CHAPTER 10

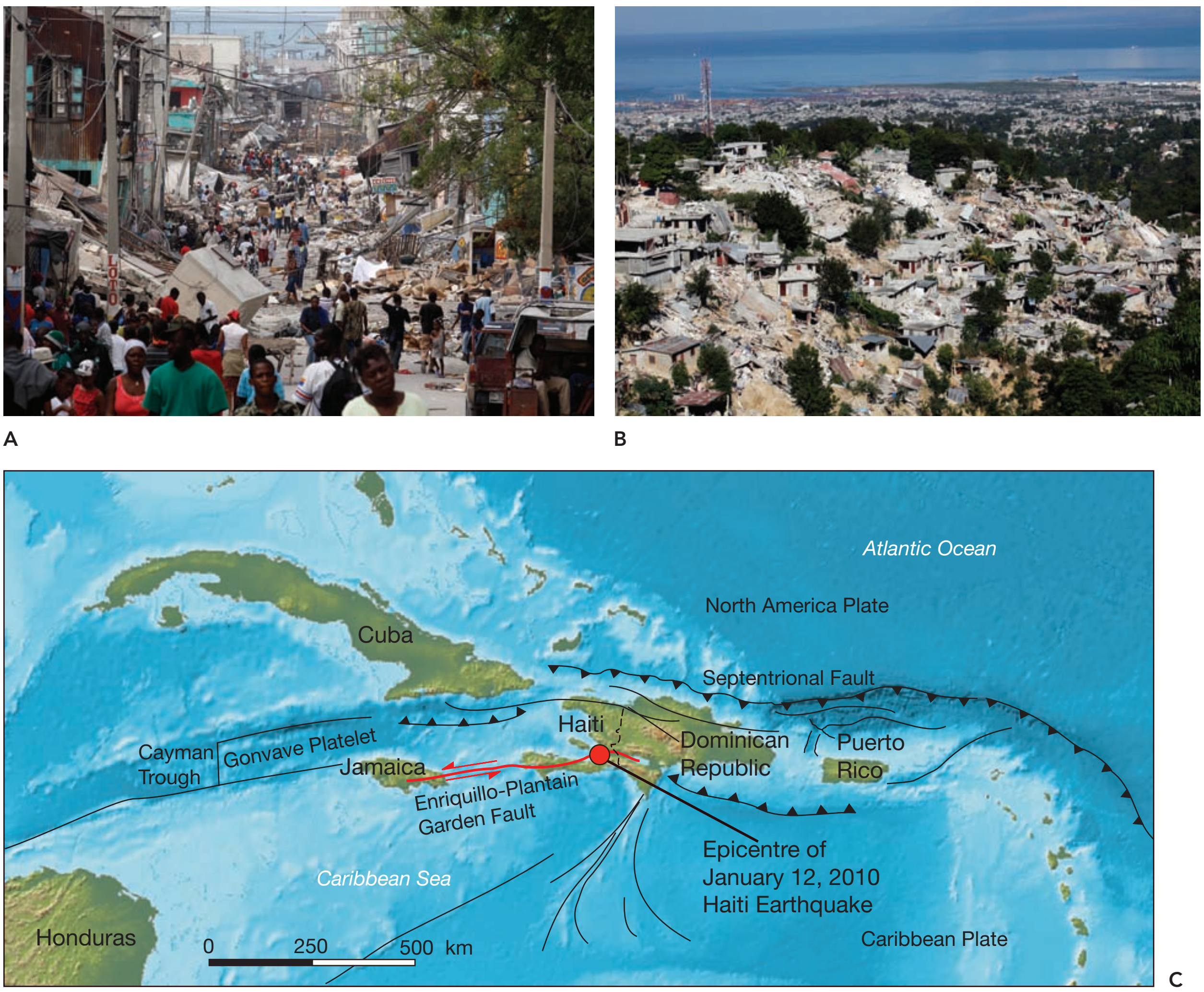
EARTHQUAKES AND EARTH'S INTERIOR

CHAPTER OUTLINE

Destruction Destruction caused by a severe earthquake that struck Haiti on January 12, 2010. About 100,000 people perished and 600,000 were left homeless.
(Photo by Thony Belizaire/AFP/Getty Images)

"They are pulling people out of the rubble, literally with blood running in the gutter like water," sighed an exhausted aid worker in Haiti, the western hemisphere's poorest country, which was devastated by a magnitude 7.0 earthquake in January 2010 (chapter opening photo; Figure 10.1). Some survivors huddled together; others lay in the street, unable to pick themselves up. Many spent desolate hours wandering in shock while others cowered under makeshift shelters as night fell.

Such was the scene late in the afternoon of January 12, 2010, around the capital city of Port-au-Prince (Figure 10.1A, B). It was difficult to take in the sheer scale of destruction as poorly constructed buildings crumpled like paper with people still in them, and shantytowns were swept away by landslides. The quake flattened hillsides, knocked out communications, blocked roads, destroyed most hospitals, and clogged Haiti's main seaport with debris. The disaster was made even worse by strong aftershocks that rumbled for days after the main event. "It was a worst-case scenario. It should be a wake-up call for the entire Caribbean," lamented an earthquake expert. Haiti, Dominican Republic, and Jamaica are sandwiched between strike-slip faults (Figure 10.1C), like the famous San Andreas. The quake occurred along a 65-kilometre stretch of the southern fault (near Port-au-Prince), where the Caribbean plate had been stuck against the smaller Gonvave platelet for 240 years. On January 12 the



◆ **Figure 10.1** **A.** Devastation in downtown Port-au-Prince after the January 12, 2010, earthquake. **B.** Poorly built houses crumble on a hillside above Port-au-Prince; others were swept away by landslides. The image in **C** shows the geology around Haiti, which is sandwiched between left-lateral strike-slip faults. The quake occurred near Port-au-Prince along the Enriquillo-Plantain Garden fault that separates the Caribbean plate and Gonvave platelet. When the plates moved in opposite directions, built-up energy was violently released.
(Photo **A** by REUTERS/Jorge Silva; photo **B** by REUTERS/Eduardo Munoz; **C** modified from an image courtesy of *Science Daily*)

two plates jerked forward about 2 metres in opposite directions to relieve built-up stress, thereby releasing massive amounts of energy in Haiti. Because it was a land-based quake, no major tsunami was generated like the famous one triggered by the Boxing Day earthquake under the Indian Ocean in 2004. In 1946 a more powerful 8.1 quake struck the northeast corner of Haiti, but the damage was greater in 2010 because the quake was near the capital city, it was shallow, and buildings were not constructed to withstand shaking. This is also why Haiti's 7.0 quake was more devastating than Chile's 8.8 quake the following month.

Earthquakes and associated phenomena, such as landslides and tsunami, are tangible reminders that Earth is a highly dynamic system, both above and below its surface. Each year, more than a thousand earthquakes occur in Canada alone, although the majority of these produce vibrations that are simply too weak to be felt by humans and can be detected only by sensitive instruments.

Occasionally, media reports alert us to particularly strong earthquake events. The most severe earthquake disasters in recent times have included those centred in Izmit, Turkey (August 17, 1999, with more than 17,000 deaths); Gujarat, India (January 26, 2001, with more than 20,000 deaths); the Bam region of Iran (December 26, 2003, with more than 25,000 deaths); and Sumatra, Indonesia (December 26, 2004, with 230,000 deaths, many by tsunami). Disasters like these are difficult for most Canadians to comprehend; indeed, many Canadians have never even *felt* an earthquake in their lifetime. The events that occurred in the Indian Ocean and Haiti are particularly significant in that they changed the consciousness of millions of people worldwide. Victims were reported in at least eight countries bordering the Indian Ocean after the 2004 event. In a sense, these disasters were truly felt by the global community. Before the Boxing Day disaster, few people had even heard the term *tsunami*; by the next day, it had become a household word. It was a strong reminder that earthquake hazards can extend far beyond the area in which an earthquake itself is generated. These events were also wake-up calls to the many nations that, having not experienced an earthquake for some time, had become complacent about preparing for such a disaster.

Ancient stories of the Aboriginal peoples of Vancouver Island, historical documents from Japan, computer-modelling methods, and a variety of geologic evidence indicate that an earthquake of similar magnitude to the Sumatra-Andaman earthquake occurred in 1700. Canada is not immune to tsunami: a rare Atlantic tsunami associated with the 1929 Grand Banks, Newfoundland and Labrador, earthquake was responsible for at least 10 deaths—historically, the largest number of deaths attributed to a single Canadian earthquake. Thus, the question is not "will another earthquake occur in Canada?" but, rather, "when will the next big earthquake happen—and where?" Answering this question will require knowledge of past earthquake events and the study of current earthquake activity.

In this chapter, we examine what earthquakes are, how they are produced, the damage they cause, and how they are studied. We also explore the positive side of earthquakes—how the behaviour of seismic waves has been used to deduce the structure of Earth's interior.

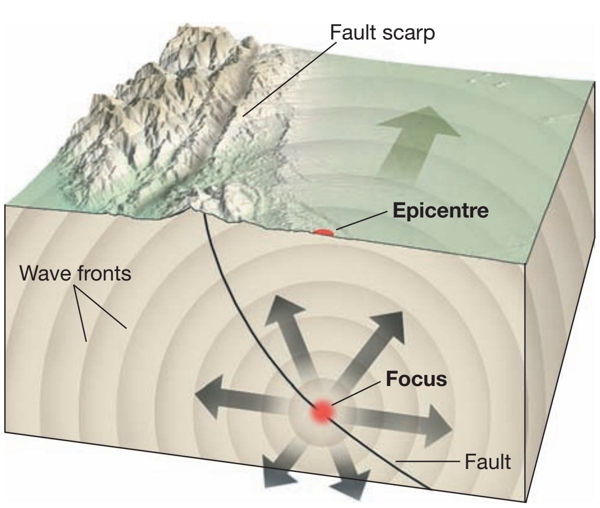
WHAT IS AN EARTHQUAKE?

GEODe:Earth

Earthquakes

What Is an Earthquake?

An **earthquake** is the vibration of Earth produced by the rapid release of energy. Most often, earthquakes are caused by slippage along faults in Earth's crust. The energy released radiates in all directions from its source—the **focus** (*focus* = a point)—in the form of seismic waves. These waves are analogous to those produced when a stone is dropped into a calm pond (Figure 10.2). Just as the impact of the stone sets water waves in motion, an earthquake generates seismic waves that radiate throughout Earth. Even though the energy dissipates rapidly with increasing distance from the focus, sensitive instruments located around the world record the event.



◆ **Figure 10.2** Earthquake focus and epicentre. The focus is the place within Earth at which the initial displacement occurs. The epicentre is the surface location directly above the focus.

Earthquakes and Faults

The tremendous energy released by atomic explosions or by volcanic eruptions can produce earthquakes, but these events are relatively weak and infrequent. What mechanism produces a destructive earthquake? Ample evidence exists that Earth is not a static planet. We know that Earth's crust has been uplifted at times, because we have found numerous ancient wave-cut benches many metres above the level of the highest tides. Other regions exhibit evidence of extensive subsidence. In addition to these vertical displacements, offsets in fence lines, roads, and other structures indicate that horizontal movement is common (Figure 10.3). These movements are usually associated with faults (Chapter 9).

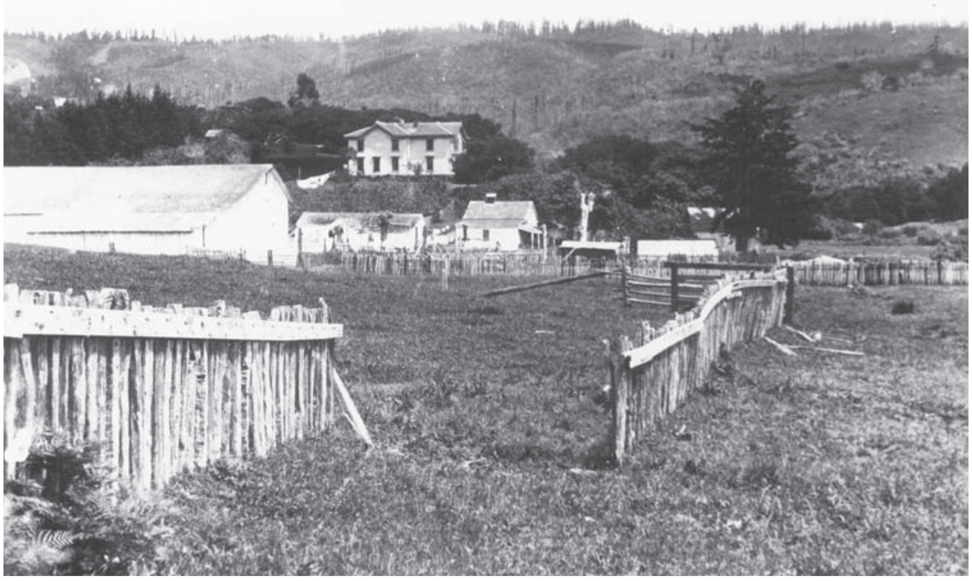
Most of the motion along faults can be satisfactorily explained by plate tectonic theory, as introduced in Chapter 1. Mobile plates of lithosphere interact with neighbouring plates, straining and deforming the rocks at their margins. In fact, it is along the faults associated with plate boundaries that most earthquakes occur. Furthermore, earthquakes are repetitive: as soon as one is over, the continuous motion of the plates resumes, adding stress to the rocks until they fail again.

Elastic Rebound

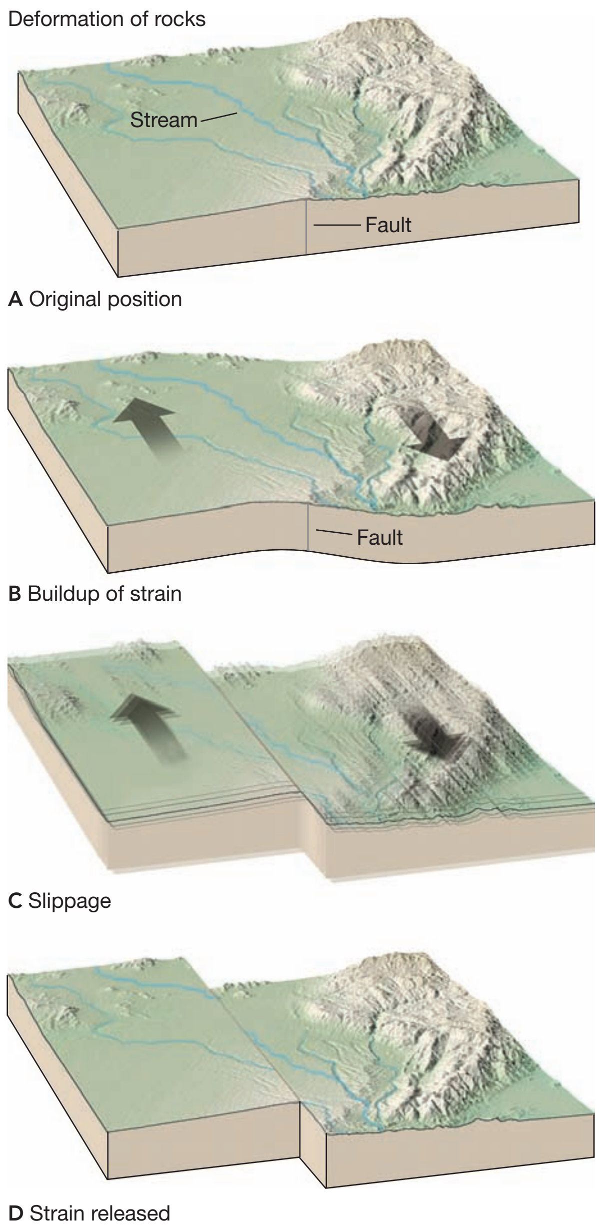
The actual mechanism of earthquake generation eluded geologists until H. F. Reid of Johns Hopkins University conducted a study following the great 1906 San Francisco earthquake. The earthquake was accompanied by horizontal surface displacements of several metres along the northern portion of the San Andreas Fault. This 1300-kilometre fracture extends northwest–southeast through southern California. It is a large fault zone that separates two great sections of Earth's crust: the North American plate and the Pacific plate. Field investigations determined that during this single earthquake, the Pacific plate lurched as much as 5 metres northward past the adjacent North American plate.

