



Plate Tectonics: Earthquake Epicenter

Introduction

History

Nothing can strike fear in the hearts of people quite like a natural disaster. The awesome power unleashed by a hurricane, tornado, volcano, or earthquake can cause large losses in life and property. Even the Earth, which we often think of as solid and unchanging, can be irrevocably scarred by one of these events. In ancient days, the occurrence of a natural disaster was often attributed to the anger of the gods who were displeased with some action of mankind. Today, most people do not hold with such ideas, relying on our knowledge of science to explain their appearance. In each case, we are able to describe the physical situations that produce these natural disasters. For instance, earthquakes occur when stresses that have built up in crust of the Earth are relieved by large blocks of rock moving very rapidly. However, for all of our knowledge of these disasters, we have yet to be able to accurately predict them.

Of these four natural disasters, the one that has the greatest reputation for loss of life and societal damage is an earthquake. History is replete with stories of massive damage from these events. Until very recently, the recording of earthquakes required that a civilization be near the epicenter of the earthquake and that this society have some way of recording it. It is, therefore, not surprising that the earliest recordings of earthquakes that we have today come from China in 1177 B.C. China, which has had a literate, recording society for many millennia, is situated in a location that has many large and lethal earthquakes. Throughout time, they have recorded a large amount of death and destruction, from the 1556 Shensi earthquake that killed 830,000 to the 1976 Tangshan earthquake that killed over 250,000 people. Other countries have also recorded vast amounts of death and damage, as seen in Table 1.

Year	Location	Dead
1290	Hopeh Province, China	100,000
1556	Shensi Province, China	830,000
1667	Shemaka, Russia	80,000
1737	Calcutta, India	300,000
1908	Messina, Italy	75,000
1920	Kansu Province, China	180,000
1923	Tokyo, Japan	142,807
1932	Kansu Province, China	70,000
1935	Quetta, India	60,000
1976	Tangshan, China	255,000

Table 1: Historical earthquakes²



Figure 1: 1964 Alaskan earthquake

The history of recorded earthquakes in the U.S. is filled with many large earthquakes, but few that rival the loss of life as those already mentioned. In 1811-1812, a series of large earthquakes that have been estimated above an 8 on the Richter scale hit near the area of New Madrid, Missouri. The vibrations from these earthquakes were felt as far away as the New York and Boston. In 1886, a devastating earthquake hit Charleston, South Carolina that killed over 60 people and destroyed almost all buildings in the city. Possibly the most destructive earthquake in the U.S. occurred in San Francisco during 1906. This earthquake caused a series of large fires that burned large sections of the city and killed over 700 people.

Most of these large earthquakes occurred before we were able to accurately measure their magnitudes. One of the largest that has occurred since we have been able to measure magnitudes happened in Alaska on March 27, 1964. It had a magnitude of 9.2 and created vertical displacement of over 11 meters high. The focus was situated under the Prince William

Sound, and resulted in a tidal wave that killed 110 people in Alaska, Canada, Hawaii, and the continental U.S. It was so large that it caused the water in pools in Texas and Louisiana to slosh.³

Structure of the Crust

In order to understand in better detail the reasons for earthquakes, we must come to a greater understanding of how our crust operates. In 1912, Alfred Wegener, a German meteorologist, proposed what seemed to be a rather preposterous scientific theory. He posited that the crust of the Earth was split up into giant plates of rock that were free to move about on the top of the mantle. This theory, known as continental drift, had several key pieces of evidence to support it. The first of these was that easternmost margins of the continental shelves of North and South America seemed to fit into the westernmost margins of those of Europe and Africa. Furthermore, the geographic features, such as rock layers, mountain ranges, and glacial scoring, lined up if one were to place the continents together as such. The fossil record also showed some anomalies that were easily explained if these continents had been placed together some time in the past. Although there was much to support this theory, it suffered from one great problem: there was no known mechanism that could have driven the rock layers to their locations.

After World War II, more evidence was discovered that helped support the idea that the crust was made of moving plates. The discoveries of the Mid-Atlantic Ridge and the mirrored striping of magnetic fields across the Ridge led to a re-thinking of the continental drift theory and efforts to actually measure relative movements of the plates. The later discoveries of this movement and evidence for mantle convection brought the idea to the forefront of geological thinking. The theory of plate tectonics is now widely accepted and used as a basis for all work in geology and paleontology.

