What

An earthquake is the **catastrophic failure** within a body of rock due to the application of forces which exceed the rock's intrinsic strength. This failure usually occurs along a surface called a **fault** (Figure 1), resulting in the relative motion of rocks on either side. One of the most famous earthquake faults in the world is the San Andreas fault of California. The rapid slip along an earthquake fault generates **elastic waves** which radiate outward in all directions. Shaking associated with these waves can result in **serious damage** to structures at the earth's surface, as well as attendant loss of life. Earthquake waves can also be used to study both the **earthquake mechanism** itself, as well as the **properties of the earth** through which they propagate.

Figure 1. (Bolt, 1999)
Elastic energy can propagate through the earth in three distinct modes (Figure 2):

**Compressional:** particle motion back and forth in the direction of wave propagation.

**Shear:** particle motion perpendicular to the direction of wave propagation.

**Surface:** particle motion restricted to the zone near the surface.

Since compressional waves travel fastest, they are usually the first waves to be detected and are thus also called primary, or P waves.

Shear waves are next fastest, and are therefore called secondary, or S waves.

Surface waves are the slowest of all. There are two flavors of surface wave: Love waves, in which the particle motion is horizontal and perpendicular to the direction of propagation, and Rayleigh waves, in which the ground motion describes a retrograde ellipse in the vertical plane perpendicular to the direction of motion. Whereas P and S waves are usually impulsive in nature, surface waves are dispersive, that is, the energy tends to "spread out" as it travels, resulting in motion that tends to "ring" for a considerable period of time.

**Seismometers** (Figure 3) are used to detect ground motions generated by earthquakes, even those that are located at great distances. Analysis of seismograph recordings (Figure 4) of such ground motions can yield estimates of earthquake location, size and the nature of the faulting involved.
Figure 2. (Bolt, 1999)
Where

Locating earthquakes with seismograms takes advantage of the fact that the different earthquake waves travel at different speeds. If we were to plot the recordings from a single earthquake as a function of distance between the earthquake and seismic station (Figure 4), we would obviously see that the vibrations due to P and S waves arrive at greater and greater travel times as the distance increases. Analysis of such travel times can be used to measure the speed of seismic waves as a function of depth in the earth. Such analysis is our primary source of information on the internal structure of the earth (Figures 5 and 6).

Of particular utility however, is the fact that the difference in the arrival time between the P and S waves also increases with increasing distance from the earthquake (Figure 4). By measuring the S-P travel time and comparing it to standard travel time curves (Figure 7) or tables (Figure 8), we can estimate the distance to a given earthquake. Note that distance is usually measured in terms of
degrees of arc along the earth's surface when looking at earthquakes on the global scale. 1 degree of arc is approximately 110 km.

However, identifying the P and the S wave is complicated by the fact that seismic waves can follow different raypaths through the earth (Figure 5). For example, some P and S wave energy that is emitted by an earthquake may reflect off major interfaces in the earth, such as the core-mantle boundary. In fact, it is the recognition of such waves that is our primary evidence that such boundaries exist. The expected travel time behavior of such seismic phases is also shown on the travel time curve (Figure 7). Note that there is a standard nomenclature for labeling the different possible raypaths in the earth (e.g. PcP for a P wave that travels downward, reflects from the core and then returns upward as a P wave).
6. a) First order structure of the earth’s interior, as deduced from seismology (Skinner and Porter, 1995). b) Modern studies using tomographic methods have revealed that this basic spherical symmetry is perturbed by considerably second order complexity. Here is a visualization of velocity anomalies in the lower mantle.
Figure 7. Global Travel Time Curve
## JEFFREYS-BULLEN TABLES

**S - P Arrival Times as a Function of Delta.**

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Figure 8. (from Simon, 1972).
Exercise I: Identify the P and S waves for the earthquake (EQK I) seismograms recorded here at INY. Measure the S-P travel time difference. Use either the travel time curve or travel time table to determine the distance from INY to the earthquake.

\[ S-P \text{ (minutes)} = \underline{} \]

\[ Distance \text{ (degrees)} = \underline{} \]

Now we know that the earthquake **epicenter** (or position at the surface above the earthquake) must lie somewhere on a circle drawn around INY at the computed distance (Figure 9).

