Chapter 12.

Earthquakes

# What is an Earthquake?

### Earthquake Shaking Comes from Elastic Deformation

Earthquakes occur when rock **ruptures** (breaks), causing rocks on one side of a fault to move relative to the rocks on the other side. Although motion along a fault is part of what happens when an earthquake occurs, rocks grinding past each other is not what creates the shaking. In fact, it could be said that the earthquake happens *after* rocks have undergone most of the displacement. Consider this: if rocks slide a few centimeters or even meters along a fault, would that motion *alone* explain the incredible damage caused by some earthquakes? If you were in a car that suddenly accelerated then stopped, you would feel a jolt. But earthquakes are not a single jolt. Buildings can swing back and forth until they shake themselves to pieces, train tracks can buckle and twist into s-shapes, and roads can roll up and down like waves on the ocean. During an earthquake, rock is not only slipping. It is also vibrating like a plucked guitar string.

Rocks might seem rigid, but when stress is applied they may

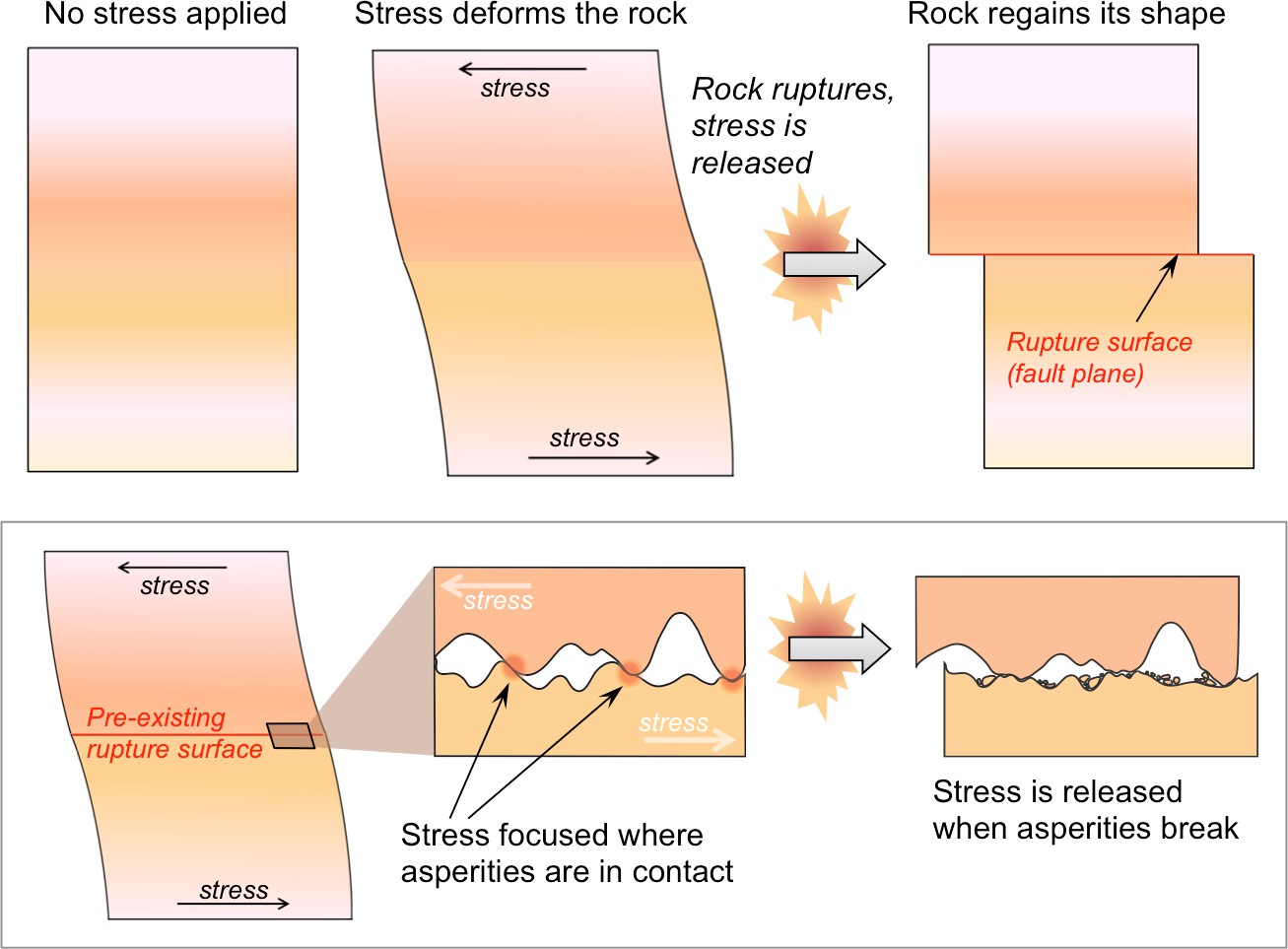
stretch. If there hasn’t been too much stretching, a rock will snap back to its original shape once the stress is removed. Deformation that is reversible is called **elastic deformation**. Rocks that are stressed beyond their ability to stretch can rupture, allowing the rest of the rock to snap back to its original shape. The snapping back of the rock returning to its original shape causes the rock to vibrate, and this is what causes the shaking during an earthquake. The snapping back is called **elastic rebound**.

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Figure 12.1 (top) shows this sequence of events. Stress is applied to a rock and deforms it. The deformed rock ruptures, forming a fault

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After rupturing, the rock above and below the fault snaps back to the shape it had before deformation.



***Figure 12.1*** *Elastic deformation, rupture, and elastic rebound. Top: Stress applied to a rock causes it to deform by stretching. When the stress becomes too much for the rock, it ruptures, forming a fault. The rock snaps back to its original shape in a process called elastic rebound. Bottom: On an existing fault, asperities keep rocks on either side of the fault from sliding. Stress deforms the rock until the asperities break, releasing the stress, and causing the rocks to spring back to their original shape. Source: Karla Panchuk (2017) CC BY 4.0. Modified after Steven Earle (2015) CC BY 4.0 view original.*

Ruptures can also occur along pre-existing faults (Figure 12.1, bottom). The rocks on either side of the fault are locked together because bumps along the fault, called **asperities**, prevent the rocks from moving relative to each other. When the stress is great enough to break the asperities, the rocks on either side of the fault can slide again. While the rocks are locked together, stress can cause elastic deformation. When asperities break and release the stress, the rocks undergo elastic rebound and return to their original shape.

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### Rupture Surfaces Are Where the Action Happens

Images like 12.2 are useful for illustrating elastic deformation and rupture, but they can be misleading. The rupture that happens doesn’t occur as in 12.2, with the block being ruptured through and through. The rupture and displacement only happen along a subsection of a fault, called the **rupture surface**. In Figure 12.3, the rupture surface is the dark pink patch. It takes up only a part of the **fault plane** (lighter pink). The fault plane represents the surface where the fault exists, and where ruptures have happened in the past. Although the fault plane is drawn as being flat in Figure 12.4, faults are not actually perfectly flat.

The location on the fault plane where the rupture happens is called the **hypocenter** or **focus** of the earthquake (Figure 12.4, right). The location on Earth’s surface immediately above the hypocenter is the **epicenter** of the earthquake.

