18 Seismicity and Earth's Interior

In defiance of nature, and the rule of plate tectonics, the king of Antiochus boasted that his mausoleum and these monuments would be "unravaged by the outrages of time." A few decades later, the great statues of Nemrud Dagh, in eastern Turkey, were toppled by an earthquake. Little did this king know, but Turkey's northern borderlands are sliced through by a great strike-slip fault and sliding underneath its southern shores is a great plate of oceanic lithosphere. In fact, Turkey projects westward toward Europe exactly because it is being squeezed out of Asia by the ongoing collision of Africa and Arabia with the Eurasian plate. Earthquakes occur with tragic regularity in Turkey; its last great earthquake shook Istanbul and the rest of northern Turkey in 1999, leaving cities in rubble and killing 20,000 people. Landslides, tsunamis, groundshaking, and liquefaction all took their toll.

Earthquakes, perhaps more than any other phenomenon, demonstrate that Earth continues to be a dynamic planet, changing each day by internal, tectonic forces. Most earthquakes occur along plate boundaries. As the plates move, these boundaries—ocean ridges, continental rifts, subduction zones, and transform faults—are the sites of the most intense earthquake activity on Earth. Earthquakes occur during sudden movements along faults.

Every year, more than a million earthquakes are recorded by the worldwide network of seismic stations and are analyzed with the aid of computers such as those at the Earthquake Information Center in Golden, Colorado. With this network, the exact location, depth, and magnitude of all detectable earthquakes are plotted on regional maps. As a result, we can monitor the details of present plate motion. But that is not all. Seismic waves also provide our most effective probe of Earth's interior, and they constitute the main method of collecting data upon which we base our present concepts of Earth's internal structure.

Indeed, earthquakes are human disasters, as the power released by a single event is staggering. When the energy stored up in deforming rocks is suddenly released, the consequences may be devastating. Many large cities lie along major faults. Thus, it is imperative that we learn as much as we can about earthquakes so that their damage can be lessened.

MAJOR CONCEPTS

1. Seismic waves are vibrations in Earth caused by the rupture and sudden movement of rock.
2. Three types of seismic waves are produced by an earthquake shock: (a) P waves, (b) S waves, and (c) surface waves.
3. The primary effect of an earthquake is ground motion. Secondary effects include (a) landslides, (b) tsunamis, and (c) regional or local uplift or subsidence.
4. The exact location and timing of an earthquake cannot be predicted. However, seismic risk can be evaluated and, in areas with high risk, preparations for future earthquakes made.
5. Most earthquakes occur along plate boundaries. Divergent plate boundaries and transform fault boundaries produce shallow-focus earthquakes. Convergent plate boundaries produce an inclined zone of shallow-focus, intermediate-focus, and deep-focus earthquakes.
6. The velocities at which P waves and S waves travel through Earth indicate that Earth has a layered internal structure based on composition—crust, mantle, and core. It also has a solid inner core, a liquid outer core, a weak asthenosphere, and a rigid lithosphere.
7. Plate tectonics and upwelling and downwelling plumes are the most important manifestations of Earth's internal convection. The magnetic field is probably caused by convection of the molten iron core.

CHARACTERISTICS OF EARTHQUAKES

Earthquakes are vibrations of Earth, caused by the rupture and sudden movement of rocks that have been strained beyond their elastic limits. Three types of seismic waves are generated by an earthquake shock: (1) primary waves, (2) secondary waves, and (3) surface waves.

Elastic-Rebound Theory

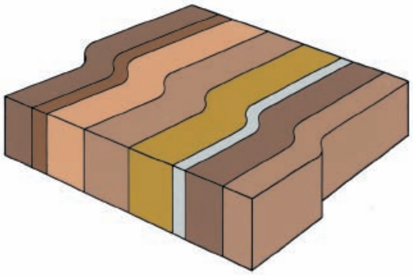
What causes earthquakes?

The origin of an **earthquake** can be illustrated by a simple experiment. Bend a stick until it snaps. Energy is stored in the elastic bending and is released if rupture occurs, causing the fractured ends to vibrate and send out sound waves. Detailed studies of active faults show that this model, known as the **elastic-rebound theory,** applies to all major earthquakes (Figure 18.1). Precision surveys across the San Andreas Fault in California show that railroads, fence lines, and streets are slowly deformed at first, as strain builds up, and are offset when movement occurs along the fault, releasing the elastic strain. The San Andreas fault is the boundary between the Pacific and North American plates. Its movement is horizontal, with the Pacific plate moving toward the northwest. On a long-term basis, the plates move quite steadily, at a rate of roughly 3 cm/yr. On a short-term however, much of the movement occurs in a series of jerks. Sections of the fault can be "locked" together until enough strain accumulates to exceed the rock's **elastic limit** and cause displacement.

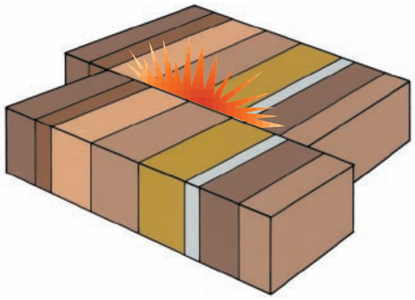
The point within Earth where the initial slippage generates earthquake energy is the **focus.** The point on Earth's surface directly above the focus is the **epicenter** (Figure 18.2).

Types of Seismic Waves

Several different types of **seismic waves** are generated by an earthquake shock (Figure 18.3). Each type travels at a different speed, and each therefore arrives at a **seismograph** that might be hundreds of kilometers away at a different time. The



(A) Strain builds up in rocks in seismically active areas until the rocks rupture or move along preexisting fractures.



(B) Energy is released when the rocks rupture, and seismic waves move out from the point of rupture.

FIGURE 18.1 **Earthquakes originate** where rocks are strained beyond their elastic limits and rupture.

first waves to arrive are known as **primary waves** (**P waves;** Figure 18.4). These are a kind of **compressional wave,** identical in character to sound waves passing through a liquid or gas. The wave transmits energy by compressing and dilating the material through which it moves. Thus, the particles involved in these waves move short distances forward and backward in the direction of wave travel. P waves commonly have smaller amplitudes than the later waves. The next waves to arrive are **secondary waves (S waves).** In these, particles oscillate back and forth at right angles to the direction of wave travel. In other words, the particles shear or slide past one another. These **shear waves** cannot move through liquids. S waves cause a second burst of strong movements to be recorded on a seismograph (Figure 18.4A). The last waves to arrive are **surface waves,** which travel relatively slowly over Earth's surface. Particles involved in one type of surface wave move in orbits, similar to particles in water waves. They may have amplitudes up to 0.5 m and wavelengths of about 8 m.

Earthquake Locations

The location of an earthquake's focus is important in the study of plate tectonics because it indicates the depth at which rupture and movement occur. Although the movement of material within Earth occurs throughout the mantle and core, earthquakes are concentrated in its upper 700 km.

Within the 700-km range, earthquakes can be grouped according to depth of focus. **Shallow-focus earthquakes** occur from the surface to a depth of 70 km. They occur in all seismic belts and produce the largest percentage of earthquakes. **Intermediate-focus earthquakes** occur between 70 and 300 km below the surface, and **deep-focus earthquakes** between 300 and 700 km. Both intermediate-focus

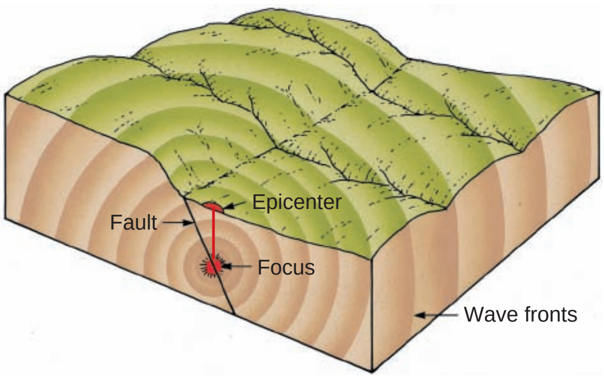
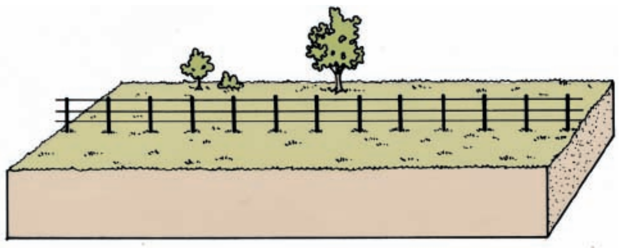
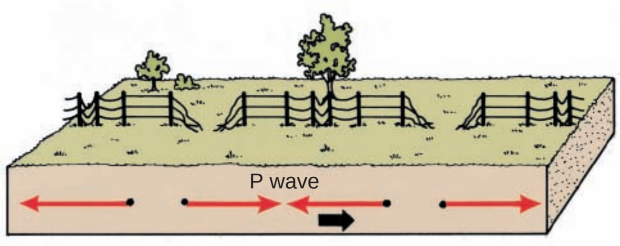


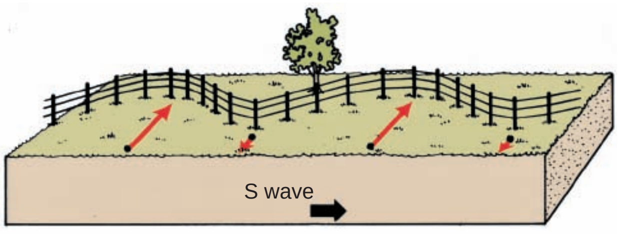
FIGURE 18.2 **The relationship between an earthquake's focus, its epicenter, and seismic wave fronts** is depicted in this diagram. The focus is the point of initial movement on the fault. Seismic waves radiate from the focus. The epicenter is the point on Earth's surface directly above the focus.



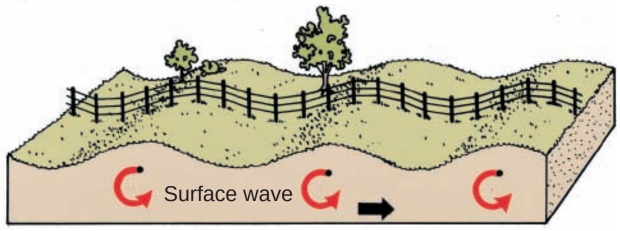
(A) Before seismic disturbance. A regular grid or a straight fence line provide good reference markers for future movement.



(B) Motion produced by a P wave. Particles are compressed and then are expanded in the line of wave progression. P waves can travel through any Earth material.



(C) Motion produced by an S wave. Particles move back and forth at right angles to the line of wave progression and are commonly called shear waves. S waves travel only through solids.



(D) Motion produced by one type of surface wave. Particles move in nearly circular paths at the surface. The motion diminishes with depth, like that produced by surface waves in the ocean.

FIGURE 18.3 **Motion produced by various types of seismic waves** can be illustrated by the distortions they produce in a regular grid. For comparison, the motion of one type of surface wave is shown.

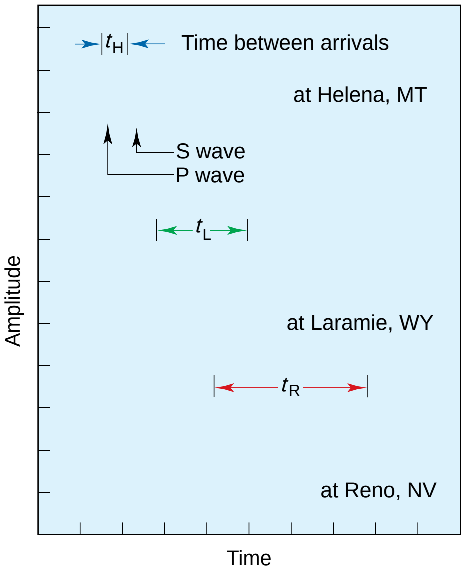
and deep-focus earthquakes are limited in number and distribution. In general, they are confined to convergent plate margins. The maximum energy released by an earthquake tends to become progressively smaller as the depth of focus increases. Also, seismic energy from a source deeper than 70 km is largely dissipated by the time it reaches the surface. Most large earthquakes therefore have a shallow focus, originating in the crust. The depth of an earthquake's focus is calculated from the time that elapses between the arrivals of the three major types of seismic waves.

The method of locating an earthquake's epicenter is relatively simple and can be understood easily by referring to Figure 18.4. The P wave, traveling faster than the S wave, is the first to be recorded at the seismic station. The time interval between the arrival of the P wave and the arrival of the S wave is a function of the station's distance from the epicenter. By tabulating the travel times of P and S waves from earthquakes of known sources, seismologists have constructed time-distance graphs, which can be used to determine the distance to the epicenter of a new quake. The seismic records show the distance, but not the direction, to the epicenter. Records from at least three stations are therefore necessary to determine the epicenter's precise location.

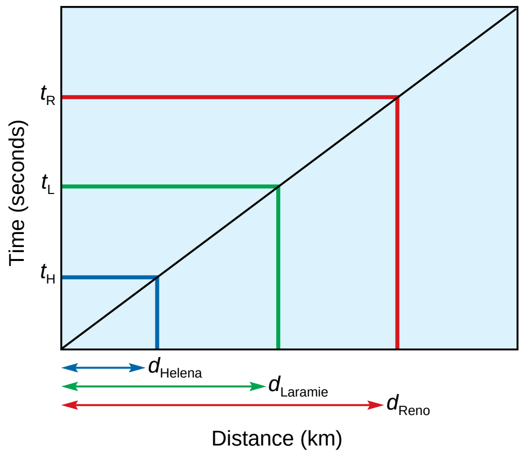
Intensity

The **intensity,** or destructive power, of an earthquake is an evaluation of the severity of ground motion at a given location. It is measured in relation to the effects of the earthquake on humans. In general, destruction is described in subjective terms for the damage caused to buildings, dams, bridges, and other structures, as reported by witnesses.