The mechanism for earthquake formation that Reid deduced from this information is illustrated in Figure 10.4. In part A of the figure, you see an existing fault. In part B, tectonic forces slowly deform the crustal rocks on both sides of the fault, as demonstrated by the bent features. Under these conditions, rocks bend and store elastic energy, much like a wooden stick does if bent. Eventually, the frictional resistance holding the rocks together is overcome. As slippage occurs at the weakest point (the focus), displacement exerts stress farther along the fault, where additional slippage occurs (Figure 10.4C) until most of the built-up stress is released (Figure 10.4D). This slippage allows the deformed rock to "snap back." The vibrations we know as an earthquake occur as the rock elastically returns to its original shape. The springing back of the rock was termed **elastic rebound** by Reid because the rock behaves elastically, much like a stretched rubber band does when it is released. After release, stress builds up again along the fault as the earthquake cycle continues.

In summary, most earthquakes are produced by the rapid release of elastic energy stored in rock that has been subjected to great stress. Once the strength of the rock is exceeded, it suddenly ruptures, causing the vibrations of an earthquake. Earthquakes most commonly occur along existing faults whenever the frictional forces on the fault surfaces are overcome.



◆ **Figure 10.3** This fence was offset 2.5 metres during the 1906 San Francisco earthquake. (Photo by G. K. Gilbert, U.S. Geological Survey)



◆ **Figure 10.4** Elastic rebound. As rock is deformed, it bends, storing elastic energy. Once strained beyond its breaking point, the rock fails (breaks), releasing the stored-up energy as earthquake waves.

Elastic Rebound

Foreshocks and Aftershocks

The intense vibrations of the 1906 San Francisco earthquake lasted about 40 seconds. Although most of the displacement along the fault occurred in this rather short span, additional movements along this and other nearby faults lasted for several days following the main quake. The adjustments that follow a major earthquake often generate smaller earthquakes called **aftershocks**. Although these aftershocks are usually much weaker than the main earthquake, they can sometimes destroy already badly weakened structures. This occurred during a 1988 earthquake in Armenia and the 2010 Haitian quake (chapter opening photo; Figure 10.1). A large aftershock collapsed many structures that had been weakened by the main tremor.

In addition, small earthquakes called **foreshocks** often precede a major earthquake by days or, in some cases, by as much as several years. Monitoring of these foreshocks has been used as a means of predicting forthcoming major earthquakes, with mixed success. We will consider the topic of earthquake prediction later in this chapter.

SEISMOLOGY

GEODe:Earth **Earthquakes**
Seismology

The study of earthquake waves, **seismology** (*seismos* = shake, *ology* = the study of) dates back to attempts made by the Chinese almost 2000 years ago to determine the direction from which these waves originated. The seismic instrument used by the Chinese was a large hollow jar that contained a suspended mass (similar to a clock pendulum) connected in some fashion to the jaws of several large dragon figurines that encircled the container (Figure 10.5). The jaws of each dragon held a metal ball. When earthquake waves reached the instrument, the relative motion between the suspended mass and the jar would dislodge some of the metal balls into the waiting mouths of frog figurines directly below.

The Chinese were probably aware that the first strong ground motion from an earthquake is directional, and when it is strong enough, all poorly supported items will topple over in the same direction. Apparently the Chinese used this fact, plus the position of the dislodged balls, to detect the direction to an earthquake's source. However, the complex motion of seismic waves makes it unlikely that the actual direction to an earthquake was determined with any regularity.

In principle, modern **seismographs** (*seismos* = shake, *graph* = write), instruments that record seismic waves, are not unlike the device used by the early Chinese. Seismographs have a mass freely suspended from a support that is attached to the ground (Figure 10.6). When the vibration from a distant earthquake reaches the instrument, the **inertia***

**Inertia: Simply stated, objects at rest tend to stay at rest, and objects in motion tend to remain in motion, unless acted on by an outside force. You probably have experienced this phenomenon when an automobile you were in stopped quickly and your body continued to move forward.*

◆ **Figure 10.5** Ancient Chinese seismograph. During an Earth tremor, the dragons located in the direction of the main vibrations would each drop a ball into the mouth of the frog below it.
(Photo by James E. Patterson)

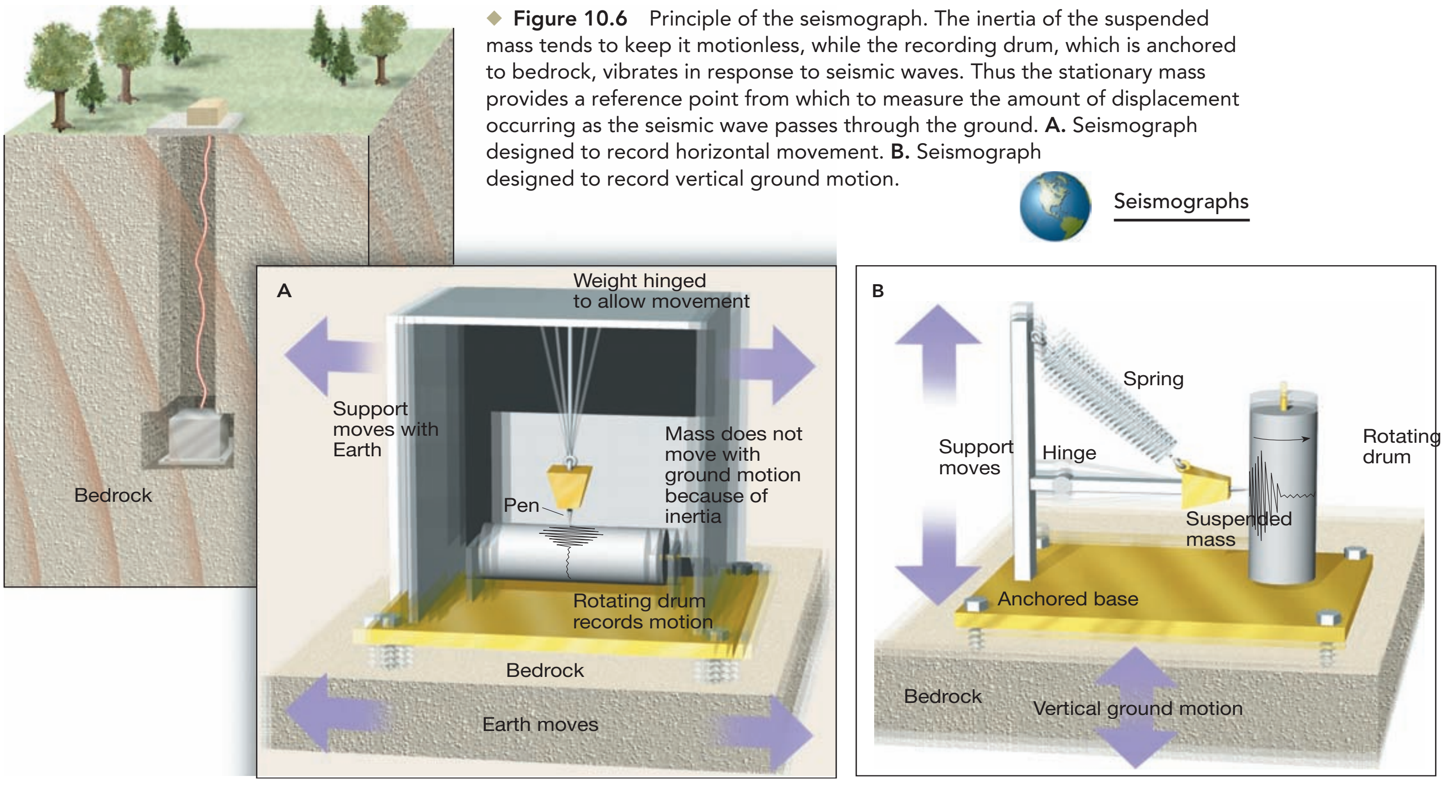
(*iners* = idle) of the mass keeps it relatively stationary while Earth and the support move. The movement of Earth in relation to the stationary mass is recorded on a rotating drum, magnetic tape, or digital device.

The records obtained from seismographs, called **seismograms** (*seismos* = shake, *gramma* = what is written), provide a great deal of information concerning the behaviour of seismic waves. Seismic waves are energy forms that radiate out in all directions from the focus. The propagation (transmission) of this energy can be compared to the shaking of gelatine in a bowl that results as some is spooned out. Whereas the gelatine will have one mode of vibration, seismograms reveal that two main groups of seismic waves are generated by the slippage of a rock mass. The wave types that travel through Earth's interior are called **body waves**. The types that travel along the outer part of Earth are called **surface waves**. Body waves are further divided into two types, called **primary waves,** or **P waves**, and **secondary waves**, or **S waves**.

Body waves are divided into P and S waves by their mode of travel through intervening materials. P waves are "push–pull" waves—they push (compress) and pull (expand) rocks in the direction the wave is travelling (Figure 10.7A). Imagine holding someone by the shoulders and shaking that person. This push–pull movement is how P waves move through Earth. Solids, liquids, and gases resist

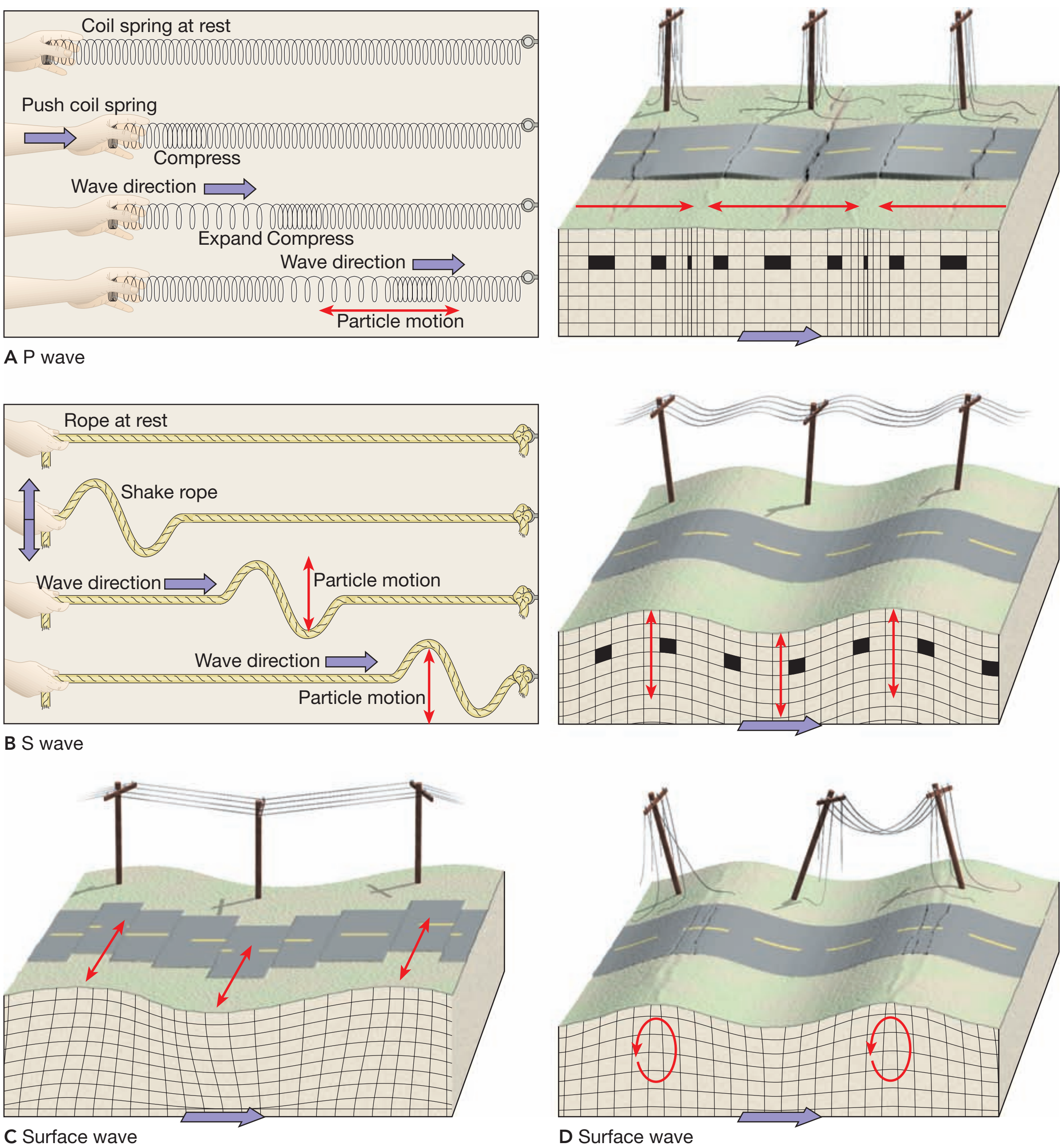
◆ **Figure 10.6** Principle of the seismograph. The inertia of the suspended mass tends to keep it motionless, while the recording drum, which is anchored to bedrock, vibrates in response to seismic waves. Thus the stationary mass provides a reference point from which to measure the amount of displacement occurring as the seismic wave passes through the ground. **A.** Seismograph designed to record horizontal movement. **B.** Seismograph designed to record vertical ground motion.

