Seismic Waves and the Slinky

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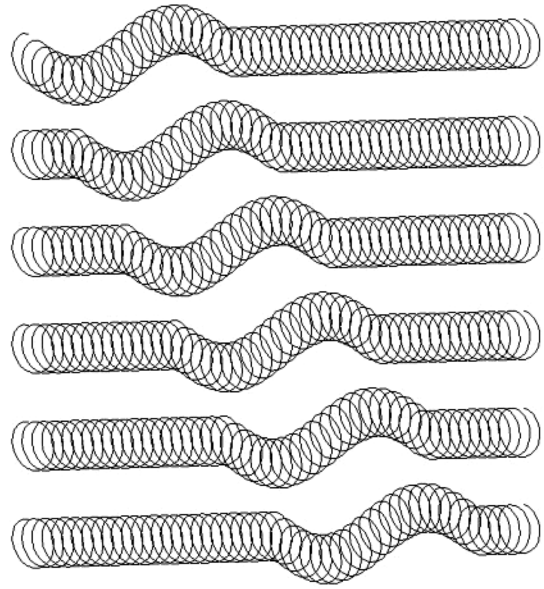
Seismic Waves: Because of the elastic properties of Earth materials (rocks) and the presence of the Earth's surface, four main types of seismic waves propagate within the Earth. *Compressional* (P) and *Shear* (S) waves propagate through the Earth's interior and are known as body waves. *Love* and *Rayleigh* waves propagate primarily at and near the Earth's surface and are called surface waves. Wave propagation and particle motion characteristics for the P, S, Rayleigh and Love waves are illustrated in Figures 1-4. Additional illustrations of P, S, Rayleigh and Love waves are contained in Bolt (1993, p. 27 and 37; 2004, p. 22) and in Shearer (1999, p. 32 and 152). Effective animations of P and S waves are contained in the Nova video "Earthquake" (1990; about 13 minutes into the program) and of P, S, Rayleigh and Love waves in the Discovery Channel video "Living with Violent Earth: We Live on Somewhat Shaky Ground" (1989, about 3 minutes into the program), and at: http://web.ics.purdue.edu/~braile/edumod/waves/WaveDemo.htm.

Slinky Demonstrations of P and S Waves: The P and S waves have distinctive particle motions (Figures 1-4) and travel at different speeds. P and S waves can be demonstrated effectively with a slinky (the original metal slinky works best). For the P or compressional wave, have two people hold the ends of the slinky about 3-4 meters apart. One person should cup his or her hand over the end (the last 3-4 coils) of the slinky and, when the slinky is nearly at rest, hit that hand with the fist of the other hand. The compressional disturbance that is transmitted to the slinky will propagate along the slinky to the other person. Note that the motion of each coil is either compressional or extensional with the movement parallel to the direction of propagation. Because the other person is holding the slinky firmly, the P wave will reflect at that end and travel back along the slinky. The propagation and reflection will continue until the wave energy dies out. The propagation of the P wave by the slinky is illustrated in Figure 5.

Demonstrating the S or Shear wave is performed in a similar fashion except that the person who creates the shear disturbance does so by moving his or her hand quickly up and then down. This motion generates a motion of the coils that is perpendicular to the direction of propagation, which is along the slinky. Note that the particle motion is not only perpendicular to the direction of motion but also



in the vertical plane. One can also produce Shear waves with the slinky in which the motion is in the horizontal plane by the person creating the source moving his or her hand quickly left and then right. The propagation of the S wave by the slinky is illustrated in Figure 6. Notice that, although the motion of the disturbance was purely perpendicular to the direction of propagation (no motion in the disturbing source was directed along the slinky), the disturbance still propagates away from the source, along the slinky. The reason for this phenomenon (a good challenge question for students) is because the material is elastic and the individual coils are connected (like the individual particles of a solid) and thus transmit their motion (disturbance or deformation) to the adjacent coils. As this process continues, the shear disturbance travels along the entire slinky (elastic medium).

P and S waves can also be generated in the slinky by an additional method that reinforces the concept of elasticity and the elastic rebound theory which explains the generation of earthquakes by plate tectonic movements (Bolt, 1993, p. 74-77; 2004, p. 113-116). In this method, for the P wave, one person should slowly gather a few of the end coils of the slinky into his or her hand. This process stores elastic energy in the coils of the slinky that are compressed (as compared to the other coils in the stretched slinky) similar to the storage of elastic energy in rocks adjacent to a fault that are deformed by plate motions prior to slip along a fault plane in the elastic rebound process. When a few coils have been compressed, release them suddenly (holding on to the end coil of the slinky) and a compressional wave disturbance will propagate along the slinky. This method helps illustrate the concept of the elastic properties of the slinky and the storage of energy in the elastic rebound process. However, the compressional wave that it generates is not as simple or visible as the wave produced by using a blow of one's fist, so it is suggested that this method be demonstrated after the previously described method using the fist.

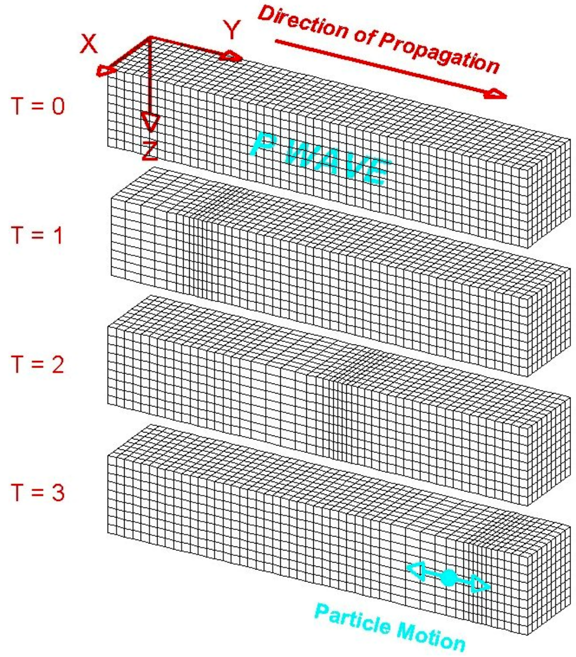


Figure 1. Perspective view of elastic P-wave propagation through a grid representing a volume of material. The directions X and Y are parallel to the Earth's surface and the Z direction is depth. T = 0 through T = 3 indicate successive times. The disturbance that is propagated is a compression (grid lines are closer together) followed by a dilatation or extension (grid lines are farther apart). The particle motion is in the direction of propagation. The material returns to its original shape after the wave has passed.

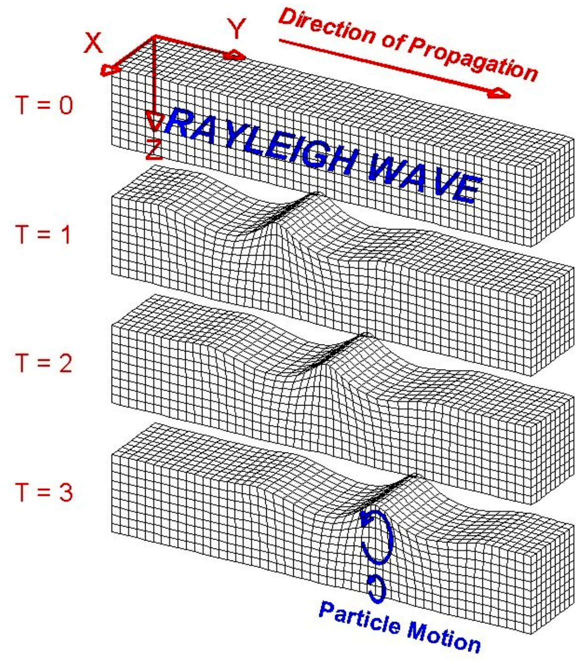


Figure 3. Perspective view of Rayleigh-wave propagation through a grid representing a volume of elastic material. Rayleigh waves are surface waves. The disturbance that is propagated is, in general, an elliptical motion which consists of both vertical (shear; perpendicular to the direction of propagation but in the plane of the raypath) and horizontal (compression; in the direction of propagation) particle motion. The amplitudes of the Rayleigh wave motion decrease with distance away from the surface. The material returns to its original shape after the wave has passed.

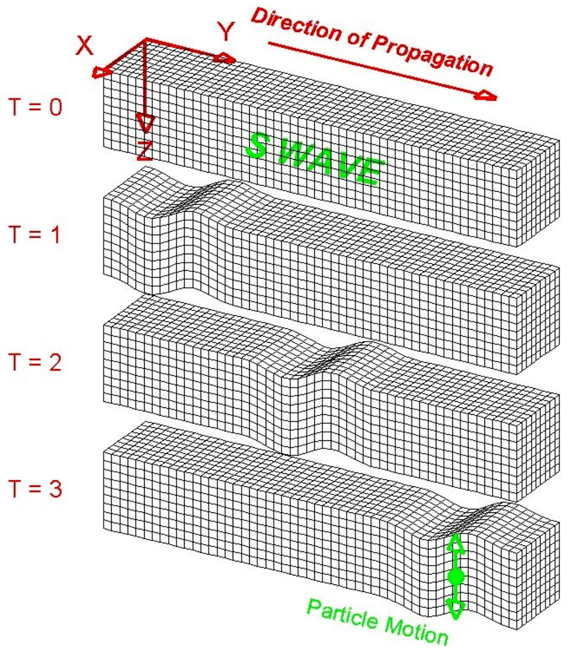


Figure 2. Perspective view of S-wave propagation through a grid representing a volume of elastic material. The disturbance that is propagated is an up motion followed by a down motion (the shear motion could also be directed horizontally or any direction that is perpendicular to the direction of propagation). The particle motion is perpendicular to the direction of propagation. The material returns to its original shape after the wave has passed.

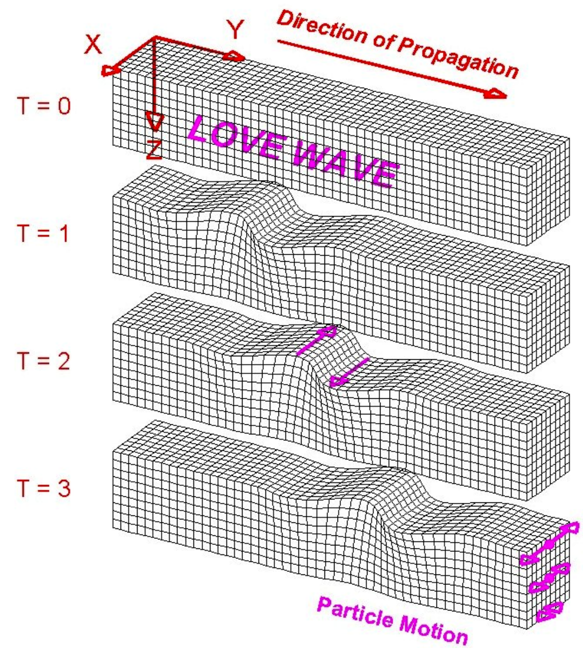


Figure 4. Perspective view of Love-wave propagation through a grid representing a volume of elastic material. Love waves are surface waves. The disturbance that is propagated is horizontal and perpendicular to the direction of propagation. The amplitudes of the Love wave motion decrease with distance away from the surface. The material returns to its original shape after the wave has passed.

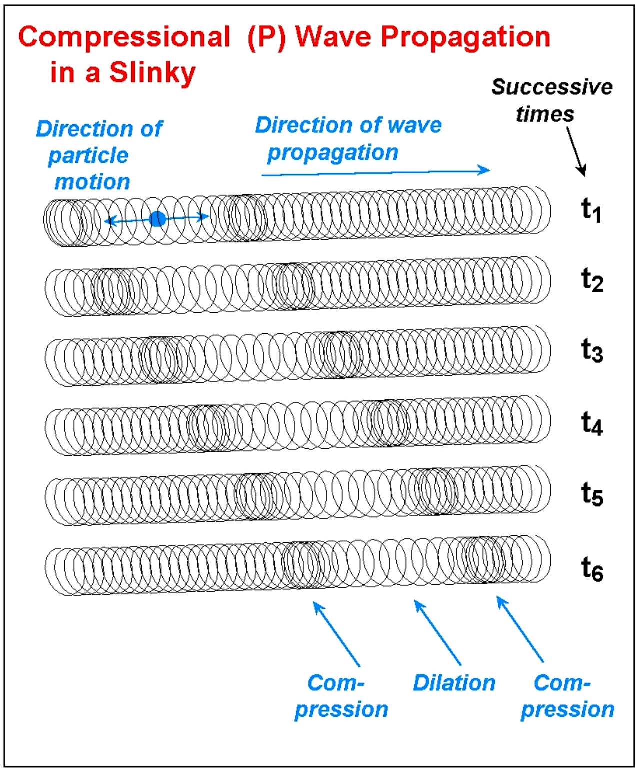


Figure 5. Compressional (P) wave propagation in a slinky. A disturbance at one end results in a compression of the coils followed by dilation (extension), and then another compression. With time (successive times are shown by the diagrams of the slinky at times t_1 through t_6), the disturbance propagates along the slinky. After the energy passes, the coils of the slinky return to their original, undisturbed position. The direction of particle motion is in the direction of propagation.

Similarly, using this "elastic rebound" method for the S waves, one person holding the end of the stretched slinky should use their other hand to grab one of the coils about 10-12 coils away from the end of the slinky. Slowly pull on this coil in a direction perpendicular to the direction defined by the stretched slinky. This process applies a shearing displacement to this end of the slinky and stores elastic energy (strain) in the slinky similar to the storage of strain energy in rocks adjacent to a fault or plate boundary by plate tectonic movements. After the coil has been displaced about 10 cm or so, release it suddenly (similar to the sudden slip along a fault plane in the elastic rebound process) and an S wave disturbance will propagate along the slinky away from the source.

Surface Waves: The Love wave (Figure 4) is easy to demonstrate with a slinky or a double length slinky. Stretch the slinky out on the floor or on a tabletop and have one person at each end hold on to the end of the slinky. Generate the Love wave motion by quickly moving one end of the slinky to the left and then to the right. The horizontal shearing motion will propagate along the slinky. Below the surface, the Love wave motion is the same except that the amplitudes decrease with depth. Using the slinky for the Rayleigh wave (Figure 3) is more difficult. With a regular slinky suspended between two people, one person can generate the motion of the Rayleigh wave by rapidly moving his or her hand in a circular or elliptical motion. The motion should be up, back (toward the person generating the motion), down, and then forward (away from the person), coming back to the original location and forming an ellipse or circle with the motion of the hand. This complex pattern will propagate along the slinky but will look very similar to an S wave. Rayleigh wave motion also decreases with depth below the surface. Excellent illustrations of the wave motion of Love and Rayleigh waves can also be found in Bolt (1993, p. 37). Further details on the characteristics and propagation of Love and Rayleigh waves can be found in Bolt (1993, p. 37-41).

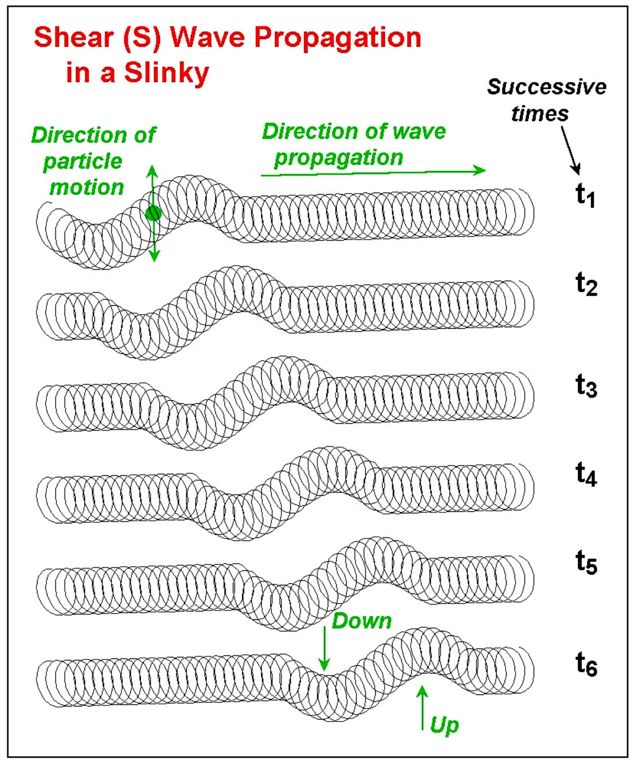


Figure 6. Shear (S) wave propagation in a slinky. A disturbance at one end results in an up motion of the coils followed by a down motion of the coils. With time (successive times are shown by the diagrams of the slinky at times t_1 through t_6), the disturbance propagates along the slinky. After the energy passes, the coils of the slinky return to their original, undisturbed position. The direction of particle motion is perpendicular (for example, up and down or side to side) to the direction of propagation.

Wave Propagation in All Directions: An additional demonstration with P and S waves can be performed with the 5-slinky model. By attaching 5 slinkys to a wood block as shown in Figure 7, 5 people can hold the ends of the 5 slinkys (stretched out in different directions to about 3-4 m each). One person holds the wood block and can generate P or S waves (or even a combination of both) by hitting the wood block with a closed fist or causing the block to move quickly up and then down or left and then right. The purpose of this demonstration is to show that the waves propagate in all directions in the Earth from the source (not just in the direction of a single slinky).

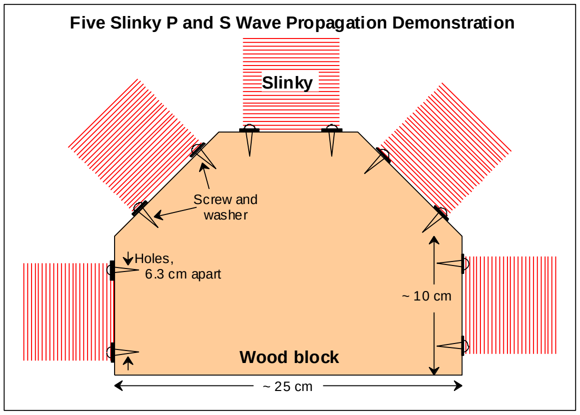


Figure 7. Diagram showing how five slinkys can be attached to the edge of a wood block. Photographs of the five slinky model are shown in Figures 8 and 9. When the slinkys are stretched out to different positions (five people hold the end of one slinky each) and a P or S wave is generated at the wood block, the waves propagate out in all directions. The five slinky model can also be used to show that the travel times to different locations (such as to seismograph stations) will be different. To demonstrate this effect, wrap a small piece of tape around a coil near the middle one of the slinkys. Have the person holding that slinky compress all of the coils from the outer end to the coil with the tape so that only one half of the slinky is extended. Also, attach an additional (sixth) slinky, using plastic electrical tape, to the end of one of the slinkys. Have the person holding this double slinky stand farther away from the wood block so that the double slinky is fully extended. When a P or S wave is generated at the wood block, the waves that travel along the slinky will arrive at the end of the half slinky first, then at approximately the same time at the three regular slinkys, and finally, last at the double length slinky

Attaching an additional slinky (with small pieces of plastic electrical tape) to one of the five slinkys attached to the wood block makes one slinky into a double length slinky which can be stretched out to 6-8 m. For one of the other four slinkys, have the person holding it collapse about half of the coils and hold them in his or her hands, forming a half slinky, stretched out about 1½ - 2 m. Now when a source is created at the wood block, one can see that the waves take different amounts of time to travel the different distances to the ends of the various slinkys. An effective way to demonstrate the different arrival times is to have the person holding each slinky call out the word "now" when the wave arrives at their location (if the people holding the slinkys close their eyes and call out when they feel the wave arrive, their responses may be more accurate). The difference in arrival times for the different distances will be obvious from the sequence of the call of "now." This variation in travel time is similar to what is observed for an earthquake whose waves travel to various seismograph stations that are different distances from the source (epicenter). Although these two demonstrations with the five slinky model represent fairly simple concepts, we have found the demonstrations to be very effective with all age groups. In fact, the five slinky demonstrations are often identified the "favorite activities" of participants.