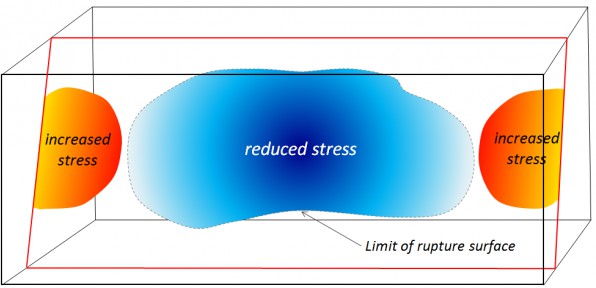
### Shifting Stress Causes Foreshocks and Aftershocks

Earthquakes don’t usually occur in isolation. There is often a sequence in which smaller earthquakes occur prior to a larger one, and then progressively smaller earthquakes occur after. The largest earthquake in the series is the **mainshock**. The smaller ones that come before are **foreshocks**, and the smaller ones that come after are **aftershocks**. These descriptions are relative, so it can be necessary to reclassify an earthquake. For example, the strongest earthquake in a series is classified as the mainshock, but if another even bigger one comes after it, the bigger one is called the mainshock, and the earlier one is reclassified as an aftershock.

A rupture surface does not fail all at once. A rupture in one place leads to another, which leads to another. Aftershocks and foreshocks represent the same thing, except on a much larger scale.

The rupture illustrated in Figure 9.4 reduced stress in one area, but in doing so, transferred stress to others (Figure 9.5). Imagine a frayed rope breaking strand by strand. When a strand breaks, the tension on that strand is released, but the remaining strands must still hold up the same amount of weight. If another strand breaks under the increased burden, the remaining strands have an even greater burden than before. In the same way that the stress causes one strand after another to fail, a rupture can trigger subsequent ruptures nearby.

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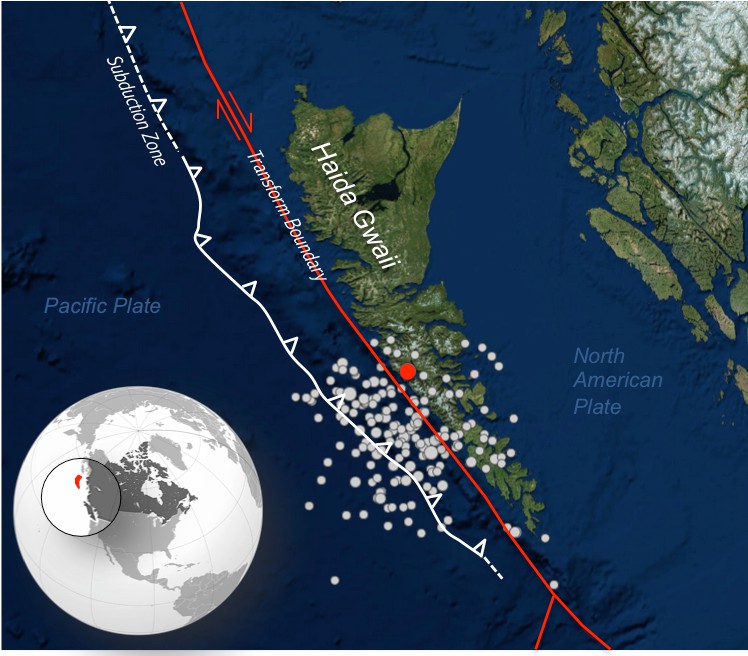


***Figure 12.2*** *Stress changes related to an earthquake. Stress decreases in the area of the rupture surface, but increases on adjacent parts of the fault.*

*Source: Steven Earle (2015) CC BY 4.0 view source.*

Numerous aftershocks were associated with the magnitude 7.8 earthquake that struck Haida Gwaii, B.C., in October of 2012 (Figure 12.3; mainshock in red, aftershocks in white). Some of the stress released by the mainshock was transferred to other nearby parts of the fault, and contributed to a cascade of smaller ruptures. But stress transfer need not be 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 aftershocks from the Haida Gwaii earthquake are scattered rather than located only on the main faults.

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***Figure 12.3*** *Magnitude 7.8 Haida Gwaii earthquake and aftershocks. Mainshock (red circle marks the epicenter) occurred on October 28th, 2012. Aftershocks are for the period from October 28th to November 10th of 2012. Although the epicenter is near a transform boundary, the rupture was influenced more by compression related to the subduction zone. Source: Karla Panchuk (2017) CC BY 4.0. Base map with epicenters from the U. S. Geological Survey Latest Earthquakes tool view interactive map. Subduction zone after Wang et al. (2015). Click the image for more attributions.*

The effects of stress transfer may not show immediately. Aftershocks can be delayed for hours, days, weeks, or even years. Because stress transfer affects a region, not just a single fault, and because there can be delays between the event that transferred stress and the one that was triggered by the transfer, it can sometimes be hard to be know whether one earthquake is actually associated with another, and whether a **foreshock** or aftershock should be assigned to a particular mainshock.

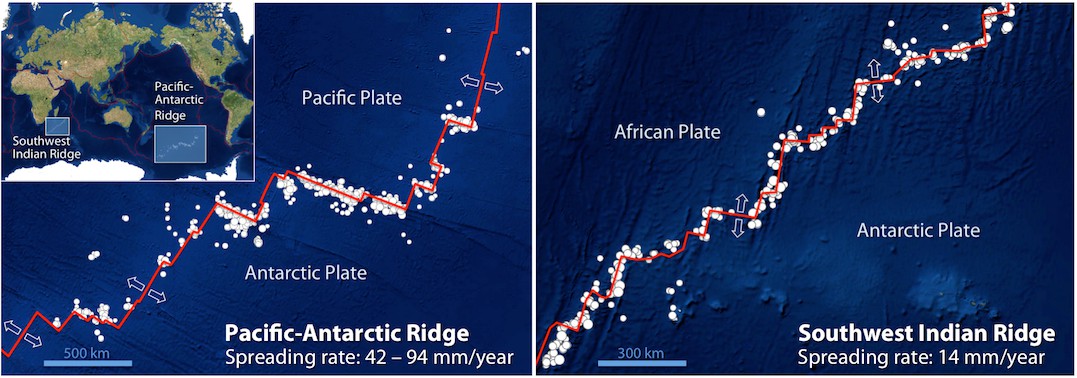
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### Earthquakes at Divergent and Transform Plate Boundaries

Earthquakes along divergent and transform plate margins are shallow (usually less than 30 km deep) because below those depths, rock is too hot and weak to avoid being permanently deformed by the stresses in those settings. If deformation is permanent, then removing the stress does not result in the rocks snapping back to their original shape. No snapping back means no shaking.

Mid-ocean ridge divergent plate margins are offset by numerous transform faults (Figure 12.4). The locations of earthquakes along mid-ocean ridges, and the mechanisms for causing them, depend on how rapidly the mid-ocean ridges are spreading. The Pacific- Antarctic Ridge (left) is spreading relatively rapidly at 42 to 94 mm/ year, depending on the location along the ridge. Rapid spreading causes rocks near the axis of the spreading center to be hot and weak. As a result, most of the earthquakes (white dots) are located along transform faults, where rocks are cooler and stronger. Along rapidly spreading ridges, new ocean crust is bent upward into wide, high ridges. As spreading proceeds and crust moves away from the ridge, the bend is relaxed, and the crust stretches and breaks. This triggers earthquakes many kilometers away from the ridge.

