Seismic Waves

Seismic waves are messengers that convey information about the earth's interior. Basically, these waves test the extent to which earth materials can be stretched or squeezed somewhat as you can squeeze a sponge. They cause the particles of material to vibrate, which means that these particles are temporarily stretched out of position as they move back and forth. The capacity of a material to be temporarily deformed by passing seismic waves can be described by its properties of *elasticity*. These physical properties

can be used to distinguish different materials. They influence the speeds of se
through those materials. In this chapter, we will discuss the elastic properties of ea
and the different kinds of seismic waves that can travel in these materials. Then we
how elastic properties influence the paths of seismic waves. Finally, we will look at
for recording seismic waves and the typical patterns of vibration displayed on seism

ELASTICITY

Stress and Strain

We can deform an object in different ways by stretching, squeezing, and twisting it. To accomplish this distortion, we must apply force on the sides of the object. Such application of force on the surface of an object is called *stress,* and it is expressed in units of force divided by area. The SI unit of stress is called a *pascal* and is equivalent to one newton (N) of force applied over a surface one meter square: $1\ Pa = 1\ N/m^2$.

When an object is placed at some depth in a pond or lake, the water presses against its sides from all directions. This is a *nondirected* stress which we call hydrostatic pressure. It is the pressure felt in the ears of a swimmer who dives deeply into the water. The fact that the swimmer feels the same pressure on both ears is evidence of equal stress in all directions. Other kinds of stress can be applied in a particular direction. An example of such a *directed* stress is the force of a hammer striking the head of a nail.

What is the effect of stress on an object? It deforms that object by changing its shape and size. The deformation is called *strain.* The different kinds of strain that can be produced depend on the strength and direction of stress and the nature of the substance being deformed. First, let us consider *elastic* strain. This kind of strain is proportional to the applied stress, and it disappears when that stress ceases. For example, we can apply stress by pulling a rubber band, and th
which it stretches is a measure o
The harder we pull, the farther
But when we release one end, th
ishes, and the rubber band snap
unstretched size. This is the natu
strain.

Something different happens
stretch a lump of soft clay.
stretched after we release it. This
deformation is evidence of *plastic*
substances ordinarily respond elas
atively weak, short-term stresses, a
hibit plastic strain if stronger or
longed stress is applied. Finally,
exceeds the strength of the mate
occurs. By pulling very hard, we c
rubber band or the lump of clay.

Exploration seismologists are
principally with elastic strain. To b
tic deformation and rupture are c
plosions or movements along
generate seismic waves. But this s
mation is confined to a small zone
point of energy release. As the se
travel away from this small sourc
produce only elastic strain. There
portant for us to examine the for
describing elastic strain.

Bulk Modulus

Suppose we placed a specimen o
stance in a fluid-filled container
that hydrostatic pressure could

Figure 2–1
Testing the bulk modulus in a pressure cell. An increase in pressure from P_1 to P_2 causes the volume of the specimen to decrease from V_1 to V_2.

(Figure 2–1). If the pressure is maintained at a value of P_1, the volume of the specimen is observed to be V_1. Now, if the pressure is increased by a small amount to P_2, the specimen becomes compressed into a slightly smaller volume V_2. This indicates the strain produced

by the change in stress. The ma
of the specimen remain unchang
Therefore, if the density of the sp
under pressure of P_1, an increase
to P_2 causes an increase in densit
on elastic substances show that th
volume and density are propor
change in pressure. The constan
tionality *(k)* is called the *bulk modu*
defined as

$$k = \frac{-\Delta P}{\Delta V/V_1} = \frac{\Delta P}{\Delta \rho/\rho_1}$$

where $\Delta P = P_1 - P_2$, $\Delta V = V_1 -$
$\rho_1 - \rho_2$. The bulk modulus is a m
capacity of a substance to be com
ferent values of this physical pro
used to distinguish one substan
other (Table 2–1). From Equation
that it is expressed in the same u
The value of bulk modulus can
for any kind of solid, liquid, or ga

Shear Modulus

Let us next consider a test of
which the shape of a specimen

TABLE 2–1 Elastic Properties of Selected Rock Specimens

SPECIMEN	BULK MODULUS (N/m²)	SHEAR MODULUS (N/m²)	YOUNG'S MODULUS (N/m²)
Sandstone, quartzitic	4.17×10^{10}	4.28×10^{10}	9.6×10^{10}
Limestone, Solenhofen, Bavaria, West Germany	4.67×10^{10}	2.47×10^{10}	6.3×10^{10}
Granite, Quincy, Mass.	5.21×10^{10}	3.45×10^{10}	8.49×10^{10}
Gabbro, French Creek, Penn.	8.85×10^{10}	4.80×10^{10}	10.43×10^{10}
Marble, Vermont	7.19×10^{10}	3.33×10^{10}	8.7×10^{10}

From Francis Birch, J. F. Schairer, and H. Cecil Spicer (editors), *Handbook of Physical Constants*, Geologica
America, Special Papers, No. 36, Table 5–8, p. 80, January 31, 1942, reprinted 1961.

