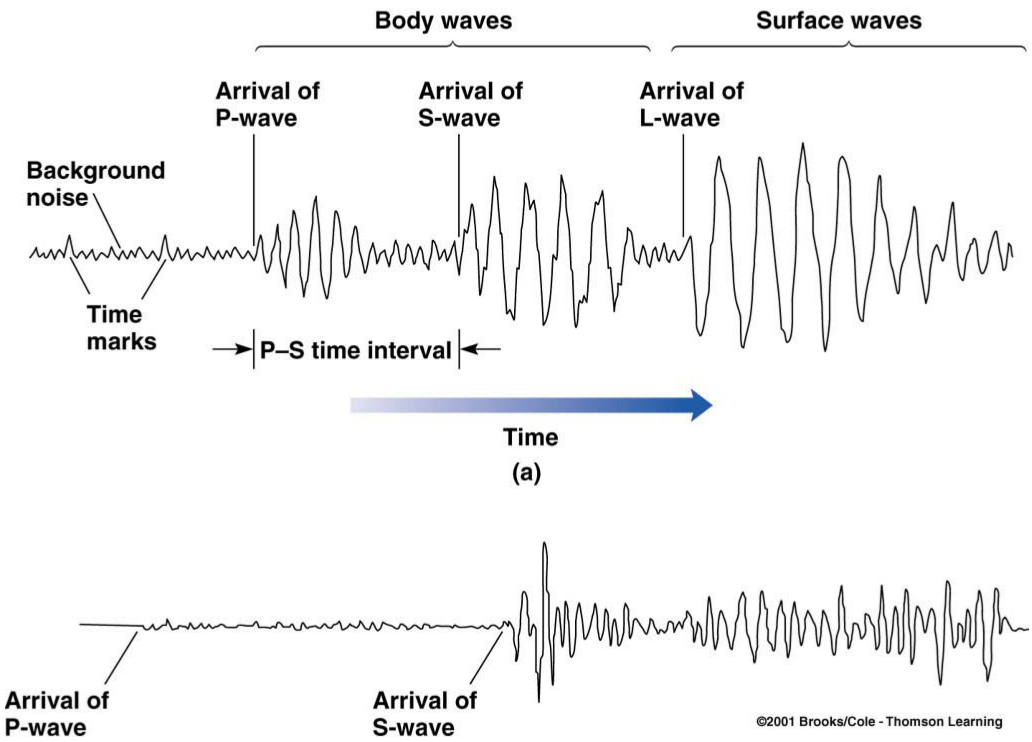
Ideal Answer of Geophysics Examination (Seismics)

4th Year Special Geology

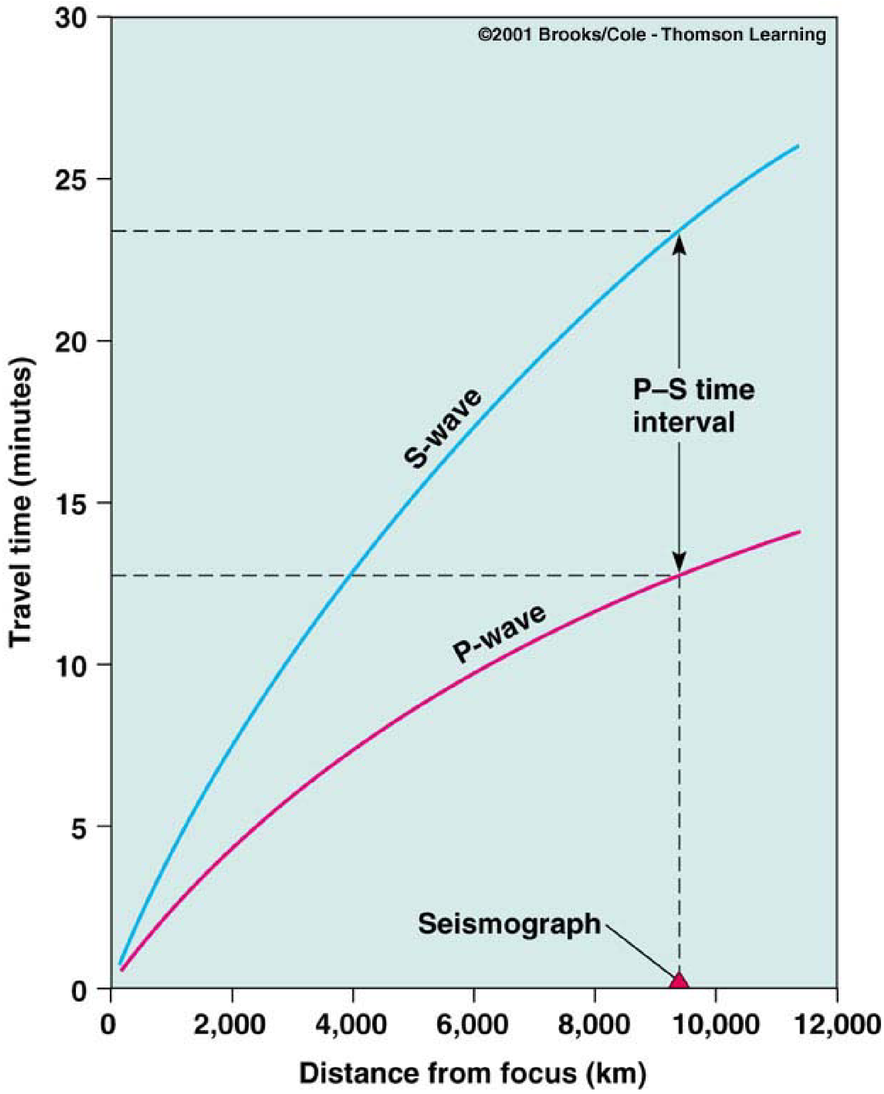
How is an Earthquake's Epicenter Located?