The next step in location is to determine the **direction** to the earthquake. If we had data from multiple seismic stations we could triangulate the location of the earthquake by drawing circles around each station corresponding to the distances computed from S-P for each station. These circles should intersect at the earthquake's epicentral location.

Sometimes, however, we can get rough direction from the information recorded at a single station. We do this by taking advantage of the earthquake's **first motions** as recorded on the different directional components of the seismograph.

For example, consider a P wave incident upon a 3 component seismic station from below. Because of its nature, the motion of an upcoming P wave can either be a **push** or a **pull**. If it is a **push**, the first deflection of seismograms should be **upward** and **away from the earthquake**. On the vertical component this will appear as an **up** motion. On the horizontal components, the first motion will then indicate the directions that are **opposite** to the direction the earthquake waves are arriving. For example, if the earthquake is **from the northwest**, it would result in **southward** and **eastward** first deflections. The relative amplitudes of the P wave on the two horizontal components gives some idea of the primary direction to the earthquake. For example, if in this case the amplitude of the P wave on the NS component is larger than that on the EW component, the earthquake would be from the NNW.
If the P wave happens to arrive with a pull motion, it will appear on the vertical component as a downward first deflection. In this case, the first deflections of the two horizontal components would then indicated the direction toward the earthquake.

Identify the first motions on the 3 components of the earthquake seismogram:

Vertical: _______________

North-South: _______________

East-West: _______________

Based on these first motions, and the relative amplitudes of the P wave on the two horizontal components, what is the approximate direction to the earthquake: _______________

Plot your epicentral location on the Ithaca-centric map (Figure 9).

What is the geographic location of your earthquake? ___________

____________________________

When

Once you have determined the distance to the earthquake using the S-P travel times, you can determine the absolute time of earthquake occurrence. This is simply the measured P arrival time (absolute clock time on the seismogram) minus the P wave travel time as determined from the travel time curve, i.e.

P arrival time (clock time) from seismogram: ____________ GMT

P travel time from TT table/curve: - ____________ GMT

Time of earthquake: = ____________ GMT
Once we know where an earthquake has occurred, we would like to know how "large" the earthquake is. Earthquake size is usually ranked on a logarithmic scale which assigns each event a magnitude (Figure 10). There are several types of magnitude scale in use; the best known is the Richter magnitude scale. This is an empirical ranking based on the measured amplitude of vibration of a standard instrument at a standard (reference) distance. Scientists nowadays tend to use a more physical measure, the seismic moment or moment-magnitude, to evaluate the energy released during an earthquake. The seismic moment is defined as:
\[ M_o = \mu A S \]

where \( S \) is the slip on a fault of area \( A \), and \( \mu \) is the shear modulus of the rocks involved. The moment magnitude is a number that is derived from the moment but is more comparable to the Richter magnitude.

Because the magnitude scale is logarithmic, a unit change in magnitude corresponds to a factor of 10 change in ground motion, or a factor of 30 in the amount of energy released.

Another measure of an earthquake's "size" is its intensity. Intensity is based on the observed damage caused by an earthquake. One example of an intensity scale is the Modified Mercalli Scale. Unlike magnitude, intensive varies as a function of both the distance from an earthquake and the geology...
underlying the area of measurement (e.g. Figure 11). It is thus a measure of an earthquakes impact on a given area, rather than an estimate of the intrinsic energy released by the earthquake.

The following is an abbreviated description of the 12 levels of Modified Mercalli intensity.

I. Not felt except by a very few under especially favorable conditions.
II. Felt only by a few persons at rest, especially on upper floors of buildings.
III. 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.
IV. 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.
V. Felt by nearly everyone; many awakened. Some dishes, windows broken. Unstable objects overturned. Pendulum clocks may stop.
VI. Felt by all, many frightened. Some heavy furniture moved; a few instances of fallen plaster. Damage slight.
VII. 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.
VIII. 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.
IX. 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.
X. Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations. Rails bent.
XI. Few, if any (masonry) structures remain standing. Bridges destroyed. Rails bent greatly.
XII. Damage total. Lines of sight and level are distorted. Objects thrown into the air.


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Exercise II: Compute the Richter magnitude of the earthquake on the INY seismograms.

