CHAPTER

10

Earthquakes and the Earth's Structure

Continents glide continuously around the globe, but we cannot feel the motion because it is too slow. Occasionally, however, the Earth trembles noticeably. The ground rises and falls and undulates back and forth, as if it were an ocean wave. Buildings topple, bridges fail, roadways and pipelines snap. An **earthquake** is a sudden motion or trembling of the Earth caused by the abrupt release of energy that is stored in rocks.

Before the plate tectonics theory was developed, geologists recognized that earthquakes occur frequently in some regions and infrequently in others, but they did not understand why. Modern geologists know that most earthquakes occur along plate boundaries, where huge tectonic plates separate, converge, or slip past one another.

The January 1995 Kobe earthquake destroyed this portion of the Kobe-Osaka Highway in western Japan and killed nearly 5000 people. (Atsushi Tsukada/AP Wide World)

10.1 WHAT IS AN EARTHQUAKE?

How do rocks store energy, and why do they suddenly release it as an earthquake?

Stress is a force exerted against an object.[1] You stress a cable when you use it to tow a neighbor's car. Tectonic forces stress rocks. The movement of lithospheric plates is the most common source of tectonic stress.

When an object is stressed, it changes volume and shape. If a solid object is stressed slowly, it first deforms in an elastic manner: When the stress is removed, the object springs back to its original size and shape. A rubber band exhibits elastic deformation. The energy used to stretch a rubber band is stored in the elongated rubber. When the stress is removed, the rubber band springs back to its initial size and shape and releases the stored energy. Rocks also deform elastically when tectonic stress is applied (Fig. 10–1).

Every rock has a limit beyond which it cannot deform elastically. Under certain conditions, when its elastic limit is exceeded, a rock continues to deform like putty. This behavior is called **plastic deformation**. A rock that has deformed plastically retains its new shape when the stress is released (Fig. 10–2). Earthquakes do not occur when rocks deform plastically.

Under other conditions, an elastically stressed rock may rupture by **brittle fracture** (Fig. 10–3). The fracture releases the elastic energy, and the surrounding rock springs back to its original shape. This rapid motion creates vibrations that travel through the Earth and are felt as an earthquake.

Earthquakes also occur when rock slips along previously established faults. Tectonic plate boundaries are huge faults that have moved many times in the past and will move again in the future (Fig. 10–4).

Although tectonic plates move at rates between 1 and 16 centimeters per year, friction prevents the plates from slipping past one another continuously. As a result, rock near a plate boundary stretches or compresses. When its accumulated elastic energy overcomes the friction that binds plates together, the rock suddenly slips along

[1]More precisely, stress is defined as force per unit area and is measured in units of newtons per square meter (N/m^2).

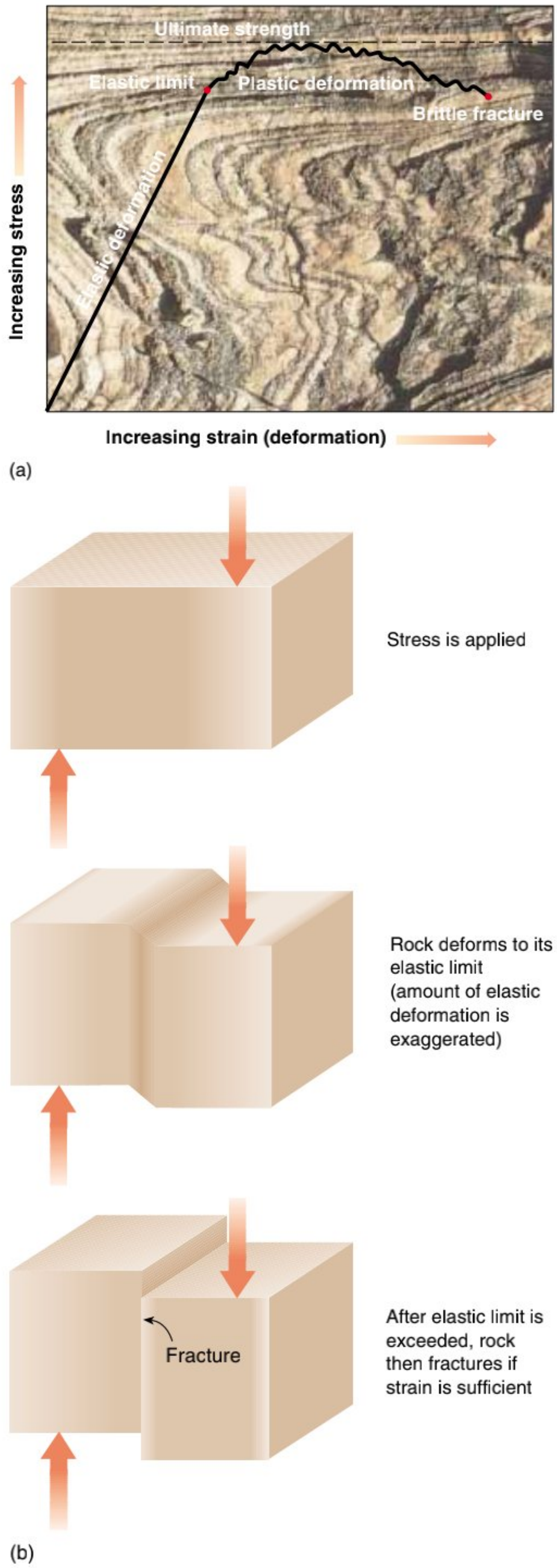
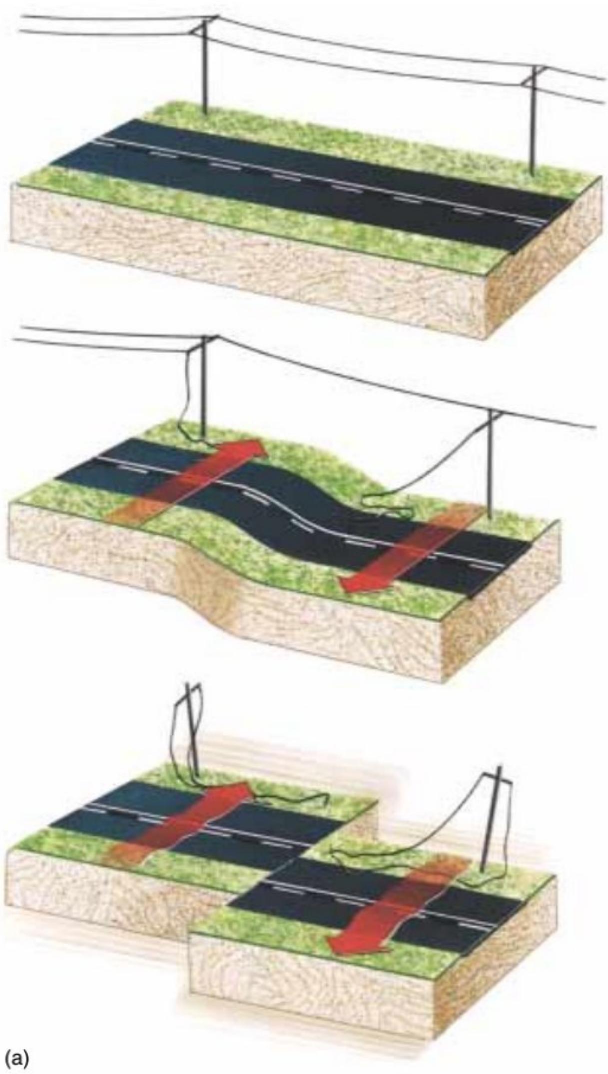


Figure 10–1 The behavior of a rock as stress increases in graphical form (a), in schematic form (b). At first the rock deforms by elastic deformation in which the amount of deformation is directly proportional to the amount of stress. Beyond the elastic limit, the rock deforms plastically and a small amount of additional stress causes a large increase in distortion. Finally, at the yield point, the rock fractures. Many stressed rocks deform elastically and then rupture, with little or no intermediate plastic deformation. The factors that control rock behavior are discussed further in Chapter 12.

Figure 10–2 Rocks may deform plastically when stressed. Plastic deformation contorted the layering in metamorphic rocks in Connecticut.



(a)

(b)

Figure 10–3 (a) A rock stores elastic energy when it is distorted by a tectonic force. When the rock fractures, it snaps back to its original shape, creating an earthquake. In the process, the rock moves along the fracture. (b) Moving rock and soil fractured and displaced this roadway during the Loma Prieta earthquake in 1989.

Figure 10–4 California's San Andreas fault, the source of many earthquakes, is the boundary between the Pacific plate, on the left in this photo, and the North American plate, on the right. (R.E. Wallace/USGS)

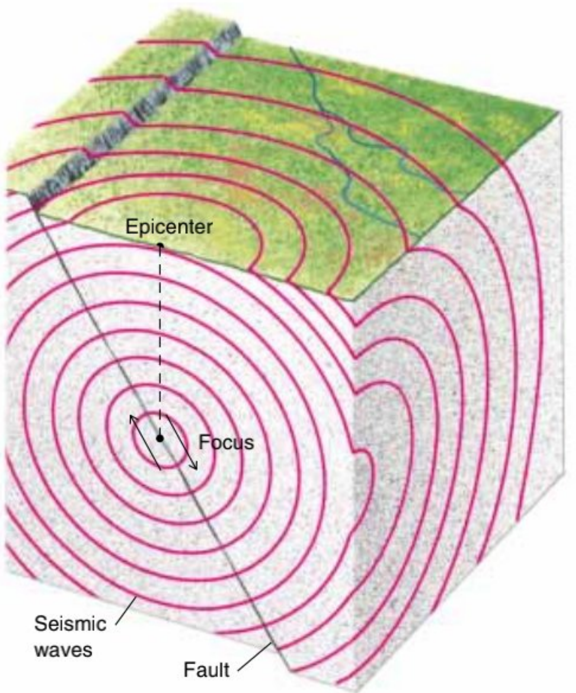


Figure 10–5 Body waves radiate outward from the focus of an earthquake.

the fault, generating an earthquake. The rocks may move from a few centimeters to a few meters, depending on the amount of stored energy.

10.2 EARTHQUAKE WAVES

If you have ever bought a watermelon, you know the challenge of picking out a ripe, juicy one without being able to look inside. One trick is to tap the melon gently with your knuckle. If you hear a sharp, clean sound, it is probably ripe; a dull thud indicates that it may be overripe and mushy. The watermelon illustrates two points that can be applied to the Earth: (1) The energy of your tap travels through the melon, and (2) the nature of the melon's interior affects the quality of the sound.

A wave transmits energy from one place to another. Thus, a drumbeat travels through air as a sequence of waves, the Sun's heat travels to Earth as waves, and a tap travels through a watermelon in waves. Waves that travel through rock are called **seismic waves**. Earthquakes and explosions produce seismic waves. **Seismology** is the study of earthquakes and the nature of the Earth's interior based on evidence from seismic waves.