Seismographs



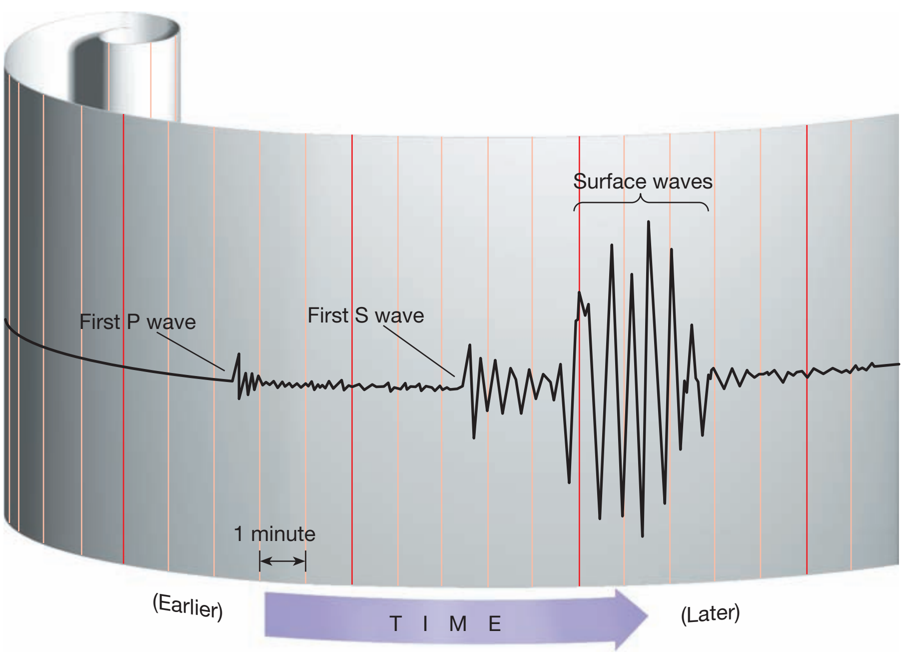
a change in volume when compressed and will elastically spring back once the force is removed. Therefore, P waves, which are compressional waves, can travel through all these materials.

Conversely, S waves "shake" the particles at right angles to their direction of travel. This can be illustrated by fastening one end of a rope and shaking the other end, as shown in Figure 10.7B. Unlike



◆ **Figure 10.7** Basic types of seismic waves and their characteristic motion. (Note that during a typical earthquake, ground shaking consists of a combination of various kinds of seismic waves that have complex movements in three dimensions.) **A.** As illustrated by a coil spring, P waves are compressional waves that alternately compress and expand the material through which they pass. The back-and-forth motion produced as compressional waves travel along the surface can cause the ground to buckle and fracture, and they can cause power lines to break. **B.** S waves cause material to oscillate at right angles to the direction of wave motion. Because S waves can travel in any plane, they produce up-and-down and sideways shaking of the ground. **C.** One type of surface wave is essentially the same as that of an S wave that exhibits only horizontal motion. This kind of surface wave moves the ground from side to side and can be particularly damaging to the foundations of buildings. **D.** Another type of surface wave travels along Earth's surface, much like rolling ocean waves. The arrows show the elliptical movement of rock as the wave passes.

Seismic Wave Motion



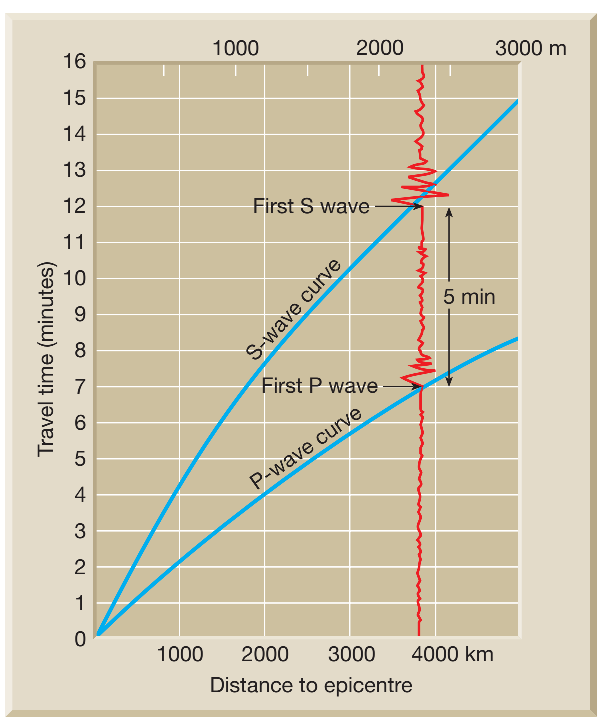
◆ **Figure 10.8** Typical seismogram. Note the time interval (about 5 minutes) between the arrival of the first P wave and the arrival of the first S wave.

P waves, which temporarily change the *volume* of intervening material by alternately compressing and expanding it, S waves temporarily change the *shape* of the material that transmits them. Because fluids (gases and liquids) do not respond elastically to changes in shape, they will not transmit S waves.

The motion of surface waves is somewhat more complex (Figure 10.7C). As surface waves travel along the ground, they cause the ground and anything resting on it to move, much as ocean swells toss a ship. In addition to their up-and-down motion, surface waves have a side-to-side motion similar to an S wave oriented in a horizontal plane. This latter motion is particularly damaging to the foundations of structures.

By observing a "typical" seismic record, as shown in Figure 10.8, you can see major differences among these seismic waves: P waves arrive at the recording station first, then S waves, and then surface waves. This is a consequence of their speeds. To illustrate, the velocity of P waves through granite within the crust is about 6 kilometres per second. S waves under the same conditions travel at 3.6 kilometres per second. Differences in density and elastic properties of the rock greatly influence the velocities of these waves. Generally, in any solid material, P waves travel 1.7 times as fast as S waves, and S waves travel 1.1 times as fast as surface waves.

A system was developed for locating earthquake epicentres by using seismograms from earthquakes whose epicentres could be easily pinpointed from physical evidence. From these seismograms, travel-time graphs were constructed (Figure 10.9). The first travel-time graphs were greatly improved when seismograms became available from nuclear explosions, because the precise location and time of detonation were known.



◆ **Figure 10.9** A travel-time graph is used to determine the distance to the epicentre. The difference in arrival times of the first P and S waves in the example is 5 minutes. Thus, the epicentre is roughly 3800 kilometres away.

In addition to velocity differences, also notice in Figure 10.8 that the height or, more correctly, the amplitude of these wave types varies. The S waves tend to have a slightly greater amplitude than the P waves, while the surface waves, which cause the greatest destruction, tend to exhibit an even greater amplitude. Because surface waves are confined to a narrow region near the surface and are not spread throughout Earth as P and S waves are, they retain their maximum amplitude longer. Surface waves also have longer periods (time intervals between crests); therefore, they are commonly referred to as **long waves**, or **L waves**.

As you will see, seismic waves are useful in determining the location and magnitude of earthquakes, in addition to providing information on the nature of Earth's internal characteristics.

LOCATING THE SOURCE OF AN EARTHQUAKE

Earthquakes
Locating the Source of an Earthquake

Recall that the *focus* is the place within Earth where earthquake waves originate. The **epicentre** (*epi* = upon, *centr* = a point) is the location on the surface directly above the focus (see Figure 10.2).

The difference in velocities of P and S waves provides a method for locating the epicentre. The principle used is analogous to a race between two vehicles, one faster than the other. The P wave always wins the race, arriving ahead of the S wave. But the greater the length of the race, the greater will be the difference in the arrival times at the finish line (the seismic station). Therefore, the greater the interval measured on a seismogram between the arrival of the first P wave and the first S wave, the greater the distance to the earthquake source. By using the sample seismogram in Figure 10.8 and the travel-time curves in Figure 10.9, we can determine the distance separating the recording station from the earthquake in two steps: (1) determine the time interval between the arrival of the first P wave and the first S wave, and (2) find on the travel-time graph the equivalent time spread between the P and S wave curves. From this information, we can determine that this earthquake occurred 3800 kilometres from the recording instrument.

Now we know the *distance* but what *direction*? The epicentre could be in any direction from the seismic station. As shown in Figure 10.10, the precise location can be found when the distance is known from three or more different seismic stations. On a globe, we draw a circle around each seismic station. Each circle represents the epicentre distance for each station. The point at which the three circles intersect is the epicentre of the quake. This method is called *triangulation*.

The study of earthquakes was greatly bolstered during the 1960s through efforts to discriminate between underground nuclear explosions and natural

◆ **Figure 10.10** An earthquake epicentre is located by using the distances obtained from three or more seismic stations.

earthquakes. A worldwide network of more than 100 seismic stations is coordinated through Golden, Colorado. The largest of these, located in Billings, Montana, consists of an array of 525 instruments grouped in 21 clusters covering a region 200 kilometres in diameter. By using data from this array and high-speed computers, seismologists are able to differentiate between nuclear blasts and natural earthquakes, as well as determine the position of an earthquake's epicentre.

Earthquake Belts

About 95 percent of the energy released by earthquakes originates in a few relatively narrow zones that wind around the globe (Figure 10.11). The greatest energy is released along a path around the outer edge of the Pacific Ocean known as the *circum-Pacific belt*. Included in this zone are regions of great seismic activity, such as Japan, the Philippines, Chile, and numerous volcanic island chains, as exemplified by the Aleutian Islands.

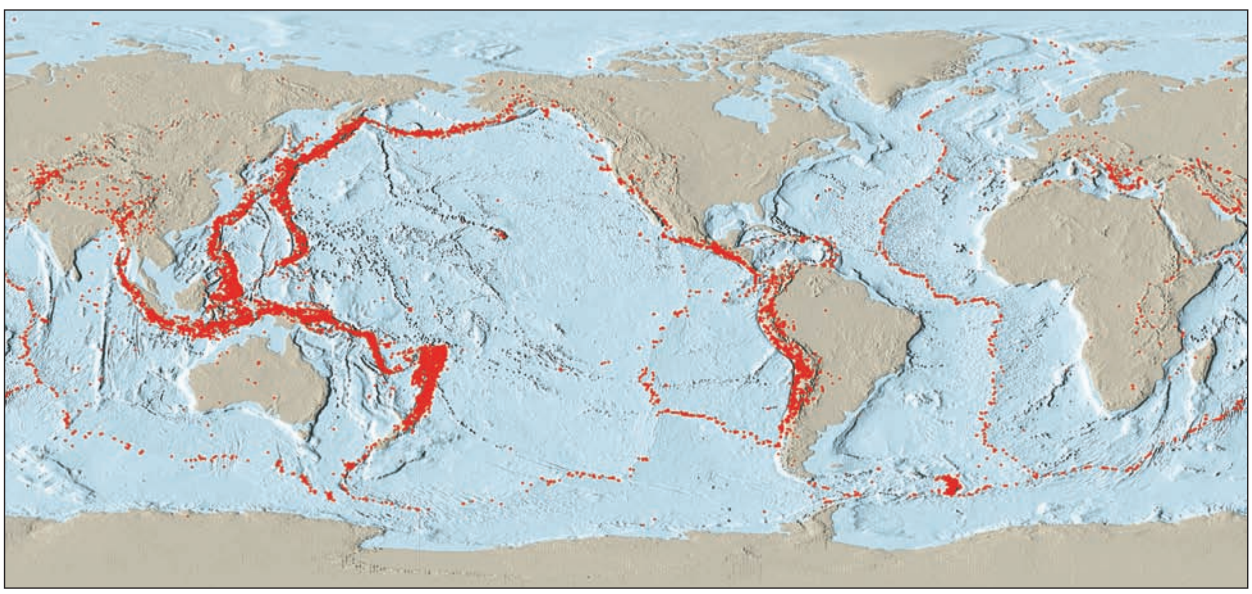
Another major concentration of strong seismic activity extends through the mountainous regions that flank the Mediterranean Sea and continues through Iran and on past the Himalayan complex. Figure 10.11 indicates that yet another continuous belt extends for thousands of kilometres through the world's oceans. This zone coincides with the oceanic ridge system, an area of frequent but weaker seismic activity.

Areas of North America included in the circum-Pacific belt lie adjacent to California's San Andreas Fault and along the western coastal regions extending to the Aleutian Islands. In addition to these high-risk areas, other sections of North America are regarded as regions where strong earthquakes are likely to occur. In Canada earthquakes are concentrated in British Columbia but do occur elsewhere (Box 10.1).

Earthquake Depths

Areas of North America included in the circum-Pacific belt lie adjacent to California's San Andreas Fault and along the western coastal regions extending to the Aleutian Islands. In addition to these high-risk areas, other sections of North America are regarded as regions where strong earthquakes are likely to occur. In Canada earthquakes are concentrated in British Columbia but do occur elsewhere (Box 10.1).

Evidence from seismic records reveals that earthquakes originate at depths ranging from 5 to nearly 700 kilometres. In a somewhat arbitrary fashion, earthquake foci have been classified according to their depth of occurrence. Those with points of origin within 70 kilometres of the surface are referred to as *shallow*, while those generated between 70 and 300 kilometres are considered *intermediate*, and those with a focus deeper than 300 kilometres are classified as *deep*. About 90 percent of all earthquakes occur at depths of less than 100 kilometres, and nearly all very damaging earthquakes appear to originate at shallow depths (the 2010 Haitian one originated 13 kilometres deep).



◆ **Figure 10.11** Distribution of the 14,229 earthquakes with magnitudes of at least 5 between 1980 and 1990.
(Data from National Geophysical Data Center/NOAA)

Plate Boundary Features

BOX 10.1 CANADIAN PROFILE

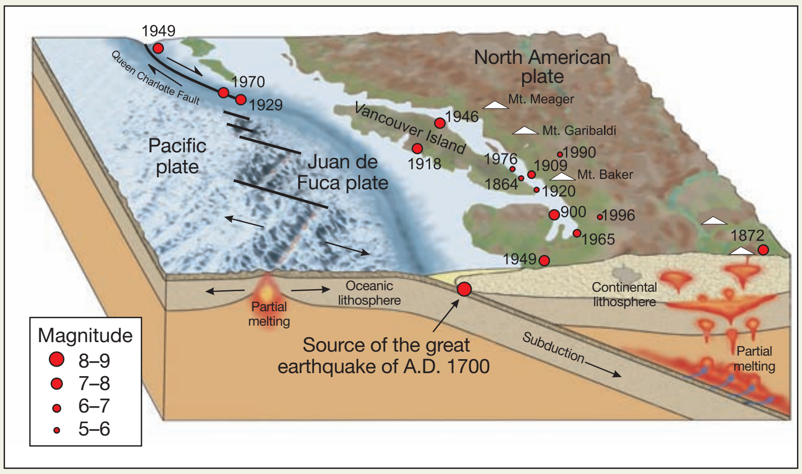
Earthquakes in Canada

David Eaton*

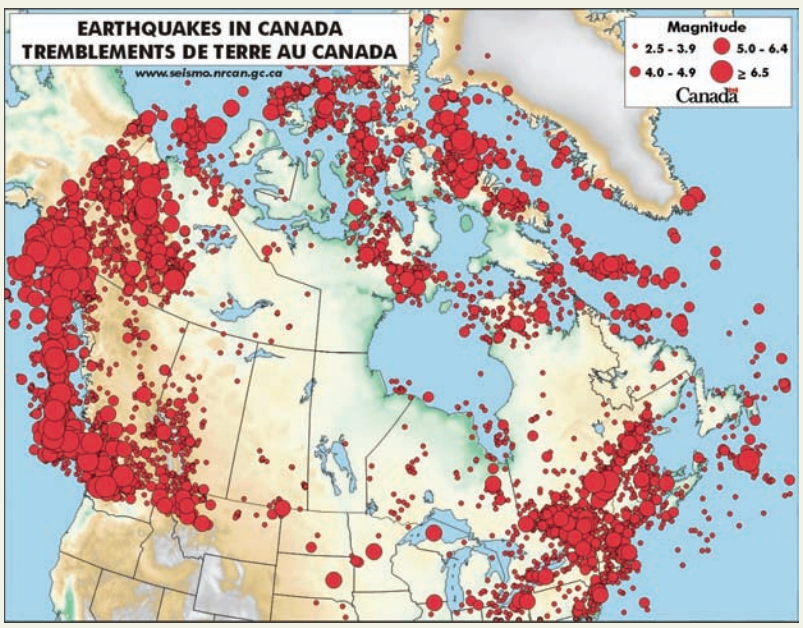
On August 22, 1949, residents of the Queen Charlotte Islands were jolted by Canada's largest measured earthquake. With a magnitude of 8.1, this offshore tremor ranks as one of the great earthquakes of the twentieth century, well recorded around the globe. The earthquake knocked cows off their feet and caused an oil tank at Cumshewa Inlet to collapse. Even 300 kilometres away in Terrace, British Columbia, the ground shaking was so intense that standing on the street was described as "like being on the heaving deck of a ship at sea." Fortunately, the sparse population of the region meant that casualties and property damage were light in comparison with other earthquakes of similar magnitude, such as the 1946 Vancouver Island earthquake (Figure 10.12B).

