Chapter 3
Seismic Fundamentals

Basic Concepts

It is necessary to introduce some basic concepts before discussing seismic methods. That is the purpose of this chapter

Seismic Waves

The principle of sound propagation, while it can be very complex, is familiar. Consider a pebble dropped in still water. When it hits the water's surface, ripples can be seen propagating away from the center in circular patterns that get progressively larger in diameter. A close look shows that the water particles do not physically travel away from where the pebble was dropped. Instead they displace adjacent particles vertically then return to their original positions. The energy imparted to the water by the pebble's dropping is transmitted along the surface of the water by continuous and progressive displacement of adjacent water particles. A similar process can be visualized in the vertical plane, indicating that wave propagation is a three-dimensional phenomenon.

Types of Seismic Waves

Sound propagates through the air as changes in air pressure. Air molecules are alternately compressed (compressions) and pulled apart (rarefactions) as sound travels through the air. This phenomenon is often called a sound wave but also as a compressional wave, a longitudinal wave, or a P-wave. The latter designation will be used most often in this book.

Figure 3.1 illustrates P-wave propagation. Darkened areas indicate compressions. The positions of the compression at times t_1 through $6t_1$ are shown from top to bottom. Note that the pulse propagates a distance d_p over a time of $6t_1 - t_1 = 5t_1$. The distance traveled divided by the time taken is the propagation velocity, symbolized V_p for P-waves.

M.R. Gadallah, R. Fisher, *Exploration Geophysics*,
DOI 10.1007/978-3-540-85160-8_3,

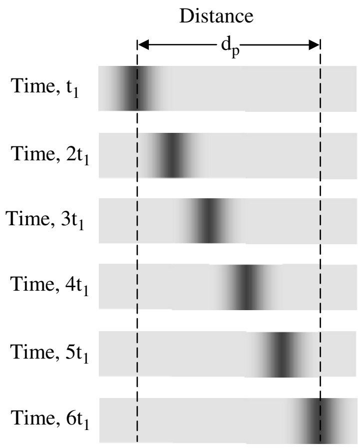

Fig. 3.1 Propagation of a P-wave pulse

P-waves can propagate in solids, liquids, and gasses. There is another kind of seismic wave that propagates only in solids. This is called a *shear wave* or an *S-wave*. The latter term is preferred in this book. Motion induced by the S-wave is perpendicular to the direction of propagation, i.e. – up and down or side-to-side.

Figure 3.2 illustrates propagation of an S-wave pulse. Note that the S-wave propagates a distance d_s in the time $5t_1$. The S-wave velocity, designated as V_s, is $d_s/5t_1$. Since d_s is less than d_p, it can be seen that $V_s, < V_p$. That is, S-waves propagate more slowly than P-waves.

Surface waves are another kind of seismic waves that exist at the boundary of the propagating medium. The *Rayleigh wave* is one kind of a surface wave. It exhibits a retrograde elliptical particle motion. Figure 3.3 shows motion of a particle over one period as a Rayleigh waves propagates from left to right. The Rayleigh wave is often recorded on seismic records taken on land. It is then usually called *ground roll*. Love waves are similar surface wave in which the particle motion is similar to S-waves. However, Love wave motion is only parallel to the surface.

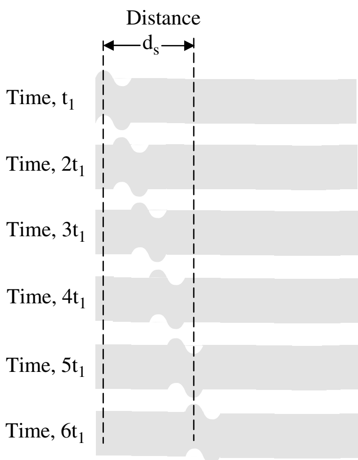

Fig. 3.2 Propagation of an S-wave pulse

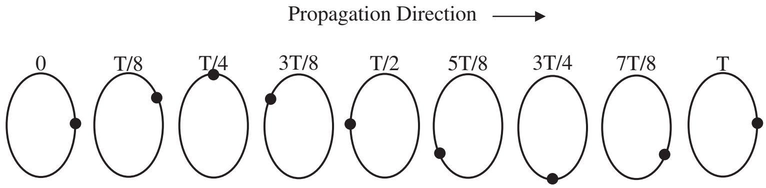

Fig. 3.3 Rayleigh wave particle motion

Seismic Wave Propagation

In comparing seismic wave propagation to the wave generated around a pebble thrown in the water, replace the pebble with a device such as an explosive or vibrator that introduces energy into the ground. This energy initially propagates as expanding spherical shells through the earth. A photograph of the traveling wave motion taken at a particular time would show a connected set of disturbances a certain distance from the source. This leading edge of the energy is called a *wave front.* Many investigations of seismic wave propagation in three dimensions are best done by the use of wavefronts.

