layering of planets?
2. What are the two major reasons for the increase in density with depth within Earth's mantle?
3. Why is seismology responsible for gasoline prices being much more affordable than they otherwise might be (without seismology)?
4. Explain three different ways that oceanic crust and continental crust are different. Where would you go to find very thick crust? Very thin crust?
5. What is the importance of determining the *cross-over* distance? (See Figure 12.7.)
6. How do S waves allow us to determine that the mantle is solid?
7. If there were a lot of water in Earth's mantle, in what layer would it most likely reside?
8. What mineral phase changes occur at the top and bottom of the transition zone?
9. What layer of Earth has the greatest volume?
10. How is the D'' layer similar to the lithosphere?
11. True or False: No seismic waves arrive in the *shadow zone*? Explain.
12. Why is Earth's core one-sixth of Earth's volume but one-third of its mass?
13. Why is Earth's inner core growing in size?
14. Why is heat flow from Earth's surface not evenly distributed?
15. What were the sources of heat that caused Earth to get so hot early in its history?
16. What prevents Earth from being a cold, motionless sphere of totally solid rock and metal?
17. Explain the difference between conduction and convection.
18. Why is convection a less efficient means of heat transfer in materials with high viscosity?
19. Why is conduction more important than convection within Earth's crust?
20. What happens to rock as the geotherm approaches the melting point temperature?
21. What happens to rock in regions where the geotherm crosses above the melting point curve?
22. Why would tectonic plates have a hard time moving if it were not for the existence of the asthenosphere?
23. Why is the lithosphere stiffer than the asthenosphere?
24. Earth once rotated much faster than it currently does. How would Earth's shape have been different in the past?
25. Would you expect to find a large layer of iron ore underground in a region with a positive or negative gravity anomaly? Explain.
26. Why does the mid-Atlantic ridge appear as a slow seismic velocity anomaly in Figure 12.20?
27. What are the three sources of convection in the outer core?
28. What happens during a magnetic reversal?
29. Why might a magnetic reversal be dangerous to humans?
30. If plate tectonics got suddenly much faster and a large volume of subducted oceanic lithosphere sank to the bottom of the mantle, do you think magnetic reversals would occur more or less frequently? Explain.

Key Terms

asthenosphere (p. 331)
 conduction (p. 336)
 convection (p. 335)
 core (p. 334)
 crust (p. 330)
 D'' layer (p. 332)

geothermal gradient,
 or geotherm (p. 337)
 geoid (p. 346)
 inner core (p. 334)
 lithosphere (p. 331)
 lower mantle (p. 332)

mantle (p. 331)
 Moho (p. 330)
 outer core (p. 334)
 seismic anisotropy (p. 331)
 seismic tomography
 (p. 340)

tectosphere (p. 341)
 transition zone (p. 332)
 viscosity (p. 336)

Web Resources



The *Earth* Website uses the resources and flexibility of the Internet to aid in your study of the topics in this chapter. Written and developed by geology instructors, this site will help improve your understanding of geology. Visit <http://www.prenhall.com/tarbuck> and click on the cover of *Earth 9e* to find:

- Online review quizzes.
- Critical thinking exercises.
- Links to chapter-specific Web resources.
- Internet-wide key-term searches.

<http://www.prenhall.com/tarbuck>

GEODe: Earth

GEODe: Earth makes studying faster and more effective by reinforcing key concepts using animation, video, narration, interactive exercises and practice quizzes. A copy is included with every copy of *Earth*.

Chapter 12 Earth's Interior Earth's Layered Structure

At Earth's center is the **inner core**, a solid, iron-rich sphere having a diameter of 2432 kilometers (1511 miles).

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Chapter 12 Earth's Interior Earth's Layered Structure

In most regions, the continental crust is about 35 kilometers (20 miles) thick, but it may exceed 70 kilometers (40 miles) in areas of prominent mountains.

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