Exploration and Assessment: After demonstrating seismic waves with the slinky, have students use the slinky to explore wave propagation and generation of different wave types and wave characteristics. One can also use slinky activities for authentic assessment by asking students to show their understanding by performing the demonstrations in class.

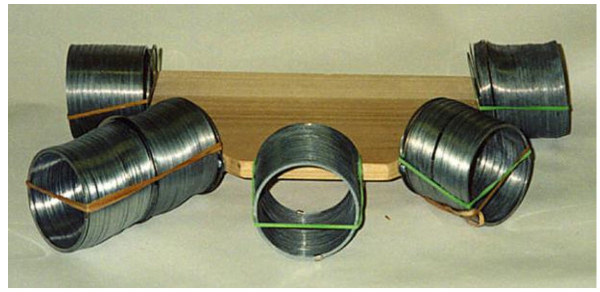


Figure 8. Photograph of the five slinky model.

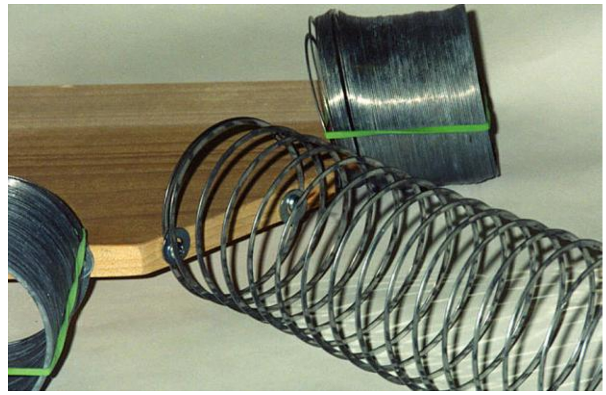


Figure 9. Close-up view of five slinky model showing attachment of slinky using screws and washers (holes are 6.3 cm apart, screws are #6, ½" long, washers are 3/16").

Connections to National Science Education Standards (National Research Council, 1996): *Teaching Standards:* inquiry-based (A,B); opportunity for assessment (C). *Professional Development Standards:* opportunity for learning new Earth science content (A,C); suggestions for effective teaching strategies (B). *Assessment Standards:* authentic assessment (C). *Content Standards: Science as Inquiry* – practice inquiry and fundamental science skills (grades 5-8 and 9-12, A); *Physical Science* – properties of matter, motion, transfer of energy (grades 5-8, B), structure of matter, motion, interactions of energy and matter (grades 9-12, B); *Earth and Space Science* – relate to energy in the Earth system (grades 9-12, D).

References:

Bolt, B.A., *Earthquakes and Geological Discovery*, Scientific American Library, W.H. Freeman, New York, 229 pp., 1993.

Bolt, B.A., *Earthquakes*, (5th edition; similar material is included in earlier editions), W.H. Freeman, New York, 378 pp., 2004.

Earthquake, NOVA series videotape, 58 minutes, available from 800-255-9424; http://www.pbs.org, 1990.

Living with Violent Earth: We Live on Somewhat Shaky Ground, Assignment Discovery series videotape, Discovery Channel, 25 minutes, http://www.dsc.discovery.com, 1989.

National Research Council, *National Science Education Standards*, National Academy of Sciences, Washington, D.C., 262 pp., 1996.

Shearer, P. M., *Introduction to Seismology*, Cambridge University Press, Cambridge, UK, 260pp, 1999.

(This description of seismic waves and the slinky is excerpted from **Seismic Waves and the Slinky: A Guide for Teachers** available at: web.ics.purdue.edu/~braile. Developed in cooperation with the IRIS Consortium [Incorporated Research Institutions for Seismology, www.iris.edu], supported by the National Science Foundation.)

Title: Human Wave Demonstration (DRAFT WRITE UP)

Version: 1.0

Last Revision: March, 2013

What's included:
- Overview
- Materials List
- Activity Flow
- Teacher Background

URL Slug: http://www.iris.edu/hq/inclass/lesson/human_wave

Thumbnail:

***Suggested Level*:** Novice

Time: 15 minutes

Learning Objectives:
Students will be able to
1. Distinguish between a P wave and an S wave based on
   a. Speed
   b. Direction of particle motion relative to wave propagation
   c. Materials it can propagate through
2. Explain molecularly, why S waves are not able to travel in a liquid, while P waves are able to travel in a liquid.
3. Explain how P and S waves provide evidence for Earth's interior

Material List
1. Space in the front of the room for 10-15 students to line-up shoulder-to-shoulder.
2. Optional: Stopwatch
3. Projection system and computer
4. http://web.ics.purdue.edu/~braile/edumod/waves/Pwave.htm
5. http://web.ics.purdue.edu/~braile/edumod/waves/Swave.htm
6. Venn Diagram worksheet

Abstract (200 words or less)
Lined up shoulder-to-shoulder, students to "become" the material that P and S waves travel through in this demonstration. Once "performed," the principles of P and S waves will not be easily forgotten.

Overview:
This demonstration explores two of the four (P and S waves) main ways energy propagates from the hypocenter of an earthquake. In this demo, students "experience" the waves as they line up shoulder-to-shoulder to "become" the solid and then liquid material the waves travel through. The physical nature of the "Human wave" demonstration makes it a highly engaging activity for most students. Some find that this tactile/kinesthetic learning activity stretches them personally; while for others it channels disruptive energy into a creative endeavor. Either way, developing ways to physically involve students in learning helps students grasp, internalize and maintain abstract information. Once "performed," the principles of P and S waves will not be easily forgotten.

Position in a teaching sequence or other related resources
This activity is part of something larger… to see this as part of a teaching sequence visit www.IRIS:InClass/sequences

Teacher Preparation – None

Safety – Have a "spotter" at one end of the line to ensure that the last person is not bumped over.

Vocabulary
Elastic deformation
Primary (P) Wave
Propagate
Secondary (S) Wave
Seismic Waves
Shearing
Wave

Activity Flow
There are a number of ways to use this demo. It can precede or follow a slinky demonstration depending on the rest of your instruction. As written here it follows and reinforces instruction of seismic waves with the slinky.

Open –
1. Today we are going to have an earthquake in class!

2. Ask for approximately 10 to 12 students to come to the front of the room and lineup tallest to shortest.

TIP – Depending on the maturity of your students, this demonstration may work best with homogeneous grouping by gender.

TIP – Since this is a kinesthetic demonstration, it may not be appropriate for some students with physical disabilities to participate.

Prior Knowledge
1. Tell the students that you are the earthquake and that they are a line of molecules in a material that the seismic energy will propagate through.

2. Instruct the line of "molecules" to become an elastic solid. If students have difficulty, coach them to the answer.

> ANSWER: The line should stand shoulder-to-shoulder with their feet shoulder-width apart. Molecules are tightly packed and are rigid in a solid. They can't slide past one another. Therefore, the students be very close together and should place their arms over the shoulders of the person next to them… chorus-line style.

3. Remind students that solids are elastic. Ask: What does that mean in terms of how the "molecules" should respond when seismic wave passes?

> ANSWER: The group should be rigid, but not overly so. Since they are an elastic material they should deform to the force that they feel and return to their original position. It is important that the molecules not be too rigid (e.g. not move when bumped) nor too limp (e.g. fall into the person next to them) for the demonstration to be effective.

Explore/**E**xplain
Seated students should create a Venn diagram about P and S waves during the demo and discussions.

Modeling a P wave in a solid
1. From the tall end of the line, lightly push first student's shoulder so that they bump the student next to them. This causes them to move closer together temporarily, a compression, followed by spreading farther apart temporarily, a dilation. This pattern to propagates down the line.

2. Ask "I was the earthquake that released energy into the system. What was transmitted? Energy? Material? Neither? Both?" Be sure to encourage students to provide evidence for their claim.

> ANSWER: Energy was transmitted from one end of the line to the other end. The molecules returned to their original position and were NOT transmitted.

3. Ask "How did the molecules move as compared to the direction of the energy?"

> ANSWER: Parallel to the direction of wave propagation and since the material is elastic the molecules were deformed as the wave passed but they returned to their original position.

4. Ask "Did it take some time for the energy to move from one end of the material to the other?"

> ANSWER: Yes

5. Since it took x amount of time for the energy to move from one end of the material to the other, what does that mean about the wave?

ANSWER: It has a definite and measurable velocity

Modeling an S wave in a solid
1. Starting at the tallest end of the line, place one hand in the center of the first person's back, and a second hand on their waist. Bend them forward at the waist and then back up again.

2. This molecules resistance to shearing (e.g. arms over each others shoulders) will cause the shearing to propagate down the line of "molecules".

3. Again note that the wave has a velocity and that this time the deformation of the molecules was perpendicular to the direction of wave propagation.

4. Many students will note that the S wave is slower than the P wave though repeating the demo and having observers time the waves can emphasize this point dramatically. (Because the shear motion in the demonstration is more complicated than the compression. This velocity difference could be measured with a stopwatch if the P and the S demo were repeated.)

5. Ask "How was the S Wave different from the P wave?"
ANSWER: Particles moved perpendicular to the direction of wave propagation and the S wave was slower than the P wave

6. Ask "How was the P Wave similar to the S wave?"
ANSWER: They both had a definite velocity and transferred energy from on end to the other.

Figure 1. An S Wave traveling through the tightly spaced molecules of a solid

Modeling A P Wave In A Liquid
1. Instruct the line of "molecules" to now become a liquid. If students have difficulty, coach them to the answer.
ANSWER: The line should stand shoulder-to-shoulder with their feet shoulder-width apart. Molecules in a liquid can slide past one another but there is little free space between them. Therefore, the students' shoulders should touch one another but they should no longer be linked chorus-line style.

2. Remind the “molecules” that they are still elastic. This means that even though they are liquids they should deform to the force that they feel and return to their original position. It is important that the molecules not be too rigid (e.g. not move when bumped) nor too limp (e.g. fall into the person next to them) for the demonstration to be effective.

3. From the tall end of the line, lightly push first student’s shoulder so that they bump the student next to them. This causes them to move closer together temporarily, a compression, followed by spreading farther apart temporarily, a dilation. This pattern to propagates down the line.

4. Ask “Was there a difference between a P Wave in a liquid vs a solid?”
    ANSWER: There was no noticeable difference (though it should be noted that in the earth, the energy is transferred more slowly in a liquid. If time allows you might explore why this occurs with your students.)

Figure 2. A P wave traveling through the molecules of a liquid

Modeling An S Wave In A Liquid

1. Starting at the tallest end of the line, place one hand in the center of the first person’s back, and a second hand on their waist. Bend them forward at the waist and then back up again. This time the shear disturbance cannot propagate down the line.
TIP - When doing this demo, the second person in line frequently bends at the waist “sympathetically”. This should be anticipated and you should clarify their movement to the class by asking them if they “felt or saw” the disturbance. Once clarified, the demo should be repeated so that this time they only move when they feel the disturbance.

2. Ask “ Was there a difference between an S Wave in a liquid vs. a solid? What was it?
    ANSWER: Yes, the wave did not travel through the liquid!

3. Ask “Based on these demonstrations, why can’t an S wave travel in a liquid?”
    ANSWER: Because the molecules in a liquid are not rigid, they can slip past one another. As a result sharing motion is not resisted and does not propagate.

Reflect

1. Quickly reverse the students and repeat the demo with the remaining half of the class

2. Document this second demonstration by creating a Venn Diagram for P and S Waves on the board. Students should also complete their version at their seats.

3. Ask “We have examined this model (the line of students) for how can behave like seismic waves moving through Earth. Do you think it is completely acute? How could it be different from reality?”

ANSWER: The Human wave demo is a functional model. Thus, it is both like and unlike the target it represents. Accept answers generated by students but be sure to supplement these with the ideas mentioned in the teacher’s background section.

Apply

1. Before the second group return to their seats, divide the students in the following way; the first quarter of the students should be solids, the middle half of the students should be liquids, and the final quarter should be solids.

2. Ask students to predict… “What do you think will happen if a P wave were to propagate through the line? An S wave?”

ANSWER: Accept all responses

3. Test their hypothesis by sending a P wave and an S wave through the line. Because P waves propagate in both solids and liquids, the P wave will propagate from one end to the other. However, the S wave will stop when it reaches the solid/liquid boundary because liquids do not resist shearing. This final demo models the S wave shadow zone; evidence that Earth’s core is a liquid.

TIP – If you have not already covered Earth’s internal structure, don’t mention the connection to Earth’s structure yet. When get to this in a few weeks, refer back to or even repeat this demo for students when you discuss seismic evidence. However, if you have already provided instruction on Earth’s interior structure, a discussion connecting the demo to the S-wave shadow zone is warranted.

Teacher Background

This demonstration explores two of the four (P and S waves) main ways energy propagates from the hypocenter of an earthquake. In this demo, students “experience” the waves as they line up shoulder to shoulder to “become” the material the waves travel through. The physical nature of the “Human wave” demonstration makes it a highly engaging activity for most students. Some find that this tactile/kinesthetic learning activity stretches them personally; while for others it channels disruptive energy into a creative endeavor. Either way, developing ways to physically involve students in learning helps students grasp, internalize and maintain abstract information (Griss,1994). Once "performed," the principles of P and S waves will not be easily forgotten.

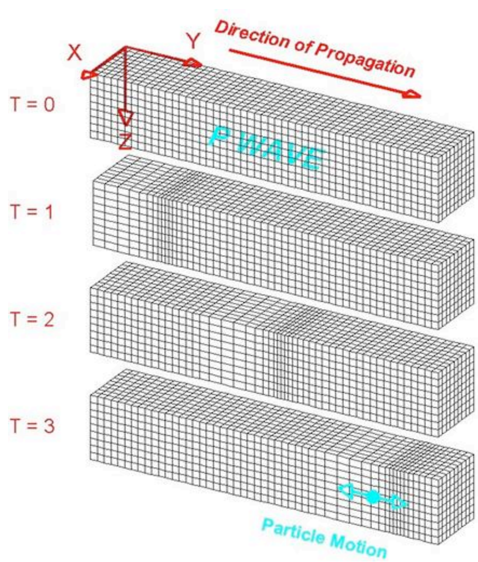
Seismic Waves

The energy from an earthquake radiates outwards in all directions. Because of the elastic properties of Earth materials (rocks) and the presence of the Earth's surface this energy propagates as four main types of seismic waves. Compressional or Primary (P) and Shear or Secondary (S) waves propagate through the Earth’s interior and are known as body waves. Love and Rayleigh waves propagate primarily at and near the Earth's surface and are called surface waves. Table 1 below summarizes detailed characteristics of the P, S Rayleigh and Love waves.

Table 1: Seismic Waves

Wave Type (and names)	Particle Motion	Typical Velocity	Other Characteristics
P, **Compressional**, Primary, Longitudinal	Alternating compressions (“pushes”) and dilations (“pulls”) which are directed in the same direction as the wave is propagating (along the ray path); and therefore, perpendicular to the wavefront.	**V_P ~ 5 – 7 km/s in typical Earth’s crust; >~ 8 km/s in Earth’s mantle and core; ~1.5 km/s in water; ~0.3 km/s in air.**	P motion travels fastest in materials, so the P-wave is the first-arriving energy on a seismogram. Generally smaller and higher frequency than the S and Surface-waves. P waves in a liquid or gas are pressure waves, including sound waves.
S, **Shear**, Secondary, Transverse	Alternating transverse motions (perpendicular to the direction of propagation, and the ray path); commonly approximately polarized such that particle motion is in	**V_S ~ 3 – 4 km/s in typical Earth’s crust;** >~ 4.5 km/s in Earth’s mantle; ~ 2.5-3.0 km/s in (solid) inner core.	S-waves do not travel through fluids, so do not exist in Earth’s outer core (inferred to be primarily liquid iron) or in air or water or molten rock (magma). S waves travel slower than P waves in a solid and, therefore, arrive after the P wave.

	vertical or horizontal planes.		
L, **Love**, Surface waves, Long waves	Transverse horizontal motion, perpendicular to the direction of propagation and generally parallel to the Earth's surface.	V_L ~ 2.0 - 4.4 km/s in the Earth depending on frequency of the propagating wave, and therefore the depth of penetration of the waves. In general, the Love waves travel slightly faster than the Rayleigh waves.	Love waves exist because of the Earth's surface. They are largest at the surface and decrease in amplitude with depth. Love waves are dispersive, that is, the wave velocity is dependent on frequency, generally with low frequencies propagating at higher velocity. Depth of penetration of the Love waves is also dependent on frequency, with lower frequencies penetrating to greater depth.
R, **Rayleigh**, Surface waves, Long waves, Ground roll	Motion is both in the direction of propagation and perpendicular (in a vertical plane), and "phased" so that the motion is generally elliptical – either prograde or retrograde.	V_R ~ 2.0 - 4.2 km/s in the Earth depending on frequency of the propagating wave, and therefore the depth of penetration of the waves.	Rayleigh waves are also dispersive and the amplitudes generally decrease with depth in the Earth. Appearance and particle motion are similar to water waves. Depth of penetration of the Rayleigh waves is also dependent on frequency, with lower frequencies penetrating to greater depth.



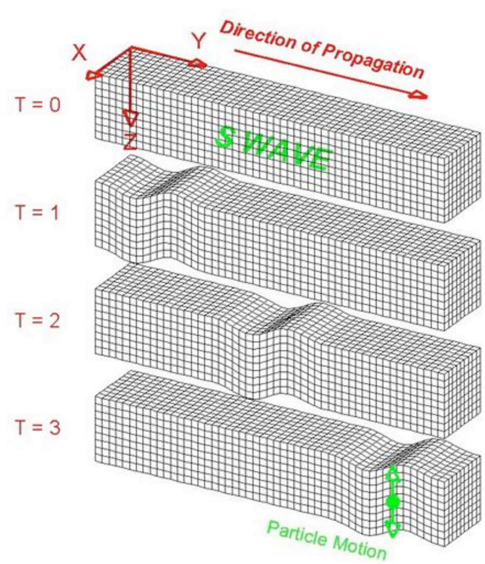


Figure 3. Perspective views of P-wave (left) and S wave (right) propagation through a grid representing a volume of material. The directions X and Y are parallel to the Earth's surface and the Z direction is depth. T = 0 through T = 3 indicate successive times. (Left) The disturbance that is propagated is a compression (grid lines are closer together) followed by a dilatation or extension (grid lines are farther apart). The particle motion is in the direction of propagation. The material returns to its original shape after the wave has passed. (Right) The disturbance that is propagated is an up motion followed by a down motion (though the motion could also be directed horizontally or any direction that is perpendicular to the direction of propagation). The particle motion is perpendicular to the direction of propagation. The material returns to its original shape after the wave has passed.