An earthquake produces several different types of seismic waves. **Body waves** travel through the Earth's interior. They radiate from the initial rupture point of an earthquake, called the **focus** (Fig. 10–5).

The point on the Earth's surface directly above the focus is the **epicenter**. During an earthquake, body waves carry some of the energy from the focus to the surface. **Surface waves** then radiate from the epicenter along the Earth's surface. Although the mechanism is different, surface waves undulate across the ground like the waves that ripple across the water after you throw a rock into a calm lake.

BODY WAVES

Two main types of body waves travel through the Earth's interior. A **P wave** (also called a compressional wave) is an elastic wave that causes alternate compression and expansion of the rock (Fig. 10–6). Consider a long spring such as the popular Slinky™ toy. If you stretch a Slinky

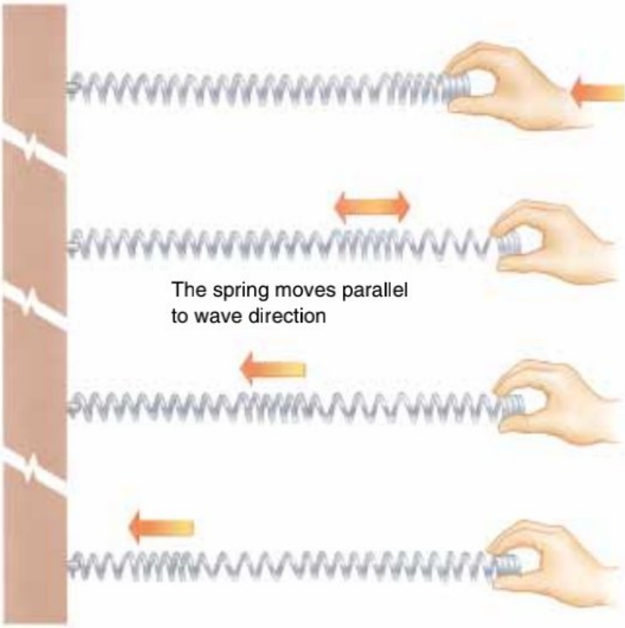


Figure 10–6 Model of a P wave; a compressional wave. The wave is propagated along the spring. The particles in the spring move parallel to the direction of wave propagation.

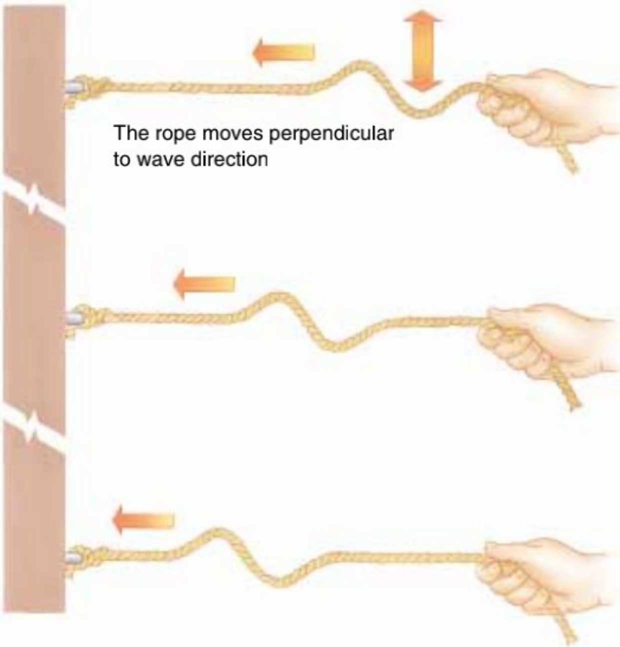


Figure 10–7 Model of an S wave; a shear wave. The wave is propagated along the rope. The particles in the rope move perpendicular to the direction of wave propagation.

and strike one end, a compressional wave travels along its length. P waves travel through air, liquid, and solid material. Next time you take a bath, immerse your head until your ears are under water and listen as you tap the sides of the tub with your knuckles. You are hearing P waves.

P waves travel at speeds between 4 and 7 kilometers per second in the Earth's crust and at about 8 kilometers per second in the uppermost mantle. As a comparison, the speed of sound in air is only 0.34 kilometer per second, and the fastest jet fighters fly at about 0.85 kilometer per second. P waves are called primary waves because they are so fast that they are the first waves to reach an observer.

A second type of body wave, called an **S wave**, is a **shear wave**. An S wave can be illustrated by tying a rope to a wall, holding the end, and giving it a sharp up-and-down jerk (Fig. 10–7). Although the wave travels parallel to the rope, the individual particles in the rope move at right angles to the rope length. A similar motion in an S wave produces shear stress in rock and gives the wave its name. S waves are slower than P waves and travel at speeds between 3 and 4 kilometers per second in the crust. As a result, S waves arrive after P waves and are the secondary waves to reach an observer.

Unlike P waves, S waves move only through solids. Because molecules in liquids and gases are only weakly bound to one another, they slip past each other and thus cannot transmit a shear wave.

SURFACE WAVES

Surface waves travel more slowly than body waves. Two types of surface waves occur simultaneously in the Earth (Fig. 10–8). A **Rayleigh wave** moves with an up-and-down rolling motion like an ocean wave. **Love waves** produce a side-to-side vibration. Thus, during an earth-

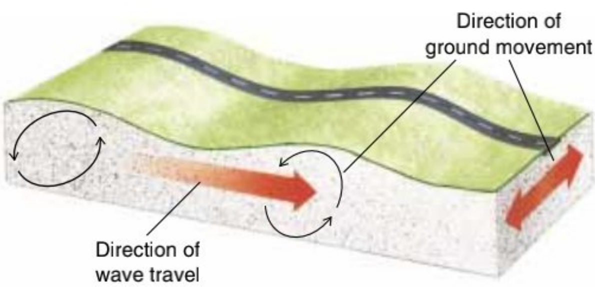


Figure 10–8 Surface waves. Surface motion includes up-and-down movement like that of an ocean wave and also a side-to-side sway.

Figure 10–9 Surface waves cause a large proportion of earthquake damage. Collapse of Interstate Highway 880 during the 1989 Loma Prieta, California, earthquake. (Paul Scott/Sygma)

quake, the Earth's surface rolls like ocean waves and writhes from side to side like a snake (Fig. 10–9).

MEASUREMENT OF SEISMIC WAVES

A **seismograph** is a device that records seismic waves. To understand how a seismograph works, consider the act of writing a letter while riding in an airplane. If the plane hits turbulence, inertia keeps your hand relatively stationary as the plane moves back and forth beneath it, and your handwriting becomes erratic.

Early seismographs worked on the same principle. A weight was suspended from a spring. A pen attached to the weight was aimed at the zero mark on a piece of graph paper (Fig. 10–10). The graph paper was mounted on a rotary drum that was attached firmly to bedrock. During an earthquake, the graph paper jiggled up and down, but inertia kept the pen stationary. As a result, the paper moved up and down beneath the pen. The rotating drum recorded earthquake motion over time. This record of Earth vibration is called a **seismogram** (Fig. 10–11). Modern seismographs use electronic motion detectors which transmit the signal to a computer.

MEASUREMENT OF EARTHQUAKE STRENGTH

Over the past century, geologists have devised several scales to express the size of an earthquake. Before seismographs were in common use, earthquakes were evaluated on the basis of structural damage. One that destroyed many buildings was rated as more intense than one that destroyed only a few. This system did not accurately measure the energy released by a quake, however, because structural damage depends on distance from the focus, the rock or soil beneath the structure, and the quality of construction.

In 1935 Charles Richter devised the **Richter scale** to express earthquake magnitude. Richter magnitude is

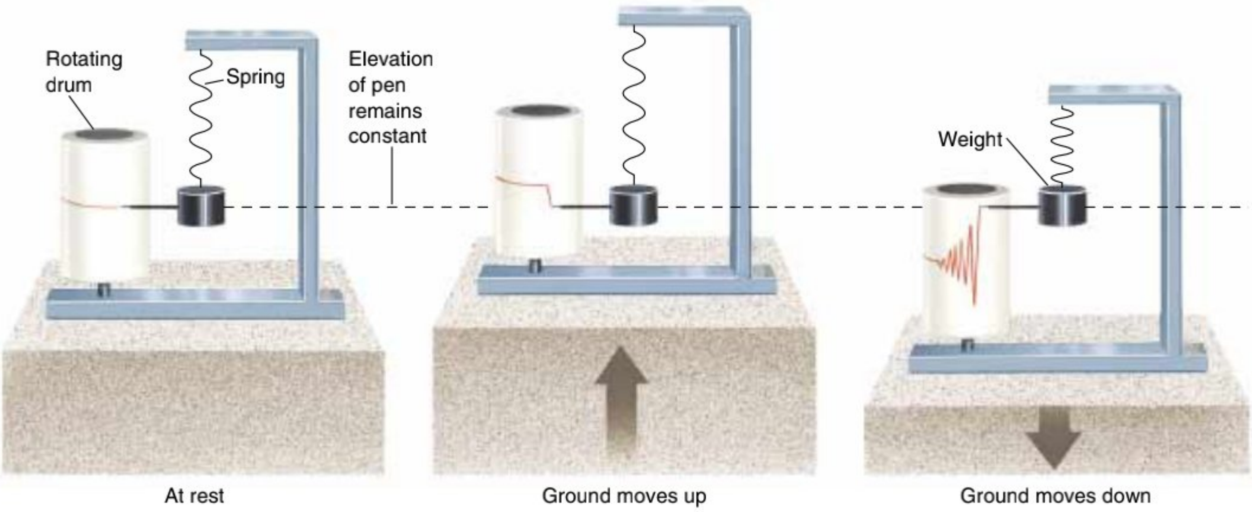


Figure 10–10 A seismograph records ground motion during an earthquake. When the ground is stationary, the pen draws a straight line across the rotating drum. When the ground rises abruptly during an earthquake, it carries the drum up with it. But the spring stretches, so the weight and pen hardly move. Therefore, the pen marks a line lower on the drum. Conversely, when the ground sinks, the pen marks a line higher on the drum. During an earthquake, the pen traces a jagged line as the drum rises and falls.