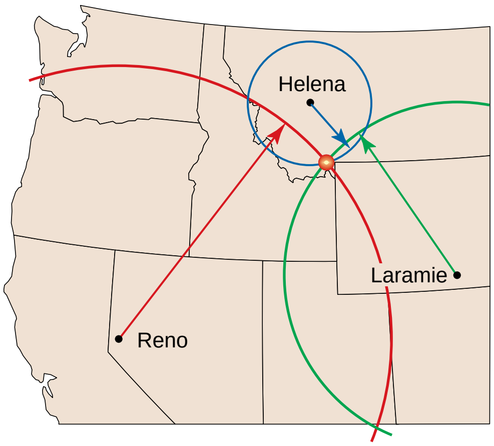
The intensity of an earthquake at a specific location depends on several factors. Foremost among these are (1) the total amount of energy released, (2) the distance from the epicenter, and (3) the type of rock and degree of consolidation. In general, wave amplitude and destruction are greater in soft, unconsolidated material than in dense, crystalline rock (Figure 18.5).



(A) The greater the distance between the seismic event and a recording seismograph station, the more time it takes for the first wave to arrive. Also, the greater the distance, the longer the interval between the arrival of P waves and S waves.



(B) The time between the arrivals of P waves and S waves is correlated with the distance between the seismic event and the recording station. For example, time at Helena, t_H, yields distance from Helena, d_{Helena}.



(C) The direction of the event from any single station is not known, but simply plotting the intersection of three arcs that have radii the respective distances from the three stations identifies a common point. That point lies at the epicenter of the seismic event.

FIGURE 18.4 **Locating the epicenter of an earthquake** is accomplished by comparing the arrival times of P waves and S waves at three seismic stations.

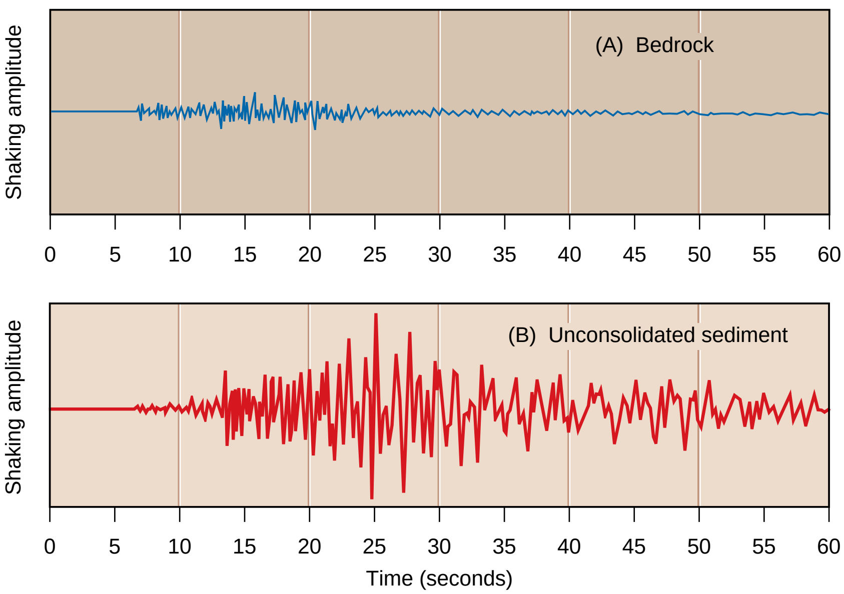


FIGURE 18.5 **The intensity of an earthquake can be magnified** in locales with unconsolidated sediment or landfill. The seismograms shown here are for the same aftershock of the 1992 Landers earthquake in California. Seismographs located on unconsolidated sediment measured much greater wave amplitudes and durations than seismometers on solid bedrock. Amplification can be by a factor of 10 or more.

TABLE 18.1
Earthquake Magnitude Scale

Magnitude	Approximate Number per Year
1	700,000
2	300,000
3	300,000
4	50,000
5	6000
6	800
7	120
8	20
>8	1 every few years

Magnitude

The **magnitude** of an earthquake is an objective measure of the amount of energy released. It is a much more precise measure than intensity. Earthquake magnitudes are based on direct measurements of the size (amplitude) of seismic waves, made with recording instruments, rather than on subjective observations of destruction. The total energy released by an earthquake can be calculated from the amplitude of the waves and the distance from the epicenter. The first attempt by seismologists to express magnitudes of earthquakes resulted in the Richter scale (Table 18.1), which assigns a single number to an earthquake. Each step on the scale represents an increase in wave amplitude by a factor of 10. The vibrations of an earthquake with a magnitude of 2 are therefore 10 times as great in amplitude as those of an earthquake with a magnitude of 1, and the vibrations of an earthquake with a magnitude of 8 are 1 million times as great in amplitude as those of an earthquake with a magnitude of 2. The increase in total energy released, however, is about 30 times for each step on the scale. The largest earthquake ever recorded had a magnitude of approximately 8.8 on the Richter scale. Significantly larger earthquakes are not likely to occur because rocks are not strong enough to accumulate more energy.

Recent refinements of the earthquake magnitude scale attempt to better distinguish the differences in large earthquakes. One such modification, called the **moment magnitude** scale, is the most widely used measure of earthquake magnitude today. The moment magnitude scale is designed to reflect the amount of *energy* released by an earthquake. The magnitudes reported by the media are commonly from this scale.

What is the difference between the magnitude and intensity of an earthquake?

An added advantage of the moment magnitude scale is that the magnitude of an ancient earthquake can be calculated from measurements of the amount of slip along the fault scarp. Like the standard Richter scale, moment magnitudes are logarithmic and range to about 10, but the absolute values are slightly different on the two scales, especially at the high end of the scale. Thus, the 1906 San Francisco earthquake had a Richter magnitude of 7.9 and a moment magnitude of 8.2. The largest known earthquake on the moment magnitude scale had a magnitude of 9.5. Moment magnitudes are used in the rest of this chapter.

EARTHQUAKE HAZARDS

> In addition to ground shaking and surface faulting, earthquake hazards include submergence, liquefaction, tsunamis, and landslides. Fires and floods caused by breakage of water lines or dam failures are also important.

Earthquakes pose a significant threat to much of the world's population. On average, several tens of thousands of people die each year because of large earthquakes. Figure 18.6 shows the expected probability of earthquakes in the conterminous United States. The principal areas exposed to earthquake risk are in the western United States—California, Nevada, Utah, and Montana—but there are significant risks elsewhere in the country as well. The primary effect of earthquakes is the violent ground motion caused by movement along a fault. This motion can shear and collapse buildings, dams, tunnels, and other rigid structures (Figure 18.7). Secondary effects include soil liquefaction, landslides, tsunamis, and submergence of the land. Following are a few examples of well-documented earthquakes.

What kinds of secondary damage are caused by earthquakes?

San Francisco, 1906. The most destructive earthquake in the history of the United States was the San Francisco earthquake of 1906. This shallow earthquake was located on a transform plate boundary. It lasted only a minute but had a magnitude of 8.2. The fire that followed it caused most of the destruction (an estimated $400 million in damage and a reported loss of 700 lives). From a scientific point of view,

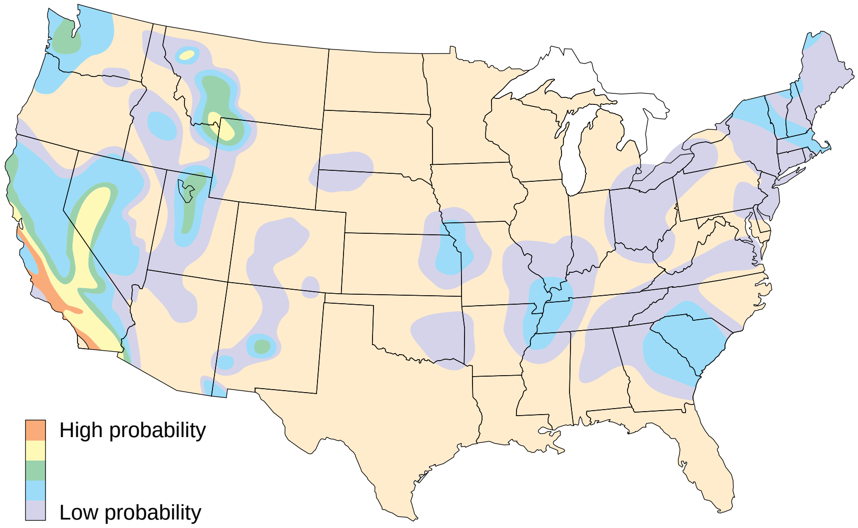


FIGURE 18.6 **A seismic risk map** shows the likelihood that an earthquake of a certain magnitude will occur. These predictions are based on the distributions and sizes of past earthquakes.

(A) Earthquake of June 16, 1964, Niigata, Japan. Apartment houses tilted by liquefaction. About one-third of the city subsided as much as 2 m as a result of sand compaction. *(Courtesy of U.S. Geological Survey)*

(B) Earthquake of January 1994, Northridge, California. Extensive damage to freeway overpasses occurred. *(Photograph by Geo-Tech Imagery)*

(C) Earthquake of September 19, 1985, Mexico City, Mexico. A 15-story reinforced concrete building collapsed. *(Courtesy of U.S. Geological Survey)*

(D) Earthquake of February 9, 1971, San Fernando, California. Southern Pacific railroad tracks near Los Angeles were laterally displaced. *(Courtesy of U.S. Geological Survey)*

FIGURE 18.7 **The effects of historic earthquakes** are dramatically displayed in the types of damage rendered to buildings and other structures.

the earthquake was important because of the visible effects it produced along the San Andreas fault zone. Horizontal displacement occurred over a distance of about 400 km and offset roads, fences, and buildings by as much as 7 m.

Alaska, 1964. The earthquake that devastated southern Alaska late in the afternoon of March 27, 1964, was one of the largest tectonic events of modern times. This convergent boundary earthquake had a moment magnitude of 9.2, and its duration ranged from 3 to 4 minutes at the epicenter. Despite its magnitude and severe effects, this quake caused far less property damage and loss of life than other national disasters (114 lives were lost and property worth $311 million was damaged) because, fortunately, much of the affected area was uninhabited. The crustal deformation associated with the Alaskan earthquake was the most extensive ever documented. The level of the land was changed in a zone 1000 km long and 500 km wide. Submarine and terrestrial landslides triggered by the earthquake caused spectacular damage to communities, and the shaking spontaneously liquefied deltaic materials along the coast, causing slumping of the waterfronts of Valdez and Seward. These landslides triggered destructive tsunamis that swept the state's southern shoreline (Figure 18.8). Fires started when oil tanks at Seward harbor ruptured.

Northridge, California, 1994. Early in the morning of January 17, 1994, an earthquake of magnitude 6.6 struck Northridge, California, just north of Los Angeles. It was followed by thousands of smaller aftershocks, including at least two with magnitudes greater than 5. This earthquake, although modest compared with the much larger magnitude-8 earthquakes that can occur along the San Andreas Fault, was one of the most destructive in the region. It caused the deaths of more than 50 people and $30 billion in property damage. The earthquake occurred on a previously unknown reverse fault that does not reach the surface but dips beneath the heavily populated San Fernando Valley. The focus of the quake was at a depth of about 14 km. Shaking caused considerable damage, including the collapse of several apartment buildings (where most of the people died), a parking garage, and several freeway overpasses (Figure 18.7B). The ground surface was ruptured along a 15-km stretch, with offsets as much as 20 cm. Overall, the ground was uplifted by about 1 m, locally disrupting sidewalks, streets, and buildings.

What is the difference between the tectonic setting of the earthquakes in Japan versus those in California?

FIGURE 18.8 Tsunamis from the 1964 Alaskan earthquake devastated the bay at Seward. An overturned ship, demolished truck, and torn-up dock strewn with logs and scrap metal attest to the power of the wave. A section of the waterfront slid into the bay. Waves spread in all directions, destroying railroad docks, washing out railroad and highway bridges. Flaming petroleum spread over the water, igniting homes and an electrical generation plant. *(Courtesy of National Geophysical Data Center)*.

FIGURE 18.9 The Northridge, California, earthquake of January 1994 caused several types of damage. An aftershock caused landslides and associated dust plumes on Santa Paula Ridge. *(Courtesy of P. Morton)*

The earthquake was also responsible for much secondary damage. Landslides were common on steep mountain slopes, stripped bare of their vegetation by earlier brush fires (Figure 18.9). Consequently, the Pacific Coast Highway and many local streets were closed, hampering rescue efforts. Much of the secondary damage was the result of destruction to built structures. Electrical power was cut off for more than 3 million people. An oil main and more than 250 natural gas lines were ruptured. Many homes and some businesses and university buildings burned, as sparks ignited gas leaking from the broken lines. Several homes were caught in a fireball ignited when someone tried to start a truck moments after the earthquake. Others were flooded when water lines were ruptured. Striking views of towering flames and billowing smoke above flooded streets were shown on television and in newspapers, vividly recording the significance of these secondary effects.

Kobe, Japan, 1995. The worst earthquake to hit Japan since 1923, this tremor destroyed much of the port city of Kobe on January 17, 1995. In this city of 1.4 million, nearly 5500 died, largely because of building collapse. The homes of 300,000 people were rendered unsafe. More than 600 fires started from broken gas lines and burned out of control because of ruptured water lines. The Kobe earthquake had a moment magnitude of about 7.2, and its focus was about 20 km deep.