<u>S Wave Shadow Zone</u>
Most of the direct evidence that we have about Earth's deep interior comes from the study of seismic waves that penetrate the Earth and are recorded on the other side. Years of study of travel times from earthquakes to stations at various distances suggest the speeds at which P- and S-waves traverse different regions of Earth's interior. From this seismologists infer the path the seismic waves takes to reach any point on Earth's surface. As illustrated in Figure 4b below, S waves are unable this energy follows a curing path through the Earth. As a result of this, and the below, there are points on Earth's surface were direct S waves don't arrive (S waves can't propagate through Earth's liquid core).

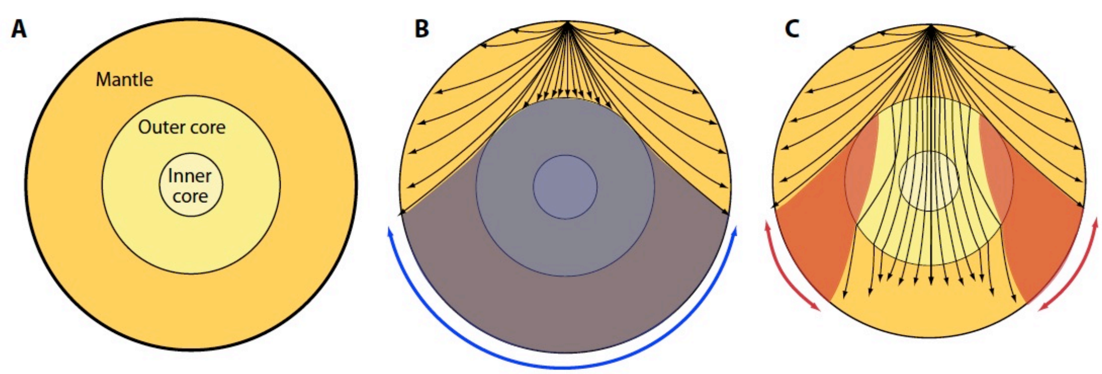
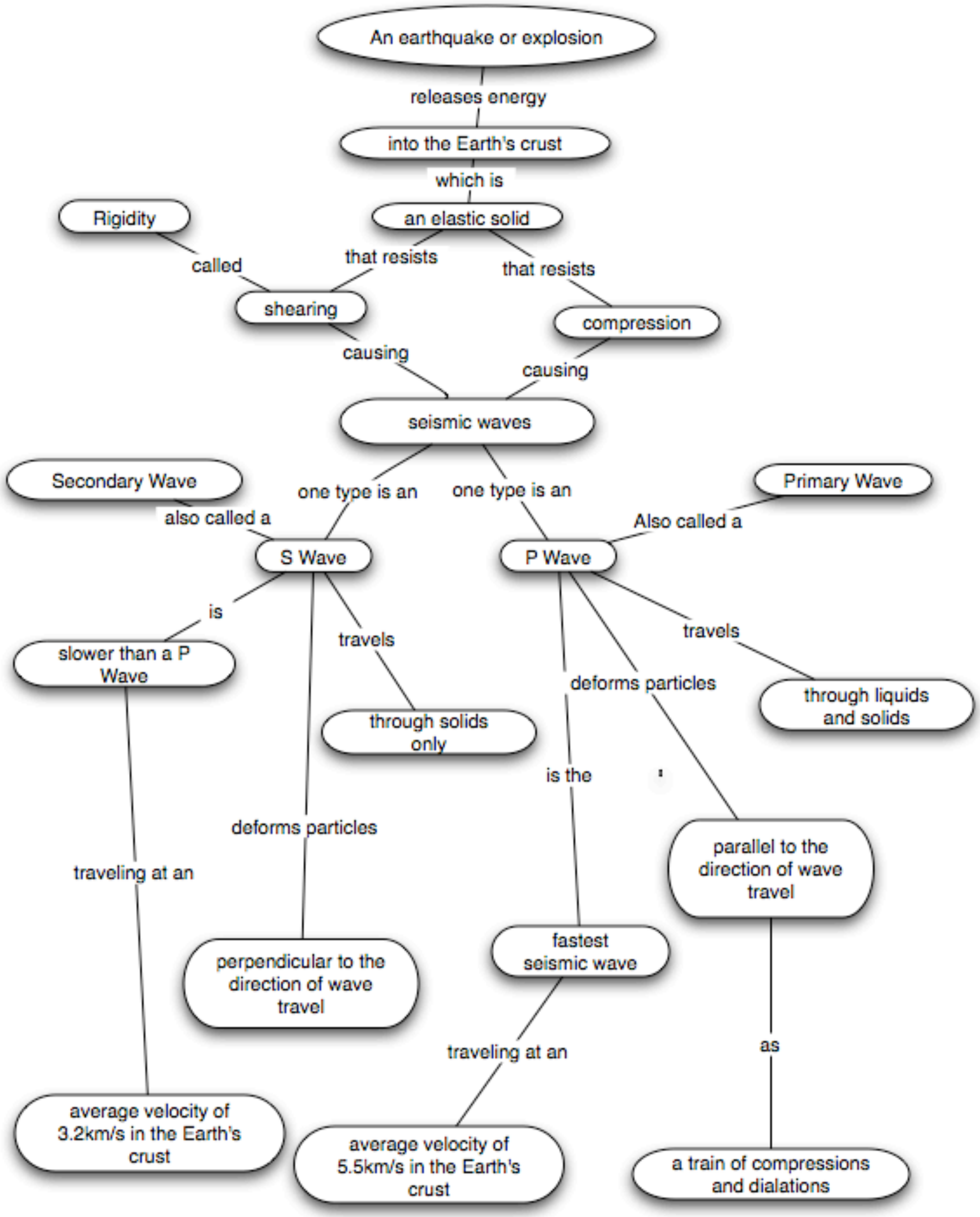


Figure 1—Simplifications of :
A) Cross section of the Earth, **B**) S-wave paths and shadow zone; and **C**) P-wave paths and shadow zone

Earthquake ray paths and arrival times are more complex than illustrated in the animations, because velocity in the Earth does not simply increase with depth. Velocities generally increase downward, according to Snell's Law, bending rays away from the vertical between layers on their downward journey; velocity generally decreases upward in layers, so that rays bend toward the vertical as they travel out of the Earth (See Hot Link above to learn more about why they travel a curving path.) Snell's Law also dictates that rays bend abruptly inward at the mantle/outercore boundary (sharp velocity decrease in the liquid) and outward at the outer core/inner core boundary (sharp velocity increase).

<u>Concept Map</u>



Limitations of the Model

While engaging, this demo is a simplified model of natural phenomena. As such it is especially important to emphasize both the strengths and weaknesses of the model to students. Such an explicit discussion helps students focus on the model as a conceptual representations rather than a concrete copy of reality. Beyond the expected, scale and compositional limitations of the model, an instructor should point out that seismic waves only travel outward all directions; not just in one direction like the line of students. Seismic waves also travel at speeds that are much faster, about 3000x faster, than the waves in the Human Wave model. Next, the instructor should also point out that the particle motion of an S wave can be in any direction that is perpendicular to the direction of wave propagation; not just up and down as shown in the demo. Finally, even though kinetic theory is not the targeted learning outcome of this demonstration, instructors should be specific about pointing out that in this demo the student “particles” are stationary for the purposes of the demo, but actual particles are constantly in motion, even in a solid.

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What's inside of the Earth?

(Black Box Activity)

Earthquakes in NC workshop

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Background:

Humans have explored from the depths of the sea, to the highest mountains on the planet, we have even explored our solar system neighbor, the Moon. We have sent robotic spacecraft to other planets, moons, asteroids, and even comets, yet we have never sent probes, sampled or drilled through the crust of our home planet. The deepest drilling project humans have ever undertaken was Kola Superdeep Borehole at 12.2 km (7.6 miles). Considering that the crust is less than one percent of the distance to the center of the Earth, How do scientists know about the composition and internal structure of the Earth?

Remote Sensing using Seismic Waves.

Seismic waves are vibrations that are caused by earthquakes and large explosions. These waves travel at different velocities through different material. Some waves (p-waves) travel through both solid and liquid material while others (s-waves) only travel through solid material. By carefully measuring and mapping the arrival of seismic waves, scientists can create models of the internal structure of the Earth.

This is a rather abstract concept to explain to middle and high school students. However, since a young age, these students all have used similar techniques to explore the world in which they live.

Materials needed:

4 opaque plastic containers with tight lids (preferably black)

Plaster of Paris, Sand, water

Mix plaster of Paris and fill one container. Fill the others with sand, water and with air (just put lid on the container).

Procedure in class:

Invite students to come to a demonstration table where the four "black-box" containers are placed. Instruct the student to place one hand on the side of the container and to tap the container with the other hand. Do not allow them to look in or lift the container. Have students guess the contents of each container.

Ask students how they were able to identify the composition of the materials in each container. (Most will respond with comments about "Feeling the vibrations.") While fingers do not have the sensitivity of seismometers, most will have sufficient real world experience to correctly identify the contents of each container.

Explain that seismic waves are vibrations in the Earth and that seismometers can record these vibrations. With careful analysis of seismograms, we can model the internal structure and composition of the Earth in the same way they "felt" and correctly identified the composition of the mystery containers.

Journey to the Center of the Earth©

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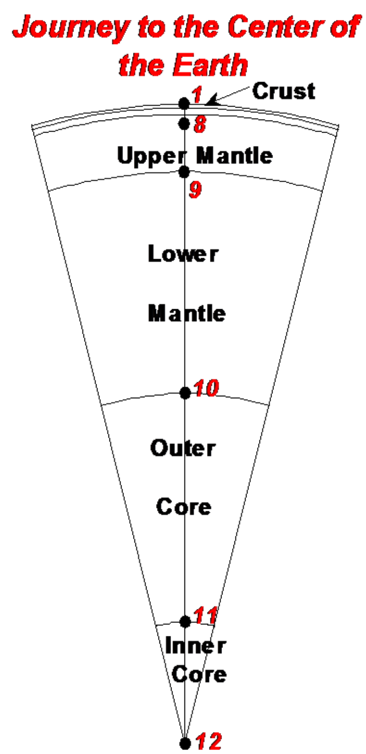
(January 25, 2002; updated April 9, 2004; October 15, 2011)

Explorations in Earth Science



"But in the cause of science men are expected to suffer." (p. 28, *A Journey to the Center of the Earth*, Jules Verne, 1864)

Objectives: This virtual journey to the center of the Earth introduces the traveler to the structure, material properties and conditions within the Earth's interior. The size and scale of the Earth and of the Earth's internal structure are also emphasized because the journey utilizes a scale model of the depths within the Earth. Opportunities for creative writing and connections to literature are also provided through Jules Verne's 1864 science fiction novel, *A Journey to the Center of the Earth*, and the 20th Century Fox 1959 movie adaptation (titled *Journey to the Center of the Earth*) starring James Mason, Pat Boone, Arlene Dahl, and Diane Baker.

Background: In the 1800's there was considerable scientific and popular interest in what was in the interior of the Earth. The details of the internal structure (crust, mantle, outer core, and inner core; and their composition and thicknesses; Figure 1) had not yet been discovered. And, although volcanic eruptions demonstrated that at least part of the interior of the Earth was hot enough to melt rocks, temperatures within the Earth and the existence of radioactivity were unknown. Jules Verne's book, *A Journey to the Center of the Earth* (1864, 272 pages; originally published in France as *Voyage au Centre de la Terre*), capitalized on this interest in the Earth and in adventure with an exciting science fiction story that is still popular today. Verne introduces us to a dedicated, and somewhat eccentric professor, and his nephew through whom the story is told (see selected quotations below), who eventually travel into the Earth's deep interior by entering into an opening in the crater of a volcano in Iceland.

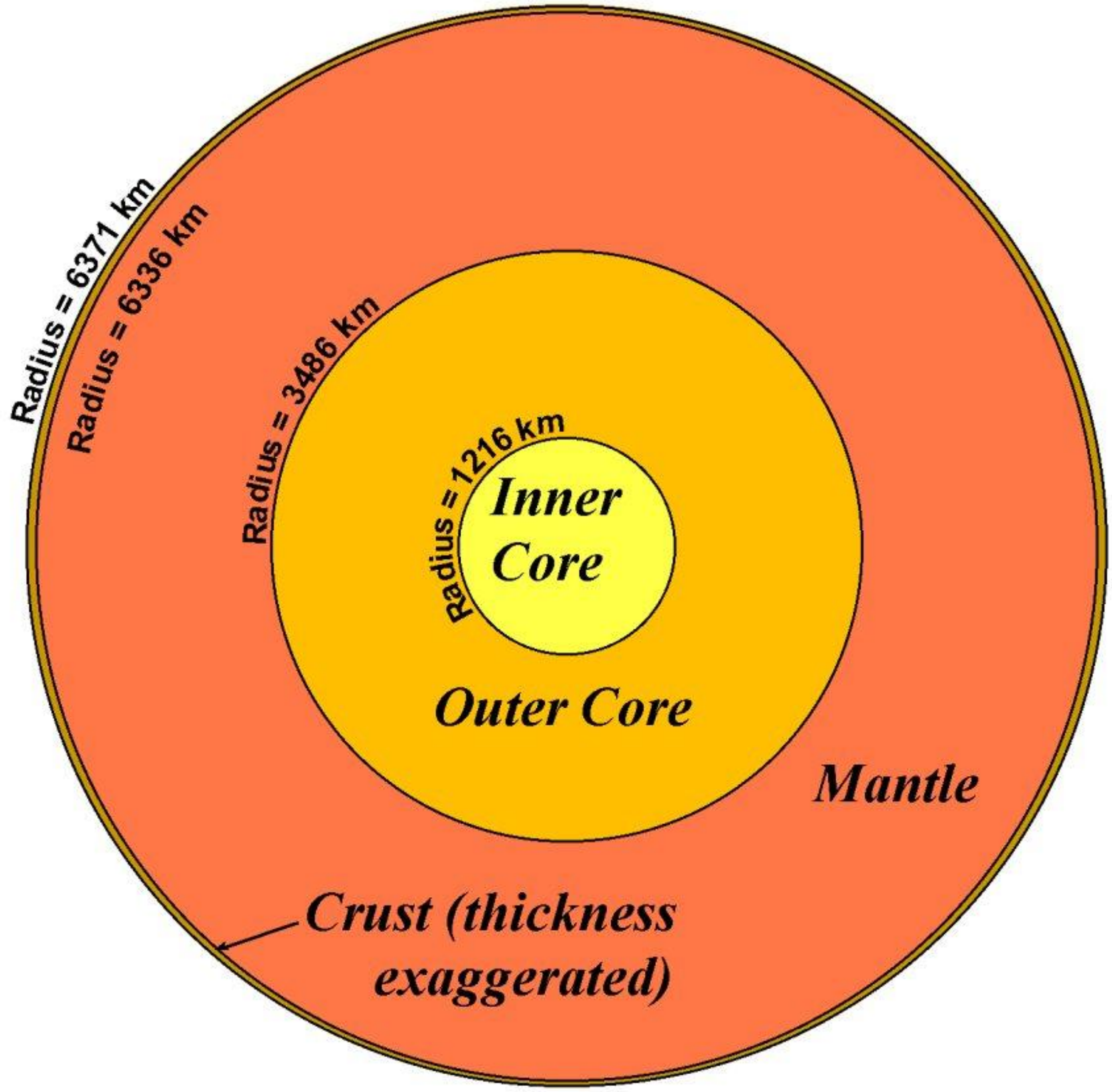


Figure 1. Earth's interior structure. The Earth's crust is made up primarily of silicic (high percentage of Silicon and Oxygen) crystalline (distinct crystals of individual minerals are visible) rocks. The mantle makes up about 82% of the Earth by volume and consists of Iron- and magnesium-rich silicate rocks. The core is mostly iron, with a small percentage of nickel. The outer core is molten and the inner core is solid.

"…and my uncle a professor of philosophy, chemistry, geology, mineralogy, and many other ologies." (p.1, Jules Verne, 1864)

"I loved mineralogy, I loved geology. To me there was nothing like pebbles—and if my uncle had been in a little less of a fury, we should have been the happiest of families." (p.3, Jules Verne, 1864)

"His imagination is a perfect volcano, and to make discoveries in the interest of geology he would sacrifice his life." (p. 14, Jules Verne, 1864)

Verne's novel is science fiction. We know today that such a journey would be impossible. The temperature and pressure conditions within the Earth are so extreme that humans could not survive below a few kilometers depth within the 6371 km radius Earth. Furthermore, we know of no significant openings that would provide access to the deep interior of the planet, and caves or cavities at great depth are nearly impossible based on our knowledge of temperature and pressure within the Earth and the properties of Earth materials. However, Verne's story is an interesting one and it is the inspiration (along with the desire to provide materials for learning about the Earth's interior) for this Earth science educational activity.

By the late 1800's, observations of temperature in mines and drill holes had demonstrated that temperature within the Earth increased with depth, and thus it is possible that the Earth's interior is very hot. Seismographic recordings in the early 1900's were used to identify the Earth's thin (about 5 – 75 km thick) crust (in 1909) and the existence of the core (in 1906). In 1936, Danish seismologist Inge Lehman presented evidence for the existence of a solid inner core. Since then, seismology and other geological and geophysical studies have provided considerably more detailed information about the structure, composition and conditions of the interior of the Earth. These features will be highlighted during our virtual "Journey to the Center of the Earth".

As it is commonly done, we have represented (Figure 1 and Table 1) the Earth as a layered sphere of 6371 km radius. The Earth is actually not quite spherical. Because of the rotation on its axis, the Earth is approximately an ellipsoid with the equatorial radius being about 21 km larger than the polar radius. Also, in detail, the Earth is not exactly spherically symmetric. Lateral as well as vertical variations in composition and rock properties have been recognized from seismological and other geophysical observations. Finally, because of plate tectonics, there are significant differences in shallow Earth structure in continental versus oceanic areas, near plate boundaries, and at different locations on the surface. For these reasons, the depths to the boundaries that we will encounter in our journey would be slightly different if we chose a different location for the start of our journey. The depths, properties and other descriptions listed in the scale model for our journey are reasonable average values for a continental region.