Beginning at the source and connecting equivalent points on successive wave fronts by perpendicular lines, gives the directional description of wave propagation. The connecting lines form a *ray,* which is a simple representation of a three-dimensional phenomenon. Remember, when we use a ray diagram we are referring to the wave propagation in that particular direction; that is, the wave fronts are perpendicular to the ray at all points (see Fig. 3.4).

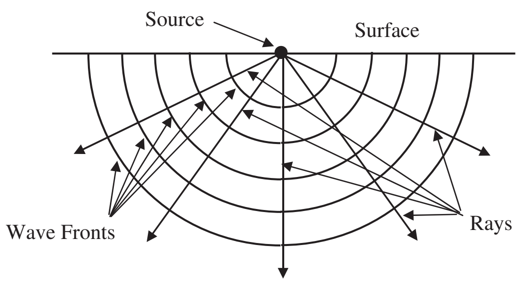

Fig. 3.4 Wave fronts and rays

Reflection and Refraction

As a first departure from the simplest earth model, consider a layered earth. What happens when an incident compressional wave strikes a boundary between two media with different velocities of wave propagation and/or different densities? Answer: Part of the energy is reflected from the boundary and the rest is transmitted into the next layer. The sum of the reflected and transmitted amplitudes is equal to the incident amplitude.

The relative sizes of the transmitted and reflected amplitudes depend on the contrast in *acoustic impedances* of the rocks on each side of the interface. While it is difficult to precisely relate acoustic impedance to actual rock properties, usually the harder the rocks the larger the acoustic impedance at their interface.

The acoustic impedance of a rock is determined by multiplying its density by its P-wave velocity, i.e., V. Acoustic impedance is generally designated as Z.

Consider a P-wave of amplitude A_0 that is normally incident on an interface between two layers having seismic impedances (product of velocity and density) of Z_1 and Z_2 (See Fig. 3.5). The result is a transmitted ray of amplitude A_2 that travels on through the interface in the same direction as the incident ray, and a reflected ray of amplitude A_1 that returns to the source along the path of the incident ray.

The reflection coefficient R is the ratio of the amplitude A_1 of the reflected ray to the amplitude A_o of the incident ray,

$$R = \frac{A_1}{A_0} \tag{3.1}$$

The magnitude and polarity of the reflection coefficient depends on the difference between seismic impedances of layers 1 and 2, Z_1 and Z_2. Large differences $(Z_2 - Z_1)$ in seismic impedances results in relatively large reflection coefficients. If the seismic impedance of layer 1 is larger than that of layer 2, the reflection coefficient is negative and the polarity of the reflected wave is reversed. Some Typical values of reflection coefficients for near-surface reflectors and some good subsurface reflectors are shown below:

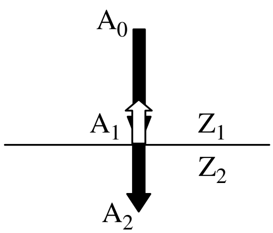

Fig. 3.5 Normal reflection and transmission

It can be seen in Table 3.1 that a soft, muddy ocean bottom reflects only about one-third of the incident energy, while a hard bottom reflects about two thirds of the energy.

