Geography Optional 2024

EARTHQUAKE AND INTERIOR OF EARTH

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An **earthquake** is the vibration of Earth produced by the rapid release of energy. Most often earthquakes are caused by slippage along a fault in earth's crust. The energy released radiates in all directions from its source, the **focus**, in the form of waves. These waves are analogous to those produced when a stone is dropped into a calm pond. Just as the impact of the stone sets water waves in motion, an earthquake generates seismic waves that radiate throughout Earth. Even though the energy dissipates rapidly with increasing distance from the focus, sensitive instruments located around the world record the event.

Earthquakes and Faults

The tremendous energy released by atomic explosions or by volcanic eruptions can produce an earthquake, but these events are relatively weak and infrequent. Ample evidence exists that Earth is not a static planet. We know that Earth's crust has been uplifted at times, because we have found numerous ancient wave-cut benches many meters above the level of the highest tides. Other regions exhibit evidence of extensive subsistence. In addition to these vertical displacements, offsets on fence lines, roads, and other structures indicate that horizontal movement is common. These movements are usually associated with large fractures in Earth's crust called **faults**.

Most of the motion along faults can be satisfactorily explained by the plate tectonics theory.

This theory states that large slabs of Earth's crust are in continual slow motion. These mobile plates interact with neighboring plates, straining and deforming the rocks at their edges. In fact, it is along faults associated with plate boundaries that most earthquakes occur. Furthermore, earthquakes are repetitive: as soon as one is over, the continuous motion of the plate's resumes, adding strain to the rocks until they fail again.

Elastic Rebound

The actual mechanism of earthquake generation eluded geologists until H.F. Reid of Johns Hopkins University conducted a study following the great 1906 San Francisco earthquake.

As slippage occurs at the weakest point (the focus), the displacement will exert stress farther along the fault, where additional slippage will occur until most of the built-up strain is released. This slippage allows the deformed rock to "snap back".

The vibrations we know as an earthquake occur as the rock elastically returns to its original shape. The "springing back" of the rock was termed **elastic rebound** by Reid, because the rock behaves elastically, much like a stretched rubber band does when it is released.

In summary, most earthquakes are produced by the rapid release of elastic energy stored in rock that has been subjected to great stress. Once the strength of the rock is exceeded, it suddenly ruptures, causing the vibrations of an earthquake. Earthquakes also occur along existing fault surfaces whenever the frictional forces on the fault surfaces are overcome.

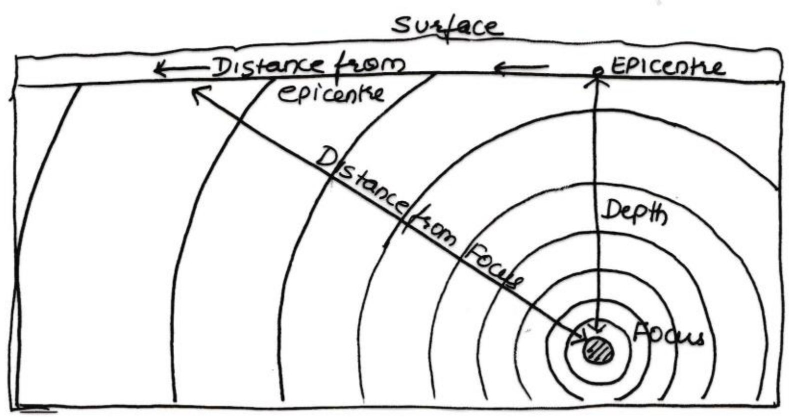
Foreshocks and Aftershocks

The adjustments that follow a major earthquake often generate smaller earthquakes called **aftershocks**. Although these aftershocks are usually much weaker than the main earthquake, they can sometimes destroy already badly weakened structures. A large aftershock of magnitude 5.8 collapsed many structures that had been weakened by the main tremor.

In addition, small earthquakes called **foreshocks** often precede a major earthquake by days or, in some cases, by as much as several years. Monitoring of these foreshocks has been used a means of predicting forthcoming major earthquakes, with mixed success.

Seismology

The study of earthquake waves, **seismology**, dates back to attempts made by the Chinese almost 2000 years ago to determine the direction from which these waves originated. The seismic instrument used by the Chinese was a large hollow jar that probably contained a mass suspended from the top. This suspended mass (similar to a clock pendulum) was connected in some fashion to the jaws of several large dragon figurines that encircled the container. The jaws of each dragon held a metal ball. When earthquake waves reached the instrument, the relative motion between the suspended mass and the jar would dislodge some of the metal balls into the waiting mouths of frogs directly below.



The Chinese were probably aware that the first strong ground motion from an earthquake is directional, and when it is strong enough, all poorly supported items will topple over in the same direction. Apparently the Chinese used this fact plus the position of the dislodged balls to detect the direction to an earthquake's source. However, the complex motion of seismic waves makes it unlikely that the actual direction to an earthquake was determined with any regularity.

In principle at least, modern **seismographs**, instruments that record seismic waves, are not unlike the device used by the early Chinese. Seismographs have a mass freely suspended from a support that is attached to the ground. When the vibration from a distant earthquake reaches the instrument, the inertia of the mass keeps it relatively stationary, while Earth and support move. The movement of Earth in relation to the stationary mass is recorded on a rotating drum or magnetic tape.

Earthquakes cause both vertical and horizontal ground motion; therefore, more than one type of seismograph is needed. The instrument is designed so that the mass is permitted to swing from side-to-side and thus it detects horizontal ground motion. Usually two horizontal seismographs are employed, one oriented north-south and the other placed with, an east-west orientation. Vertical ground motion can be detected if the mass is suspended from a spring.