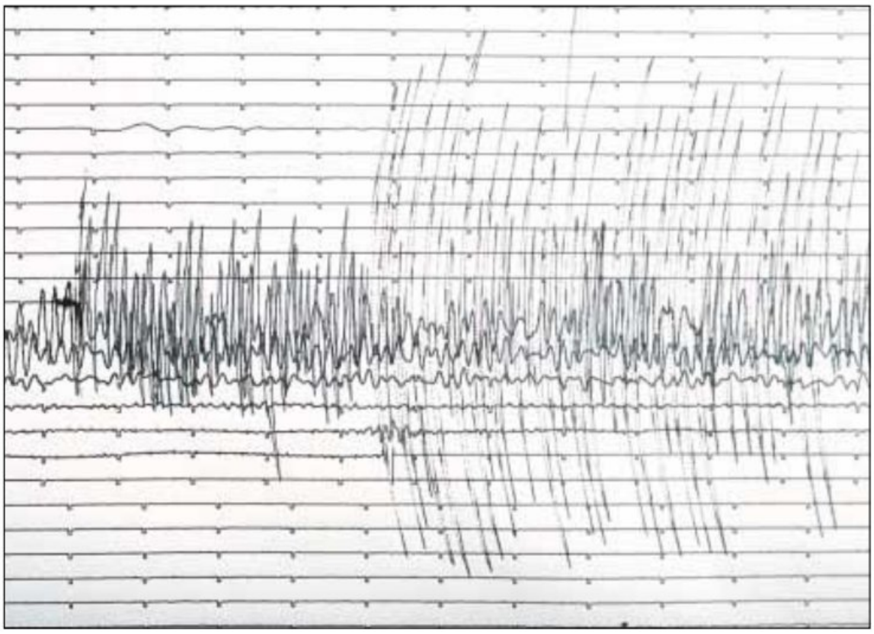


Figure 10–11 This seismogram records north-south ground movements during the October 1989 Loma Prieta earthquake. (Russell D. Curtis/USGS)

calculated from the height of the largest earthquake body wave recorded on a specific type of seismograph. The Richter scale is more quantitative than earlier intensity scales, but it is not a precise measure of earthquake energy. A sharp, quick jolt would register as a high peak on a Richter seismograph, but a very large earthquake can shake the ground for a long time without generating extremely high peaks. In this way, a great earthquake can release a huge amount of energy that is not reflected in the height of a single peak, and thus is not adequately expressed by Richter magnitude.

Modern equipment and methods enable seismologists to measure the amount of slip and the surface area of a fault that moved during a quake. The product of these two values allows them to calculate the **moment magnitude**. Most seismologists now use moment magnitude rather than Richter magnitude because it more closely reflects the total amount of energy released during an earthquake. An earthquake with a moment magnitude of 6.5 has an energy of about 10^{25} (10 followed by 25 zeros) ergs.[2] The atomic bomb dropped on the Japanese city of Hiroshima at the end of World War II released about that much energy.

[2]An erg is the standard unit of energy in scientific usage. One erg is a small amount of energy. Approximately 3×10^{12} ergs are needed to light a 100-watt light bulb for 1 hour. However, 10^{25} is a very large number, and 10^{25} ergs represents a considerable amount of energy.

On both the moment magnitude and Richter scales, the energy of the quake increases by about a factor of 30 for each successive increment on the scale. Thus, a magnitude 6 earthquake releases roughly 30 times more energy than a magnitude 5 earthquake.

The largest possible earthquake is determined by the strength of rocks. A strong rock can store more elastic energy before it fractures than a weak rock. The largest earthquakes ever measured had magnitudes of 8.5 to 8.7, about 900 times greater than the energy released by the Hiroshima bomb.

LOCATING THE SOURCE OF AN EARTHQUAKE

If you have ever watched an electrical storm, you may have used a simple technique for estimating the distance between you and the place where the lightning strikes. After the flash of a lightning bolt, count the seconds that pass before you hear thunder. Although the electrical discharge produces thunder and lightning simultaneously, light travels much faster than sound. Therefore, light reaches you virtually instantaneously, whereas sound travels much more slowly, at 340 meters per second. If the time interval between the flash and the thunder is 1 second, then the lightning struck 340 meters away and was very close.

The same principle is used to determine the distance from a recording station to both the epicenter and focus of an earthquake. Recall that P waves travel faster than

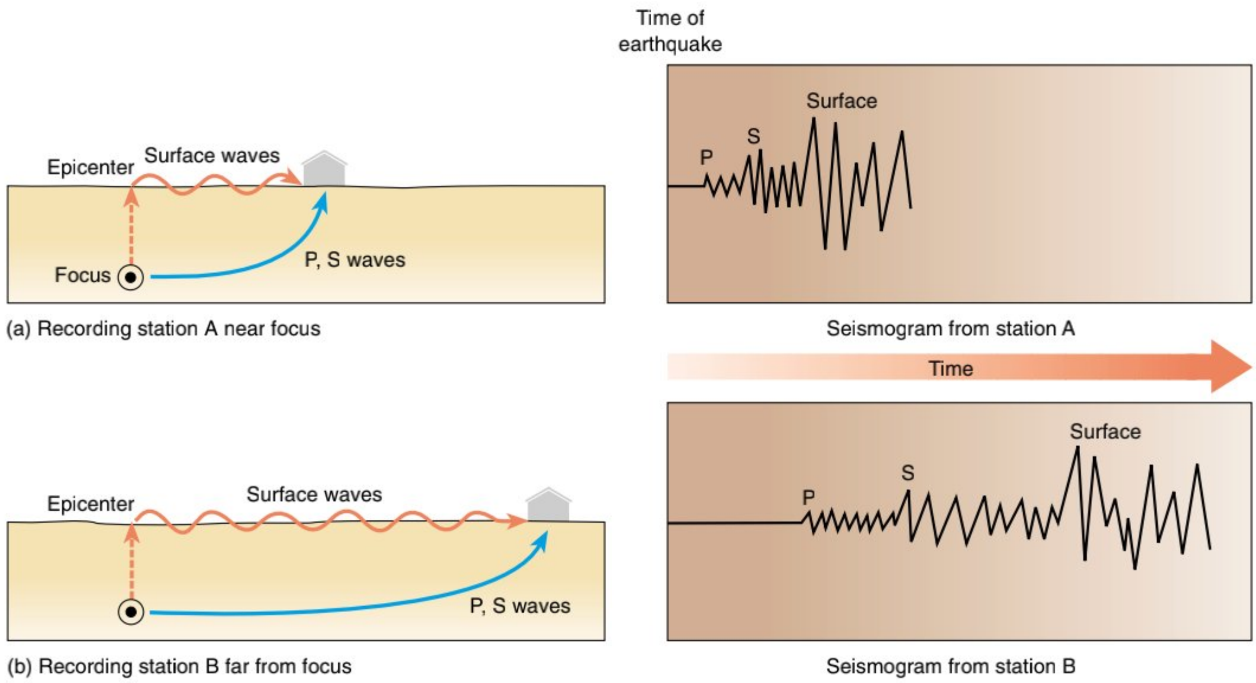


Figure 10–12 The time intervals between arrivals of P, S, and L waves at a recording station increase with distance from the focus of an earthquake.

S waves and that surface waves are slower yet. If a seismograph is located close to an earthquake epicenter, the different waves will arrive in rapid succession for the same reason that the thunder and lightning come close together when a storm is close. On the other hand, if a seismograph is located far from the epicenter, the S waves arrive at correspondingly later times after the P waves arrive, and the surface waves are even farther behind, as shown in Figure 10–12.

Geologists use a **time-travel curve** to calculate the distance between an earthquake epicenter and a seismograph. To make a time-travel curve, a number of seismic stations at different locations record the times of arrival of seismic waves from an earthquake with a known epicenter and occurrence time. Then a graph such as Figure 10–13 is drawn. This graph can then be used to measure the distance between a recording station and an earthquake whose epicenter is unknown.

Time-travel curves were first constructed from data obtained from natural earthquakes. However, scientists do not always know precisely when and where an earthquake occurred. In the 1950s and 1960s, geologists studied seismic waves from atomic bomb tests to improve the time-travel curves because they knew both the locations and timing of the explosions.

Figure 10–13 shows us that if the first P wave arrives 3 minutes before the first S wave, the recording sta-

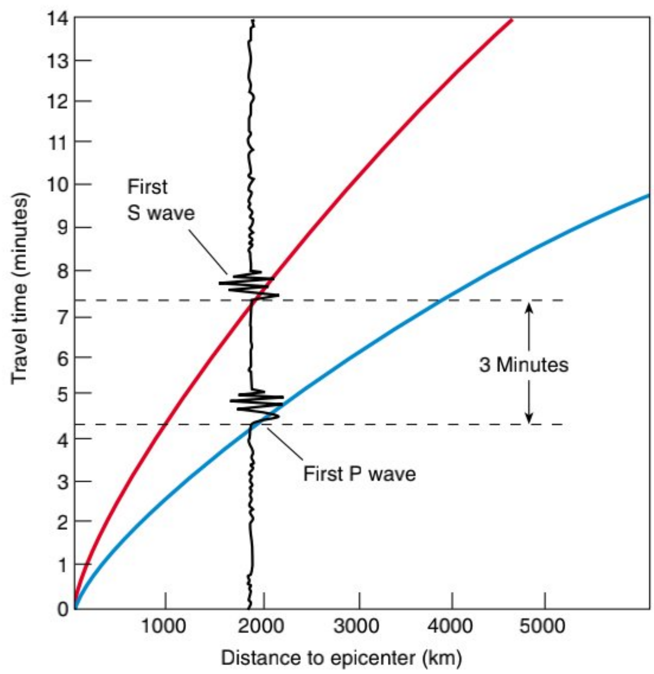


Figure 10–13 A time-travel curve. With this graph you can calculate the distance from a seismic station to the source of an earthquake. In the example shown, a 3-minute delay between the first arrivals of P waves and S waves corresponds to an earthquake with an epicenter 1900 kilometers from the seismic station.

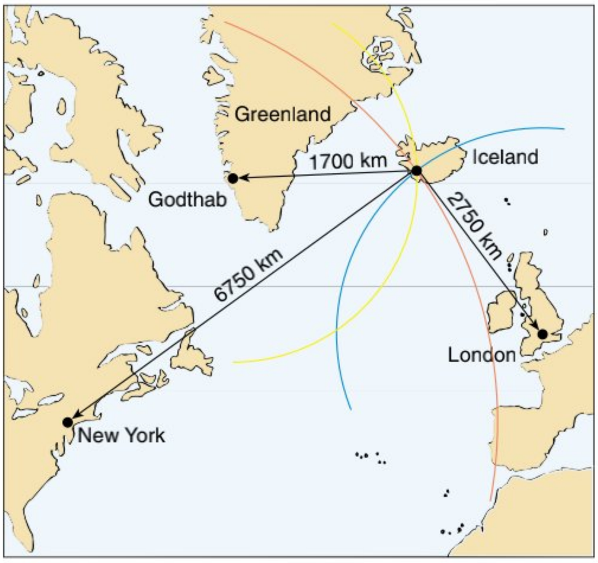


Figure 10–14 Locating an earthquake. The distance from each of three seismic stations to the earthquake is determined from time-travel curves. The three arcs are drawn. They intersect at only one point, which is the epicenter of the earthquake.

tion is about 1900 kilometers from the epicenter. But this distance does not indicate whether the earthquake originated to the north, south, east, or west. To pinpoint the location of an earthquake, geologists compare data from three or more recording stations. If a seismic station in New York City records an earthquake with an epicenter 6750 kilometers away, geologists know that the epicenter lies somewhere on a circle 6750 kilometers from New York City (Fig. 10–14). The same epicenter is reported to be 2750 kilometers from a seismic station in London and 1700 kilometers from one in Godthab, Greenland. If one circle is drawn for each recording station, the arcs intersect at the epicenter of the quake.