The transmission coefficient is the ratio of the amplitude transmitted to the incident amplitude:

$$T = \frac{A_2}{A_0} = 1 - R \tag{3.2}$$

When a P-ray strikes an interface at an angle other than 90°, reflected and transmitted P-rays are generated as in the case of normal incidence. In such cases, however, some of the incident P-wave energy is converted into reflected and transmitted S-waves (see Fig. 3.6). The resulting S-waves, called SV waves, are polarized in the vertical plane. The Zoeppritz' equations are a relatively complex set of equations that allow calculation of the amplitudes of the two reflected and the two transmitted

Table 3.1 Typical reflection coefficients

Near-Surface Reflectors:	
Soft ocean bottom (sand/shale)	0.33
Hard ocean bottom	0.67
Base of weathered layer	0.63
Good Subsurface Reflectors	
Sand/shale versus limestone at 4,000 ft	0.21
Shale versus basement at 12,000 ft.	0.29
Gas sand versus shale at 4,000 ft.	0.23
Gas sand versus shale at 12,000 ft.	0.125

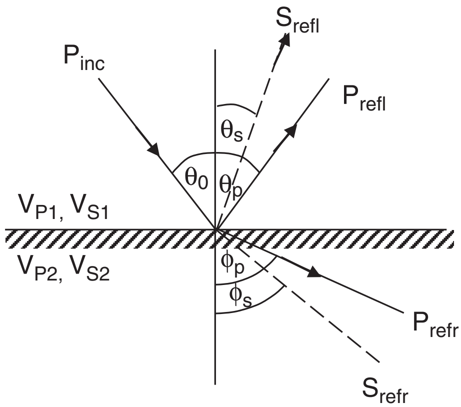

Fig. 3.6 Reflection and refraction of an incident P-wave. $V_{P2} > V_{S2} > V_{P1} > V_{S1}$

waves as functions of the angle of incidence. The equations require P- and S-wave velocities (V_{P2}, V_{S2}, V_{P1}, and V_{S1} in Fig. 3.6) plus densities on both sides of the boundary. The S-waves that are called converted rays contain information that can help identify fractured zones in reservoir rocks but this book will discuss compressional waves only.

Snell's Law

This relationship was originally developed in the study of optics. It does, however, apply equally well to seismic waves. Its major application is to determine angles of reflection and refraction from the incidence of seismic waves on layer boundaries at angles other than 90°.

Snell's law of reflection states that the angle at which a ray is reflected is equal to the angle of incidence. Both the angle of incidence and the angle of reflection are measured from the normal to the boundary between two layers having different seismic impedances.

The portion of incident energy that is transmitted through the boundary and into the second layer with changed direction of propagation is called *a refracted ray.* The direction of the refracted ray depends upon the ratio of the velocities in the two layers. If the velocity in layer 2 is faster than that of layer 1, the refracted ray is bent toward the horizontal. If the velocity in layer 2 is slower than that of layer 1, the refracted ray is bent toward the vertical.

Table 3.2 Snell's law relationships

Velocity relationship	Angle relationship
$V_{P2} > V_{S2} > V_{P1} > V_{S1}$	$\phi_p > \phi_s > \theta_0 > \theta_s$
$V_{P2} > V_{P1} > V_{S2} > V_{S1}$	$\phi_p > \theta_0 > \phi_s > \theta_s$
$V_{P1} > V_{P2} > V_{S1} > V_{S2}$	$\theta_0 > \phi_p > \theta_s > \phi_s$
$V_{P1} > V_{S1} > V_{P2} > V_{S2}$	$\theta_0 > \theta_s > \phi_p > \phi_s$

Figure 3.6 illustrates the more general condition for reflection and refraction. In this case both P- and S-wave velocities on each side of the interface are specified because reflected P- and S-waves and refracted P- and S-waves are generated from the incident P-wave. The two angles of reflection depend on the ratios V_{P1}/V_{P1} and V_{S1}/V_{P1}. The ratio of 1 for the reflected P-wave is a restatement of the angle of reflection equaling the angle of refraction for the P-wave. Since S-wave velocity is always slower than P-wave velocity the reflected S-wave always reflects at an angle less than that of the P-wave. The two angles of refraction depend on the ratios V_{P2}/V_{P1} and V_{S2}/V_{P1}. The relationships between angles of reflection and refraction with velocity ratio are not simple ones but depend upon the trigonometric function sine of the angles.

In Fig. 3.6 the relationships among the various velocities are: $V_{P2} > V_{S2} > V_{P1} > V_{S1}$. As a result the angles of refraction for both P- and S-waves are greater than the angle of incidence. There are, however, three other possible relationships. They are shown in Table 3.2, along with the corresponding relationships among angles of refraction. (Angles of reflection are not affected).