To detect very weak earthquakes, or a great earthquake that occurred in another part of the world, seismic instruments are typically designed to magnify ground motion. Conversely, some instruments are designed to withstand the violent shaking that occurs very near the earthquake source.

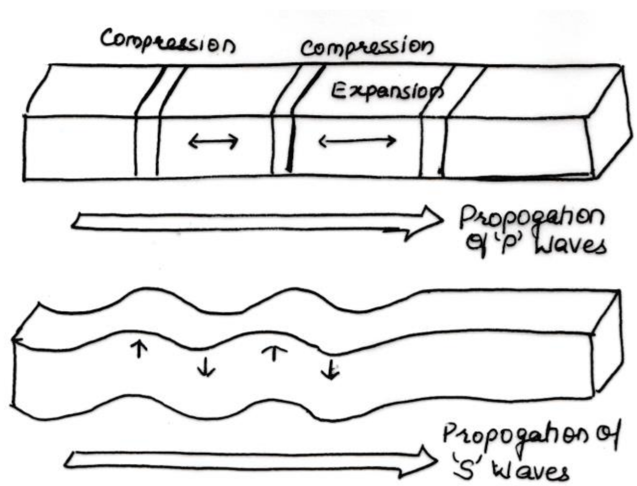
The records obtained from seismographs, called **seismograms**, provide a great deal of information concerning the behavior of seismic waves. Simply stated, seismic waves are elastic energy that radiates out in all directions from the focus. The propagation (transmission) of this energy can be compared to the shaking of gelatin in bowl, which results as some is spooned out. Whereas the gelatin will have one mode of vibration, seismograms reveal that two main groups of seismic waves are generated by the slippage of a rock mass. One of these wave types travels along the outer part of Earth. There are called **surface waves**. Others travel through Earth's interior and are called **body waves**. Body waves are further divided into two types called **primary** of **P waves** and **secondary** or **S waves**.

Body waves are divided into P and S waves by their mode of travel through intervening materials. P waves are "push-pull" waves - they push (compress) and pull (expand) rocks in the direction the wave is traveling like holding someone by the shoulders and shaking that person. This push-pull movement is how P waves move through Earth.

This wave motion is analogous to that generated by human vocal cords as they move air to create sound. Solids, liquids, and gases resist a change in volume when compressed and will elastically spring back once the force is removed. Therefore, P waves, which are compression waves, can travel through all these materials.

On the other hand, S waves "shake" the particles at right angles to their direction of travel which is like fastening one end of a rope and shaking the other end. Unlike P waves, which temporarily change the *volume* of intervening material by alternately compressing and expanding it, S waves temporarily change the *shape* of the material that transmits them. Because fluids (gases and liquids) do not respond elastically to changes in shape, they will not transmit S waves.

The motion of surface waves is somewhat more complex. As surface waves travel along the ground, they cause the ground and anything resting upon it to move, much like ocean swells toss a ship. In addition to their up-and-down motion, surface waves have a side-to-side motion similar to an S wave oriented in a horizontal plane. This latter motion is particularly damaging to the foundations of structures.



By observing a "typical" seismic record, you can see a major difference among these seismic waves: P waves arrive at the recording station first; then S waves; and then surface waves. This is a consequence of their speeds. To illustrate, the velocity of P waves through granite within the crust is about 6 kilometers per second. S waves under the same condition travel at 3.6 kilometers per second. Differences in density and elastic properties of the rock greatly influence the velocities of these waves. Generally, in any solid material, P waves travel about 1.7 times faster than S waves, and surface waves can be expected to travel at 90 percent of the velocity of the S waves. In addition to velocity differences, the height, or more correctly, the amplitude, of these wave types varies. The S waves have slightly greater amplitude than do the P waves, while the surface waves, which cause the greatest destruction, exhibit even greater amplitude.

Because surface waves are confined to a narrow region near the surface and are not spread throughout Earth as P and S waves are, they retain their maximum amplitude longer. Surface waves also have longer periods (time interval between crests); therefore, they are often referred to as **long waves**, or **L waves**.

Seismic waves are useful in determining the location and magnitude of earthquakes. In addition, they provide a tool for probing Earth's interior.

Locating the Source of an Earthquake

Recall that the focus is the place within Earth where earthquake waves originate. The epicenter is the location on the surface directly above the focus.

The difference in velocities of P and S waves provides a method for locating the epicenter. The principle used is analogous to a race between two autos, one faster that the other. The P wave always wins the race, arriving ahead of the S wave. But the greater the length of the race, the greater will be the difference in the arrival times at the finish line (the seismic station). Therefore, the greater the interval measured on a seismogram between the arrival of the first P wave and the first S wave, the greater the distance to the earthquake source.

A system for locating earthquake epicenters was developed by using seismograms from earthquakes whose epicenters could be easily pinpointed from physical evidence. From these seismograms, travel-time graphs were constructed. The first travel-time graphs were greatly improved when seismograms became available from nuclear explosions, because the precise location and time of detonation were known.

Using the sample seismogram and the travel-time curves, we can determine the distance separating the recording station from the earthquake in two steps: (1) determine the time interval between the arrival of the first P wave and the first S wave, and (2) find on the travel-time graph the equivalent time spread between the P and S wave curves. From this information, we can determine that this earthquake occurred 3800 kilometers (2350 miles) from the recording instrument.

The epicenter could be in any direction from the seismic station. The precise location can be found when the distance is known from three or more different seismic stations. On a globe, we draw a circle around each seismic station. Each circle represents the epicenter distance for each station. The point where the three circles intersect is the epicenter of the quake. This method is called *triangulation*.