In monetary terms, this was the largest earthquake disaster in history, causing direct losses of $140 billion. If an earthquake of similar magnitude had shaken the more densely populated Tokyo area, it could have caused direct losses of more than $1 trillion, to say nothing of casualties.

The cause of the Kobe earthquake was the subduction of the oceanic Philippine plate beneath southern Japan (Figure 18.10). Note in the figure that the directions of plate movement are not head-on but converge at an oblique angle. Because of this oblique collision, part of the displacement is taken up by movement along a long strike-slip fault zone. The main shock occurred along a fault in this zone.

The actual rupture of rock caused by the earthquake started 20 km from Kobe and moved rapidly toward the city. It broke the surface along a northeast-trending zone at least 9 km long. Thousands of aftershocks continued on this same trend in a zone 60 km long for several days after the quake. Both vertical and horizontal movements were about 1 or 2 m. An earthquake of this size probably occurs every 1000 to 1500 years along this strand of the main Kobe fault zone.

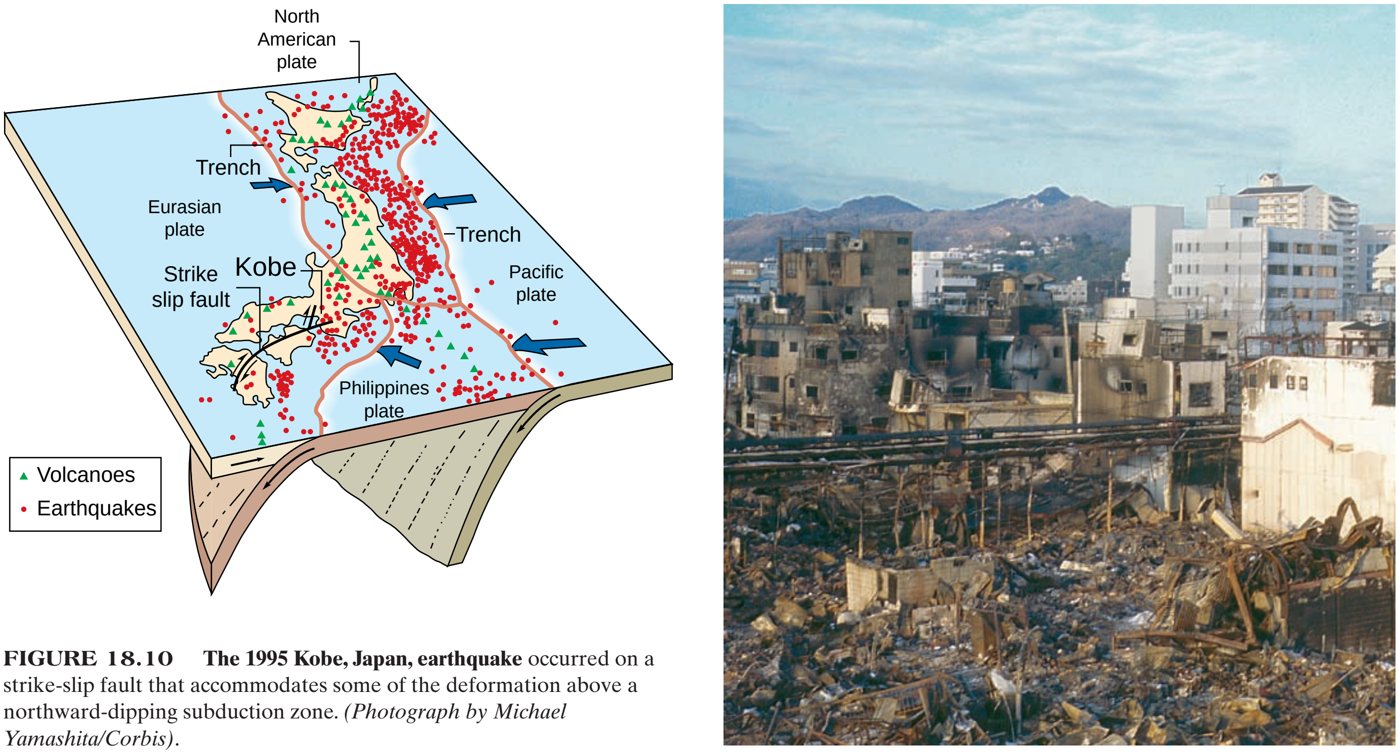


FIGURE 18.10 **The 1995 Kobe, Japan, earthquake** occurred on a strike-slip fault that accommodates some of the deformation above a northward-dipping subduction zone. *(Photograph by Michael Yamashita/Corbis)*.

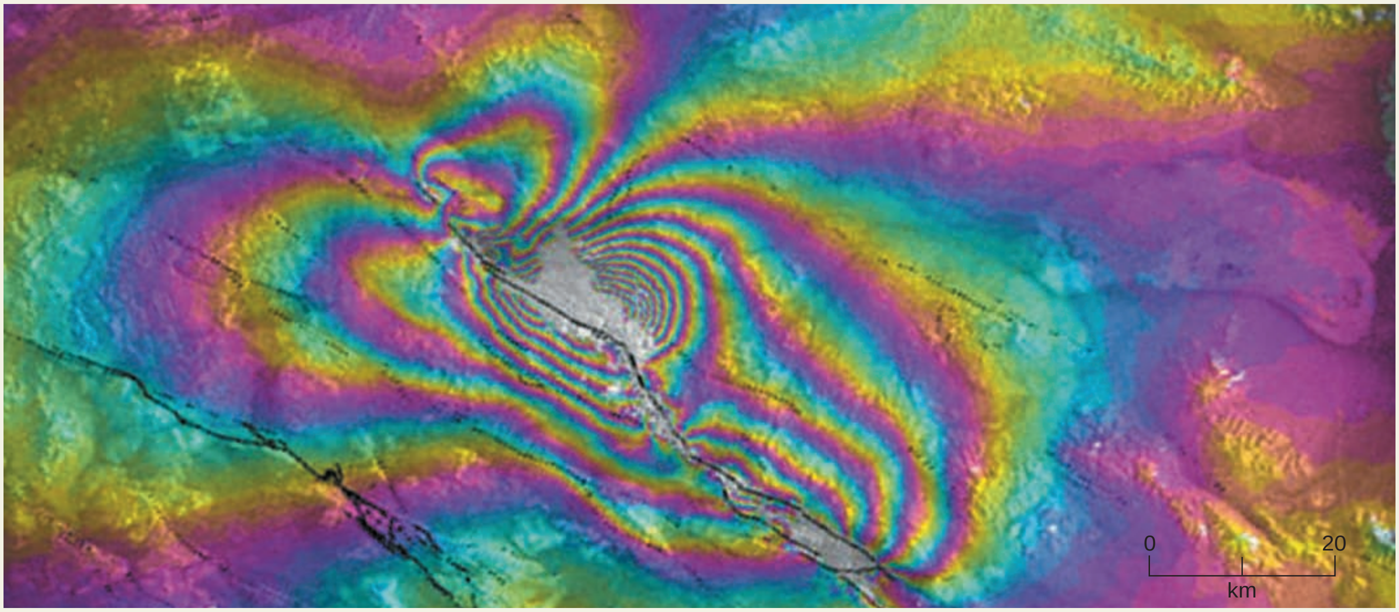
Shaking and liquefaction were the main causes of damage. **Liquefaction** occurs when unconsolidated, water-saturated regolith, soil, or landfill loses its strength and behaves like a fluid when shaken by an earthquake. Liquefied soils are unable to support buildings and other structures. In the Kobe earthquake, extensive liquefaction destroyed many buildings, bridges, highways, and utilities (sewer, gas, and water lines). Much of the city is built on human-made islands and landfill placed atop a granite basement and thus was susceptible to failure. Liquefaction most strongly affected areas with shallow water tables. As a result, geysers of wet sand erupted from fissures and covered many islands in the city. Subsidence, also a result of liquefaction, ranged from 0.5 to 3 m. Many buildings on landfills tilted because of ground settlement. However, buildings supported on deep piles sustained little or no damage.

Why does Japan have so many earthquakes?

This earthquake was about the same size as the Northridge, California, earthquake. However, many more people were killed in Kobe, and more damage was done. The likely reason was not any difference between the two earthquakes but a great difference in the human aspect: Kobe has an extremely dense population, older buildings, and construction on liquefiable landfill materials.

Izmit, Turkey, 1999. A long strike-slip fault slices across northern Turkey, connecting two convergent plate boundaries (Figure 18.11). Its size and deadly potential are comparable to that of California's San Andreas fault. During the early morning hours of August 17, 1999, one segment of the fault broke, producing a magnitude 7.4 earthquake. The epicenter was near the city of Izmit, 80 km south of Istanbul. A single gigantic heave created displacements of 3 to 4 m along a rupture 160 km long. The earthquake destroyed hundreds of buildings, damaged industrial and port facilities, a military base, pipelines, and roads, and was responsible for the collapse of bridges. As a result, nearly 20,000 people died and 600,000 were homeless. In some cities near the epicenter, 70% of the buildings collapsed or were uninhabitable. Surface faulting, ground shaking, subsidence, liquefaction, and even a small tsunami were responsible for most of the damage. Thousands of aftershocks plagued the area for several months after the earthquake. The large number of deaths in this heavily populated region brought the practices of contractors and building inspectors under intense scrutiny.

STATE OF THE ART Radar Interferometry Reveals Earthquake Deformation



(Courtesy of NASA/JPL/Caltech)

On October 16, 1999, a huge earthquake was triggered at a depth of about 6 km along the infamous San Andreas fault system of southern California. With a moment magnitude of 7.1, it set off a series of waves that rumbled through the crust of the sparsely populated region. Moments later, the tear reached the surface and rapidly spread along the fault system. The maximum strike-slip displacement along the fault was 5 m, and the rocks on one side of the strike-slip fault heaved upward, forming a scarp 3 m high. Breakage extended for a distance of 50 km and to a depth of about 15 km. Beyond these limits, the earthquake energy was insufficient to actually break the rocks.

High above this scene an orbiting satellite (European Remote Sensing satellite ERS-2) snapped a picture of the Mojave desert with its radar acting as a flashlight to illuminate the scene below. By comparing this image with one taken a month earlier, a map of the extent of disruption was made—without ever setting foot in the desert.

This technique is called *satellite interferometry* and is based on the simple notion of a before-and-after picture—with an ingenious twist. Optical sensors record only the brightness (or amplitude) of light waves reflected off a surface. Radar instruments measure both the amplitude and the exact point in the oscillation cycle where the radar wave hits the surface and bounces back to the satellite. This point is called the *phase* of the returned wave. Since radar waves have wavelengths of a few tens of centimeters, a surface change of only a few millimeters causes a significant phase change. A deviation of a centimeter results in a phase change of 40%, an amount that can be easily measured, even by a satellite orbiting hundreds of kilometers above. If two satellite images are taken from exactly the same position, there should be no phase difference for any spot on the two images. But if the ground changes ever so slightly between the two radar scans, then the phase of the wave returned from some spots in the second image are different. A map of these phase changes, called an *interferogram,* is displayed as a series of colored "fringes" that are like contour lines on a topographic map. Each fringe marks 10 cm of vertical ground motion.

The map shows a nearly continuous picture of the magnitude and distribution of the deformation caused by the Hector Mine earthquake. By starting at edges and counting the number of color bands and multiplying by 10 cm, you can see that deformation was strongest right along the surface rupture (black lines) and amounted to as much as 5 m. Clearly, displacement was not limited to a small area along the fault scarp. However, deformation declines with distance away from the fault.

This technique works best in regions where vegetation and other changes are minor. Thus, polar and arid regions are ideal. It provides many advantages over slow and expensive ground surveying, especially in remote areas. Even thick cloud cover does not obscure the surface from a probing radar system—water droplets and ice crystals do not impede the radio signals. Images can be taken in the dead of night because radar systems have their own illumination source.

Satellite radar interferometry can be used to rapidly map deformation along faults that have ruptured in earthquakes, to follow swelling volcanoes as magma accumulates beneath them, or to map their sagging as magma drains away. Other geologists use interferometry to quietly monitor landslides in remote areas or to watch the direction and amount of ice movement in glaciers or in sea ice. Ocean currents can also be measured by interferometry.