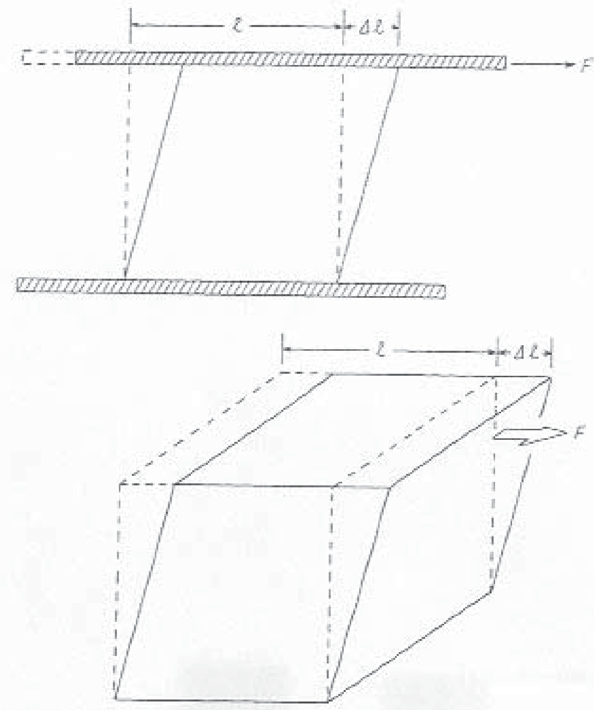

Figure 2–2
Testing shear modulus by distortion of a cube. One face of the cube is fixed, and the opposite parallel face is displaced by the small distance $\Delta\ell$ in response to the force of traction acting on that face.

We begin by attaching a cube of some substance between two parallel plates. Then force (F) is applied to displace one plate in a direction parallel to the other. Because one surface of the cube is attached to the plate, the force is applied over the entire area (A) of that surface. This creates a directed stress, $\tau = F/A$, which we call a *traction* or a *shear stress.* The cube becomes distorted as one side is shifted by the small distance $\Delta\ell$ relative to the opposite side (Figure 2–2). This distortion is an indication of shear strain. In an elastic cube of length ℓ, the small displacement will be proportional to the strength of the shear stress. The constant of proportionality (μ), called the *shear modulus,* is defined as

$$\mu = \frac{\tau}{\Delta\ell/\ell}$$

Shear stress cannot be applied t
and gases. For these substances
solids possess the physical prop
by the shear modulus. This pr
used to distinguish different sc
(Table 2–1).

Young's Modulus and Poisso

What happens to a specimen th
or compressed by a directed stre
can be tested by placing a cylindr
in a press (Figure 2–3). Here, fo
applied against the area (A) of t
specimen in the direction of the
This force produces a directed c
stress: $\eta = F/A$. Suppose that wh
has a value η_1, the length of the
ℓ_1, and its diameter is d_1. Now,
stress to a slightly higher value
that the length of the specimen
ℓ_2 and its diameter increases
change in length is proportional t
in stress for an elastic substance.
the constant of proportionality *Yo*
(E). It is defined as

$$E = \frac{-\Delta\eta}{\Delta\ell/\ell_1}$$

where $\Delta\eta = \eta_1 - \eta_2$ and $\Delta\ell =$
other constant, which we call *Poiss*
is used to show that the change
$\Delta d = d_1 - d_2$ is proportional to th
length:

$$\sigma = \frac{\Delta d/d_1}{\Delta\ell/\ell_1}$$

Young's modulus and Poisson'
the shear modulus, are constants t

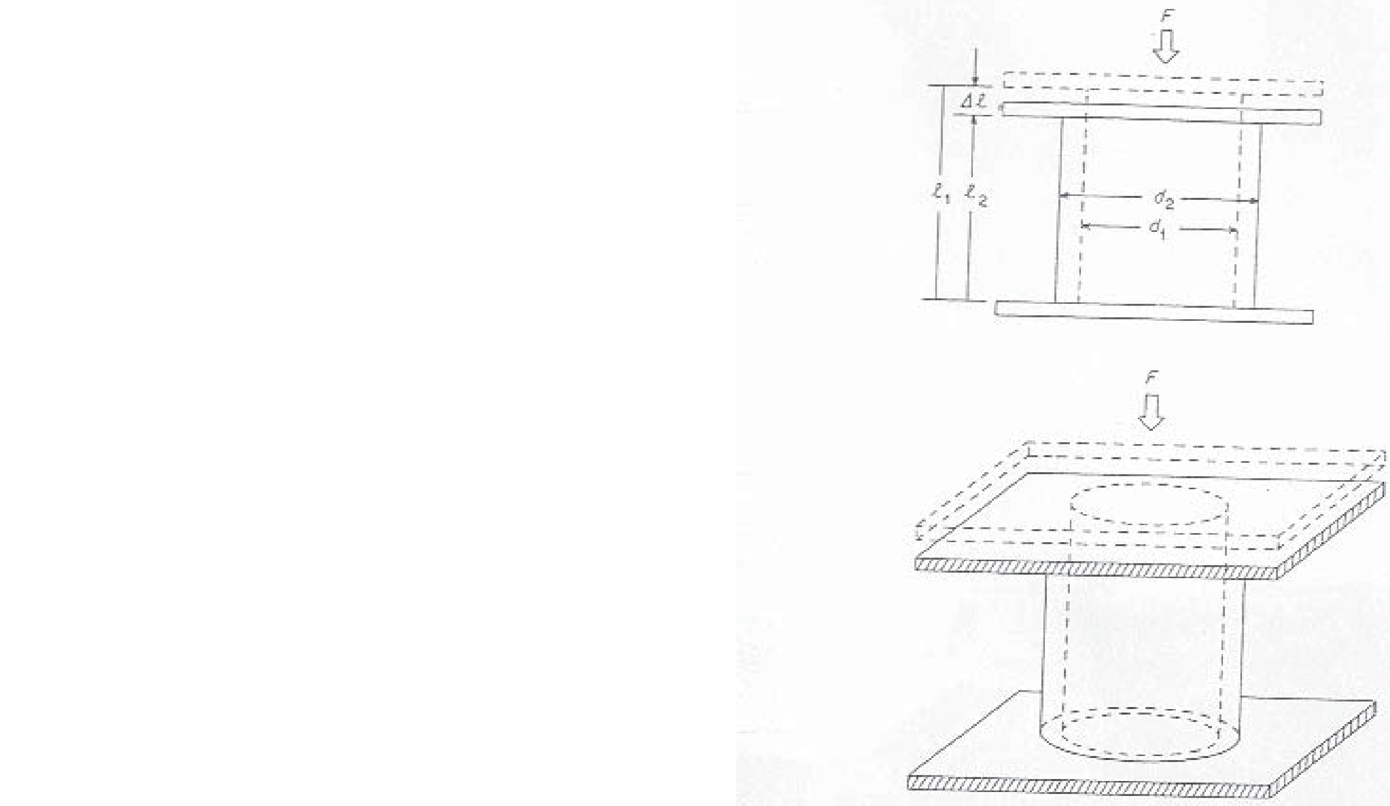

Figure 2–3

Testing Young's modulus and Poisson's ratio in a press. Young's modulus is determined from the change in length $\Delta\ell$, and Poisson's ratio is determined from the change in diameter d_2–d_1.

ful for describing the elastic properties of solids (Table 2–1). These constants cannot be used to describe liquids and gases for which elasticity is expressed solely by the bulk modulus.

SEISMIC BODY WAVES

We are more familiar with waves on ponds, lakes, and the ocean than with seismic waves. Tossing a pebble into a quiet pond produces ripples with certain features that are common to all waves. We will look at these features first

and then describe the particula
mic waves. The ripples that
pond are illustrated by the pr
2–4, which extends outward fro
impact of the pebble. The shap
surface changes from time t_1
cause the ripples are advancing
distance between two successive
some particular time is called
(λ). Observe the advance of one
crests from position r_1 at time t_1
at time t_2. This shows that the
wave is

$$V = \frac{r_2 - r_1}{t_2 - t_1}$$

The wave amplitude *(H)* is the
of a water particle above or belo
turbed surface of the pond. The
ing graph shows how the amplit
tion r_1 changes with time. The le
required for one cycle of oscilla
the wave period *(T)*. This is the
for a wave crest to advance a dis
wavelength (λ). Therefore, we
wavelength, period, and speed:

$$V = \lambda/T$$

The *frequency (f)* of the waves is th
oscillations that occur during a sta
val of time. It is the reciprocal of

$$f = 1/T$$

Waves on lakes and oceans com
periods of several seconds or tens
and wavelengths of several meter
Some earthquake waves have simi
and wavelengths. But exploration
gists usually work with wave perio
fractions of a second. These waves
quencies of tens or hundreds of cyc
ond. The unit of frequency commo
the *hertz,* which is one cycle of osc