Earthquake Belts About 95 percent of the energy released by earthquakes originates in a few relatively narrow zones that wind around the globe. The greatest energy is released along a path around the outer edge of the Pacific Ocean know as the *circum-Pacific belt*. Included in this zone are regions of great seismic activity such as Japan, the Philippines, Chile, and numerous volcanic island chains, as exemplified by the Aleutian Islands.

Another major concentration of strong seismic activity runs through the mountainous regions that flank the Mediterranean Sea and continues through Iran and on past the Himalayan complex and another continuous belt extends for thousands of kilometers through the world's oceans. This zone coincides with the oceanic ridge system, which is an area of frequent but low-intensity seismic activity.

Earthquake Depths

Evidence from seismic records reveals that earthquakes originate at depths ranging from 5 to nearly 700 kilometers. In a somewhat arbitrary fashion earthquake foci have been classified by their depth of occurrence. Those with points of origin within 70 kilometers of the surface are referred to as *shallow*, while those generated between 70 and 300 km are considered intermediate, and those with a focus greater than 300 km are classified as *deep*. About 90 percent of all earthquakes occur at depths of less than 100 km, and nearly all very damaging earthquakes appear to originate at shallow depths.

When earthquake data were plotted according to geographic location and depth, several interesting observations were noted. Rather than a random mixture of shallow and deep earthquakes, some very definite distribution patterns emerged. Earthquakes generated along the oceanic ridge system always have a shallow focus and none are very strong. Further, it was noted that almost all deep-focus earthquakes occurred in the circum-Pacific belt, particularly in regions situated landward of deep-ocean trenches.

In a study conducted in the southwestern Pacific near the Tonga trench, it was discovered that foci depths increased with distance from the trench. These seismic regions, called **Wadati-Benioff zones** after the two scientists who were the first to extensively study them, are oriented about 35 to nearly 90 degrees to the surface.

Earthquake Intensity and Magnitude

Until a century ago, earthquake size and strength were described subjectively, making accurate classification of earthquake intensity difficult. Then, in 1902, Giuseppe Mercalli developed a fairly reliable intensity scale based on damage to various types of structures. The U.S. Coast and Geodetic Survey uses a modification of this scale today.

The **Mercalli intensity scale** assesses the damage from a quake at a specific location. The earthquake intensity depends not only on the strength of the earthquake but also on other factors, such as distance from the epicenter, the nature of surface materials, and building design.

Today, earthquakes are ranked according to their **magnitude**, a measure of the amount of energy released during the event. Ideally, the magnitude of an earthquake can be determined from the amount of material that slides along the fault and the distance it is displaced. However, even in an ideal setting such as that of the 1906 San Francisco earthquake, where the fault trace is visible and displacement can be measured from physical evidence, this method can provide only a crude estimate of the forces involved. In most earthquakes, the fault does not penetrate the surface; therefore, the amount of displacement cannot be measured directly.

In 1935, Charles Richter of the California Institute of Technology attempted to rank the earthquakes of southern California into groups of large, medium, and small magnitude. The system he developed determines earthquake magnitudes from the deflections recorded on seismograms.

Today a refined **Richter scale** is used worldwide to describe earthquake magnitude. Richter magnitude is determined by measuring the amplitude of the largest wave recorded on the seismogram. For seismic stations worldwide to obtain the same magnitude for a given earthquake, adjustments must be made for the weakening of the seismic waves as they move from the focus, and for the sensitivity of the recording instrument.

Modified Mercalli Scale		Moment Magnitude Scale
I	Detected only by sensitive instruments	1.5
II	Felt by few persons at rest, especially on upper floors; delicately suspended objects may swing	2
III	Felt noticeably indoors, but not always recognized as earthquake; standing autos rock slightly, vibration like passing truck	2.5
IV	Felt indoors by many, outdoors by few, at night some may awaken; dishes, windows, doors disturbed; motor cars rock noticeably	3
V	Felt by most people; some breakage of dishes, windows, and plaster; disturbance of tall objects	3.5 4
VI	Felt by all, many frightened and run outdoors; falling plaster and chimneys, damage small	4.5
VII	Everybody runs outdoors; damage to buildings varies depending on quality of construction; noticed by drivers of automobiles	5
VIII	Panel walls thrown out of frames; fall of walls, monuments, chimneys; sand and mud ejected; drivers of autos disturbed	5.5
IX	Buildings shifted off foundations, cracked, thrown out of plumb; ground cracked; underground pipes broken	6
X	Most masonry and frame structures destroyed; ground cracked, rails bent, landslides	6.5 7
XI	Few structures remain standing; bridges destroyed, fissures in ground, pipes broken, landslides, rails bent	7.5
XII	Damage total; waves seen on ground surface, lines of sight and level distorted, objects thrown up into air	8

Richter established 100 kilometers as the standard distance and the Wood-Anderson instrument as the standard recording device.

Earthquakes vary enormously in strength, and great earthquakes produce traces having wave amplitudes that are thousands of times larger than those generated by weak tremors. To accommodate this wide variation, Richter could not use a linear scale, but instead used a *logarithmic* scale to express magnitude. On this scale a *tenfold* increase in wave amplitude corresponds to an increase of 1 on the magnitude scale. Thus, the amplitude of the largest surface wave for a 5-magnitude earthquake is 10 times greater than the wave amplitude produced by an earthquake having a magnitude of 4.

More important, each unit of Richter magnitude equates to roughly a 32-*fold energy increase*. Thus, an earthquake with a magnitude of 6.5 releases 32 times more energy than one with a magnitude of 5.5, and roughly 1000 times more energy than a 4.5-magnitude quake.

A major earthquake with a magnitude of 8.5 releases millions of times more energy than the smallest earthquakes felt by humans. This should dispel the notion that a moderate earthquake decreases the chances for the occurrence of a major quake in the same region. Thousands of moderate tremors would be needed to release the vast amount of energy equal to one “great” earthquake.