Critical Angle and Head Waves

From Table 3.2 it can be seen that when the P-wave velocity is higher in the underlying layer, the refracted P-ray is "bent" toward the boundary. As the angle of incidence increases the refracted P-ray will be bent to where it is just below and along the boundary, which means that the angle of refraction is 90°. The particular angle of incidence at which this occurs is known as the *critical angle,* usually designated θ_c. The sine of the critical angle is equal to the ratio of velocities across the boundary or interface.

This wave, known as *a head wave,* passes up obliquely through the upper layer toward the surface, as shown in Fig. 3.7.

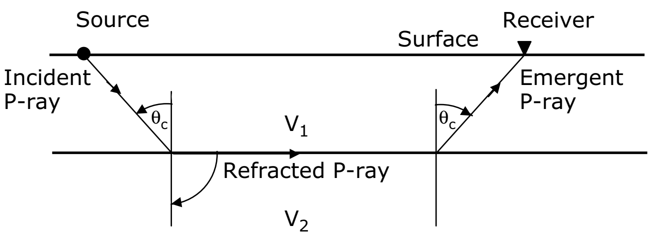

Fig. 3.7 Critical refraction/head wave

Fermat's Principle

A seismic pulse that travels in a medium follows a connected path between the source and a particular receiver. However, according to Fermat's principle there is the possibility of multiple travel paths. That means there may be more than one primary reflection event. The *buried focus* ("bow tie") effect is a classic example of Fermat's principle. On the left of Fig. 3.8 is the representation of a deep syncline and ray paths to and from seven coincident receivers and sources. There is only one path for rays numbed 1 and 7. There are two paths for rays 2, 3, 5 and 6. There are three paths for ray 4. On the right the arrival times are plotted vertically below the source/receivers. Note the crossing images and apparent anticline that results. This feature could be mistaken for a real anticline and a well that results in a dry hole.

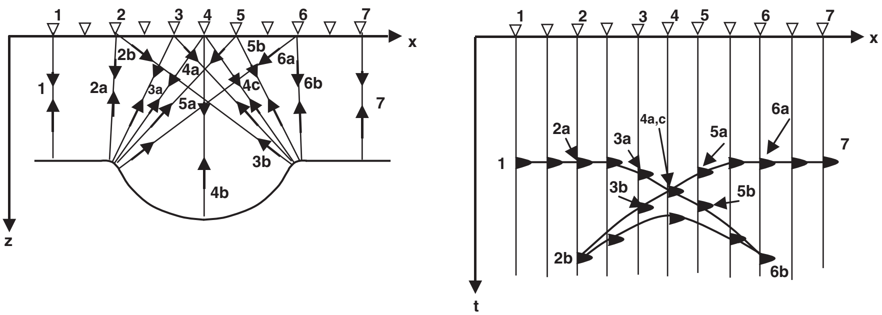

Fig. 3.8 On the *left* is a sketch of a deep syncline (buried focus) and reflection ray paths. On the *right* is its appearance on a seismic section (bowtie effect)

Huygens's Principle

This principle states that "Every point on an advancing wavefront is a new source of spherical waves". The position of the wave front at a later instant can be found by constructing a surface tangent to all secondary wavelets. See Fig. 3.9. Huygen's Principle provides a mechanism by which a propagating seismic pulse loses energy with depth.

Attenuation of Seismic Waves

As seismic waves propagate over greater and greater distances the amplitudes become smaller and smaller. That is, seismic waves are attenuated with the distance traveled. On a seismic record, this appears as attenuation with record time.

Even in a perfect medium, seismic waves are attenuated with distance. Consider the analogy of a balloon. Initially, the balloon is opaque. As the balloon becomes

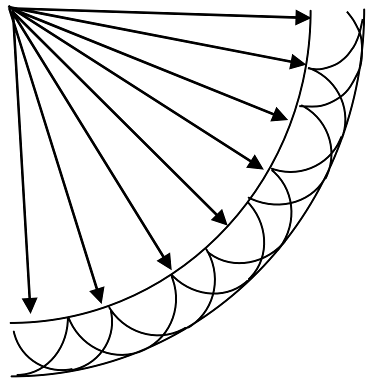

Fig. 3.9 Huygen's principle

more fully inflated its color becomes lighter until it is almost transparent. (See Fig. 3.10). This is because the balloon gets thinner and thinner as it gets bigger and bigger. There is just as much material in the balloon as before it was inflated, it's just thinner.