Seismic wave behavior

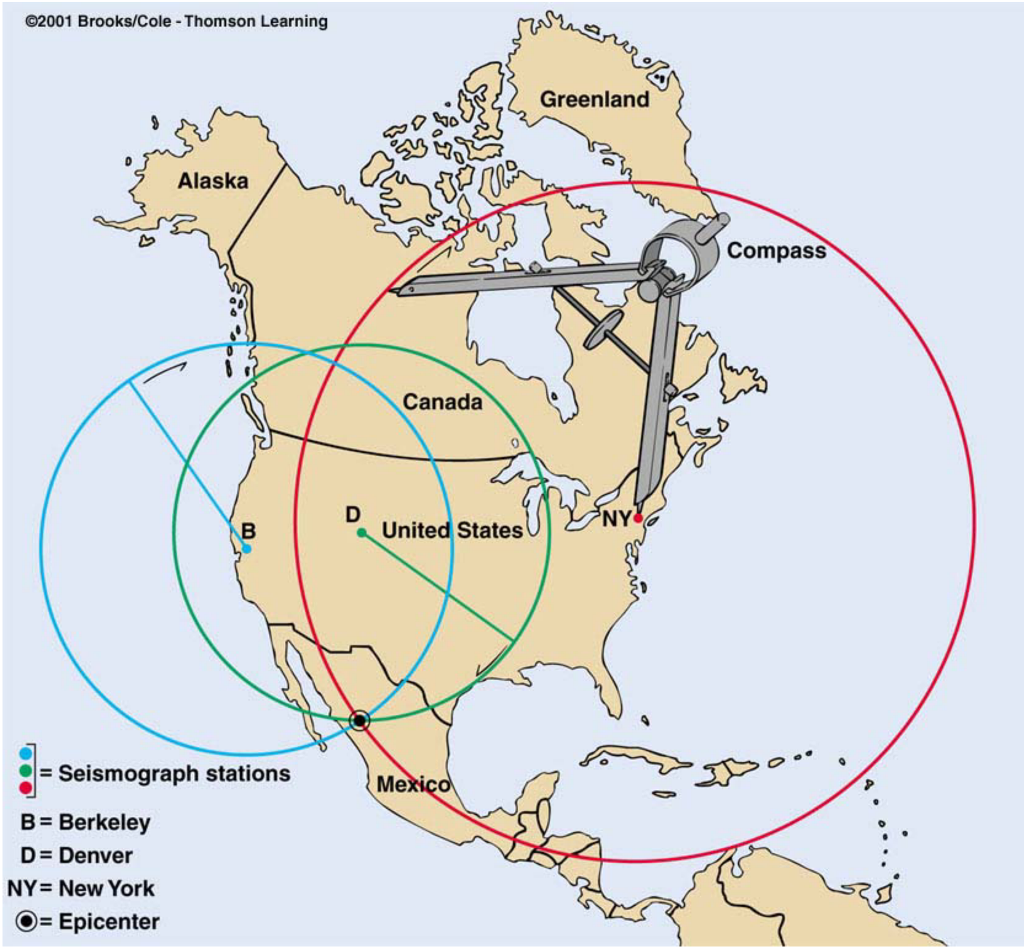
- P waves arrive first, then S waves, then L and R
- Average speeds for all these waves is known
- After an earthquake, the difference in arrival times at a seismograph station can be used to calculate the distance from the seismograph to the epicenter.



Time-distance graph showing the average travel times for P- and S-waves. The farther away a seismograph is from the focus of an earthquake, the longer the interval between the arrivals of the P- and S-waves.



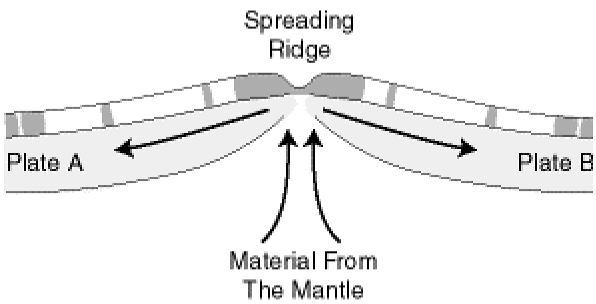
- A circle where the radius equals the distance to the epicenter is drawn.
- Three seismograph stations are needed to locate the epicenter of an earthquake.
- The intersection of the circles locates the epicenter.



2- Most earthquakes occur along the edge of the oceanic and continental plates as the following:

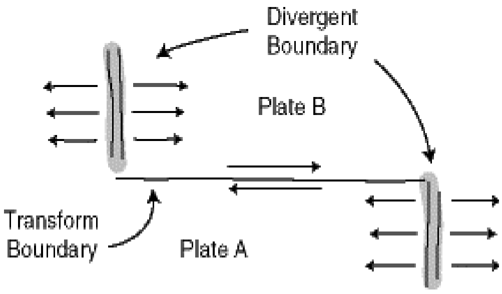
Divergent boundaries

- Movement of plates at a divergent boundary normally produces small, shallow earthquakes
- Mid-Atlantic ridge is an example of a divergent boundary