The 1949 earthquake ruptured a major segment of the Queen Charlotte fault system, an active right-lateral transform fault along Canada's west coast (Figure 10.A). Across the fault, the Pacific plate is creeping slowly northward relative to North America, in a manner analogous to lithospheric movement along the San Andreas Fault system in California. Friction across the fault resists this motion, causing stress to accumulate during aseismic intervals. The sudden release of stored energy during an earthquake (elastic rebound) starts the earthquake cycle once again.

In southwest British Columbia, a different plate-tectonic setting exists, more akin to Japan than California (Figure 10.A). From southern Vancouver Island to northern California, the oceanic Juan de Fuca plate is subducting beneath the North American plate within a system known as the Cascadia subduction zone. Although large earthquakes have taken place, such as the widely felt Nisqually earthquake near Seattle, Washington, on February 28, 2001, the Cascadia subduction zone is unusual because no large historical earthquake has occurred along the dipping plate boundary. (The Nisqually earthquake took place within the oceanic slab as it broke apart during descent into the mantle). There is now compelling scientific evidence, however, that a



◆ **Figure 10.A** Configuration of lithospheric plates and locations of significant earthquakes that have been recorded in British Columbia. (Modified from R. J. W. Turner, J. J. Clague, and B. J. Groulx, *Geoscape Vancouver, Living with Our Geological Landscape III*, Geological Survey of Canada, 1998, Open File 3309; reproduced with the permission of the Minister of Public Works and Government Services Canada, 2008, and courtesy of Natural Resources Canada)



◆ **Figure 10.B** The distribution and magnitudes of earthquakes in Canada since 1980. (Reproduced with the permission of Natural Resources Canada, 2010, and courtesy of Earthquake Canada)

**David Eaton is a professor in the Department of Earth Sciences at the University of Calgary.*

so-called megathrust earthquake took place with an estimated magnitude of 9.0 along the plate boundary on January 26, 1700, before settlement of the region by Europeans. The timing of this earthquake is known from the oral traditions of the Cowichan people of Vancouver Island, from the study of tree rings in remnant stumps of drowned forests in estuaries along the Oregon and Washington coast, and from meticulous Japanese written records of a devastating tsunami. By using computer simulations, scientists have demonstrated that the tsunami must have originated near the west coast of North America.

The giant Cascadia earthquake of 1700 was not a unique event. Studies of shoreline sediments from Oregon to Vancouver Island reveal a cyclic pattern of sudden downdrop of the coast, followed by a slow rebound in elevation. In the last 6000 years, 13 such great earthquakes have been documented. We are presently in the midst of a long span of episodic stress buildup between megathrust earthquake events, producing coastal uplift and a squeezing of Vancouver Island that is measurable by using modern survey techniques. The next offshore megathrust earthquake will happen, perhaps hundreds of years in the future, and residents of southwest British Columbia and adjacent areas need to be ready for it.

◆ **Figure 10.C** On June 23, 2010, tremors from a magnitude 5.0 earthquake centred near Val-des-Bois, Quebec (approximately 60 km north of Ottawa), were felt over much of eastern and southern Ontario and northeastern United States. Minor damage to buildings and roads were experienced within tens of kilometres of the epicentre. This bridge, located a few kilometres south of Val-des-Bois, collapsed as a direct result of the earthquake. (Photo by Jean Levac, *The Ottawa Citizen*)

The tectonically active west coast is not the only part of Canada that experiences earthquakes. Figure 10.B shows a map of earthquakes in Canada since 1980, catalogued as part of the National Earthquake Hazards Program of the Geological Survey of Canada. In western Canada, zones of earthquake activity are found in the Rocky Mountains of southwest Alberta and the Mackenzie Mountains of the Northwest Territories, where earthquakes of magnitudes up to 6.6 have been recorded. These bands of seismicity are thought to be the result of transmission of plate boundary stresses from convergent margins (Cascadia and Alaska), far into the North American plate.

Although very far from any active plate boundary, parts of eastern Canada also experience moderate earthquakes, such as the magnitude 5.9 Saguenay earthquake on November 25, 1988, and more the recent magnitude 5.0 Val-des-Bois earthquake on June 23, 2010 (Figure 10.C). The seismicity of eastern Canada is mainly concentrated along the Ottawa and St. Lawrence River valleys, and in the Appalachians of Quebec and New Brunswick. Like intraplate earthquakes in the United States, including the New Madrid earthquake sequence in 1811–1812, the cause of these earthquakes is not well understood. The zones of seismicity appear to be associated with the locations of ancient rifted margins or failed rift zones, which may rupture (reactivate) in response to the slow movement of North America away from the spreading mid-Atlantic ridge.

MEASURING THE SIZE OF EARTHQUAKES

Historically, seismologists have employed a variety of methods to obtain two fundamentally different measures that describe the size of an earthquake: intensity and magnitude. The first to be used was **intensity**—a measure of the degree of earthquake shaking at a given locale based on the amount of damage. With the development of seismographs, it was possible to obtain a quantitative measure of an earthquake based on seismic records rather than uncertain personal estimates of damage. The measurement that was developed, called **magnitude**, relies on calculations that use data provided by seismic records (and other techniques) to estimate the amount of energy released at the source of the earthquake.

As it turns out, both intensity and magnitude provide useful, although quite different, information about earthquake strength. Consequently, both measures are still used to describe the relative sizes of earthquakes.

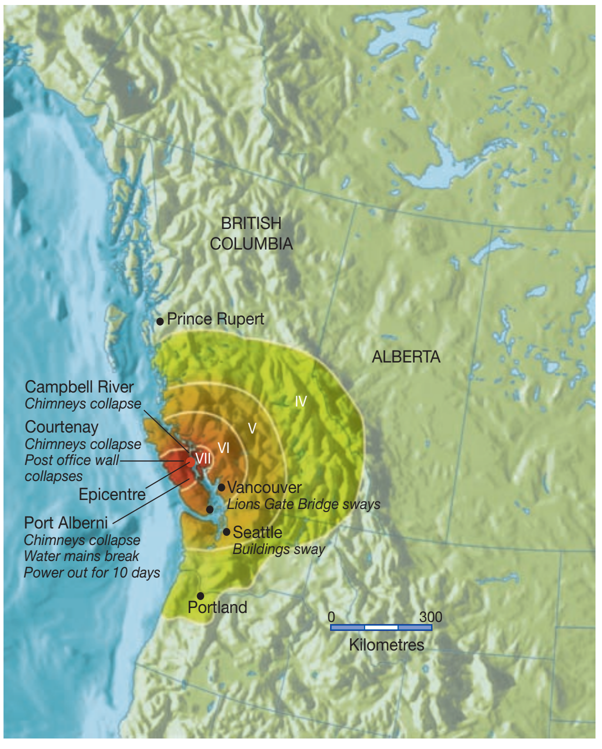
Intensity Scales

Until a little more than a century ago, historical records provided the only accounts of the severity of earthquake shaking and destruction. The use of these descriptions—which were compiled without any established standards for reporting—made accurate comparisons of earthquake sizes difficult,

TABLE 10.1 Modified Mercalli Intensity Scale

I	Not felt except by a very few under especially favourable circumstances.
II	Felt by a few persons at rest, especially on upper floors of buildings.
III	Felt quite noticeably indoors, especially on upper floors of buildings.
IV	During the day, felt indoors by many, outdoors by few. Sensation like heavy truck striking building.
V	Felt by most people, many awakened. Disturbances of trees, poles, and other tall objects sometimes noticed.
VI	Felt by all; many frightened and run outdoors. Some heavy furniture moved; few instances of fallen plaster or damaged chimneys. Damage slight.
VII	Everybody runs outdoors. Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable in poorly built or badly designed structures.
VIII	Damage slight in specially designed structures; considerable in ordinary substantial buildings with partial collapse; great in poorly built structures (fall of chimneys, factory stacks, columns, monuments, walls).
IX	Damage considerable in specially designed structures. Buildings shifted off foundations. Ground cracked conspicuously.
X	Some well-built wooden structures destroyed. Most masonry and frame structures destroyed. Ground badly cracked.
XI	Few, if any (masonry) structures remain standing. Bridges destroyed. Broad fissures in ground.
XII	Waves seen on ground surfaces. Objects thrown upward into air. Damage total.

at best. To standardize the study of earthquake severity, early earthquake investigators developed various intensity scales that considered damage done to buildings, as well as individual descriptions of the event, and secondary effects—landslides and the extent of ground rupture. By 1902, Giuseppe Mercalli had developed a relatively reliable intensity scale, which in a modified form is still used today. The **Modified Mercalli Intensity Scale** (Table 10.1) was developed by using California buildings as its standard, but it is appropriate for use throughout most of the United States and Canada to estimate the strength of an earthquake (see Figure 10.12). For example, if some well-built wooden structures and most masonry buildings are destroyed by an earthquake, a region would be assigned an intensity of X on the Mercalli scale (Table 10.1).



A

B

◆ **Figure 10.12** **A.** Modified Mercalli Intensity map showing intensity of shaking associated with an earthquake (magnitude 7.3) that occurred at 10:15 A.M. on June 23, 1946. The epicentre was in central Vancouver Island, near the communities of Courtenay and Campbell River. **B.** Damage to a school in Courtenay, British Columbia, resulting from the 1946 earthquake.
(Part **A** © Department of Natural Resources Canada. All rights reserved. Photo **B** reproduced with the permission of Natural Resources Canada, 2010)

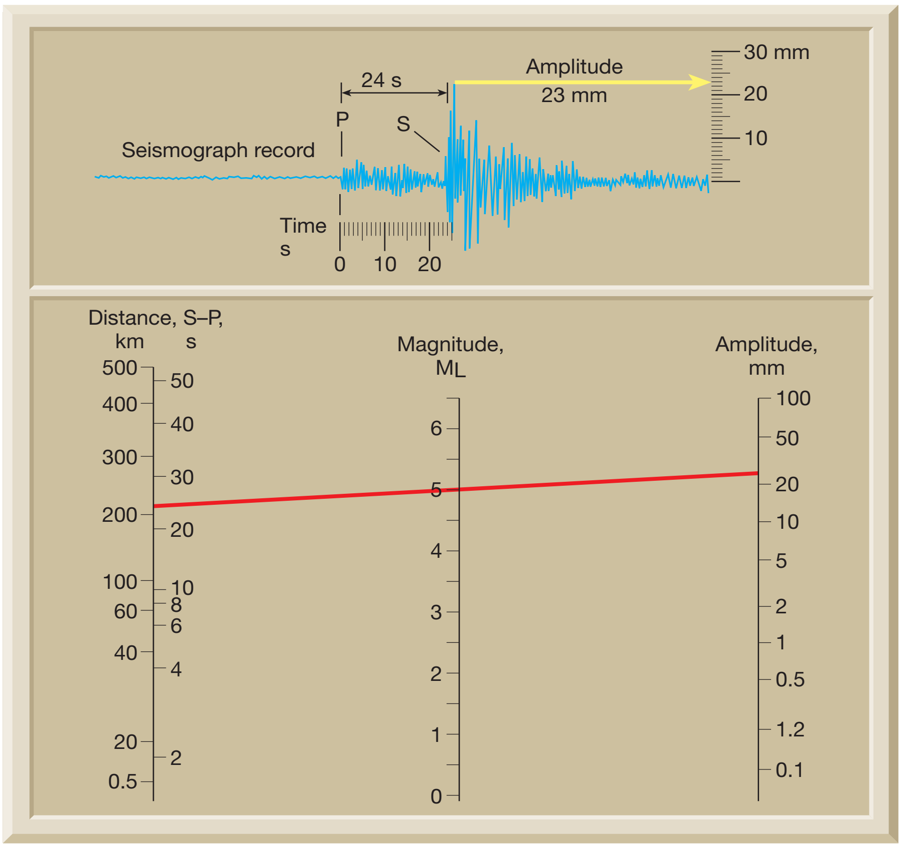
Despite their usefulness in providing seismologists with a tool to compare earthquake severity, particularly in regions without seismographs, intensity scales have severe drawbacks. In particular, intensity scales are based on effects of earthquakes (largely destruction) that depend not only on the severity of ground shaking but also on external factors, such as population density, building design, and the nature of surface materials. In British Columbia, for example, the 1946 Vancouver Island earthquake caused more property damage (Figure 10.12B) than the much stronger earthquake that occurred west of the Queen Charlotte Islands in 1949, largely because it affected a more densely populated area. Haiti (2010) experienced intensity of VII-IX on the Mercalli scale.

Magnitude Scales

To compare earthquakes across the globe, a measure was needed that does not rely on parameters that vary considerably from one part of the world to another, such as types of construction. As a consequence, the Richter scale was developed.

RICHTER MAGNITUDE In 1935, Charles Richter of the California Institute of Technology developed the first magnitude scale by using seismic records to estimate the relative sizes of earthquakes. As shown in Figure 10.13, the **Richter scale** is based on the amplitude of the largest seismic wave (P, S, or surface wave) recorded on a seismogram. Because seismic waves weaken as the distance between the earthquake epicentre and the seismograph increases (in a manner similar to light), Richter developed a method that accounted for the decrease in wave amplitude with increased distance. Theoretically, as long as the same, or equivalent, instruments were used, monitoring stations at various locations would obtain the same Richter magnitude for every recorded earthquake (Richter selected the Wood-Anderson seismograph as the standard recording device). In practice, however, different recording stations often obtained slightly different Richter magnitudes for the same earthquake—a consequence of the variations in rock types through which the waves travelled.