Destruction from Seismic Vibrations

The amount of structural damage attributable to the vibrations depends on several factors, including (1) the intensity and (2) duration of the vibrations; (3) the nature of the material upon which the structure rests; and (4) the design of the structure.

Amplification of Seismic Waves

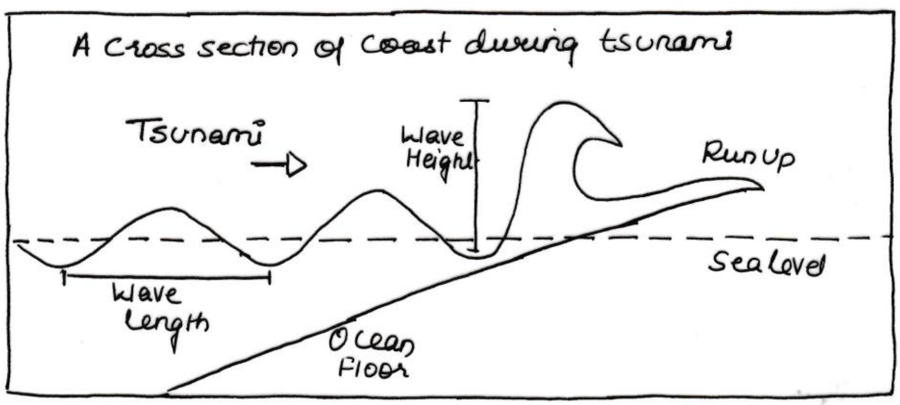
Although the region within 20 to 50 kilometers of the epicenter will experience about the same intensity of ground shaking, the destruction varies considerably within this area. This difference is mainly attributable to the nature of the ground on which the structures are built. Soft sediments, for example, generally amplify the vibrations more than solid bedrock. Thus, the buildings located in Anchorage, which were situated on unconsolidated sediments, experienced heavy structural damage. By contrast, most of the town of Whittier, although much nearer the epicenter, rests on a firm foundation of granite and hence suffered much less damage. However, Whittier was damaged by a seismic sea wave.

Liquefaction

In areas where unconsolidated materials are saturated with water, earthquake vibrations can generate a phenomenon known as **liquefaction**. Under these conditions, what had been a stable soil turns into a mobile fluid that is not capable of supporting buildings or other structures. As a result, underground objects such as storage tanks and sewer lines may literally float toward the surface of their newly liquefied environment. Buildings and other structures may settle and collapse.

Seiches

The effects of great earthquake may be felt thousands of kilometers from their source. Ground motion may generate *seiches*, the rhythmic sloshing of water in lakes, reservoirs, and enclosed basins such as the Gulf of Mexico. Seiches can be particularly dangerous when they occur in reservoirs retained by earthen dams. These waves have been known to slosh over reservoir walls and weaken the structure, thereby endangering the lives of those downstream.



Tsunami

Most deaths associated with the 1964 Alaskan quake were caused by **seismic sea waves**, or **tsunami**. These destructive waves often are called "tidal waves" by the media. However, this name is inappropriate, for these waves are generated by earthquakes, not the tidal effect of the Moon or Sun.

Most tsunamis result from vertical displacement of the ocean floor during an earthquake. Once created a tsunami resembles the ripples formed when a pebble is dropped into a pond. In contrast to ripples, tsunami advance across the ocean at amazing speeds between 500 and 950 kilometers per hour. Despite this striking characteristic, a tsunami in the open ocean can pass undetected because its height is usually less than 1 meter and the distance between wave crests is great, ranging from 100 to 700 kilometers.

However, upon entering shallower coastal water, these destructive waves are slowed down and the water begins to pile up to heights that occasionally exceed 30 meters. As the crest of a tsunami approaches the shore, it appears as a rapid rise in sea level with a turbulent and chaotic surface. Tsunami can be very destructive.

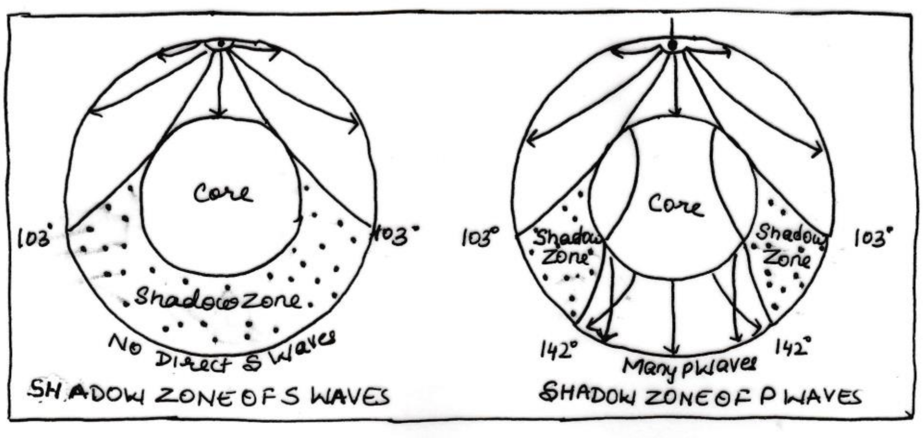
Usually the first warning of an approaching tsunami is a relatively rapid withdrawal of water from beaches. Coastal residents have learned to heed this warning and move to higher ground, for about 5 to 30 minutes later, the retreat of water is followed by a surge capable of extending hundreds of meters inland. In a successive fashion, each surge is followed by rapid oceanward retreat of the water.

Tsunamis are able to traverse large stretches of the ocean before their energy is totally dissipated.