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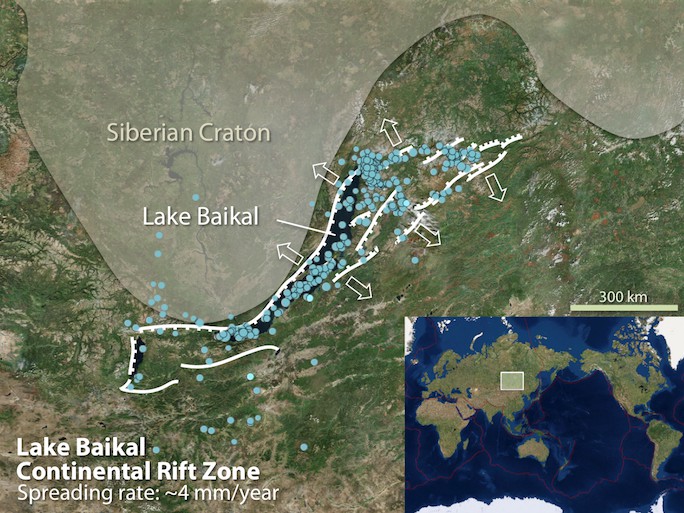


***Figure 12.4*** *Locations of earthquakes of magnitude 4 and greater from 1990 to 2010 along two mid-ocean ridges. Plate boundaries are marked in red. Arrows show the direction of plate motion. Left: Rapidly spreading Pacific-Antarctic ridge with earthquakes concentrated along transform faults. Right: Slowly spreading Southwest Indian Ridge, with earthquakes along both spreading segments and transform faults. Source: Karla Panchuk (2017) CC BY 4.0. Base maps with epicentres generated using the U. S. Geological Survey Latest Earthquakes website. Visit Latest Earthquakes*

The Southwest Indian Ridge (right) spreads very slowly, at approximately 14 mm/year. Rocks are cooler and stronger along the slowly spreading ridge than along the rapidly spreading one. In the slow-spreading environment, earthquakes are generated when rocks along the ridge axis stretch and break. Earthquakes are more evenly distributed between divergent and transform segments of the boundary than they are along fast-spreading ridges.

Earthquakes in continental rift zones are also shallow, but scattered more broadly than those along mid-ocean ridges. Lake Baikal (Figure 12.5), the world’s oldest, deepest, and largest freshwater lake, formed 25 million years ago because of continental rifting. Note the scale in Figure 12.18, and compare how widely the earthquakes (blue dots) are spread in the Lake Baikal region, versus along the mid-ocean ridges in Figure 12.5.

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***Figure 11.5*** *Small circles mark the locations of earthquakes of M4 and greater from 1990 to 2010 along the Lake Baikal rift zone. White lines show some of the faults in the region. White lines with tick marks are normal faults related to spreading. Arrows show the direction of spreading. White lines without tick marks are transform faults. The Siberian Craton (shaded region) is strong 2 billion year old crust. Source: Karla Panchuk (2017) CC BY 4.0. Faults after U.*

*S. Geological Survey (see references). Base maps (inset and rift views) with epicenters generated using the U. S. Geological Survey Latest Earthquakes website. Visit Latest Earthquakes*

One reason for the difference in earthquake distribution in continental rift zones is that the rifts are only beginning to form. Faulting is “disorganized” within the continental crust. There is no well-established spreading center, unlike mid-ocean ridges. Another reason is that the locations of faults, and thus earthquakes, in continental rift zones are affected by pre-existing geological structures within continental crust. In the case of the Lake Baikal rift, the strong, ancient crust of the Siberian Craton influences the orientation of the faults forming the rift. Faults run parallel to the **craton** near Lake Baikal. As rifting extends to the east, the part

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of the craton in the upper right of Figure 12.18 may deflect rifting southward.

# Measuring Earthquakes

The shaking from an earthquake travels away from the rupture in the form of **seismic waves**. Seismic waves are measured to determine the location of the earthquake, and to estimate the amount of energy released by the earthquake (its **magnitude**).

### Types of Seismic Waves

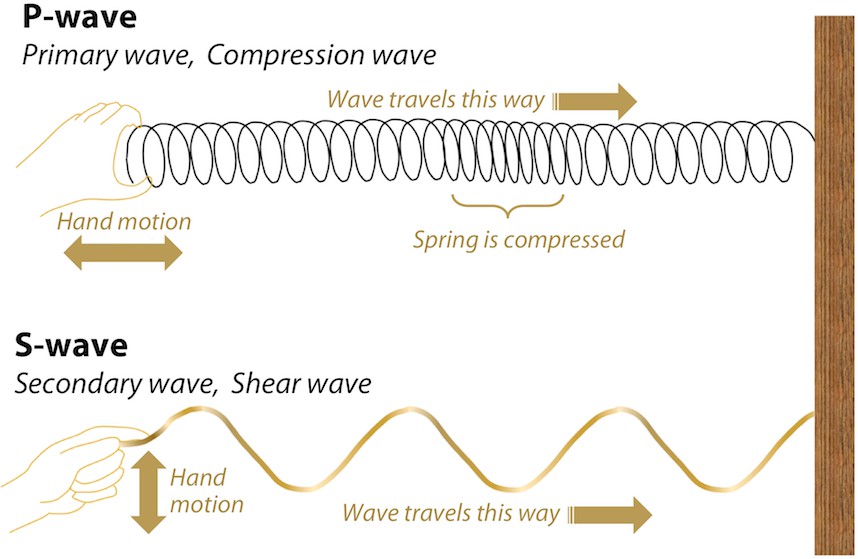
Seismic waves are classified according to where they travel, and how they move particles.

#### Body Waves

Seismic waves that travel through Earth’s interior are called **body waves**. **P-waves** are body waves that move by alternately compressing and stretching materials in the direction the wave moves. For this reason, P-waves are also called compression waves. The “P” in P-wave stands for primary, because P-waves are the fastest of the seismic waves. They are the first to be detected when an earthquake happens.

A P-wave can be simulated by fixing one end of a spring to a solid surface, then giving the other end a sharp push toward the surface (Figure 12.6, top). The compression will propagate (travel) along the length of the spring. Some parts of the spring will be stretched, and others compressed. Any one point on the spring will jiggle forward and backward as the compression travels along the spring.

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***Figure 12.6*** *Seismic waves simulated using a spring and rope attached to a fixed surface. Top: P-waves travel as pulses of compression. Bottom:*

*S-waves move particles at right angles to the direction of motion.*

*Source: Karla Panchuk (2018) CC BY*

*4.0 modified after Steven Earle (2015)*

*CC BY 4.0*

*view original.*

**S-waves** are body waves that move with a shearing motion, shaking particles from side to side. S-waves can be simulated by fixing one end of a rope to a solid surface, then giving the other end a flick (Figure 12.6, bottom). Any one point on the rope will move from side to side at a right angle to the direction in which the snaking motion is traveling. The “S” in S-wave stands for secondary, because S- waves are slower than P-waves, and are detected after the P-waves are measured. S-waves cannot travel through liquids.