Although the Richter scale has no upper limit, the largest magnitude recorded on a Wood-Anderson seismograph was 8.9. These great shocks released an amount of energy that is roughly equivalent to the detonation of 1 billion tonnes of TNT. Conversely, earthquakes with a Richter magnitude of less than 2.0 are not felt by humans. Earthquakes vary enormously



◆ **Figure 10.13** Illustration showing how the Richter magnitude of an earthquake can be determined graphically by using a seismograph record from a Wood-Anderson instrument. First, measure the height (amplitude) of the largest wave on the seismogram (23 millimetres) and then the distance to the focus by using the time interval between S and P waves (24 seconds). Next, draw a line between the distance scale (left) and the wave amplitude scale (right). By doing this, you obtain the Richter magnitude (M_L) of 5.
(Data from California Institute of Technology)

TABLE 10.2 Earthquake Magnitude and Energy Equivalence

Earthquake Magnitude	Energy Released (millions of ergs)	Approximate Energy Equivalence
0	630,000	0.5 kilograms of explosives
1	20,000,000	
2	630,000,000	Energy of lightning bolt
3	20,000,000,000	
4	630,000,000,000	450 kilograms of explosives
5	20,000,000,000,000	
6	630,000,000,000,000	1946 Bikini atomic bomb test
		1994 Northridge earthquake
7	20,000,000,000,000,000	1989 Loma Prieta earthquake
8	630,000,000,000,000,000	1906 San Francisco earthquake
		1980 Eruption of Mount St. Helens
9	20,000,000,000,000,000,000	1964 Alaskan earthquake
		1960 Chilean earthquake
10	630,000,000,000,000,000,000	Annual U.S. energy consumption

Source: U.S. Geological Survey.
Note: For each unit increase in magnitude, the energy released increases 31.6 times.

in strength, and great earthquakes produce wave amplitudes that are thousands of times larger than those generated by weak tremors. To accommodate this wide variation, Richter used a *logarithmic scale* to express magnitude, where a *tenfold* increase in wave amplitude corresponds to an increase of 1 on the magnitude scale. Thus, the amount of ground shaking for a magnitude 5 earthquake is 10 times greater than that produced by a magnitude 4 earthquake.

In addition, each unit of Richter magnitude equates to roughly a *32-fold energy increase*. Thus, an earthquake with a magnitude of 6.5 releases 32 times as much energy as one with a magnitude of 5.5, and roughly 1000 times as much energy as one with a magnitude of 4.5 (Table 10.2). A major earthquake with a magnitude of 8.5 releases millions of times as much energy as the smallest earthquakes felt by humans.

OTHER MAGNITUDE SCALES Richter's original goal was modest in that he only attempted to rank the earthquakes of southern California (shallow-focus earthquakes) into groups of large, medium, and small magnitude. Hence, Richter magnitude was designed to study nearby (or local) earthquakes and is denoted by the symbol (M_L)—where M is for *magnitude* and L is for *local.*

The convenience of describing the size of an earthquake by a single number that could be calculated quickly from seismograms makes the Richter scale a powerful tool. Further, unlike intensity scales that could only be applied to populated areas of the globe, Richter magnitudes could be assigned to earthquakes in more remote regions and even to events that occurred in the ocean basins. As a result, the method devised by Richter was adapted to a number of different seismographs located throughout the world. In time, seismologists modified Richter's work and developed new magnitude scales. These modified scales were initially calibrated to be equivalent to the Richter scale and have contributed to efforts to measure the size of earthquakes. However, despite their usefulness, none of these scales is adequate for describing very large earthquakes.

MOMENT MAGNITUDE In recent years seismologists have been employing a more precise measure called **moment magnitude** (M_W), which is derived from the amount of displacement that occurs along a fault zone rather than by measuring the ground motion at some distant point. Moment magnitude is calculated by using a combination of factors that include the average amount of displacement along the fault, the area of the rupture surface, and the shear strength of the faulted rock—a measure of how much strain energy a rock can store before it suddenly slips and releases this energy in the form of an earthquake (and heat). For example, the energy involved in a 3-metre displacement of a rock body along a rupture a few hundred kilometres long would be much larger than that produced by a 1-metre displacement along a 10-kilometre-long rupture (assuming comparable rupture depths).

The moment magnitude can also be readily calculated from seismograms by examining very long period seismic waves. The values obtained have been calibrated so that small- and moderate-sized earthquakes have moment magnitudes that are roughly equivalent to Richter magnitudes. However, moment magnitudes are much better for describing very large earthquakes. For example, on the moment magnitude scale, the 1906 San Francisco earthquake, which had a surface-wave magnitude of 8.3, would be demoted to 7.9 on the moment magnitude scale, whereas the 1964 Alaskan earthquake with an 8.3 Richter magnitude would be increased to 9.2. The strongest earthquake on record is the 1960 Chilean earthquake with a moment magnitude of 9.5.

Moment magnitude has gained wide acceptance among seismologists and engineers because (1) it is the only magnitude scale that estimates adequately the size of very large earthquakes; (2) it is a measure established from the size of the rupture surface and the amount of displacement; thus, it better reflects the total energy released during an earthquake; and (3) it can be verified by two independent methods: field studies that are based on measurements of fault displacement and seismic methods by using long-period waves.

EARTHQUAKE DESTRUCTION

Many factors determine the degree of destruction that will accompany an earthquake. The most obvious is the magnitude of the earthquake and its proximity to a populated area. Fortunately, most earthquakes are small and occur in remote regions. However, about 20 major earthquakes are reported annually, one or two of which can be catastrophic.

Destruction from Seismic Vibrations

The 1964 Alaskan earthquake provided geologists with new insights into the role of ground shaking as a destructive force. As the energy released by an earthquake travels along Earth's surface, it causes the ground to vibrate in a complex manner by moving up and down as well as from side to side. The amount of structural damage attributable to the vibrations depends on several factors, including (1) the intensity of the vibrations, (2) the duration of the vibrations, (3) the nature of the material on which the structure rests, and (4) the design of the structure.

All of the multi-storey structures in Anchorage were damaged by the vibrations. The more flexible wood-frame residential buildings fared best. However, many homes were destroyed when the ground failed. A striking example of how construction variations affect earthquake damage is shown in Figure 10.14. You can see that the steel-frame building on the left withstood the vibrations, whereas the poorly designed J. C. Penney building was badly damaged. Engineers have learned that non-reinforced masonry buildings are the most serious safety threat in earthquakes (see also Chapter opening photo and Figure 10.1A, B).

Most large structures in Anchorage were damaged, even though they were built according to the earthquake building standards of the day. Perhaps some of that destruction can be attributed to the unusually long duration of this earthquake. Most quakes consist of tremors that last less than a minute. The Alaska quake was a notable exception: it reverberated for three to four minutes.

AMPLIFICATION OF SEISMIC WAVES Although the region within 20 to 50 kilometres of the epicentre will experience about the same intensity of ground shaking, the destruction varies considerably within

◆ **Figure 10.14** Damage caused to the five-storey J. C. Penney Co. building, Anchorage, Alaska. Very little structural damage was incurred by the adjacent building.
(Courtesy of NOAA/Seattle)

this area. This difference is mainly attributable to the nature of the ground on which the structures are built. Soft sediments, for example, generally amplify the vibrations more than solid bedrock. Thus, the buildings located in Anchorage, which were situated on unconsolidated sediments, experienced heavy structural damage. By contrast, most of the nearby town of Whittier, although much closer to the epicentre, rests on a firm foundation of granite and hence suffered much less damage. However, Whittier was damaged by a *tsunami* (described in the next section).

An earthquake that occurred in Mexico in 1985 gave seismologists and engineers a vivid reminder of what had been learned from the 1964 Alaskan earthquake. The Mexican coast, where the earthquake was centred, experienced unusually mild tremors despite the strength of the quake. As expected, the seismic waves became progressively weaker with increasing distance from the epicentre. However, in the central section of Mexico City, nearly 400 kilometres from the source, the vibrations intensified to five times that experienced in outlying districts. Much of this amplified ground motion can be attributed to soft sediments, remnants of an ancient lake bed, that underlie portions of the city. The lake bed sediments were shaken by the earthquake, causing buildings to also shake and collapse (Figure 10.15).

LIQUEFACTION In areas where unconsolidated materials are saturated with water, earthquake vibrations can generate a phenomenon known as **liquefaction** (*liqueo* = to be fluid, *facio* = to make). Under these conditions, what had been a stable soil turns into a mobile fluid that is not capable of supporting buildings or other structures (Figure 10.15). As a result, underground objects, such as storage tanks and sewer lines, can literally float to the surface of their newly liquefied environment. Buildings and other structures can settle and collapse. During the 1989 Loma Prieta earthquake in California, foundations failed and geysers of sand and water shot from the ground, indicating that liquefaction had occurred (Figure 10.16).

Tsunami

Most deaths associated with the 1964 Alaskan quake were caused by a tsunami* (*tsu* = harbour, *nami* = waves) or seismic sea wave. Before the catastrophic tsunami of December 26, 2004 (Box 10.2), these

**Seismic sea waves were given the name* tsunami *by the Japanese, who have suffered a great deal from them. The term* tsunami *is now used worldwide.*

◆ **Figure 10.15** Effects of liquefaction. This tilted building rests on unconsolidated sediment that behaved like quicksand during the 1985 Mexican earthquake. (Photo by James L. Beck)

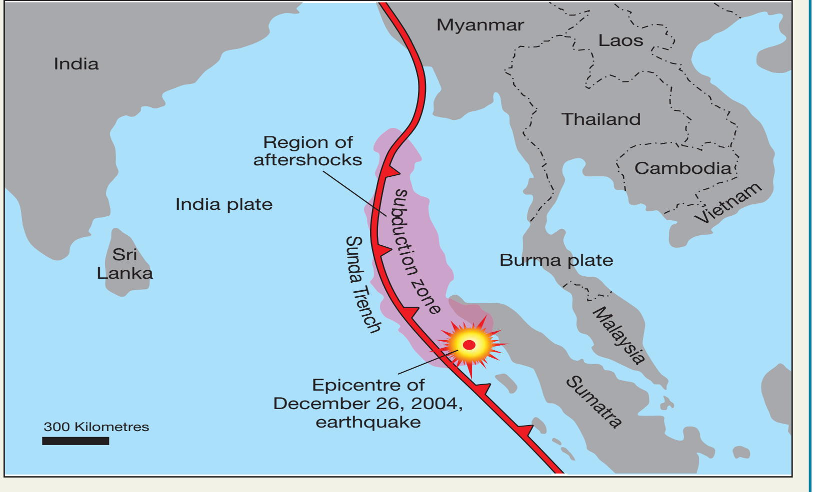
◆ **Figure 10.16** These "mud volcanoes" were produced by the Loma Prieta earthquake of 1989 in California. They formed when fountains of sand and water shot from the ground, an indication that liquefaction had occurred. (Photo by Richard Hilton, courtesy of Dennis Fox)

BOX 10.2 UNDERSTANDING EARTH

2004 Boxing Day Tsunami

On December 26, 2004, the second-largest earthquake ever recorded by seismographs (9.3 M_W) occurred 150 kilometres off the northwest coast of Sumatra (Figure 10.D). The resulting tsunami damaged 11 countries around the Indian Ocean and Andaman Sea, claimed more than 200,000 human lives, and left more than 1 million people homeless (Figure 10.E). At the epicentre, located near the Sunda Trench, the ocean floor was thrust up to 5 metres and generated a wave up to 50 metres high. This wall of water smashed into the coasts of the Seychelles Islands and Somalia, some 2000 kilometres away, and was even recorded by tide gauges in California and New Jersey. The south and east coasts of Sri Lanka and India were hit hard as their shorelines suddenly pulled back (Figure 10.F), exposing a large swath of sea bed as the tsunami crested then surged its way 2 kilometres inland, destroying property and people.

The earthquake focus was approximately 10 kilometres beneath the seafloor, generated by slippage deep beyond the Sunda Trench where the Australian-Indian plate subducts under the Burma subplate, part of the Eurasian plate. The epicentre is located near the intersection of two plates (Eurasian and Australian-Indian) that has generated several major earthquakes and tsunami, including the 1941 tsunami that killed 5000 people along the east coast of India. Interactions between these plates were also responsible for major volcanic eruptions including Krakatoa, Tambora, and Toba. In spite of this record,



◆ **Figure 10.D** This image highlights the crests (red) and troughs (blue) of the tsunami generated by the great Sumatra earthquake of 2004. To the east of the Andaman and Nicobar Islands, the waves travelled across the Andaman Sea to hit Thailand and Malaysia. Simultaneously, the waves that were generated westward into the Indian Ocean hit Sri Lanka (as depicted here),India, and the eastern coast of Africa.
(Modified from an image courtesy of the Geological Survey of Japan, AIST).

◆ **Figure 10.E** The village of Banda Aceh on the north shore of Sumatra before **(A)** and after **(B)** being hit by the 2004 tsunami.
(Photo by National Oceanic and Atmospheric Administration)

A

B

◆ **Figure 10.F** Aerial view of Kalutara (south coast of Sri Lanka) before **(A)** and during **(B)** shoreline withdrawal just before the tsunami surged inland. (Photos courtesy of Digital Globe/Getty Images)

countries were taken by surprise when the Boxing Day tsunami hit. This region lacked the tide gauges, wave sensors, regional tsunami warning system, and communications network that the Pacific region has (see Box 10.3). By the time the Pacific network became aware of the disaster, it was too late to warn the Indian region, which underscored the need for a similar network. In 2006 a DART tsunami buoy sensor was deployed in the Indian Ocean to collect wave data; by 2010 three buoys and 17 seismic stations were installed around the region.

destructive waves were commonly called tidal waves by the media. However, that name is incorrect; the waves are generated by earthquakes, not the tidal effect of the Moon or Sun.

