2-5. Strong Ground Motions

Shaking of ground on the Earth's surface is a net consequence of motions caused by seismic waves generated by energy release at each material point within the three-dimensional volume that ruptures at the fault. These waves arrive at various instants of time, have different amplitudes and carry different levels of energy. Thus, the motion at any site on ground is random in nature with its amplitude and direction varying randomly with time.

Large earthquakes at great distances can produce weak motions that may not damage structures or even be felt by humans. But, sensitive instruments can record these. This makes it possible to locate distant earthquakes. However, from engineering viewpoint, strong motions that can possibly damage structures are of interest. This can happen with earthquakes in the vicinity or even with large earthquakes at reasonable medium to large distances.

2-6. Earthquake Fault Sources

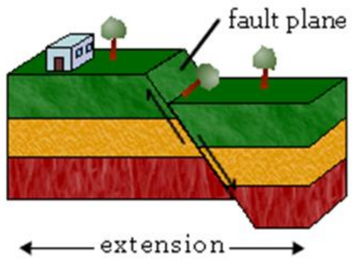
A fault is a fracture or zone of fractures between two blocks of rock. Faults allow the blocks to move relative to each other. This movement may occur rapidly, in the form of an earthquake - or may occur slowly, in the form of creep. Faults may range in length from a few millimeters to thousands of kilometers. Most faults produce repeated displacements over geologic time.

During an earthquake, the rock on one side of the fault suddenly slips with respect to the other. The fault surface can be horizontal or vertical or some arbitrary angle in between. Earth scientists use the angle of the fault with respect to the surface (known as the dip) and the direction of slip along the fault to classify faults. Faults which move along the direction of the dip plane are dip-slip faults and described as either normal or reverse, depending on their motion. Faults which move horizontally are known as strike-slip faults and are classified as either right-lateral or left-lateral. Faults which show both dip-slip and strike-slip motion are known as oblique-slip faults.

1] **DIP-SLIP FAULTS**

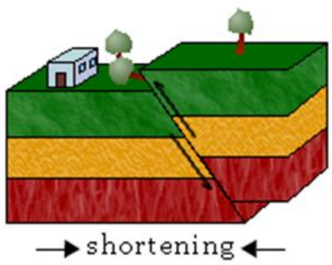
a) Normal Fault

In a normal fault, the block above the fault moves down relative to the block below the fault. This fault motion is caused by tensional forces and results in extension. [Other names: normal-slip fault, tensional fault or gravity fault]



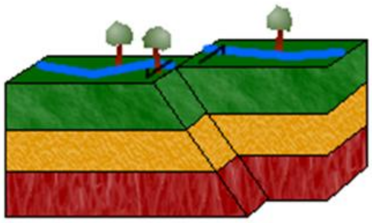
b) Reverse Fault

In a reverse fault, the block above the fault moves up relative to the block below the fault. This fault motion is caused by compressional forces and results in shortening. A reverse fault is called a thrust fault if the dip of the fault plane is small. [Other names: thrust fault, reverse-slip fault or compressional fault]



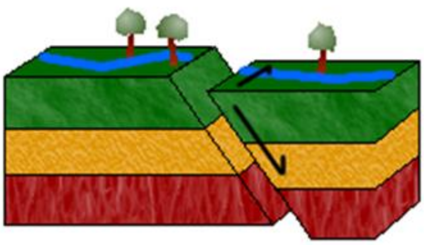
2] **STRIKE-SLIP FAULT**

In a strike-slip fault, the movement of blocks along a fault is horizontal. If the block on the far side of the fault moves to the left, the fault is called left-lateral. If the block on the far side moves to the right, the fault is called right-lateral. The fault motion of a strike-slip fault is caused by shearing forces. [Other names: Tran's current fault, lateral fault, tear fault or wrench fault]



3] **OBLIQUE-SLIP FAULT**

Oblique-slip faulting suggests both dip-slip faulting and strike-slip faulting. It is caused by a combination of shearing and tension of compressional forces.