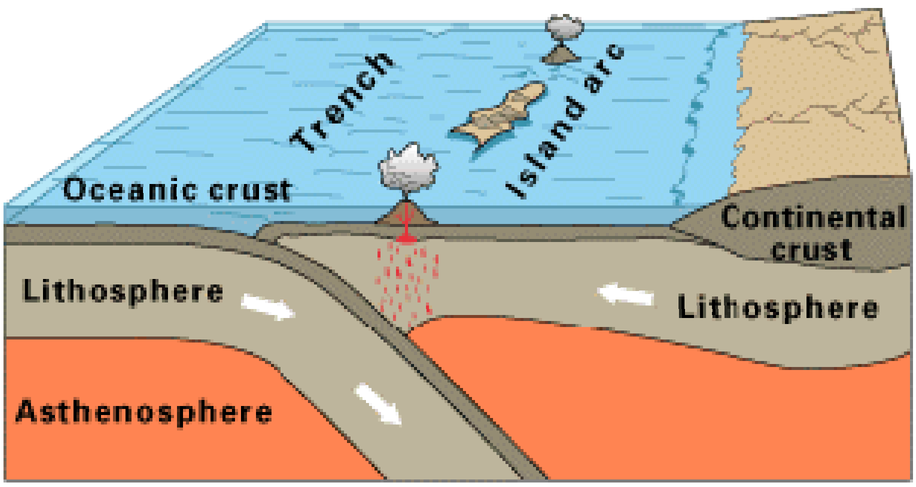
Conservative (transform) boundaries

- Movement of plates at a transform boundary can produce large, shallow to intemediate deeps (<300 km), earthquakes
- San-Andreas fault (USA) is an example of a transform fault.



Convergent boundaries (b)

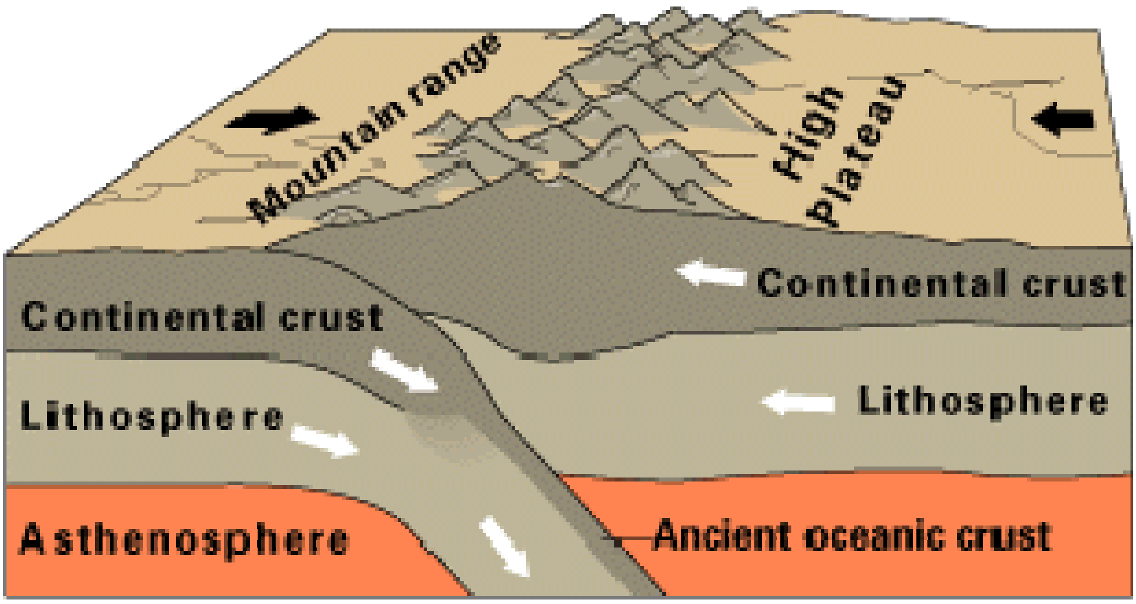
- One oceanic plate subducts under the other plate forming a deep oceanic trench at the boundary.
- Island volcanoes are produced by released water, at high temp. and pressure, from subducting plate.
- Large, deep (>300km), earthquakes are produced.



Oceanic-oceanic convergence

Convergent boundaries (c)

- One continental plate subducts under the other continental plate forming a mountain ranges and high plateaux,
- Himalayan mountain range (about 8.9km high) is an example a feature caused by of convergent boundary of the Indian and Eurasian plates
- Large, deep (>300km), earthquakes are produced



Continental-continental convergence

3-

a) Seismic Travel-Time Curves

The *travel time* of a seismic wave from a source to a receiver depends on the seismic velocities of the Earth materials traversed, the distance from the source to the receiver, and the geometry of boundaries separating Earth materials. A receiver commonly records more than one arrival of seismic energy because the energy may radiate from the source as various body and surface waves, and because the body waves are refracted and reflected along different paths when they encounter boundaries.

Fig. 9 shows that the bulk and shear moduli (k and μ) are generally of the same order of magnitude. The equations:

$$\boxed{V_P = \sqrt{\frac{k + 4/3\mu}{\rho}}}$$

and

$$\boxed{V_S = \sqrt{\frac{\mu}{\rho}}}$$

Therefore illustrate that, for many Earth materials, $V_S \approx 0.6\ V_P$. Rayleigh waves are slightly slower than shear waves, so that the Rayleigh wave velocity $V_R \approx 0.9\ V_S$. Rayleigh waves thus travel about half the speed of compressional waves ($V_R \approx 0.5\ V_P$).