Earthquake magnitudes are usually based on the amplitudes of either the P wave or the S wave. Here we will use the P wave and formula (Bolt, 1999):

\[ m_b = \log_{10} \left( \frac{A}{T} \right) + 0.01D + 5.9 \]

First measure the amplitude (zero to peak) of the P wave on the vertical component seismogram.

\[ P \text{ Amplitude} = \quad \text{mm} \]

Convert this amplitude to microns (1 mm = 1000 microns)

\[ P \text{ Amplitude} = \quad \text{microns} \]

Now we must convert this seismograph amplitude to ground motion amplitude. To do this, divide the observed deflection (in microns) by the Magnification of the instrument.

\[ \text{Magnification} = \quad 2500 \]

\[ \text{Ground motion amplitude} \ (Ag) = \quad \text{microns} \]
Now to compensate for variations in the frequency of the P wave, we must divide this ground motion by the period (T) of the P wave in seconds. The period of a wave is simply the time between successive peaks of the wave, or twice the time between successive zero crossings. Measure the period of the P wave from your seismogram:

\[
\text{Period (T)} = \underline{\hspace{2cm}} \text{ sec}
\]

\[
\frac{A_g}{T} = \underline{\hspace{2cm}}
\]

Now we need to put this on a logarithmic scale. Take the base 10 logarithm of \( \frac{A_g}{T} \):

\[
\log(\frac{A_g}{T}) = \underline{\hspace{2cm}}
\]

Now we must correct for distance from the earthquake.

\[
\Delta = \underline{\hspace{2cm}}
\]

Distance correction = \( 0.01 \Delta \) = \( \underline{\hspace{2cm}} \)

Now put it all together:

\[
m_b = \log(\frac{A_g}{T}) + 0.01\Delta + 5.9 = \underline{\hspace{2cm}}
\]

This magnitude is labelled \( m_b \), with the \( b \) standing for body wave. If the magnitude is computed from surface waves, it is called \( M_s \).

**How close did you get?**

Did we get the right answer? Let's compare our location and magnitude with that determined by the experts. A good location to find out information about recent earthquakes is the website maintained by the National Earthquake Information Center of the U.S. Geological Survey (see web address below). On that site we can find information about current and recent earthquakes.
Surf the NEIC site to find out what major earthquakes occurred near the time you determined for the INY seismogram.

Is there a listing for an earthquake that matches your time and location?

If so, what is it's location?

How close did your location estimate come?

What was its "official time"? ________________

What is the difference between your estimated time and the listed time? ________________ (min, sec)

What could be the sources of error involved in this difference?

What is the listed body wave magnitude for this event? ______

What is the difference between your magnitude estimate and the "official" magnitude? ____________

What could be the sources of error involved in this difference?

What about our neighborhood?

The eastern U.S. is not normally thought of as “earthquake country”. However, some of the most damaging events in history have occurred east of the Rocky Mountains. The Charleston, S.C., earthquake of 1886 and the New Madrid Missouri earthquakes of 1811-12 are thought to have been the largest in recorded U.S. history (Figure 11).

Consider the following earthquake seismograms.
EQK II

Z \uparrow

N \uparrow

\text{OHIO EQK 9 25 99 NY}
How do they differ from those of Earthquake I?

These seismograms correspond to an earthquake in western Pennsylvania (which is usually called, for perverse reasons, the Ohio earthquake of 1998).

Identify the P and S waves for the Ohio event. Measure the S-P time.

\[ S-P = \] \[ \] km

Because this earthquake was so close, the global travel time curve is not very useful. Use the regional travel-time curve (Figure 12) to compute the distance from INY to this local earthquake.

\[ \text{Distance} = \] \[ \] km

How does this distance compare with the distance from INY to the known location of the Ohio (Pennsylvania) earthquake. Hint: check the web for the earthquake location.

\[ \text{Map distance to published epicenter location=} \] \[ \] km

\[ \text{Difference between map distance and your computed distance } = \] \[ \] km
Figure 12.
References

Books:


On the web:

The National Earthquake Information Center:
http://wwwneic.cr.usgs.gov/

Seismo-Surfing the Web:
http://www.geophys.washington.edu/seismosurfing.html