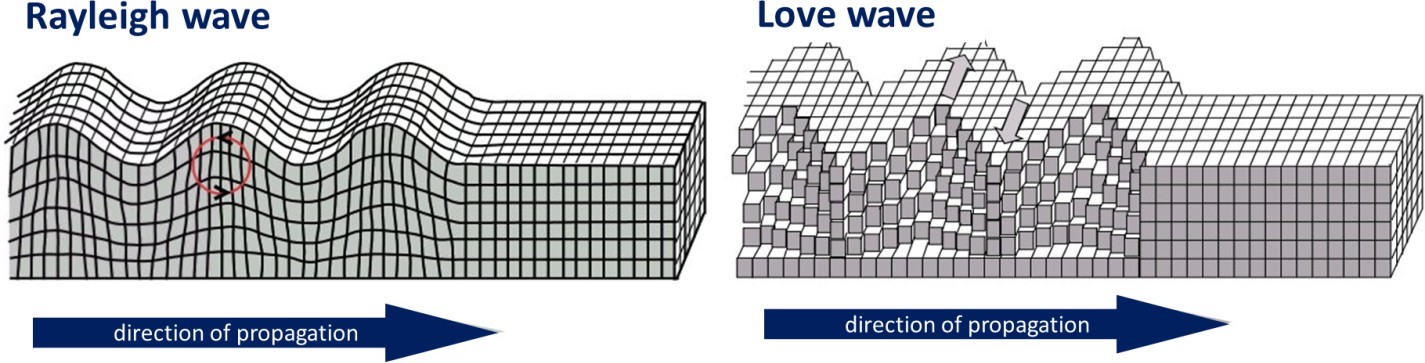
P-waves and S-waves can travel rapidly through geological

materials, at speeds many times the speed of sound in air.

#### Surface Waves

When body waves reach Earth’s surface, some of their energy is transformed into surface waves, which travel along Earth’s surface. Two types of surface waves are **Rayleigh waves** and **Love waves** (Figure 12.7). Rayleigh waves (R-waves) are characterized by vertical motion of the ground surface, like waves rolling on water. Love waves (L-waves) are characterized by side-to-side motion. Notice that the effects of both kinds of surface waves diminish with depth in Figure 12.7.

Surface waves are slower than body waves, and are detected after the body waves. Surface waves typically cause more ground motion than body waves, and therefore do more damage than body waves.



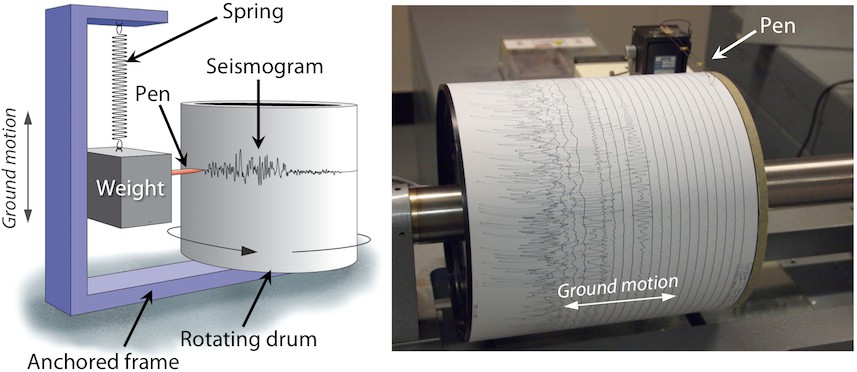
***Figure 12.7*** *Surface waves travel along Earth’s surface and have a diminished impact with depth. Rayleigh waves (left) cause a rolling motion, and Love waves (right) cause the ground to shift from side to side. Source: Steven Earle (2015) CC BY 4.0 view source. Click the image for more attributions.*

### Recording Seismic Waves Using a Seismograph

A **seismometer** is an instrument that detects seismic waves. An instrument that combines a seismometer with a device for recording the waves is called a s**eismograph**. The graphical output from a seismograph is called a **seismogram**. Figure 12.8 (right) shows how a seismograph works. The instrument consists of a frame or

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housing that is firmly anchored to the ground. A mass is suspended from the housing, and can move freely on a spring. When the ground shakes, the housing shakes with it, but the mass remains fixed. A pen attached to the mass moves up and down on a rotating drum of paper, drawing a wavy line, the seismogram. The seismograph in Figure 12.8 (right) is oriented to measure vertical ground motion. The photo on the left shows a seismograph oriented to record horizontal ground motion.



***Figure 12.8*** *How a seismograph records earthquakes. Source: Left- Karla Panchuk (2018) CC BY-NC-SA 4.0 modified after IRIS (2012) “How Does a Seismometer Work?” view source; Right: Karla Panchuk (2018) CC BY-SA 4.0, photo by Z22 (2014) CC BY-SA 3.0 view source. Click the image for more attributions.*

The pen and drum of a mechanical seismograph record the motion of the ground relative to the mass. However, unless an earthquake causes a large amount of ground motion directly beneath the seismograph, the height of the wave recorded on paper might be very small, making the seismogram difficult to analyze. The seismograph on the right has a device to amplify the ground motion, drawing larger waves that are easier to study.

Modern seismographs record shaking as electrical signals, and are able to transmit those signals. This means seismologists need not return to the instrument to collect recordings before the records can be examined.

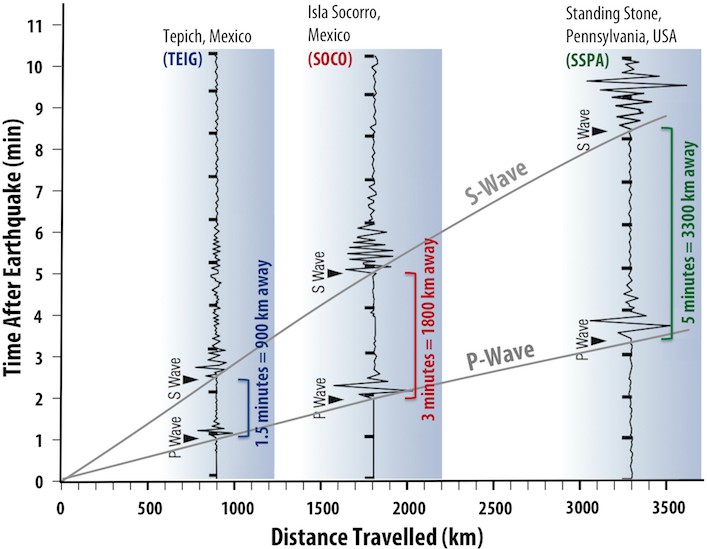
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### Finding The Location of an Earthquake

P-waves travel faster than S-waves. As the waves travel away from the location of an earthquake, the P-wave gets farther and father ahead of the S-wave. Therefore, the farther a seismograph is from the location of an earthquake, the longer the delay between when the P-wave arrival is recorded, and the S-wave arrival is recorded. The delay between the P-wave and S-wave arrival appears as a widening gap in a diagram of P-wave and S-wave travel times (Figure 12.9, grey lines).

P-wave and S-wave arrival times can be identified on seismograms. In the three seismograms in Figure 12.9, the arrivals of the P-waves and S-waves are marked with arrows, and the interval in minutes between the P-wave and S-wave arrivals are noted. The seismograms were recorded at three different **seismic stations** (earthquake monitoring locations equipped with seismographs). The distance of each station from the earthquake is determined by finding the distance along the graph where the gap between the P- wave and S-wave travel-time curves matches the delay between P- wave and S-wave arrivals on the seismogram.