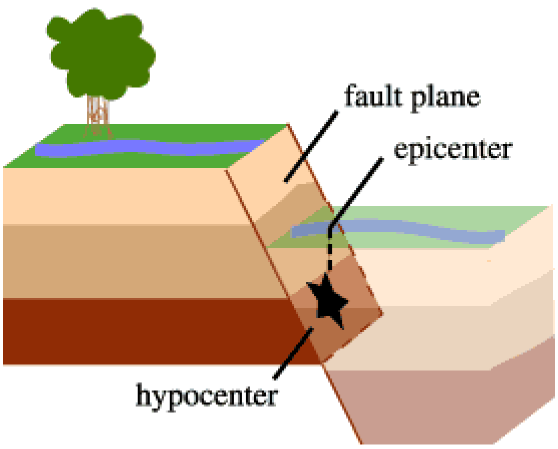
2-7. How the Ground Shake?

The term earthquake describes both the sudden slip on a fault and the radiated seismic energy and ground shaking caused by the slip. It also covers ground shaking caused by volcanic or magmatic activity and other sudden movement due to stress changes in the earth.

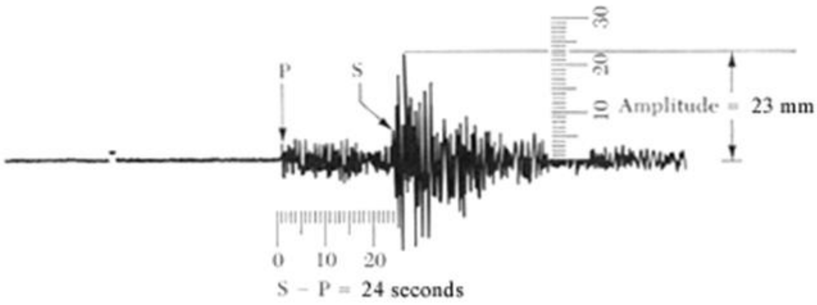
2-8. What are Aftershocks and Foreshocks?

Aftershocks and foreshocks are related terms describing earthquakes that happen before or after a "main shock", that is to say, an earthquake. Foreshocks are earthquakes that occur in the same location as a main shock, but before the earthquake happens, whereas aftershocks are smaller earthquakes that occur after the main shock in the same area, but not necessarily the same exact location. Aftershocks generally decrease with time after the main event and are much less common in deep earthquakes.

2-9. How Do I Locate that Earthquake's Epicenter?



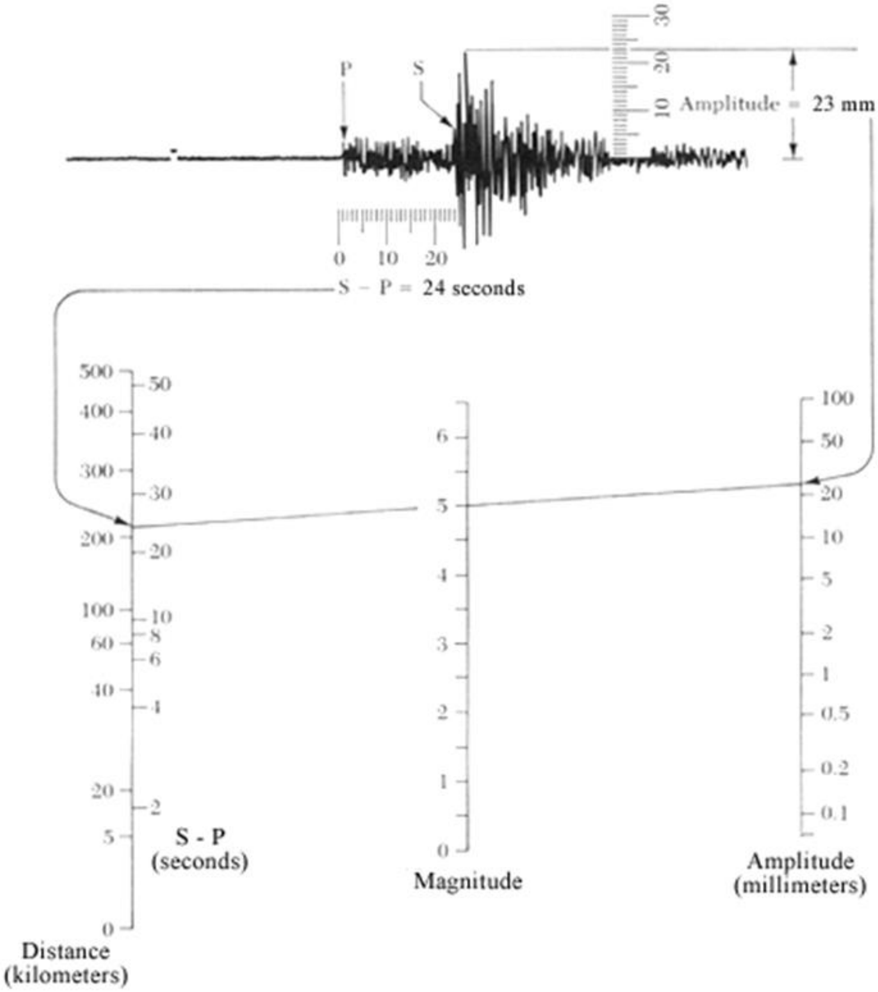
To figure out just where that earthquake happened, you need to look at your seismogram and you need to know what at least two other seismographs recorded for the same earthquake. You will also need a map of the world, a ruler, a pencil, and a compass for drawing circles on the map. Here's an example of a seismogram:



Our typical seismogram from before, this time marked for this exercise (from Bolt, 1978).

One minute intervals are marked by the small lines printed just above the squiggles made by the seismic waves (the time may be marked differently on some seismographs). The distance between the beginning of the first P wave and the first S wave tells you how many seconds the waves are apart. This number will be used to tell you how far your seismograph is from the epicenter of the earthquake.

Finding the Distance to the Epicenter and the Earthquake's Magnitude



Use the amplitude to derive the magnitude of the earthquake, and the distance from the earthquake to the station. (From Bolt, 1978)

1. Measure the distance between the first P wave and the first S wave. In this case, the first P and S waves are 24 seconds apart.
2. Find the point for 24 seconds on the left side of the chart below and mark that point. According to the chart, this earthquake's epicenter was 215 kilometers away.
3. Measure the amplitude of the strongest wave. The amplitude is the height (on paper) of the strongest wave. On this seismogram, the amplitude is 23 millimeters. Find 23 millimeters on the right side of the chart and mark that point.

4. Place a ruler (or straight edge) on the chart between the points you marked for the distance to the epicenter and the amplitude. The point where your ruler crosses the middle line on the chart marks the magnitude (strength) of the earthquake. This earthquake had a magnitude of 5.0.

2-10. How Can Scientists Tell where the Earthquake Happened?

You have just figured out how far your seismograph is from the epicenter and how strong the earthquake was, but you still don't know exactly where the earthquake occurred. This is where the compass, the map, and the other seismograph records come in.

