Chapter 11 Earthquakes

Learning Objectives

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

- Explain how the principle of elastic deformation applies to earthquakes.
- Describe how the main shock and the immediate aftershocks define the rupture surface of an earthquake, and explain how stress transfer is related to aftershocks.
- Explain the process of episodic tremor and slip.
- Describe the relationship between earthquakes and plate tectonics, including where we should expect earthquakes to happen at different types of plate boundaries and at what depths.
- Distinguish between earthquake magnitude and intensity, and explain some of the ways of estimating magnitude.
- Explain the importance of collecting intensity data following an earthquake.
- Describe how earthquakes lead to the destruction of buildings and other infrastructure, fires, slope failures, liquefaction, and tsunami.
- Discuss the value of earthquake forecasting, and describe some of the steps that governments and individuals can take to minimize the impacts of large earthquakes.

Earthquakes scare people … a lot! That's not surprising because time and time again earthquakes have caused massive damage and many injuries and deaths. Anyone who has lived through a strong earthquake cannot forget the experience (Figure 11.0.1). But geoscientists and engineers are getting better at understanding earthquakes, minimizing the amount of damage they cause, and reducing the number of people affected. People living in western Canada don't need to be frightened by earthquakes, but they do need to be prepared.

Figure 11.0.1 A schoolroom in Courtenay damaged by the 1946 Vancouver Island earthquake. If the earthquake had not happened on a Sunday, the casualties would have been much greater.

Media Attributions

Figure 11.0.1: "Courtenay B.C., Damage to interior of Elementary School" by Earthquakes Canada. Permission to reproduce for public, non-commercial purposes.

11.1 What is an Earthquake?

An earthquake is the shaking caused by the **rupture** (breaking) and subsequent displacement of rocks (one body of rock moving with respect to another) beneath Earth's surface.

A body of rock that is under stress becomes deformed. When the rock can no longer withstand the deformation, it breaks and the two sides slide past each other. Most earthquakes take place near plate boundaries—but not necessarily right on a boundary—and not necessarily even on a pre-existing fault.

The engineering principle of **elastic deformation**, which can be used to understand earthquakes, is illustrated in Figure 11.1.1. The stress applied to a rock—typically because of ongoing plate movement—results in strain or deformation of the rock (Figure 11.1.1b). Because most rock is strong (unlike loose sand, for example), it can withstand a significant amount of deformation without breaking. But every rock has a deformation limit and will rupture (break) once that limit is reached. At that point, in the case of rocks within the crust, the rock breaks and there is displacement along the **rupture surface** (Figure 11.1.1c). The magnitude of the earthquake depends on the extent of the area that breaks (the area of the rupture surface) and the average amount of displacement (sliding).

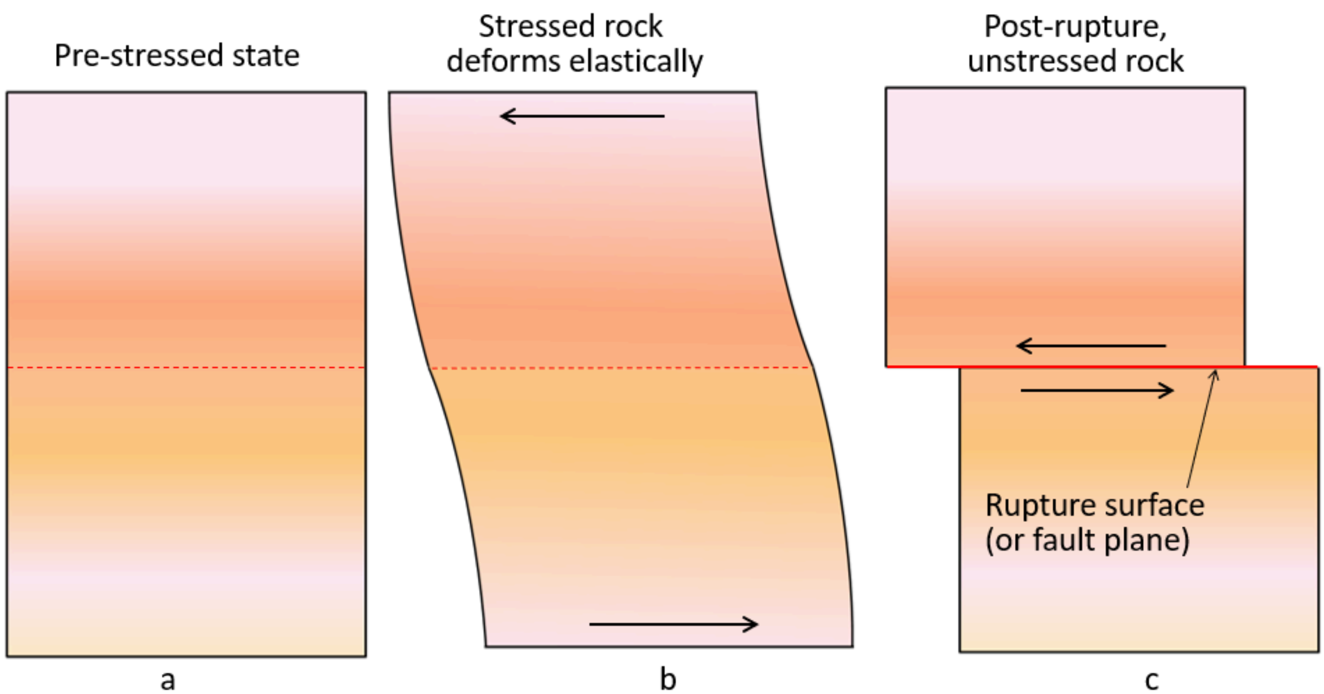

Figure 11.1.1 Depiction of the concept of elastic deformation and rupture. The plate boundary is shown as a dashed red line. In b the two plates are moving as shown by the arrows, but they are locked against each along the plate boundary so they are both deforming and the rocks are stressed. In c there has been a rupture along the boundary and the stress is released.

The concept of a rupture surface, which is critical to understanding earthquakes, is illustrated in Figure 11.1.2. An earthquake does not happen at a point, it happens over an area within a plane, although not necessarily a *flat* plane. Within the area of the rupture surface, the amount of displacement is variable (Figure 11.1.2), and, by definition, it decreases to zero at the edges of the rupture surface because the rock beyond that point isn't displaced at all. The extent of a rupture surface and the amount of displacement will depend on a number of factors, including the type and strength of the rock, and the degree to which it was stressed beforehand.

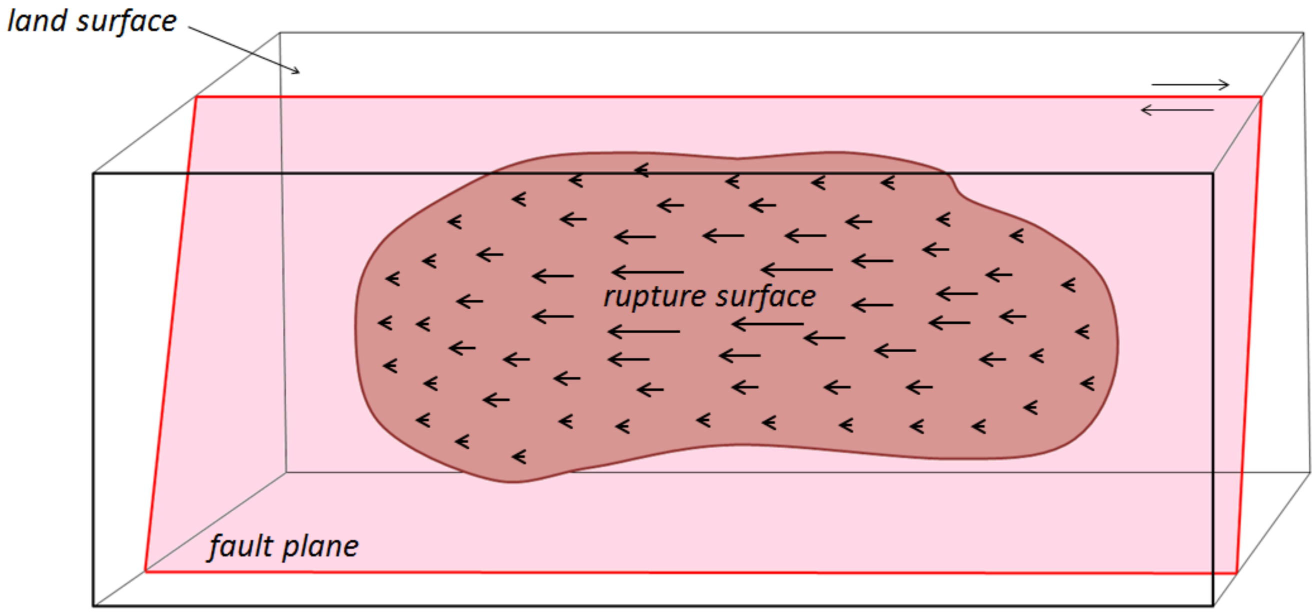

Figure 11.1.2 A rupture surface (dark pink), on a steeply dipping fault plane (light pink). The diagram represents a part of the crust that may be up to tens or hundreds of kilometres long. The rupture surface is the part of the fault plane along which displacement occurred. In this example, the near side of the fault is moving to the left, and the lengths of the arrows within the rupture surface represent relative amounts of displacement.

Earthquake rupture doesn't happen all at once; it starts at a single point and spreads rapidly from there. Depending on the extent of the rupture surface, the propagation of failures out from the point of initiation is typically completed within seconds to several tens of seconds (Figure 11.1.3). The initiation point isn't necessarily in the centre of the rupture surface; it may be close to one end, near the top, or near the bottom.

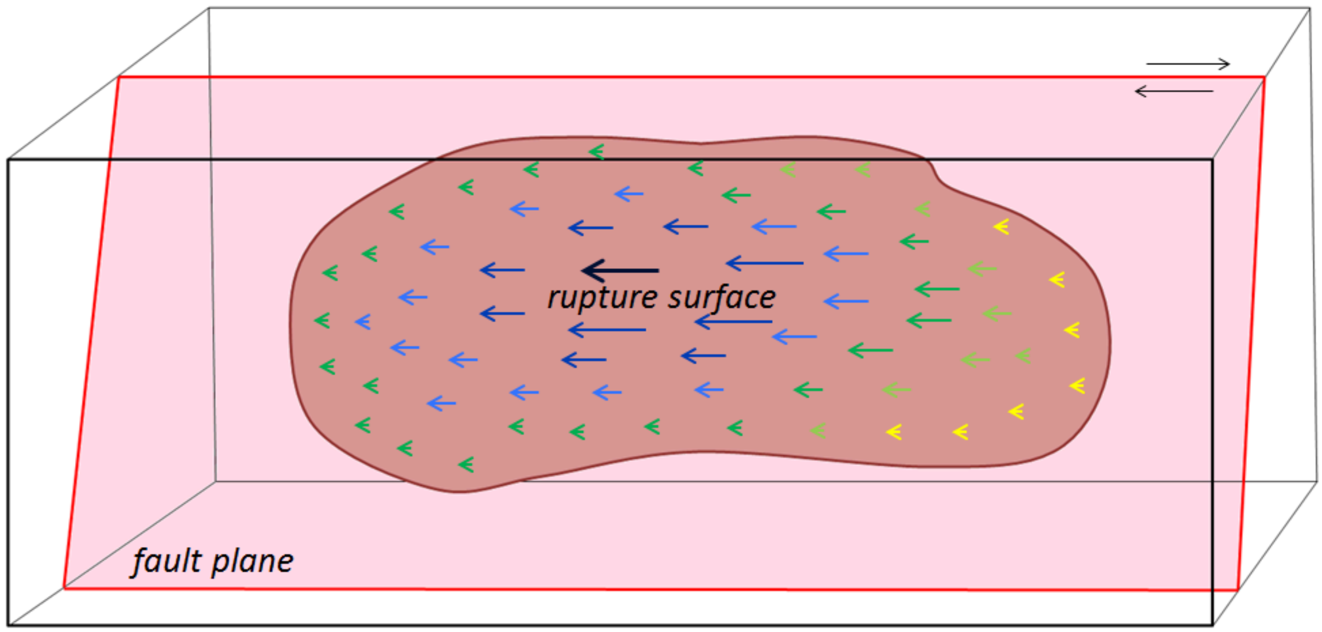

Figure 11.1.3 Propagation of failure on a rupture surface. In this case, the failure starts at the dark blue heavy arrow in the centre and propagates outward, reaching the left side first (green arrows) and the right side last (yellow arrows).