Once one realizes that the interior of the Earth is hot, it is natural to ask, why is it hot? Because the Earth is 4.5 billion years old, it would seem logical that the planet would have cooled by now. The heat within the Earth results primarily from two sources – original heat from the Earth's formation and radiogenic heat (Poirier, 2000). The largest of these sources, radiogenic heat, is mostly produced by three, naturally occurring, radioactive elements, Uranium, Thorium and Potassium. These elements are present in the mantle at concentrations of about 0.015 ppm (parts per million; meaning that only about 15 of every billion atoms in the mantle are Uranium) for

Uranium, 0.080 ppm for Thorium and 0.1% for Potassium (Brown and Mussett, 1981). Spontaneous radioactive decay of these elements releases heat. Although the major radioactive elements are more concentrated (10 to 100 times as abundant) in the Earth's crust, most of the radiogenic heat production comes from the mantle because of the much greater volume. The original heat from formation of the Earth dates from the accretion of the Earth from planetesimals that bombarded the early planet converting gravitational energy into heat.

Modern scientific information about the interior of the Earth comes from a variety of studies including: seismology in which seismic waves from earthquakes and other sources are used to generate images of the interior structure and determine the physical properties of Earth materials; analysis of the Earth's gravity field indicates density variations; high-pressure mineral experiments that are used to infer the composition of deep layers; thermal modeling of temperature measurements in drill holes; modeling of the Earth's magnetic field that is produced by convection currents in the electrically-conductive outer core; and chemical analysis of rock samples (called xenoliths) from deep within the Earth that are brought to the surface in volcanic eruptions. More information about the deep Earth and the methods of study of the Earth's interior can be found in the references listed below. A good starting point is the book by Bolt (1993), the American Scientist article by Wysession (1995) or a chapter on the Earth's interior by Wysession from an introductory geology textbook (see reference list). More advanced readers may wish to refer to Brown and Mussett (1981), Jeanloz (1993), Ahrens (1995), Wysession (1996), Poirier (2000) and Gurnis (2001). For younger readers, examine the children's book by Harris (1999). Much of the information about deep Earth properties and conditions given in Table 1 comes from Ahrens (1995). Information about microbes in the Earth's crust (mentioned in the Narrative, Stop number 3) is from Fredrickson and Onstott (1996).

Procedure and Teaching Strategies: A scale model (either a "classroom" scale or a "playground or hallway" scale; Figures 1 and 2 and Table 1) is used to provide the depths and locations of stops for a virtual journey to the center of the Earth.

Using a meter stick or meter wheel, mark out the locations of the 12 stops in the classroom (1:1,000,000 scale model; 6.37 m long) or playground or hallway (1:100,000 scale model; 63.7 m long). Masking tape placed on the floor or pavement is a convenient method for marking the stops. A felt pen can be used to label the stop number on the strip of masking tape. Folded index cards, labeled with the stop number, can also be used and have the advantage that the numbers can be seen from a distance (looking forward or backward to stops along the journey. Depths and the names of the locations can also be labeled using the masking tape, if desired. Provide each student in the class with a copy of the "Tour Guide" that can be produced as described near the end of this document. Folding the page in "thirds" creates a small brochure that each student can use on the tour and take home to help them remember the information that they learned and their experiences on the Journey to the Center of the Earth.

1. With the class, start at stop number 1 (the Earth's surface) and read the first part of the "Journey to the Center of the Earth Narrative" (below). Proceed to the other stops and read the appropriate section of the narrative at each stop. Be sure to point out the

distance that you've traveled in each move (by looking forward and backward along the model and using the scaled and actual distances from Table 1) and the distance that is remaining to travel to the Earth's center. Answer student (traveler) questions at each stop. The information in Table 1 may be useful for answering questions. Other questions may form the basis of class or individual student research ("let's find out") using the references listed below or library or Internet searches.

2. When back in the classroom, use transparencies (or copies) of Table 1 and Figures 1, 2 and 3 to review with the students the main features of the Earth's interior and the properties and conditions at various depths within the Earth. Note the increases in density, temperature and pressure with depth within the Earth and the abrupt changes in density at the major boundaries between layers. Additional questions can be answered or used to prompt additional study (such as other activities related to the Earth's interior structure or plate tectonics) or research or to provide an assessment of student learning from the activity.

3. As an extension, or to connect to reading, writing and literature study, have the class read Jules Verne's *A Journey to the Center to the Earth* (or selected chapters) or watch the movie (it is about 2 hours long, although one could skip the first approximately 30 minutes; starting as the explorers begin to climb the volcano). Relevant writing assignments for the students could be to write their own brief version of *A Journey to the Center of the Earth* based on the more accurate information about the nature of the Earth's interior; write a review of the book or movie, or write about the inaccuracies and misconceptions that are evident in the book and movie. The accuracies and misconceptions also can provide material for an effective class discussion and assessment of student learning after reading Verne's book or viewing the movie.

4. For younger students, reading *Journey to the Center of the Earth* (Harris, 1999) or *The Magic School Bus Inside the Earth* (Cole, 1987) before or after completing the journey is a useful extension and connection to literature.

5. Related Earth structure activities include Earth's Interior Structure (Braile, 2000) and Three-D Earth Structure Model (Braile and Braile, 2000). A useful and attractive color poster (Earth Anatomy poster) illustrating Earth's interior structure is available from the Wright Center for Science Education, Tufts University. A page size version of the poster can be downloaded from http://www.tufts.edu/as/wright_center/svl/posters/erth.html.

6. Additional extensions are also possible. An interesting assignment is to have each student or pair of students select one stop (depth) along the journey. Have the student or student team learn about the materials and conditions at that depth (some additional reading from the references provided below or from online sources would be necessary) and then draw an illustration that can be used to help describe each stop on the journey. Rock samples, if available (even photographs of rock or mineral samples from a book or from the Internet*), could also be placed at each stop to help illustrate the materials that

make up the Earth's interior. A piece of iron or steel can be used for the Earth's core remembering that it will be liquid iron in the outer core. The student experts from one class, stationed at each stop, could also be the tour guides that would provide information, show their illustration and rock sample, and read the appropriate section of the journey narrative for another class or group of students. The experience of students learning in-depth information about one area of the tour and serving as "experts" can be an excellent "students teaching students" approach to learning. To emphasize the long journey or tour experience in the "Journey to the Center of the Earth" activity, a glass of water, a piece of candy or other refreshments could be served at one of the stops, probably the core/mantle boundary (stop 10) which is a little less than half way along the journey in terms of depth.

7. Connections of this activity to the National Science Education Standards (National Research Council, 1996) are listed in Table 3 below.

* Photographs of appropriate rocks and minerals can be found at several online sources, including: http://www.soes.soton.ac.uk/resources/collection/minerals/ (these photos can be enlarged by clicking on the photo until the photograph is almost full screen size); examples of sedimentary rocks are appropriate for the surface stop, number 1, click on "Sedimentary Rocks" at top of web page; for example, see sample #8, a sandstone; Granite samples from the "Igneous Rocks" link can be used for stops 2, 3, 4, and 5, alternatively, Gneiss samples could be used to represent crustal rocks, particularly for stops 4 and 5 that are deeper in the upper continental crust; Gabbro or Basalt samples, also from the "Igneous Rocks" link can be used to represent lower crustal rocks; a photograph of Olivine, an iron-magnesium silicate that is a common mineral in the Earth's mantle – stops 6 – 10 – can be found in the "Minerals" section of the above web site or at: http://www.musee.ensmp.fr/gm/836.html; for the Earth's core, a photo of an iron-nickel meteorite (http://www-curator.jsc.nasa.gov/outreach1/expmetmys/slideset/IronMet.JPG) is a good representation of the material that forms the core. A selection of photos that are useful for representing typical rocks from the Earth's interior is provided in Table 2 below.

Table 1. ***Journey to the Center of the Earth***

Stop Num.	Depth (km)	Scaled Depth (m) 1:1 million	Scaled Depth (m) 1:100,000	Name or Location	Rock/ Material	Density (g/cm^3)	Pressure (MPa)	Temp. (Deg C)
1	0	0	0	Earth's Surface	Atmosphere Sediments	0.001 1.5	0.1	~10
2	1	0.001 (1 mm)	0.01 (1 cm)	Top of "Basement"	Sed. Rocks Granitic Rk.	2.0 2.6	20	~16
3	3.6	0.0036 (3.6 mm)	0.036 (3.6 cm)	Deepest Mine	Granitic Rock	2.7	100	~50
4	10	0.01 (1 cm)	0.1 (10 cm)	Upper Crust	Granitic Rock	2.7	300	~180
5	12	0.012 (1.2 cm)	0.12 (12 cm)	Deepest Drill Hole	Granitic Rock	2.7	360	~200
6	35	0.035 (3.5 cm)	0.35 (35 cm)	Base of Crust ("Moho")	Mafic Rock Olivine-rich Rk.	3.0 3.3	1100	~600
7	100	0.1 (10 cm)	1	Base of Lithosphere	Olivine-rich Rock	3.4	3200	~1200
8	150	0.15 (15 cm)	1.5	Astheno-sphere	Olivine-rich Rock	3.35	4800	~1300
9	670	0.67 (67 cm)	6.7	Upper Mantle Transition	Fe-Mg Silicate	4.1	23800	~1700
10	2885	2.885	28.85	Core/Mantle Boundary	Fe-Mg Silicate Liquid Iron	5.6 9.9	135800	~3500
11	5155	5.155	51.55	Inner Core/Outer Core Bound.	Liquid Iron Solid Iron	12.2 12.8	329000	~5200
12	6371	6.37	63.7	Center of Earth	Solid Iron	13.1	364000	~5500

Table 1. (cont.) *Journey to the Center of the Earth*

Stop Num.	Description/Comments
1	The Earth's surface is a marked boundary, between the solid or liquid Earth below and the Atmosphere above, with distinct changes in properties. Surface materials on land are usually soil, sediments, sedimentary rocks or weathered crystalline rocks.
2	Beneath surface sedimentary rocks, lies a crystalline "basement" made up of igneous or metamorphic rocks, usually of granitic composition. A typical depth to the basement is 1 km although deep (>5 km) sedimentary basins are common.
3	The deepest depth that humans have explored on land is in a gold mine in South Africa -- almost 3.6 km deep. In the oceans, a special submarine carried explorers to the bottom of the Mariana trench at over 11 km below the Pacific Ocean's surface.
4	Upper layer of continental crust consists of granitic (high % of Silicon and Oxygen) rocks. Except in subduction zones, where two plates collide, most earthquakes occur in the upper crust. Lower crust is more mafic (higher % of Mg and Fe).
5	The deepest drill holes in the Earth are about 12 km deep. Rock samples have been recovered from these depths. The holes have been drilled for scientific study of the crust and to explore for petroleum in deep sedimentary basins.
6	The crust-mantle boundary, or "Moho", separates mafic rocks of the lower crust from Olivine-rich rocks that make up the Earth's mantle. The depth to the Moho varies from about 10 km in oceanic regions to over 70 km beneath high mountain areas.
7	The depth of this boundary is controlled by temperature. It is a gradual rather than an abupt boundary. The lithosphere (tectonic plates) above is relatively cool, rigid and brittle. Lower lithosphere is mantle. Beneath is the "soft" asthenosphere.
8	Partial melting of mantle rocks in this layer produces magma for volcanic eruptions and intrusions. Although a solid, asthenosphere is hot enough to flow in convection currents. Lithosphere/asthenosphere boundary is shallower in hot regions.
9	As pressure increases with depth in mantle, Fe and Mg silicate minerals compress into more dense crystalline forms in the transition zone and below. Mantle is relatively homogeneous chemically and forms ~82% of Earth by volume. Deep earthquakes in subduction zones are found to a depth of about 670 km.
10	Boundary separates liquid iron core from the silicate rock mantle. A transition zone (~200 km thick) exists just above the core-mantle boundary that may represent areas of partially melted mantle (the bottom of mantle plumes) from heat flowing from the outer core, or old lithospheric slabs that have descended to the bottom of the mantle. The core is ~16.5% of the Earth by volume but about 33% of the Earth by mass. No seismic shear waves travel in outer core. Convection currents in the electrically conductive outer core produce Earth's magnetic field.
11	This boundary separates the solid inner core from the liquid iron (and ~10% nickel, sulfur, silicon and oxygen) outer core. Although the radius of the inner core is about 1216 km, the inner core includes only about 0.7% of the volume of the Earth.
12	Earth's center is within the dense, iron inner core. Although the temperature is very high, the pressure is so great (~3.6 million times the pressure at the surface), that the inner core is solid.

Table 1. (cont.) ***Journey to the Center of the Earth*** **Description of Column Headings:**
1. Stop Number -- The stop number for our virtual "Journey to the Center of the Earth", in which we will travel from the Earth's surface to the Earth's center (using a scale model).
2. Depth -- The depth (in the Earth) in kilometers corresponding to each stop in our journey. Many of the depths are approximate and will vary by location.
3. Scaled Depth -- The depth (in meters) for each stop in the 1:1 million scale model. Total depth (surface to center) in the scale model is 6.37 m. "Classroom scale" model.
4. Scaled Depth -- The depth (in meters) for each stop in the 1:100,000 scale model. Total depth (surface to center) in the scale model is 63.7 m. "Playground or hallway scale" model.
5. Name or Location -- Description or name of the location of each stop.
5. Rock/Material -- Rock type, description or composition of the material at each stop. Two entries separated by a line give the rock type or material both above and below a boundary at the corresponding depth.
6. Density -- The approximate density (in grams per cubic centimeter; for comparison, the density of water is 1 g/cm^3) of the material at each stop. Two entries separated by a line give the density of the rock or material above and below the boundary.
7. Pressure -- The approximate pressure (in Mega-Pascals) at each stop (depth in the Earth). One atmosphere of pressure (the pressure at the Earth's surface due to the weight of the atmosphere above us) is about 0.1 MPa (1 Kg/cm^2 or ~14 lbs/in^2). The pressure in the tires of a car (and at about 10 meters depth under water) is about 2 atmospheres or about 0.2 MPa.
8. Temperature -- The approximate temperature in degrees Celsius at each stop (depth in the Earth).
9. Description/Comments -- Description and comments about the material and conditions at each stop in the journey.

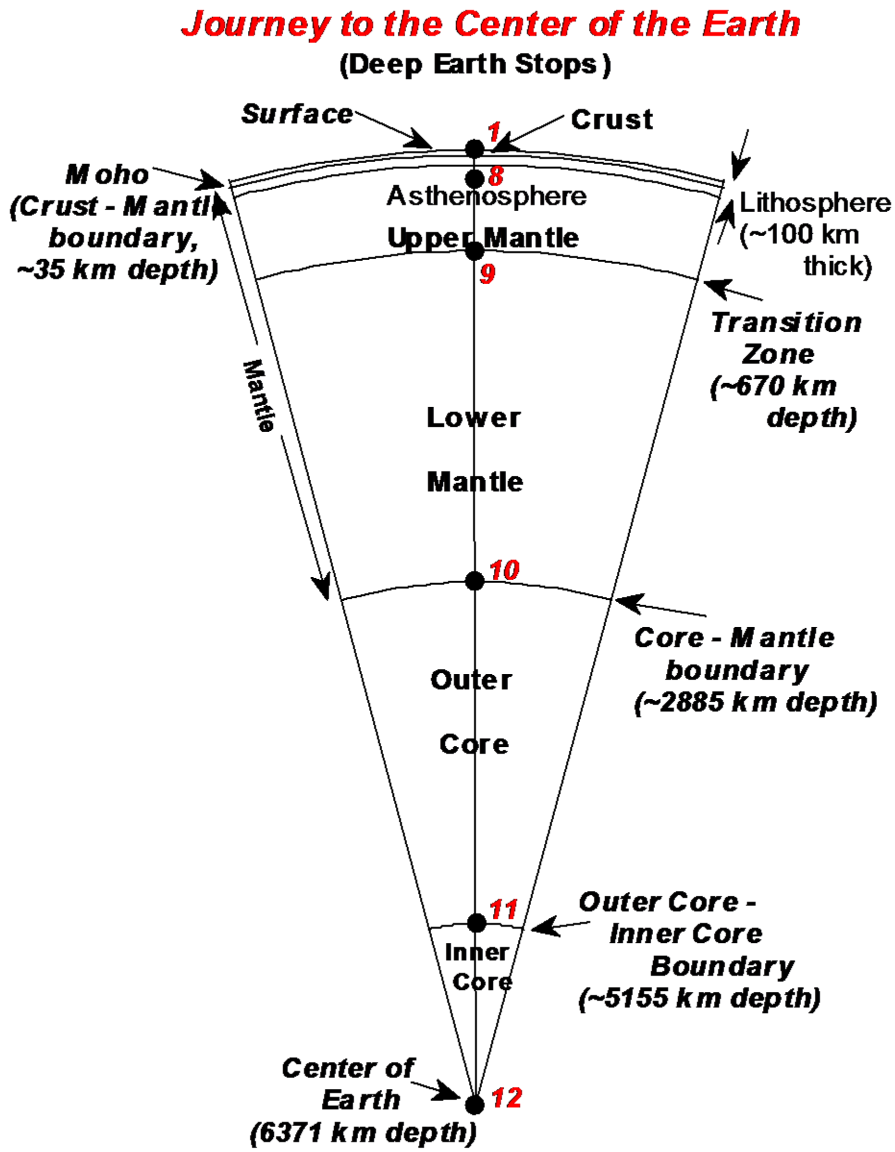


Figure 2. Earth's interior (to scale) showing the depths to the major boundaries between the Earth's layers (spherical shells). The numbered dots indicate the locations of the stops (Table 1) in our virtual journey. A close-up view (Figure 3) of the upper 150 km of the Earth's interior shows the locations of the first eight stops.

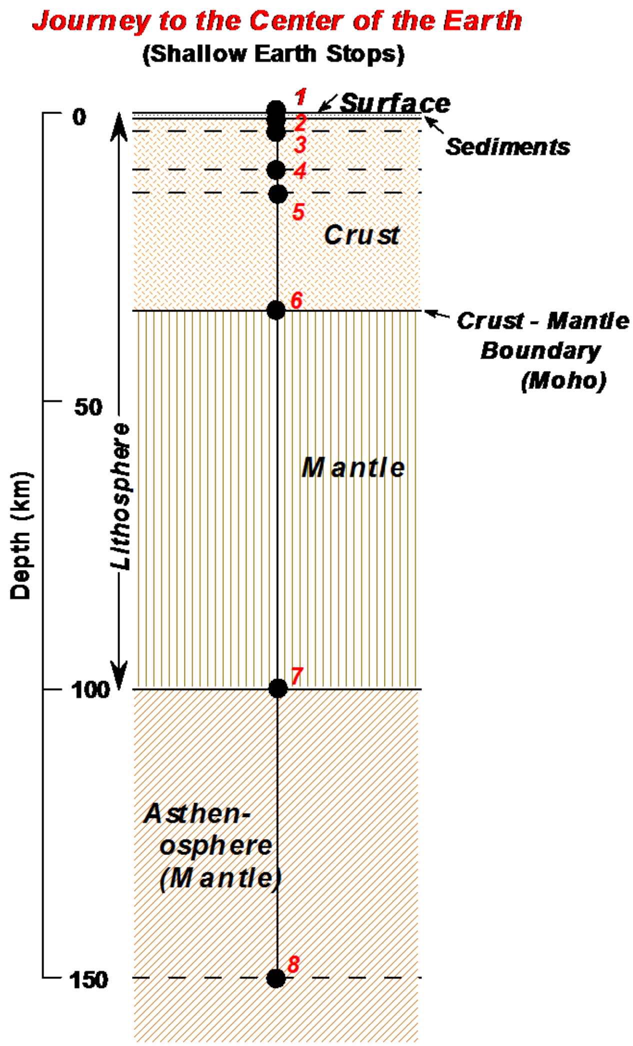


Figure 3. Shallow Earth structure showing the depths to boundaries in the upper 150 km of the Earth. The numbered dots indicate the locations of the first 8 stops (Table 1) in our virtual journey.