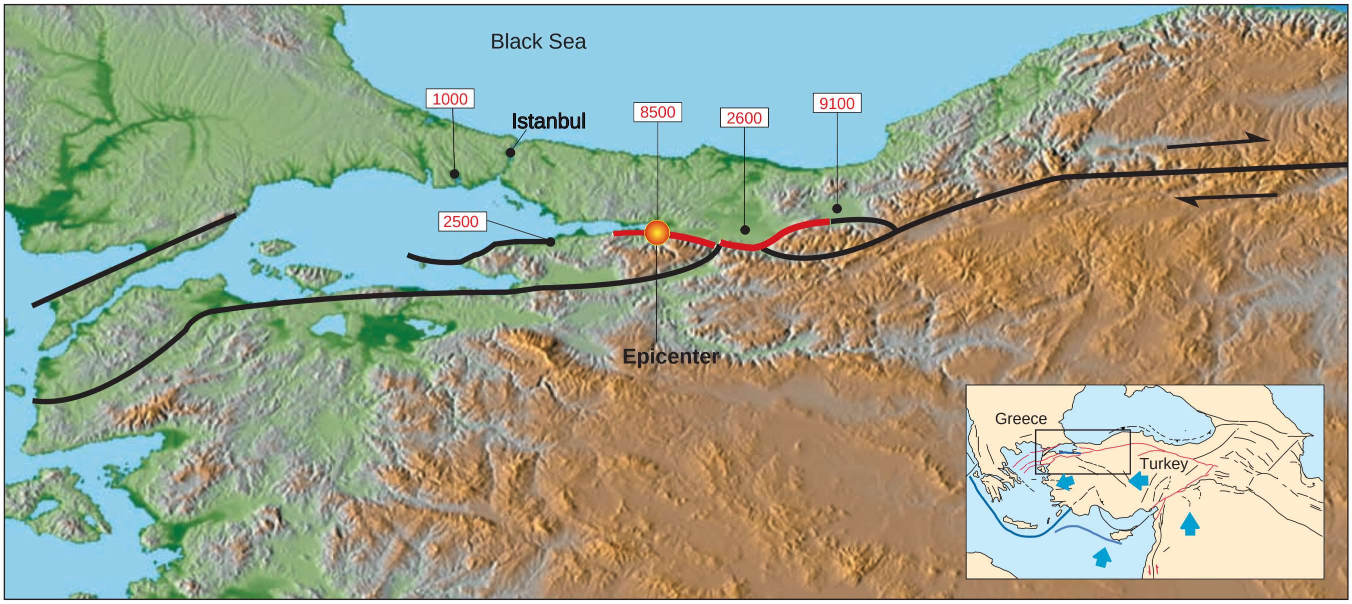


FIGURE 18.11 **The 1999 Izmit, Turkey, earthquake** occurred on a strike-slip fault system that connects two convergent plate boundaries. The red line shows the length of the fault that actually ruptured. The trace of the fault is marked by lakes, an inlet of the sea, and a line separating different types of landforms. It was responsible for the deaths of 20,000 people. The number of deaths at each location is given in the white boxes. (*Courtesy of Ken Perry, Chalk Butte, Inc.*)

Tangshan, China, 1976. The great Tangshan quake, which shook China in the early morning hours of July 28, 1976, was probably the second most devastating earthquake in recorded history. (The most destructive also occurred in China, in 1556.) In a matter of seconds, a large industrial city was reduced to rubble. The enormous shock registered 7.8 on the magnitude scale; an aftershock with a magnitude of 7.1 struck late in the afternoon, destroying structures that had withstood the main quake. The total amount of energy released by the earthquake and its aftershocks is almost unbelievable: the equivalent of 400 atomic bombs of the size dropped on Japan at the end of World War II. When the quake struck, many people were catapulted into the air, some as high as 2 m, by what were described as violent, hammer-like blows. The Tangshan quake killed 240,000 people, and many more were seriously injured. Most people were killed when their houses collapsed on them. All of the city's lifelines were destroyed, including bridges, railroads, telephone, electricity, water, and sewers. Fortunately, there were no natural gas lines in the city. Of the 680,000 residential buildings near the quake, 650,000 suffered serious damage. Eighty percent of the water reservoirs were damaged. Before the earthquake, little attention had been paid to earthquake resistance in the local building codes, and the buildings were highly vulnerable to seismic damage. Moreover, much of the city was built on young, unconsolidated sediments, which, as noted earlier, tend to amplify earthquake intensity (Figure 18.5).

The earthquake's focus was 11 km deep and caused movement in a fault zone 120 km long and 20 km wide. Ground fissures opened up, some with a lateral slip of 1.5 m. Liquefaction of wet sediments drove sand gushers from holes as wide as 1 m across. There were no foreshocks; in fact, no earthquakes of any size were detected in more than 2 months before the devastating blow came in July. In contrast, aftershocks continued for 4 years, and some 24,000 separate events were identified.

Other Earthquakes. The earthquake that shook Peru on May 31, 1970, had a magnitude of 7.8. Claiming the lives of 50,000 people, it was the deadliest earthquake in Latin American history. Much of the damage to towns and villages was caused by the collapse of adobe buildings, which are easily destroyed by ground motion. Eighty percent of the adobe houses in an area of 65,000 km^2 were destroyed. Vibrations caused most of the destruction to buildings, but the massive landslides triggered by the Andes quake were a second major cause of fatalities. In a matter of a few minutes, a huge debris avalanche buried 90% of the resort town of Yungay, with a population of 20,000. In 1797 similar earthquake-triggered debris avalanches killed 41,000 in Ecuador and Peru and, in 1939, killed 40,000 people in Chile.

Another type of earthquake hazard is exemplified by the 1960 earthquake that occurred in the Andes of Chile and caused extensive damage from a tsunami, in addition to ground motion, landslides, and flooding. It triggered a spectacular tsunami that devastated seaports with a series of waves 7 m high. This tsunami crossed the Pacific at approximately 1000 km/hr and built up to 11 m in height at Hilo, Hawaii. Nearly a day after the quake, the tsunami reached Japan, causing damage to property estimated at $70 million (Figure 15.29).

EARTHQUAKE PREDICTION

Effective short-term earthquake prediction, which could save many lives and millions of dollars in property damage, is proving to be elusive. Preparation is a more achievable goal.

Considerable effort has been made to find methods of predicting earthquakes. More than a million earthquakes occur around the world each year; about 50 are large enough to cause property damage and loss of lives. Some people have tried to relate earthquakes to sunspots, tides, changes in the weather, the alignment of planetary bodies, and other phenomena but have ignored many facts about Earth's seismicity. As a result, their predictions fail. Scientists have struggled with the problem, and significant strides have been made. Yet, predicting the location and magnitude of an earthquake is still an elusive goal.

Chinese scientists claim to have been successful in predicting about 15 earthquakes in recent years. In spite of the fact that there is no scientific evidence that animals can sense the advent of an earthquake, Chinese predictions rely heavily on the centuries-old idea that animals sense various underground changes before an earthquake and hence behave abnormally. Cattle, sheep, and horses refuse to enter their corrals; rats leave their hideouts and march fearlessly through houses; shrimp crawl on dry land; ants pick up their eggs and migrate en masse; fish jump above the surface of the water; and rabbits hop about aimlessly. Chinese scientists, using a variety of precursory activities, successfully predicted the large (magnitude 7.3) Haicheng earthquake that occurred on February 4, 1975. In late 1974 the water table in this region periodically rose and fell. Well water became turbid and the ground tilted in some places. The most bizarre events occurred in mid-December, when snakes came out of hibernation and froze to death, and groups of rats appeared and scurried about. These events were followed by a swarm of small earthquakes that led to a prediction of a magnitude-6 or larger earthquake by January or February. Continued abnormal animal behavior, and another swarm of earthquakes on February 1, were accompanied by variations in electrical conductivity of the ground. Some water wells spouted into the air; others stopped producing water. Finally, at 10 A.M. on February 4, seismologists convinced the provincial government to issue a warning by telephone to begin the evacuation of Haicheng. Schools, factories, and businesses were closed. Medical rescue teams were mobilized. The earthquake struck at 7:36 P.M. the same day. In some areas, more than 90% of the houses collapsed. Thousands of lives were spared, and many people were convinced that earthquakes could be predicted. This hope was dashed 2 years later by the unpredicted Tangshan earthquake already described.

Why is earthquake prediction so difficult?

An ambitious attempt to predict earthquakes has also been made along an active strand of the San Andreas Fault system near the small town of Parkfield, California (see Figure 20.19). This area was the site of a string of regularly spaced

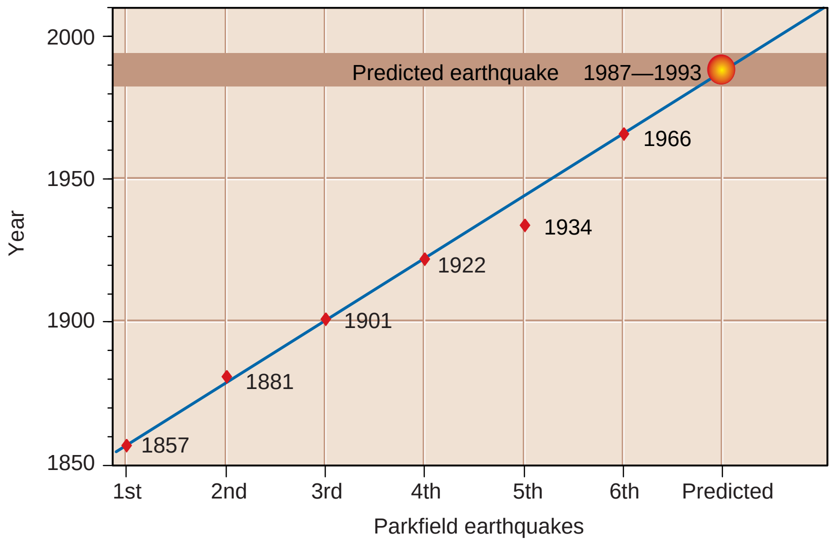


FIGURE 18.12 Six Parkfield earthquakes between 1857 and 1966 happened almost as regularly as the ticks of a clock. This regularity led to a prediction that a major earthquake would occur between 1988 and 1993 (yellow circle). That earthquake is now overdue showing just how difficult they are to predict.

earthquakes beginning in 1857 (Figure 18.12). Six large earthquakes, with magnitudes of about 6, occurred at Parkfield between 1857 and 1966. An earthquake occurred on average every 21 or 22 years. Recognizing this regular spacing, geophysicists predicted that a major earthquake should occur in about 1987, and that certainly one would occur before the end of 1993. As a result, the strike-slip fault near Parkfield has become a natural laboratory, with a host of instruments arrayed along the fault to measure strain across it, as well as small changes in elevation, release of gas from the soil, magnetic-field variations, electrical properties, and small earthquakes that might be foreshocks. In theory, a fault can begin to move weeks or months before the sudden rupture that marks a major earthquake.

The 1993 "deadline" has now come and gone and no major earthquake has struck Parkfield. Warnings were issued several times when small earthquakes, less than magnitude 4.7, were triggered on the fault. The warnings were based on the hypothesis that the small quakes could be foreshocks. Aside from those events, however, no sign of slip or other changes that might precede a major earthquake has been detected. The hypothesis that a fault is a simple system that gradually stores stress and then releases it at a specific threshold may need to be reconsidered. Some computer models of stress accumulation and relief through faulting suggest that fault behavior is likely to be so chaotic that the long-term prediction of earthquakes may be impossible. Have the monitoring and instrumentation at Parkfield been a waste of money? Certainly not. The information gleaned, even if it is negative with regard to earthquake prediction, has greatly expanded our understanding of the complex response of Earth's crust to tectonic stress.

Instead of attempting to predict the time, place, and magnitude of an expected earthquake, geologists are now concentrating on the more modest goal of forecasting which areas may be most susceptible to significant quakes. One approach is to calculate the probability that an earthquake of a certain magnitude will occur in a certain period ranging from decades to centuries.

Another contribution to forecasting has been the compilation of maps showing the seismic potential of major plate boundaries. Figure 18.13 essentially shows locations along the plate boundaries, where major quakes are most likely to occur in the near future. Along the plate margins are several gaps in seismic activity, where stress may be building to a critical level. The most susceptible areas are those where major tremors have occurred in the past, but not in the last 100 years. At various times, these gaps included such heavily populated areas as southern California, central Japan, central Chile, Taiwan, and the west coast of Sumatra. Some of the gaps have now been filled with recent earthquakes. For example, the

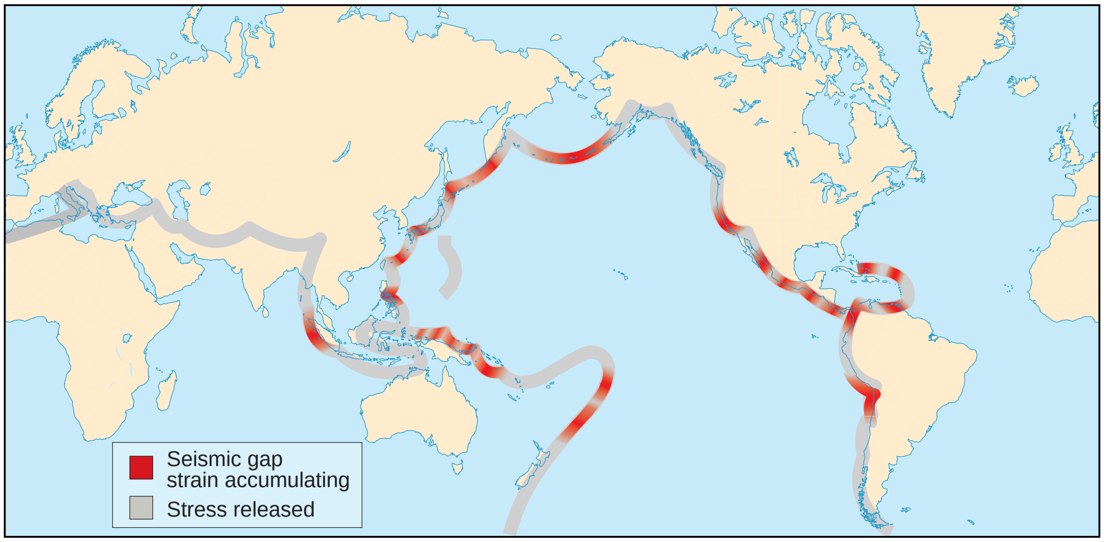


FIGURE 18.13 **Seismic gaps** are important in earthquake forecasting. Areas along plate margins that are not seismically active are believed to be building up stress and may be sites of significant seismic activity in the future. In the gray areas, earthquakes have relieved strain within the last 40 years, but in the red-shaded areas, no large quakes have occurred and strain is still building.

1989 Loma Prieta earthquake in northern California occurred in a seismic gap identified by the U.S. Geological Survey. The forecast called for a 30% probability of a magnitude-6.5 earthquake within a 30-year window. Other large earthquakes occurred in seismic gaps along the western coast of Mexico in 1979 and another off the coast of Nicaragua in 1992 (magnitude 7.6). The rest of these areas appear likely to experience a major earthquake (magnitude of 7 or greater) in the next few decades.