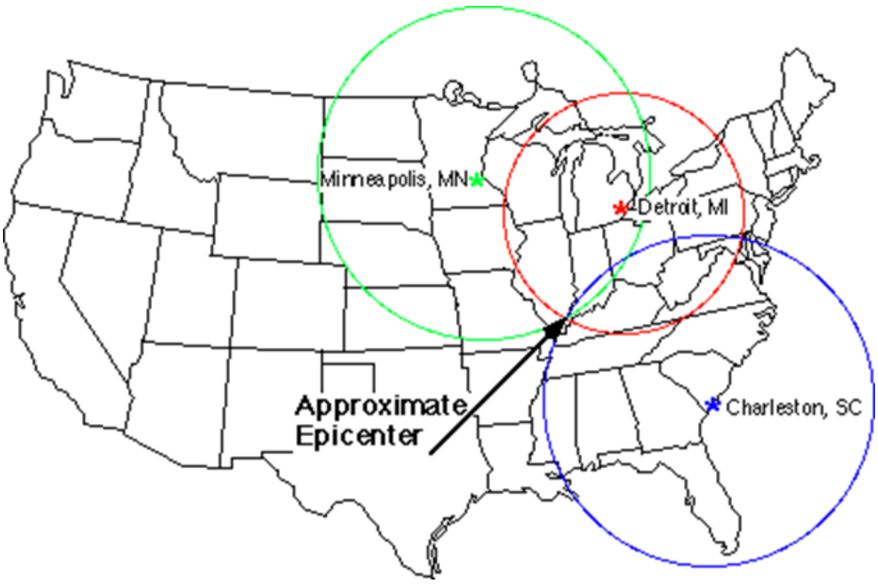


Figure 3 - The point where the three circles intersect is the epicenter of the earthquake. This technique is called 'triangulation.'

1. Check the scale on your map. It should look something like a piece of a ruler. All maps are different. On your map, one centimeter could be equal to 100 kilometers or something like that.
2. Figure out how long the distance to the epicenter (in centimeters) is on

your map. For example, say your map has a scale where one centimeter is equal to 100 kilometers. If the epicenter of the earthquake is 215 kilometers away, that equals 2.15 centimeters on the map.

3. Using your compass, draw a circle with a radius equal to the number you came up with in Step 2 (the radius is the distance from the center of a circle to its edge). The center of the circle will be the location of your seismograph. The epicenter of the earthquake is somewhere on the edge of that circle.
4. Do the same thing for the distance to the epicenter that the other seismograms recorded (with the location of those seismographs at the center of their circles). All of the circles should overlap. The point where all of the circles overlap is the approximate epicenter of the earthquake.

2-11. Seismic Waves

Large strain energy released during an earthquake travels as seismic waves in all directions through the Earth's layers, reflecting and refracting at each interface. These waves are of two types - *body waves* and *surface waves*; the latter are restricted to near the Earth's surface (Figure 1).

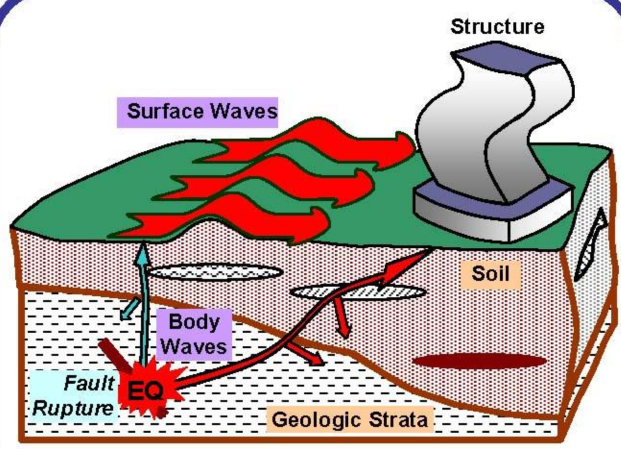


Figure 1: Arrival of Seismic Waves at a Site

Body waves consist of *Primary Waves (P-waves)* and *Secondary Waves (S--waves)*, and surface waves consist of *Love waves* and *Rayleigh waves*. Under P-waves, material particles undergo extensional and compressional strains along direction of energy transmission, but under S-waves, oscillate at right angles to it (Figure 2). Love waves cause surface motions similar to that by S-waves, but with no vertical component. Rayleigh wave makes a material particle oscillate in an elliptic path in the vertical plane (with horizontal motion along direction of energy transmission). P-waves are fastest, followed in sequence by S-, Love and Rayleigh waves.