Narrative for "Journey to the Center of the Earth":

Attention! Attention! We are ready to start our journey to the center of the Earth. My name is Mrs. Braile[1] and I will be your tour guide today. We are going to learn many fascinating things about the interior of the Earth. Please feel free to ask questions along the way and I will try to answer them. Our journey is long, so I hope that you've had a big breakfast. Also, there are no rest rooms along the way. So, prepare yourself for an exciting journey!

We will begin at the Earth's surface – our familiar home. Except for natural caves, tunnels, mine shafts, and drill holes that extend from the surface to depths of a few kilometers, we know of no large openings that could provide access to the Earth's deep interior. Furthermore, the very high temperature and pressure and the lack of air in the deep Earth create conditions that we could not survive. In addition, it is a long journey – it is 6371 kilometers to the Earth's center. If we were able to walk directly to the Earth's center, it would take about 53 days (at 5 km/hr, 24 hours per day) of walking. And then, we'd have to walk back! Even if there were a very fast elevator that would take us to the Earth's center, the time that our journey would take would only be reduced to about 4 ½ days. If there was a highway to the Earth's center, it would take about 64 hours to drive there at 100 km/hr. Because of these facts, we will be taking a *virtual* journey to the center of the Earth using a scale model. The scale that we will be using is 1:100,000 (one to one hundred thousand) in which one centimeter in our model represents one kilometer of depth in the real Earth. Using this scale, our model of the distance from the Earth's surface to its center is 63.7 meters long.[2] Another way to understand the concept of scale is to realize that we would have to multiply the depths in our scale model by *100,000* in order to produce the actual depths in the Earth! We will begin at the surface and make 12 stops along our journey to the center of the Earth. At each stop, we'll observe the relative distance that we've traveled in our scale model, and learn about the materials and conditions that exist at these locations within the Earth.

Stop Number 1 – Earth's Surface: We're already at our first stop – the Earth's surface. If we began our journey at a different location we would probably find different geological materials at the surface. For example, if we started in a desert region, we might find sand deposits or sandstones at the surface, or if we started in Hawaii, we would most likely begin on volcanic rock. If we began our journey in the middle of the ocean, we would find a layer of ocean water at the surface. Here in Indiana, the near-surface deposits are mostly glacial sediments deposited during the period of 15,000 to about a million years ago.[1] Beneath the glacial sediments are layers of Paleozoic sandstones, shales and limestones that were deposited in a shallow ocean over 300 million years ago. Of course, above us is the Earth's atmosphere, consisting of about 21% oxygen, which we breathe in order to live. Deep in the Earth, we would not have sufficient air to breath. However, that won't be a problem in our virtual journey.

Let's go to our next stop – it isn't far.

Stop Number 2 – Top of the Crystalline Basement: Here we are about 1 km beneath the surface. In continental regions, this is a typical depth to the bottom of the near-surface sediments and sedimentary rocks. The crystalline basement is immediately beneath us. In areas where there are deep sedimentary basins, or in ocean basins, the depth to the basement is significantly greater – up to 10 km or more. In continental areas, the crystalline basement usually consists of igneous and metamorphic rocks of granitic composition. These rocks have interlocking crystals made of minerals that are easily visible and have a composition that includes about 70% Silicon and Oxygen. You may be familiar with the common igneous rock of this type, called granite. The crystalline basement is the top of a layer that makes up most of the Earth's crust.

Our next stop is even deeper in the crust.

Stop Number 3 – Depth of Deepest Mine: This stop is at a depth of 3.6 km (3600 meters) below the surface and is the depth of the deepest mine in the world. It is a gold mine in South Africa, and it's the greatest depth that humans have gone beneath the continents. However, it is not the greatest depth where life exists. Subsurface bacteria (microbes) have been found in drill holes about 3 km beneath the surface and have been shown to be able to survive temperatures as high as 110 degrees Celsius. At this temperature, it is likely that microbes exist in the Earth's crust as deep as about 7 km beneath oceans and about 4-5 km beneath continents.

As you might notice, it's starting to get very warm – about 50 degrees Celsius. You can touch the rocks but don't leave your hand on the rocks for very long or you will burn your hand!

Let's move on.

Stop Number 4 – Upper Crust: We're now deep within the crust at a depth of 10 km. Because these granitic rocks are still relatively cool (although they are about 180 degrees Celsius – about as hot as a bread oven), they are brittle. Except in subduction zones, where two tectonic plates collide, most of the world's earthquakes occur within the upper crust within a few kilometers of our present depth. If we had begun our journey above a deep sedimentary basin, we might find petroleum deposits (oil or natural gas within the pore spaces of sandstones or other porous rocks) at this depth. If we had begun our journey at the surface of the ocean, we would be near the base of the oceanic crust at this depth. The oceanic crust consists of marine sediments overlying rocks of approximately basaltic composition.

If there are no questions, we'll go to the next depth.

Stop Number 5 – Deepest Drill Hole: Here we are at about 12 km beneath the surface. This is the depth of the world's deepest drill hole (in the Kola peninsula of Russia). The pressures and temperatures are so great that it is difficult to build drill bits and drilling equipment that will penetrate these rocks. The rocks are also so compacted that there is almost no space between the crystals or grains that make up the rocks. We definitely couldn't survive here.

Although our next stop is the base of the continental crust, it's only about one half of one percent of the way along our journey! We shouldn't delay.

Stop Number 6 – Base of the Crust: We've reached the base of the crust. It is also called the crust/mantle boundary, or "Moho", after Andrija Mohorovicic the Croatian seismologist who discovered this prominent boundary in 1909. If you'll look back toward the Earth's surface, you'll notice that we really haven't gone very far on our journey to the center of the Earth. The depth to the Moho averages about 35 km beneath continents but is about 10 – 15 km depth beneath oceans. The Moho is an abrupt boundary in composition and properties. Just above the Moho, the lowest layer of the crust consists of more mafic (higher in Magnesium and Iron) rocks than the granitic rocks that we've been traveling through in the upper crust. Below is the mantle – a thick layer that forms about 82% of the Earth's volume; so we'll be traveling through the mantle for a long time. Like the crust, the mantle is also made up of silicate (high percentage of Silicon and Oxygen) rocks. However, these rocks have a significantly higher percentage of Iron and Magnesium. A common material in the mantle is Olivine – an olive green mineral that is commonly found as large crystals in basaltic volcanic rocks such as in Hawaii.

Let's go to the next stop.

Stop Number 7 – Base of the Lithosphere: Here we are at the base of the lithosphere. Notice that the lithosphere consists of the crust *and* the uppermost part of the mantle. This boundary is gradual with depth, not an abrupt "discontinuity". The depth (~50 – 300 km) to the base of the lithosphere is controlled by temperature. Where temperatures in the upper mantle are higher than average, such as beneath mid ocean ridges and in active tectonic zones in continental areas, the lithosphere is thinner. Old, relatively cool lithosphere is much thicker. The lithosphere forms the tectonic plates that separate, collide, and slide past each other to create the Earth's landscape and produce mountain ranges, faults, earthquakes and volcanic eruptions. Below the lithosphere, temperatures are hot enough to partially melt the mantle rocks, forming the asthenosphere – the primary source of magma that erupts from volcanoes on the surface.

The asthenosphere is our next stop.

Stop Number 8 – Asthenosphere: Except beneath areas that are very old (over about one billion years) and have relatively cool upper mantle, at our current depth of 150 km, we would find ourselves within a very hot (about 1300 degrees Celsius) mantle that is partially (probably less than 1-2 percent) molten and flowing. Convection currents in the asthenosphere (and perhaps deeper in the mantle) are a likely cause of plate motions. Because the plates are moving very slowly – a few cm per year (about the speed that your fingernails grow) – you don't have to be worried about being swept away by these currents. Because seismic shear waves travel through the asthenosphere, we classify this part of the mantle (as well as the rest of the mantle) as a solid even though it flows. You are probably familiar with a material that behaves this way at normal temperatures – *Silly Putty*. Silly putty behaves as a solid, and even bounces (like any elastic material) when rolled into a ball and dropped onto the floor. However, it can be stretched, and slowly flows over longer periods of time. It even flows slowly into the form of the plastic egg-shaped container that it is sold in. This behavior, over longer periods of time, is more like a liquid.

Take a close look at the rocks here. You might find diamonds! Diamonds form in the upper mantle from Carbon atoms at high pressure at depths greater than about 150 km. The diamonds can be deposited closer to the surface in "Kimberlite pipes" – narrow vents that are created in brief explosive eruptions.

Well, we've got a long distance to go to our next stop, so we'd better start walking.

Stop Number 9 – Upper Mantle Transition Zone: We're well below the asthenosphere now at about 670 km depth. The pressure is so great at this depth that some of the minerals that form mantle rocks undergo a transformation in their crystal structure that results in a tighter packing of the atoms that make up the mineral. Because of this tighter packing, mantle rocks in the upper mantle transition zone (about 400 – 700 km) become denser with depth even though the chemical composition of the rocks is virtually the same. Therefore, lower mantle rocks are similar in composition to the olivine-rich rocks of the upper mantle but are of higher density. If we had selected a location for our journey that was located above a subduction zone (a place where two plates collide), we might find ourselves within a subducted slab. These parts of lithospheric plates descend, normally at steep angles and at typical plate tectonic velocities – about 2-10 cm/year, from the collision zone at the surface into the mantle. Therefore, these slabs formerly were near the Earth's surface. Because the slabs remain cooler than the surrounding mantle for tens of millions of years, deep earthquakes occur within or at the edges of these slabs. The deepest earthquakes occur at about 670 km depth.

I know that you can feel the intense heat and pressure that are present at this depth, so we need to move on. Our destination, the center of the Earth, is still very distant; in fact, we've only traveled just over 10 percent of our journey. We'll make fewer stops for the rest of the journey.

Stop Number 10 – Core/Mantle Boundary: We're now 2885 km below the surface and at the core/mantle boundary. Let's turn around and look at the Earth's surface to see how far we've gone and to see how much of the Earth is mantle. Let's also look further down in depth to the Earth's center to see how far we have to go. This boundary is the most prominent boundary in the Earth's interior. It is a dramatic boundary in composition, and therefore density, with silicate rocks of the mantle above and dense iron and nickel below. In addition, the mantle above is solid and the outer core below is liquid. The boundary probably varies laterally, and in detail is a transition zone above the liquid outer core that is about 200 kilometers thick. The transition zone has been interpreted to consist of the bottom of mantle plumes where heat flowing from the outer core causes partial melting of the mantle rocks above the core mantle boundary and lithospheric plates (old subducted slabs) that have descended to the bottom of the mantle. The temperature here is about 3500 degrees Celsius, about 2 –3 times hotter than a blast furnace and hot enough to melt iron even under the great pressure that exists at this depth. Because of the dense rocks and high pressure, compressional seismic waves (P-waves) travel at nearly 14 km/s in the mantle just above this boundary. Because the outer core is liquid, the P-wave velocity decreases to about 8 km/s and shear (S) waves cannot propagate in the outer core. Also, the hot, electrically-conductive outer core liquid flows by convection, generating the Earth's magnetic field. It is this magnetic field that aligns the needle on our compass at the Earth's surface.

You may wonder why the temperature is so high in the Earth's interior. Most of the heat comes from radioactive decay of Uranium, Thorium and Potassium atoms that are found in the mantle. These elements are of fairly small concentration in the mantle, so the level of radioactivity is low. However, there are enough radioactive atoms in the mantle to generate significant heat. Some of the Earth's heat was also generated at the time of formation of the planet by bombardment of planetesimals (causing melting) during the accretion of the Earth. Because rocks are not good conductors of heat, the temperature in the interior has remained high.

Well, it's really getting hot, so I'm sure that you're anxious to complete our journey and get back to the Earth's surface. Let's hurry to our next stop.

Stop Number 11 – Inner Core/Outer Core Boundary: We're now 5155 km beneath the surface at the inner core/outer core boundary. The material both above and below us is iron, along with a small percentage of nickel and probably oxygen or sulfur. Above us the iron-nickel outer core is molten. Below us the pressure is so high that, even though it is very hot, the iron-nickel inner core is solid. Although the radius of the inner core is 1216 km (look toward the center of the Earth in our model; that's how far we have to go), the inner core is only 0.7 percent of the Earth by volume.

Let's hurry; only one more stop!

Stop Number 12 – Center of the Earth: Well, we made it. Congratulations, we're at the center of the Earth! It's 6371 km back to the surface. Take a look at how far we traveled from the surface. The temperature is about 5500 degrees Celsius. The pressure is over 3.6 million times the pressure at the Earth's surface. HOWEVER, I MUST WARN YOU TO HOLD ON! Because there is approximately the same amount of Earth all around us (we're in the center of a nearly spherical planet), Earth's gravity here is ZERO. *If* there was an opening here, we would feel weightless! However, the pressure and temperature are very high, so we could not survive. It's a good thing this is a *virtual* journey!

It's now time to go back to the surface. It's been a long journey, so let's go directly back.

Back at Stop Number 1 – Earth's Surface: Thank you for being such a good tour group! I hope that you've enjoyed our Journey to the Center of the Earth and that you've learned some interesting things about the Earth's interior. If you have any additional questions about our journey or about the interior of the Earth, I'd be glad to try to answer them for you.

[1] There are several places in the narrative that can be personalized for your use.
[2] The narrative is written assuming the playground or hallway (1:100,000, or 63.7 m long) scale model (Table 1). If the classroom (1:1,000,000, or 6.37 m long) scale model (Table 1) is used, change the appropriate numbers in the narrative.

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Journey to the Center of the Earth, video, Twentieth Century Fox, 1959, $9.98, available at www.amazon.com. [As would be expected, the movie version of *A Journey to the Center of the Earth* is somewhat different from the story in the book. There are additional characters and the story line has been changed, although the main feature of the story – the journey to the center of the Earth – remains, and the personalities of the main characters are, in general, retained.]

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Acknowledgments: We thank Michael Wysession, Barry Marsh and John and Kathy Taber for providing information or assistance. The development of this activity was partially supported by the National Science Foundation.

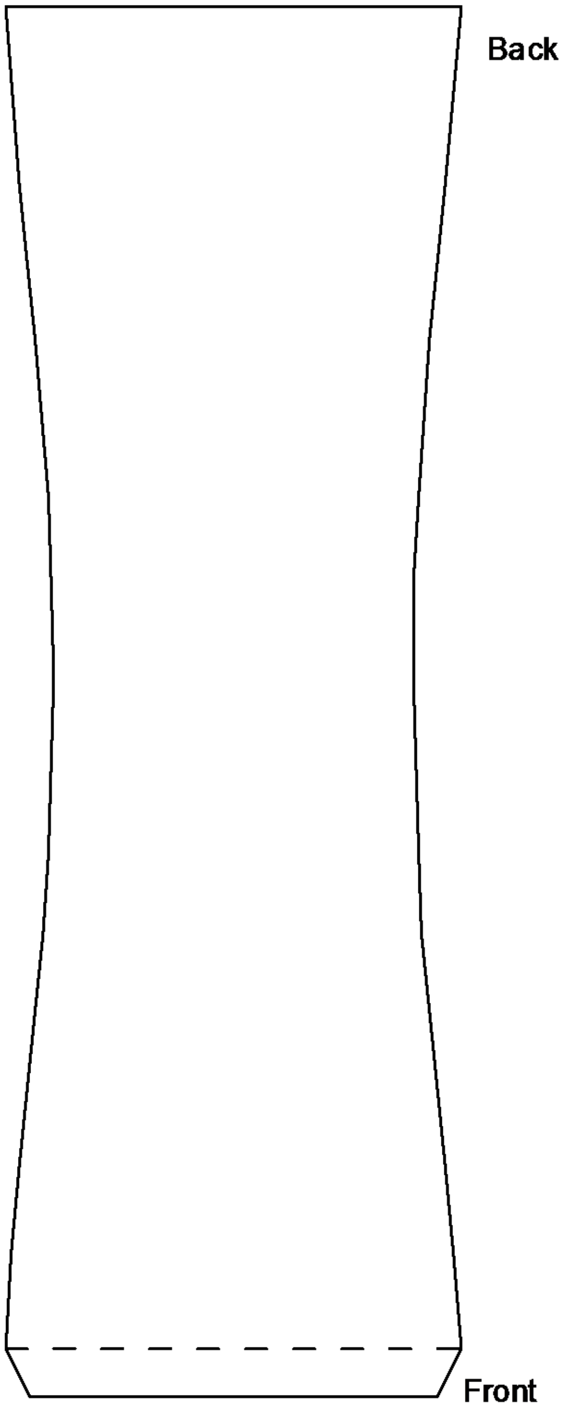
Tour Guide: Provide each student in the class with a copy of the "Tour Guide" that can be produced by printing the next two pages of this document. Trim each image and enlarge each one to fit on one page. Or, print the Tour Guide from the .doc (recommended; margins are better) or .pdf files: guide.doc or guide.pdf. Copy (in color or black and white) on the front and back of a sheet of paper. Make copies for the class. Folding the page in "thirds" creates a small brochure that each student can use on the tour and take home to help them remember the information that they learned and their experiences on the Journey to the Center of the Earth.

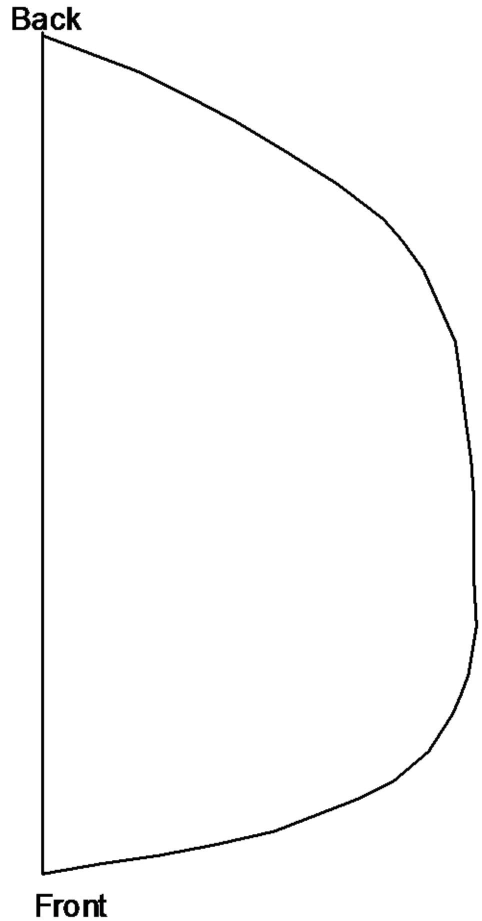
To generate additional interest in the Journey to the Center of the Earth tour, one can make a construction paper "hardhat" (Figures 4 and 5) for the tour leader or for each participant in the tour (the idea of wearing a hat for the tour was provided by John and Kathy Taber).