Landslides and Ground Subsidence and Fire can be other effects of earthquake.

Seismic waves and earth's interior

Earth's interior lies just below us: however, direct access to it remains very limited. Wells drilled into the crust in search of oil, gas, and other natural resources have generally been confined to the upper 7 kilometers (4 miles) - only a small fraction of Earth's 6370-kilometer radius. Even the Kola well, a super-deep research well located in a remote northern outpost of Russia, has penetrated to a depth of only 13 kilometers. Although volcanic activity is considered a window into Earth's interior because materials are brought up from below, it allows only a glimpse of the outer 200 kilometers of our planet.



Fortunately, geologists have learned a great deal about Earth's composition and structure through computer modeling, by high-pressure laboratory experiments, and from samples of the solar system (meteorites) that collide with Earth.

In addition, many clues to the physical conditions inside our planet have been acquired through the study of seismic waves pass generated by earthquakes and nuclear explosions. As seismic waves pass through Earth, they carry information to the surface about the materials through which they were transmitted. Hence, when carefully analyzed, seismic records provide an "X-ray" image of Earth's interior.

Probing Earth's Interior

Much of our knowledge of Earth's interior comes from the study of earthquake waves that penetrate Earth and emerge at some distant point. Simply stated, the technique involves accurately measuring the time required for P (*compressional*) and S (*shear*) waves to travel from an earthquake or nuclear explosion to a seismographic station. Because the time required for P and S waves to travel through Earth depends on the properties of the materials encountered, seismologists search for variations in travel times that cannot be accounted for simply by differences in the distances traveled. These variations correspond to changes in the properties of the materials encountered.

One major problem is that to obtain accurate travel, time seismologists must establish the exact location and time of an earthquake. This is often a difficult task because most earthquakes occur in remote areas. By contrast, the exact time and location of a nuclear test explosion is always known exactly. Despite the limitations of studying seismic waves generated by earthquakes, seismologists in the first half of the twentieth century were able to use them to detect the major layers of Earth. It was not until the early 1960s, when nuclear testing was in its heyday and networks consisting of hundreds of sensitive seismographs were deployed, that the finer structures of Earth's interior were established with certainty. (Testing of nuclear devices has been banned by international agreement for several years).

The Nature of Seismic Waves

To examine Earth's composition and structure, we must first study some of the basic properties of wave transmission, or propagation. Seismic energy travels out from its source in all directions as waves. Significant characteristics of seismic waves include the following:

1. The velocity of seismic waves depends on the density and elasticity of the intervening material. Seismic waves travel most rapidly in rigid materials that elastically spring back to their original shapes when the stress caused by a seismic wave is removed. For instance, crystalline rock transmits seismic waves more rapidly than does a layer of unconsolidated mud.

2. Within a given layer the speed of seismic waves generally increases with depth because pressure increases and squeezes the rock into a more compact elastic material.
3. Compressional waves (P waves), which vibrate back and forth in the same plane as their direction of travel, are able to propagate through liquids as well as solids because, when compressed, these materials behave elastically: that is, they resist a change in volume and, like a rubber band, return to their original shape as a wave passes.
4. Shear waves (S waves), which vibrate at right angles to their direction of travel, cannot propagate through liquids because, unlike solids, liquids have no shear strength. That is, when liquids are subjected to forces that act to change their shapes, they simply flow.
5. In all materials, P waves travel faster than do S waves.
6. When seismic waves pass from one material to another, the path of the wave is refracted (bent). In addition, some of the energy is reflected from the **discontinuity** (the boundary between the two dissimilar materials). This is similar to what happens to light when it passes from air into water.

Thus, depending on the nature of the layers through which they pass, seismic waves speed up or slow down, and may be refracted (bent), or reflected. These measurable changes in seismic wave motions enable seismologists to probe Earth's interior.

Seismic Waves and Earth's Structure

If Earth were a perfectly homogeneous body, seismic waves would spread through it in all directions. Such seismic waves would travel in a straight line at a constant speed. However, this is not the case for Earth. It so happens that the seismic waves reaching seismographs located farther from an earthquake travel at faster average speeds than do those recorded at locations closer to the event. This general increase in speed with depth is a consequence of increased pressure, which enhances the elastic properties of deeply buried rock. As a result, the paths of seismic rays through Earth are refracted.

As more sensitive seismographs were developed, it became apparent that in addition to gradual changes in seismic-wave velocities, rather abrupt velocity changes also occur at particular depths. Because these discontinuities were detected worldwide, seismologists concluded that Earth must be composed of distinct shells having varying compositions and/or mechanical properties.

Compositional Layers

Compositional layering likely resulted from density sorting that took place during an early molten period in Earth's history. During this period the heavier elements, principally iron and nickel, sank as the lighter rocky components floated upward. This segregation of material is still occurring, but at a much reduced rate. Because of this chemical differentiation, Earth's interior is not homogeneous. Rather, it consists of three major regions that have markedly different chemical compositions.

The principal compositional layers of earth include (1) the **crust**, Earth's comparatively thin outer skin that ranges in thickness from 3 kilometers (2 miles) at the oceanic ridges to over 70 kilometers (40 miles) in some mountain belts such as the Andes and Himalayas; (2) the **mantle**, a solid rocky (silica-rich) shell that extends to a depth of about 2900 kilometers (1800 miles); and (3) the **core**, an iron-rich sphere having a radius of 3486 kilometers (2166 miles).

Mechanical Layers

Because both pressure and temperature greatly affect the mechanical behavior (strength) as well as the density of Earth materials, other structural divisions exist. For example the core, which is composed mostly of an iron-nickel-alloy, is divided into two regions that exhibit different mechanical behavior.