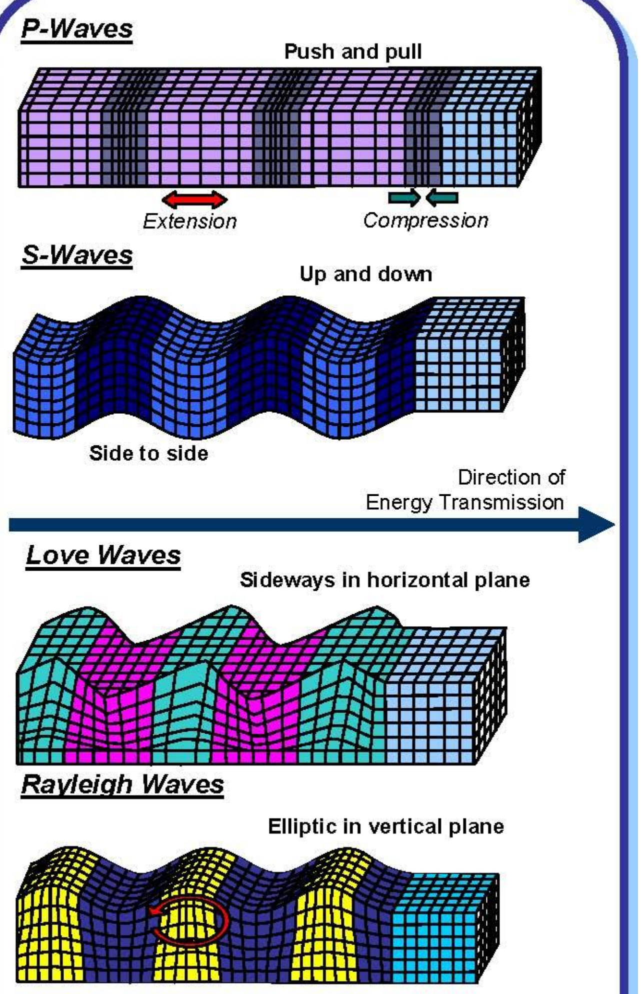


Figure 2:
Motions caused by Body and Surface Waves
(Adapted from FEMA 99, Non-Technical Explanation of the NEHRP Recommended Provisions)

For example, in granites, P- and S-waves have speeds ~4.8 km/sec and ~3.0km/sec, respectively. S-waves do not travel through liquids. S-waves in association with effects of Love waves cause maximum damage to structures by their racking motion on the surface in both vertical and horizontal directions. When P- and S-waves reach the Earth's surface, most of their energy is reflected back. Some of this energy is returned back to the surface by reflections at different layers of soil and rock. Shaking is more severe (about twice as much) at the Earth's surface than at substantial depths. This is often the basis for designing structures buried underground for smaller levels of acceleration than those above the ground.

2-12. How Are Earthquake Magnitudes Measured?

The magnitude of most earthquakes is measured on the Richter scale, invented by Charles F. Richter in 1934. The Richter magnitude is calculated from the amplitude of the largest seismic wave recorded for the earthquake, no matter what type of wave was the strongest.

The Richter magnitudes are based on a logarithmic scale (base 10). What this means is that for each whole number you go up on the Richter scale, the amplitude of the ground motion recorded by a seismograph goes up ten times. Using this scale, a magnitude 5 earthquake would result in ten times the level of ground shaking as a magnitude 4 earthquake (and 32 times as much energy would be released). To give you an idea how these numbers can add up, think of it in terms of the energy released by explosives: a magnitude 1 seismic wave releases as much energy as blowing up 6 ounces of TNT. A magnitude 8 earthquake releases as much energy as detonating 6 million tons of TNT. Pretty impressive, huh? Fortunately, most of the earthquakes that occur each year are magnitude 2.5 or less, too small to be felt by most people.

The Richter magnitude scale can be used to describe earthquakes so small that they are expressed in negative numbers. The scale also has no upper limit, so it can describe earthquakes of unimaginable and (so far) unexperienced intensity, such as magnitude 10.0 and beyond.