As seismic waves propagate away from the source the wavefront that describes the wave's advance becomes larger and larger. The energy gets spread over an ever-larger surface area. As a result, energy per unit area becomes smaller. Seismic amplitudes are proportional to the square root of energy per unit area so amplitudes get smaller even at a greater rate than the decrease in energy per unit area. This type of amplitude attenuation is called *spherical spreading* or *geometrical spreading.*

Another reason that seismic amplitudes get smaller is that rocks are not perfect conductors of seismic energy. Rocks are made up of individual particles or crystals. As a result, some of the energy becomes scattered. It does not all go in the main direction of propagation. In addition, because seismic wave propagation involves motion of particles, there is some "rubbing" of rock particles against one another. This results in some seismic energy being converted to heat. The higher the frequency of the seismic waves the greater the heat loss, and scattering, that occurs. This means that seismic wavelets become lower in frequency and longer in duration the farther they travel and hence, the later they arrive at the seismic detectors. This type of amplitude attenuation is called *inelastic attenuation.* Figure 3.11 illustrates the effect of both geometrical spreading and inelastic attenuation.

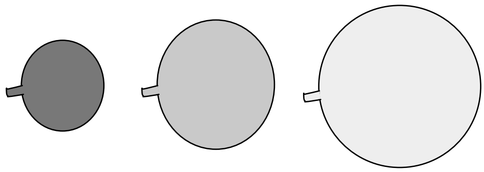

Fig. 3.10 Effect of balloon inflation

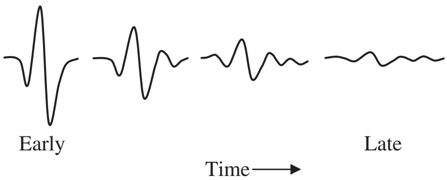

Fig. 3.11 Change in reflection amplitude with record time

Propagation Model for Exploration Seismology

Exploration seismology was developed to explore sedimentary basins that have gentle dip and layered structure with horizontal continuity over a large area. Simple models that include these essential features and propagating seismic pulses in these models enhance the understanding and interpreting of seismic records and sections.

The models adopted here assume that the seismic energy propagates along paths involving multiple receivers and multiple sources. The following propagation models will make it clear that the redundancy in sources and receivers allow estimation of needed velocity information.

Figure 3.12 is the simplest model considered. It consists of a single layer overlying a semi-infinite medium with the layer boundary being flat and horizontal. The thickness of the layer is Z and its propagation velocity has a constant value of V. This model can be used to calculate time required for energy to travel from the source to the receiver via reflection from the base of the layer.

There is an energy source at S and 12 receivers laid out at equal intervals, or offsets, from the source. Reflection raypaths are straight lines down to the base of the layer and straight lines up to the receivers. Reflection points are midway between source and receiver on the reflector. Reflection times are simply the total lengths of these pairs of lines divided by the velocity, V.

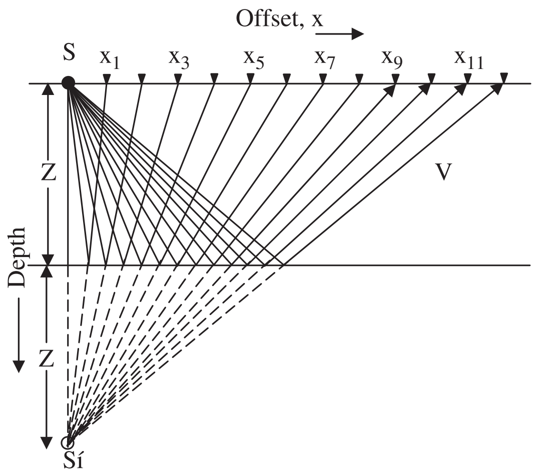

Fig. 3.12 Simple earth model

Constructing the image point of the source at S', which is at a depth Z below the boundary or a distance 2Z from the surface perpendicular to the surface, allows drawing of the dashed lines shown in Fig. 3.12. The lines from S' to x_1 through x_{12} are the same lengths as the two segments of the ray paths from S to x_1 through x_{12}. Figure 3.13(a) shows the lengths of the raypaths S'x_1 through S'x_{12} designated by d_1 through d_{12}. Note that while the lines increase in length with increasing offset, the rate of increase is not linear. If a curve is drawn connecting the ends of the lines representing reflection path length, it is found to be a curved line called a hyperbola.