Figure 11.1.4 shows the distribution of immediate **aftershocks** associated with the 1989 Loma Prieta earthquake. Panel (b) is a section along the San Andreas Fault; this view is equivalent to what is shown in Figures 11.1.2 and 11.1.3. The area of red dots is the rupture surface; each red dot is a specific aftershock that was recorded on a seismometer. The hexagon labelled "main earthquake" represents the first or main shock. When that happened, the rock at that location broke and was displaced. That released the stress on that particular part of the fault, but it resulted in an increase of the stress on other nearby parts of the fault, and contributed to a cascade of smaller ruptures (aftershocks), in this case, over an area about 60 kilometres long and 15 kilometres wide.

So, what exactly is an aftershock then? An aftershock is an earthquake just like any other, but it is one that can be shown to have been triggered by **stress transfer** from a preceding earthquake. Within a few tens of seconds of the main Loma Prieta earthquake, there were hundreds of smaller aftershocks; their distribution defines the area of the rupture surface.

Aftershocks can be of any magnitude. Most are smaller than the earthquake that triggered them, but they can be bigger. The aftershocks shown in Figure 11.1.4 all happened within seconds or minutes of the main shock, but aftershocks can be delayed for hours, days, weeks, or even years. As already noted, aftershocks are related to stress transfer. For example, the main shock of the Loma Prieta earthquake triggered aftershocks in the immediate area, which triggered more in the

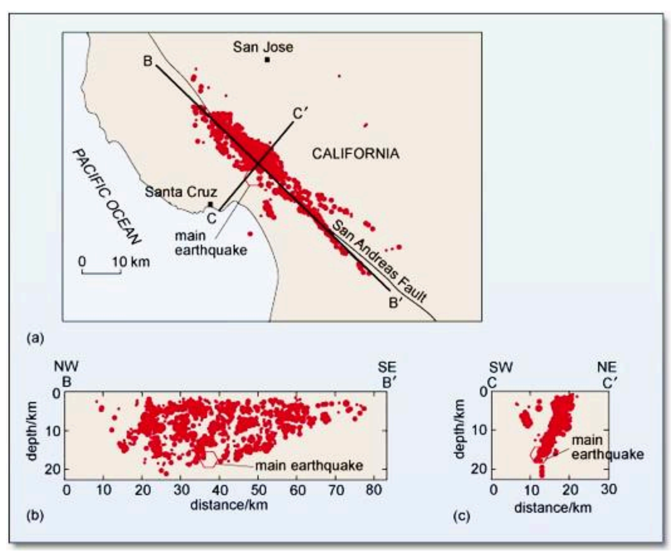

Figure 11.1.4 Distribution of the aftershocks of the 1989 M 6.9 Loma Prieta earthquake (a: plan view, b: section along the fault, c: section across the fault.)

surrounding area, eventually extending for 30 kilometres along the fault in each direction and for 15 kilometres toward the surface. But the earthquake as a whole also changed the stress on adjacent parts of the San Andreas Fault. This effect, which has been modelled for numerous earthquakes and active faults around the world, is depicted in Figure 11.1.5. Stress was reduced in the area of the rupture (blue), but was increased at either end of the rupture surface (red and yellow).

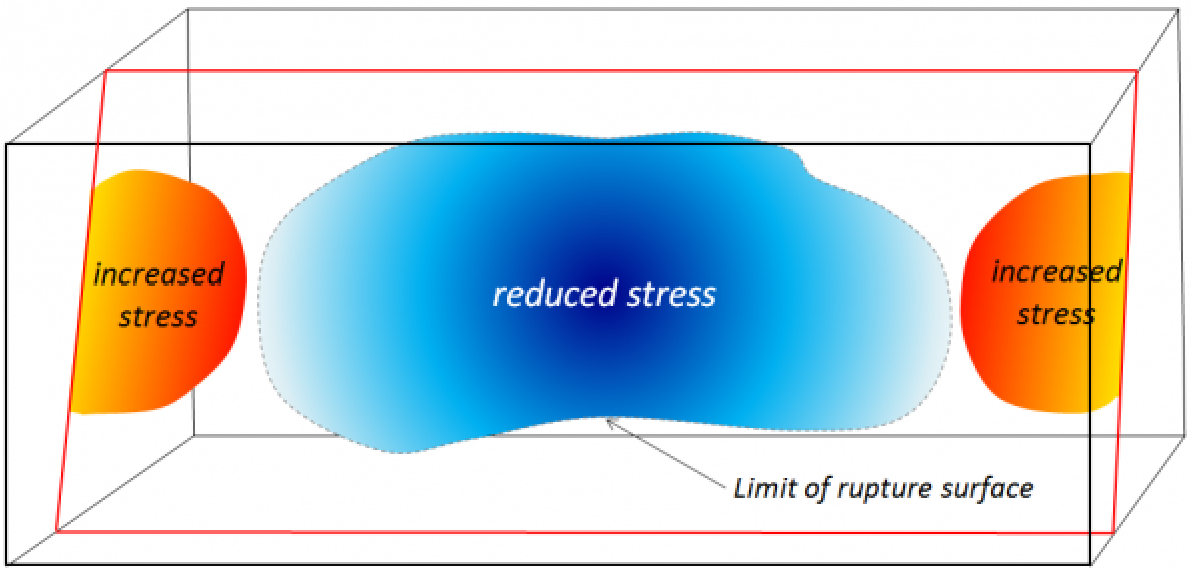

Figure 11.1.5 Depiction of stress changes related to an earthquake. Stress decreases in the area of the rupture surface, but increases on adjacent parts of the fault.

Stress transfer isn't necessarily restricted to the fault along which an earthquake happened. It will affect the rocks in general around the site of the earthquake and may lead to increased stress on other faults in the region. The effects of stress transfer don't necessarily show up right away. Segments of faults are typically in some state of stress, and the transfer of stress from another area is only rarely enough to push a fault segment beyond its limits to the point of rupture. The stress that is added by stress transfer accumulates along with the ongoing buildup of stress from plate motion and eventually leads to another earthquake.

Episodic Tremor and Slip

Episodic tremor and slip (ETS) is periodic slow sliding along part of a subduction boundary. It does not produce recognizable earthquakes, but does produce seismic tremor (rapid seismic vibrations on a seismometer). It was first discovered on the Vancouver Island part of the Cascadia subduction zone by Geological Survey of Canada geologists Herb Dragert and Garry Rogers.[1]

The boundary between the subducting Juan de Fuca Plate and the North America Plate can be divided into three segments (Figure 11.1.6). The cold upper part of the Juan de Fuca Plate boundary is locked. The plates are stuck and don't move, except with very large earthquakes that happen *approximately* every 500 years (the last one was approximately M9 on January 26, 1700). The warm lower part of the boundary is sliding continuously because the warm rock is weaker. The central part of the boundary

1. Rogers, G. and Dragert, H., 2003, Episodic tremor and slip on the Cascadia subduction zone: the chatter of silent slip, Science, V. 300, p. 1942-1943.

isn't cold enough to be stuck, but isn't warm enough to slide continuously. Instead it slips episodically, approximately every 14 months for about 2 weeks, moving a few centimetres each time.

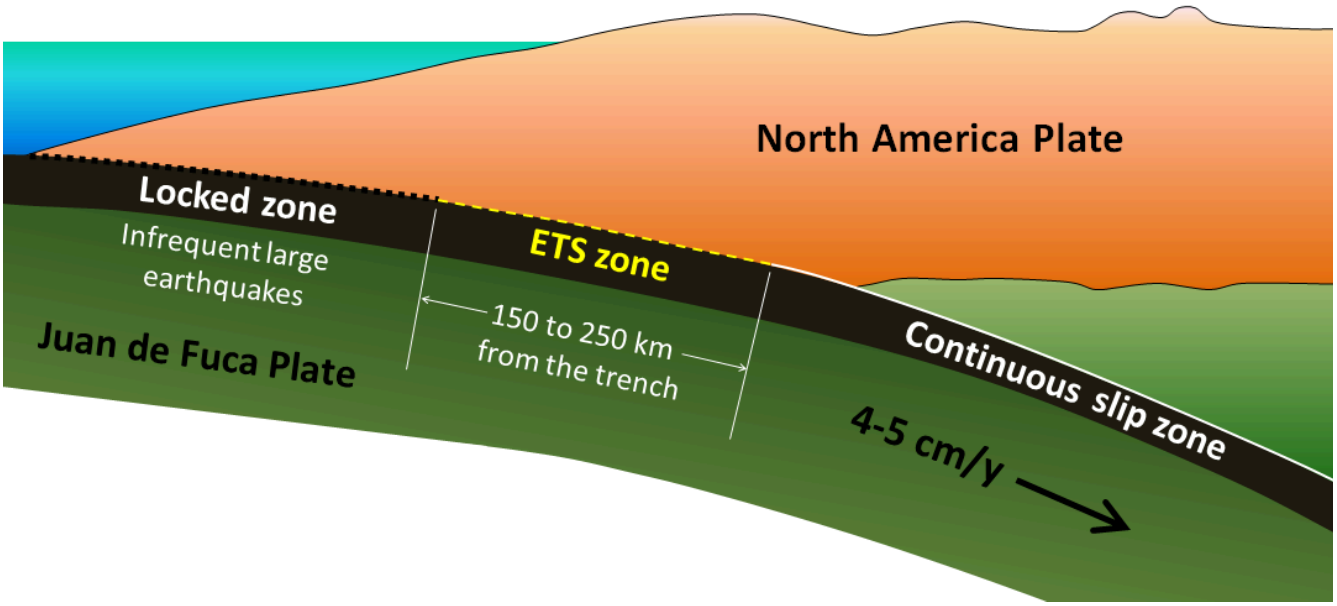

Figure 11.1.6 The boundary between the subducting Juan de Fuca Plate and the North America Plate is locked in the upper part, slides continuously in the lower part and slides episodically in the middle part.

You might be inclined to think that it's a good thing that there is periodic slip on this part of the plate because it releases some of the tension and reduces the risk of a large earthquake. In fact, the opposite is likely the case. The movement along the ETS part of the plate boundary acts like a medium-sized earthquake and leads to stress transfer to the adjacent locked part of the plate. Approximately every 14 months, during the two-week ETS period, there is a transfer of stress to the shallow locked part of the Cascadia subduction zone, and therefore an increased chance of a large earthquake.

Since 2003, ETS processes have also been observed on subduction zones in Mexico, New Zealand and Japan.

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- Figure 11.1.5: © Steven Earle. CC BY. Based on data from 2010 Laguna Salada earthquake by Stein and Toda.

11.2 Earthquakes and Plate Tectonics

The distribution of earthquakes across the globe is shown in Figure 11.2.1. It is relatively easy to see the relationships between earthquakes and the plate boundaries. Along divergent boundaries like the mid-Atlantic ridge and the East Pacific Rise, earthquakes are common, but restricted to a narrow zone close to the ridge, and consistently at less than a 30 kilometre depth. Shallow earthquakes are also common along transform faults, such as the San Andreas Fault. Along subduction zones, as we saw in Chapter 10, earthquakes are very abundant, and they are increasingly deep on the landward side of the subduction zone.