Labels: A set of labels for the stops along the Journey tour is available in the MS Word document: http://web.ics.purdue.edu/~braile/edumod/journey/labels.doc. The labels include a single page to mark the first five stops. Labels that can be placed at the appropriate scaled distance for Stop 1 and Stops 6-12 are also included. The label for the first five stops is designed for the 1:100,000 scale (63.7 m scale model). The labels should be printed on card stock paper. Labels for Stops 1 and 6-12 can be folded to make a convenient and visible sign.

This document at:
MS Word format: http://web.ics.purdue.edu/~braile/edumod/journey/journey.doc
HTML format: http://web.ics.purdue.edu/~braile/edumod/journey/journey.htm
PDF format: http://web.ics.purdue.edu/~braile/edumod/journey/journey.pdf

http://web.ics.purdue.edu/~braile
braile@purdue.edu





Template for paper hardhat sides. Enlarge 200% and cut out two sides using red, yellow or orange construction paper.

Template for paper hardhat top. Enlarge 200% and cut out one top using red, yellow or orange construction paper. Fold along dotted line to form the "bill" of the hat.

Figure 4. Template for making a "hardhat" (Figure 5) out of construction paper for the Journey to the Center of the Earth tour.

Figure 5. Photograph of completed construction paper "hardhat." The templates for the sides and top are given in Figure 4. The two sides are joined to the top with small pieces of transparent tape in the inside of the hat. The Journey to the Center of the Earth logo (below) is taped to the front of the hat with two-sided tape.

Table 2. Photographs of rock samples that can be used to represent possible rock types for selected stops in the Earth's interior*.

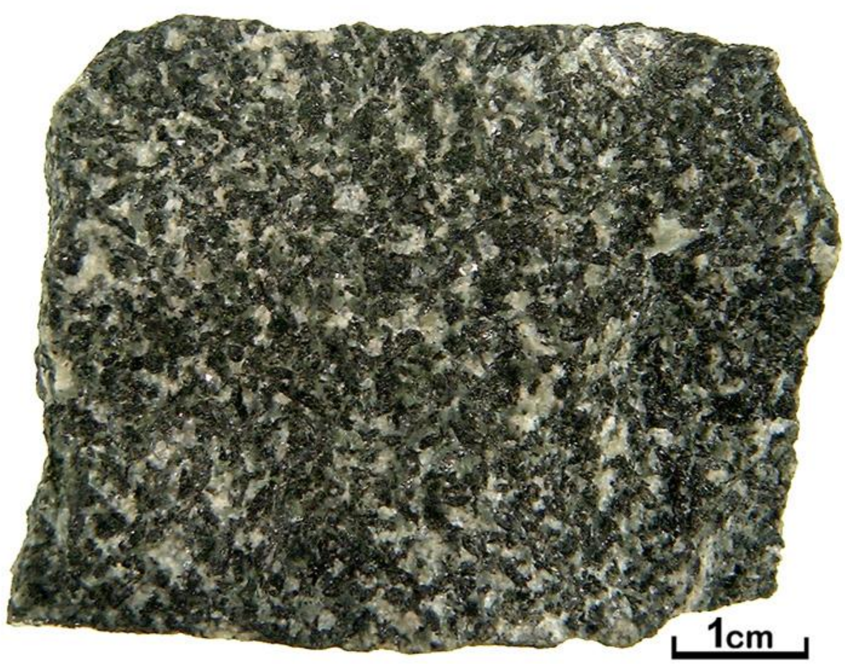
1. Sandstone


Close up of central area

2. Granite

2. Gneiss

3. Gabbro



5. Olivine

6. Iron-Nickel Meteorite

* Photos 1 – 5 courtesy of Barry Marsh, School of Ocean and Earth Science, Southampton Oceanography Center, University of Southampton, UK; used with permission. For more information and additional rock and mineral sample photographs, see http://www.soes.soton.ac.uk/resources/collection/minerals/. Photo 6 is from NASA, http://www-curator.jsc.nasa.gov/outreach1/expmetmys/slideset/Slides35-42.htm.

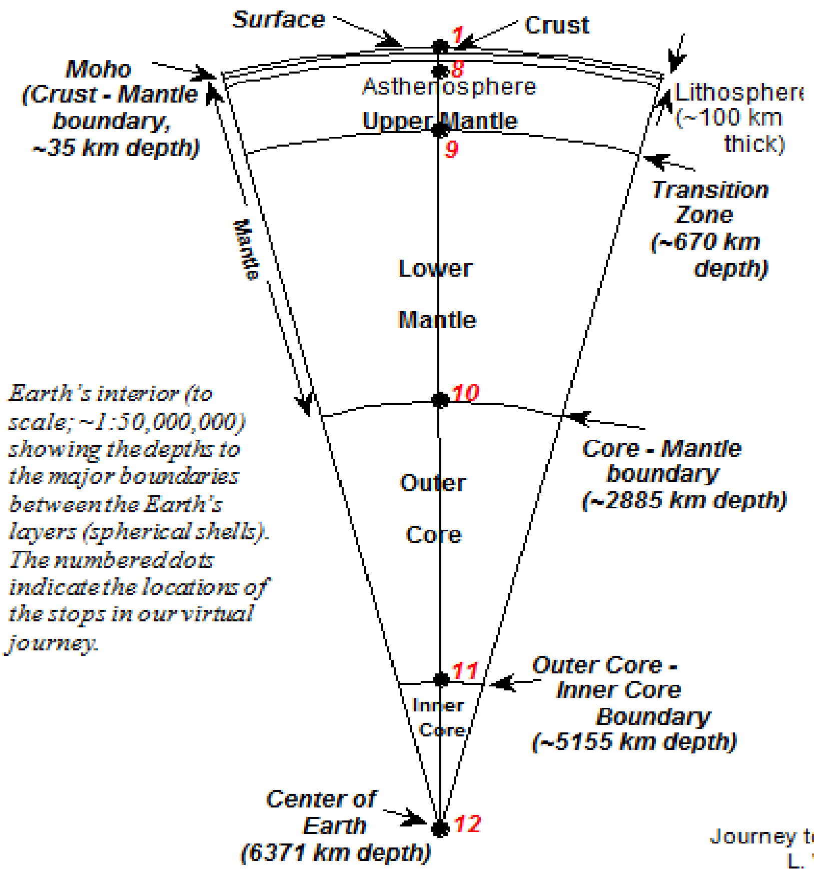
Table 3. "Journey to the Center of the Earth" and the National Science Education Standards (NSES; National Research Council, 1996).

NSES Standard	How standard is addressed in Journey to the Center of the Earth activities (see Procedure and Teaching Strategies section for more details)*
Science Teaching Standards	Activities include inquiry and opportunities for student involvement (A, B) and provide opportunities for ongoing assessment of student learning (C).
Professional Development Standards	The activity provides opportunities and appropriate resource material for teachers to learn about an Earth science topic that is not likely to have been included in their previous educational experiences and that build on their previous knowledge (A, C) and includes suggestions for effective teaching strategies (B).
Assessment Standards	Authentic assessment activities are suggested (C).
Science Content Standards	
- Unifying Concepts and Processes in Science	Activity provides experience with observation, evidence and explanation, and constancy, change and measurement.
- Science as Inquiry	Includes discussion of observations and evidence that result in conclusions about properties and conditions in the Earth's interior.
- Physical Science Standards	Activity explores properties and changes of properties in matter (Grades 5-8, B). Activity explores structure and properties of matter (Grades 9-12, B).
- Earth and Space Science	Activity explores structure of the Earth system (Grades 5-8, D). Activity relates to energy in the Earth system, origin and evolution of the Earth system (Grades 9-12, D).
- History and Nature of Science	Activity includes discussion of history of science, science as a human endeavor (and a connection to literature) (Grades 5-8, G). Activity includes discussion of science as a human endeavor and historical perspectives (and a connection to literature) (Grades 9-12, G).

Return to Braile's Earth Science Education Activities page:

Journey to the Center of the Earth

(Deep Earth Stops)



Earth's interior (to scale; ~1:50,000,000) showing the depths to the major boundaries between the Earth's layers (spherical shells). The numbered dots indicate the locations of the stops in our virtual journey.

"To conclude, I may say that our journey into the interior of the earth created an enormous sensation throughout the civilized world." (Jules Verne, *A Journey to the Center of the Earth*, 1864)

Journey to the Center of the Earth©
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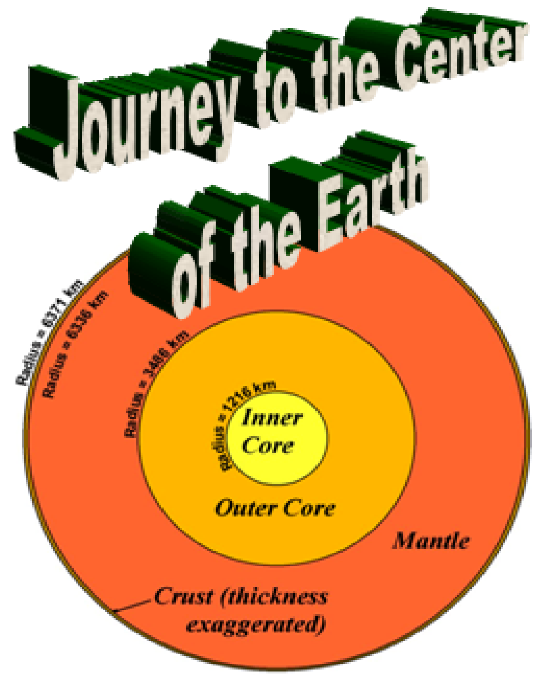
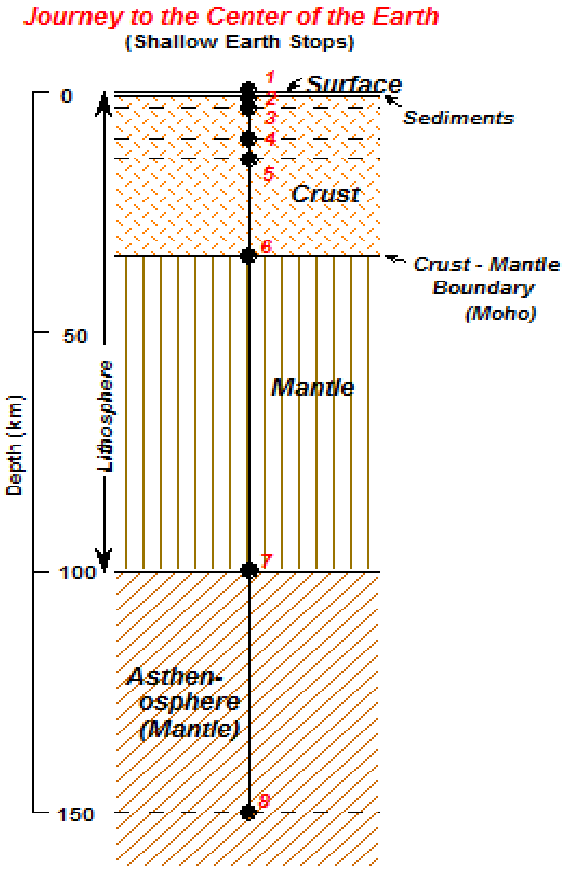
Explorations in Earth Science

A once in a lifetime opportunity!!! Join us in a…



Tour the inside of the planet in under one hour! Discover the structure of the Earth's interior! Experience the conditions deep below the surface!



Shallow Earth structure showing the depths to boundaries in the upper 150 km of the Earth. The numbered dots indicate the locations of the first 8 stops in our virtual journey.

Scheduled stops on our *Journey to the Center of the Earth* (Depth in kilometers in parentheses):

Shallow Earth stops:

1. Earth's surface (0 km) – Atmosphere above, Earth below.
2. Top of crystalline basement (~1 km) – Granitic igneous and metamorphic rocks.
3. Depth of deepest mine (3.6 km) – Temperature is ~50° C here.
4. Upper crust (10 km) – Many earthquakes occur near this depth.
5. Depth of deepest drill hole (12 km) – Drilling used for scientific study and oil exploration.
6. The Moho – crust/mantle boundary (~35 km [beneath continents]) – Crust is a thin shell; mantle is ~82% of Earth.
7. Base of the lithosphere (~100 km) – The Earth's plates (lithosphere) are moving at centimeters per year!
8. The asthenosphere (150 km) – Partially molten mantle and convection currents here.

Deep Earth stops (see diagram on back page):

9. Upper mantle transition zone (~670 km) – Increased pressure transforms minerals to more compact crystal structure and higher density. This depth is only a little more than 10% of our journey.
10. Core/mantle boundary (2885 km) – Solid mantle (iron/magnesium silicate rock) above; liquid iron and nickel below in outer core.
11. Inner core/outer core boundary (5155 km) – Pressure is so great that the iron inner core is solid. Density is about 13 g/cm^3.
12. Center of the Earth (6371 km) – Temperature is ~5500° C, pressure is over 3.6 million times the pressure at the surface.

Thanks for joining us on our Journey to the Center of the Earth! We hope you've enjoyed the tour and will come back again soon!

Measuring the Size of an Earthquake

Seismic waves are the vibrations from earthquakes that travel through the Earth; they are recorded on instruments called seismographs. Seismographs record a zig-zag trace that shows the varying amplitude of ground oscillations beneath the instrument. Sensitive seismographs, which greatly magnify these ground motions, can detect strong earthquakes from sources anywhere in the world. The time, locations, and magnitude of an earthquake can be determined from the data recorded by seismograph stations.

The Richter Scale

The Richter magnitude scale was developed in 1935 by Charles F. Richter of the California Institute of Technology as a mathematical device to compare the size of earthquakes. The magnitude of an earthquake is determined from the logarithm of the amplitude of waves recorded by seismographs. Adjustments are included for the variation in the distance between the various seismographs and the epicenter of the earthquakes. On the Richter Scale, magnitude is expressed in whole numbers and decimal fractions. For example, a magnitude 5.3 might be computed for a moderate earthquake, and a strong earthquake might be rated as magnitude 6.3. Because of the logarithmic basis of the scale, each whole number increase in magnitude represents a tenfold increase in measured amplitude; as an estimate of energy, each whole number step in the magnitude scale corresponds to the release of about 31 times more energy than the amount associated with the preceding whole number value.

At first, the Richter Scale could be applied only to the records from instruments of identical manufacture. Now, instruments are carefully calibrated with respect to each other. Thus, magnitude can be computed from the record of any calibrated seismograph.

Earthquakes with magnitude of about 2.0 or less are usually called microearthquakes; they are not commonly felt by people and are generally recorded only on local seismographs. Events with magnitudes of about 4.5 or greater - there are several thousand such shocks annually - are strong enough to be recorded by sensitive seismographs all over the world. Great earthquakes, such as the 1964 Good Friday earthquake in Alaska, have magnitudes of 8.0 or higher. On the average, one earthquake of such size occurs somewhere in the world each year.

The Richter Scale is not commonly used anymore, as it has been replaced by another scale called the moment magnitude scale which is a more accurate measure of the earthquake size.

Magnitude

Modern seismographic systems precisely amplify and record ground motion (typically at periods of between 0.1 and 100 seconds) as a function of time. This amplification and recording as a function of time is the source of instrumental amplitude and arrival-time data on near and distant earthquakes. Although similar seismographs have existed since the 1890's, it was only in the 1930's that Charles F. Richter, a California seismologist, introduced the concept of earthquake magnitude. His original definition held only for California earthquakes occurring within 600 km of a particular type of seismograph (the Woods-Anderson torsion instrument). His basic idea was quite simple: by knowing the distance from a seismograph to an earthquake and observing the maximum signal amplitude recorded on the seismograph, an empirical quantitative ranking of the earthquake's inherent size or strength could be made. Most California earthquakes occur within the top 16 km of the crust; to a first approximation, corrections for variations in earthquake focal depth were, therefore, unnecessary.

Richter's original magnitude scale (M_L) was then extended to observations of earthquakes of any distance and of focal depths ranging between 0 and 700 km. Because earthquakes excite both body waves, which travel into and through the Earth, and surface waves, which are constrained to follow the natural wave guide of the Earth's uppermost layers, two magnitude scales evolved - the m_b and M_S scales.

The standard body-wave magnitude formula is

$$m_b = \log_{10}(A/T) + Q(D,h) ,$$

where *A* is the amplitude of ground motion (in microns); *T* is the corresponding period (in seconds); and $Q(D,h)$ is a correction factor that is a function of distance, *D* (degrees), between epicenter and station and focal depth, *h* (in kilometers), of the earthquake. The standard surface-wave formula is

$$M_S = \log_{10} (A/T) + 1.66 \log_{10} (D) + 3.30 .$$

There are many variations of these formulas that take into account effects of specific geographic regions, so that the final computed magnitude is reasonably consistent with Richter's original definition of M_L. Negative magnitude values are permissible.

A rough idea of frequency of occurrence of large earthquakes is given by the following table:

M_S	Earthquakes per year
8.5 - 8.9	0.3
8.0 - 8.4	1.1
7.5 - 7.9	3.1
7.0 - 7.4	15
6.5 - 6.9	56
6.0 - 6.4	210

This table is based on data for a recent 47 year period. Perhaps the rates of earthquake occurrence are highly variable and some other 47 year period could give quite different results.

The original m_b scale utilized compressional body P-wave amplitudes with periods of 4-5 s, but recent observations are generally of 1 s-period P waves. The M_S scale has consistently used Rayleigh surface waves in the period range from 18 to 22 s.

When initially developed, these magnitude scales were considered to be equivalent; in other words, earthquakes of all sizes were thought to radiate fixed proportions of energy at different periods. But it turns out that larger earthquakes, which have larger rupture surfaces, systematically radiate more long-period energy. Thus, for very large earthquakes, body-wave magnitudes badly underestimate true earthquake size; the maximum body-wave magnitudes are about 6.5 - 6.8. In fact, the surface-wave magnitudes underestimate the size of very large earthquakes; the maximum observed values are about 8.3 - 8.7. Some investigators have suggested that the 100 s mantle Love waves (a type of surface wave) should be used to estimate magnitude of great earthquakes. However, even this approach ignores the fact that damage to structure is often caused by energy at shorter periods. Thus, modern seismologists are increasingly turning to two separate parameters to describe the physical effects of an earthquake: seismic moment and radiated energy.