It is this fact that the plates of the crust are moving that accounts for almost all earthquakes. The plates are not greasy pucks sliding around on frictionless ice. They are giant slabs of solid material that are moving past other solid pieces of rock with great deals of friction, stress, and strain. As they move, the plates deform, warp, and stretch, sometimes buckling in two or cracking in place. This means that earthquakes can occur in any location on the Earth. However, they are much more likely to occur at the boundaries between plates, as this is where the most amount of stress and buckling occur. To understand this, let us look at the types of boundaries that there are.

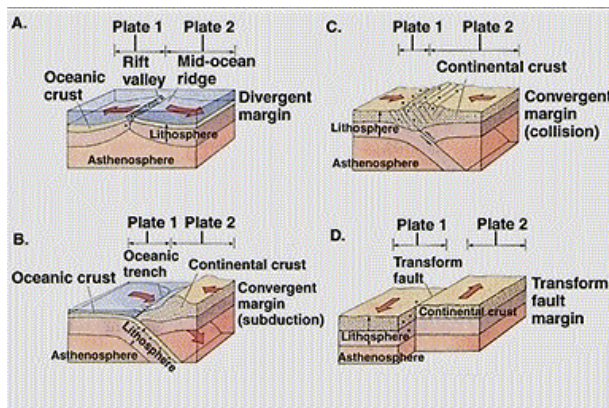


Figure 2: Plate margins

Since the Earth is not changing in size, the fact that there are moving plates means that there will be three types of boundaries between the plates. There will be boundaries where plates are moving apart, where plates are moving together, and where plates are moving past each other. Divergent zones (Figure 2A) are characterized by large-scale volcanoes and shallow, small magnitude earthquakes. As the plates come apart, molten magma from near the crust-mantle interface is allowed to seep to the surface and fill in the cracks. This magma pushing up causes some stress and cracking in the surrounding rock, thus creating small earthquakes. The Mid-Atlantic Ridge is a perfect example of this type of zone.

Convergent zones are separated into two categories depending upon the types of rocks on either side of the boundary. When continental crust (thicker and less dense) crashes into oceanic crust (thinner and denser), the oceanic crust goes under the continental crust to form a subduction zone (Figure 2C). The oceanic crust that is driven down toward the mantle melts as it encounters higher temperatures and pressure. As this magma is driven up through cracks created in the continental crust, dissolved gases within it begin to expand out of solution due to the reduced pressure. This forces the magma up through the cracks faster, which often results in explosive volcanism. These zones are characterized by deep-seated earthquakes and stratovolcanoes that form island arc near the continental shelf.

The other type of convergent zone occurs when continental crust crashes into continental crust (Figure 2B). When this occurs, the two plates fold and crumple at the boundary, uplifting material into mountain chains. This results in few, if any, volcanoes and numerous shallow earthquakes. The best example of this is the Himalaya Mountains that form the boundary between the Indian Plate and Asia.

The last type of boundary is a transform fault (Figure 2D) that will form between two plates that are sliding past one another. This type of boundary has never resulted in volcanoes in recorded history. The most likely explanation of this is that there is no crustal plate melting, nor any exposure to the mantle, with this type of boundary. However, earthquakes are very prevalent in this type of zone, as there is a great deal of friction between the plates that will allow cause no movement for long periods of time while stress builds up. Earthquakes in this zone are usually shallow. In the U.S., our favorite example of this type of boundary is the San Andreas fault that runs through California.

As previously stated, earthquakes do occur in other locations than plate boundaries. They will occur as stress is relieved slowly over time from the entire plate moving in the direction it is going. Oftentimes, these earthquakes occur at former plate boundaries between the "welded" sections of the plate. For instance, the Appalachian Mountains are the site of a former boundary between North America and Europe over 300 million years ago. They were formed when the two proto-continent crashed into one another and welded together. Because of their history, they are a weak area of the plate. Possibly the most famous intracontinental earthquake zone in the U.S. exists near New Madrid, Missouri, where the famous 1811-1812 earthquakes occurred.