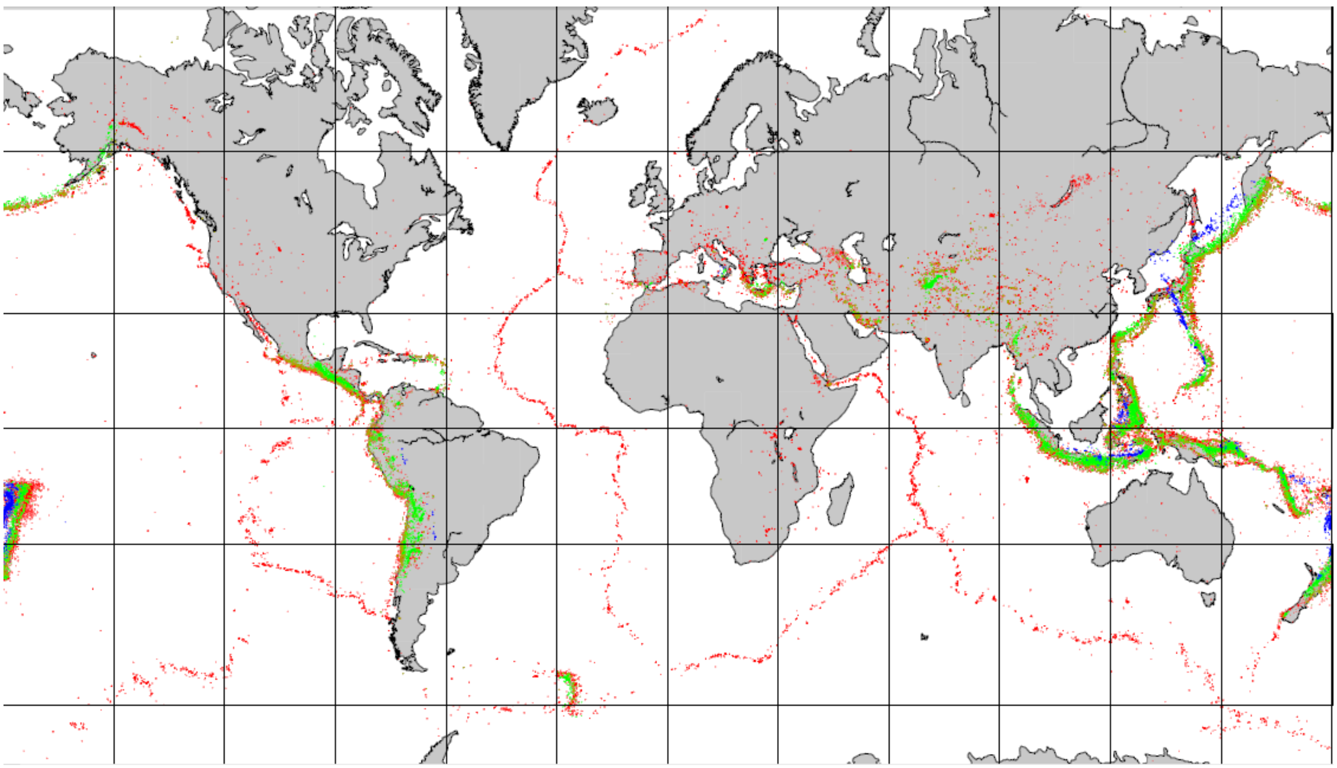

Figure 11.2.1 General distribution of global earthquakes of magnitude 4 and greater from 2004 to 2011, colour coded by depth (red: 0 to 33 kilometres, orange 33 to 70 kilometres, green: 70 to 300 kilometres, blue: 300 to 700 kilometres).

Earthquakes are also relatively common at a few intraplate locations. Some are related to the buildup of stress due to continental rifting or the transfer of stress from other regions, and some are not well understood. Examples of intraplate earthquake regions include the Great Rift Valley area of Africa, the Tibet region of China, and the Lake Baikal area of Russia.

Earthquakes at Divergent and Transform Boundaries

Figure 11.2.2 provides a closer look at magnitude (M) 4 and larger earthquakes in an area of divergent boundaries in the mid-Atlantic region near the equator. Here, as we saw in Chapter 10, the segments of the mid-Atlantic ridge are offset by some long transform faults. Most of the earthquakes are located along the transform faults, rather than along the spreading segments, although there are clusters of earthquakes at some of the ridge-transform boundaries. Some earthquakes do occur on spreading ridges, but they tend to be small and infrequent because of the relatively high rock temperatures in the areas where spreading is taking place.

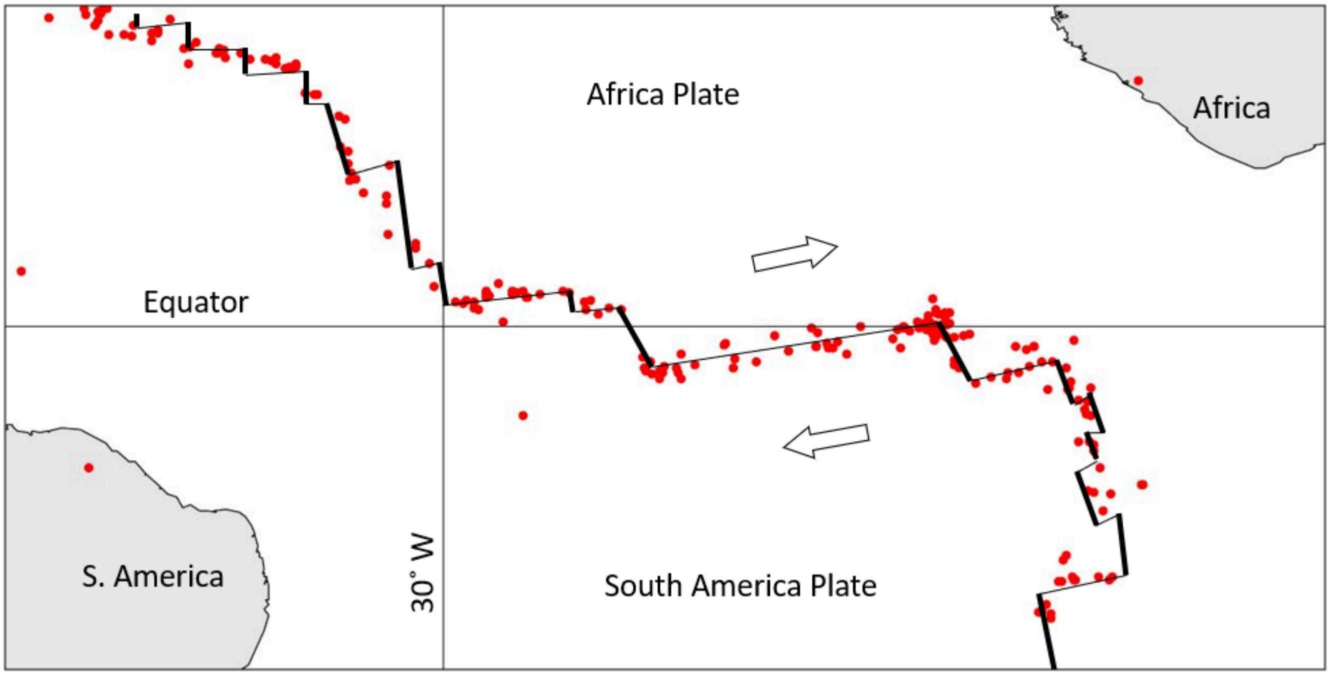

Figure 11.2.2 Distribution of earthquakes of M4 and greater in the area of the mid-Atlantic ridge near the equator from 1990 to 1996. All are at a depth of 0 to 33 kilometres.

Earthquakes at Convergent Boundaries

The distribution and depths of earthquakes in the Caribbean and Central America area are shown in Figure 11.2.3. In this region, the Cocos Plate is subducting beneath the North America and Caribbean Plates (ocean-continent convergence), and the South and North America Plates are subducting beneath the Caribbean Plate (ocean-ocean convergence). In both cases, the earthquakes get deeper with distance from the trench. In Figure 11.2.3, the South America Plate is shown as being subducted beneath the Caribbean Plate in the area north of Colombia, but since there is almost no earthquake activity along this zone, it is questionable whether subduction is actually taking place.

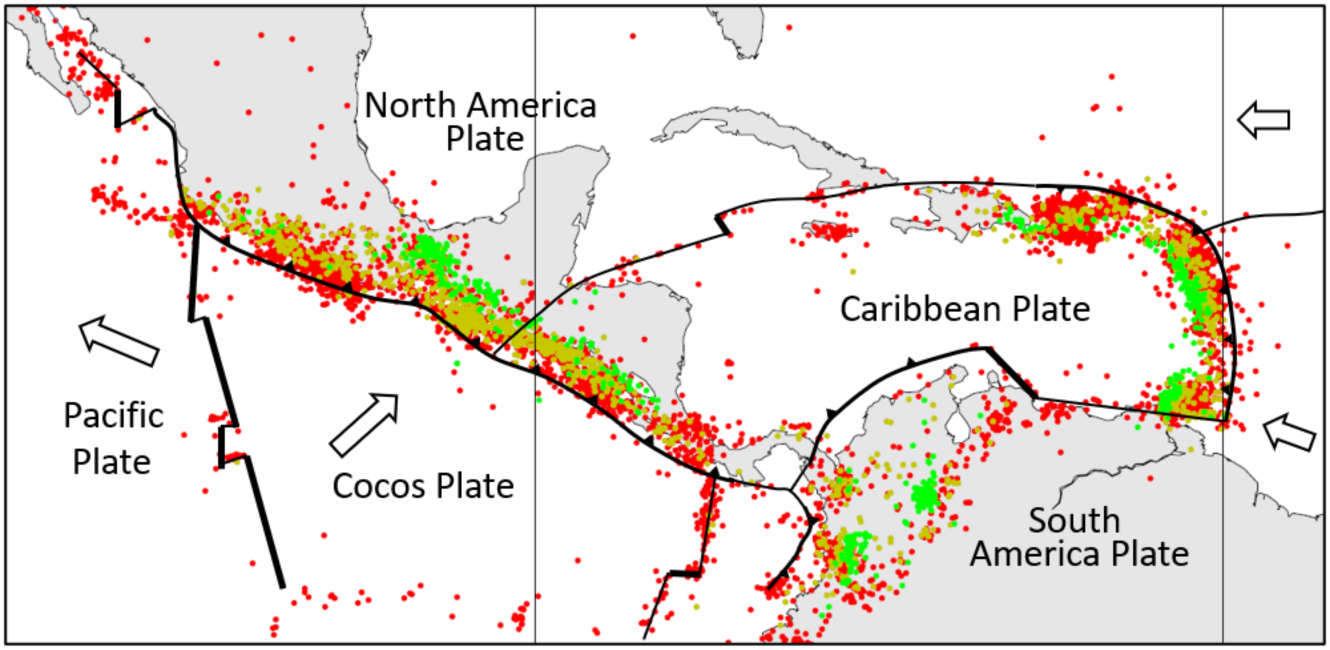

Figure 11.2.3 Distribution of earthquakes of M4 and greater in the Central America region from 1990 to 1996 (red: 0 to 33 kilometres, orange: 33 to 70 kilometres, green: 70 to 300 kilometres, blue: 300 to 700 kilometres) (Spreading ridges are heavy lines, subduction zones are toothed lines, and transform faults are light lines.)

There are also various divergent and transform boundaries in the area shown in Figure 11.2.3, and as we've seen in the mid-Atlantic area, most of these earthquakes occur along the transform faults.

The distribution of earthquakes with depth in the Kuril Islands of Russia in the northwest Pacific is shown in Figure 11.2.4. This is an ocean-ocean convergent boundary. The small red and yellow dots show background seismicity over a number of years, while the larger white dots are individual shocks associated with a M6.9 earthquake in April 2009. The relatively large earthquake took place on the upper part of the plate boundary between 60 kilometres and 140 kilometres inland from the trench. As we saw for the Cascadia subduction zone, this is where large subduction earthquakes are expected to occur.

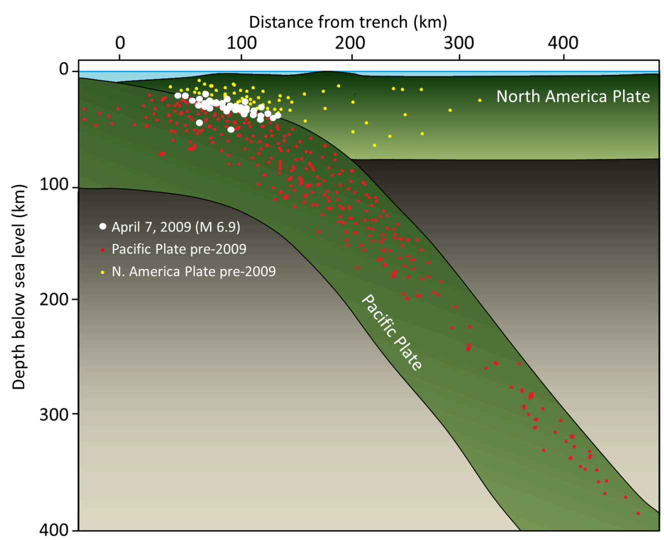

Figure 11.2.4 Distribution of earthquakes in the area of the Kuril Islands, Russia (just north of Japan) (White dots represent the April 2009 M6.9 earthquake. Red and yellow dots are from background seismicity over several years prior to 2009.)