Earthquake Preparation

Because earthquakes are still difficult, if not impossible, to predict, societies around the world need to prepare for them in an attempt to minimize destruction. Preparation involves several important steps. The first is to acquire a thorough understanding of where past earthquakes occurred, under the assumption that they will strike there in the future. Such a **seismic risk map** is shown for the United States in Figure 18.6. Maps are also made for much smaller areas that show a specific type of earthquake hazard (surface rupture, landslides, and soil liquefaction, for example). Such maps can then be used as the basis for community zoning laws. In high-risk areas, stiffer building codes need to be in place. Critical facilities (hospitals, fire stations, and schools) should be placed where seismic hazards are lowest. Zoning laws should prohibit development in some areas. In other areas, even in the same city, where seismic hazards are smaller, zoning requirements and building codes can be more lenient, reflecting the smaller risk.

What is the best way to prepare for an earthquake?

To see how important and effective this approach is, compare the results of four earthquakes. The 1989 Loma Prieta (San Francisco) earthquake occurred in an area with strict building codes. The earthquake had a magnitude of 6.9, and about 65 people died. The 1988 earthquake in Armenia also had a magnitude of 6.9, but 28,000 people died when their poorly constructed homes and apartment buildings collapsed on them. In Latur, India, a magnitude-6.4 earthquake killed more than 11,000 people in 1993, in an area where little thought had been given to earthquake preparation. The walls in homes there were as much as 1.5 m thick, but the walls were made by stacking boulders together and filling the gaps between them with mud and pebbles. The walls crumbled as the earth shook. In the 1999 earthquake in Turkey where 20,000 people died, the importance of building with rigorous standards was revealed when whole neighborhoods built in violation of existing codes collapsed and adjacent ones built to the required standard survived. One expert concluded, "Almost all of the casualties could be attributed to buildings that collapsed because they were not built to code. Building codes save lives."

EARTHQUAKES AND PLATE TECTONICS

> The distribution of earthquakes delineates plate boundaries. Shallow-focus earthquakes coincide with the crest of the oceanic ridge and with transform faults between ridge segments. Earthquakes at convergent plate margins occur in a zone inclined downward beneath the adjacent continent or island arc.

How are earthquakes related to plate boundaries?

A worldwide network of 125 sensitive seismic stations was established in 1961 by the U.S. Coast and Geodetic Survey. Since then, the network has been expanded greatly, and the data received are processed by computers at many seismic data centers such as the one in Golden, Colorado. From the worldwide network of seismic stations, seismologists have compiled an amazing amount of data concerning earthquakes and plate tectonics. Not only are the locations and magnitudes of thousands of earthquakes established and plotted on regional maps each year, but other information, such as the direction of displacement on earthquake faults, is collected. The result is a new and important insight into the details of current plate motion.

Global Patterns of Earth's Seismicity

Tens of thousands of earthquakes have been recorded since the establishment of the worldwide network of seismic observation stations. Their locations and depths are summarized in the seismicity map in Figure 18.14. From the standpoint of Earth's dynamics, this map is an extremely significant compilation because it shows where and how the lithosphere of Earth is moving at the present time.

Why aren't there as many damaging earthquakes in Iceland, which sits astride a plate boundary, as in Japan?

Divergent Plate Boundaries. The global patterns of Earth's seismicity show a narrow belt of shallow-focus earthquakes that coincides almost exactly with the crest of the oceanic ridge and marks the boundaries between divergent plates. This zone is remarkably narrow compared with the zone of seismicity that follows the trends of young mountain belts and island arcs. The shallow earthquakes along divergent plate boundaries are usually less than 15 km deep and typically are small in magnitude. Earthquakes associated with the crest of the oceanic ridge occur within, or near, the rift valley. They appear to be associated with normal faulting and intrusions of basaltic magmas. Detailed studies indicate that earthquakes associated with the ridge crest are produced by normal faulting.

Transform Plate Boundaries. Shallow-focus earthquakes also follow the transform faults that connect offset segments of the ridge (Figure 18.14). Studies of fault motion indicate horizontal (strike-slip) displacement away from the ridge crest. Moreover, as is predicted by the plate tectonic theory, earthquakes are restricted to the active transform fault zone—the area between ridge axes—and do not occur in inactive fracture zones. The most damaging transform zone earthquakes are those found on land—including the San Andreas fault and the Anatolian fault of northern Turkey, where several strong earthquakes occurred in 1999.

Subduction Zones. On Earth, the most widespread and intense earthquake activity occurs along subduction zones at convergent plate boundaries. This belt of

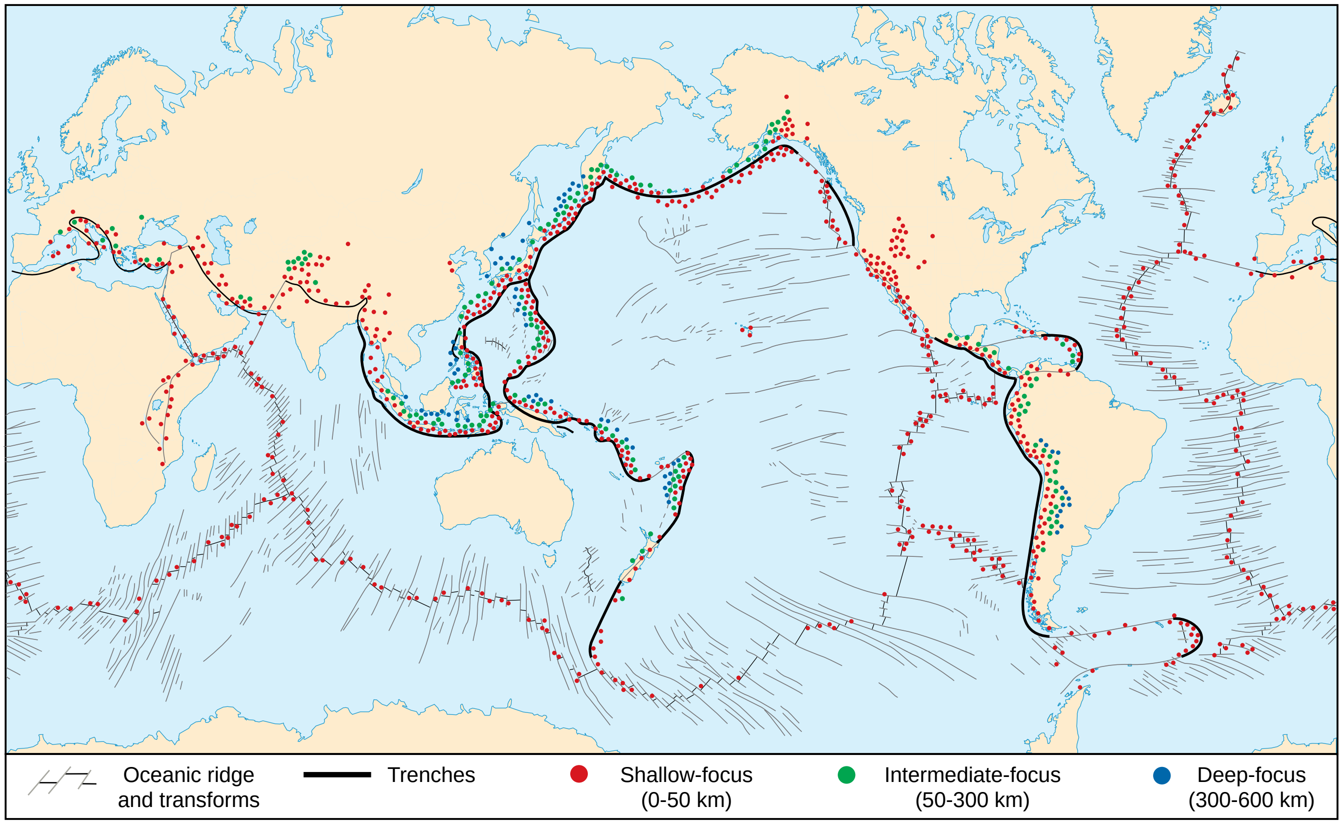


FIGURE 18.14 **Earth's seismicity** is clearly related to plate margins. This map shows the locations of many earthquakes that occurred during a 5-year period. Shallow-focus earthquakes occur at both divergent and convergent plate margins, whereas intermediate-focus and deep-focus earthquakes are restricted to the subduction zones of converging plates.

seismic activity is immediately apparent from the world seismicity map (Figure 18.14), which shows a strong concentration of shallow, intermediate, and deep earthquakes coinciding with the subduction zones of the Pacific Ocean. The three-dimensional distribution of earthquakes in this belt defines seismic zones that are inclined, at moderate to steep angles from the trenches, and extend down under the adjacent island arcs or continents. Along convergent plate boundaries, earthquakes occur at depths as great as 660 km. No other tectonic setting produces earthquakes as deep as this.

How are earthquakes at convergent plate boundaries different from those at divergent boundaries?

Collision Zones. The Himalayas and the Tibetan Plateau define a wide belt of shallow earthquakes. In this area, two continents collided—India and Asia. This convergence produced the wide zone of exceptionally high topography in the Himalayas and the Tibetan Plateau, but no deep earthquakes because there is no longer a subduction zone.

Intraplate Seismicity. Although most of the world's seismicity occurs along plate boundaries, the continental platforms also experience infrequent and scattered shallow-focus earthquakes. The zones of seismicity in East Africa and the western United States are most striking. They are probably associated with incomplete rifting. The minor shallow earthquakes, in the eastern United States (including New Madrid, Missouri, and South Carolina) and Australia, are more difficult to explain. Apparently, lateral motion of a plate across the asthenosphere involves slight vertical movement. Built-up stress can exceed the strength of the rocks within the lithospheric plate, causing infrequent faulting and seismicity along old lines of weakness such as ancient rifts. Although these continental intraplate earthquakes may be large, they are infrequent. In terms of the total energy released by seismicity each year, they account for only 0.5%.

SEISMIC WAVES AS PROBES OF EARTH'S INTERIOR

Seismic waves passing through Earth are refracted in ways that show distinct discontinuities within Earth's interior and provide the basis for the belief that Earth has a distinctive core.

How can seismic waves "X ray" Earth's internal structure?

Speculations about the interior of Earth have stimulated the imagination of humans for centuries, but only after we learned how to use seismic waves to obtain an "X-ray" picture of Earth were we able to probe the deep interior and formulate models of its structure and composition. Seismic waves—both P waves and S waves—travel faster through rigid material than through soft or plastic material. The velocities of these waves traveling through a specific part of Earth thus give an indication of the type of rock there. Abrupt changes in seismic wave velocities indicate significant changes in Earth's interior.

Seismic waves are similar in many respects to light waves, and their paths are governed by laws similar to those of optics. Both seismic rays and and light rays have velocities that depend on the kind of material through which they are transmitted. Both move in straight lines through homogeneous bodies. If the waves encounter a boundary between different substances, however, they are either reflected or refracted (bent). Familiar examples are light waves reflected from a mirror or refracted as they pass from air to water.

If Earth were a homogeneous solid, seismic waves would travel through it at a constant speed in all directions. A **seismic ray** (a line perpendicular to the wave front) would then be a straight line, like the ones shown in Figure 18.15. Early investigations, however, found that seismic waves arrive progressively sooner than was expected at stations progressively farther from an earthquake's source. The rays arriving at a distant station travel deeper through Earth than those reaching stations closer to the epicenter. Obviously, then, if the travel times of long-distance waves are progressively shortened as they go deeper into Earth, they must travel more rapidly at depth than they do near the surface. The significant conclusion drawn from these studies is that Earth is not a homogeneous, uniform mass, but has physical properties that change with depth.

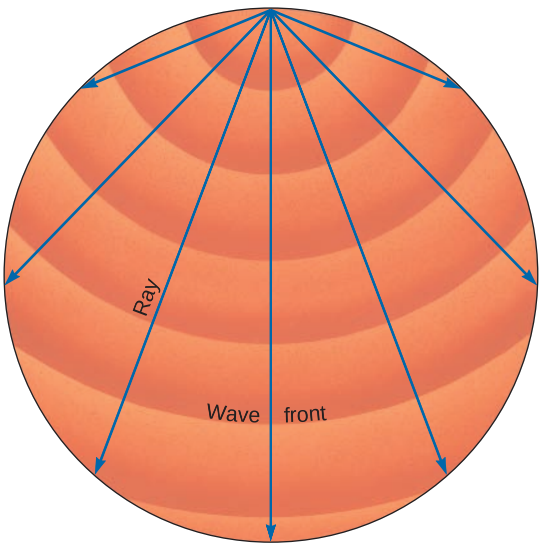


FIGURE 18.15 Seismic waves in a homogeneous planet would be neither reflected nor refracted. Lines drawn perpendicular to the wave fronts (rays) would follow linear paths.

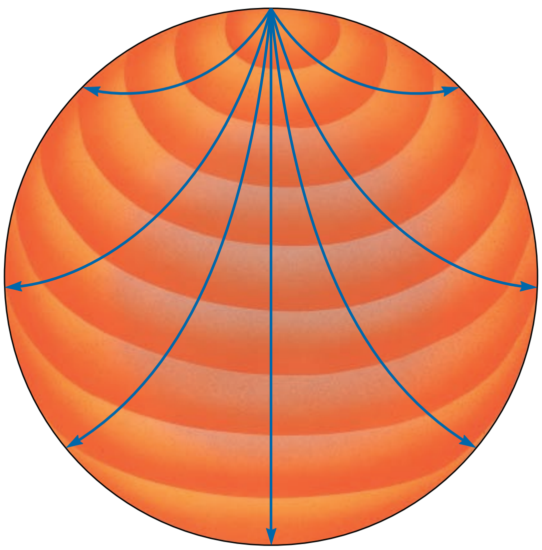


FIGURE 18.16 Seismic waves in a differentiated planet would pass through material that gradually increases in rigidity with depth. As a result, wave velocities would increase steadily with depth, and rays would follow curved path.