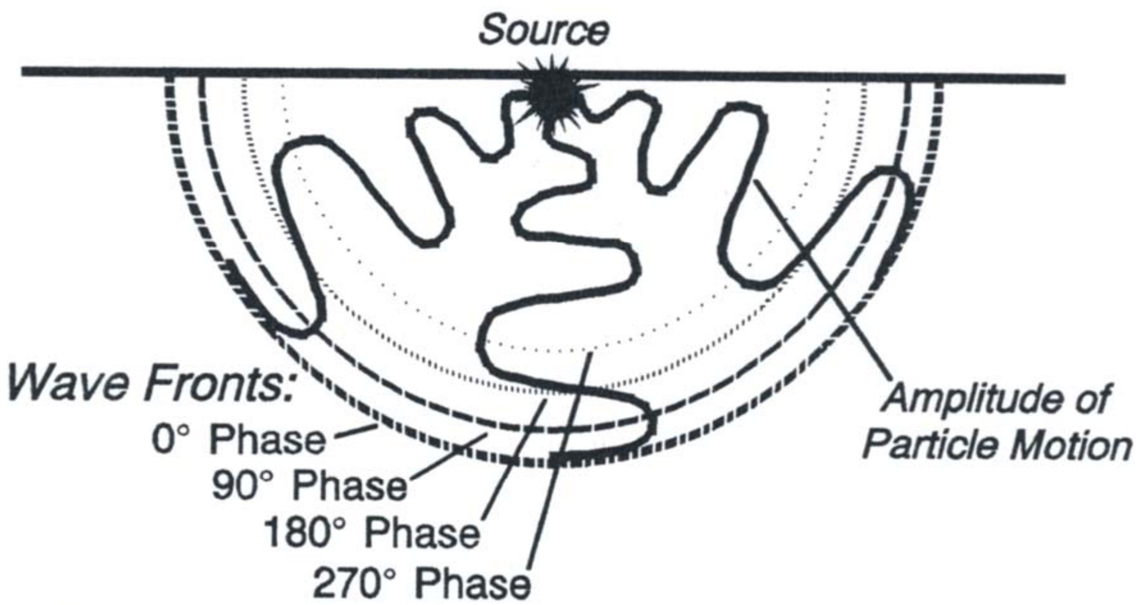


FIGURE 14 Wave fronts are surfaces along which particle motions of the propagating wave are in phase (one complete oscillation is 360° of phase). For example, a surface where particle motions reach their maximum positive amplitude is 90° phase; where they are maximum negative amplitude is 270° phase.

A *wave front* is a surface along which portions of a propagating wave are *in phase.* For example, arrival of the wave in Fig. 14 occurs where particles first move as the wave approaches (wave front at 0° phase). The maximum amplitude of particle motion occurs along the 90° phase wave front. Other wave fronts correspond to positions where the wave goes from positive to negative amplitude (180°) and at minimum amplitude (270°).

Consider the wave fronts that represent the leading edges of oncoming P, S, and Rayleigh waves (Fig. 15a). In a homogeneous medium (constant seismic velocities), the body waves (P and S) radiate outward along spherical wavefronts, while Rayleigh wave (R) roll along the surface.

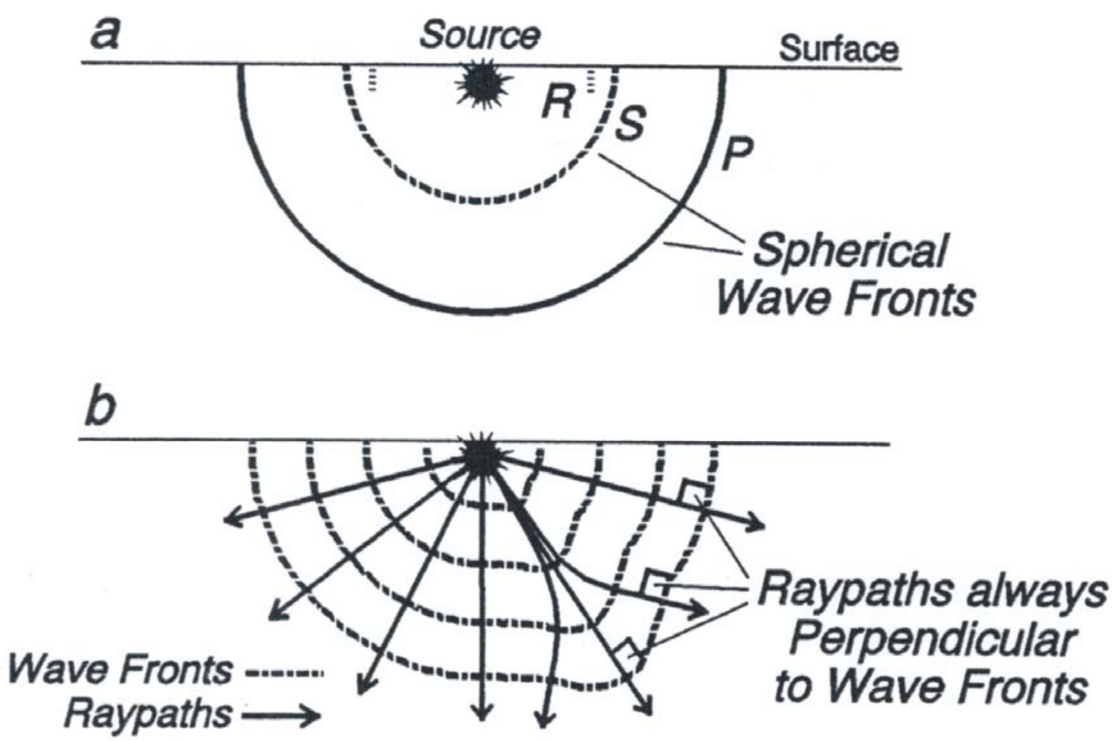


FIGURE 15 a) Initial wave fronts for compressional (P), shear (S), and Rayleigh (R) waves. b) Wave fronts for propagating P-wave. Changes in velocity cause segments of wave fronts to speed up or slow down, distorting the wave fronts from perfect spheres. Raypaths thus bend (refract) as velocity changes.