In fact, all of the very large earthquakes — M9 or higher — take place at subduction boundaries because there is the potential for a greater width of rupture zone on a gently dipping boundary than on a steep transform boundary. The largest earthquakes on transform boundaries are in the order of M8.

The background seismicity at this convergent boundary, and on other similar ones, is predominantly near the upper side of the subducting plate. The frequency of earthquakes is greatest near the surface and especially around the area where large subduction quakes happen, but it extends to at least a 400 kilometre depth. There is also significant seismic activity in the overriding North America Plate, again most commonly near the region of large quakes, but also extending for a few hundred kilometres away from the plate boundary.

The distribution of earthquakes in the area of the India-Eurasia plate boundary is shown in Figure 11.2.5. This is a continent-continent convergent boundary, and it is generally assumed that although the India Plate continues to move north toward the Asia Plate, there is no actual subduction taking place. There are transform faults on either side of the India Plate in this area.

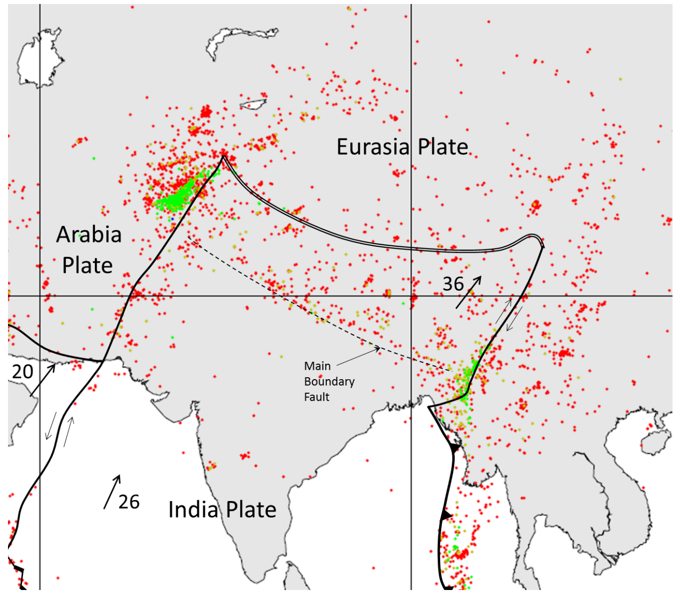

Figure 11.2.5 Distribution of earthquakes in the area where the India Plate is converging with the Asia Plate (data from 1990 to 1996, red: 0 to 33 kilometres, orange: 33 to 70 kilometres, green: 70 to 300 kilometres). (Spreading ridges are heavy lines, subduction zones are toothed lines, and transform faults are light lines. The double line along the northern edge of the India Plate indicates convergence, but not subduction. Plate motions are shown in millimetres per year).

The entire northern India and southern Asia region is very seismically active. Earthquakes are common in northern India, Nepal, Bhutan, Bangladesh and adjacent parts of China, and throughout Pakistan and Afghanistan. Many of the earthquakes are related to the transform faults on either side of the India Plate, and most of the others are related to the significant tectonic squeezing caused by the continued convergence of the India and Asia Plates. That squeezing has caused the Asia Plate to be thrust over top of the India Plate, building the Himalayas and the Tibet Plateau to enormous heights. Most of the earthquakes of Figure 11.2.5 are related to the thrust faults shown in Figure 11.2.6 (and to hundreds of other similar ones that cannot be shown at this scale). The southernmost thrust fault in Figure 11.2.6 is equivalent to the Main Boundary Fault in Figure 11.2.5.

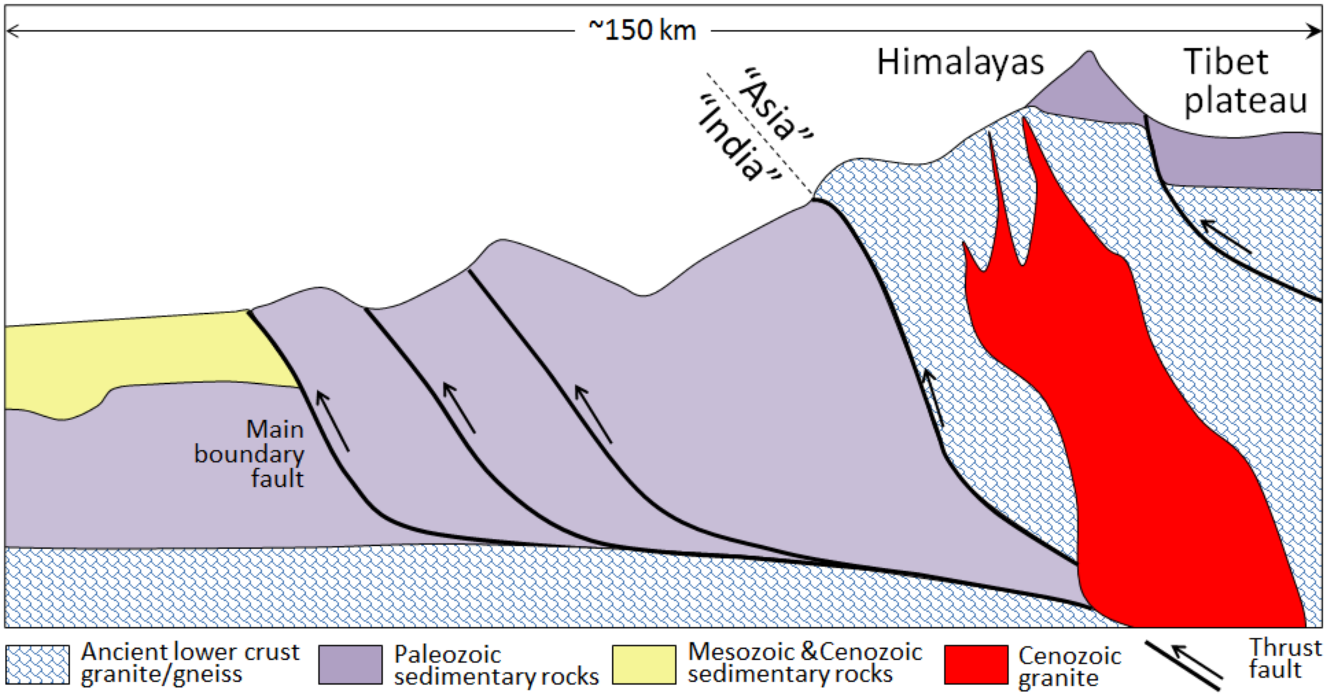

Figure 11.2.6 Schematic diagram of the India-Asia convergent boundary, showing examples of the types of faults along which earthquakes are focused. The devastating Nepal earthquake of May 2015 took place along one of these thrust faults.

There is a very significant concentration of both shallow and deep (greater than 70 kilometres) earthquakes in the northwestern part of Figure 11.2.5. This is northern Afghanistan, and at depths of more than 70 kilometres, many of these earthquakes are within the mantle as opposed to the crust. It is interpreted that these deep earthquakes are caused by northwestward subduction of part of the India Plate beneath the Asia Plate in this area.

Exercise 11.1 Earthquakes in British Columbia

This map shows the incidence and magnitude of earthquakes in British Columbia over a one-month period in March and April 2015.

1. What is the likely origin of the earthquakes between the Juan de Fuca (JDF) and Explorer Plates?
2. The string of small earthquakes adjacent to Haida Gwaii (H.G.) coincides closely with the rupture surface of the 2012 M7.8 earthquake in that area. How might these earthquakes be related to that one?
3. Most of the earthquakes around Vancouver Island (V.I.) are relatively shallow. What is their likely origin?
4. Some of the earthquakes in B.C. are interpreted as being caused by natural gas extraction (including fracking). Which of the earthquakes here could fall into this category?

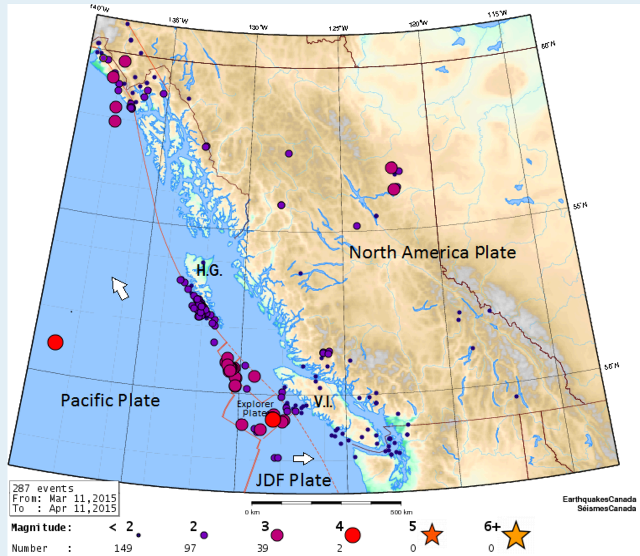

Figure 11.2.7 [Image Description]

See Appendix 3 for Exercise 11.1 answers.

Image descriptions

Figure 11.2.7 image description: The incidence and magnitude of earthquakes in British Columbia over a one-month period in March and April 2015: There were a few dozen smaller earthquakes spread out around Vancouver Island and the sunshine coast with a magnitude of 2. Farther west along the Explorer Plate, which is between the North American plate, the Juan de Fuca Plate, and the Pacific Plate, there were quite a few earthquakes with a magnitude of 3 and at least one earthquake with a magnitude of 4. Between the North American Plate and the Pacific Plate off the south-west coast of Haida Gwaii, there was a large cluster of earthquakes with magnitudes of 2. Along the Alaskan panhandle, there was a collection of 2- and 3-magnitude earthquakes. In addition, there were two 3-magnitude earthquakes west of Fort St. John in northern British Columbia and one or two 2-magnitude earthquakes. In total, this map shows one hundred and forty-nine earthquakes with a magnitude less than 2, ninety-seven earthquakes with a magnitude of 2, thirty-nine earthquakes with a magnitude of 3, and two earthquakes with a magnitude of 4 [Return to Figure 11.2.7]

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- Figure 11.2.1: Global Earthquakes © Dale Sawyer, Rice University. Used with permission.
- Figure 11.2.2: Earthquakes Around the Mid-Atlantic Ridge © Steven Earle after Dale Sawyer, Rice University.
- Figure 11.2.3: Earthquakes Around the Central-American Region © Steven Earle after Dale Sawyer, Rice University.
- Figure 11.2.4: Earthquakes Around the Kuril Islands. © Steven Earle after Gavin Hayes,

from data from the USGS [PDF].

- Figure 11.2.5: Earthquakes Around the India Plate © Steven Earle after Dale Sawyer, Rice University.
- Figure 11.2.6: India-Asia Convergent Boundary © Steven Earle based on D. Vouichard, from a United Nations University document.

11.3 Measuring Earthquakes

There are two main ways to measure earthquakes. The first of these is an estimate of the energy released, and the value is referred to as **magnitude**. This is the number that is typically used by the press when a big earthquake happens. It is often referred to as "Richter magnitude," but that is a misnomer, and it should be just "magnitude." There are many ways to measure magnitude—including Charles Richter's method developed in 1935—but they are all ways to estimate the same number, which is proportional to the amount of energy released.

The other way of assessing the impact of an earthquake is to assess what people felt and how much damage was done. This is known as **intensity**. Intensity values are assigned to locations, rather than to the earthquake itself, and therefore intensity can vary widely, depending on the proximity to the earthquake and the types of materials underneath and the local conditions.

Earthquake Magnitude

Before we look more closely at magnitude we need to review what we know about body waves, and look at surface waves. Body waves are of two types, P waves, or primary or compression waves (like the compression of the coils of a spring), and S waves, or secondary or shear waves (like the flick of a rope). An example of P and S seismic wave records is shown in Figure 11.3.1. The critical parameters for the measurement of magnitude are labelled, including the time interval between the arrival of the P- and S-waves—which is used to determine the distance from the earthquake to the seismic station, and the amplitude of the S-waves—which is used to estimate the magnitude of the earthquake.