As a result of these differences with depth, seismic rays are believed to bend and follow curved paths through Earth (Figure 18.16).

In 1906 scientists first recognized that whenever an earthquake occurs, there is a large region on the opposite side of the planet where the seismic waves are not detectable. To better understand the nature and significance of this **shadow zone,** refer to Figure 18.17. For an earthquake at a particular spot (labeled 0°), a shadow zone for S waves invariably exists beyond 103° from the earthquake's focus. This huge S wave shadow zone extends almost halfway around Earth, opposite the earthquake's focus (Figure 18.17). Evidently, something stops the waves so that they do not reach the other side of Earth. This was the first evidence that Earth had a core made of something distinctly different from the rest of the planet. S waves simply do not pass through this core. One of the important properties of shear waves (like S waves) is their inability to move through liquids. S waves are transmitted only through solids that have enough elastic strength to return to their former shapes after being distorted by the wave motion. The fact that S waves will not travel through the core, therefore, is generally taken as evidence that the outer core is liquid. This, combined with Earth's magnetic field and high density, implied that the core was made of molten iron.

What is the principal evidence that Earth has a core?

The effect of the core on P waves is also informative but more complex (Figure 18.18). The shadow zone for P waves forms a belt around the planet between 103° and 143° away from the earthquake's focus (Figure 18.19). Evidently, the P waves are deflected but not completely stopped by Earth's core. Consequently, they are not detected in the shadow zone. Seismic rays traveling through the mantle follow curved paths from the earthquake's focus and emerge at the surface between 0° and 103° from the focus (slightly more than a quarter of the distance around Earth). In Figure 18.18, ray 1 just misses the core and is received by a station located 103° from the focus. Ray 2, however, being steeper than ray 1, encounters the core's boundary, where it is refracted. It travels more slowly through the core, is refracted again at the core's boundary, and is finally

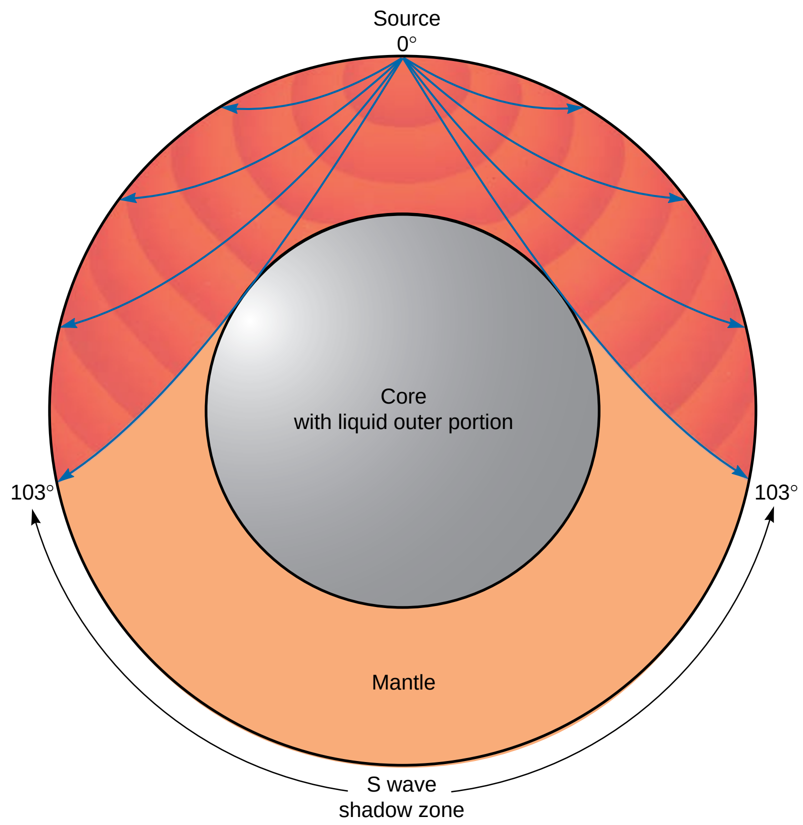


FIGURE 18.17 The shadow zone of S waves extends almost halfway around the globe from the earthquake's focus. This phenomenon can be explained if the outer core of Earth is liquid. Because S waves cannot travel through liquid, they do not pass through the core.

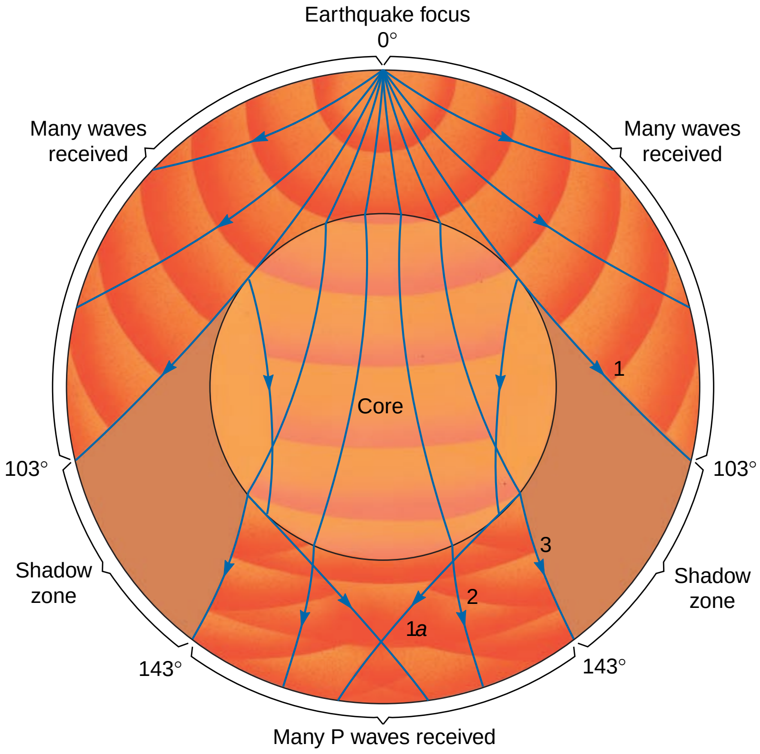


FIGURE 18.18 A P wave shadow zone occurs in the area between 103° and 143° from an earthquake's focus. The best way to explain the P wave shadow zone is to postulate that Earth has a central core through which P waves travel relatively slowly. Ray 1 just misses the core and is received at a station located 103° from the earthquake's focus. A steeper ray, such as ray 2, encounters the boundary of the core and is refracted. It travels through the core, is refracted again at the core's boundary, and is received at a station fewer than 180° from the focus. Similarly, ray 3 is refracted and emerges at the surface 143° from the focus. Other rays that are steeper than ray 1 are severely bent by the core, so that no P waves are directly received in the shadow zone. From shadow zones, seismologists calculate that the boundary of the core is 2900 km below the surface.

received at a station on the opposite side of Earth. Ray 3 is similarly refracted and emerges on the opposite side, 143° from the focus. Other rays that are steeper than ray 1 are also refracted through the core and emerge between 143° and 180° from the focus. Thus, refraction at the boundary between the core and the mantle causes the P wave shadow zone.

More recent studies of the P wave shadow zone show that some weak P waves are received in this zone. This is the evidence for the presence of a solid inner core, which deflects the deep, penetrating P waves in the manner shown in Figure 18.19.

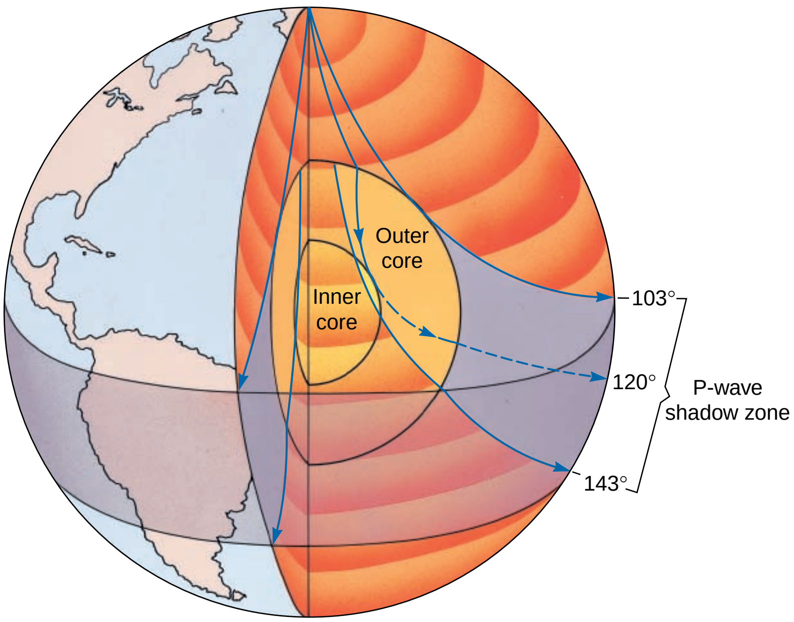


FIGURE 18.19 P waves are deflected by the inner core and are received in the shadow zone as weak, indirect signals. This deflection suggests that the inner core is solid.

SEISMIC WAVE VELOCITY DISCONTINUITIES

Seismic discontinuities reveal the size of Earth's crust, mantle, and core and show that they have different chemical compositions. In addition, seismic studies reveal much about the physical nature of the interior, revealing a solid inner core, a liquid outer core, a soft asthenosphere, and a rigid lithosphere. Seismic tomography is beginning to reveal the pattern of convection in the mantle.

With the present worldwide network of recording stations, even minor variations in seismic velocities with depth, known as **seismic discontinuities,** can be determined with considerable accuracy. Seismic wave velocity versus depth curves, like the one in Figure 18.20, provide a huge amount of significant information about Earth's interior. The first seismic discontinuity occurs between 5 and 70 km below the surface. This is known as the **Mohorovičić discontinuity,** or simply **Moho,** after Andrija Mohorovičić, the Croatian seismologist who first recognized it. The discontinuity is considered to represent the base of the crust and heralds an important compositional change from the feldspar-rich crust to the olivine-rich mantle. Seismic wave velocity studies also show that the continental crust is much thicker (25 to 70 km) than oceanic crust (about 8 km).

What are seismic discontinuities? What do they reveal about Earth's interior?

Perhaps the most significant discontinuity, however, is the low-velocity zone from 100 to 250 km below the surface (Figure 18.21). Beno Gutenberg, a German seismologist, recognized this zone in the 1920s. The normal trend is for seismic wave velocities to increase with depth in the mantle. In the low-velocity zone, however, the trend is reversed, and seismic waves travel about 6% slower than they do in adjacent regions. The generally accepted explanation for the low seismic wave velocities is that the mantle is very near its melting point or even partially molten, with perhaps 1% to 5% liquid. A thin film of liquid around the mineral grains may

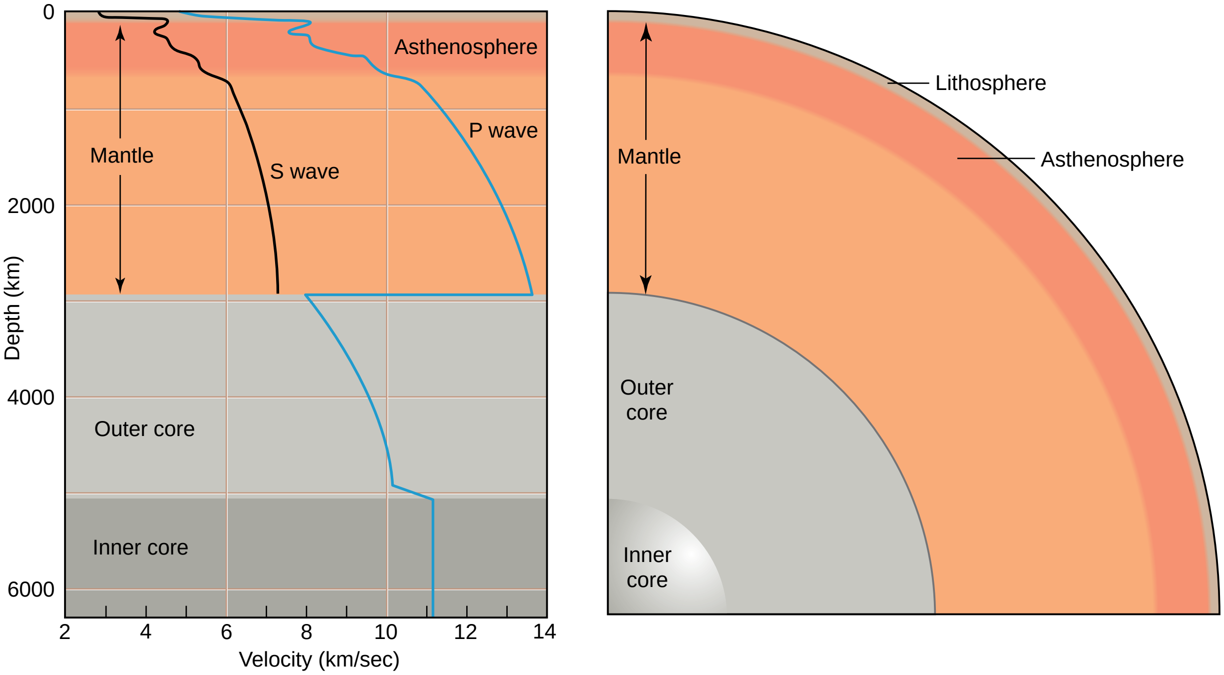


FIGURE 18.20 **The internal structure of Earth** is deduced from variations in the velocity of seismic waves at depth. The velocity of both P waves and S waves increases until they reach a depth of approximately 100 km. There the waves are slow until they have traveled to a depth of about 250 km. This low-velocity layer lies within the asthenosphere. Below this, the velocity of P waves and S waves increases until a depth of about 2900 km, where both velocities change abruptly. S waves do not travel through the central part of Earth, and the velocity of the P waves decreases drastically. This variation is the most striking discontinuity and indicates the boundary between the liquid outer core and the mantle. Another discontinuity in P wave velocity, at a depth of 5000 km, indicates the surface of the solid inner core.