Earthquake Waves

When a violent release of energy occurs because of rock movement, vibrations are generated as the rock moves past each other. These vibrations create waves, which travel away from the focus of the earthquake. Some of these waves only travel along the surface of the earth and are, therefore, called surface waves. Those that travel through the interior of the earth are called body waves, and fall into two different categories: compressional and shear waves. The frequencies of some of these waves are high enough to be heard, while others fall below our hearing level of about 10 hertz. As they travel, they obey all of the principles of other types of wave motion, meaning that they will reflect and refract at boundaries between parts of the Earth that have different wave speeds.

The surface waves are designated as either Rayleigh or Love waves, depending upon whether their motion is vertical or horizontal (Figure 3). These types of waves have the slowest wave speed, but are usually responsible for most of the damage that occurs during an earthquake, as their motion drives the ground up and down, and side to side. Since these waves spread in all surface directions, their amplitudes decrease with distance away from the epicenter, making them truly dangerous only within a hundred miles or less of the epicenter.

Of the two types of body waves, shear or S-waves have the slowest speed. These waves travel in a transverse fashion, as the material through which it travels is moved at right angles to the direction that the wave is travelling. The analog to this type of wave is the type of wave that travels on a plucked, taut string. As the wave travels down the string, the string vibrates back and forth. This is different from the compressional or P-waves, which travel by material movement in the same direction as the wave is moving. To visualize how this wave travels, imagine a Slinky that is placed on a floor and pulled into a straight line. If the Slinky is moved back and forth in the direction of the line that the Slinky makes, a

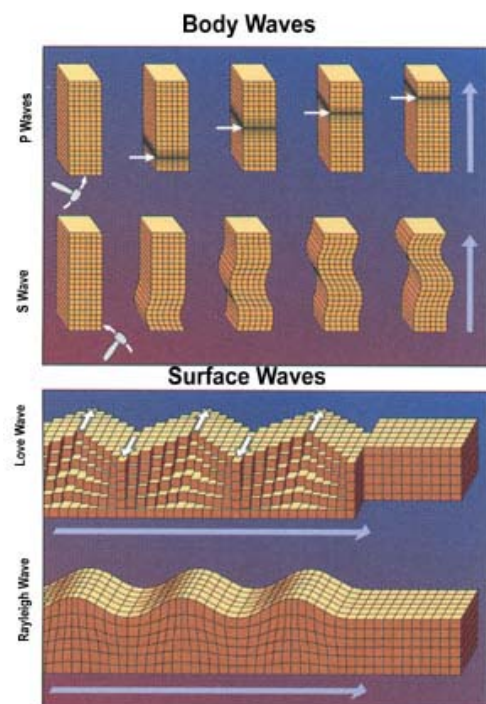


Fig 3: Seismic waves (USGS)

compressional wave will travel down its length.

This difference between P and S-waves in how they travel is very important. In order for S-waves to propagate, they require a fairly stiff material matrix, i.e. they need for the molecules in the material through which they are travelling to be tightly bound to their nearest neighbors. This is so that there will be internal forces that will pull material back to its original location as it is perturbed away from it. Because of this, S-waves cannot travel through a liquid or a gas. P-waves, on the other hand, have no such restriction, and can travel through all types of media (sound waves in air are an example of a compressional wave). This difference allows us to determine a very important fact about the interior of the Earth: the outer core is liquid. S-waves are not able to penetrate this layer and are reflected from the core/mantle boundary.

These waves can be detected through the use of a seismograph. As the waves pass by the seismograph, they will jiggle the base of device, which causes the pens to mark out the amplitude of the wave on the paper as it moves underneath it. The amplitude of the wave will depend upon the strength of the earthquake and the distance it is from the seismograph, as the wave energy is spread over a greater region as it travels and loses intensity due to reflections. The time that it takes for the waves to reach the seismograph also depends on this distance, as well as the type of wave. The differences in their wave speeds means that P-waves hit a spot first, followed by S-waves and then surface waves.

Earthquake Magnitude

The earliest attempts to quantify the magnitude of an earthquake relied upon assessing the damage near its epicenter. Different numbers were assigned to an earthquake depending upon whether buildings were knocked down, trees toppled over, or things fell off of shelves. The scale that was developed for doing this was known as the Mercalli scale, after its inventor Giuseppe Mercalli. The problem with using this scale to quantify earthquakes was that it was too subjective. Depending upon many factors such as the quality of the buildings, the population density of the region, and the investigator doing the survey, different numbers could be assigned to the same earthquake.