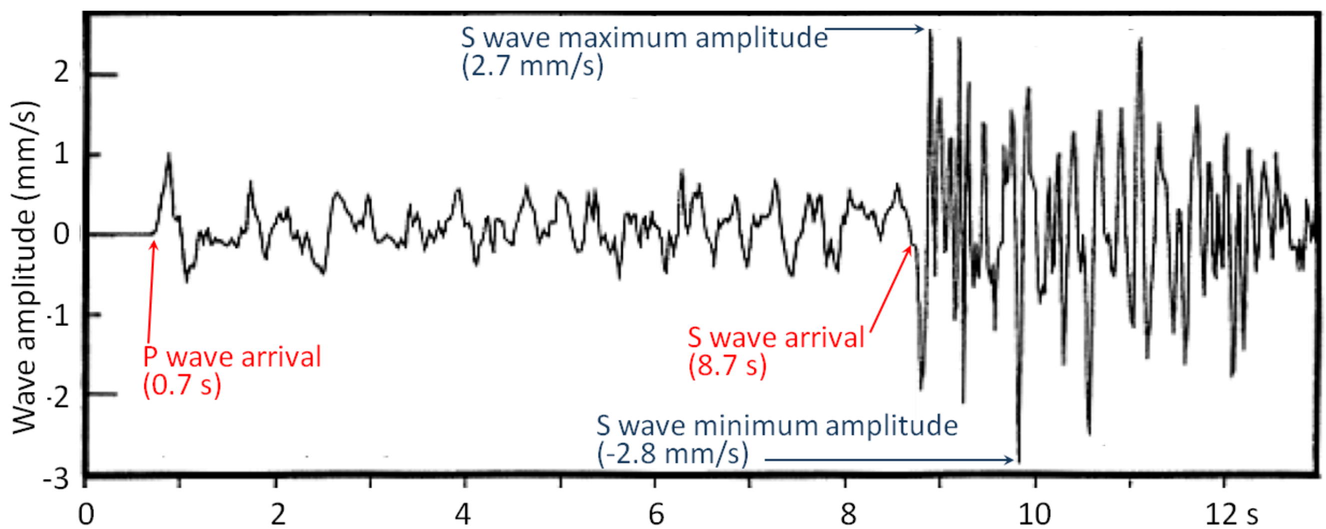

Figure 11.3.1 P-waves and S-waves from a small (M4) earthquake that took place near Vancouver Island in 1997. [Image Description]

When body waves (P or S) reach Earth's surface, some of their energy is transformed into surface waves, of which there are two main types, as illustrated in Figure 11.3.2. **Rayleigh waves** are characterized by vertical motion of the ground surface, like waves on water, while **Love waves** are characterized by horizontal motion. Both Rayleigh and Love waves are about 10% slower than S-waves (so they arrive later at a seismic station). Surface waves typically have greater amplitudes than body waves, and they do more damage.

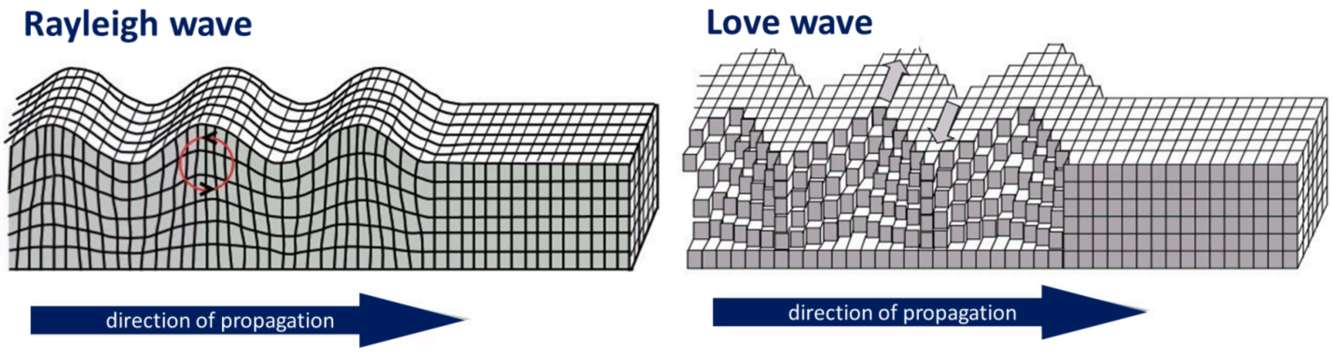

Figure 11.3.2 Depiction of seismic surface waves.

Other important terms for describing earthquakes are **hypocentre** (or **focus**) and **epicentre**. The hypocentre is the actual location of an individual earthquake shock at depth in the ground, and the epicentre is the point on the land surface vertically above the hypocentre (Figure 11.3.3).

A number of methods for estimating magnitude are listed in Table 11.1. Local magnitude (ML) was widely used until late in the 20th century, but **moment magnitude** (MW) is now more commonly used because it gives more accurate estimates (especially with larger earthquakes) and can be applied to earthquakes at any distance from a seismometer. Surface-wave magnitudes can also be applied to measure distant large earthquakes.

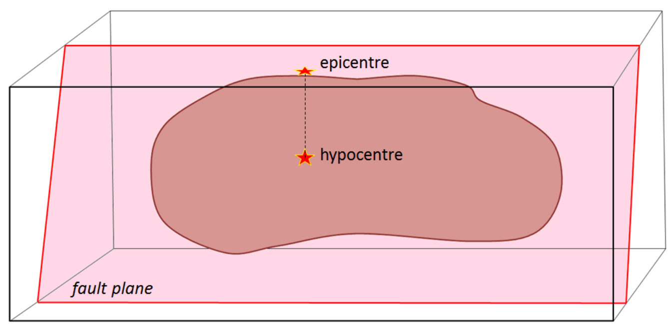

Figure 11.3.3 Epicentre and hypocentre.

Because of the increasing size of cities in earthquake-prone areas (e.g., China, Japan, California) and the increasing sophistication of infrastructure, it is becoming important to have very rapid warnings and magnitude estimates of earthquakes that have already happened. This can be achieved by using P-wave data to determine magnitude because P-waves arrive first at seismic stations, in many cases several seconds ahead of the more damaging S-waves and surface waves. Operators of electrical grids, pipelines, trains, and other infrastructure can use the information to automatically shut down systems so that damage and casualties can be limited.

Table 11.1 A summary of some of the different methods for estimating earthquake magnitude.[1]

[Skip Table]

Type	M Range	Dist. Range	Comments
Local or Richter (M_L)	2 to 6	0 to 400 kilometres	The original magnitude relationship defined in 1935 by Richter and Gutenberg. It is based on the maximum amplitude of S-waves recorded on a Wood-Anderson torsion seismograph. ML values can be calculated using data from modern instruments. L stands for local because it only applies to earthquakes relatively close to the seismic station.
Moment (M_W)	Greater than 3.5	All	Based on the **seismic moment** of the earthquake, which is equal to the average amount of displacement on the fault times the fault area that slipped. It can also be estimated from seismic data if the seismometer is tuned to detect long-period body waves.
Surface wave (M_S)	5 to 8	20 to 180°	A magnitude for distant earthquakes based on the amplitude of surface waves measured at a period near 20 seconds.
P-wave	2 to 8	Local	Based on the amplitude of P-waves. This technique is being increasingly used to provide very rapid magnitude estimates so that early warnings can be sent to utility and transportation operators to shut down equipment before the larger (but slower) S-waves and surface waves arrive.

Exercise 11.2 Moment magnitude estimates from earthquake parameters

Use this moment magnitude calculation tool to estimate the moment magnitude based on the approximate length, width, and displacement values provided in the following table:

Table 11.2 Calculate Moment Magnitude Based on Length, Width, and Displacement Values

[Skip Table]

Length (kilometres)	Width (kilometres)	Displacement (metres)	Earthquake	M W?
60	15	4	The 1946 Vancouver Island earthquake	
0.4	0.2	.5	The small Vancouver Island earthquake shown in Figure 11.3.1	
20	8	4	The 2001 Nisqually earthquake described in Exercise 11.3	
1,100	120	10	The 2004 Indian Ocean earthquake	
30	11	4	The 2010 Haiti earthquake	

1. Table 11.1 by Steven Earle.

The largest recorded earthquake had a magnitude of 9.5. Could there be a 10? You can answer that question using this tool. See what numbers are needed to make MW = 10. Are they reasonable?

See Appendix 3 for Exercise 11.2 answers.

The magnitude scale is logarithmic; in fact, the amount of energy released by an earthquake of M4 is 32 times higher than that released by one of M3, and this ratio applies to all intervals in the scale. If we assign an arbitrary energy level of 1 unit to a M1 earthquake the energy for quakes up to M8 will be as shown on the following table:

Table 11.3 The energy of an earthquake increases by 32 times at each magnitude level.

Magnitude	Energy
1	1
2	32
3	1,024
4	32,768
5	1,048,576
6	33,554,432
7	1,073,741,824
8	34,359,738,368

In any given year, when there is a large earthquake on Earth (M8 or M9), the amount of energy released by that one event will likely exceed the energy released by all smaller earthquake events combined.

Earthquake Intensity

The intensity of earthquake shaking at any location is determined not only by the magnitude of the earthquake and its distance, but also by the type of underlying rock or unconsolidated materials. If buildings are present, the size and type of buildings (and their inherent natural vibrations) are also important.

Intensity scales were first used in the late 19th century, and then adapted in the early 20th century by Giuseppe Mercalli and modified later by others to form what we know call the modified Mercalli intensity scale (Table 11.4). Intensity estimates are important because they allow us to characterize parts of any region into areas that are especially prone to strong shaking versus those that are not. The key factor in this regard is the nature of the underlying geological materials, and the weaker those are, the more likely it is that there will be strong shaking. Areas underlain by strong solid bedrock tend to experience much less shaking than those underlain by unconsolidated river or lake sediments.

Table 11.4 The modified Mercalli intensity scale.

[Skip Table]

Level of intensity	Description
Not felt (1)	Not felt except by a very few under especially favourable conditions
Weak (2)	Felt only by a few persons at rest, especially on upper floors of buildings
Weak (3)	Felt quite noticeably by persons indoors, especially on upper floors of buildings; many people do not recognize it as an earthquake; standing motor cars may rock slightly; vibrations similar to the passing of a truck; duration estimated
Light (4)	Felt indoors by many, outdoors by few during the day; at night, some awakened; dishes, windows, doors disturbed; walls make cracking sound; sensation like heavy truck striking building; standing motor cars rocked noticeably
Moderate (5)	Felt by nearly everyone; many awakened; some dishes, windows broken; unstable objects overturned; pendulum clocks may stop
Strong (6)	Felt by all, many frightened; some heavy furniture moved; a few instances of fallen plaster; damage slight
Very Strong (7)	Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable damage in poorly built or badly designed structures; some chimneys broken
Severe (8)	Damage slight in specially designed structures; considerable damage in ordinary substantial buildings with partial collapse; damage great in poorly built structures; fall of chimneys, factory stacks, columns, monuments, walls; heavy furniture overturned
Violent (9)	Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb; damage great in substantial buildings, with partial collapse; buildings shifted off foundations
Extreme (10)	Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; rails bent
Extreme (11)	Few, if any (masonry), structures remain standing; bridges destroyed; broad fissures in ground; underground pipelines completely out of service; earth slumps and land slips in soft ground; rails bent greatly
Extreme (12)	Damage total; waves seen on ground surfaces; lines of sight and level distorted; objects thrown upward into the air

An example of this effect is the 1985 M8 earthquake that struck the Michoacán region of western Mexico, southwest of Mexico City. There was relatively little damage in the area around the epicentre, but there was tremendous damage and about 5,000 deaths in heavily populated Mexico City some 350 kilometres from the epicentre. The key reason for this is that Mexico City was built largely on the unconsolidated and water-saturated sediment of former Lake Texcoco. These sediments resonate at a frequency of about two seconds, which was similar to the frequency of the body waves that reached the city. For the same reason that a powerful opera singer can break a wine glass by singing the right note, the amplitude of the seismic waves was amplified by the lake sediments. Survivors of the disaster recounted that the ground in some areas moved up and down by about 20 centimetres every two seconds for over two minutes. Damage was greatest to buildings between 5 and 15 storeys tall, because they also resonated at around two seconds, which amplified the shaking.

Exercise 11.3 Estimating intensity from personal observations

The following observations were made by residents of the Nanaimo area during the M6.8 Nisqually earthquake near Olympia, Washington in 2001. Estimate the Mercalli intensities using Table 11.4.