Modified Mercalli Intensity Scale

Mercalli Intensity	Equivalent Richter Magnitude	Witness Observations
I	1.0 to 2.0	Felt by very few people; barely noticeable.
II	2.0 to 3.0	Felt by a few people, especially on upper floors.
III	3.0 to 4.0 **Minor**	Noticeable indoors, especially on upper floors, but may not be recognized as an earthquake.
IV	4.0	Felt by many indoors, few outdoors. May feel like heavy truck passing by.
V	4.0 to 5.0 **Light**	Felt by almost everyone, some people awakened. Small objects moved trees and poles may shake.
VI	5.0 to 6.0 **Moderate**	Felt by everyone. Difficult to stand. Some heavy furniture moved, some plaster falls. Chimneys may be slightly damaged
VII	6.0	Slight to moderate damage in well built, ordinary structures. Considerable damage to poorly built structures. Some walls may fall.
VIII	6.0 to 7.0 **Strong**	Little damage in specially built structures. Considerable damage to ordinary buildings, severe damage to poorly built structures. Some walls collapse.
IX	7.0	Considerable damage to specially built structures, buildings shifted off foundations. Ground cracked noticeably. Wholesale destruction. Landslides.
X	7.0 to 8.0 **Major**	Most masonry and frame structures and their foundations destroyed. Ground badly cracked. Landslides. Wholesale destruction.
XI	8.0	Total damage. Few, if any, structures standing. Bridges destroyed. Wide cracks in ground. Waves seen on ground.
XII	8.0 or greater **Great**	Total damage. Waves seen on ground. Objects thrown up into air.

2-13. Measuring Instruments

The instrument that measures earthquake shaking, a *seismograph*, has three components – the *sensor*, the *recorder* and the *timer*. The principle on which it works is simple and is explicitly reflected in the early seismograph (Figure 3) – a pen attached at the tip of an oscillating simple pendulum (a mass hung by a string from a support) marks on a chart paper that is held on a drum rotating at a constant speed. A magnet around the string provides required damping to control the amplitude of oscillations. The pendulum mass, string, magnet and support together constitute the *sensor*; the drum, pen and chart paper constitute the *recorder*; and the motor that rotates the drum at constant speed forms the *timer*.

One such instrument is required in each of the two orthogonal horizontal directions. Of course, for measuring vertical oscillations, the *string* pendulum (Figure 3) is replaced with a *spring* pendulum oscillating about a fulcrum. Some instruments do not have a timer device (*i.e.,* the drum holding the chart paper does not rotate). Such instruments provide only the maximum extent (or scope) of motion during the earthquake; for this reason they are called *seismoscopes*.

The analog instruments have evolved over time, but today, *digital instruments* using modern computer technology are more commonly used.

The digital instrument records the ground motion on the memory of the microprocessor that is in-built in the instrument.

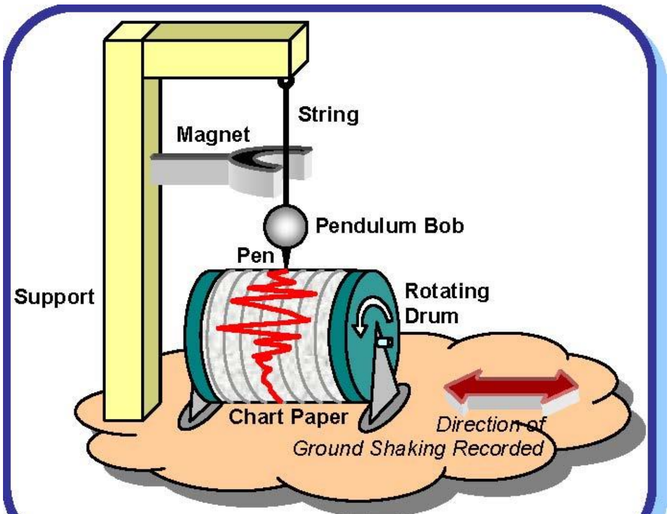


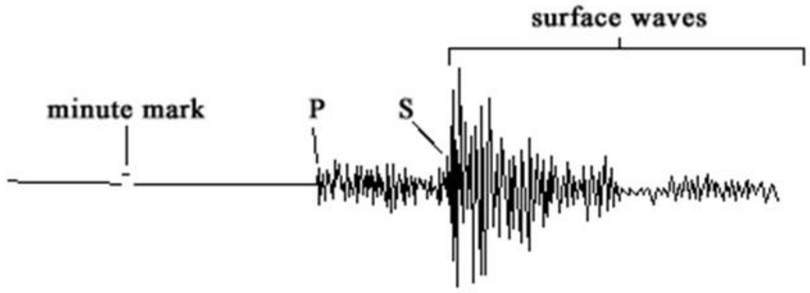
Figure 3: Schematic of *Early Seismograph*

2-14. How do Scientists Measure the Size of Earthquakes?