The **outer core** is a *liquid* metallic layer 2270 kilometers (1410 miles) thick. This zone, which is capable of convective flow, surrounds the **inner core**, a *solid* sphere having a radius of 1216 kilometers (756 miles).

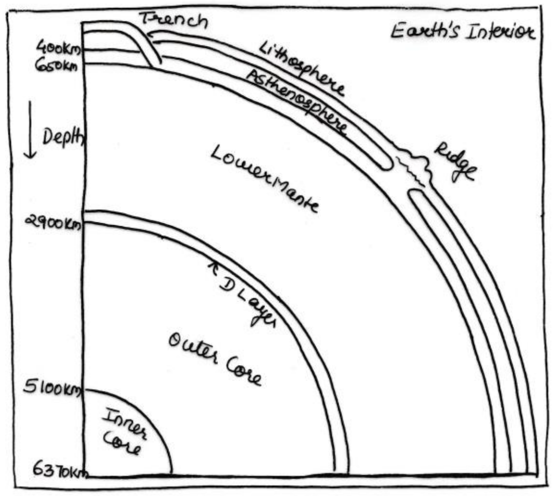
Other structural divisions have been discovered in which materials have undergone a phase change. Phase changes occur, for example, when rock melts, or when the atoms in minerals rearrange themselves into tighter crystalline structures in response to the enormous pressures at great depths. The latter type of phase change occurs at depths between 400 and 660 kilometers.

Discovering Earth's Major Boundaries

Over the past century, seismological data gathered from many seismographic stations have been complied and analyzed. From this information, seismologists have developed a detailed image of Earth's interior. This model is continually being fine-tuned as more data became available and as new seismic techniques are employed. Furthermore, laboratory studies that experimentally determine the properties of various Earth materials under the extreme environments found deep in Earth add to this body of knowledge.

The Moho

In 1909, a pioneering Yugoslavian seismologist, Andrija Mohorovicic, presented the first convincing evidence for layering within Earth. The boundary he discovered separates crustal materials from rocks of different composition in the underlying mantle and was named the **Mohorovicic discontinuity** in his honor. For obvious reasons, the name for this boundary was quickly shortened to **Moho**.



By carefully examining the seismograms of shallow earthquakes, Mohorovicic found that seismographic stations located more than 200 kilometers from an earthquake obtained appreciably faster average travel velocities for P waves than did stations located neared the quake. In particular, P waves that reached the closest stations first had velocities that averaged about 6 kilometers per second. By contrast, the seismic energy recorded at more distant stations traveled at speeds that approached 8 kilometers per second. This abrupt jump in velocity did not fit the general pattern that had been previously observed. From these data, Mohorovicic concluded that below 50 kilometers there exists a layer with properties markedly different from those of Earth's other shell.

The first wave to reach the seismograph located 100 kilometers from the epicenter traveled the shortest route directly through the crust. However, at the seismograph located 300 kilometers from the epicenter, the first P wave to arrive traveled through the mantle, a zone of higher velocity. Thus, although this wave traveled a greater distance, it reached the recording instrument sooner than did the rays taking the more direct route. This is because a large portion of its journey was through a region having a composition where seismic waves travel more rapidly. This principle is analogous to a driver taking a bypass route around a large city during rush hour. Although this alternate route is longer, it may be faster.

The Core-Mantle Boundary

A few years later, in 1914, the location of another major boundary was established by the German seismologist Beno Gutenberg. This discovery was based primarily on the observation that P waves diminish and eventually die out completely about 105 degrees from an earthquake.

Then, about 140 degrees away, the P waves reappear, but about 2 minutes later than would be expected based on the distance traveled. This belt, where direct seismic waves are absent, is about 35 degrees wide and has been named the **P wave shadow zone**.

Gutenberg, and others before him, realized that the P wave shadow zone could be explained if Earth contained a core that was composed of material unlike the overlying mantle. The core, which Gutenberg calculated to be located at a depth of 2900 kilometers, must somehow hinder the transmission of P waves similar to the way in which light rays are blocked by an object that casts a shadow. However, rather than actually stopping the P waves, the shadow zone is produced by bending (refracting) of the P waves, which enter the core.

It was further determined that S waves do not travel through the core. This fact led geologists to conclude that atleast a portion of this region is liquid. This conclusion was further supported by the observation that P-wave velocities suddenly decrease by about 40 percent as they enter the core. Because melting reduces the elasticity of rock, this evidence pointed to the existence of a liquid layer below the rocky mantle.

Discovery of the Inner Core

In 1936, the last major subdivision of Earth's interior was predicted by Inge Lehmann, a Danish seismologist. Lehmann discovered a new region of seismic reflection and refraction within the core. Hence, a core within a core was discovered. The size of the inner core was not precisely established until the early 1960s when underground nuclear tests were conducted in Nevada. Because the exact location and time of the explosions were known, echoes from seismic waves that bounced off the inner core provided a precise means of determining its size.

From these data, the inner core was found to have a radius of about 1216 kilometers. Furthermore, P waves passing through the inner core have appreciably faster average velocities than do those penetrating only the outer core. The apparent increase in the elasticity of the inner core is evidence that this innermost region is solid.

EARTH'S INTERIOR

Details of interior layers

Over the past few decades, advances in seismology and rock mechanics have allowed for much refinement of the gross view of Earth's interior that has been presented to this point.

The Crust

Earth's crust averages less than 20 kilometers thick, making it the thinnest of Earth's divisions. Along this eggshell-thin layer great variations in thickness exist. Crustal rocks of the stable continental interiors are roughly 30 kilometers thick. However, in a few exceptionally prominent mountainous regions, the crust obtains its greatest thickness, exceeding 70 kilometers. The oceanic crust is much thinner, ranging from 3 to 15 kilometers thick. Further, crustal rocks of the deep-ocean basins are compositionally different from those of the continents.