Table 11.5

[Skip Table]

Building Type	Floor	Shaking Felt	How long it lasted (in seconds)	Description of Motion	Intensity?
House	1	no	10	Heard a large rumble lasting not even 10 seconds, mirror swayed	
House	2	moderate	60	Candles, pictures and CDs on bookshelf moved, towels fell off racks	
House	1	no		Pots hanging over stove moved and crashed together	
House	1	weak		Rolling feeling with a sudden stop, picture fell off mantle, chair moved	
Apartment	1	weak	10	Sounded like a big truck then everything shook for a short period	
House	1	moderate	20-30	Teacups rattled but didn't fall off	
Institution	2	moderate	15	Creaking sounds, swaying movement of shelving	
House	1	moderate	15-30	Bed banging against the wall with me in it, dog barking aggressively	

See Appendix 3 for Exercise 11.3 answers.

An intensity map for the 1946 M7.3 Vancouver Island earthquake is shown in Figure 11.3.4. The intensity was greatest in the central island region where, in some communities, chimneys were damaged on more than 75% of buildings, some roads were made impassable, and a major rock slide occurred. The earthquake was felt as far north as Prince Rupert, as far south as Portland Oregon, and as far east as the Rockies.

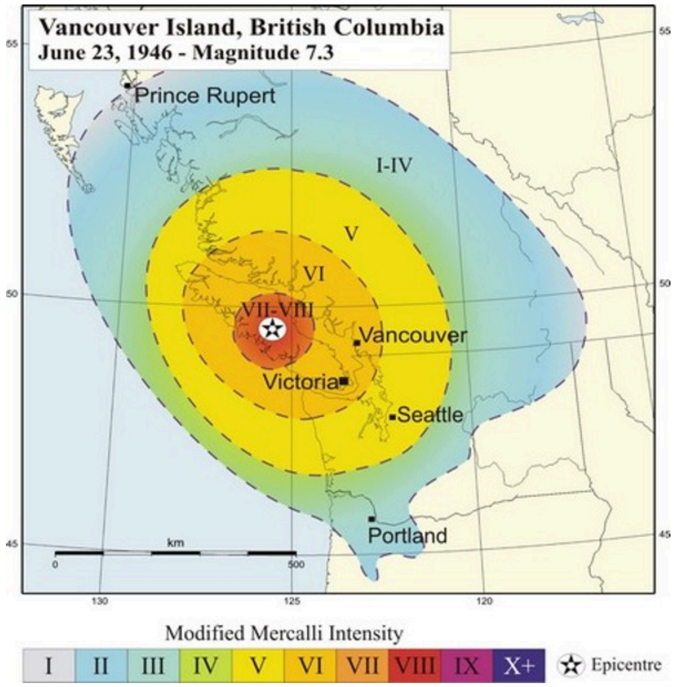

Figure 11.3.4 Intensity map for the 1946 M7.3 Vancouver Island earthquake. [Long Decsription]

Image Descriptions

Figure 11.3.1 image description: P-waves and S-waves from a small (M4) earthquake near Vancouver Island in 1997. The P-wave arrived in 0.7 seconds with an amplitude ranging from negative 0.7 millimetres per second to 1.1 millimetres per second and lasting until the arrival of the S-wave. The S-wave arrived at 8.7 seconds, with a minimum amplitude of negative 2.8 millimetres per second and a maximum amplitude of 2.7 millimetres per second. The S-wave's net amplitude gradually decreased over the next 5 seconds. [Return to Figure 11.3.1]

Figure 11.3.4 image description: The graduated intensity of the 1945 M7.3 Vancouver Island earthquake based on the modified Mercalli intensity scale. The area surrounding the epicentre of the earthquake which included central Vancouver Island ranged between a very strong (7) and severe (8) intensity. The next ring included the northern and southern parts of Vancouver Island, as well as a part of the main land coast including Vancouver and much of the Sunshine coast a strong (6) intensity. The next ring, which reached experienced a moderate (5) intensity, included Seattle and much of the BC interior. The outermost ring ranged between not felt (1) and light (4) intensity. It was felt as far north as Prince Rupert and the southern tip of Haida Gwaii, south eastern BC, and as far south as north western Oregon. [Return to Figure 11.3.4]

Media Attributions

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- **Figure 11.3.2 (right):** "Love Wave" © Nicoguaro. Adapted by Steven Earle. CC BY.
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Text Attributions

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11.4 The Impacts of Earthquakes

Some of the common impacts of earthquakes include structural damage to buildings, fires, damage to bridges, highways, pipelines and electrical transmission lines, initiation of slope failures, liquefaction, and tsunami. The types of impacts depend to a large degree on where the earthquake is located: whether it is predominantly urban or rural, densely or sparsely populated, highly developed or underdeveloped, and of course on the ability of the infrastructure to withstand shaking.

As we've seen from the example of the 1985 Mexico earthquake, the geological foundations on which structures are built can have a significant impact on earthquake shaking. When an earthquake happens, the seismic waves produced have a wide range of frequencies. The energy of the higher frequency waves tends to be absorbed by solid rock, while the lower frequency waves (with periods slower than one second) pass through the solid rock without being absorbed, but are eventually absorbed and amplified by soft sediments. It is therefore very common to see much worse earthquake damage in areas underlain by soft sediments than in areas of solid rock. A good example of this is in the Oakland area near San Francisco, where parts of a two-layer highway built on soft sediments collapsed during the 1989 Loma Prieta earthquake (Figure 11.4.1).

Figure 11.4.1 A part of the Cypress Freeway in Oakland California that collapsed during the 1989 Loma Prieta earthquake.

Building damage is also greatest in areas of soft sediments, and multi-storey buildings tend to be more seriously damaged than smaller ones. Buildings can be designed to withstand most earthquakes, and this practice is increasingly applied in earthquake-prone regions. Turkey is one such region, and even though Turkey had a relatively strong building code in the 1990s, adherence to the code was poor, as builders did whatever they could to save costs, including using inappropriate materials in concrete and reducing the amount of steel reinforcing. The result was that there were over 17,000 deaths in the 1999 M7.6 Izmit earthquake (Figure 11.4.2). After two devastating earthquakes that year, Turkish authorities strengthened the building code further, but the new code has been applied only in a few regions, and enforcement of the code is still weak, as revealed by the amount of damage from a M7.1 earthquake in eastern Turkey in 2011.

Figure 11.4.2 Buildings damaged by the 1999 earthquake in the Izmit area, Turkey.

Fires are commonly associated with earthquakes because fuel pipelines rupture and electrical lines are damaged when the ground shakes (Figure 11.4.3). Most of the damage in the great 1906 San Francisco earthquake was caused by massive fires in the downtown area of the city (Figure 11.4.4). Some 25,000 buildings were destroyed by those fires, which were fuelled by broken gas pipes. Fighting the fires was difficult because water mains had also ruptured. The risk of fires can be reduced through P-wave early warning systems if utility operators can reduce pipeline pressure and close electrical circuits.

Figure 11.4.3 Some of the effects of the 2011 Tohoku earthquake in the Sendai area of Japan. An oil refinery is on fire, and a vast area has been flooded by a tsunami.

Figure 11.4.4 Fires in San Francisco following the 1906 earthquake.

Earthquakes are important triggers for failures on slopes that are already weak. An example is the Las Colinas slide in the city of Santa Tecla, El Salvador, which was triggered by a M7.6 offshore earthquake in January 2001 (Figure 11.4.5).

Ground shaking during an earthquake can be enough to weaken rock and unconsolidated materials to the point of failure, but in many cases the shaking also contributes to a process known as **liquefaction**, in which an otherwise solid body of sediment is transformed into a liquid mass that can flow. When water-saturated sediments are shaken, the grains become rearranged to the point where they are no longer supporting one another. Instead, the water between the grains is holding them apart and the material can flow. Liquefaction can lead to the collapse of buildings and other structures that might be otherwise undamaged. A good example is the collapse of apartment buildings during the 1964 Niigata earthquake (M7.6) in Japan (Figure 11.4.6). Liquefaction can also contribute to slope failures and to fountains of sandy mud (sand volcanoes) in areas where there is loose saturated sand beneath a layer of more cohesive clay.

Figure 11.4.5 The Las Colinas debris flow at Santa Tecla (a suburb of the capital San Salvador) triggered by the January 2001 El Salvador earthquake. This is just one of many hundreds of slope failures that resulted from that earthquake. Over 500 people died in the area affected by this slide.

Figure 11.4.6 Collapsed apartment buildings in the Niigata area of Japan. The material beneath the buildings was liquefied to varying degrees by the 1964 earthquake.

Parts of the Fraser River delta are prone to liquefaction-related damage because the region is characterized by a 2 metre to 3 metre thick layer of fluvial silt and clay over top of at least 10 metres of water-saturated fluvial sand (Figure 11.4.7). Under these conditions, it can be expected that seismic shaking will be amplified and that the sandy sediments will liquefy. This could lead to subsidence and tilting of buildings, and to failure and sliding of the silt and clay layer. Current building-code regulations in the Fraser delta area require that measures be taken to strengthen the ground underneath multi-storey buildings prior to construction.

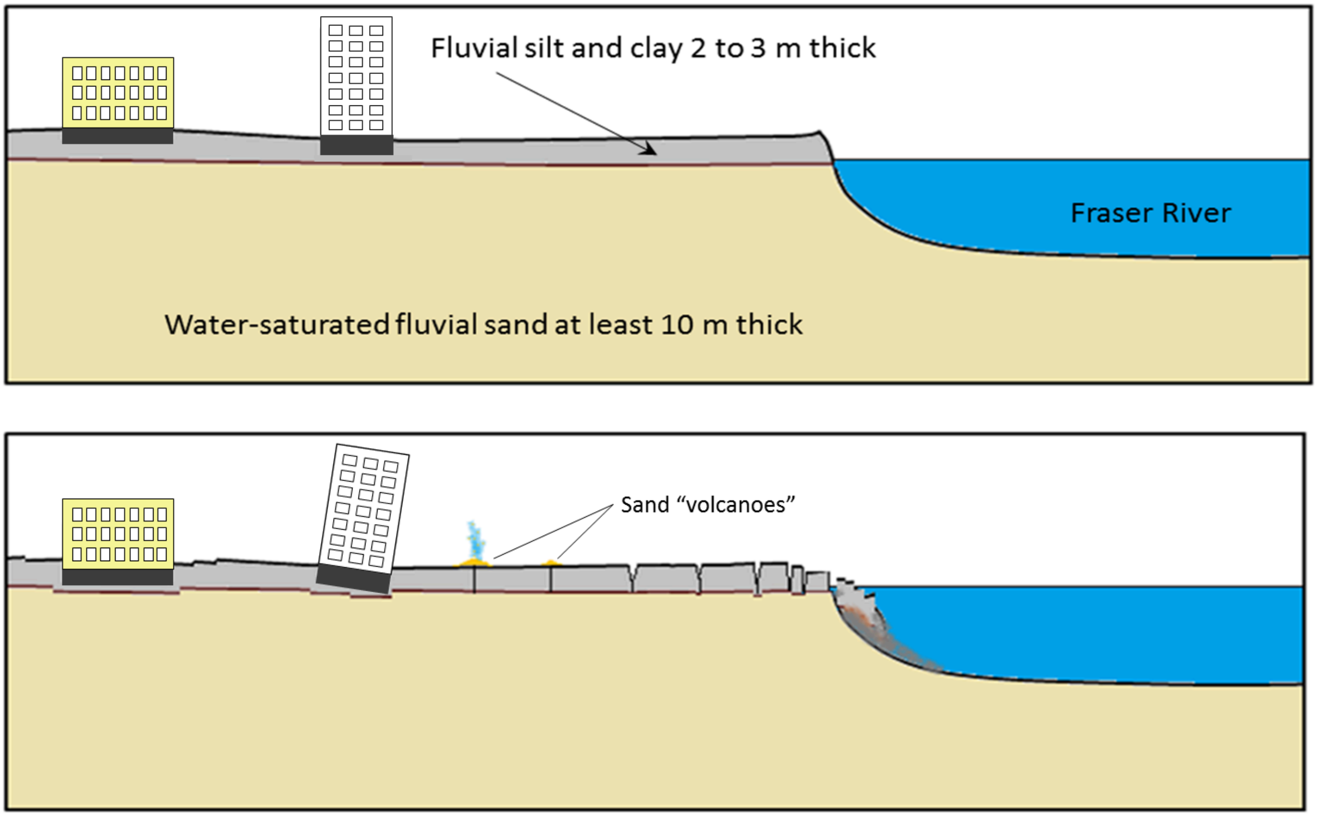

Figure 11.4.7 Recent unconsolidated sedimentary layers in the Fraser River delta area (top) and the potential consequences in the event of a damaging earthquake.