The size of an earthquake depends on the size of the fault and the amount of slip on the fault, but that’s not something scientists can simply measure with a measuring tape since faults are many kilometers deep beneath the earth’s surface. So how do they measure an earthquake? They use the seismogram recordings made on the seismographs at the surface of the earth to determine how large the earthquake was. A short wiggly line that doesn’t wiggle very much means a small earthquake, and a long wiggly line that wiggles a lot means a large earthquake. The length of the wiggle depends on the size of the fault, and the size of the wiggle depends on the amount of slip.

2-15. How Do I Read a Seismogram?

When you look at a seismogram, there will be wiggly lines all across it. These are all the seismic waves that the seismograph has recorded. Most of these waves were so small that nobody felt them. These tiny microseisms can be caused by heavy traffic near the seismograph, waves hitting a beach, the wind, and any number of other ordinary things that cause some shaking of the seismograph. There may also be some little dots or marks evenly spaced along the paper. These are marks for every minute that the drum of the seismograph has been turning. How far apart these minute marks are will depend on what kind of seismograph you have.



A typical seismogram.

So which wiggles are the earthquake? The P wave will be the first wiggle that is bigger than the rest of the little ones (the microseisms). Because P waves are the fastest seismic waves, they will usually be the first ones that your seismograph records. The next set of seismic waves on your seismogram will be the S waves. These are usually bigger than the P waves.

2-16. Characteristics of Strong Ground Motions

The motion of the ground can be described in terms of displacement, velocity or acceleration. The variation of ground acceleration with time recorded at a point on ground during an earthquake is called an *accelerogram*. The nature of accelerograms may vary (Figure 4) depending on energy released at source, type of slip at fault rupture, geology along the travel path from fault rupture to the Earth's surface, and local soil (Fig. 1). They carry distinct information regarding ground shaking; *peak amplitude*, *duration of strong shaking*, *frequency content* (*e.g.,* amplitude of shaking associated with each frequency) and *energy content* (*i.e.,* energy carried by ground shaking at each frequency) are often used to distinguish them.

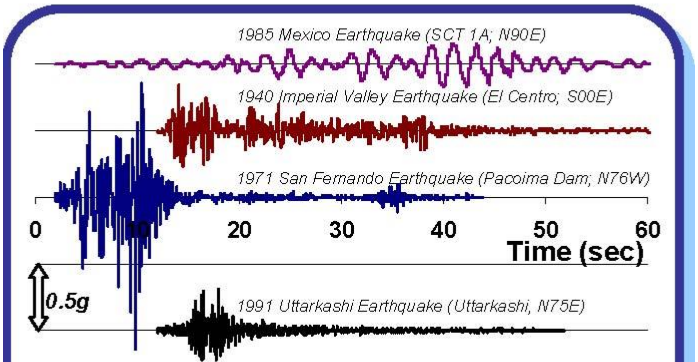


Figure 4::
Some typical recorded accelerograms

Peak amplitude (*peak ground acceleration, PGA*) is physically intuitive. For instance, a horizontal PGA value of *0.6g* (= *0.6* times the acceleration due to gravity) suggests that the movement of the ground can cause a maximum horizontal force on a rigid structure equal to *60%* of its weight. In a rigid structure, all points in it move with the ground by the same amount, and hence experience the same maximum acceleration of PGA. Horizontal PGA values greater than *1.0g* were recorded during the 1994 Northridge Earthquake in USA. Usually, strong ground motions carry significant energy associated with shaking of frequencies in the range *0.03-30Hz (i.e., cycles per sec)*.