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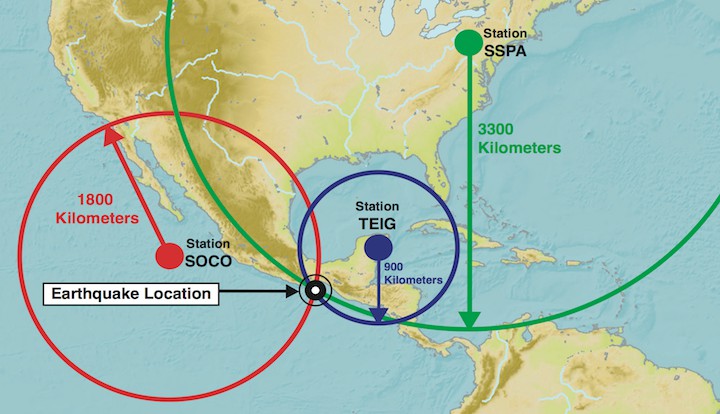


***Figure 12.9*** *Using P-wave and S-wave travel times to determine how far seismic waves have traveled. Grey curves show the distance traveled by*

*P-waves and S-waves after an earthquake occurs. P-waves are faster than S-waves, and the gap between them increases with time and distance. The*

*delays between P-wave and S-wave arrivals on seismograms are matched to the curve to find the distances of seismic stations from the source of the seismic waves. Source: Karla Panchuk (2018) CC BY-NC-SA 4.0 modified after IRIS (n.d.) “How Are Earthquakes Located?” view source*

The delay between the P-wave and S-wave arrival at a seismic station can indicate how far the station is from the source of the earthquake, but not the direction from which the seismic waves travelled. The possible locations of the earthquake can be represented on a map by drawing a circle around the seismic station, with the radius of the circle being the distance determined from the P-wave and S-wave travel times (Figure 12.10). If this is done for at least three seismic stations, the circles will intersect at the origin of the earthquake. 380



***Figure 12.10*** *Locating earthquakes by drawing three circles with radii of lengths determined from P-wave and S-wave travel times. Where the radii intersect from all three circles is the epicenter of the earthquake. Station names (SOCO, TEIG, SSPA) correspond to seismograms in Figure 12.11. Source: IRIS (n.d.) “How Are Earthquakes Located?” view source Click the image for terms of use.*

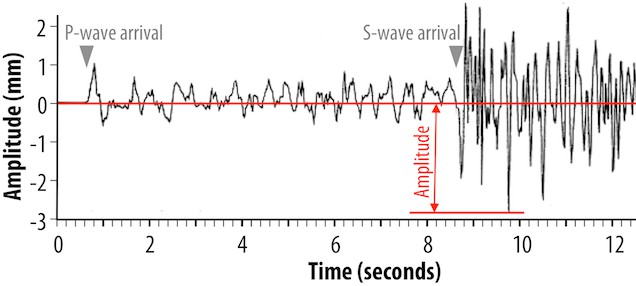
### How Big Was It?

Earthquakes can be described in terms of their **magnitude**, which reflects the amount of energy released by the shaking. They can also be described in terms of **intensity**, which characterizes the impact of the shaking on people and their surroundings.

#### Earthquake Magnitude

Earthquake magnitudes are determined by measuring the amplitudes of seismic waves. The **amplitude** is the height of the wave relative to the baseline (Figure 12.11). Wave amplitude depends on the amount of energy carried by the wave. The amplitudes of 381

seismic waves reflect the amount of energy released by earthquakes.



***Figure 12.11*** *Seismogram for a small earthquake that occurred near Vancouver Island in 1997. The maximum amplitude of the S-wave is indicated. Source: Karla Panchuk (2018) CC BY 4.0 modified after Steven Earle (2015) CC BY 4.0 view source*

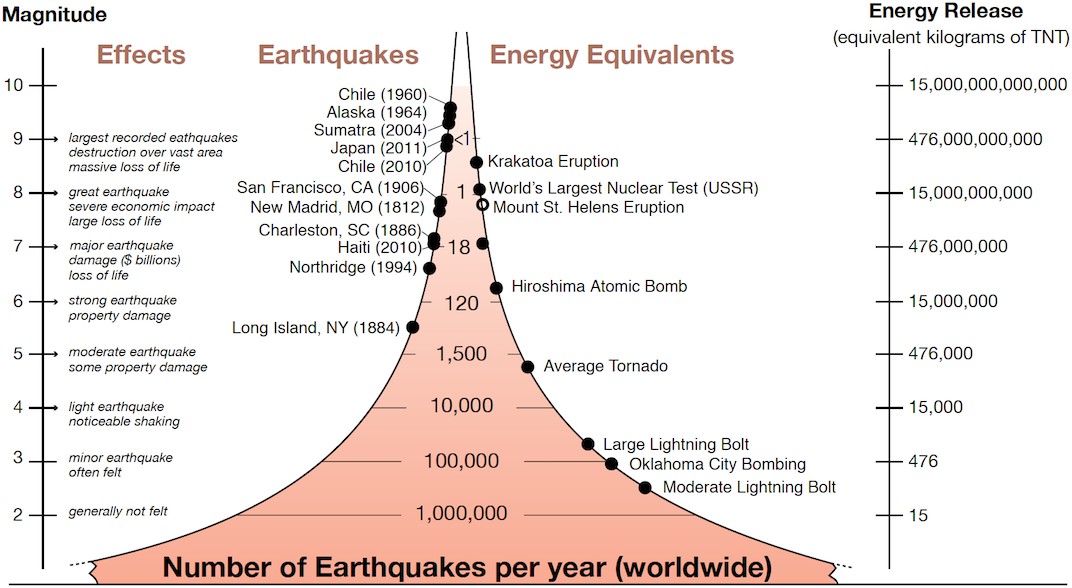
The **Richter magnitude** scale uses the amplitudes of S-waves, and corrects for the decrease in amplitude that happens as the waves travel away from their source. The correction depends on how seismic waves interact with the specific rock types through which they travel, and therefore on local conditions, so the Richter magnitude is also referred to as the **local magnitude**.

While news reports about earthquakes might still refer to the “Richter scale” when describing magnitudes, the number they report is most likely the **moment magnitude**. The moment magnitude is calculated from the **seismic moment** of an earthquake. The seismic moment takes into account the surface area of the region that ruptured, how much displacement occurred, and the stiffness of the rocks. Moment magnitude can capture the difference between short earthquakes and longer ones resulting from larger ruptures, even of both types of earthquakes generate the same amplitude of waves. The moment magnitude scale is also better for earthquakes that are far from the seismic station. Seismic

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wave measurements are still used to determine the moment magnitude, however different waves are used than for the local magnitude scale.

The magnitude scale is a logarithmic one rather than a linear one- an increase of one unit of magnitude corresponds to a 32 times increase in energy release (Figure 12.12). There are far more low-magnitude earthquakes than high-magnitude earthquakes. In 2017 there were 7 earthquakes of M7 (magnitude 7) or greater, but millions of tiny earthquakes.



***Figure 12.12*** *Earthquake magnitude and corresponding energy release. Energy release increases by approximately 32 times for each unit change in magnitude. Source: IRIS (n.d.) “How Often Do Earthquakes Occur?” view source*

#### Earthquake Intensity

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 now call the **Modified Mercalli Intensity Scale** (Table 12.1). To determine the intensity of an earthquake, reports are collected about what people felt and

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how much damage was done. The reports are then used to assign intensity ratings to regions where the earthquake was felt.