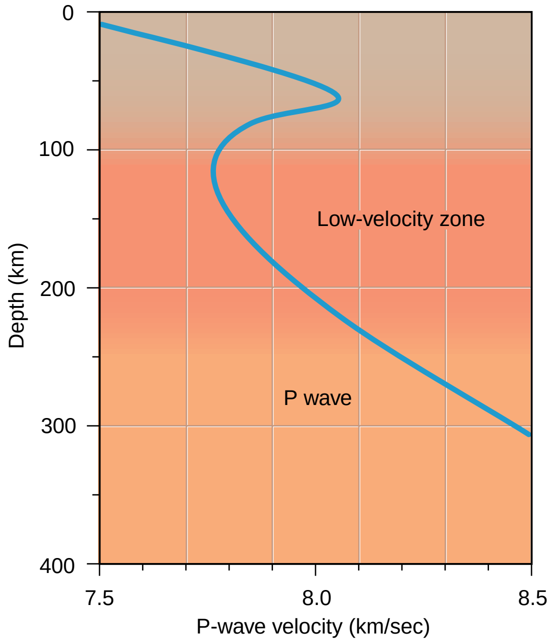


FIGURE 18.21 **The low velocity zone** is revealed by a drop in the velocities of both P waves (shown here) and S waves. This marks a zone of low strength in the upper mantle between about 100 and 250 km below the surface. The low-velocity zone is contained in the asthenosphere and marks part of the mantle that is very near its melting point and may be a zone of partial melting.

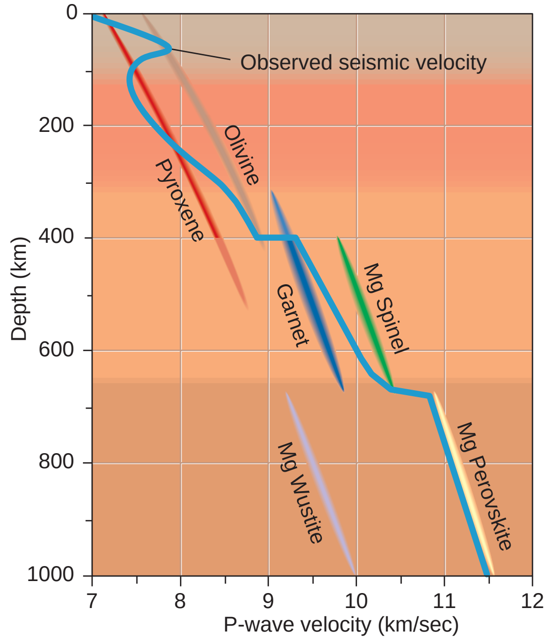


FIGURE 18.22 **Discontinuities** in seismic wave velocities may correspond to phase changes. The blue line shows how P wave velocities change with depth. The other lines show velocities for various minerals. The uppermost mantle is dominated by olivine. Below 400 km a velocity increase implies olivine is replaced by spinel. At greater depths, spinel is probably replaced by magnesium-perovskite. Each change increases the density of the mantle.

slow both S and P waves. Moreover, rocks near their melting points are very weak and ductile. The low-velocity zone is embedded within the asthenosphere. The asthenosphere plays a key role in the motion of tectonic plates at Earth's surface. If Earth lacked a weak, ductile asthenosphere, the upper part of the mantle would be directly tied—frozen, if you will—to the lower part of the mantle, and plate motions would be prohibited. Apparently, the asthenosphere effectively decouples the moving lithosphere from the lower part of the mantle.

Other changes in seismic velocities occur in the mantle at depths of about 400 and 660 km to create a transition zone between the upper and lower mantle (Figure 18.22). These discontinuities are probably caused by phase changes: a metamorphic transformation of the minerals in the mantle (unlike the compositional change that marks the Moho). Seismic velocities suggest olivine and pyroxene are the dominant minerals in the upper mantle. With increasing depth and pressure, denser phases, with more-compact atomic structures, become stable. For example, garnet gradually replaces pyroxene between 300 and 500 km in depth. Olivine is unstable at depths greater than about 400 km, where a denser mineral (magnesium spinel), with the same composition but a different internal structure, replaces it. This metamorphic transformation probably causes the seismic discontinuity at 400 km depth (Figure 18.22). The discontinuity at 660 km depth may be caused by the replacement of spinel by two different magnesium-rich minerals (magnesiowustite [(Mg,Fe)O] and perovskite [$(Mg,Fe)SiO_3$]). The seismic wave velocities suggest that perovskite is much more abundant (Figure 18.22).

The most striking variation in seismic wave velocities occurs at the core-mantle boundary, at a depth of 2900 km (Figure 18.20). There, S waves stop, and the velocities of P waves are drastically reduced. The seismic wave velocities and density of the outer core can be explained if there is a striking change in composition and physical state. Laboratory studies of seismic wave velocities and comparisons with meteorites show convincingly that the core must be made mostly of iron. The outer core is most likely made of molten iron mixed with nickel, together with some lower density elements such as silicon or sulfur. The light component is needed to explain the density of the core which is about 10% less than pure iron. The deepest discontinuity is a strong increase in seismic velocity. The increase shows that the inner core is less compressible and more rigid than the outer core. Apparently, the inner core is solid (Figure 18.20). Even though the temperature must be higher, the extremely high pressures found inside the deep Earth make the more compact solid phase of iron more stable.

CONVECTION INSIDE EARTH

Convection of the core and mantle is the most important mechanism of heat transfer in Earth. Convection in the iron core probably creates the magnetic field, and convection in the mantle creates mantle plumes and plate tectonics.

Observations such as the three-dimensional seismic tomographs are driving a revolution in our understanding about convection in Earth's interior. Our newly acquired ability to construct three-dimensional tomographic images of Earth's deep interior opens up the breathtaking prospect of tracing tectonic plates as they plunge below the surface in a subduction zone and descend into the deep mantle. In addition, calculations of how the mantle and the core flow are becoming increasingly more realistic as more complete data about the interior are used.

Convection in the Core

Seismic velocity studies clearly show Earth's core is made mostly of iron and divided into a liquid outer core and a solid inner core. With the help of three-dimensional computer models, we are beginning to chart the flow of the molten iron alloy that forms the outer core and speculate about the origin of Earth's magnetic field. At 5000°C, the temperature of Earth's core approaches the temperature of the surface of the Sun, but it is slowly cooling as the inner core crystallizes. Consequently, a system of convective currents is established in the molten outer core. Moving metals can produce an electric current if the metal passes through a magnetic field. This principle is used to make electrical generators (dynamos) wherein conductive metallic wires are spun inside a magnetic field. In turn, an electrical current can generate a magnetic field (see Figure 17.7). The magnetic field may be caused by Earth's rotation combined with convection of the molten metal in a shell surrounding the inner core (Figure 18.23). This creates a self-sustaining dynamo. Changes in these convective flow patterns may cause periodic polarity reversals (see Chapter 17).

Seismic maps of the core-mantle boundary show that the surface of the core is not smooth but is marked by broad swells and depressions with a difference in height of up to 20 km. A rough boundary could disturb the flow of the liquid iron in the outer core, much as a mountain influences the flow pattern of winds. Some seismic studies also suggest that Earth's solid inner core spins a bit faster than the rest of the planet.

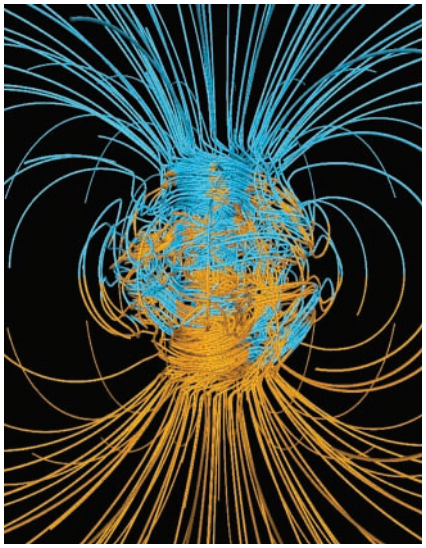


FIGURE 18.23 Earth's magnetic field probably forms by convection of the outer core, which is made of molten iron. A computer model of convection shows magnetic field lines as a smooth dipole with blue lines directed inward and gold field lines directed outward. The field inside the core is much more complex. *(Courtesy of Gary A. Glatzmaier, University of California, Santa Cruz.)*

Convection in the Mantle

Another way Earth shows its dynamic nature is by the large-scale convection of its mantle. Indeed, Earth is a large heat engine constantly churning by internal convection. The changes in mineral assemblage and density of the mantle outlined above may help control the way the interior of Earth convects. The result of combining seismic data about the mantle's layered structure with computer simulations of temperature distribution and internal convection are shown in Figure 18.24.

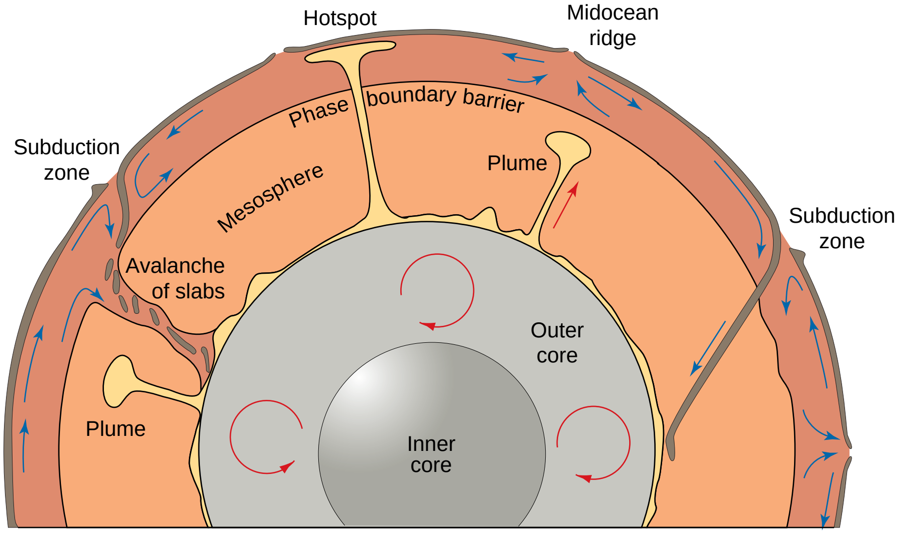


FIGURE 18.24 Earth's thermal structure and convection can be modeled using computers to complement the observations of seismic tomography. In one model, subducted slabs pass without pausing through the phase boundary at 660 km. In another model, the phase boundary is a temporary barrier that is broken down when enough subducted material accumulates and then flushes rapidly through the lower mantle. The lower mantle may convect by generating thin plumes that rise off of the core-mantle boundary. Some of the plumes may be triggered by the sinking of the dense overlying mantle.

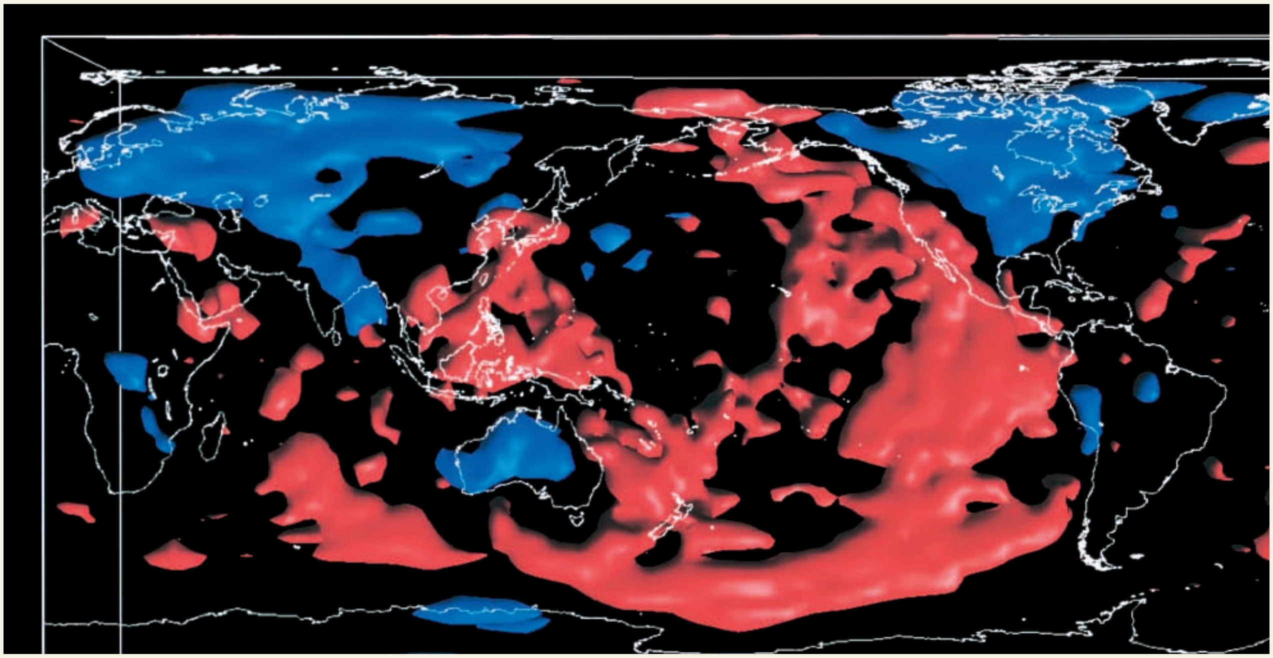
One possibility is that the upper mantle convects separately from the lower mantle because of the differences identified by seismic discontinuities (Figure 18.24). This model suggests that the mantle may convect in two more or less distinct layers that are usually separated by the 660 km discontinuity. The lower mantle may convect by generating narrow cylindrical plumes, shown in the model as yellowish mushroom-like fingers projecting from the core-mantle boundary. In the computer models, some of these plumes penetrate the upper mantle and reach all the way to the surface, supporting the notion that volcanic hotspots are related to deep upwellings in the mantle. Convection of the upper mantle may be driven by the sinking of cold, dense slabs of subducted lithosphere. Perhaps strong rock in the lower mantle prohibits the slabs from penetrating the phase boundary at 660 km. However, the cold slabs slowly accumulating above the boundary could eventually have enough weight to break through the barrier and "flush" into the lower mantle. To understand this process, think of what happens if you place a small piece of iron on a coffee table. Iron is denser than wood, but it would not immediately sink through the table because of the strength of the wood. If you continue stacking more and more iron on the table, it will eventually break and the iron will fall to the next barrier to its movement, presumably the floor. It may take several hundred million years of subduction to amass a cold sinker that could break through the boundary. Three-dimensional numerical models show that these avalanches take the shape of broad cylindrical pipes with greatly enlarged bases, created when the cold rock hits the dense, impenetrable core. Each flushing event may trigger an upward counterflow as material from the lowest part of the mantle moves upward to balance the downward flow of the cold avalanche (Figure 18.24). Periods of enhanced volcanism on the real Earth could be a plume response to a flushing event.