Seismic energy travels along trajectories perpendicular to wavefronts, known as *raypaths* (Fig. 15b). Variations in body wave velocity cause wave fronts to deviate from perfect spheres, thus bending or *refracting* the raypaths. Raypaths can be used to analyse portions of seismic waves that make it back to the surface, as discussed below for direct, critically refracted, and reflected waves.

A *seismic trace* is the recording of ground motion by a receiver, plotted as a function of time (Fig. 16). A seismic wave takes a certain amount of time to travel from the source to a receiver, depending on the distance to the receiver, the path taken by the wave, and the wave's velocity. Arrival of each of the P, S, and Rayleigh waves starts as initial movement of the ground, followed by reverberations that die out with time (Fig. 16b). A body wave arrives as a relatively short burst of energy. Surface waves, however, are commonly *dispersed;* broad (low-frequency) waves arrive first, followed by progressively narrower (higher frequency) arrivals.

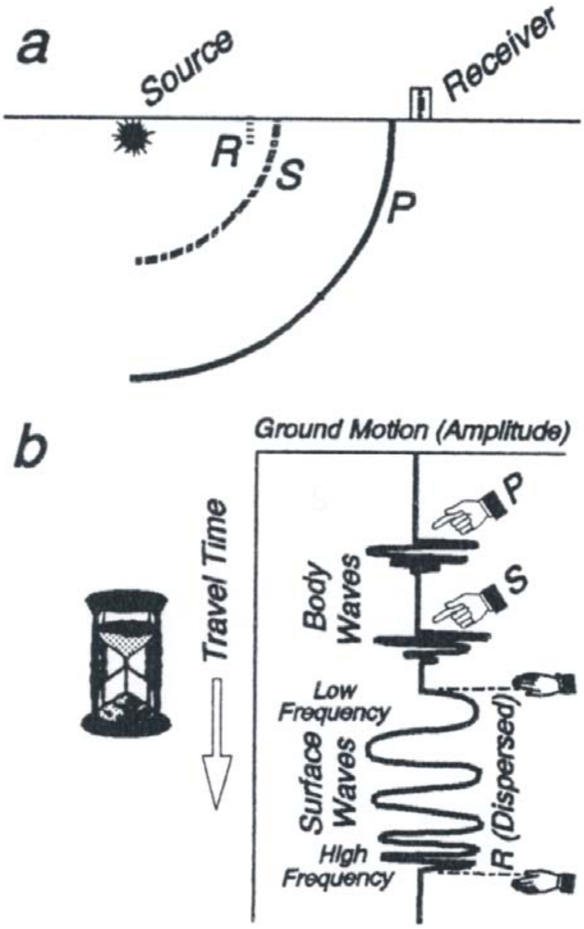


FIGURE 16 a) Seismic waves radiating from a source to one receiver. b) Seismic trace recording ground motion by the receiver, as a function of the travel time from the source to the receiver. For controlled source studies (seismic refraction and reflection), the travel time is commonly plotted positive downward.

On *travel-time graph* seismic traces from several receivers are plotted side by side, according to the horizontal distance (X) from the source to each receiver (Fig. 17). Travel time (T) is commonly plotted as *increasing downward* in refraction and reflection studies, because T often relates to depth within the Earth. For each of the initial P-wave, S-wave, or R-wave arrivals, the travel time from the source to a receiver is linear, expressed by the *travel-time curve*:

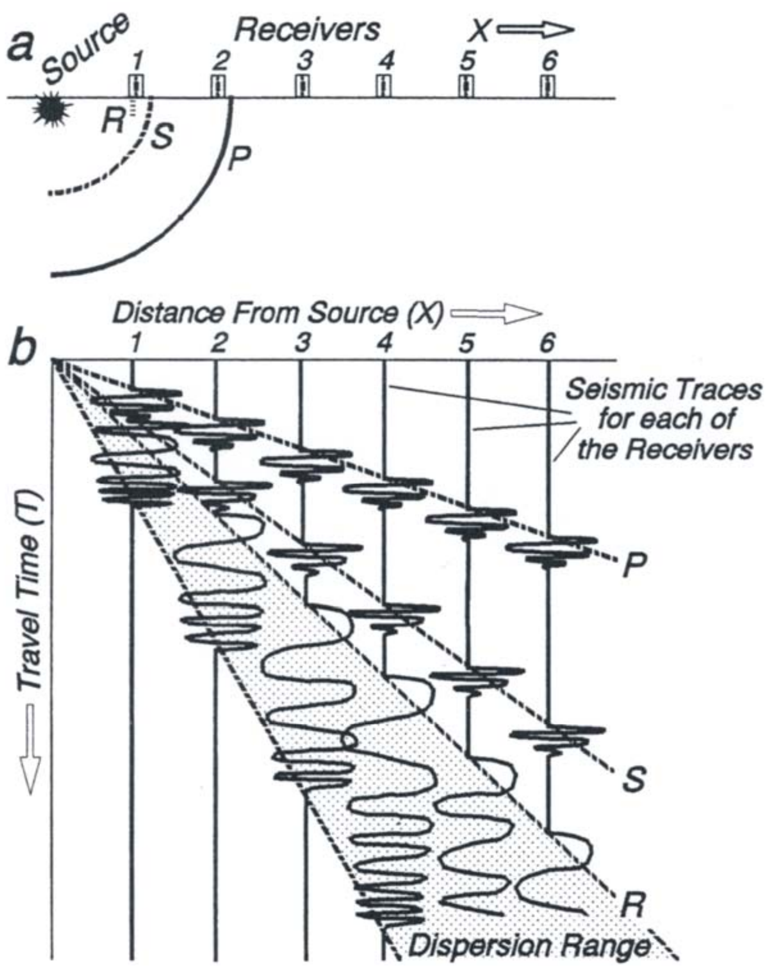


FIGURE 17 a) Initial wave fronts for P, S, and R waves, propagating across several receivers at increasing distance from the source. b) Travel-time graph. The seismic traces are plotted according to the distance (X) from the source to each receiver. The elapsed time after the source is fired is the travel time (T).

$$\boxed{T = \frac{X}{V}}$$

Where:

T = total time for the wave to travel from the source to the receiver.