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**I Not felt**

**II Weak**

III Weak

IV Light

V Moderate

VI Strong

VII Very Strong

VIII

Severe

IX

Violent

X Extreme

XI

Extreme

XII Extreme**Table 12.1 Modified Mercalli Intensity Scale**

Not felt except by a very few under especially favorable conditions

Felt only by a few persons at rest, especially on upper floors of buildings

Felt quite noticeably by persons indoors, especially on upper earthquake; standing motor cars may rock slightly; vibrations similar to the passing of a truck; duration estimated

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

Felt by nearly everyone; many awakened; some dishes, windows broken; unstable objects overturned; pendulum clocks may stop

Felt by all, many frightened; some heavy furniture moved; a few instances of fallen plaster; damage slight

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

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

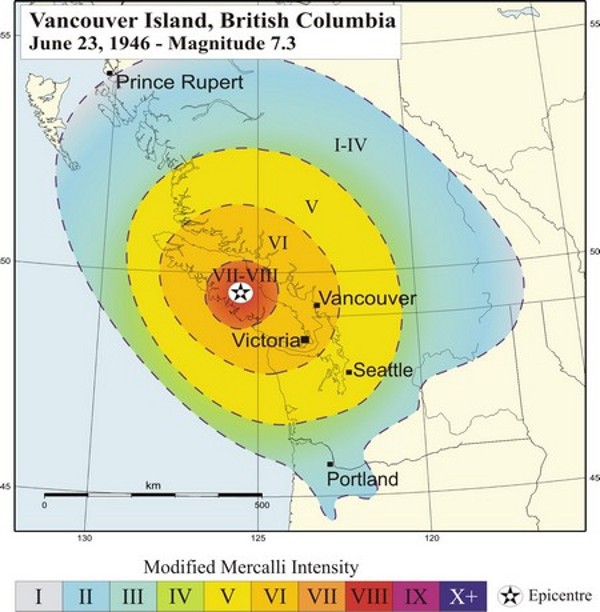
Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb; damage great in substantial buildings, with partial collapse; buildings shifted off foundations

Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; rails bent

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

Damage total; waves seen on ground surfaces; lines of sight and level distorted; objects thrown upward into the air

Source: U. S. Geological Survey (1989). *The Severity of an Earthquake*. USGS General Interest Publication 1989-288-913 view source

Intensity values are assigned to locations, rather than to the earthquake itself. This means that intensity can vary for a given earthquake, depending on the proximity to the epicenter and local conditions. For the 1946 M7.3 Vancouver Island earthquake, intensity was greatest in the central island region (Figure 12.13). In some communities within this region, 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, but with less intensity.

***Figure 12.13*** *Intensity map for the M7.3 Vancouver Island earthquake on June 23, 1946. Source: Earthquakes Canada, Natural Resources Canada (2016) view source. Click the image for terms of use.*

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Intensity estimates are important as a way to identify regions that are especially prone to strong shaking. A key factor is the nature of the underlying geological materials. The weaker the underlying materials, the more likely it is that there will be strong shaking. Areas underlain by strong solid bedrock tend to experience far less shaking than those underlain by unconsolidated river or lake sediments.

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 near the epicenter, but 350 km away in heavily populated Mexico City there was tremendous damage and approximately 5,000 deaths. The reason 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. Consequently, the shaking was amplified. Survivors of the disaster recounted that the ground in some areas moved up and down by approximately 20 cm every two seconds for over two minutes. Damage was greatest to buildings between 5 and 15 stories tall, because they also resonated at around two seconds, which amplified the shaking.

# The Impacts of Earthquakes

Earthquakes can have direct impacts, such as structural damage to buildings from shaking, and secondary impacts, such as triggering landslides, fires, and tsunami. The types and extent of impacts will depend on local conditions where the earthquake strikes. The geological materials in the area matter, as does the type of terrain, and whether the region is near the coast or not. The extent of impact and type of damage will depend on whether the area is predominantly urban or rural, densely or sparsely populated, highly developed or underdeveloped. It will depend on whether the infrastructure has been designed to withstand shaking.

### Damage to Structures from Shaking

As with the example of the 1985 Mexico earthquake, the geological foundations on which structures are built will affect the amount of shaking that occurs. Earthquakes produce seismic waves that vibrate at different rates, or **frequencies**. Waves with rapid vibrations have a high frequency, and waves with slower vibrations have lower frequencies.

The energy of the higher frequency waves tends to be absorbed by solid rock. Lower frequency waves pass through solid rock without being absorbed, but are 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. During the 1989 Loma Prieta earthquake, parts of a two-layer highway in the Oakland area near San Francisco collapsed where they were built on soft sediments (Figure 12.14).

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***Figure 12.14*** *A collapsed section of the Cypress Freeway in Oakland California. Source: U. S. Geological Survey (1989) Public Domain view source*

Building damage is also greatest in areas of soft sediments, and multi-story 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, but even though Turkey had a relatively strong building code in the 1990s, adherence to the code was poor. Builders did whatever they could to save costs, including using inappropriate materials in concrete, and reducing the amount of steel reinforcing. The result was more than 17,000 deaths in the 1999 M7.6 Izmit earthquake (Figure 12.15). 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.

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***Figure 12.15*** *Damage from the 1999 M7.6 Izmit, Turkey earthquake. Source: Left; USGS (1999) Public Domain view source; Right: USGS (1999) Public Domain view source*

Structures underlain by sediments may be at risk of another hazard, called **liquefaction**, in which sediment is transformed into a fluid. When water-saturated sediments are shaken, the grains may lose contact with each other, and no longer support one another. Water between the grains holds them apart, causing the sediment to turn to mud and flow. The loss of support can lead to the collapse of buildings or other structures that might otherwise have sustained little damage. During the 1964 M7.6 earthquake in Niigata, Japan, liquefaction caused buildings to sink into the sediments (Figure 12.16).

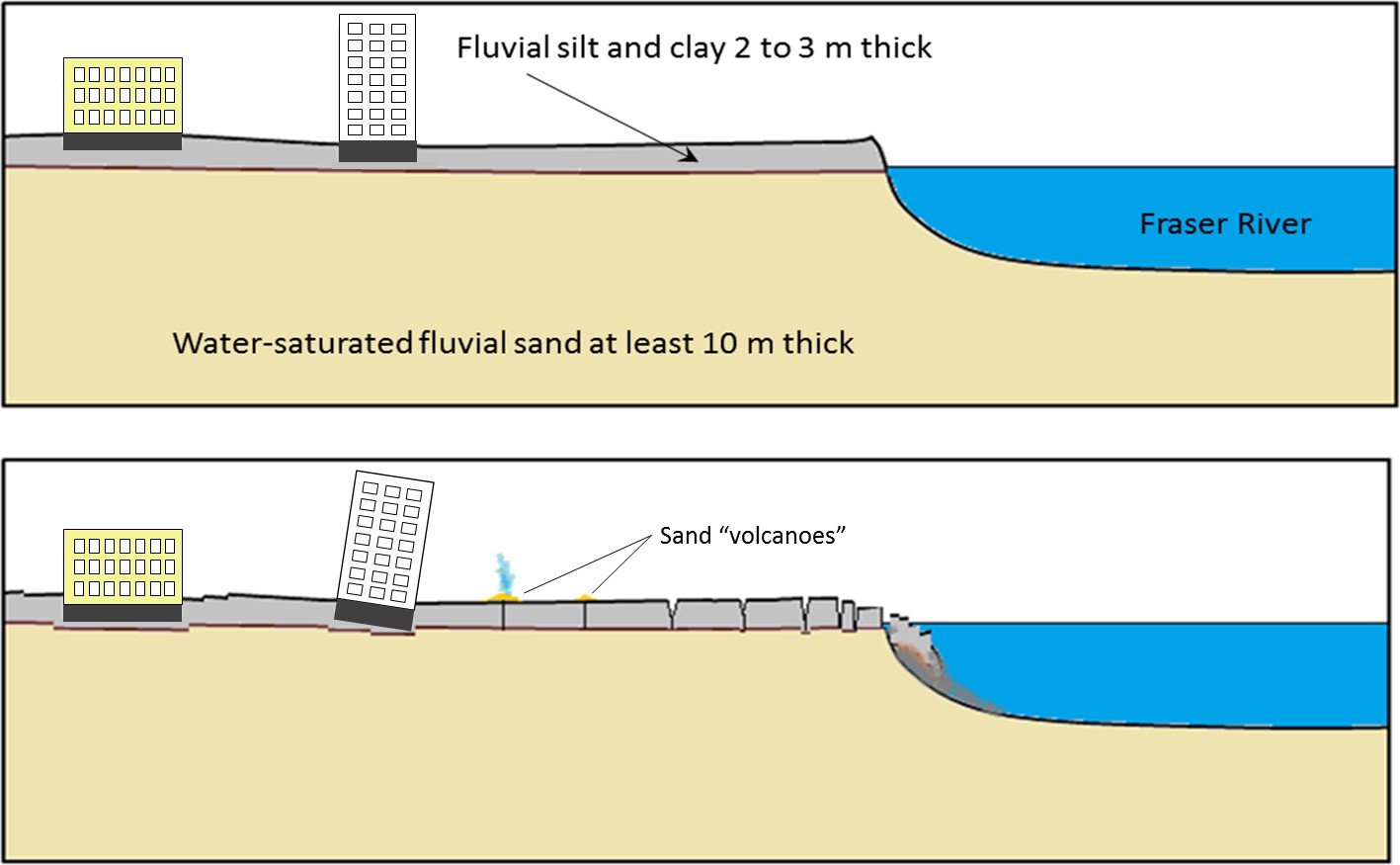
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***Figure 12.16*** *Collapsed apartment buildings in the Niigata area of Japan. The material beneath the buildings was liquefied by the 1964 earthquake. Source: DOC/NOAA/NESDIS/NCEI (1964) Public Domain view source*