10.3 EARTHQUAKE DAMAGE

Large earthquakes can displace rock and alter the Earth's surface (Fig. 10–15). The New Madrid, Missouri, earthquake of 1811 changed the course of the Mississippi River. During the 1964 Alaskan earthquake, some beaches rose 12 meters, leaving harbors high and dry, while other beaches sank 2 meters, causing coastal flooding.

Most earthquake fatalities and injuries occur when falling structures crush people. Structural damage, injury, and death depend on the magnitude of the quake, its proximity to population centers, rock and soil types, topography, and the quality of construction in the region.

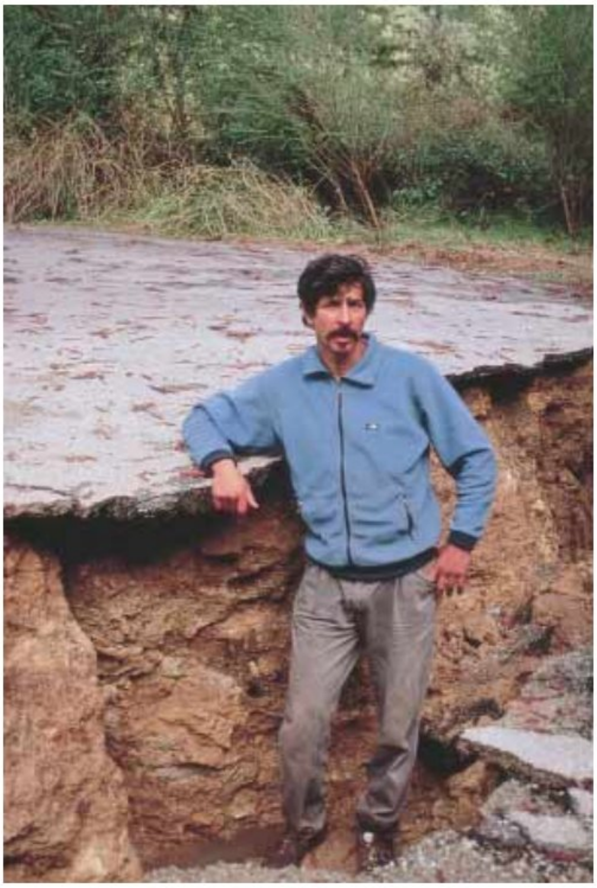


Figure 10–15 Jon Turk stands in an area of permanent ground displacement caused by the Loma Prieta, California, earthquake of 1989. (Christine Seashore)

HOW ROCK AND SOIL INFLUENCE EARTHQUAKE DAMAGE

In many regions, bedrock lies at or near the Earth's surface and buildings are anchored directly to the rock. Bedrock vibrates during an earthquake and buildings may fail if the motion is violent enough. However, most bedrock returns to its original shape when the earthquake is over, so if structures can withstand the shaking, they will survive. Thus, bedrock forms a desirable foundation in earthquake hazard areas.

In many places, structures are built on sand, clay, or silt. Sandy sediment and soil commonly settle during an earthquake. This displacement tilts buildings, breaks pipelines and roadways, and fractures dams. To avert structural failure in such soils, engineers drive steel or concrete pilings through the sand to the bedrock below. These pilings anchor and support the structures even if the ground beneath them settles.

Mexico City provides one example of what can happen to clay-rich soils during an earthquake. The city is built on a high plateau ringed by even higher mountains. When the Spaniards invaded central Mexico, lakes dotted the plateau and the Aztec capital lay on an island at the end of a long causeway in one of the lakes. Over the following centuries, European settlers drained the lake and built the modern city on the water-soaked, clay-rich lake-bed sediment. On September 19, 1985, an earthquake with a magnitude of 8.1 struck about 500 kilometers west of the city. Seismic waves shook the wet clay beneath the city and reflected back and forth between the bedrock sides and bottom of the basin, just as waves in a bowl of Jell-O™ bounce off the side and bottom of the bowl. The reflections amplified the waves, which destroyed more than 500 buildings and killed between 8000 and 10,000 people (Fig. 10–16). Meanwhile, there was comparatively little damage in Acapulco, which was much closer to the epicenter but is built on bedrock.

If soil is saturated with water, the sudden shock of an earthquake can cause the grains to shift closer together, expelling some of the water. When this occurs, increased stress is transferred to the pore water, and the pore pressure may rise sufficiently to suspend the grains in the water. In this case, the soil loses its shear strength and behaves as a fluid. This process is called **liquefaction.** When soils liquefy on a hillside, the slurry flows downslope, carrying structures along with it. During the 1964 earthquake near Anchorage, Alaska, a clay-rich bluff 2.8 kilometers long, 300 meters wide, and 22 meters high liquefied. The slurry carried houses into the ocean and buried some so deeply that bodies were never recovered.

CONSTRUCTION DESIGN AND EARTHQUAKE DAMAGE

A magnitude 6.4 earthquake struck central India in 1993, killing 30,000 people. In contrast, the 1994 magnitude 6.6 quake in Northridge (near Los Angeles) killed only 55. The tremendous mortality in India occurred because buildings were not engineered to withstand earthquakes.

Figure 10–16 The 1985 Mexico City earthquake had a magnitude of 8.1 and killed between 8000 and 10,000 people. Earthquake waves amplified within the soil so that the effects were greater in Mexico City than they were in Acapulco, which was much closer to the epicenter. (AP/Wide World Photos)

Some common framing materials used in buildings, such as wood and steel, bend and sway during an earthquake but resist failure. However, brick, stone, concrete, adobe (dried mud), and other masonry products are brittle and likely to fail during an earthquake. Although masonry can be reinforced with steel, in many regions of the world people cannot afford such reinforcement.

FIRE

Earthquakes commonly rupture buried gas pipes and electrical wires, leading to fire, explosions, and electrocutions (Fig. 10–17). Water pipes may also break, so fire fighters cannot fight the blazes effectively. Most of the damage from the 1906 San Francisco earthquake resulted from fires.

LANDSLIDES

Landslides are common when the Earth trembles. Earthquake-related landslides are discussed in more detail in Chapter 13.

TSUNAMIS

When an earthquake occurs beneath the sea, part of the sea floor rises or falls (Fig. 10–18). Water is displaced in response to the rock movement, forming a wave. Sea waves produced by an earthquake are often called tidal waves, but they have nothing to do with tides. Therefore, geologists call them by their Japanese name, **tsunami**.

In the open sea, a tsunami is so flat that it is barely detectable. Typically, the crest may be only 1 to 3 meters high, and successive crests may be more than 100 to 150 kilometers apart. However, a tsunami may travel at 750 kilometers per hour. When the wave approaches the shallow water near shore, the base of the wave drags against the bottom and the water stacks up, increasing the height of the wave. The rising wall of water then flows inland. A tsunami can flood the land for as long as 5 to 10 minutes.

Figure 10–17 Ruptured gas and electric lines often cause fires during earthquakes in urban areas. This blaze followed the 1989 San Francisco earthquake. (Lysaght/Gamma Liaison)

10.4 EARTHQUAKES AND TECTONIC PLATE BOUNDARIES

Although many faults are located within tectonic plates, the largest and most active faults are the boundaries between tectonic plates. Therefore, as Figure 10–19 shows, earthquakes occur most frequently along plate boundaries.

EARTHQUAKES AT A TRANSFORM PLATE BOUNDARY: THE SAN ANDREAS FAULT ZONE

The populous region from San Francisco to San Diego straddles the San Andreas fault zone, which is a transform boundary between the Pacific plate and the North American plate (Fig. 10–20). The fault itself is vertical and the rocks on opposite sides move horizontally. A fault of this type is called a **strike–slip fault** (Fig. 10–21). Plate motion stresses rock adjacent to the fault, generating numerous smaller faults, shown by the solid lines in the figure. The San Andreas fault and its satellites form a broad region called the **San Andreas fault zone**.

In the past few centuries, hundreds of thousands of earthquakes have occurred in this zone. Geologists of the United States Geological Survey recorded 10,000 earthquakes in 1984 alone, although most could be detected only with seismographs. Severe quakes occur periodically. One shook Los Angeles in 1857, and another destroyed San Francisco in 1906. A large quake in 1989 occurred south of San Francisco, and another rocked Northridge, just outside Los Angeles, in January 1994. The fact that the San Andreas fault zone is part of a major plate boundary tells us that more earthquakes are inevitable.

The plates move past one another in three different ways along different segments of the San Andreas fault zone:

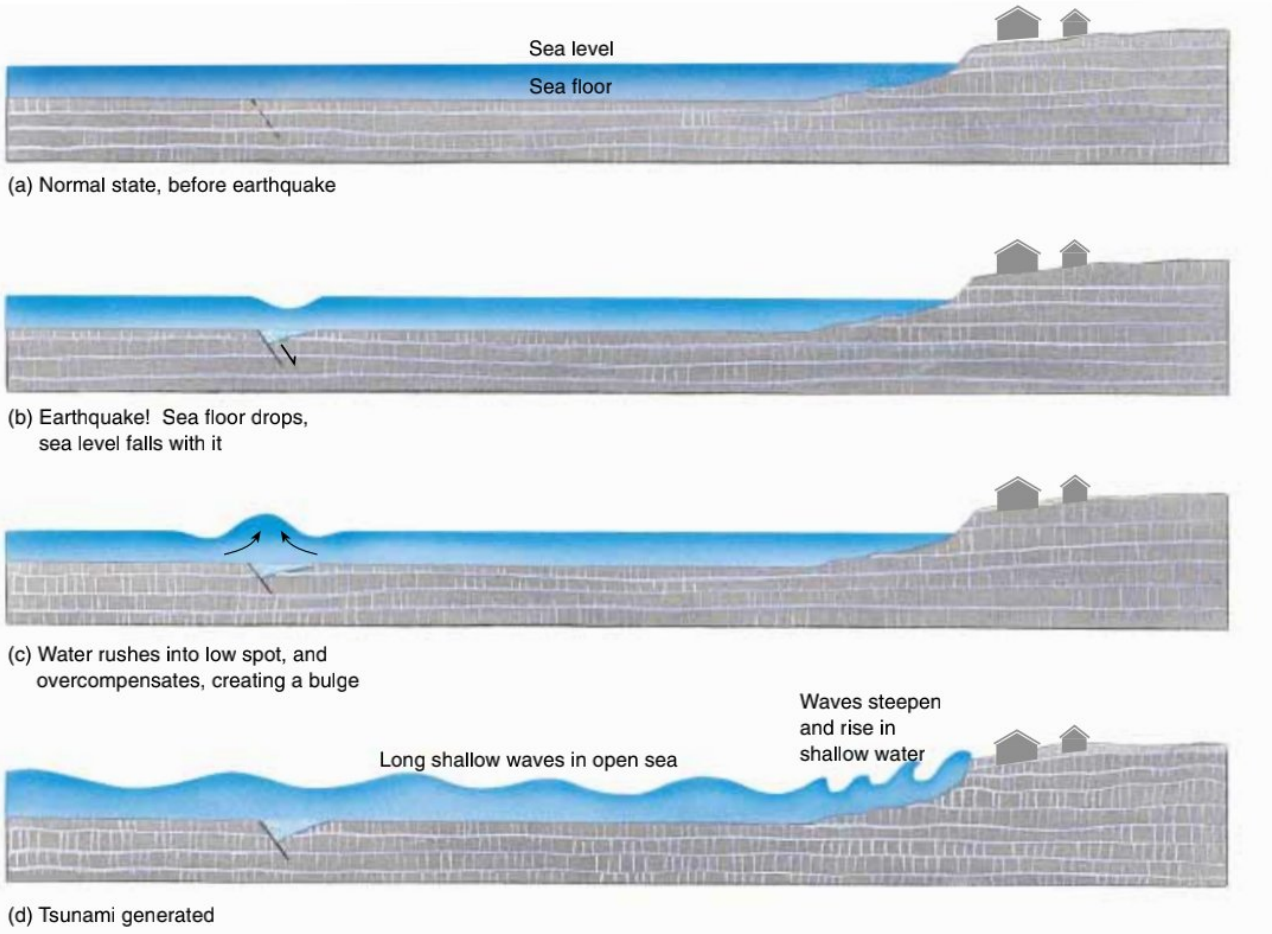


Figure 10–18 Formation of a tsunami. If a portion of the sea floor drops during an earthquake, the sea level falls with it. Water rushes into the low spot and overcompensates, creating a bulge. The long, shallow waves build up when they reach land.