X = distance from the source to the receiver, measured along the surface.

V = seismic velocity of the P, S, or R arrival.

The slope of the line is the elapsed time (ΔT) divided by the distance traveled during that time (ΔX):

$$\text{Slope} = \frac{\Delta T}{\Delta X}$$

The slope at a given distance (X) can also be determined by taking the first derivative at that point on the travel time curve (Fig. 3 8 a):

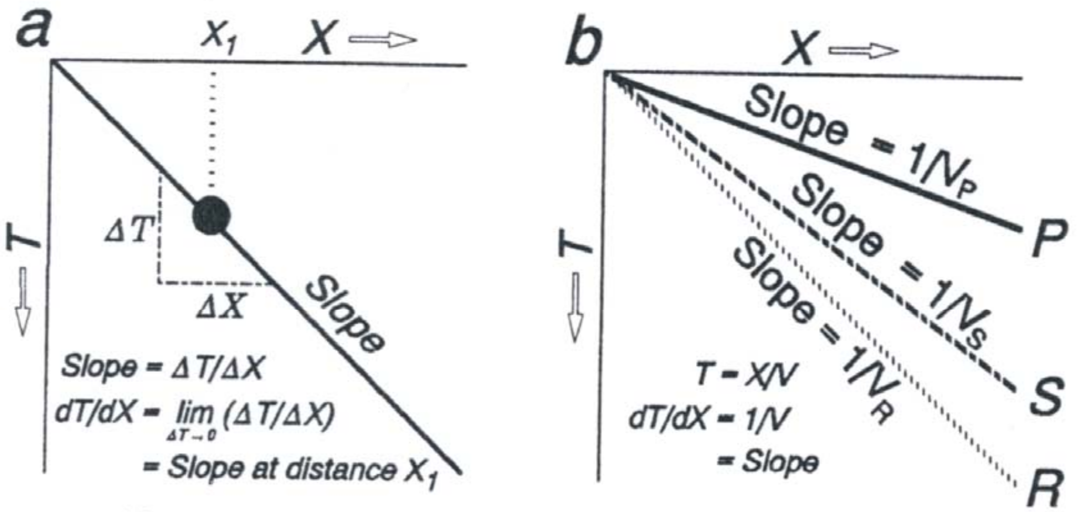


FIGURE : 18 Travel-time curves for wave traveling in a homogenous medium. a) The slope of the line for each arrival is the first derivative (dT/dX). b) The slope of the travel time for each of the P, S, and R arrivals (Figure 17) is the inverse of the velocity.

$$\frac{dT}{dX} = \lim_{\Delta t \to 0} \left\{\frac{\Delta T}{\Delta X}\right\}$$

The travel time can be written:

$$T = \frac{1}{V}(X)$$

So that:

$$\frac{dT}{dX} = \frac{1}{V}$$

The velocities for each type of wave can thus be calculated by taking the inverse of the slope (Fig. 18b):

$$V = \frac{1}{dT/dX}$$

The first derivative, or slope, is thus useful in determining the velocity represented at any point on travel-time curves for different arrivals.

b) Elastic Constants An *elastic constant* describes the strain of a material under a certain type of stress. The *bulk modulus* (or *incompressibility)* describes the ability to resist being compressed (Fig. 5). Under pressure that is equal in all directions (hydrostatic for material under water; lithostatic for material within the Earth), the stress is the change in pressure (ΔP). The strain is the change in volume (ΔV) divided by the original volume (V). The bulk modulus (k) is the stress divided by the strain:

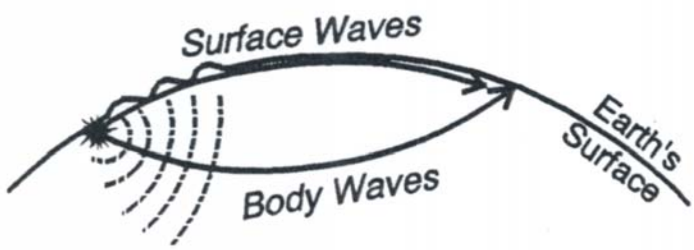


FIGURE 4 Surface wave energy is confined to a thin, outer shell of the Earth. Body wave energy radiates in three dimensions, like the surface of a balloon expanding through the Earth.