In order to get a more objective measure of the magnitude of an earthquake, Dr. Charles Richter developed the Richter scale in 1935. This system assigned each earthquake a number dependent upon the amplitude of the waves generated by it. Since there was such a broad range of values of amplitudes for earthquakes measured, Richter made the scale logarithmic. This means that each increase of one on the Richter scale corresponds to an increase of 10 in the amplitude of the waves. Thus, on the Richter scale, a magnitude-8 earthquake is 10 times greater in amplitude than a magnitude-7 and 100 times greater than a magnitude-6.

While the Richter scale does allow one to objectively measure an earthquake, it still does not give one a sense for the damage that an earthquake can render. For instance, the 1964 Alaska earthquake that we mentioned before was 9.2 on the Richter scale, the second highest value every recorded. While Figure 1 attests to the damage to buildings on the fault line being severe, only 110 people died in this earthquake. This is far less than the numbers of people killed in the earthquakes listed in Table 1, even though each of these earthquakes was probably less than a magnitude 9 on the Richter scale. In order to give one a sense of how much damage is done by an earthquake, scientists have developed the Modified Mercalli scale. Earthquakes are given a letter from Roman numeral I up to Roman numeral XII based upon eyewitness accounts and observations of the epicenter after the earthquake. While not as precise as the Richter scale, its narrowed classification schemes over the old Mercalli scale do allow for a reasonable estimation as to the damage to humans and their buildings.

References

- 1 <http://pubs.usgs.gov/gip/earthq1/>, September 13, 2003.
- 2 The World Book Encyclopedia, Volume 6, p. 19, Field Enterprises Educational Corporation, Chicago, 1974.
- 3 http://neic.usgs.gov/neis/eqlists/USA/1964_03_28.html, September 13, 2003.

Activity

The proximity of Jonesboro to the New Madrid Fault Zone makes an understanding of plate tectonics and earthquakes a necessity. In this week's activity, we are going to try to locate the epicenter of an earthquake using readings from three different seismograph stations. This will be done by measuring the difference in time between the arrival of the P and S-waves. The difference in time of their arrivals is due to the difference in speeds for both waves. In particular, the difference in time is given by $(\text{distance to epicenter}) / (V_p - V_s)$. Thus, we can find out how far away a particular seismograph is from an earthquake by solving this equation for distance. Since there are three stations, we can triangulate between all three to find the exact spot.

Before we use actual data, we will first do a "test run" using a simulator from California State University – Los Angeles. This simulator will walk you through the steps of epicenter triangulation that we will then apply to some real data recorded by seismographs. To get to the simulator, click on

<http://sciencecourseware.com/VirtualEarthquake/>

When you get to the simulator site, select "San Francisco area" at the bottom of the page, and follow the instructions on the succeeding pages.

Once you have completed this activity and understand how to use triangulation to find epicenters, you are ready to test out your skills on a real set of data. The [USGS Earthquake website](#) maintains a running record from seismographs that are stationed around the world. For your activity, we have captured this data for September 11, 2003. The data for three different stations is attached on the last sheet of this activity. Follow the steps below:

1. Print out the seismographic records from three different stations on that day, and determine the arrivals of the P and S waves on them.
2. Measure the differences in the arrival times for each station, and plot these on the Travel Time Curve plot on the Activity Sheet to determine the distance from the earthquake's epicenter to the station.
3. Using a scale of 1 inch = 2,500 miles, determine the radius of the circle to draw from each station on the attached world map
4. For each station, set a compass to draw a circle of the corresponding radius in inches. Placing the sharp end of the compass on the station, draw a circle to represent the distance that the epicenter is away from the station.
5. Repeat this process for the other two stations.
6. Determine the epicenter by the intersection of the three circles.
7. Answer the questions on the activity sheet.