Parts of the Fraser River delta are also prone to liquefaction-related damage. The region is characterized by a 2 m to 3 m thick layer of fluvial silt and clay above a layer of water-saturated fluvial sand that is at least 10 m thick (Figure 12.17). Under these conditions, seismic shaking can be amplified, and the sandy sediments will liquefy. This could lead to subsidence and tilting of buildings. 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. Current building-code regulations in the Fraser delta area require that measures be taken to strengthen the ground beneath multi-story buildings prior to construction.

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***Figure 12.17*** *Recent unconsolidated sedimentary layers in the Fraser River delta area (top) and the potential consequences in the event of a damaging earthquake. Source: Steven Earle (2015) CC BY 4.0 view source*

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### Slope Failure

Ground shaking during an earthquake can be enough to weaken rock and loose materials to the point of failure. Earthquakes can also trigger failures on slopes that are already weak. In January of 2011 a M7.6 earthquake offshore of El Salvador triggered slope failures that killed nearly 600 people (Figure 12.18).

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***Figure 12.18*** *A slope gives way in a suburb of San Salvador after the January 2001 earthquake offshore of El Salvador. This is one of hundreds of slope failures resulting from the earthquake. Source: U. S. Geological Survey (2001) Public Domain view source*

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### Fires

Fires are commonly associated with earthquakes because gas lines rupture and electrical lines are damaged when the ground shakes. Most of the damage in the great 1906 San Francisco earthquake was caused by massive fires in the downtown area of the city (Figure 12.19). Some 25,000 buildings were destroyed by those fires, which were fueled by gas leaking from broken pipes. Fighting the fires was difficult because water mains had also ruptured. Today the risk of fires can be reduced through P-wave early warning systems if utility operators can decrease pipeline pressure and break electrical circuits.

***Figure 12.19*** *Fires in San Francisco following the 1906*

*earthquake. Source: Pillsbury Picture Co. (1906) Public Domain, courtesy of the Library of Congress Prints and Photographs Division view source*

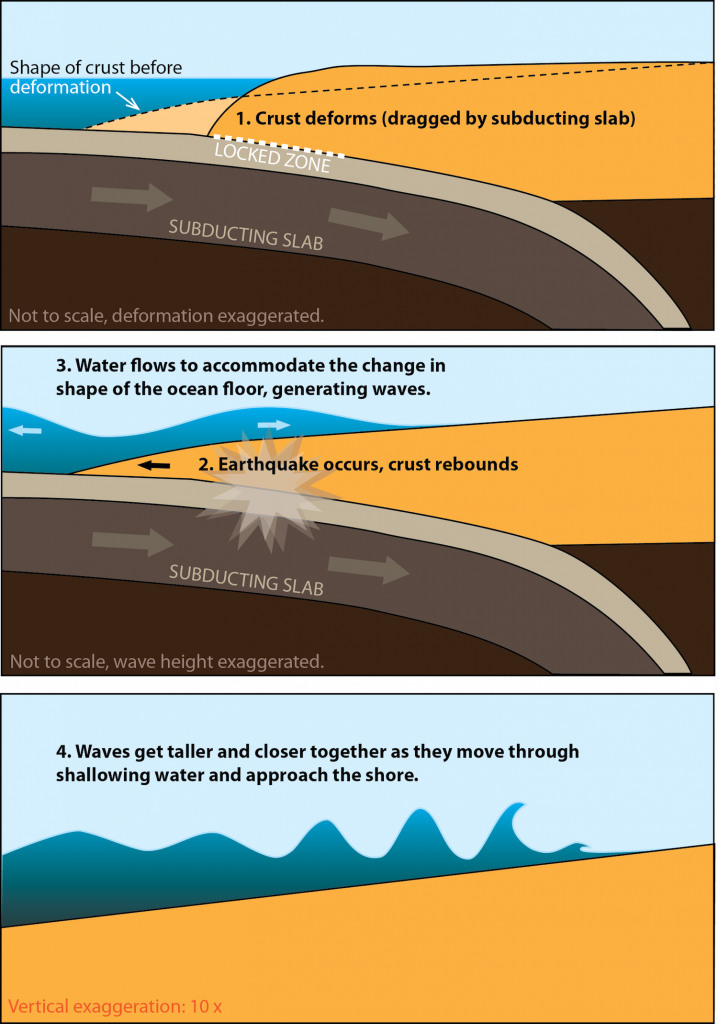
### Tsunami

Large earthquakes that take place beneath the ocean have the otential to displace large volumes of water. In a subduction zone,

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for example, the overriding plate becomes distorted by elastic deformation. It is squeezed laterally and pushed up (Figure 12.20 top). When an earthquake happens, the plate rebounds over an area of thousands of square kilometers, generating waves- a **tsunami** (Figure 12.20, middle). The waves spread across the ocean at velocities of several hundred kilometers per hour. Tsunami can make it to the far side of an ocean in about the same time as a passenger jet.