Continental rocks have an average density of about 2.8 g/cm3 and some have been discovered that exceed 3.8 billion years in age. From both seismic studies and direct observations, the average compositions of continental rocks are estimated to be comparable to the felsic igneous rock *granodiorite*. Like granodiorite, the continental crust is enriched in the elements potassium, sodium, and silicon. Although numerous granitic intrusions and chemically equivalent metamorphic rocks are very abundant, large outpourings of basaltic and andesitic rocks are also commonly found on the continents.

The rocks of the oceanic crust are younger (180 million years or less) and more dense (about 3.0 g/cm3) than continental rocks. The deep-ocean basins lie beneath 4 kilometers of seawater as well as hundreds of meters of sediment. Thus, until recently, geologists had to rely on indirect evidence (such as slivers of what was thought to be oceanic crust that were thrust unto land) to estimate the composition of this inaccessible region. With the development of deep-sea drilling ships, the recovery of core samples from the ocean floor became possible. As predicted, the samples obtained were predominantly *basalt*. Volcanic eruptions of basaltic lave are known to have generated many islands, such as the Hawaiian chain, located within the deep-ocean basins.

The Mantle

Over 82 percent of Earth's volume is contained within the mantle, a nearly 2900-kilometer-thick shell of silicate rock extending from the base of the crust (Moho) to the liquid outer core. Our knowledge of the mantle's composition comes from experimental data and from the examination of material carried to the surface by volcanic activity. In particular, the rocks composing kimberlite pipes, in which diamonds are sometimes found, are often thought to have originated at depths approaching 200 kilometers, well within the mantle. Kimberlite deposits are composed of *peridotite*, a rock that contains iron and magnesium-rich silicate minerals, mainly olivine and pyroxene, plus lesser amounts of garnet. Further, because S waves readily travel through the mantle, we know that it behaves as an elastic solid.

Thus, the mantle is described as a solid rocky layer, the upper portion of which has the composition of the ultramafic rock peridotite.

The mantle is divided into the **mesosphere** or *lower mantle*, which extends from the core-mantle boundary to a depth of 660 kilometers, and the *upper mantle*, which continues to the base of the crust. In addition, other subdivisions have been identified. At the depth of about 400 kilometers a relatively abrupt increase in seismic velocity occurs. Whereas the crust-mantle boundary represents a compositional change, the zone of seismic velocity increase at the 400-kilometer level is the result of a *phase change*. (A phase change occurs when the crystalline structure of a mineral is altered in response to changes in temperature and/or pressure.) Laboratory studies show that the mineral olivine, (Mg, Fe)2SiO4, which is one of the main constituents in the rock peridotite, will collapse to a more compact, high-pressure mineral (spinel) at the pressure experienced at this depth. This change to a denser crystal form explains the increased seismic velocities observed.

Another boundary within the mantle has been detected from variations in seismic velocity at a depth of 660 kilometers. At this depth the mineral spinal is believed to undergo a transformation to the mineral perovskite (Mg, Fe)SiO3. Because perovskite is thought to be pervasive in the lower mantle, it is perhaps Earth's most abundant mineral.

In the lowermost roughly 200 kilometers of the mantle, there exists an important region known as the D" layer. Recently, researchers reported that seismic waves traveling through some parts of the D" layer experience a sharp decrease in P-wave velocities. So far, the best explanation for this phenomenon is that the lowermost layer of the mantle is partially molten in places.

If these zones of partially melted rock exist, they are very important because they would be capable of transporting heat from the core to the lower mantle much more efficiently than solid rock. A high rate of heat flow would, in turn, cause the solid mantle located above these partly molten zones to be warmed sufficiently to become buoyant and slowly rise toward the surface.

Such rising plumes of super-heated rock may be the source of volcanic activity associated with hotspots, such as that experienced at Hawaii and Iceland. If these observations are accurate, some of the volcanic activity that we see at the surface is a manifestation processes occurring 2900 kilometers (1800 miles) below our feet.

The Lithosphere and Asthenosphere

Earth's outer layer, consisting of the uppermost mantle and crust, forms a relatively cool, rigid shell. Although this layer consists of materials with markedly different chemical compositions, it tends to act as a unit that behaves similarly to mechanical deformation. This outermost rigid unit of Earth is called the **lithosphere** (*sphere of rock*). Averaging about 100 kilometers in thickness, the lithosphere may be 250 kilometers or more in thickness below the older portions (shields) of the continents. Within the ocean basins the lithosphere is only a few kilometers thick along the oceanic ridges and increases to perhaps 100 kilometers in regions of older and cooler crustal rocks.

Beneath the lithosphere (to a depth of about 660 kilometers) lies a soft, relatively weak layer located in the upper mantle known as the **asthenosphere** ("*weak sphere*"). The upper 150 kilometers or so of the asthenophere has the temperature/pressure regime in which a small amount of melting takes place (perhaps 1 to 5 percent). This region of partial melting within the upper asthenosphere is known as the **low-velocity zone**, because seismic waves show a marked decrease in velocity. Within this very weak zone, the lithosphere is effectively detached from the asthenosphere located below. The result is that the lithosphere is able to move independently of the asthenosphere.

It is important to emphasize that the strength of various Earth materials is a function of their composition, as well as the temperature and pressure of their environment. You should not get the idea that the entire lithosphere is brittle, like the rocks found at the surface. Rather, the rocks of the lithosphere get progressively weaker (more easily deformed) with increasing depth. At the depth of the upper asthenosphere (low-velocity zone) the rocks are very close to their melting temperature (some melting is thought to occur) so that they are easily deformed. Thus, the asthenosphere is weak because it is near its melting point, just as hot wax is weaker than cold wax. However, in the material located below this weak zone, increased pressure offsets the effects of increased temperature. Therefore, these materials gradually stiffen with depth, forming the more rigid lower mantle. Despite their greater strength, the materials of the lower mantle are still capable of very gradual flow.