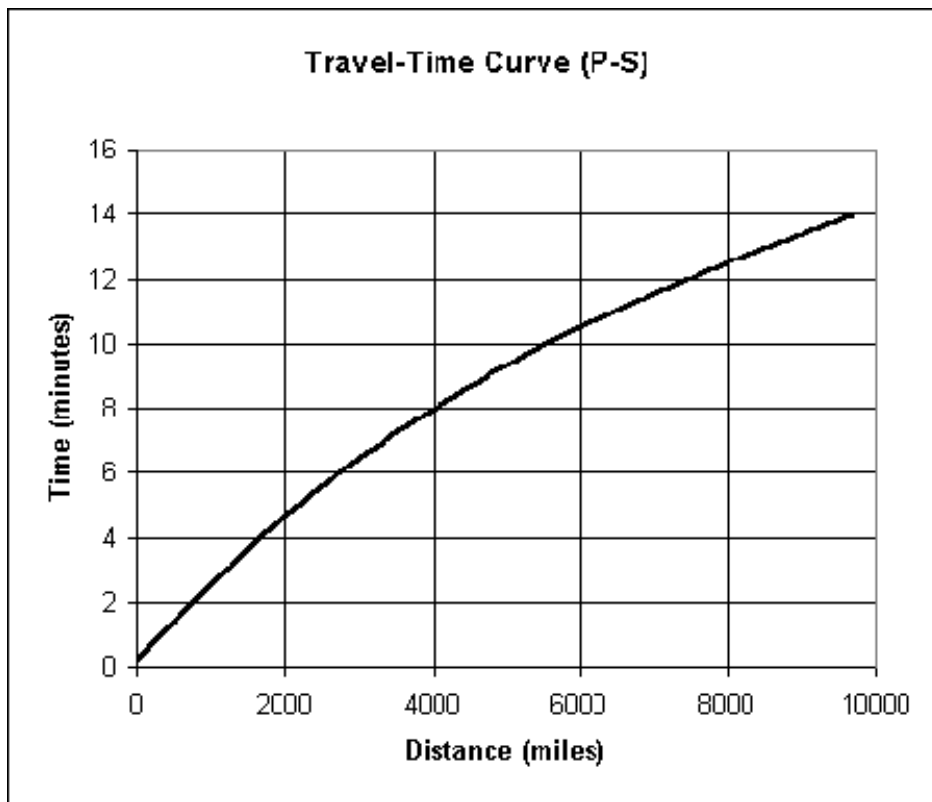
Links

USGS Seismograms - <http://earthquake.usgs.gov/eqcenter/helicorders.php>

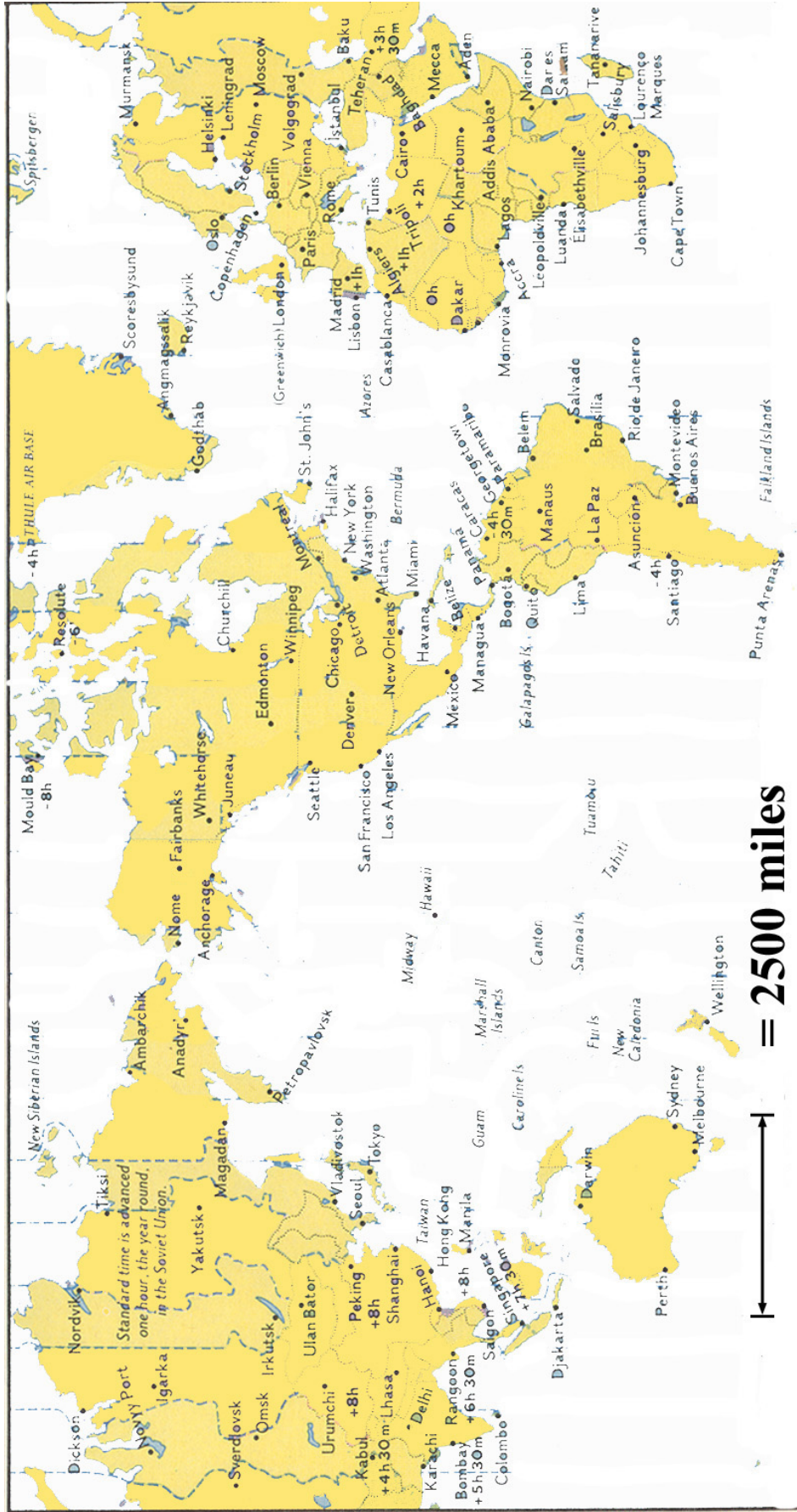
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Name:

1. For the "Virtual Earthquake" activity, what is the epicenter of the earthquake?