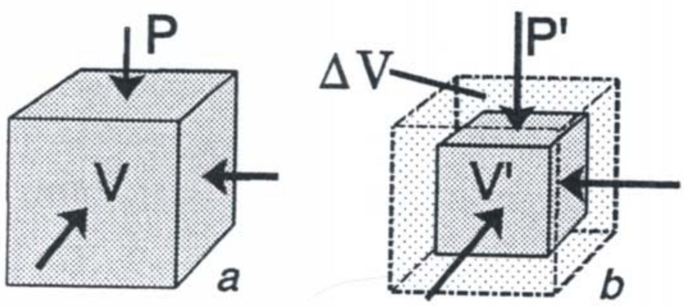


FIGURE 5 *Bulk modulus.* a) Material of volume V under confining pressure P, shown by equal arrows. b) Material compressed to volume V′ as pressure increases to P′ (longer arrows). The bulk modulus (k) determines the change in volume (ΔV) of the material.

$$K = \frac{Stress}{Strain} = \frac{\Delta P}{\Delta V/V}$$

Where:

$\Delta P = P' - P$ = pressure change (applied stress)

P = original confining pressure

P' = confining pressure under the applied stress

$\Delta V = V - V'$ = change in volume caused by ΔP

V = original volume V' = volume under the applied stress.

The above equation illustrates that if a material undergoes no volume change ($\Delta V = 0$) when subjected to compressive stress (ΔP), the material is said to be incompressible ($k = \infty$). Conversely, materials that are easy to compress (k very small) undergo large changes in volume (large ΔV) when subjected to relatively small compressive stresses (small ΔV) when subjected to relatively small compressive stresses (small ΔP).

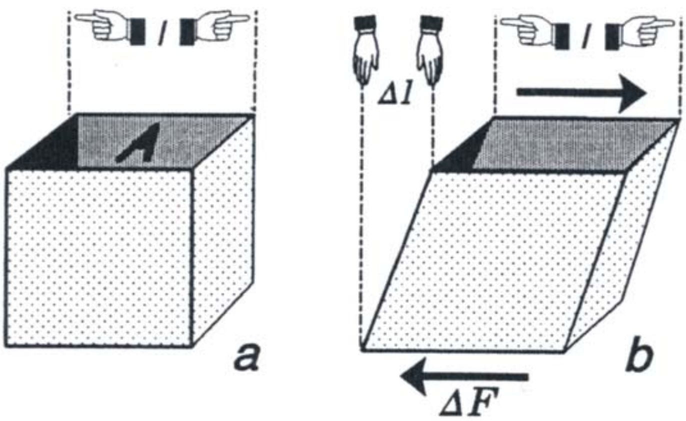


FIGURE 6 *Shear modulus.* a) Configuration of material before change in shear force. Note cube with sides of area A and length l. b) The change in shear force ΔF acts across the area A. One side of the cube (A) is displaced a distance Δl relative to the opposite side, according to the shear modulus of the material.

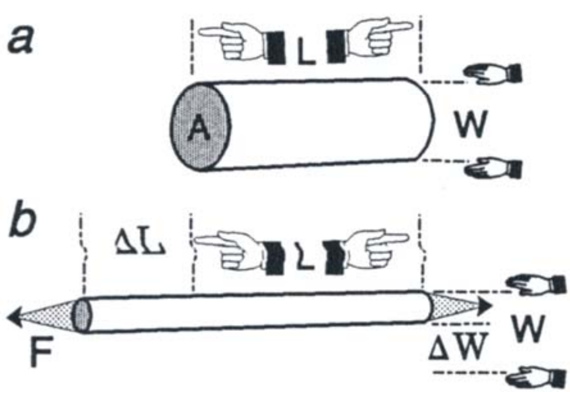


FIGURE 7 a) A rod of length (L), width (W) and cross-sectional area (A) can be used to measure Young's modulus and Poisson's ratio. The rod is subjected to a longitudinal stress (force, F, acting over the cross-sectional area, A). *Young's modulus* determines the resulting longitudinal strain (change in length, ΔL, divided by the original length, L). *Poisson's ratio* is the transverse strain (ΔW/W) divided by the longitudinal strain (ΔL/L).

The ***shear modulus* (or *rigidity)*** refers to the ability of a material to resist shearing (Fig. 6). When a cube of material is subjected to shearing, the stress is the tangential force (ΔF) divided by the area over which the force is applied (A). The strain is the shear displacement (Δ1) divided by the length (1) of the area acted upon by ΔF. For such a stress, the shear modulus (μ) is:

$$\boxed{\mu = \frac{Stress}{Strain} = \frac{\Delta F/A}{\Delta l/l}}$$

A material that shows strong resistance to shearing ($\Delta l = 0$) is very rigid ($\mu = \infty$). A fluid, on the other hand, has no resistance to shearing ($\Delta l = \infty$) and therefore lacks rigidity ($\mu = 0$). For an unbounded, isotropic material, the elastic constants k and μ, along with the density *(ρ),* determine how fast body waves travel through the material. It may not be practical, however, to measure those two elastic constants directly. Other constants may be more readily measured and used to calculate k and μ (Fig. 7).

Young's modulus (or *the stretch modulus)* describes the behavior of a rod that is pulled or compressed, according to the equation:

$$\boxed{E = \frac{Stress}{Strain} = \frac{F/A}{\Delta L/L}}$$

Where:

E = Young's modulus

F/A = force per unit area applied to end of rod

L = original length of rod

ΔL = change in length of rod.

Poisson's ratio states that, for a stretched rod, the ratio of transverse strain (ΔW/W) to longitudinal strain (ΔL/L) is:

$$\boxed{\upsilon = \frac{\Delta W/W}{\Delta L/L}}$$

Where:

υ = Poisson's ratio

W = original width of rod

ΔW = amount by which width contracts.

<u>**Lame's constant**</u> **(λ)** illustrates the relationship between the four constants discussed above, according to:

$$\boxed{\lambda = k - \frac{2\mu}{3} = \frac{\upsilon E}{(1+\upsilon)(1-2\upsilon)}}$$

Relatively easy measurements of E and υ can be used to determine λ for a material; λ can then be used as one of the parameters describing the velocity of seismic waves through the material.

c) CONTROLLED SOURCE SEISMIC TECHNIQUES

Seismic waves can be used to determine depths to interfaces within the Earth and velocities of layers between the interfaces. The velocities in turn may be used as one parameter to interpret the nature of Earth materials.