What does seismic tomography tell us about Earth's internal structure?

In a competing model, the whole mantle convects as a single unit. Subducting slabs of oceanic lithosphere may be dense enough to pass unobstructed through the boundary between the upper and lower mantle (Figure 18.24). Recently constructed tomographic sections give some support to the suggestion. They show that inclined sheets of anomalously cold rock extend through the 660 km discontinuity and into the lowermost mantle. Ultimately, deeply subducted oceanic lithosphere must stall at the core-mantle boundary because it is less dense than the metallic core. And indeed, a large concentration of anomalously "cold" rock has been found at the core-mantle boundary. Why would cold material be present in a place where we would expect to find hot mantle created by the flow of heat from the convecting iron core? These deep-mantle cold spots lie below subduction zones. Could the cold rock correspond to ancient oceanic lithosphere, still distinctive because of its composition and lower temperature? Has dense oceanic lithosphere accumulated there over billions of years, to form a "slab graveyard"?

Why do we believe the mantle is moving in a slow convection system?

STATE OF THE ART — *Seismic Tomography of Earth's Mantle*



(Courtesy of P. Morin)

During the last decade, scientists have utilized a new analytic technique, known as **seismic tomography,** that promises to greatly enhance our knowledge of the deep internal structure of Earth, including the pattern of flow in the mantle. Seismic tomography is like its medical analog, the CAT scan (computer-assisted tomography; tomograph is based on a Greek word, *tomos,* meaning "section"). In a medical CAT scan, X rays that penetrate the body from all directions are used to construct an image of a slice (cross section) through the body. Bones, organs, and tumors are identified because they have different densities and absorb X rays differently. With the aid of a computer, these images are stacked side by side to produce a three-dimensional view.

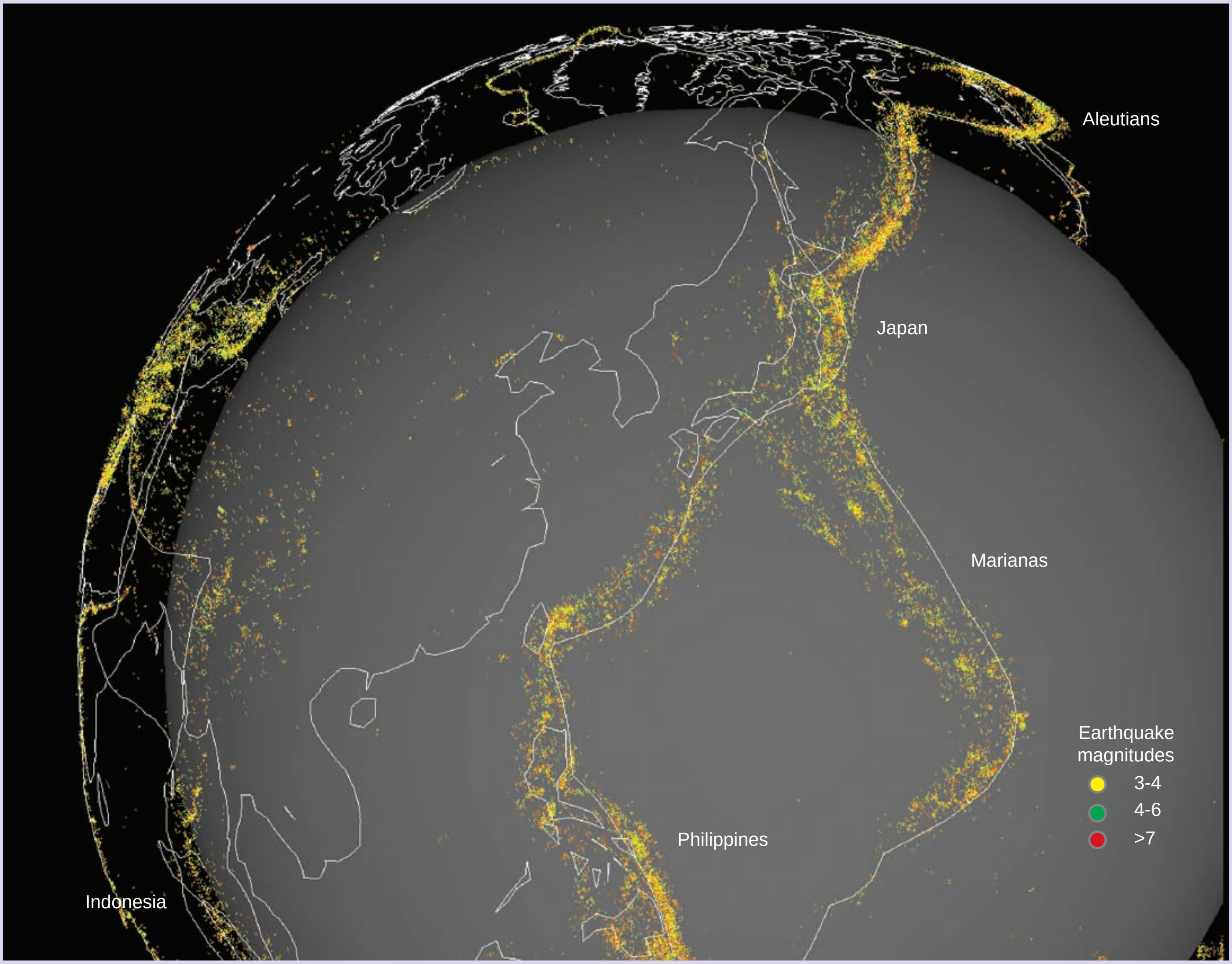
In seismic tomography of Earth's interior, natural seismic waves from earthquakes are used as X rays are. Using the vast network of seismographs around the world, seismologists analyze the velocities of hundreds of thousands of seismic waves as they pass through Earth in different directions.

Geophysicists use computers to produce three-dimensional images of Earth's interior from the data. The results show regions where seismic waves travel faster or slower than normal. Geophysicists know, from laboratory studies and from observations near volcanoes, that seismic waves travel slowly through the relatively weak, hot rock and quickly through stronger, cooler rock. The tomographs can thus be interpreted as temperature maps. In addition, hot parts of the mantle, being less dense than their surroundings, will rise, whereas cool mantle rock will sink. Thus, the tomograph can be used to outline the patterns of convective flow in the mantle.

The three-dimensional view of the mantle obtained from seismic tomography provides an unparalleled view of the effects of plate tectonics on Earth's interior. At a depth of 150 km, slow seismic zones occur under most of the volcanic regions, including the midocean ridges. This is evidence that the mantle directly beneath the midocean ridges is hotter and probably less dense than normal. As a result, it is probably rising. At still greater depths of 350 km, there is still hot rock (shown in red on the image) concentrated beneath the ocean ridge system. Beneath the mid-Atlantic ridge this zone of hot rock is not continuous, but is broken up into isolated segments. At depths even greater than those shown here, the relationship between mantle and surface features is even weaker. This indicates that the ocean ridge system is not simply the surface expression of vertical upwelling currents from the deepest mantle. Instead, midocean ridges must be fed by the movement of hot material in the *upper* mantle.

In contrast to the mantle beneath the ocean basins, the continental shields of Canada, Brazil, Siberia, Africa, and Australia are all underlain by mantle that has higher than normal velocities (blue on this tomograph) and must be some of the coldest mantle. This cold upper mantle forms a deep root beneath the continents and may not convect with the rest of the mantle, but is part of the lithosphere and moves with the continents.

GeoLogic Inclined Seismic Zones



(Courtesy of Paul Morin)

Earthquakes are like a two-edged sword. On the one hand, they are devastating and kill thousands of people and cause billions of dollars of damage each year. On the other hand, they have helped us understand many of the fine points about Earth's internal structure. For example, consider the following facts about earthquake focal depths in the northwestern Pacific region shown on this globe with a transparent crust.

Observations

1. Earthquakes occur only in brittle, cool rocks.
2. Earthquakes form narrow zones that are inclined below volcanic arcs with andesitic composite volcanoes.
3. Here, and around the Pacific Ocean, these inclined earthquake zones all dip beneath the adjacent continent.
4. The earthquake zone reaches the surface at a deep oceanic trench.
5. The deepest earthquakes on Earth occur in these inclined seismic zones.
6. Earthquakes do not occur below about 600 km depth.

Interpretations

Initially, this small set of observations led geologists and geophysicists to conclude that great slabs of oceanic crust were being thrust beneath the continents. Eventually, we realized that it was not just the crust that was involved, but an entire plate of lithosphere is descending into the interior. The oceanic lithosphere is denser than the continental lithosphere; in fact, it can become denser than the mantle beneath it and subduct back into Earth's interior. Earthquakes occur at great depths only in subduction zones, because the oceanic plate is cold and brittle as it descends into the warmer mantle. Once the plate descends to depths greater than 600 km, it is too warm to rupture and form an earthquake.

KEY TERMS

compressional wave (p. 511)
deep-focus earthquake (p. 511)
earthquake (p. 510)
elastic limit (p. 510)
elastic-rebound theory (p. 510)
epicenter (p. 510)
focus (p. 510)
intensity (p. 512)
intermediate-focus earthquake (p. 511)
liquefaction (p. 518)
magnitude (p. 514)
Moho (Mohorovičić discontinuity) (p. 529)
moment magnitude (p. 514)
P wave (primary wave) (p. 511)
S wave (secondary wave) (p. 511)
seismic discontinuity (p. 529)
seismic ray (p. 526)
seismic risk map (p. 523)
seismic tomography (p. 533)
seismic wave (p. 510)
seismograph (p. 510)
shadow zone (p. 527)
shallow-focus earthquake (p. 511)
shear wave (p. 511)
surface wave (p. 511)

REVIEW QUESTIONS

1. Explain the elastic-rebound theory of the origin of earthquakes.
2. Describe the motion and velocity of the three major types of seismic waves.
3. Explain how the location of an earthquake's epicenter is determined.
4. What secondary effects commonly accompany earthquakes?
5. Describe the difficulties geologists have encountered in trying to predict earthquakes.
6. How can a seismic gap be used to predict an earthquake?
7. Describe the global pattern of earthquakes.
8. Compare the relative earthquake hazard along a divergent, transform, and convergent plate boundary.
9. How does the depth of earthquakes indicate (a) convergent plate margins and (b) divergent plate margins?
10. Draw a diagram showing the paths that would be followed by seismic rays through Earth if the core were only half the diameter shown in Figure 18.17.
11. What do seismic velocity-depth diagrams (see Figures 18.18 and 18.20) tell us about Earth's internal structure?
12. What changes could explain the transition zone between the upper and lower mantle between 400 and 660 km?
13. How does the cause of the Mohorovičić discontinuity differ from the cause of discontinuities at 400 and 660 km?
14. Where is the low-velocity zone and why is it important for plate tectonics?
15. Where and how is Earth's magnetic field thought to be generated?
16. How does the presence of anomalously hot (low seismic wave velocity) material below the oceanic ridges of the Pacific support the plate tectonic model?
17. Why are the continents underlain by cold mantle?
18. Compare convection styles in the upper and lower mantle.
19. Trace the path of a hypothetical slab of oceanic lithosphere from its production at a midocean ridge to its ultimate demise.

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MULTIMEDIA TOOLS

Earth's Dynamic Systems Website
The Companion Website at **www.prenhall.com/hamblin** provides you with an on-line study guide and additional resources for each chapter, including:

- On-line Quizzes (Chapter Review, Visualizing Geology, Quick Review, Vocabulary Flash Cards) with instant feedback
- Quantitative Problems
- Critical Thinking Exercises
- Web Resources

Earth's Dynamic Systems CD
Examine the CD that came with your text. It is designed to help you visualize and thus understand the concepts in this chapter. It includes:

- Animations that show how earthquake waves propagate
- Models of mantle convection
- Slide shows with more examples of earthquake damage
- A direct link to the Companion Website