2. What is the epicenter for the actual earthquake you studied? Can you find any evidence of this earthquake in the news or online to support your supposition?
3. How well does this triangulation method work?

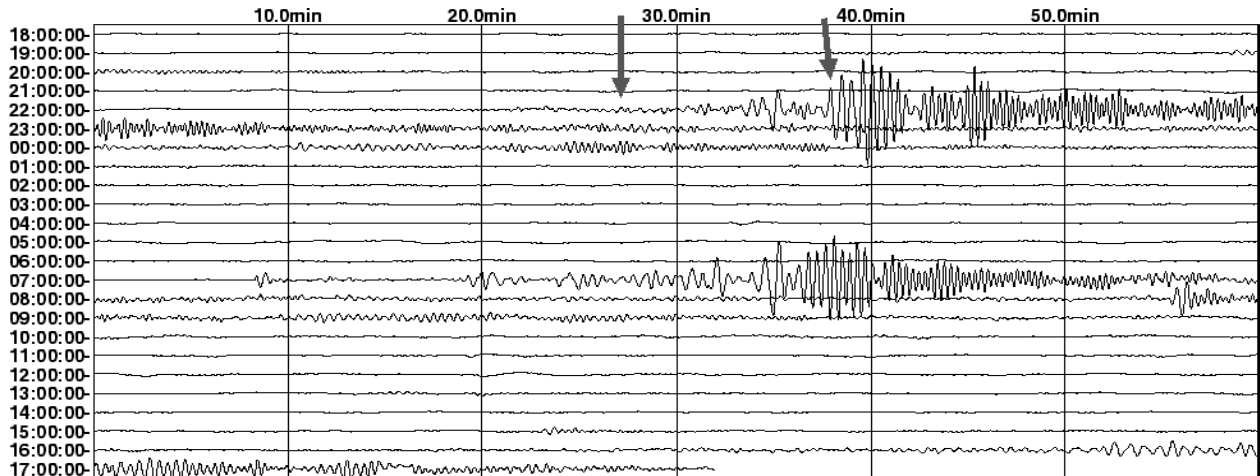


Mark your travel-times on the above chart and attach your world map to show your triangulation work.