Generally, the maximum amplitudes of horizontal motions in the two orthogonal directions are about the same. However, the maximum amplitude in the vertical direction is usually less than that in the horizontal direction. In design codes, the vertical design acceleration is taken as $1/2$ to $2/3$ horizontal design acceleration. In contrast, the maximum horizontal and vertical ground accelerations *in the vicinity* of the fault rupture do not seem to have such a correlation.

2-17. Predicting Earthquakes

The goal of earthquake prediction is to give warning of potentially damaging earthquakes early enough to allow appropriate response to the disaster, enabling people to minimize loss of life and property. The U.S. Geological Survey conducts and supports research on the likelihood of future earthquakes. This research includes field, laboratory, and theoretical investigations of earthquake mechanisms and fault zones. A primary goal of earthquake research is to increase the reliability of earthquake probability estimates. Ultimately, scientists would like to be able to specify a high probability for a specific earthquake on a particular fault within a particular year. Scientists estimate earthquake probabilities in two ways: by studying the history of large earthquakes in a specific area and the rate at which strain accumulates in the rock.

This time-exposure photograph of the electronic-laser, ground-motion movement system in operation at Parkfield, California, to track movement along the San Andreas fault.

Scientists study the past frequency of large earthquakes in order to determine the future likelihood of similar large shocks. For example, if a region has experienced four magnitude 7 or larger earthquakes during 200 years of recorded history, and if these shocks occurred randomly in time, then scientists would assign a 50 percent probability (that is, just as likely to happen as not to happen) to the occurrence of another magnitude 7 or larger quake in the region during the next 50 years.

But in many places, the assumption of random occurrence with time may not be true, because when strain is released along one part of the fault system, it may actually increase on another part. Four magnitude 6.8 or larger earthquakes and many magnitude 6 - 6.5 shocks occurred in the San Francisco Bay region during the 75 years between 1836 and 1911. For the next 68 years (until 1979), no earthquakes of magnitude 6 or larger occurred in the region. Beginning with a magnitude 6.0 shock in 1979, the earthquake activity in the region increased dramatically; between 1979 and 1989, there were four magnitude 6 or greater earthquakes, including the magnitude 7.1 Loma Prieta earthquake. This clustering of earthquakes leads scientists to estimate that the probability of a magnitude 6.8 or larger earthquake occurring during the next 30 years in the San Francisco Bay region is about 67 percent (twice as likely as not).

Another way to estimate the likelihood of future earthquakes is to study how fast strain accumulates. When plate movements build the strain in rocks to a critical level, like pulling a rubber band too tight, the rocks will suddenly break and slip to a new position. Scientists measure how much strain accumulates along a fault segment each year, how much time has passed since the last earthquake along the segment, and how much strain was released in the last earthquake. This information is then used to calculate the time required for the accumulating strain to build to the level that results in an earthquake. This simple model is complicated by the fact that such detailed information about faults is rare. In the United States, only the San Andreas fault system has adequate records for using this prediction method.

Both of these methods, and a wide array of monitoring techniques, are being tested along part of the San Andres fault. For the past 150 years, earthquakes of about magnitude 6 have occurred an average of every 22 years on the San Andreas fault near Parkfield, California. The last shock was in 1966. Because of the consistency and similarity of these earthquakes, scientists have started an experiment to "capture" the next Parkfield earthquake. A dense web of monitoring instruments was deployed in the region during the late 1980s. The main goals of the ongoing Parkfield Earthquake Prediction Experiment are to record the geophysical signals before and after the expected earthquake; to issue a short-term prediction; and to develop effective methods of communication between earthquake scientists and community officials responsible for disaster response and mitigation. This project has already made important contributions to both earth science and public policy.

Scientific understanding of earthquakes is of vital importance to the Nation. As the population increases, expanding urban development and construction works encroach upon areas susceptible to earthquakes. With a greater understanding of the causes and effects of earthquakes, we may be able to reduce damage and loss of life from this destructive phenomenon.