Exercise 11.4 Creating liquefaction and discovering the harmonic frequency

There are a few ways that you can demonstrate the process of liquefaction for yourself. The simplest is to go to a sandy beach (lake, ocean, or river) and find a place near the water’s edge where the sand is wet. This is best done with your shoes off, so let’s hope it’s not too cold! While standing in one place on a wet part of the beach, start moving your feet up and down at a frequency of about once per second. Within a few seconds the previously firm sand will start to lose strength, and you’ll gradually sink in up to your ankles.

Figure 11.4.8

If you can’t get to a beach, or if the weather isn’t cooperating, put some sand (sandbox sand will do) into a small container, saturate it with water, and then pour the excess water off. You can shake it gently to get the water to separate and then pour the excess water away, and you may have to do that more than once. Place a small rock on the surface of the sand; it should sit there for hours without sinking in. Now, holding the container in one hand gently thump the side or the bottom with your other hand, about twice a second. The rock should gradually sink in as the sand around it becomes liquefied.

As you were moving your feet up and down or thumping the pot, it's likely that you soon discovered the most effective rate for getting the sand to liquefy; this would have been close to the natural harmonic frequency for that body of material. Stepping up and down as fast as you can (several times per second) on the wet beach would not have been effective, nor would you have achieved much by stepping once every several seconds. The body of sand vibrates most readily in response to shaking that is close to its natural harmonic frequency, and liquefaction is also most likely to occur at that frequency.

See Appendix 3 for Exercise 11.4 answers.

Earthquakes that take place beneath the ocean have the potential to generate **tsunami**. (Tsunami is the Japanese word for harbour wave. It is the same in both singular and plural.) The most likely situation for a significant tsunami is a large (M7 or greater) subduction-related earthquake. As shown in Figure 11.4.9, during the time between earthquakes the overriding plate becomes distorted by elastic deformation; it is squeezed laterally (Figure 11.4.9B) and pushed up.

When an earthquake happens (Figure 11.4.9C), the plate rebounds and there is both uplift and subsidence on the sea floor, in some cases by as much as several metres vertically over an area of thousands of square kilometres. This vertical motion is transmitted through the water column where it generates a series of waves that then spread across the ocean.

Subduction earthquakes with magnitude less than 7 do not typically generate significant tsunami because the amount of vertical displacement of the sea floor is minimal. Sea-floor transform earthquakes, even large ones (M7 to M8), don't typically generate tsunami either, because the motion is mostly side to side, not vertical.

Tsunami waves travel at velocities of several hundred kilometres per hour and easily make it to the far side of an ocean in about the same time as a passenger jet. The simulated one shown in Figure 11.4.10 is similar to that created by the 1700 Cascadia earthquake off the coast of British Columbia, Washington, and Oregon, which was recorded in Japan nine hours later.

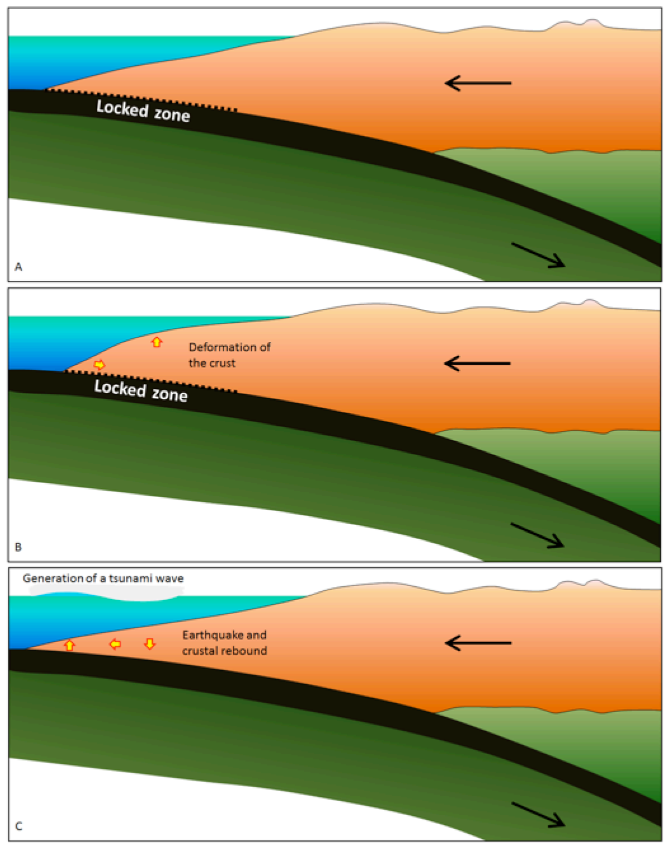

Figure 11.4.9 Elastic deformation and rebound of overriding plate at a subduction setting (B). The release of the locked zone during an earthquake (C) results in both uplift and subsidence on the sea floor, and this is transmitted to the water overhead, resulting in a tsunami.

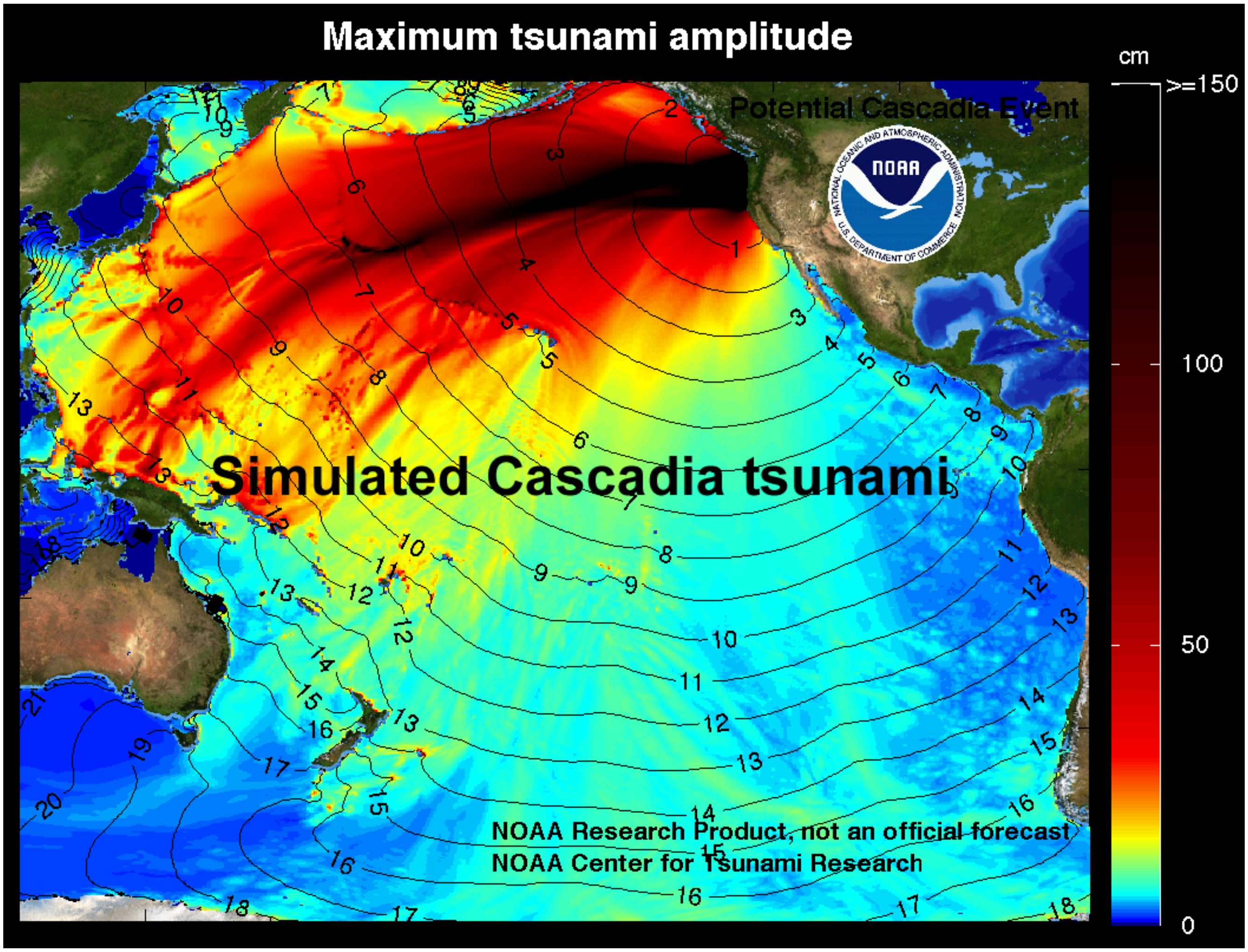

Figure 11.4.10 Model of the tsunami from the 1700 Cascadia earthquake (around M9) showing open-ocean wave heights (colours) and travel time contours. Tsunami wave amplitudes typically increase in shallow water.

Tsunami are discussed further in Chapter 17 under the topic of waves and coasts.

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- Figure 11.4.1: "Cypress Collapsed" by the USGS. Public domain.
- Figure 11.4.2: "Izmit, Turkey Earthquake" by the USGS. Public domain.
- Figure 11.4.3: "SH-60B helicopter flies over Sendai" by the United States Navy. Public domain.
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- Figure 11.4.5: "ElSalvadorslide" by the USGS. Public domain.
- Figure 11.4.6: "Liquefaction at Niigata." Public domain.
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- Figure 11.4.10: "Maximum Tsunami Amplitude" from NOAA/PMEL/Center for Tsunami Research. Public domain.

11.5 Forecasting Earthquakes and Minimizing Damage and Casualties

It has long been a dream of seismologists, geologists, and public safety officials, to be able to accurately predict the location, magnitude, and timing of earthquakes on time scales that would be useful for minimizing danger to the public and damage to infrastructure (e.g., weeks, days, hours). Many different avenues of prediction have been explored, such as using observations of warning foreshocks, changes in magnetic fields, seismic tremor, changing groundwater levels, strange animal behaviour, observed periodicity, stress transfer considerations, and others. So far, none of the research into earthquake prediction has provided a reliable method. Although there are some reports of successful earthquake predictions, they are rare, and many are surrounded by doubtful circumstances.

The problem with earthquake predictions, as with any other type of prediction, is that they have to be accurate *most* of the time, not just *some* of the time. We have come to rely on weather predictions because they are generally (and increasingly) accurate. But if we try to predict earthquakes and are only accurate 10% of the time (and even that isn't possible with the current state of knowledge), the public will lose faith in the process very quickly, and then will ignore all of the predictions. Efforts are currently focused on forecasting earthquake probabilities, rather than predicting their occurrence.

There was great hope for earthquake predictions late in the 1980s when attention was focused on part of the San Andreas Fault at Parkfield, about 200 kilometres south of San Francisco. Between 1881 and 1965 there were five earthquakes at Parkfield, most spaced at approximately 20-year intervals, all confined to the same 20 kilometre-long segment of the fault, and all very close to M6 (Figure 11.5.1). Both the 1934 and 1966 earthquakes were preceded by small foreshocks exactly 17 minutes before the main quake.