1. Along some portions of the fault, rocks slip past one another at a continuous, snail-like pace called **fault creep**. The movement occurs without violent and destructive earthquakes because the rocks move continuously and slowly.
2. In other segments of the fault, the plates pass one another in a series of small hops, causing numerous small, nondamaging earthquakes.
3. Along the remaining portions of the fault, friction prevents slippage of the fault although the plates continue to move past one another. In this case, rock near the fault deforms and stores elastic energy. Because the plates move past one another at 3.5 centimeters per year, 3.5 meters of elastic deformation accumulate over a period of 100 years. When the accumulated elastic energy exceeds friction, the rock suddenly slips along the fault and snaps back to its original shape, producing a large, destructive earthquake.

CASE STUDY The Northridge Earthquake of January 1994

In January 1994, a magnitude 6.6 earthquake struck Northridge in the San Fernando Valley just north of Los Angeles (Fig. 10–22). Fifty-five people died and property damage was estimated at $8 billion.

As explained earlier, the San Andreas fault is a strike–slip fault that is part of a transform plate boundary. In the mid-1980s, geologists also discovered buried thrust faults in Southern California. A **thrust fault** is one in which rock on one side of the fault slides up and over the rock on the other side (Fig. 10–23). While the San

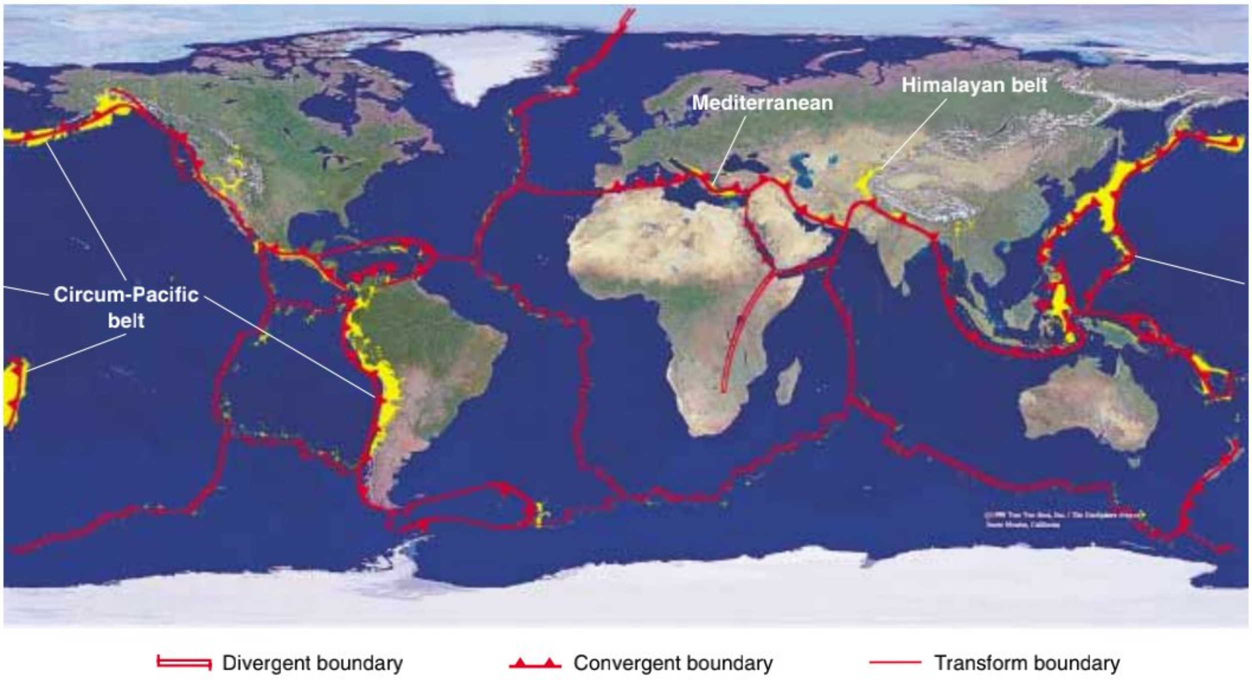


Figure 10–19 The Earth's major earthquake zones coincide with tectonic plate boundaries. Each yellow dot represents an earthquake that occurred between 1961 and 1967. (Tom Van Sant, Geosphere Project)

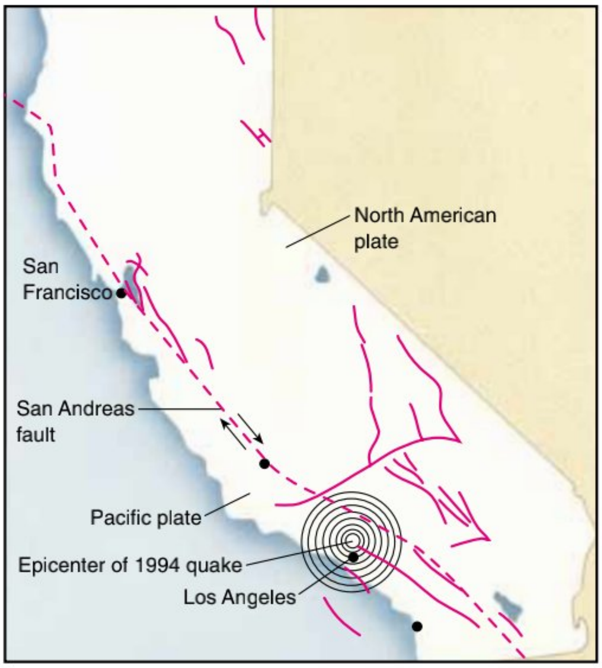


Figure 10–20 Faults and earthquakes in California. The dashed line is the San Andreas fault, and solid lines are related faults. Blue dots are epicenters of recent earthquakes. (Redrawn from USGS data)

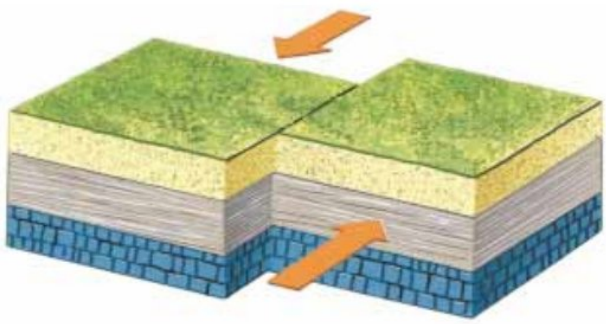


Figure 10–21 A strike–slip fault is vertical, and the rock on opposite sides of the fracture moves horizontally.

Andreas fault lies east of metropolitan Los Angeles, the Santa Monica thrust fault lies directly under the city (Fig. 10–24). A major quake on one of these faults can be more disastrous than one on the main San Andreas fault. The Northridge earthquake occurred when one of these thrust faults slipped. According to geologist James Dolan of the California Institute of Technology, "There's a whole seismic hazard from buried thrust faults that we didn't even appreciate until six years ago."

The existence of thrust faults indicates that in Southern California, both the direction of tectonic forces

Figure 10–22 The January 1994 Northridge, California, earthquake killed 55 people and caused eight billion dollars in damage. (Earthquake Engineering Institute)

and the manner in which stress is relieved are more complicated than a simple model expresses. Although many disastrous and expensive earthquakes have shaken southern California in the past few decades, none of them has been the Big One that seismologists still fear.

EARTHQUAKES AT SUBDUCTION ZONES

In a subduction zone, a relatively cold, rigid lithospheric plate dives beneath another plate and slowly sinks into the mantle. In most places, the subducting plate sinks with intermittent slips and jerks, giving rise to numerous earthquakes. The earthquakes concentrate along the upper part of the sinking plate, where it scrapes past the opposing plate (Fig. 10–25). This earthquake zone is called the **Benioff zone**, after the geologist who first recognized it. Many of the world's strongest earthquakes occur in subduction zones.

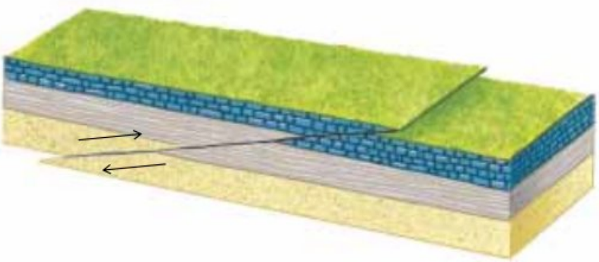


Figure 10–23 A thrust fault is a low-angle fault in which rock on one side of the fault slides up and over rock on the other side.

CASE STUDY Earthquake Activity in the Pacific Northwest

The small Juan de Fuca plate, which lies off the coasts of Oregon, Washington, and southern British Columbia, is diving beneath North America at a rate of 3 to 4 centimeters per year. Thus, the region should experience subduction zone earthquakes. Yet although small earthquakes occasionally shake the Pacific Northwest, no large ones have occurred in the past 150 to 200 years.

Why are earthquakes relatively uncommon in the Pacific Northwest? Geologists have suggested two possible answers to that question. Subduction may be occurring slowly and continuously by fault creep. If this is the case, elastic energy would not accumulate in nearby rocks, and strong earthquakes would be unlikely. Alternatively, rocks along the fault may be locked together by friction, accumulating a huge amount of elastic energy that will be released in a giant, destructive quake sometime in the future.