Fault Geometry and Seismic Moment, M_O

The orientation of the fault, direction of fault movement, and size of an earthquake can be described by the fault geometry and seismic moment. These parameters are determined from waveform analysis of the seismograms produced by an earthquake. The differing shapes and directions of motion of the waveforms recorded at different distances and azimuths from the earthquake are used to determine the fault geometry, and the wave amplitudes are used to compute moment. The seismic moment is related to fundamental parameters of the faulting process.

$$M_O = \mu S \langle d \rangle ,$$

where μ is the shear strength of the faulted rock, S is the area of the fault, and <d> is the average displacement on the fault. Because fault geometry and observer azimuth are a part of the computation, moment is a more consistent measure of earthquake size than is magnitude, and more importantly, moment does not have an intrinsic upper bound. These factors have led to the definition of a new magnitude scale M_W, based on seismic moment, where

$$M_W = 2/3 \log_{10}(M_O) - 10.7 .$$

The two largest reported moments are 2.5×10^{30} dyn·cm (dyne·centimeters) for the 1960 Chile earthquake (M_S 8.5; M_W 9.6) and 7.5×10^{29} dyn·cm for the 1964 Alaska earthquake (M_S 8.3; M_W 9.2). M_S approaches it maximum value at a moment between 10^{28} and 10^{29} dyn·cm.

Energy, *E*

The amount of energy radiated by an earthquake is a measure of the potential for damage to man-made structures. Theoretically, its computation requires summing the energy flux over a broad suite of frequencies generated by an earthquake as it ruptures a fault. Because of instrumental limitations, most estimates of energy have historically relied on the empirical relationship developed by Beno Gutenberg and Charles Richter:

$$\log_{10} E = 11.8 + 1.5 M_S$$

where energy, *E,* is expressed in ergs. The drawback of this method is that M_S is computed from an bandwidth between approximately 18 to 22 s. It is now known that the energy radiated by an earthquake is concentrated over a different bandwidth and at higher frequencies. With the worldwide deployment of modern digitally recording seismograph with broad bandwidth response, computerized methods are now able to make accurate and explicit estimates of energy on a routine basis for all major earthquakes. A magnitude based on energy radiated by an earthquake, M_e, can now be defined,

$$M_e = 2/3 \log_{10}E - 2.9.$$

For every increase in magnitude by 1 unit, the associated seismic energy increases by about 32 times.

Although M_w and M_e are both magnitudes, they describe different physical properites of the earthquake. M_w, computed from low-frequency seismic data, is a measure of the area ruptured by an earthquake. M_e, computed from high frequency seismic data, is a measure of seismic potential for damage. Consequently, M_w and M_e often do not have the same numerical value.

Intensity

The increase in the degree of surface shaking (intensity) for each unit increase of magnitude of a shallow crustal earthquake is unknown. Intensity is based on an earthquake's local accelerations and how long these persist. Intensity and magnitude thus both depend on many variables that include exactly how rock breaks and how energy travels from an earthquake to a receiver. These factors make it difficult for engineers and others who use earthquake intensity and magnitude data to evaluate the error bounds that may exist for their particular applications.

An example of how local soil conditions can greatly influence local intensity is given by catastrophic damage in Mexico City from the 1985, M_S 8.1 Mexico earthquake centered some 300 km away. Resonances of the soil-filled basin under parts of Mexico City amplified ground motions for periods of 2 seconds by a factor of 75 times. This shaking led to selective damage to buildings 15 - 25 stories high (same resonant period), resulting in losses to buildings of about $4.0 billion and at least 8,000 fatalities.

The occurrence of an earthquake is a complex physical process. When an earthquake occurs, much of the available local stress is used to power the earthquake fracture growth to produce heat rather that to generate seismic waves. Of an earthquake system's total energy, perhaps 10 percent to less that 1 percent is ultimately radiated as seismic energy. So the degree to which an earthquake lowers the Earth's available potential energy is only fractionally observed as radiated seismic energy.

Determining the Depth of an Earthquake

Earthquakes can occur anywhere between the Earth's surface and about 700 kilometers below the surface. For scientific purposes, this earthquake depth range of 0 - 700 km is divided into three zones: shallow, intermediate, and deep.

Shallow earthquakes are between 0 and 70 km deep; intermediate earthquakes, 70 - 300 km deep; and deep earthquakes, 300 - 700 km deep. In general, the term "deep-focus earthquakes" is applied to earthquakes deeper than 70 km. All earthquakes deeper than 70 km are localized within great slabs of shallow lithosphere that are sinking into the Earth's mantle.

The evidence for deep-focus earthquakes was discovered in 1922 by H.H. Turner of Oxford, England. Previously, all earthquakes were considered to have shallow focal depths. The existence of deep-focus earthquakes was confirmed in 1931 from studies of the seismograms of several earthquakes, which in turn led to the construction of travel-time curves for intermediate and deep earthquakes.

The most obvious indication on a seismogram that a large earthquake has a deep focus is the small amplitude, or height, of the recorded surface waves and the uncomplicated character of the P and S waves. Although the surface-wave pattern does generally indicate that an earthquake is either shallow or may have some depth, the most accurate method of determining the focal depth of an earthquake is to read a depth phase recorded on the seismogram. The most characteristic depth phase is pP. This is the P wave that is reflected from the surface of the Earth at a point relatively near the epicenter. At distant seismograph stations, the pP follows the P wave by a time interval that changes slowly with distance but rapidly with depth. This time interval, pP-P (pP minus P), is used to compute depth-of-focus tables. Using the time difference of pP-P as read from the seismogram and the distance between the epicenter and the seismograph station, the depth of the earthquake can be determined from published travel-time curves or depth tables.

Another seismic wave used to determine focal depth is the sP phase - an S wave reflected as a P wave from the Earth's surface at a point near the epicenter. This wave is recorded after the pP by about one-half of the pP-P time interval. The depth of an earthquake can be determined from the sP phase in the same manner as the pP phase by using the appropriate travel-time curves or depth tables for sP.

If the pP and sP waves can be identified on the seismogram, an accurate focal depth can be determined.

by William Spence, Stuart A. Sipkin, and George L. Choy
Earthquakes and Volcanoes
Volume 21, Number 1, 1989

http://earthquake.usgs.gov/learn/topics/measure.php

Pasta Quake—The San Francisco Treat

Demonstration to learn the concept of magnitude & log scale.

Activity is used with permission from Paul Doherty http://www.exo.net/~pauld/index.html. Worksheets by Roger Groom.

Time: 5-10 Minutes

Target Grade Level: 4th grade and up

Content Objective: Students will learn the earthquake magnitude scale by breaking different amounts of spaghetti. Visual scale of the pasta emphasizes the relative differences between magnitudes; each whole step in magnitude

Background

The severity of an earthquake can be expressed in terms of both intensity and magnitude. However, the two terms are quite different, and they are often confused.

Intensity is based on the observed effects of ground shaking on people, buildings, and natural features. It varies from place to place within the disturbed region depending on the location of the observer with respect to the earthquake epicenter.

Magnitude is related to the amount of seismic energy released at the hypocenter of the earthquake. It is based on the amplitude of the earthquake waves recorded on instruments which have a common calibration. The magnitude of an earthquake is thus represented by a single, instrumentally determined value.

To Do and Notice

Hold up one piece of spaghetti. Bend the piece between your hands until it breaks. Notice the work it takes to break the spaghetti. Call this a **5** on the Pasta Magnitude scale.

Hold up a bundle of 30 pieces of spaghetti. Bend the bundle until it breaks. Notice the work it takes to break the bundle. If the pasta magnitude scale were like the earthquake magnitude scale this would be a Pasta Magnitude **6** break.

Hold up 900 pieces of pasta, the remainder of the package. Bend the bundle until it breaks. Notice the work it takes to break the bundle. This is a Pasta Magnitude **7** break.

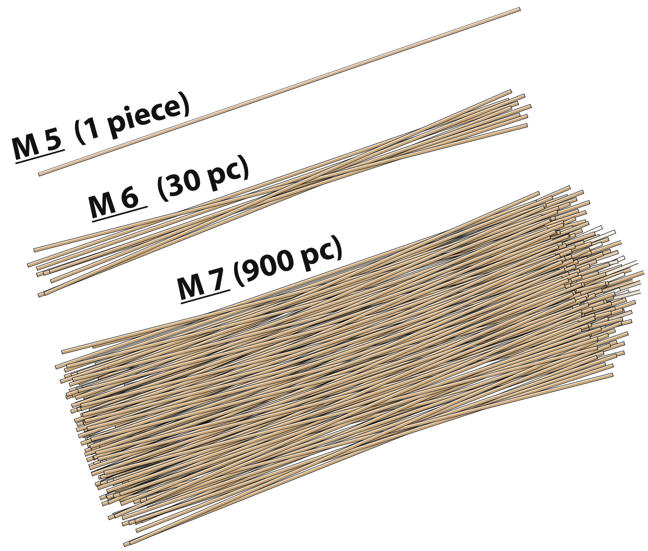
What's Going On?

The magnitude scales for earthquakes are logarithmic scales. In particular for the Richter scale, each increase of 1 unit on the scale, say from 6 to 7, represented an increase of one order of magnitude, i.e. times 10, in the amount of motion recorded on a particular type of seismograph.

The now-common *Moment Magnitude* scale was defined because the Richter scale does not adequately differentiate between the largest earthquakes. The new "moment magnitude" scale is a new technique of using the Richter scale.

Materials

1# package of thin spaghetti <u>or</u>
2# package of regular spaghetti.



In the moment-magnitude scale a magnitude increase of one unit corresponds to a factor of 30 increase in the energy released by the breaking of the fault in an earthquake. That's why we increased the number of spaghetti noodles from 1 to 30 to 900 (900 =30 x 30).

So What?—In order to release the energy of one M 7 earthquake you would have to have 30 M 6 quakes or 900 magnitude 5's. Notice also all the little "quakes" before and after the big-quake break.

In this model, *what does the spaghetti represent?* (The earth, rocks, tectonic plates) What do your hands represent? (Forces, stress, another plate) What does the breaking spaghetti represent? (An earthquake)

Discussion

Describe the energy transfers and transformations—what kind of energy does the spaghetti have when it is bent but not yet broken? (Potential) What kind of energy does the spaghetti have when it is breaking? (Sound, kinetic, heat) Is this an energy transfer or transformation? Explain your choice. (Transformation because the type of energy is changing) If energy cannot be created or destroyed, what happens to the energy released during an earthquake? (It transfers to move buildings, the ground, and the air, and transforms to sound, heat etc)

Are the forces in this investigation balanced or unbalanced? (Unbalanced) How do you know? (The spaghetti bends which is a change in direction (acceleration) and when the spaghetti breaks, it changes speed (acceleration)

Name: _________________________ Per: ____

Today's Date: __________________

Due Date: __________________

Pasta Quake!

There are 3 main ways to measure the magnitude of an earthquake. The **magnitude** is a measurement of earthquake strength based on seismic waves and movement along faults.

First, let's review. What are the 3 types of seismic waves? For each, give a brief description.

1. ____
2. ____
3. ____

What about exactly where earthquakes happen? Draw a diagram in the space below to show the difference between the **epicenter** and the **focus**.

Now, match up the 3 main scales to measure earthquakes with their descriptions:

4.	Richter Scale	Measures the intensity of an earthquake. This is a measure of the strength of ground motion and has a scale of 1 – 12, but in Roman numerals, it's I – XII. One earthquake may have different ratings depending on the damage at different locations.
5.	Moment Magnitude Scale	Measures the size of the seismic waves as shown on a seismograph. Good for small or close-by earthquakes.
6.	Modified Mercalli Scale	Measures the total energy released by an earthquake and can be used for all sizes, near or far. This is usually what is used today.

On the Moment Magnitude Scale, each different number is a measure of the total energy released by an earthquake, and it's about 30 times greater between numbers of the scale (it's actually about 32 times greater). We can demonstrate with spaghetti:

Pasta Magnitude Scale	# of spaghetti pieces broken
3	
4	
5	
6	
7	
8	

Below is a list of some major earthquakes and their Moment Magnitude ratings:

1811-12	New Madrid (Midwestern US)	8.1
1906	San Francisco, California	7.7
1960	Arauco, Chile	9.5
1964	Anchorage, Alaska	9.2
1971	San Fernando, California	6.7
1985	Mexico City, Mexico	8.1
1989	San Francisco, California	7.0
1995	Kobe, Japan	6.9

7. Which earthquake released about 30 times more energy than the Mexico City quake of 1985?

8. Two of the earthquakes released just a little more than 30 times more energy than the 1989 San Francisco quake. They were . . .

9. If the 1964 quake in Alaska released about 30 times more energy than the New Madrid quake, it released about 900 times more energy than which quake?

TEACHER ANSWER KEY—PAGE 1/2

MEASURING EARTHQUAKES

Name:________________________ Class______

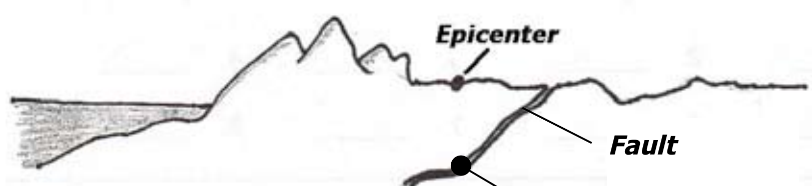
Today's Date: _________ Due Date: __________

There are 3 main ways to measure the magnitude of an earthquake. The **magnitude** is a measurement of earthquake strength based on seismic waves and movement along faults.

First, let's review. What are the 3 types of seismic waves? For each, give a brief description.

1. *P-Waves – primary waves – fastest, move straight through solids and liquids*
2. *S-Waves – secondary/shear waves – second fastest, cannot move through liquids*
3. *L-Waves – surface waves – slowest, travels along surface only*

What about exactly where earthquakes happen? Draw a diagram in the space below to show the difference between the **epicenter** and the **focus**.



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ANSWER KEY—PAGE 2/2

On the Moment Magnitude Scale, each different number is a measure of the total energy released by an earthquake, and it's about 30 times greater between numbers of the scale (it's actually about 32 times greater). We can demonstrate with spaghetti:

Pasta Magnitude Scale	**# of spaghetti pieces broken**
3	***1/30***
4	***1***
5	***30***
6	***900***
7	***27,000***
8	***810,000***

Below is a list of some major earthquakes and their Moment Magnitude ratings:

1811-12	New Madrid (Midwestern US)	8.1
1906	San Francisco, California	7.7
1960	Arauco, Chile	9.5
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   1964 Anchorage, Alaska

8. Two of the earthquakes released just a little more than 30 times more energy than the 1989 San Francisco quake. They were . . .

   1811 New Madrid and 1985 Mexico City

9. If the 1964 quake in Alaska released about 30 times more energy than the New Madrid quake, it released about 900 times more energy than which quake?

   1989 San Francisco

Human Seismometer

Earthquakes in NC workshop

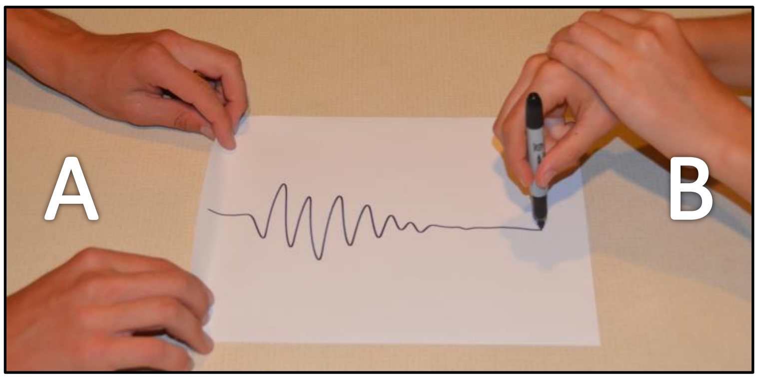
Tim Martin tmartin@greensboroday.org

Background:

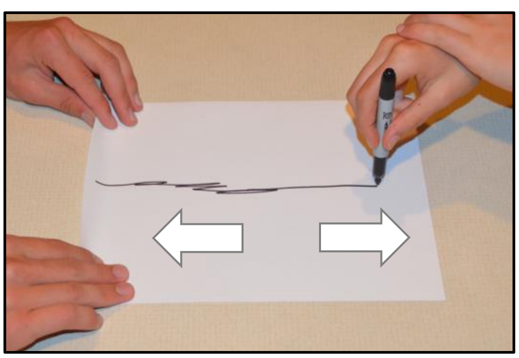
Seismometers are instruments designed to record the Earth's movement during an earthquake. Today most seismic recording stations have sensitive electronic motion and position sensors (accelerometers) as well as associated electronics to process the data. Historical seismometers consist of an isolated mass connected to levers holding a pen that touches a rotating drum. With the drum mount anchored firmly to the ground, the isolated mass will stay relatively stationary when the ground (and drum) move during an earthquake. The pen will record the drum motion as the seismic waves pass.

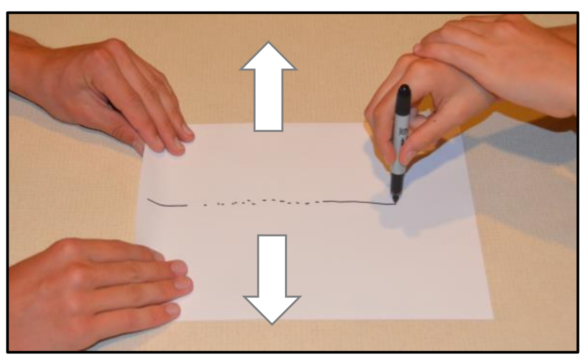
The Demonstration:

The human seismometer is a simple and effective way to introduce students to the basic principles of how an historical seismometer works. Instruct two students to sit on opposite sides of a small table. Instruct student A to slowly pull the paper while student B firmly holds a pen/pencil in contact with the paper. (Student B should avoid all contact with the table while student A should rest his/her hands on the table.) Vigorously shake the table perpendicular to the motion of the paper. The pen will record the table motion and produce a seismogram.



Ask students if this type of device will accurately record all of the motion during an earthquake. For the second demonstration, shake the table parallel with the motion of the paper. Have students discuss or explain that it is necessary to have one seismometer operating on a North/South axis and another on an East/West axis. Reiterate the same question about recording all motion during an earthquake. For the third demonstration shake the table up and down, thus demonstrating the need for a Z axis seismometer to record motion in the up/down direction as well.



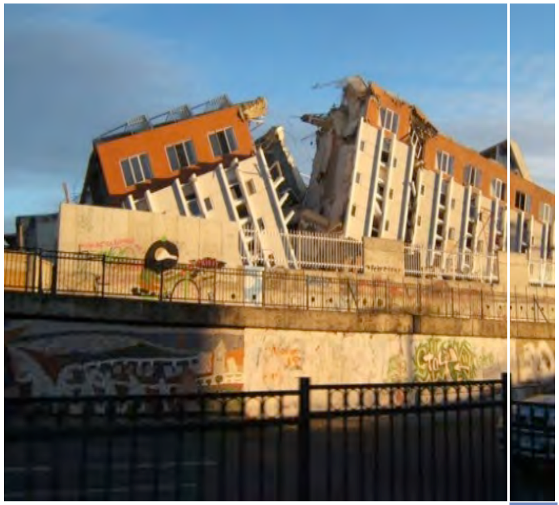


EXTRA: Explore electronic seismometers on a smart phone with free iseismometer (iphone) or Seismo (Android).

NCDENR

GEOLOGICAL SURVEY N.C. 1823

EMERGENCY MANAGEMENT NORTH CAROLINA



What causes an earthquake?