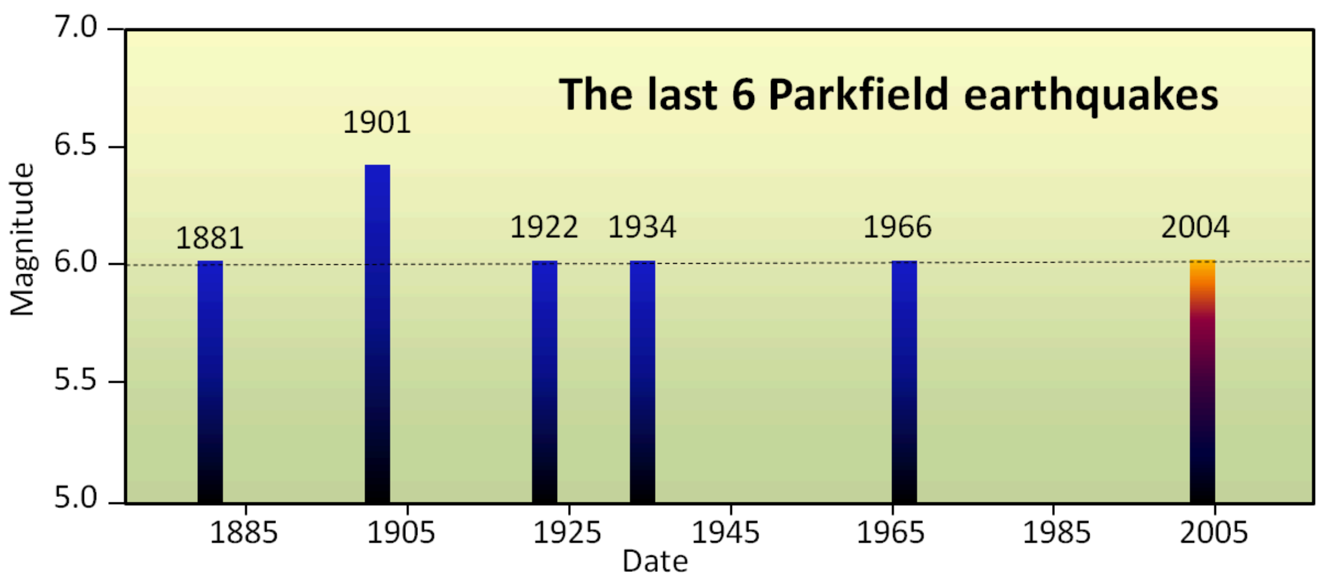

Figure 11.5.1 Earthquakes on the Parkfield segment of the San Andreas Fault between 1881 and 2004.

The U.S. Geological Survey recognized this as an excellent opportunity to understand earthquakes and earthquake prediction, so they armed the Parkfield area with a huge array of geophysical instruments and waited for the next quake, which was expected to happen around 1987. Nothing happened! The "1987 Parkfield earthquake" finally struck in September 2004. Fortunately all of the equipment was still there, but it was no help from the perspective of earthquake prediction. There were no significant precursors to the 2004 Parkfield earthquake in any of the parameters measured, including seismicity, harmonic tremor, strain (rock deformation), magnetic field, the conductivity of the rock, or creep, and there was no foreshock. In other words, even though every available technique was used to monitor it, the 2004 earthquake came as a complete surprise, with no warning whatsoever.

The hope for earthquake prediction is not dead, but it was hit hard by the Parkfield experiment. The current focus in earthquake-prone regions is to provide forecasts of the probability of an earthquake of a certain magnitude within a certain time period—typically a number of decades—while officials focus on ensuring that the population is educated about earthquake risks and that buildings and other infrastructure are as safe as can be. An example of this approach for the San Francisco Bay region of California is shown in Figure 11.5.2. Based on a wide range of information, including past earthquake history, accumulated stress from plate movement, and known stress transfer, seismologists and geologists have predicted the likelihood of a M6.7 or greater earthquake on each of eight major faults that cut through the region. The greatest probabilities are on the Hayward, Rogers Creek, Calaveras, and San Andreas Faults. As shown in Figure 11.5.2, there is a 72% chance that a major and damaging earthquake will take place somewhere in the region prior to 2043.

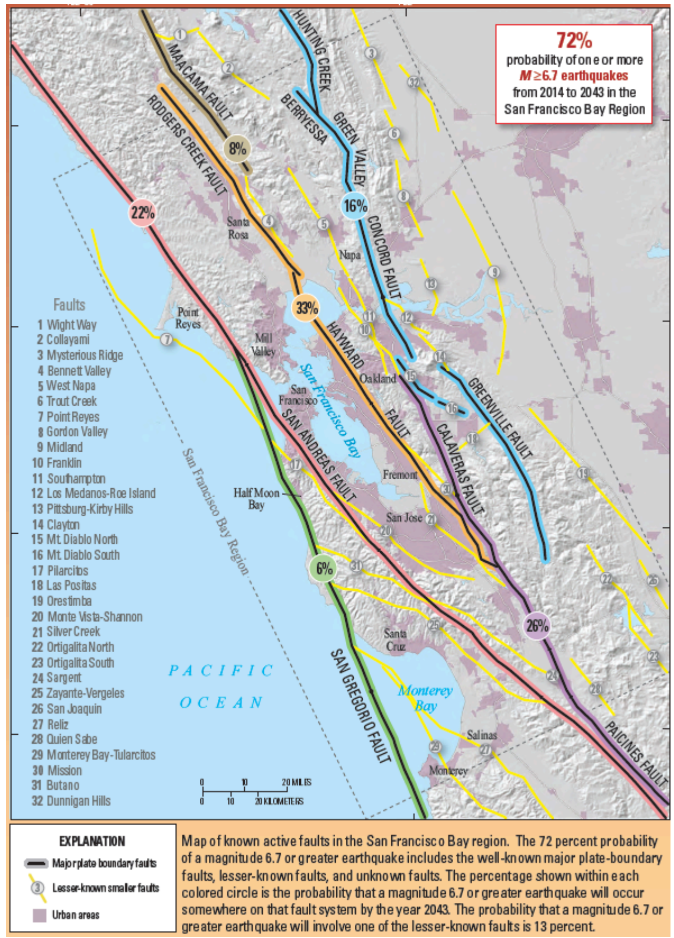

Figure 11.5.2 Probabilities of a M6.7 or larger earthquake over the period 2014 to 2043 on various faults in the San Francisco Bay region of California.

As we've discussed already, it's not sufficient to have strong building codes, they have to be enforced. Building code compliance is quite robust in most developed countries, but is sadly inadequate in many developing countries.

It's also not enough just to focus on new buildings; we have to make sure that existing buildings—especially schools and hospitals—and other structures such as bridges and dams, are as safe as they can be. An example of how this is applied to schools in B.C. is described in Exercise 11.5.

Making the Seismic Upgrade in B.C.'s Schools

British Columbia is in the middle of a multi-billion-dollar program to make schools safer for students. The program is focused on older schools, because, according to the government, those built since 1992 already comply with modern seismic codes. Some schools would require too much work to make upgrading economically feasible and they are slowly being replaced. Where upgrading is feasible, the school is assessed carefully before any upgrade work is initiated.

An example is Sangster Elementary in Colwood on southern Vancouver Island. The school was originally built in 1957, with a major addition in 1973. Ironically, the newer part of the school, built of concrete blocks, required strengthening with the addition of a steel framework, while the 1957 part, which is a wood-frame building, did not require seismic upgrading. The work was completed in 2014.

Figure 11.5.3 Sangster Elementary image from Google Maps – street view.

As of July 2019, upgrades had been completed at 168 B.C. schools, 17 were underway, and an additional 26 were ready to proceed, with funding identified. Another 274 schools were listed as needing upgrades.[1]

Exercise 11.5 Is your local school on the seismic upgrade list?

Here is The B.C. Ministry of Education's list of schools in the seismic mitigation program as of July 2019: Progress Report: Seismic Mitigation Program (BC Schools).

If you live in B.C., you can check to see if any of the schools in your area are on the list. If so, you might be able to find out, either from the school or on the Internet, what type of work has been done or is planned.

The seismic mitigation program has a strong focus on the Lower Mainland and Vancouver Island. Why do you think that is the case, and is it reasonable?

See Appendix 3 for Exercise 11.5 answers.

The final part of earthquake preparedness involves the formulation of public emergency plans, including escape routes, medical facilities, shelters, and food and water supplies. It also includes personal planning, such as emergency supplies (food, water, shelter, and warmth), escape routes from houses and offices, and communication strategies (with a focus on ones that don't involve the cellular network).

Media Attributions

- Figure 11.5.1: © Steven Earle. CC BY.
- Figure 11.5.2: "San Francisco Bay region Earthquake Probability [PDF]" by USGS. Public domain.

1. "Seismic Mitigation Program Progress Report, July 2019." Accessed July 21, 2019 from https://www2.gov.bc.ca/gov/content/education-training/k-12/administration/capital/seismic-mitigation

- Figure 11.5.3: Sangster Elementary school. © 2018 Google Maps. Street View. See Google's permissions for non-commercial use.

Summary

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

Section	Summary
11.1 What is an Earthquake?	An earthquake is the shaking that results when a body of rock that has been deformed breaks and the two sides quickly slide past each other. The rupture is initiated at a point but quickly spreads across an area of a fault, via a series of aftershocks initiated by stress transfer. Episodic tremor and slip is a periodic slow movement, accompanied by harmonic tremors, along the middle part of a subduction zone boundary.
11.2 Earthquakes and Plate Tectonics	Most earthquakes take place at or near plate boundaries, especially at transform boundaries (where most quakes are at less than a 30 kilometre depth) and at convergent boundaries (where they can be at well over a 100 kilometre depth). The largest earthquakes happen at subduction zones, typically in the upper section where the rock is relatively cool.
11.3 Measuring Earthquakes	Magnitude is a measure of the amount of energy released by an earthquake, and it is proportional to the area of the rupture surface and to the amount of displacement. Although any earthquake has only one magnitude value, it can be estimated in various ways, mostly involving seismic data. Intensity is a measure of the amount of shaking experienced and damage done at a particular location around the earthquake. Intensity will vary depending on the distance to the epicentre, the depth of the earthquake, and the geological nature of the material below surface.
11.4 The Impacts of Earthquakes	Damage to buildings is the most serious consequence of most large earthquakes. The amount of damage is related to the type and size of buildings, how they are constructed, and the nature of the material on which they are built. Other important consequences are fires, damage to bridges and highways, slope failures, liquefaction, and tsunami. Tsunami, which are almost all related to large subduction earthquakes, can be devastating.
11.5 Forecasting Earthquakes and Minimizing Damage and Casualties	There is no reliable technology for predicting earthquakes, but the probability of one happening within a certain time period can be forecast. We can minimize earthquake impacts by ensuring that citizens are aware of the risk, that building codes are enforced, that existing buildings like schools and hospitals are seismically sound, and that both public and personal emergency plans are in place.

Questions for Review

1. Define the term *earthquake*.
2. How does elastic rebound theory help to explain how earthquakes happen?
3. What is a rupture surface, and how does the area of a rupture surface relate to earthquake magnitude?

4. What is an aftershock and what is the relationship between aftershocks and stress transfer?
5. Episodic slip on the middle part of the Cascadia subduction zone is thought to result in an increase in the stress on the upper part where large earthquakes take place. Why?
6. Explain the difference between magnitude and intensity as expressions of the size of an earthquake.
7. How much more energy is released by an M7.3 earthquake compared with an M5.3 earthquake?
8. Figure A shows a map of earthquake locations with the depths coded according the colour scheme used in Figure 11.2.5. What type of plate boundary is this?
9. Draw a line on the map to show approximately where the plate boundary is situated.
10. In which directions are the plates moving, and where in the world might this be?
11. Earthquakes are relatively common along the mid-ocean ridges. At what type of plate boundary do most such quakes occur?
12. The northward motion of the Pacific Plate relative to the North America Plate takes place along two major transform faults. What are they called?

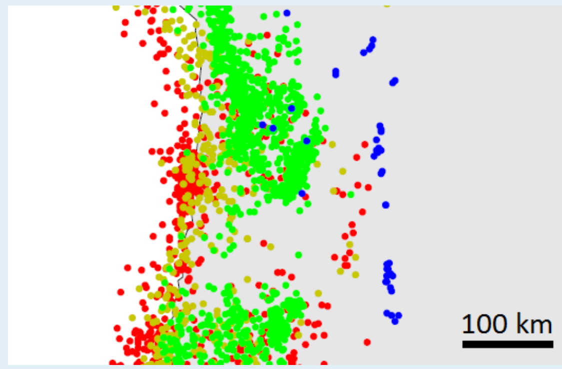

Figure A

13. Why is earthquake damage likely to be more severe for buildings built on unconsolidated sediments as opposed to solid rock?
14. Why are fires common during earthquakes?
15. What type of earthquake is likely to lead to a tsunami?
16. What did we learn about earthquake prediction from the 2004 Parkfield earthquake?
17. What are some of the things we should know about an area in order to help minimize the impacts of an earthquake?
18. What is the difference between earthquake prediction and forecasting?

Answers to Review Questions can be found in Appendix 2.

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