As previously noted, for the constant velocity layer of Fig. 3.12, reflection times are given by dividing total path length, d, by velocity V. Thus, time for the reflection recorded at x_1 is $T_1 = d_1/V$. The time for the reflection recorded at $x_2$is $T_2 = d_2/V$. Times T_3 through T_{12} are calculated by dividing d_3 through d_{12} by V. Figure 3.13(b) plots the reflection times corresponding to the reflection raypaths of Fig. 3.12. Trace number corresponds to number of the receiver from which data were recorded.

The zero-offset time, T_0, is defined as the time required for a vertical reflection from the source to the base of the layer and back. Expressed as an equation, $T_0 = 2Z/V$. The reflection times T_1 through T_{12} are all greater than T_0. Thus, these times can be expressed as

$$T_j = T_0 + \Delta T_j, \quad j = 1, 2, \ldots, 12.$$

The quantity ΔT_j is called normal moveout or NMO and it depends on both the offset and velocity. It also has a hyperbolic shape. One of the important seismic data

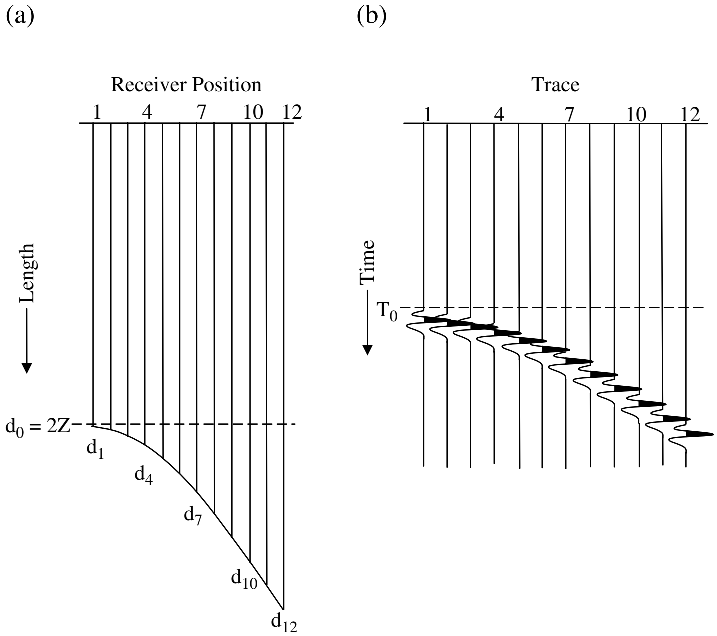

Fig. 3.13 (**a**) Reflection path lengths from Fig. 3.12 and (**b**) corresponding reflection times

processes is the correction for NMO, allowing true reflection times to be determined. Since the real earth is much more complex than the simple model of Fig. 3.12, and seismic velocities are not generally known beforehand, considerable effort is expended to extract velocity information from the data.

Summary and Discussion

Seismic waves propagate in three dimensions and following a seismic pulse through the earth is a difficult task. To better understand this propagation process, the pulses are followed through greatly simplified earth models.

Seismic waves occur as compressional waves, or P-waves, shear waves, or S-waves, and Rayleigh waves. P-waves are usually of greatest interest. S-waves can be used to obtain more detailed or special information about the subsurface. Rayleigh waves may be recorded on land seismic records as ground roll, an undesirable event and, hence referred to as "noise". P-waves propagate in solids, liquids and gasses. S-waves propagate only in solids. P-waves always have higher propagation velocities than S-waves, in the same medium.

Seismic energy that is input to the ground using an energy source such as an explosive (e.g.; dynamite) or a vibratory source (Vibroseis®*) energy propagates outward from the source in expanding spheres through the earth. Surfaces of these spheres are called wavefronts

Seismic rays indicate the paths that seismic waves take between two or more points in a medium. They are always perpendicular to the wavefronts. It should be remembered that when a ray diagram is presented, it implies wavefronts that are perpendicular to the ray at all points.