The Core Larger than the planet Mars, the core is Earth's dense central sphere with a radius of 3486 kilometers. Extending from the inner edge of the mantle to the center of Earth, the core constitutes about one-sixth of Earth's volume and nearly one-third of its total mass. Pressure at the center is millions of times greater than the air pressure at Earth's surface, and the temperatures can exceed 6700° C. As more precise seismic data became available, the core was found to consist of a liquid outer layer about 2270 kilometers thick, and a solid inner sphere with a radius of 1216 kilometers.

Density and Composition

One of the more interesting characteristics of the core is its great density. The average density of the core is about 11 g/cm3, and approaches 14 times the density of water at Earth's center. Even under the extreme pressure at these depths, the common silicate minerals found in the crust (with surface densities 2.6 to 3.5g/cm3) would not be compacted enough to account for the density calculated for the core. Consequently, attempts were undertaken to determine the material that could account for this property. Surprisingly enough, meteorites provided an important clue to Earth's internal composition. Because meteorites are part of the solar system, they are assumed to be representative samples of the material from which Earth originally accreted.

Their composition ranges from metallic types, made primarily of iron and lesser amounts of nickel, to stony meteorites composed of rocky substances that closely resemble the rock peridotite. Because Earth's crust and mantle contain a much smaller percentage of iron than is found in the debris of the solar system, geologists concluded that the interior of Earth must be enriched on this heavy metal. Further, iron is by far the most abundant substance found in the solar system that possesses the seismic properties and density resembling that measured for the core. Current estimates suggest that the core is mostly iron, with 5 to 10 percent nickel and lesser amounts of lighter elements, including perhaps sulfur and oxygen.

Origin, Although the existence of a metallic central core is well established, explanations of the core's origin are more speculative. The most widely accepted explanation suggests that the core formed early in Earth's history from what was originally a relatively homogeneous body. During the period of accretion, the entire Earth was heated by energy released by the collisions of infalling material. Sometime late in this period of growth, Earth's internal temperature was sufficiently high to melt and mobilize the accumulated material. Blobs of heavy iron-rich materials collected and sank toward the center. Simultaneously, lightly substances may have floated upward to generate the crust. In a short time, geologically speaking, Earth took on a layered configuration, not significantly different from what we find today.

In its formative stage, the entire core was probably liquid. Further, this liquid iron alloy was in a state of vigorous mixing. However, as Earth began to cool, iron in the core began to crystallize and the inner core began to form. As the core continues to cool, the inner core should grow at the expense of the outer core.

Earth's Magnetic Field

Our picture of the core with its solid inner sphere surrounded by a mobile liquid shell is further supported by the existence of Earth's magnetic field. This field behaves as though a large bar magnet were situated deep within Earth. However, we know that the source of the magnetic field cannot be permanently magnetized material, because Earth's interior is too for any material to retain its magnetism. The most widely accepted explanation of Earth's magnetic field requires that the core be made of a material that conducts electricity, such as iron, and one that is mobile. Both of these conditions are met by the model of Earth's core that was established on the basis of seismological data. One recently discovered consequence of Earth's magnetic field is that it affects the rotation of the solid inner core. Current estimates indicate that the inner core rotates in a west-to-east direction about 1 degree a year *faster* than the Earth's surface. Thus, the core makes one extra rotation for the inner core is offset about 10 degrees from Earth's rotational poles.

Earth's Internal Heat Engine

Temperature gradually increases with an increase in depth at a rate known as the **geothermal gradient**. The geothermal gradient varies considerably from place to place. In the crust, temperatures increase rapidly, averaging between 20°C and 30°C per kilometer. However, the rate of increase is much less in the mantle and core. At a depth of 100 kilometers, the temperature is estimated to exceed 1200°C, whereas the temperature at the core-mantle boundary is calculated to be about 4500°C and it may exceed 6700°C at Earth's center (hotter than the surface of the Sun).

Three major processes have contributed to Earth's internal heat: (1) heat emitted by radioactive decay of isotopes of uranium (U), thorium (Th), and potassium (K); (2) heat released as iron crystallized to form the solid inner core; and (3) heat released by colliding particles during the formation of our planet. Although the first two processes are still operating, their rate of heat generation is much less than in the geologic past. Today, our planet is radiating more of its internal heat out to space than is being replaced by these mechanisms. Therefore, Earth is slowly, but continuously, cooling.

Heat Flow in the Crust

Heat flow in the crust occurs by the familiar process called **conduction**. Anyone who has attempted to pick up a metal spoon left in a hot pan is quick to realize that heat was conducted through the spoon. *Conduction*, which is the transfer of heat through matter by molecular activity, occurs at a relatively slow rate in crustal rocks. Thus, the crust tends to act as an insulator (cool on top and hot on the bottom), which helps account for the steep temperature gradient exhibited by the crust.

Certain regions of Earth's crust have much higher rates of heat flow than do others. In particular, along the axes of mid-ocean ridges where the crust is only a few kilometers thick, heat flow rates are relatively high. By contrast, a relatively low heat flow is observed in ancient shields (such as the Canadian and Baltic Shields). This may occur because these zones have a thick lithosphere root that effectively insulates the crust from the asthenospheric heat below. Other crustal regions exhibit a high heat flow, because of shallow igneous intrusions or because of higher-than average concentrations of radioactive materials.

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