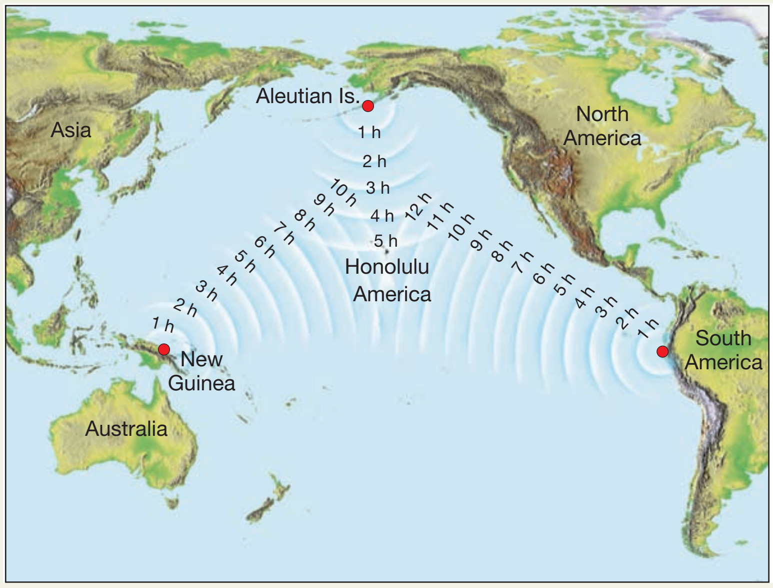
Most tsunami result from vertical displacement along a fault located on the ocean floor (Figure 10.17), or some other severe undersea disturbance, such as a large volcanic eruption or submarine landslide. Once created, a tsunami resembles the ripples formed when a pebble is dropped into a pond. In contrast to ripples, tsunami advance across the ocean at amazing speeds: between 500 and 950 kilometres per hour. Despite this striking characteristic, a tsunami in the open ocean can pass undetected because its height is usually less than 1 metre and the distance between wave crests is great, ranging from 100 to 700 kilometres. However, on entering shallower coastal waters, these destructive waves are slowed down and the water begins to pile up to heights that occasionally exceed 30 metres (Figure 10.17). As the crest of a tsunami approaches the shore, it appears as a rapid rise in sea level, with a turbulent and chaotic surface. Tsunami can be very destructive (see Boxes 10.2 and 10.3). Canadian examples include the 1929 tsunami that killed 28 people and damaged 40 villages along the Burin Peninsula of southern Newfoundland and Labrador (Figure 10.18A), and 1964 tsunami (generated by the great Alaskan earthquake) that caused extensive damage to Port Alberni

BOX 10.3 **PEOPLE AND THE ENVIRONMENT**

Tsunami Warning System

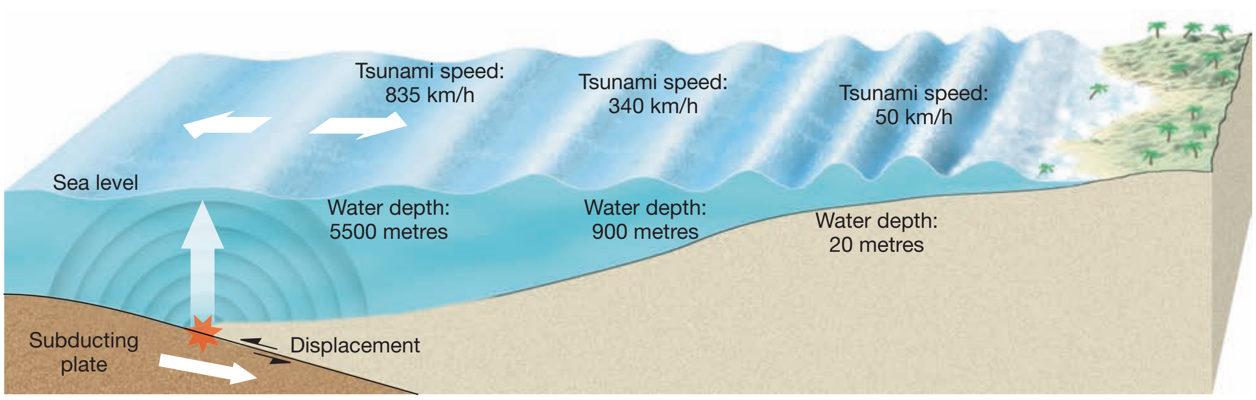
Tsunami traverse large stretches of ocean before their energy is fully dissipated. The tsunami generated by a 1960 Chilean earthquake, in addition to destroying villages along an 800-kilometre stretch of coastal South America, travelled 17,000 kilometres across the Pacific to Japan. Here, about 22 hours after the quake, considerable damage occurred in south coastal villages. For several days afterward, tidal gauges in Hilo, Hawaii, detected these diminishing waves as they reverberated like echoes about the Pacific.

In 1946 a large tsunami struck the Hawaiian Islands without warning. A wave more than 15 metres high left several coastal villages in shambles. This destruction motivated the National Oceanic and Atmospheric Administration to establish a tsunami warning system for coastal areas of the Pacific. From seismic observatories throughout the region, large earthquakes are reported to the Pacific Tsunami Warning Center at Ewa Beach (near Honolulu), Hawaii. Scientists at the Center use tidal gauges to determine whether a tsunami has been formed. Within an hour a warning is issued. Although tsunami travel very rapidly, there is sufficient time to evacuate all but the region nearest the epicentre (Figure 10.G). For example, a tsunami generated near the Aleutian Islands would take 5 hours to reach Hawaii, and one generated near the coast of Chile would travel 15 hours before reaching Hawaii.



◆ **Figure 10.G** Tsunami travel times to Honolulu, Hawaii, from selected locations throughout the Pacific.
(Data from NOAA)

Fortunately, most earthquakes do not generate tsunami. On the average, only about 1.5 destructive tsunami are generated worldwide annually. Of these, only about 1 every 10 years is catastrophic.



◆ **Figure 10.17** Schematic drawing of a tsunami generated by displacement of the ocean floor. The speed of a wave correlates with ocean depth. As shown, waves moving in deep water advance at speeds in excess of 800 kilometres per hour. Speed gradually slows to 50 kilometres per hour at depths of 20 metres. Decreasing depth slows the movement of the wave. As waves slow in shallow water, they grow in height until they topple and rush onto shore with tremendous force. The size and spacing of these swells are not to scale.

A

B

◆ **Figure 10.18** Examples of destruction by tsunami in Canada. **A.** Debris from houses, wharfs, boats, and fishing gear along the shore of the Burin Peninsula, Newfoundland and Labrador, following the tsunami generated by the Grand Banks earthquake of 1929. **B.** Damage to buildings associated with the tsunami that hit Port Alberni, British Columbia, in 1964. (Photo **A** courtesy of the Rooms Provincial Archives, A1–141/Parsons & Sons; photo **B** courtesy of UBC Library, Rare Books & Special Collections, neg # AAC-15402)

on Vancouver Island (Figure 10.18B). The Burin tsunami was triggered by a magnitude 7.2 earthquake (epicentre 250 kilometres south of Newfoundland and Labrador) that caused a turbidity current of 200 cubic kilometres to rush down the continental slope of the Grand Banks while rupturing 12 transatlantic cables and generating the tsunami. The wave rose to 27 metres and also reached the coasts of South Carolina and Portugal.

Usually, the first warning of an approaching tsunami is a relatively rapid withdrawal of water from beaches (see Box 10.2). Coastal residents have learned to heed this warning and move to higher ground because about 5 to 30 minutes later, the retreat of water is followed by a surge capable of extending hundreds of metres inland. In a successive fashion, each surge is followed by rapid oceanward retreat of the water.

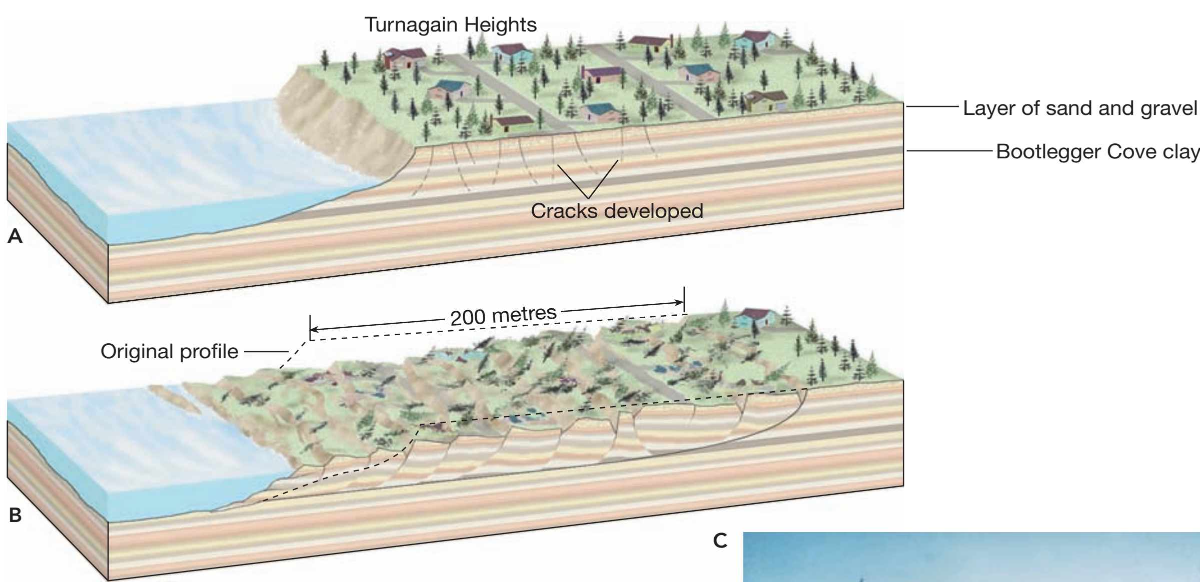
Landslides and Ground Subsidence

In the 1964 Alaskan earthquake, the greatest damage to structures was from landslides and ground subsidence triggered by the vibrations. At Valdez and Seward, the violent shaking caused deltaic materials to experience liquefaction; the subsequent slumping carried both waterfronts away. Because of the threat of recurrence, the entire town of Valdez was relocated about 7 kilometres away on more stable ground. In Valdez, 31 people on a dock died when it slid into the sea.

Most of the damage in the city of Anchorage was also attributed to landslides. Many homes were destroyed in Turnagain Heights when a layer of clay lost its strength and more than 80 hectares of land slid toward the ocean (Figure 10.19). A portion of this spectacular landslide was left in its natural condition as a reminder of this destructive event. The site was appropriately named Earthquake Park. Downtown Anchorage was also disrupted as sections of the main business district dropped by as much as 3 metres.

Fire

The famous 1906 earthquake in San Francisco reminds us of the formidable threat of fire. The



C


◆ **Figure 10.19** Turnagain Heights slide caused by the 1964 Alaskan earthquake. **A.** Vibrations from the earthquake caused cracks to appear near the edge of the bluff. **B.** Within seconds blocks of land began to slide toward the sea on a weak layer of clay. In less than five minutes, as much as 200 metres of the Turnagain Heights bluff area had been destroyed. **C.** Photo of a small portion of the Turnagain Heights slide.
(Photo **C** courtesy of U.S. Geological Survey)

central city contained mostly large older wooden structures and brick buildings. Although many nonreinforced brick buildings were extensively damaged by vibrations, the greatest destruction was caused by fires when gas and electrical lines were severed. The fires raged out of control for three days and devastated more than 500 blocks of the city. The problem was compounded by the initial ground shaking, which broke the city's water lines into hundreds of unconnected pieces.

The fire was finally contained when buildings were dynamited along a wide boulevard to provide a firebreak, the same strategy used in fighting a forest fire. Although only a few deaths were attributed to the San Francisco fire, such is not always the case. A 1923 earthquake in Japan triggered an estimated 250 fires, which devastated the city of Yokohama and destroyed more than half the homes in Tokyo. More than 100,000 deaths were attributed to the fires, which were driven by unusually high winds.

CAN EARTHQUAKES BE PREDICTED?

The earthquake that shook Northridge, California, on January 17, 1994, was brief (about 40 seconds) and only of moderate rating (M_W 6.7), but it killed 57 people and caused about US$40 billion in damage. The amount of devastation caused by the earthquake was largely due to the fact that the area it affected was densely populated and highly developed. Seismologists warn that earthquakes of comparable or greater strength will occur along the San Andreas Fault system, which cuts a 1300-kilometre path through the state. The obvious question is, can earthquakes be predicted?

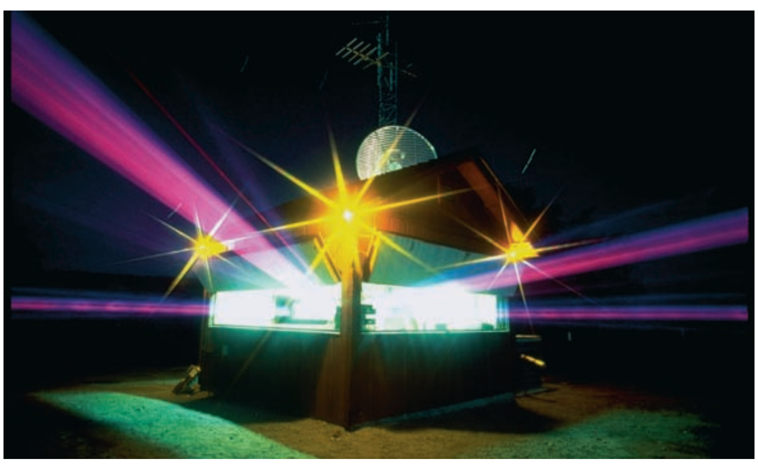
Short-Range Predictions

The goal of short-range earthquake prediction is to provide a warning of the location and magnitude of a large earthquake within a narrow time frame. We currently do not have an *absolutely* reliable method for making short-range earthquake predictions. However, substantial efforts to achieve this objective are being put forth in Japan, the United States, China, and Russia—countries where earthquake risks are high. This research has concentrated on monitoring possible *precursors*—phenomena that precede and thus provide a warning of a forthcoming earthquake. In California, for example, seismologists are measuring uplift, subsidence, and strain in the rocks near active faults. Some Japanese scientists are studying anomalous animal behaviour that may precede a quake. Other researchers are monitoring changes in groundwater levels, while still others are trying to predict earthquakes based on changes in the electrical conductivity of rocks.

Among the most ambitious earthquake experiments is one being conducted along a segment of the San Andreas Fault near the town of Parkfield in central California. Here, earthquakes of moderate intensity have occurred on a regular basis about once every 22 years since 1857. The most recent rupture was a magnitude 5.6 quake that occurred in 1966. With the next event already significantly overdue, the U.S. Geological Survey has established an elaborate monitoring network. Included are creepmeters, tiltmeters, and borehole strain meters that are used to measure the accumulation and release of strain. Seventy seismographs of various designs have been installed to record foreshocks and the main event. Finally, a network of distance-measuring devices that employ lasers measures movement across the fault (Figure 10.20). The object is to identify ground movements that may precede a sizeable rupture.

Long-Range Forecasts

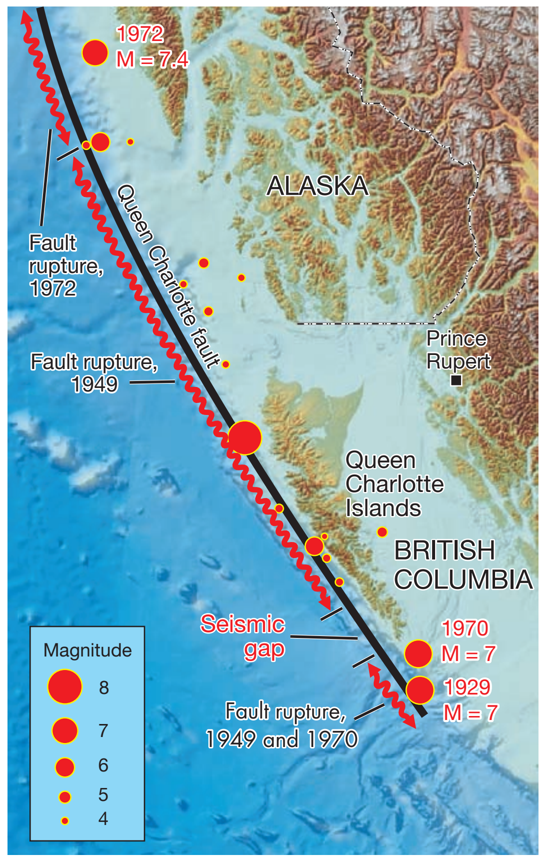
In contrast to short-range predictions, which aim to predict earthquakes within hours or at most days, long-range forecasts give the probability of a certain magnitude earthquake occurring on a time scale of 30 to 100 years or more. Long-range forecasts are based on the premise that earthquakes are repetitive or cyclic, like the weather. In other words, as soon as one earthquake is over, the continuing motions of Earth's plates begin to build stress in the rocks again until they fail once more. This has led seismologists to study historical records of earthquakes, looking for any discernible patterns so that the probability of recurrence might be established.