Recently geologists have discovered probable evidence of great prehistoric earthquakes in the Pacific Northwest. A major coastal earthquake commonly creates violent sea waves, which deposit a layer of sand along the coast. Geologists have found several such sand layers, each burying a layer of peat and mud that accumulated in coastal swamps during the quiet intervals between earthquakes. In addition, they have found submarine landslide deposits lying on the deep sea floor off the coast that formed when earthquakes triggered submarine landslides that carried sand and mud from the coast to the sea floor. These deposits show that 13 major earth-

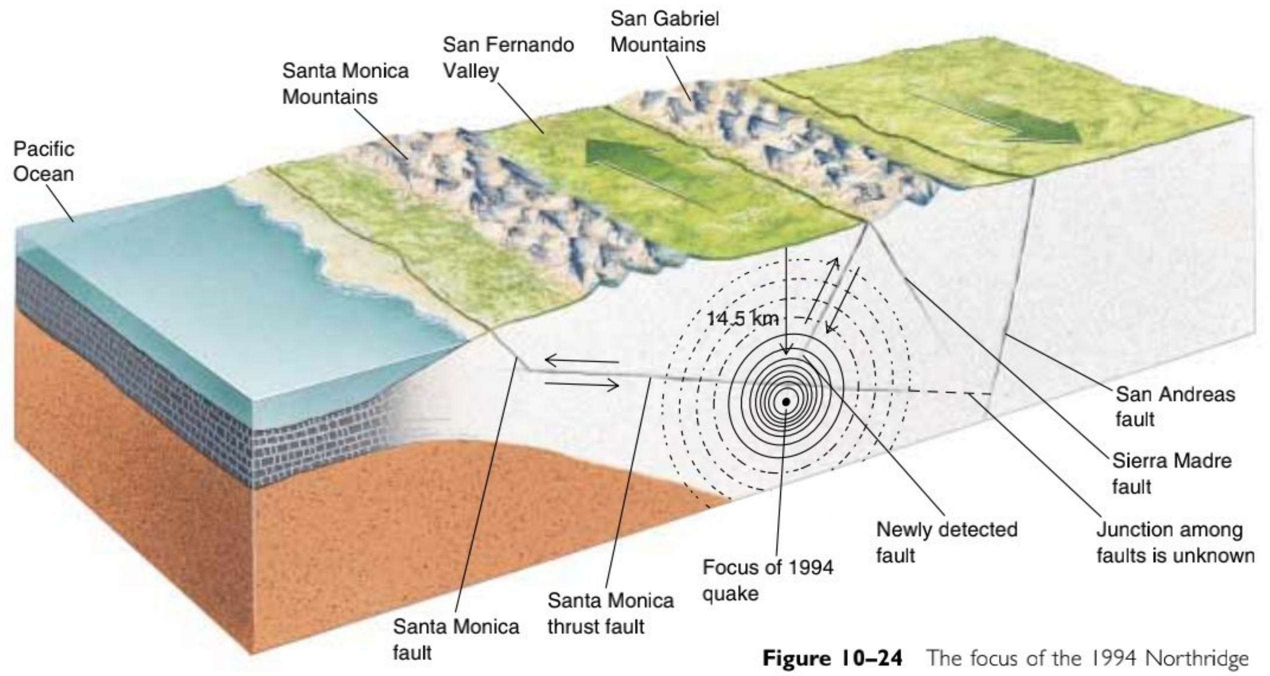


Figure 10–24 The focus of the 1994 Northridge quake was on a previously undetected thrust fault west of the San Andreas fault.

quakes, separated by 300 to 900 years, struck the coast during the past 7700 years. There is also evidence for one major historic earthquake. Oral accounts of the native inhabitants chronicle the loss of a small village in British Columbia and a significant amount of ground shaking in northern California. The same earthquake may have caused a 2-meter-high tsunami in Japan in January of 1700. Thus, many geologists anticipate another major, destructive earthquake in the Pacific Northwest during the next 600 years.

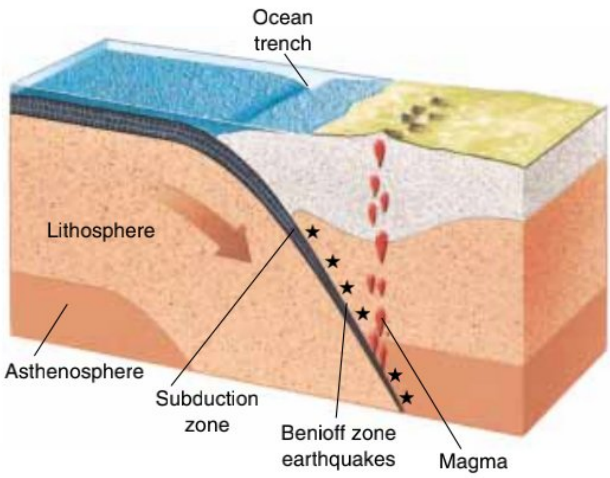


Figure 10–25 A descending lithospheric plate generates magma and earthquakes in a subduction zone. Earthquakes concentrate along the upper portion of the subducting plate, called the Benioff zone.

EARTHQUAKES AT DIVERGENT PLATE BOUNDARIES

Earthquakes frequently shake the mid-oceanic ridge system as a result of faults that form as the two plates separate. Blocks of oceanic crust drop downward along most mid-oceanic ridges, forming a rift valley in the center of the ridge. Only shallow earthquakes occur along the mid-oceanic ridge because here the asthenosphere rises to within 20 to 30 kilometers of the Earth's surface and is too hot and plastic to fracture.

EARTHQUAKES IN PLATE INTERIORS

No major earthquakes have occurred in the central or eastern United States in the past 100 years, and no lithospheric plate boundaries are known in these regions. Therefore, one might infer that earthquake danger is insignificant. However, the largest historical earthquake sequence in the contiguous 48 states occurred near New Madrid, Missouri. In 1811 and 1812, three shocks with estimated magnitudes between 7.3 and 7.8 altered the course of the Mississippi River and rang church bells 1500 kilometers away in Washington, D.C.

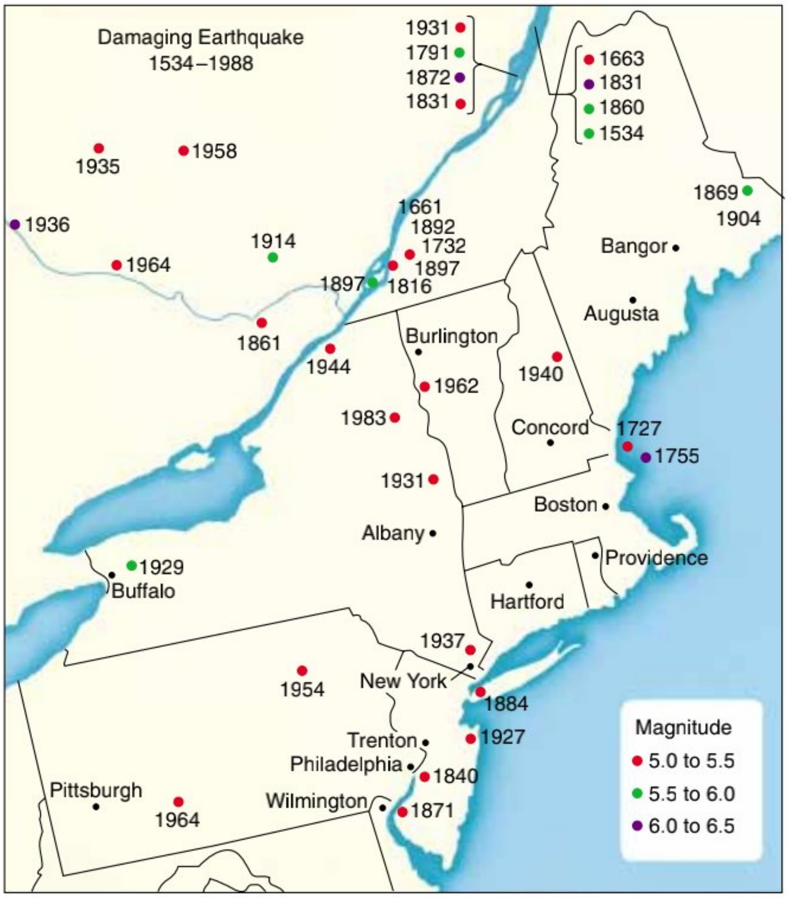


Figure 10–26 Earthquake activity in the northeastern United States and south-eastern Canada between 1534 and 1988. (Redrawn from *Geotimes*, May 1991, p. 6)

Geologists have remeasured distances between old survey bench marks near New York City and found that the marks have moved significantly during the past 50 to 100 years. This motion indicates that the crust in this region is being deformed. Historical reviews show that, although earthquakes are infrequent in the Northeast, they do occur (Fig. 10–26). If a major quake were centered near New York City or Boston today, the consequences could be disastrous.

Earthquakes in plate interiors are not as well understood as those at plate boundaries, but modern research is revealing some clues. The New Madrid region lies in an extinct continental rift zone bounded by deep faults. Although the rift failed to develop into a divergent plate boundary, the deep faults remain a weakness in the lithosphere. As the North America plate glides over the asthenosphere, it may pass over irregularities, or "bumps," in that plastic zone, causing slippage along the deep faults near New Madrid.

Some intraplate earthquakes occur where thick piles of sediment have accumulated on great river deltas such as the Mississippi River delta. The underlying lithosphere cannot support the weight of sediment, and the lithosphere fractures as it settles. Human activity may also induce intraplate earthquakes when changes in the weight of water or rock on the Earth's surface cause the lithosphere to settle. For example, water accumulating in a new reservoir is thought to have caused a magnitude 6 earthquake that killed more than 12,000 people in central India in September 1993.

10.5 EARTHQUAKE PREDICTION

LONG-TERM PREDICTION

Earthquakes occur over and over in the same places because it is easier for rocks to move along an old fracture than for a new fault to form in solid rock. Many of these faults lie along tectonic plate boundaries. Therefore, long-term earthquake prediction recognizes that earthquakes have recurred many times in a specific place and will probably occur there again (Fig. 10–27).

Figure 10–27 This map shows potential earthquake damage in the United States. The predictions are based on records of frequency and magnitude of historical earthquakes. (Ward's Natural Science Establishment, Inc.)

SHORT-TERM PREDICTION

Short-term predictions are forecasts that an earthquake may occur at a specific place and time. Short-term prediction depends on signals that immediately precede an earthquake.

Foreshocks are small earthquakes that precede a large quake by a few seconds to a few weeks. The cause of foreshocks can be explained by a simple analogy. If you try to break a stick by bending it slowly, you may hear a few small cracking sounds just before the final snap. If foreshocks consistently preceded major earthquakes, they would be a reliable tool for short-term prediction. However, foreshocks preceded only about half of a group of recent major earthquakes. At other times, swarms of small shocks were not followed by a large quake.

Another approach to short-term earthquake prediction is to measure changes in the land surface near an active fault zone. Seismologists monitor unusual Earth movements with tiltmeters and laser surveying instruments because distortions of the crust may precede a major earthquake. This method has successfully predicted some earthquakes, but in other instances predicted quakes did not occur or quakes occurred that had not been predicted.