The sources of the seismic waves can be natural *(earthquake)* or produced artificially (*controlled source).* For the latter, the source is at (or just below)

Earth's surface and an array of instruments are laid out on the surface to receive the direct, reflected, or refracted signals (Fig.13).

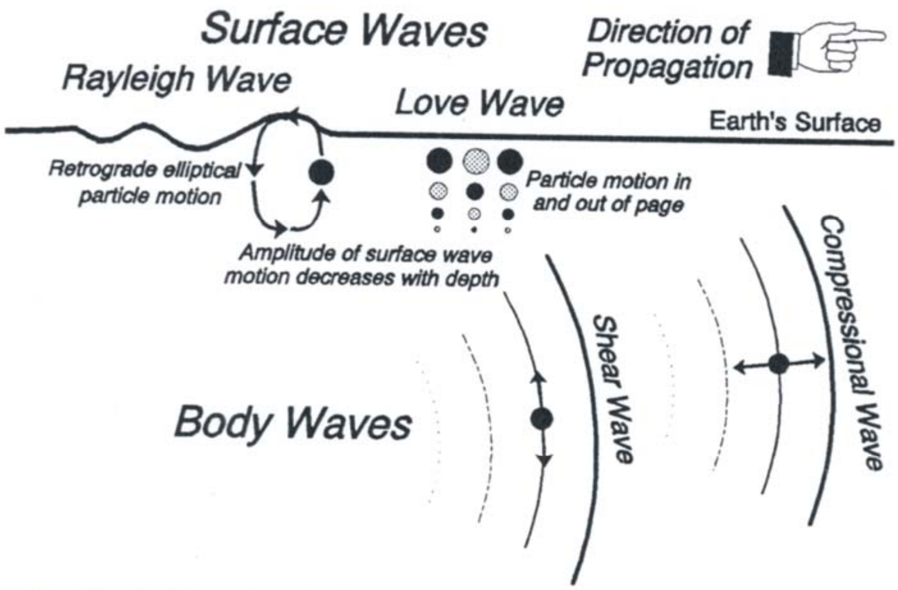


FIGURE 12 Particle motions of surface (Love and Rayleigh) waves, compared to that of body (compressional and shear) waves. Solid dots represent Love wave motions out of the page; gray dots are motions into the page.

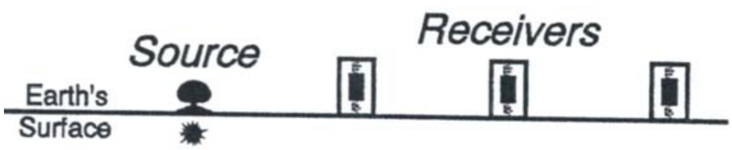


FIGURE 13 Receivers that record seismic waves are generally laid out in a line away from the artificial source. The source could be an explosion at the surface or in a shallow drillhole. The receivers depicted are geophones.

The artificial sources for surveys conducted on land may be from explosives, such as dynamite placed in drillholes, quarry blasts, or even nuclear tests. Alternatively, plates coupled to the ground beneath large trucks may massage the ground in a continuous, predetermined fashion, using a technique known as Vibroseis. Seismic studies at sea (or on rivers or lakes) commonly employ air guns towed behind a survey boat.

The receivers used on land, called *geophones,* measure ground movement (either the displacement, the velocity, or the acceleration of the ground surface). A common geophone is a magnet on springs, suspended within a coil of wire (Fig. 13); movement of the magnet through the coil generates a measurable electrical current. At sea, *hydrophones* measure changes in water pressure caused by passing seismic waves.

d) Direct wave arrival

The compressional wave that goes directly from the source to a receiver is a body wave traveling very close to the surface (Fig 20, 21a). The velocity of the wave (V_1) is the distance from the source to the receiver (X) divided by the time it takes the wave to travel directly to the receiver (T_d):

$$\boxed{V_1 = \frac{X}{T_d}}$$

The equation for the strait line representing the direct arrival on a travel-time graph (Fig. 2lb) is therefore:

$$\boxed{T_d = \frac{X}{V_1}}$$

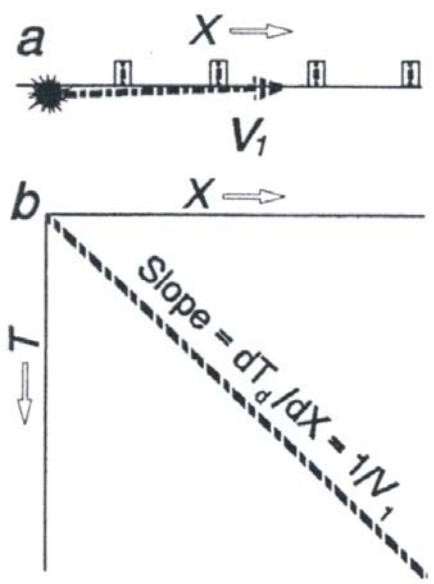


FIGURE 21 Selected raypath (a) and travel-time curve (b) for direct wave. The slope, or first derivative, is the reciprocal of the velocity (V_1).

Note that the velocity of the near-surface material can be determined by taking the inverse of the slope of the direct arrival (dT_d/dX) on the travel- time graph:

$$\boxed{\frac{dT_d}{dX} = \frac{1}{V_1}}$$

so that:

$$\boxed{V_1 = \frac{1}{dT_d/dX}}$$