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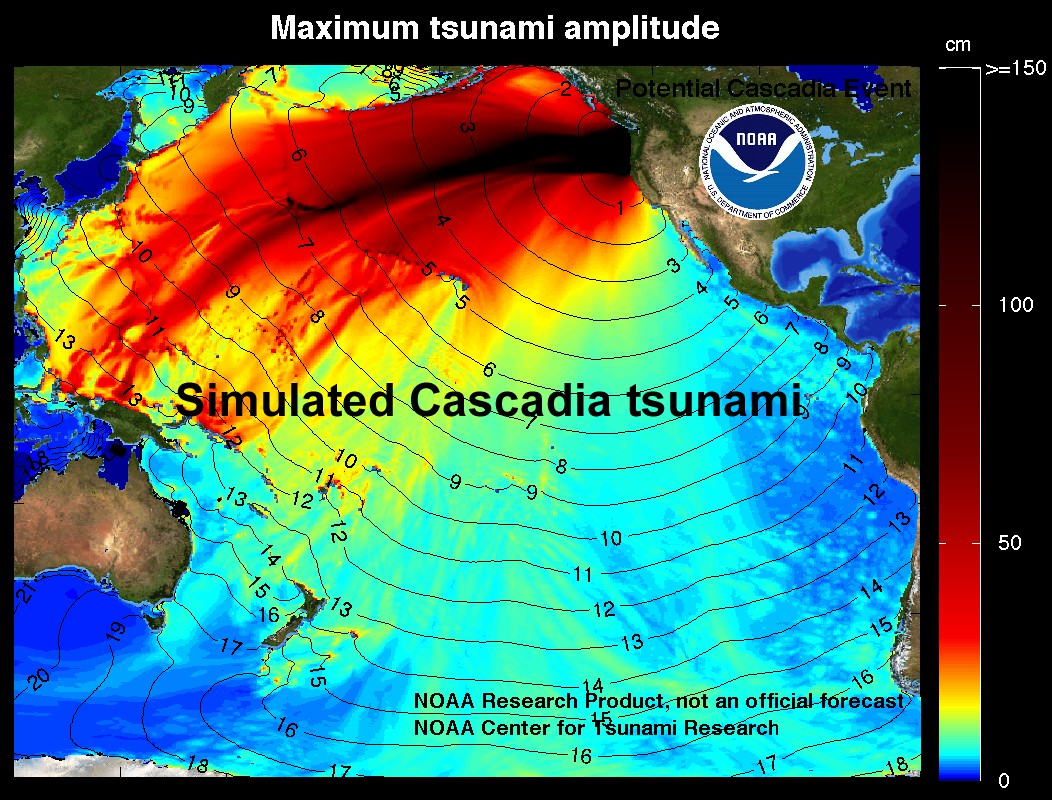
***Figure 12.20*** *Tsunami triggered along a subduction zone. Top: The overriding crust is deformed because it is locked to the subducting slab. Middle: When the locked zone ruptures, the crust rebounds, and waves are created. Bottom: Tsunami waves have small amplitudes in the deep ocean, but once in shallow water, they slow down, causing the waves to become taller and closer together. Source: Karla Panchuk (2018) CC BY-NC-SA 4.0. Top and middle modified after Steven Earle (2015) CC BY 4.0 view source. Bottom modified after COMET/UCAR (1997-2017) view source. Click the image for COMET/UCAR attribution and terms of use. 396*

Tsunami waves gain their height as they travel through shallower waters. In the deep ocean, the waves may be so small as to go undetected by ships, but when they are slowed down by interacting with the ocean floor, they can become much taller (Figure 12.20, bottom). In the tsunami following the 2004 Sumatra earthquake, the tallest waves were more than 30 m high.

Subduction earthquakes must be large to generate significant tsunami. 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 can have an impact across an entire ocean basin. They spread across the ocean at velocities of several hundred kilometers per hour, and can make it to the far side of an ocean in about the same time as a passenger jet. In 1700 a rupture along the Cascadia thrust fault running from Vancouver Island to northern California resulted in a M9 earthquake. It generated a tsunami that traveled across the Pacific Ocean, and was recorded in Japan nine hours later. A computer simulation of the tsunami (Figure 12.21) shows how long it took tsunami waves from the Cascadia earthquake to travel across the Pacific Ocean, and how high the waves were. Notice that over all, the waves decrease in height moving away from the rupture, but they increase in height again as they reach the opposite side of the Pacific Ocean.

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***Figure 12.21*** *Computer simulation of the tsunami from the 1700 M9 Cascadia earthquake. Colours show open-ocean wave heights. Contours show travel time in hours. Tsunami wave heights increase as the tsunami reached the western margin of the Pacific ocean. Source: NOAA/PMEL/Center for Tsunami Research (2011) Public Domain view source / view context*

References

Earthquakes Canada, Natural Resources Canada (2016). The M9 Cascadia Megathrust Earthquake of January 26, 1700. Visit website

University Corporation for Atmospheric Research (2010). Propagation. *Tsunamis.* Visit website

U. S. Geological Survey (2014). Indian Ocean Tsunami Remembered — Scientists reflect on the 2004 Indian Ocean that killed thousands. Visit website

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# Summary

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

##### What Is an Earthquake?

An earthquake is the shaking that results when a deformed body of rock snaps back to its original shape. The rupture is initiated at a point but quickly spreads across the area of a fault, with 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.

##### Measuring Earthquakes

Earthquakes produce seismic waves that can be measured by a seismograph. The amplitudes of seismic waves are used to determine the amount of energy released by an earthquake- its magnitude. For the moment magnitude scale used today, the amount of energy released by an earthquake is proportional to the size of the rupture surface, the amount of displacement, and the strength of the rock. Intensity is a measure of the amount of shaking that occurs, and damage done at locations that experience an earthquake. Intensity will vary depending on the distance to the epicenter, the depth of the earthquake, and the type of geological materials present.

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##### Earthquakes and Plate Tectonics

Most earthquakes happen at or near plate boundaries. Along divergent and transform boundaries earthquakes are shallow (less than 30 km depth), but at convergent boundaries they can be hundreds of kilometers beneath the surface. The largest earthquakes happen when a broad segment of the locked zone of a subduction zone ruptures. Intraplate earthquakes happen away from plate boundaries. They can be caused by human activities, or renewed motion on ancient faults.

##### 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.

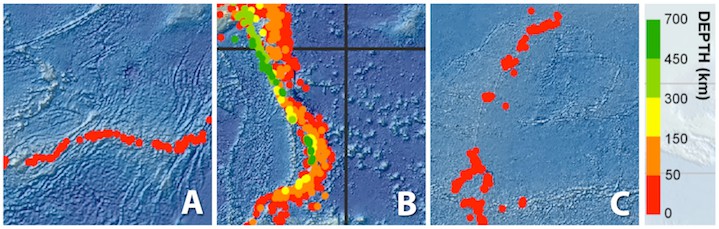
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#### Review Questions

* + 1. What causes the shaking during an earthquake?
    2. What is a rupture surface, and how does the area of a rupture surface relate to earthquake magnitude?
    3. What is an aftershock and how are aftershocks related to stress transfer?
    4. Episodic slip on the middle part of the Cascadia subduction zone is thought to increase the stress on the locked zone. Why?
    5. What is the difference between the magnitude of an earthquake and its intensity?
    6. How much more energy is released by a magnitude 7 earthquake compared to a magnitude 5 earthquake?
    7. The images below show earthquake locations for three regions of ocean lithosphere. The colour scheme indicates the depths of earthquakes. a) The images show a subduction zone, a slowly spreading mid-ocean ridge, and a rapidly spreading

mid-ocean ridge. Which is which? b) In the image with the subduction zone, which side is the subducting plate, and which is the overriding plate?

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***Figure 12.22*** *Earthquake s along plate margins.*

*Dots are color coded according to depth.*

*Source: Map details from Lisa Christianse n, Caltech Tectonics Observatory (2008) view source.*

*Click the image for terms of use.*

* + 1. Why is earthquake damage likely to be more severe for buildings built on unconsolidated sediments as opposed to on solid rock?
    2. Why are fires common during earthquakes?
    3. What type of earthquake is likely to lead to a tsunami?

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