◆ **Figure 10.20** Lasers used to measure movement along the San Andreas Fault.
(Photo by John K. Nakata/U.S. Geological Survey)

With this concept in mind, a group of seismologists plotted the distribution of rupture zones associated with great earthquakes that have occurred in the seismically active regions of the Pacific Basin. The maps revealed that individual rupture zones tended to occur adjacent to one another without appreciable overlap, thereby tracing out a plate boundary. Recall that most earthquakes are generated along plate boundaries by the relative motion of large lithospheric blocks. Because plates are in continual motion, the researchers predicted that over one or two centuries, major earthquakes would occur along each segment of the Pacific plate boundary.

When the researchers studied historical records, they discovered that some zones had not produced a great earthquake in more than a century. These quiet zones, called **seismic gaps**, were identified as probable sites for major earthquakes in the next few decades (Figure 10.21).



◆ **Figure 10.21** Map showing epicentres of large shallow earthquakes recorded from 1929 to 1972 along the northwest coast of British Columbia and the fault ruptures with which they were associated. The seismic gap located immediately west of the southern Queen Charlotte Islands denotes the most likely location for the next large earthquakes along the Queen Charlotte Fault.
(Courtesy of Tricouni Press Ltd.)

Another method of long-term forecasting, known as *paleoseismology* (*palaios* = ancient, *seismos* = shake, *ology* = the study of), has been implemented. One technique involves studying layered deposits that were offset by prehistoric seismic disturbances. Researchers recently discovered strong evidence that very powerful earthquakes (magnitude of 8 or larger) have repeatedly struck the Pacific Northwest over the past several thousand years. The most recent event occurred about 300 years ago. As a result of these findings, public officials have taken steps to strengthen some of the region's existing dams, bridges, and water systems.

In summary, it appears that the best prospects for making useful earthquake predictions involve forecasting magnitudes and locations on time scales of years or perhaps even decades. These forecasts are important because they provide information used to develop earthquake building standards and to assist in land-use planning.

PROBING EARTH'S INTERIOR

GEODe:Earth

Earth's Interior

Earth's Layered Structure

Earth's interior lies just beneath us. However, direct access to it remains very limited. Wells drilled into the crust in search of oil, gas, and other natural resources have generally been confined to the upper 7 kilometres—only a small fraction of Earth's 6370-kilometre radius. Even the Kola Well, a super-deep research well located in a remote northern outpost of Russia, has penetrated to a depth of only 12.3 kilometres. Although volcanic activity is considered a window into Earth's interior because materials are brought up from below, it allows a glimpse of only the outer 200 kilometres of our planet.

Fortunately, geologists have learned a great deal about Earth's composition and structure through computer modelling, high-pressure laboratory experiments, and samples of the solar system (meteorites) that collide with Earth. In addition, many clues to the physical conditions inside our planet have been acquired through the study of seismic waves generated by earthquakes, nuclear explosions, and weaker, human-produced vibrations. As seismic waves pass through Earth, they carry information to the surface about the materials through which they were transmitted. Hence, when carefully analyzed, seismic records provide an image of materials below the surface. The Canadian LITHOPROBE project, for example, has yielded exciting insights on the deep lithosphere from the processing of seismic data (see Box 10.4).

Much of our knowledge of Earth's interior comes from the study of earthquake waves that penetrate Earth and emerge at some distant point. The technique involves accurately measuring the time required for P and S waves to travel from an earthquake or nuclear explosion to a seismic station. Because the time required for P and S waves to

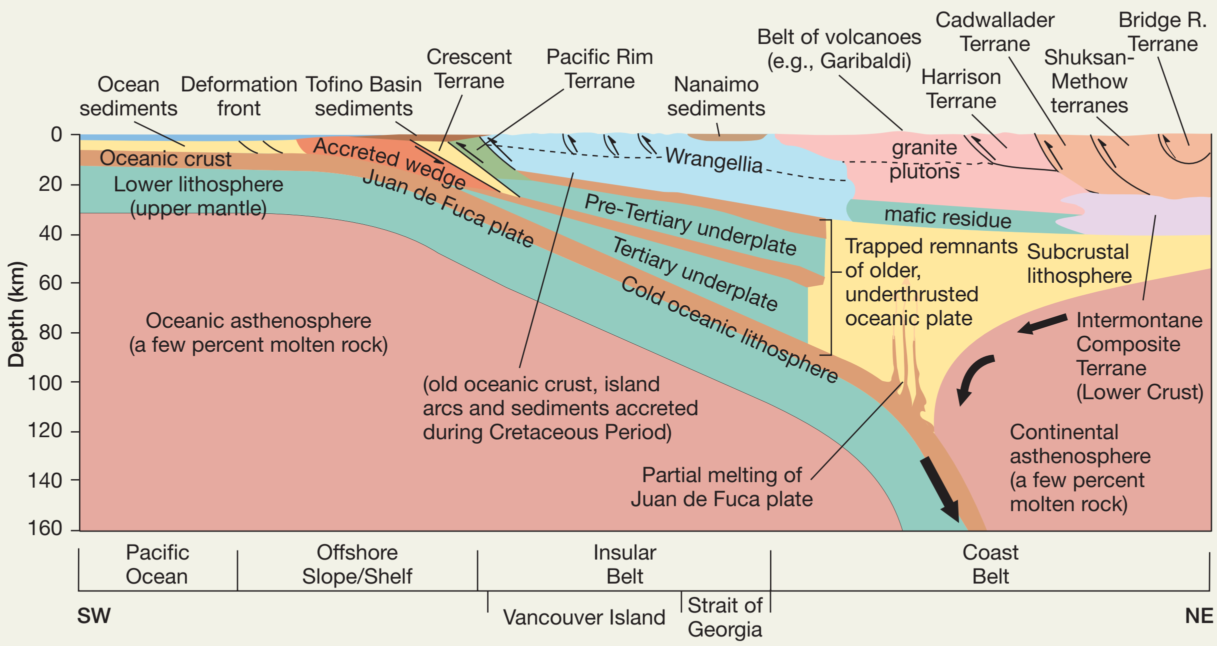
BOX 10.4 CANADIAN PROFILE

Lithoprobe: Probing the Depths of Canada

David Eaton*

Unlike the oceanic crust, continents preserve a record of Earth's early history, including how ancient mountains formed and how the continents grew and developed over time. Geoscientists are faced with a daunting challenge, however: to reconstruct this tectonic record from fragmentary evidence provided by rocks exposed at the surface. Canada's LITHOPROBE project, which commenced in 1984, is a bold multi-institutional endeavour to unravel the tectonic evolution of North America with a focus on the critical third dimension of geology: depth. With collaboration by more than 900 scientists working in academia, government agencies, and the petroleum and mining industries, LITHOPROBE has involved a significant segment of Canada's Earth science community in a coordinated research effort. The research program was divided into a series of transects that collectively span the evolution of the North American continent in time (about 4 billion years) and space.

To achieve adequate resolution of geologic features down to the base of the crust (approximately 40 kilometres) and the upper mantle, LITHOPROBE research has been spearheaded by multichannel seismic-reflection (MSR) profiling. Adapted from the petroleum industry, where it is used extensively for exploration of the top 10 kilometres of sedimentary basins, this technique makes use of P waves, usually generated mechanically by using vibroseis trucks, to record faint echoes produced at buried geologic contacts, faults, and shear zones. Much like medical imaging techniques, the seismic data are processed by computer to enhance the seismic reflections, which produces cross-sectional images that can



◆ **Figure 10.H** Part of the Canadian LITHOPROBE project involved the study of the lithospheric structure of western Canada. Among the exciting discoveries was the detection of lithospheric fragments (underplates) that remain lodged under the west coast region and have contributed to the growth of mountain roots.
(Modified from image courtesy of R. M. Clowes)

be compared with geologic data to reveal the underlying anatomy of the lithosphere. Specialists in all areas of geology, geophysics, and geochemistry have collaborated on major geotectonic problems by using techniques that include basic mapping, precise geochronology, electromagnetic studies, seismic refraction, and paleomagnetism.

LITHOPROBE's pilot study on Vancouver Island demonstrated that MSR profiling methods can be used to map layers near the top of the descending Juan de Fuca slab (Figure 10.H). It showed slabs of lithosphere welded to the continental lithosphere of western British Columbia as *underplates.* Results from this modern subduction system (Cascadia) have provided a template for understanding ancient subduction systems. Furthermore, subduction "scars" have been detected in the mantle beneath Archean rocks of the Superior craton in Quebec and the Proterozoic Great Bear Arc in the Northwest Territories. These results establish that the subduction occurred in some areas in the late Archean and support the concept that subducted slabs can be accreted to the base of continental plates as part of the process of formation of deep lithospheric roots.

LITHOPROBE's main research focused on ancient shield and platform regions where North America is known to have grown via a series of mountain-building events. LITHOPROBE profiles have allowed geoscientists to trace structural contacts down to the base of the crust (Moho). This work has led to the discovery of previously unrecognized mountain belts that lie buried beneath undeformed sedimentary rocks.

For further information visit **www.lithoprobe.ca**.

**David Eaton is a professor in the Department of Earth Sciences at the University of Calgary.*

travel through Earth depends on the properties of the materials encountered, seismologists search for variations in travel times that cannot be accounted for simply by differences in the distances travelled. These variations correspond to changes in the properties of the materials encountered.

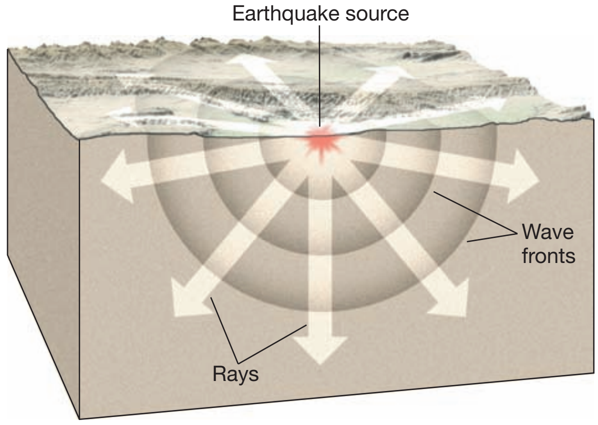
One major problem is that to obtain accurate travel times, seismologists must establish the exact location and time of an earthquake. This is commonly a difficult task because most earthquakes occur in remote areas. By contrast, the time and location of a nuclear test explosion is always known precisely. Despite the limitations of studying seismic waves generated by earthquakes, seismologists in the first half of the twentieth century were able to use them to determine the major layers of Earth. It was not until the early 1960s, when nuclear testing was in its heyday and networks consisting of hundreds of sensitive seismographs were deployed, that the finer structures of Earth's interior were established with certainty.

The Nature of Seismic Waves

To examine Earth's composition and structure, we must first study some of the basic properties of wave transmission, or *propagation*. As stated earlier in this chapter, seismic energy travels out from its source in all directions as waves. For purposes of description, the common practice is to consider the paths taken by these waves as *rays*, or lines drawn perpendicular to the wave front, as shown in Figure 10.22. Significant characteristics of seismic waves include the following:

- The velocity of seismic waves depends on the density and elasticity of the intervening material. Seismic waves travel most rapidly in rigid materials that elastically spring back to their original shapes when the stress caused by a seismic wave is removed. For instance, crystalline rock transmits seismic waves more rapidly than does a layer of unconsolidated mud.
- Within a given layer, the speed of seismic waves generally increases with depth because pressure increases and squeezes the rock into a more compact elastic material.
- Compressional waves (P waves), which vibrate back and forth in the same line as their direction of travel, are able to propagate through liquids as well as solids because when compressed, these materials behave elastically; that is, they resist a change in volume and, like a rubber band, return to their original shape as a wave passes.
- Shear waves (S waves), which vibrate at right angles to their direction of travel, cannot propagate through liquids because, unlike solids, liquids have no shear strength. That is, when liquids are subjected to forces that act to change their shapes, they simply flow.
- In all materials, P waves travel faster than S waves.
- When seismic waves pass from one material to another, the path of the wave is refracted (bent).* In addition, some of the energy is reflected from the **discontinuity** (the boundary between the two dissimilar materials). This is similar to what happens to light when it passes from air into water.

Thus, depending on the nature of the layers through which they pass, seismic waves speed up or slow down and can be refracted (bent) or reflected.

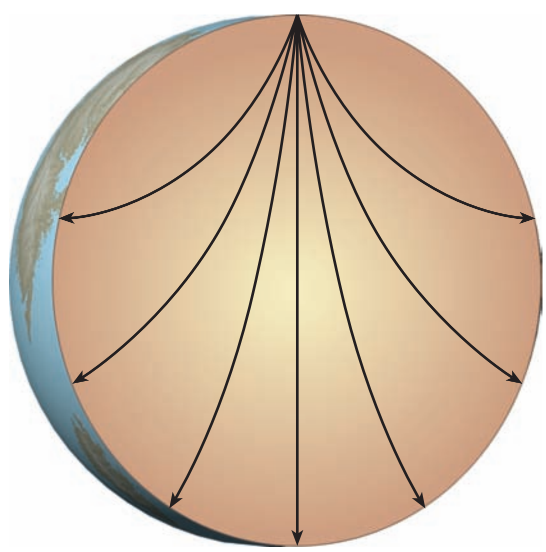


◆ **Figure 10.22** Seismic energy travels in all directions from an earthquake source (focus). The energy can be portrayed as expanding wave fronts or as rays drawn perpendicular to the wave fronts.

These measurable changes in seismic wave motions enable seismologists to probe Earth's interior.

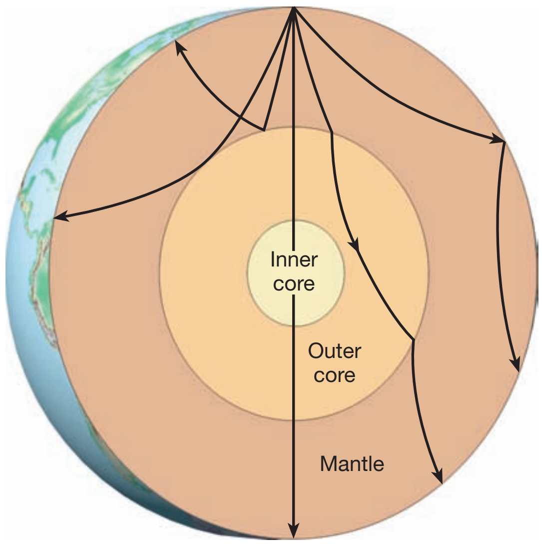
Seismic Waves and Earth's Structure

If Earth were a perfectly homogeneous body, seismic waves would spread through it in all directions. Such seismic waves would travel in a straight line at a constant speed. However, this is not the case for Earth. It so happens that the seismic waves reaching seismographs located farther from an earthquake travel at faster average speeds than those recorded at locations closer to the event. This general increase in speed with depth is a consequence of increased pressure, which enhances the elastic properties of deeply



◆ **Figure 10.23** Wave paths through a planet where velocity increases with depth.