Earthquakes happen because of a sudden slip on a fault in the earth's surface where the rock on one side moves up, down, or sideways relative to the other side. An earthquake is felt as a sudden, rapid shaking on the surface of earth. This shaking can last a few seconds or even few minutes. The motion causes waves that move through the Earth.

Earthquakes are detected with instruments that measure and record the seismic waves (like the Quake-Catcher Network sensors). The record is called a seismogram.

Thank you for participating in the Quake-Catcher Network!

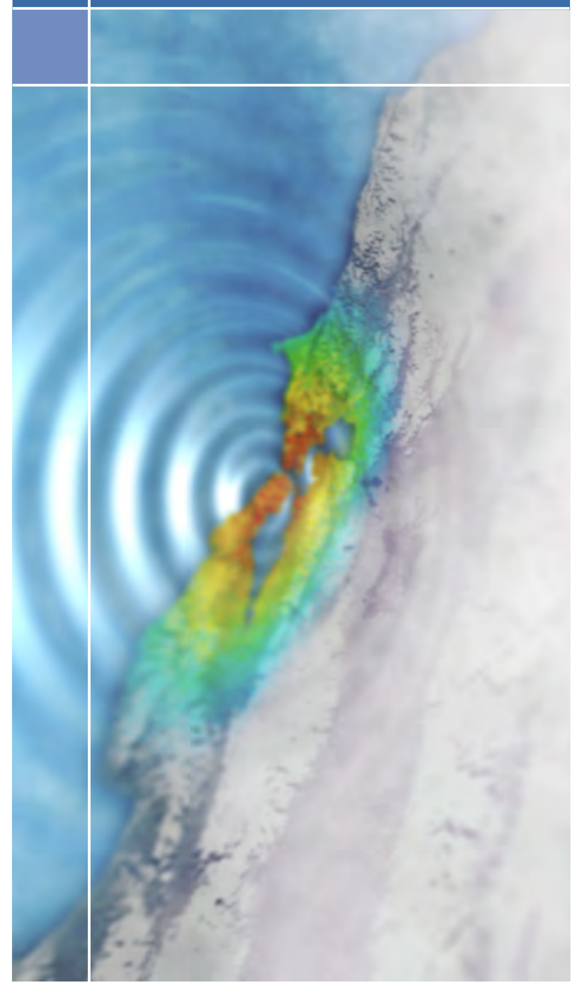
To learn more visit:
http://qcn.stanford.edu

The Quake-Catcher Network

Stanford University – Department of Geophysics
397 Panama Mall Mitchell Building
Stanford, CA 94305-2215

The Quake-Catcher Network

Bringing Seismology to Homesȟ SÃÈÏÏÌóȟ ÁÎÄ -ÕóÅÕ Í Ó



The Quake-Catcher Network
Frequently Asked Questions

What Is the Quake-Catcher Network?

The Quake-Catcher Network (QCN) is a collaborative initiative for developing the world's largest, low-cost strong motion seismic network by utilizing sensors in and attached to Internet-connected computers. With your help QCN can provide better understanding of earthquakes, give early warning to schools, emergency response systems, and others. QCN also provides educational software designed to help teach about earthquake and its hazards.

How do the sensors record an earthquake?

The sensors can measure and record acceleration (ground movement) in three directions. The easiest way to think of these directions is as the 1) up/down, 2) front/back, and 3) side to side bobbing motions of a boat. With these three components of direction it is possible to find the direction of the Earth's movement.

How does QCN know if there is an earthquake?

Sensors connect to the QCN over the Internet. Typically, when the QCN software is running, there isn't much need to transfer the data to our headquarters. Instead, the computer monitors the data locally for new high-energy signals and only sends a single time and a single significance measurement for strong new signals. If our server receives many of these signals all at once, then it is likely that an earthquake is happening.

How does the sensor know which way is North?

The USB sensors come with a mini compass so you can align the sensor with the X direction pointing toward the North Pole (note: magnetic North is not always the same as true North). The better we know the directions of motion, the more precisely we can pinpoint the earthquake.

How can I set my location?

When you install the software on a desktop you can set a permanent location through Google maps in your BOINC account. There is no GPS on the sensors. This location can be updated (in case you move, for example). By setting the location of your computer as precisely as you can, you will help seismologists much more.

Does my computer have to be on all the time?

In order to record as many earthquakes as possible, yes, your computer should be on at all times (or as much as possible). The sensors cannot record earthquakes if the power is off.

Do I have to be connected to the Internet?

Yes! The sensors can only send data to the server when connected to the Internet. The QCN program BOINC runs in the background and sends very small amounts of data over the Internet when the computer is not being used. There isn't much need to transfer data to our headquarters continuously. Instead, the computer monitors the data locally for new high-energy signals and only sends a single time and single significance measurements for strong new signals.

What happens if there is a power outage?

The power sometimes goes out for a short period after earthquakes because of broken power lines. If the power goes out, your desktop will usually power off without a proper shutdown. When you turn your computer back on, the QCN program BOINC will automatically start up.

How do I update my QCN account?

Visit http://qcn.stanford.edu and click on the "My Account" button found under "My QCN/BOINC" tab at the upper right corner. You can change your location by clicking the "My Location" button found under the same tab.

2/2/12

EPIcenter-Quake Catcher Seismic Network Fact Sheet

Background

The Quake Catcher Network (QCN) is a collaborative initiative for developing the world's largest, low-cost strong-motion seismic network by utilizing motion sensors in and attached to internet-connected computers. QCN links volunteer hosted computers into a real-time motion-sensing network. The volunteer computers monitor vibrational sensors called micro electro-mechanical systems (MEMS) accelerometers and digitally transmit "triggers" to QCN's servers whenever strong motions are observed. QCN's servers sift through these signals, and determine which ones represent earthquakes, and which ones represent cultural noise.

QCN will be providing you with a MEMS accelerometer and QCN staff will be installing the sensor and the required compatible software.

QCN Website: http://qcn.stanford.edu

Many of our participants use their QCN computer for other purposes. Several questions have been asked if QCN affects computer performance, uses a lot of memory, and/or uses significant hard disk space. The short answer is that you should see no difference in performance. At any given time QCN uses about 1% of your processor capacity. The Berkeley Open Infrastructure for Network Computing (BOINC) software and QCN sensor driver software will occupy only 2.6 MB of disk space.

BOINC is an open source middleware system for volunteer and grid computing. It was originally developed to support the SETI@home project before it became useful as a platform for other distributed applications in areas as diverse as mathematics, medicine, molecular biology, climatology, geophysics, and astrophysics. The intent of BOINC is to make it possible for researchers to tap into the enormous processing power of personal computers around the world.

If you have a metered data Internet plan (e.g. if you have 5 GB per month), QCN will not impact your data use in any significant way. QCN data is made up of text files which are not data intensive. Under normal circumstances QCN might use 25 MB of your data allotment per month. If you have a 5 GB allotment per month, this amounts to 0.5% of your data allotment.

Requirements

To successfully complete the installation, we will need a few things from your end:

1) A computer (either PC or Apple) which is either located in or can be relocated to a room with as little traffic and disturbance as possible. An ideal spot for this would be either in a basement, a library-type room, or even a very quiet office. This computer should have an available USB port for the sensor to connect to, and should be running and connected to the internet at all times.

2) An IT administrator at your site who can assist with (a) installation of the BOINC software, (b) installation of the driver for the sensor, and (c) set any permissions which will allow that

computer to send seismic data to the QCN server at Stanford. The BOINC software, or the program which uploads the MEMS accelerometer data to the servers, must be able to access the Internet at all times. Because of this, we will require someone on-hand who will be able to troubleshoot any internet-related issues we may encounter.

PC

Operating system

- Windows 2000 SP5 or XP, SP2, or later

Hardware

- Pentium 233 MHz (Recommended: Pentium 500 MHz or greater)
- 64 MB RAM (Recommended: 128 MB RAM or greater)
- 200 MB disk space (maximum)

Permissions

- You must have administrator privileges to install BOINC.

Nvidia Support

- Must have driver version 185.85 or better installed in order to use your GPU.

ATI Support

- You must have driver version 8.12 or better installed in order to use your GPU.

Apple

Operating system

Mac OS X 10.6.0 and later

Hardware

- Apple computer with an Intel x86 or PowerPC G3, G4, or G5 processor
- 28 MB RAM (Recommended: 256 MB RAM or greater)
- 200 MB hard disk space (the BOINC software and sensor driver take up 2.6 MB of hard disk space).

As of 7/2013

Quake-Catcher Network
qcn.stanford.edu

To Purchase a Sensor:
On website, right navigation under Request A Sensor, click on K-12 Program. (http://qcn.stanford.edu/join-qcn/request-a-sensor)

- Sensors are $5 for teachers (up to 3 sensors). For teachers at schools with
  - underserved students, QCN will send one free sensor.
  - Sensors can be purchased with a credit card online or with a check by printing and
  - mailing in the print form.

Two Ways to Use the Sensor
One way is as a teaching tool using the software called QCNLive.
The second way is to use the sensor as a monitoring station using the BONIC software. We recommend that you use your sensors as in both ways. When the sensors are not being handled and used by students, they should be attached to the floor and connected to the monitoring software.

Sensors and software run on Mac and Windows systems.
You may need to download a driver, depending on your computer and sensor.

1. QCN Live
On the website, right navigation, under Download Now, click on QCN Interactive Software. (http://qcn.stanford.edu/join-qcn/download)
Download the appropriate software for your computer. A manual is available at:
http://qcn.stanford.edu/downloads/QCNLive_User_Manual.pdf
This is what you should use with your students. Find activities at:
http://qcn.stanford.edu/learning-center/lessons-and-activities

2. QCN Network Software
To turn your computer and sensor into a seismic monitoring station, you need to attach the sensor to the floor, connect the computer to the internet and use the BONIC software. The software runs in the background, so you won't notice a thing.

Begin with the instructions- http://qcn.stanford.edu/join-qcn/manualsinstructions
These instructions will show you the steps to set up the monitoring station.

Message Boards - http://qcn.stanford.edu/sensor/forum_index.php
There is a QCN community of users. Many questions are answered on the message boards.

Welcome to Virtual Earthquake

Virtual Earthquake is an interactive Web-based activity designed to introduce you to the concepts of how an earthquake **EPICENTER** is located and how the **RICHTER MAGNITUDE** of an earthquake is determined. The *Virtual Earthquake* program is running on a Web Server at California State University at Los Angeles. You can interact with *Virtual Earthquake* using either a Netscape or Internet Explorer Web Browser running on Macs or PCs.

NEW: A completely revised version of Virtual Earthquake can be found HERE. This new Java applet based version is more inquiry-based than the original version and contains tools so instructors can assess student learning.

(After you complete Virtual Earthquake, check out the Geology Labs On-Line home page for the latest information about project activities. Activities about age dating, river discharge and river flooding are available.)

Instructors: here is some important information

Virtual Earthquake will show you the recordings of an earthquake's seismic waves detected by instruments far away from the earthquake. The instrument recording the seismic waves is called a **seismograph** and the recording is a **seismogram**. The point of origin of an earthquake is called its **focus** and the point on the earth's surface directly above the focus is the **epicenter**. You are to locate the epicenter of an earthquake by making simple measurement on three seismograms that will be sent to you by the *Virtual Earthquake* program. Additionally, you will be required to determine the **Richter Magnitude** of that quake from the same recordings. **Richter Magnitude** is an estimate of the amount of energy released during and earthquake.

Upon completion of this activity you will be given the opportunity to receive a personalized **Certificate as a "Virtual Seismologist."**

In order to get this certificate, you must make careful measurements throughout the activity. The actual certificate is much larger than the one displayed above.

This work was supported in part by grants from the U.S. National Science Foundation. All opinions expressed are those of the authors and not necessarily those of the NSF.

Virtual Earthquake

Here is the direct link to the good version: http://www.sciencecourseware.org/eec/earthquake/

The alternate version for slow computers is also good:
http://www.sciencecourseware.com/virtualearthquake/

Excerpt from the webpage:

"The activity, "Earthquake" is a part of the Virtual Courseware in the Earth and Environmental Science project. It is an inquiry-based activity that helps a user learn about the fundamental concepts of how earthquake (seismic) waves are used to locate an earthquake's epicenter and to determine its Richter magnitude. There are two major components to this activity: 1) an experiment about how a Travel Time graph is constructed and used and 2) an Epicenter Location and Richter magnitude activity. These activities require the user to use mqaps and seismograms to carefully record observations and measurements in a journal. At the completion of the Epicenter and Magnitude part, student learning is "assessed" with a quiz. A CERTIFICATE OF COMPLETION AS A VIRTUAL SEISMOLOGIST will be granted once the activity and quiz are successfully completed."

Virtual Courseware for Earth and Environmental Sciences is supported in part by grants from the U.S. National Science Foundation and the California State University System. All opinions expressed herein are those of the authors and not necessarily those of the NSF or the CSU.

jAmaSeis

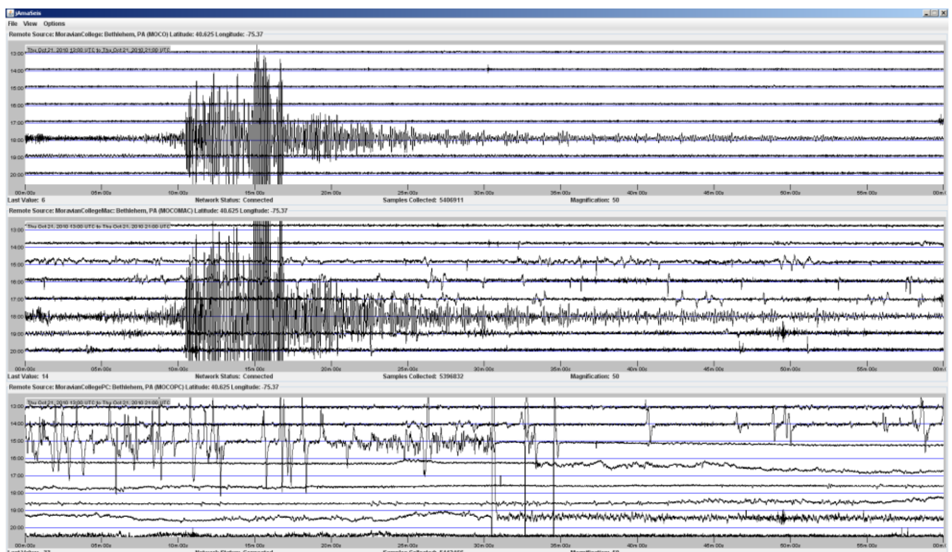
jAmaSeis is now available!

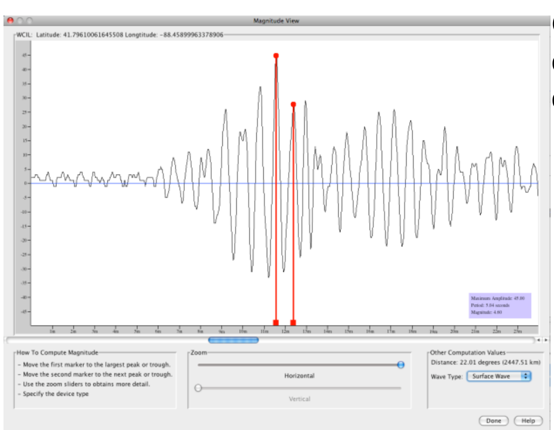
If you don't have your own educational seismometer, you can watch data from a nearby research station!

Download now! http://www.iris.edu/hq/jamaseis

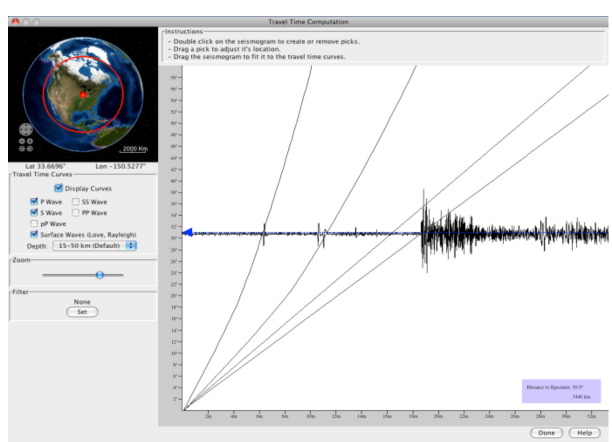
This software replaces the AmaSeis software that is currently being used in the Seismographs in Schools program. It is a significant advancement over what was previously available because it allows users to obtain data in real-time from either a local instrument or from remote stations. As a result, users without an instrument can utilize the software. Additionally, this software includes easy to use analysis tools for users to quickly extract and analyze data from either their recording device or remote data stream.

Stream View- The helicorder screen now has the flexibility to display up to three streams of data simultaneously. These can include a local educational seismometer, a remote educational seismometer over the jAmaseis network (in true real-time), or research-quality seismometers stored at the IRIS Data Management Center (in near real-time).

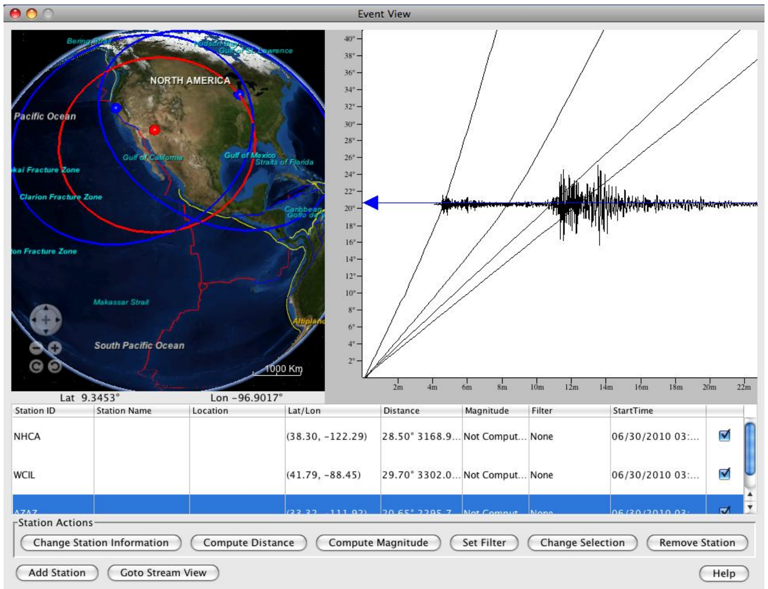




Computing Magnitude- For each stream, an event can be extracted allowing the user to pick amplitudes to calculate either a body wave or surface wave magnitude.



Computing Distance- For each stream, an event can be extracted allowing the user to pick arrivals by double-clicking on the seismogram. A travel time curve is available to align the picks, and as the seismogram is slid along the travel time curve, the numeric values update and a circle with the appropriate radius is shown on the globe.



Event View- All of the analysis for an earthquake comes together in the event view. Multiple traces can be loaded, either from the stream view or from a sac file. All of the individual distance calculations are displayed in both table and map form in addition to the individual magnitude calculations. In this view, a user can make the final determination of the location and size of the earthquake.

IRIS INCORPORATED RESEARCH INSTITUTIONS FOR SEISMOLOGY