When rock is deformed to near its rupture point prior to an earthquake, microscopic cracks may form. In some cases, the cracks release radon gas previously trapped in rocks and minerals. In addition, the cracks may fill with water and cause the water levels in wells to fluctuate. Furthermore, air-filled cracks do not conduct electricity as well as solid rock, so the electrical conductivity of rock decreases as cracks form.

Chinese scientists reported that, just prior to the 1975 quake in the city of Haicheng, snakes crawled out of their holes, chickens refused to enter their coops, cows broke their halters and ran off, and even well-trained police dogs became restless and refused to obey commands. Some researchers in the United States have attempted to quantify the relationship between animal behavior and earthquakes, but without success.

In January 1975, Chinese geophysicists recorded swarms of foreshocks and unusual bulges near the city of Haicheng, which had a previous history of earthquakes. When the foreshocks became intense on February 1, authorities evacuated portions of the city. The evacuation was completed on the morning of February 4, and in the early evening of the same day, an earthquake destroyed houses, apartments, and factories but caused few deaths.

After that success, geologists hoped that a new era of earthquake prediction had begun. But a year later, Chinese scientists failed to predict an earthquake in the adjacent city of Tangshan. This major quake was not preceded by foreshocks, so no warning was given, and at least 250,000 people died.[3] Over the past few decades, short-term prediction has not been reliable.

10.6 STUDYING THE EARTH'S INTERIOR

Recall from Chapter 2 that the Earth is composed of a thin crust, a thick mantle, and a core. The three layers are distinguished by different chemical compositions. In turn, both the mantle and core contain finer layers based on changing physical properties. Scientists have learned a remarkable amount about the Earth's structure even though the deepest well is only a 12-kilometer hole in northern Russia. Scientists deduce the composition and properties of the Earth's interior by studying the behavior of seismic waves. Some of the principles necessary for understanding the behavior of seismic waves are as follows:

1. In a uniform, homogeneous medium, a wave radiates outward in concentric spheres and at constant velocity.
2. The velocity of a seismic wave depends on the nature of the material that it travels through. Thus, seismic waves travel at different velocities in different types of rock. In addition, wave velocity varies with changing rigidity and density of a rock.

[3] Accurate reports of the death toll are unavailable. Published estimates range from 250,000 to 650,000.

3. When a wave passes from one material to another, it refracts (bends) and sometimes reflects (bounces back). Both **refraction** and **reflection** are easily seen in light waves. If you place a pencil in a glass half filled with water, the pencil appears bent. Of course the pencil does not bend; the light rays do. Light rays slow down when they pass from air to water, and as the velocity changes, the waves refract (Fig. 10–28). If you look in a mirror, the mirror reflects your image. In a similar manner, boundaries between the Earth's layers refract and reflect seismic waves.
4. P waves are compressional waves and travel through all gases, liquids, and solids, whereas S waves travel only through solids.

DISCOVERY OF THE CRUST–MANTLE BOUNDARY

Figure 10–29 shows that some waves travel directly through the crust to a nearby seismograph. Others travel downward into the mantle and then refract back upward to the same seismograph. The route through the mantle is longer than that through the crust. However, seismic waves travel faster in the mantle than they do in the crust. Over a short distance (less than 300 kilometers), waves traveling through the crust arrive at a seismograph before those following the longer route through the mantle. However, for longer distances, the longer route through the mantle is faster because waves travel more quickly in the mantle.

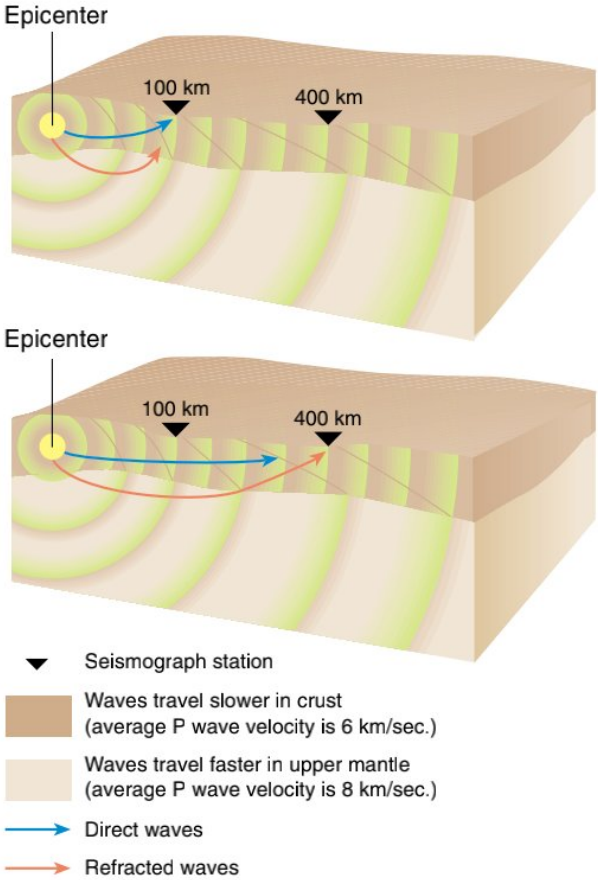


Figure 10–29 The travel path of a seismic wave. The closer station receives the direct waves first because they travel the least distance. However, a station 400 kilometers away receives the refracted wave first. Even though it travels a longer distance, most of its path is in the denser mantle, so it travels fast enough to reach the seismic station first. Think of a commuter who takes a longer route on the interstate highway rather than going via a shorter road that is choked with heavy traffic.

Figure 10–28 If you place a pencil in water, the pencil appears bent. It actually remains straight, but our eyes are fooled because light rays bend, or refract, as they cross the boundary between air and water.

The situation is analogous to the two different routes you may use to travel from your house to a friend's. The shorter route is a city street where traffic moves slowly. The longer route is an interstate highway, but you have to drive several kilometers out of your way to get to the highway and then another few kilometers from the highway to your friend's house. If your friend lives nearby, it is faster to take the city street. But if your friend lives far away, it is faster to take the longer route and make up time on the highway.

In 1909, Andrija Mohorovičić discovered that seismic waves from a distant earthquake traveled more rapidly than those from a nearby earthquake. By analyzing the arrival times of earthquake waves to many different seismographs, Mohorovičić identified the boundary between the crust and the mantle. Today, this boundary is called the **Mohorovičić discontinuity**, or the **Moho**, in honor of its discoverer.

The Moho lies at a depth ranging from 5 to 70 kilometers. Oceanic crust is thinner than continental crust, and continental crust is thicker under mountain ranges than it is under plains.

THE STRUCTURE OF THE MANTLE

The mantle is almost 2900 kilometers thick and comprises about 80 percent of the Earth's volume. Much of our knowledge of the composition and structure of the mantle comes from seismic data. As explained earlier, seismic waves speed up abruptly at the crust-mantle boundary (Fig. 10–30). Between 75 and 125 kilometers, at the base of the lithosphere, seismic waves slow down again because the high temperature at this depth causes solid rock to become plastic. Recall that the plastic layer is called the asthenosphere. The plasticity and partially melted character of the asthenosphere slow down the seismic waves. At the base of the asthenosphere 350 kilometers below the surface, seismic waves speed up again because increasing pressure overwhelms the temperature effect, and the mantle becomes less plastic.

At a depth of about 660 kilometers, seismic wave velocities increase again because pressure is great enough that the minerals in the mantle recrystallize to form denser minerals. The zone where the change occurs is called the **660-kilometer discontinuity**. The base of the mantle lies at a depth of 2900 kilometers.

DISCOVERY OF THE CORE

Using a global array of seismographs, seismologists detect direct P and S waves up to 105° from the focus of an earthquake. Between 105° and 140° is a "shadow zone" where no direct P waves arrive at the Earth's surface. This shadow zone is caused by a discontinuity, which is the mantle-core boundary. When P waves pass from the mantle into the core, they are refracted, or bent, as shown in Figure 10–31. The refraction deflects the P waves away from the shadow zone.

No S waves arrive beyond 105°. Their absence in this region shows that they do not travel through the outer core. Recall that S waves are not transmitted through liquids. The failure of S waves to pass through the outer core indicates that the outer core is liquid.

Refraction patterns of P waves, shown in Figure 10–31, shows that another boundary exists within the core. It is the boundary between the liquid outer core and

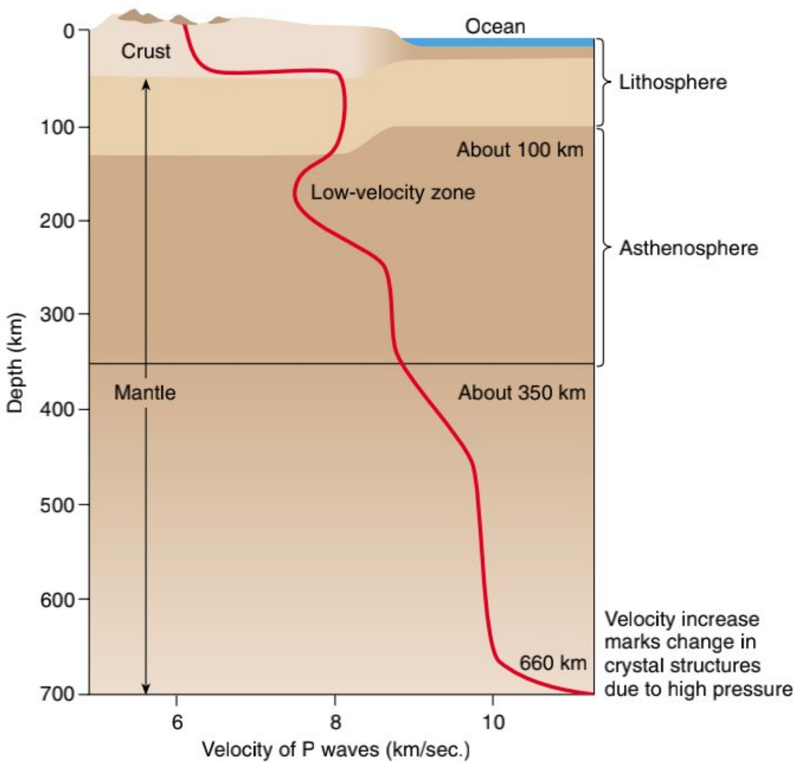


Figure 10–30 Velocities of P waves in the crust and the upper mantle. As a general rule, the velocity of P waves increases with depth. However, in the asthenosphere, the temperature is high enough that rock is plastic. As a result, seismic waves slow down in this region, called the low-velocity zone. Wave velocity increases rapidly at the 660-kilometer discontinuity, the boundary between the upper and the lower mantle, probably because of a change in mineral content due to increasing pressure.