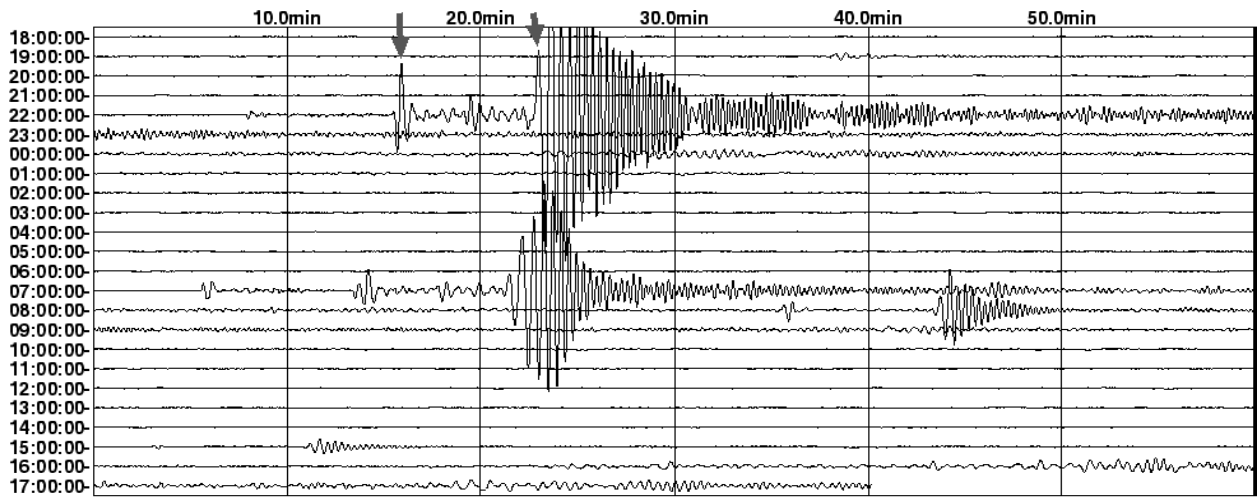


= 2500 miles

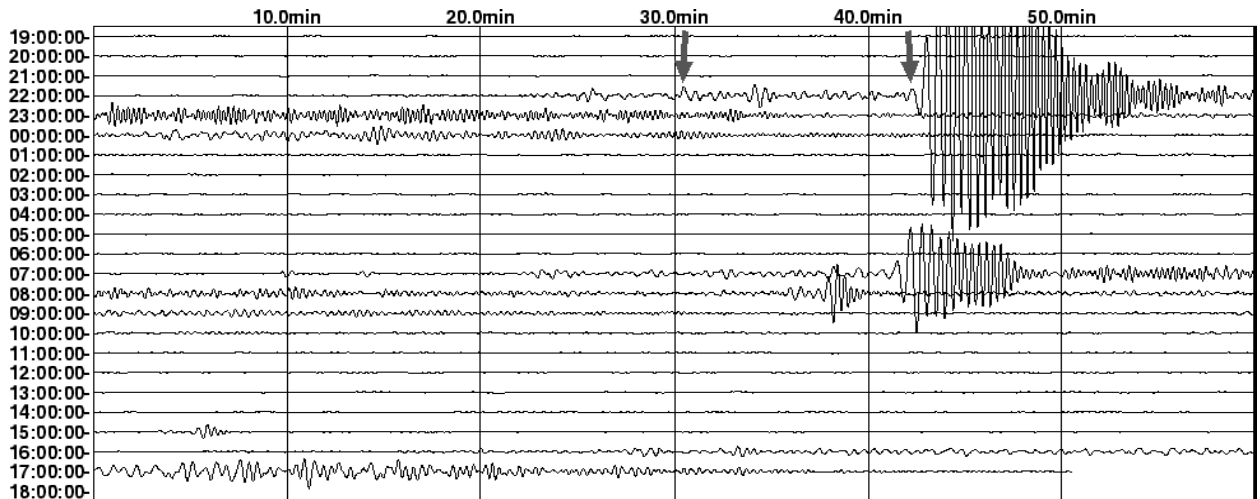




College Outpost, Alaska



Kipapa, Hawaii



Tucson, Arizona