**Refraction occurs provided that the rnay is not travelling perpendicular to the boundary.*



◆ **Figure 10.24** A few of the many possible paths that seismic rays take through Earth.

buried rock. As a result, the paths of seismic rays through Earth are refracted in the manner shown in Figure 10.23.

As more sensitive seismographs were developed, it became apparent that in addition to gradual changes in seismic-wave velocities, rather abrupt velocity changes also occur at particular depths. Because these discontinuities were detected worldwide, seismologists concluded that Earth must be composed of distinct shells that have varying compositions or mechanical properties (Figure 10.24; see also Figure 1.19).

DISCOVERING EARTH'S MAJOR BOUNDARIES

Over the past century, seismic data gathered from many seismic stations have been compiled and analyzed. From this information, seismologists have developed a detailed image of Earth's interior (Figure 1.19). This model is continually being fine-tuned as more data become available and as new seismic techniques are employed. Furthermore, laboratory studies that experimentally determine the properties of various Earth materials under the extreme environments found deep in Earth add to this body of knowledge.

The Crust–Mantle Boundary (the Moho)

In 1909 a pioneering Yugoslavian seismologist, Andrija Mohorovičić, presented the first convincing evidence for layering within Earth. The boundary he discovered separates crustal materials from rocks of different composition in the underlying mantle, and it was named the **Mohorovičić discontinuity** in his honour. The name for this boundary was quickly shortened to **Moho**.

By carefully examining the seismograms of shallow earthquakes, Mohorovičić found that seismic stations located more than 200 kilometres from an earthquake obtained appreciably faster average travel velocities for P waves than did stations located nearer the quake (Figure 10.25). In particular, P waves that reached the closest stations first had velocities that averaged about 6 kilometres per second. By contrast, the seismic energy recorded at more distant stations travelled at speeds that approached 8 kilometres per second. This abrupt jump in velocity did not fit the general pattern that had been previously observed. From these data, Mohorovičić concluded that below 50 kilometres, a layer exists with properties markedly different from those of Earth's outer shell.

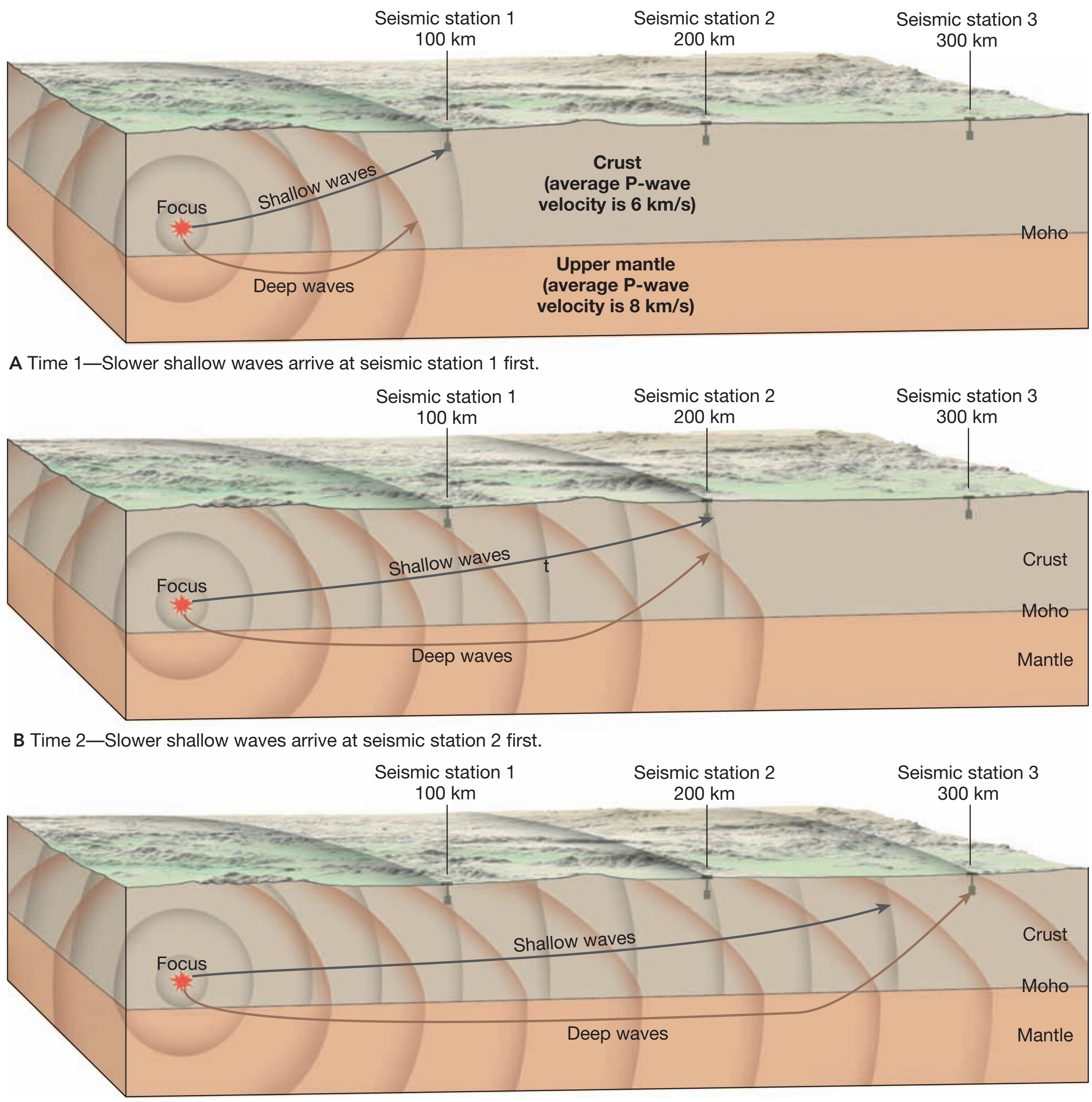
Figure 10.25 illustrates how Mohorovičić reached this important conclusion. Notice that the first wave to reach the seismograph located 100 kilometres from the epicentre travelled the shortest route directly through the crust. However, at the seismograph located 300 kilometres from the epicentre, the first P wave to arrive travelled through the mantle, a zone of higher velocity. Thus, although this wave travelled a greater distance, it reached the recording instrument sooner than did the rays taking the more direct route. This is because a large portion of its journey was through a region having a composition where seismic waves travel more rapidly. This principle is analogous to a driver taking a bypass route around a large city during rush hour. Although this alternative route is longer, it may be faster.

The Core–Mantle Boundary

A few years later, in 1914, the location of another major boundary was established by the German seismologist Beno Gutenberg.* This discovery was based primarily on the observation that P waves diminish and eventually die out completely about 105° from an earthquake (Figure 10.26). Then, about 140° away, the P waves reappear, but about two minutes later than would be expected based on the distance travelled. This belt, where direct seismic waves are absent, is about 35° wide and has been named the **P-wave shadow zone**** (Figure 10.26).

**The core–mantle boundary had been predicted by R. D. Oldham in 1906, but his arguments for a central core did not receive wide acceptance.*

***As more sensitive instruments were developed, weak and delayed P waves that enter this zone via reflection were detected.*



A Time 1—Slower shallow waves arrive at seismic station 1 first.

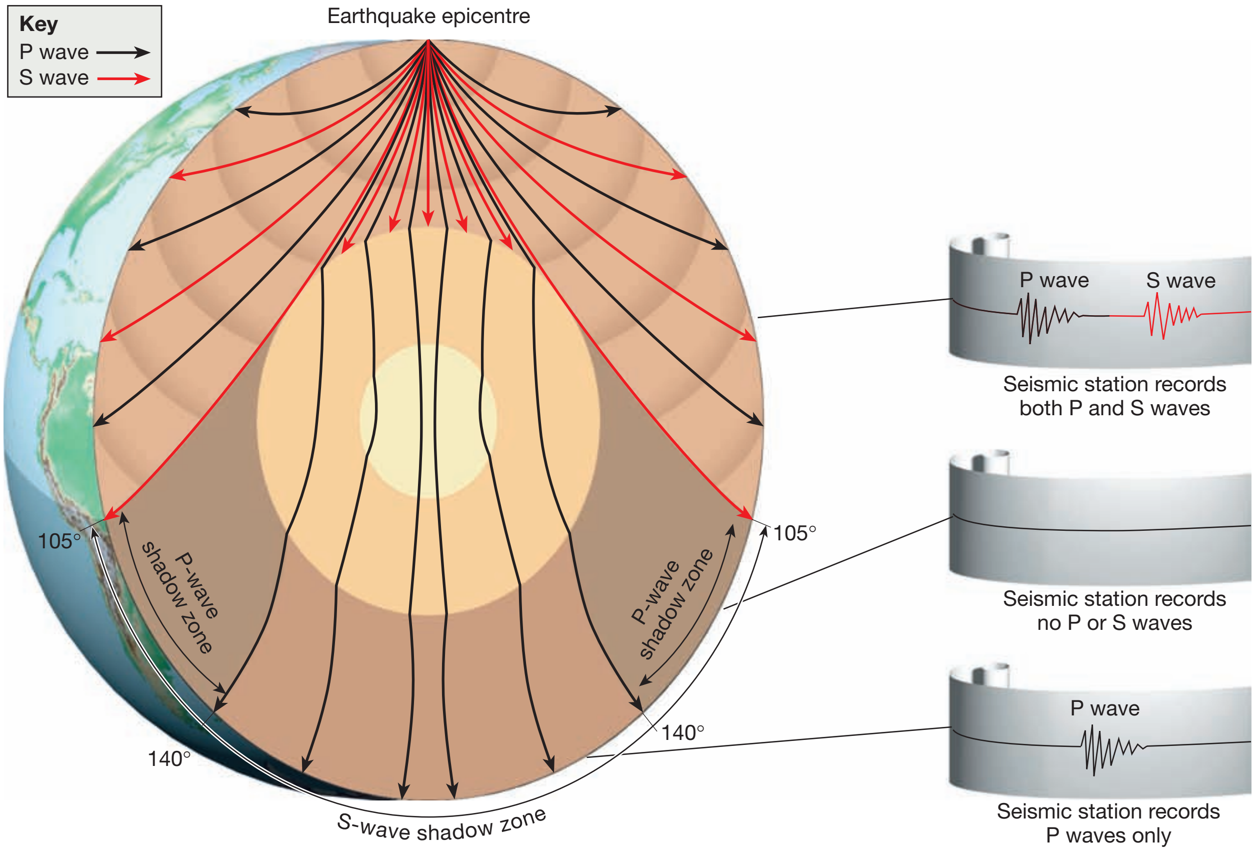
B Time 2—Slower shallow waves arrive at seismic station 2 first.

C Time 3—Faster deeper waves arrive at seismic station 3 first.

◆ **Figure 10.25** Idealized paths of seismic waves travelling from an earthquake focus to three seismic stations. In parts **A** and **B**, you can see that the two nearest recording stations receive the slower waves first because the waves travelled a shorter distance. However, as shown in part **C**, beyond 200 kilometres, the first waves received passed through the mantle, which is a zone of higher velocity.

Gutenberg, and others before him, realized that the P-wave shadow zone could be explained if Earth contained a core that was composed of material unlike the overlying mantle. The core, which Gutenberg calculated to be located at a depth of 2900 kilometres, must somehow hinder the transmission of P waves, similar to the way in which light rays are blocked by an object that casts a shadow. However, rather than actually stopping the P waves, the shadow zone is produced by bending (refracting) of the P waves, which enter the core, as shown in Figure 10.26.

It was further determined that S waves do not travel through the core. This fact led geologists to conclude that at least a portion of this region is liquid. This conclusion was further supported by the observation that P-wave velocities suddenly decrease by about 40 percent as they enter the core. Because



◆ **Figure 10.26** View of Earth's interior showing P-wave and S-wave paths. Any location more than 105° from the earthquake epicentre will not receive direct S waves because the outer core will not transmit them. Although P waves are also absent beyond 105°, they are recorded beyond 140° because of refraction.

melting reduces the elasticity of rock, this evidence pointed to the existence of a liquid layer below the rocky mantle.

Discovery of the Inner Core

In 1936 the last major subdivision of Earth's interior was predicted by Inge Lehmann, a Danish seismologist. Lehmann discovered a new region of seismic reflection and refraction within the core. Hence, a core within a core was discovered. The size of the **inner core** was not precisely established until the early 1960s, when underground nuclear tests were conducted in Nevada. Because the exact location and time of the explosions were known, echoes from seismic waves that bounced off the inner core provided a precise means of determining its size.

From these data, the inner core was found to have a radius of about 1216 kilometres. Furthermore, P waves passing through the inner core have appreciably faster average velocities than do those penetrating only the outer core. The apparent increase in the elasticity of the inner core is evidence that this innermost region is solid.

CHAPTER SUMMARY

- *Earthquakes* occur when crustal rocks rupture at the focus, below the epicentre. Rupture generates *seismic waves* and occurs by *elastic rebound* to release built-up stress along a fault. *Foreshocks* often precede a major earthquake and *aftershocks* follow it.
- Seismic waves include *body* (through the Earth) and *surface* (around the Earth). Body waves include faster *primary* or *P waves* that travel through rock along a direct path and *secondary* or *S waves* that shake rock at right angles to travel direction.

- *Epicentres* are located using travel times of P and S waves and triangulating distances from three or more seismic stations. Most epicentres occur along the circum-Pacific belt and the oceanic ridge system.
- The *Modified Mercalli Intensity Scale* is based on damage to buildings. Magnitude estimates energy released using the *Richter scale* by comparing largest wave amplitude with distance to the epicentre. Each higher number means 32 times more energy was released.
- Structural damage depends on wave amplitude, duration of vibrations, substrate (geology), and structural design. Secondary effects of quakes include *tsunami*, landslides, subsidence, and fire.
- While short-range predictions are not yet possible, earthquake-prone regions with *seismic gaps* (overdue) are well known. P and S waves travel faster in solid elastic materials, slower in weaker layers, and are reflected and refracted (bent) at material boundaries. This tells us that Earth's layers are crust, mantle, outer liquid, and inner solid core.

REVIEW QUESTIONS

1. What is an earthquake? Under what circumstances do earthquakes occur?
2. How are faults, foci, and epicentres related?
3. Explain what is meant by elastic rebound.
4. Describe the principle of a seismograph.
5. List the major differences between P (primary), S (secondary), and L (surface or long) waves. Which of these seismic wave types causes the greatest destruction?
6. Each increase of 1 on the Richter scale corresponds to a _____-fold increase in wave amplitude and a _____-fold increase in energy release.
7. In addition to seismic vibrations, what other factors contribute to the destruction associated with earthquakes?
8. What is a tsunami? How is one generated?
9. Explain how the following boundaries within Earth's interior were discovered:
   a) The crust-mantle boundary (Mohorovičić discontinuity or Moho)
   b) The core-mantle boundary
   c) The inner core–outer core boundary
10. What types of information are useful in making short-range predictions versus long-range forecasts of future earthquake activity?