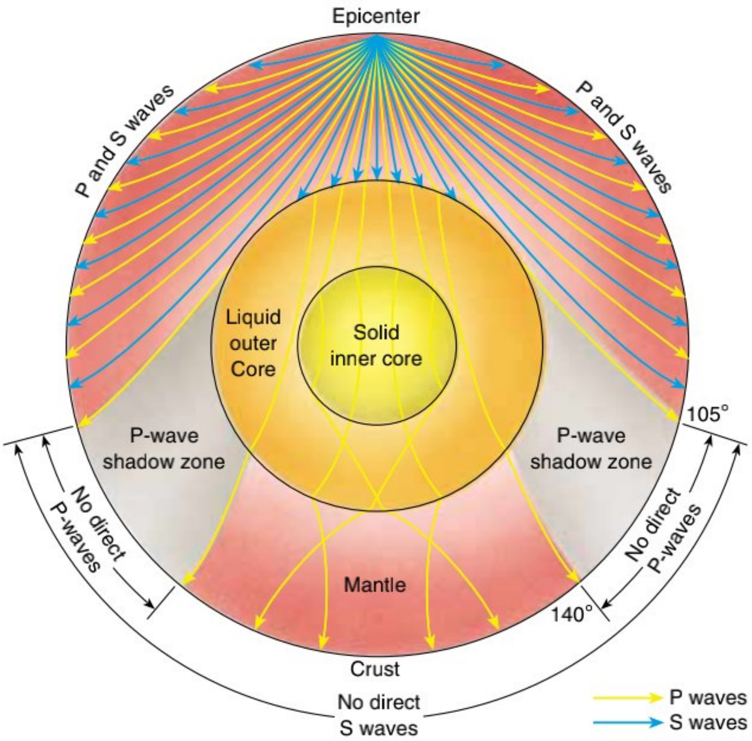


Figure 10–31 Cross section of the Earth showing paths of seismic waves. They bend gradually because of increasing pressure with depth. They also bend sharply where they cross major layer boundaries in the Earth's interior. Note that S waves do not travel through the liquid outer core, and therefore direct S waves are only observed within an arc of 105° of the epicenter. P waves are refracted sharply at the core–mantle boundary, so there is a shadow zone of no direct P waves from 105° to 140°.

the solid inner core. Although seismic waves tell us that the outer core is liquid and the inner core is solid, other evidence tells us that the core is composed of iron and nickel.

DENSITY MEASUREMENTS

The overall density of the Earth is 5.5 grams per cubic centimeter (g/cm^3); but both crust and mantle have average densities less than this value. The density of the crust ranges from 2.5 to 3 g/cm^3, and the density of the mantle varies from 3.3 g/cm^3 to 5.5 g/cm^3. Since the mantle and crust account for slightly more than 80 percent of the Earth's volume, the core must be very dense to account for the average density of the Earth. Calculations show that the density of the core must be 10 to 13 g/cm^3, which is the density of many metals under high pressure.

Many meteorites are composed mainly of iron and nickel. Cosmologists think that meteorites formed at about the same time that the Solar System did, and that they reflect the composition of the primordial Solar System. Because the Earth coalesced from meteorites and similar objects, scientists believe that iron and nickel must be abundant on Earth. Therefore, they conclude that the metallic core is composed of iron and nickel.

10.7 THE EARTH'S MAGNETISM

Early navigators learned that no matter where they sailed, a needle-shaped magnet aligned itself in a north–south orientation. Thus, they learned that the Earth has a magnetic north pole and a magnetic south pole (Fig. 10–32).

The Earth's interior is too hot for a permanent magnet to exist. Instead, the Earth's magnetic field is probably electromagnetic in origin. If you wrap a wire around a nail and connect the ends of the wire to a battery, the nail becomes magnetized and can pick up small iron objects. The battery causes electrons to flow through the wire, and this flow of electrical charges creates the electromagnetic field.

Most likely, the Earth's magnetic field is generated within the outer core. Metals are good conductors of electricity and the metals in the outer core are liquid and very mobile. Two types of motion occur in the liquid outer core. (1) Because the outer core is much hotter at its base than at its surface, convection currents cause the liquid metal to rise and fall. (2) The rising and falling metals are then deflected by the Earth's spin. These convecting, spinning liquid conductors generate the Earth's magnetic field. New research has shown that, in addition, the solid inner core rotates more rapidly than the Earth.

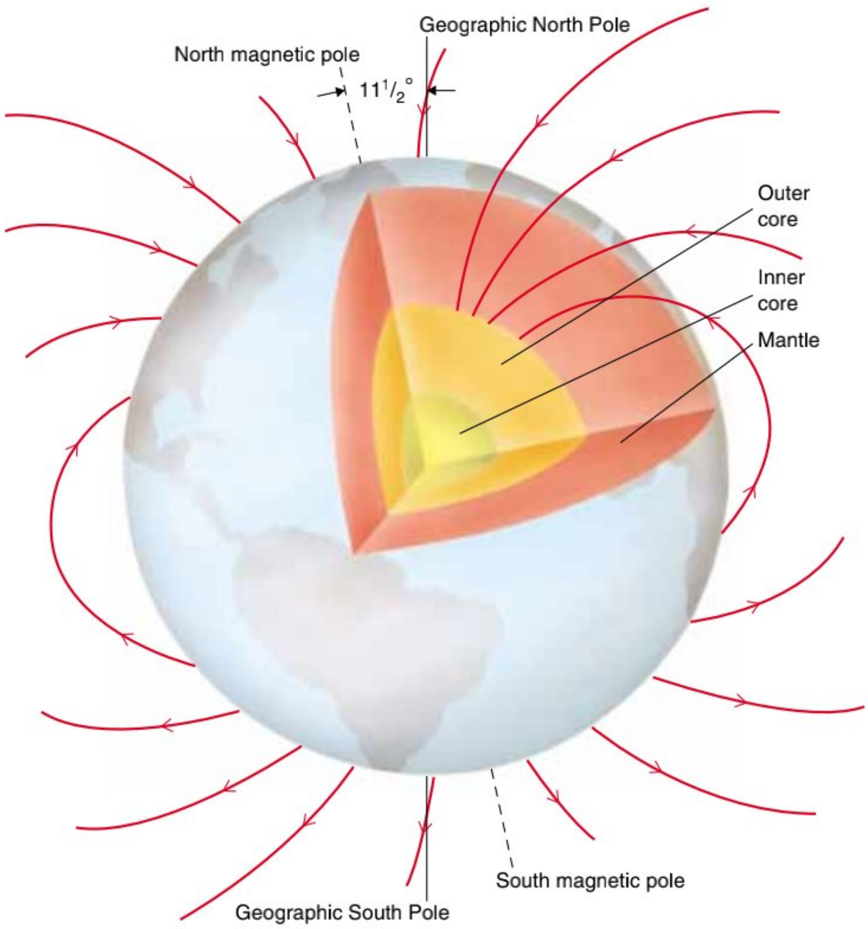


Figure 10–32 The magnetic field of the Earth. Note that the magnetic north pole is 11.5° offset from the geographic pole.

Magnetic fields are common in planets, stars, and other objects in space. Our Solar System almost certainly posessed a weak magnetic field when it first formed. The flowing metals of the liquid outer core amplified some of this original magnetic force. If we observe this magnetic field over thousands of years, its axis is approximately lined up with the Earth's rotational axis, because the Earth's spin affects the flow of metal in the outer core.

SUMMARY

An **earthquake** is a sudden motion or trembling of the Earth caused by the abrupt release of slowly accumulated energy in rocks. Most earthquakes occur along tectonic plate boundaries. Earthquakes occur either when the elastic energy accumulated in rock exceeds the friction that holds rock along a fault or when the elastic energy exceeds the strength of the rock and the rock breaks by **brittle fracture**.

An earthquake starts at the initial point of rupture, called the **focus**. The location on the Earth's surface directly above the focus is the **epicenter**. **Seismic waves** include **body waves**, which travel through the interior of the Earth, and **surface waves**, which travel on the surface. **P waves** are compressional body waves. **S waves** are body waves that travel slower than P waves. They consist of a shearing motion and travel through solids but not liquids. Surface waves travel more slowly than either type of body wave. Seismic waves are recorded on a **seismograph**. Modern geologists use the **moment magnitude** scale to record the energy released during an earthquake. The distance from a seismic station to an earthquake is calculated by recording the time between the arrival of P and S waves. The epicenter can be located by measuring the distance from three or more seismic stations. Earthquake damage is influenced by rock and soil type, construction design, and the likelihood of fires, landslides, and tsunamis.

Earthquakes are common at all three types of plate boundaries. The San Andreas fault zone is an example of a transform plate boundary, where two plates slide past one another. Subduction zone earthquakes occur when the subducting plate slips suddenly. Earthquakes occur at divergent plate boundaries as blocks of lithosphere along the fault drop downward. Earthquakes occur in plate interiors along old faults or where sediment depresses the lithosphere.

Long-term earthquake prediction is based on the observation that most earthquakes occur at tectonic plate boundaries. Short-term prediction is based on occurrences of **foreshocks**, release of radon gas, changes in the land surface, the water table, electrical conductivity, and erratic animal behavior.

The Earth's internal structure and properties are known by studies of earthquake wave velocities and **refraction** and **reflection** of seismic waves as they pass through the Earth. Flowing metal in the outer core generates the Earth's magnetic field.

KEY WORDS

earthquake *154*
stress *156*
plastic deformation *156*
brittle fracture *156*
seismic wave *158*
seismology *158*
body wave *158*
focus *158*
epicenter *158*
surface wave *158*
P wave *158*
S wave *159*
shear wave *159*
Rayleigh wave *159*
Love wave *159*
seismograph *160*
seismogram *160*
Richter scale *160*
moment magnitude scale *161*
time-travel curve *162*
liquefaction *164*
tsunami *165*
strike–slip fault *165*
San Andreas fault zone *165*
fault creep *166*
thrust fault *166*
Benioff zone *168*
foreshock *171*
Mohorovičić discontinuity (Moho) *173*

REVIEW QUESTIONS

1. Explain how energy is stored prior to and then released during an earthquake.
2. Describe the behavior of rock during elastic deformation, plastic deformation, and brittle fracture.
3. Give two mechanisms that can release accumulated elastic energy in rocks.
4. Why do most earthquakes occur at the boundaries between tectonic plates? Are there any exceptions?
5. Define *focus* and *epicenter*.
6. Discuss the differences between P waves, S waves, and surface waves.
7. Explain how a seismograph works. Sketch what an imaginary seismogram would look like before and during an earthquake.
8. Describe the similarities and differences between the Richter and moment magnitude scales. What is actually measured, and what information is obtained?
9. Describe how the epicenter of an earthquake is located.
10. List five different factors that affect earthquake damage. Discuss each briefly.
11. Discuss earthquake mechanisms at the three different types of tectonic plate boundaries.
12. Briefly discuss major faults close to Los Angeles.
13. Discuss earthquake mechanisms at plate interiors.
14. Discuss the scientific reasoning behind long-term and short-term earthquake prediction.
15. Outline the seismic gap hypothesis. Discuss modern objections to the theory.
16. What is the Moho? How was it discovered?
17. Explain how geologists learned that the core is composed of iron and nickel.
18. Briefly discuss the theories for the existence of the Earth's magnetic field.