Parts of the earth of interest in petroleum exploration are made up of many layers, or strata, that have different geological and geophysical properties. Of particular interest are propagation velocity and density. The product of these two is called acoustic impedance. When a P-wave is incident on a boundary between these layers some energy is reflected and some is transmitted. Reflection and transmission coefficients are ratios of reflected and transmitted amplitudes.

The seismic method adapts to the theory of optics to study the propagation of the seismic energy in the earth. Snell's law of reflection and refraction is fundamental to understanding the seismic energy propagation. Huygens's principle provides a view of seismic energy propagation and attenuation. Fermat's principle introduces the possibility of multiple travel paths between source and receiver that may give rise to more than one primary reflection event.

Exercises

1. Name and describe three types of seismic waves described in this chapter.
2. Define the following terms:

a. Acoustic impedance
b. Snell's law of refraction
c. Critical angle

3. Table 3.3 lists densities and velocities of three layers. What can you infer about the magnitudes and polarities of reflection coefficient 1 (for the interface between layers 1 and 2) and reflection coefficient 2 (for the interface between layers 2 and 3)?

Table 3.3 Densities and velocities for earth model

Layer	Density (gm/cm^3)	Velocity (m/s)
1	2.2	1500
2	2.9	3000
3	2.6	2500

4. Consider two reflectors, or interfaces between two layers. In the first case, the velocity of the upper layer is 2.5 km/s and the velocity of the lower layer is 5.0 km/s. In the second case, the velocity of the upper layer is 3.25 km/s and the velocity of the lower layer is 4.75 km/s if a ray travels downward through the top layer at an angle of incidence of 20° in each case, which will result in a larger angle of refraction.
5. Below a flat, horizontal surface is a layer of 1500 m thickness that has a constant velocity of 2500 m/s. Twelve detectors are placed at 100 m intervals from the source. Table 3.4 lists total path length to each detector. Determine T_0 and NMO (ΔT) for traces corresponding to each detector. List answers in ms. (1 ms = 1000 s.)

Table 3.4 Reflection path lengths

Detector	Offset (m)	Reflection path lengths (m)	ΔT (ms)
1	100	3001.7	
2	200	3006.7	
3	300	3015.0	
4	400	3026.5	
5	500	3041.4	
6	600	3059.4	
7	700	3080.6	
8	800	3104.8	
9	900	3132.1	
10	1000	3162.3	
11	1100	3195.3	
12	1200	3236.1	

Bibliography

Bath, M.*Introduction to Seismology.*Basel-Stuttgart: Birkhauser Verlag, (1973).

Birch, F. “Compressibility, Elastic Constants.” S. P. Clark, ed., *Geological Society of America Memoir 97* (1966):97–173.

Dix, C. H. “Seismic Velocities from Surface Measurements.” *Geophysics* 20 (1955):68–86.

Ewing, M., W. Jardetzky, and F. Press. *Elastic Waves in Layered Media* New York: McGraw-Hill, (1957).

Faust, L. Y. “Seismic Velocity as a Function of Depth and Geological Time.” *Geophysics* 16 (1951):192–196.

Gardner, G. H. F., L. W. Gardner, and A. R. Gregory. “Formation Velocity and Density – the Diagnostic Basis for Stratigraphic Traps.” *Geophysics 39* (1974):770–780.

Koefoed, O. “Reflection and Transmission Coefficients for Plane Longitudinal Incident Waves.” *Geophysics Prospect 10* (1962):304–351.

Muskat, M., and M. W. Meres. “Reflection and Transmission Coefficients for Plane Waves in Elastic Media.” *Geophysics 5* (1940):115–148.

Sharma, P. V. *Geophysical Methods in Geology.* Amsterdam: Elsevier, (1976).

Sheriff, R E. “Addendum to Glossary of Terms used in Geophysical Exploration.” *Geophysics 34* (1969):255–270.

Telford, W. M., L. P. Geldart, R E. Sheriff, and D. A. Keys. *Applied Geophysics.*Cambridge: Cambridge University Press, (1976).

Trorey, A. W. “A Simple Theory for Seismic Diffractions.” *Geophysics 35* (1970).