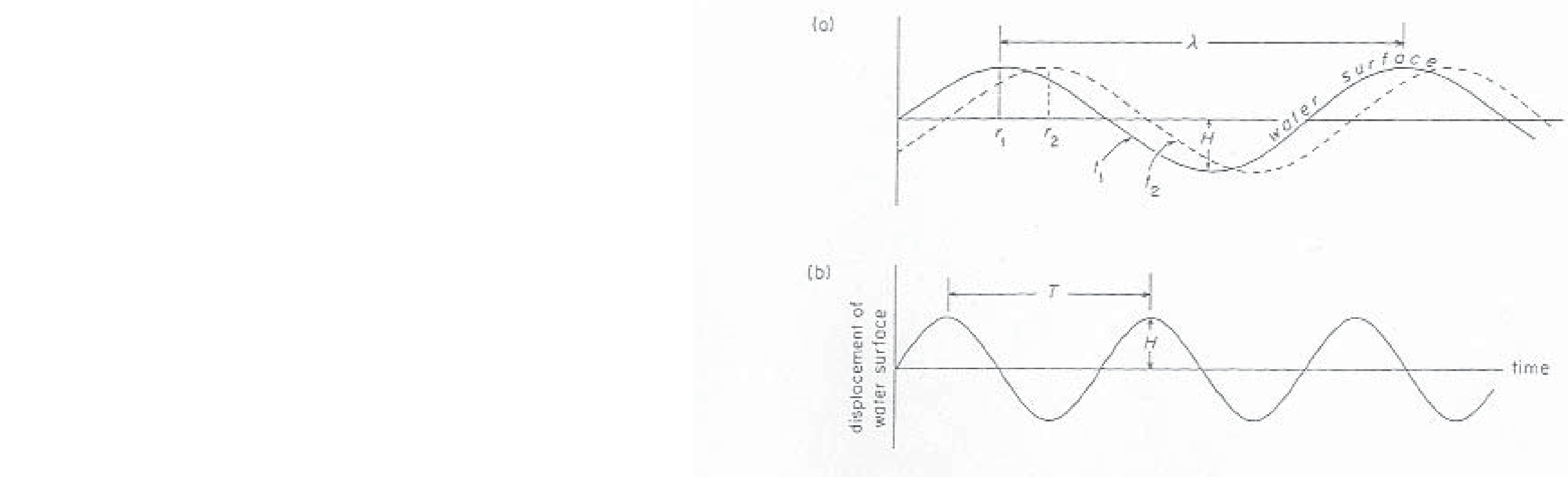

Figure 2–4
Waves produced
pebble into a po
the water surfac
extending outwa
point of impact f
(solid line), then
at time t_2 (dashe
Wavelength λ an
H describe the sh
surface. (b) Displ
point on the wat
or below the und
changes with tim
passing waves. T
T and wave heig
movement of thi

second. The particular kinds of waves of most interest to exploration seismologists can now be described.

Compressional Waves

There are different ways to make vibrations in a specimen of rock. Let us consider what vibrations will be produced by directly striking one side with a hammer. Suppose that we have some way to detect the movement of rock particles inside the specimen. Actually, tiny pressure-sensitive devices called transducers can be used for this purpose. At the instant of impact, the particles on that side will be displaced in the direction that the hammer was moving. For a brief moment, these particles will move back and forth in this direction. A little later they will cease moving, but other particles farther inside will move back and forth in the same direction. In this way, a pulse of vibration moves through the specimen, causing particles farther and farther from the point of impact to vibrate momentarily. This pulse of vibration is the seismic wave. It moves through the specimen first

compressing, and then stretch
from place to place, as we can
2–5. Observe that the vibrating
back and forth in the same di
path of the pulse through the
pulse causing this kind of vibra
compressional wave, a *longitudinal*
wave, all of which mean the sam

By placing transducers on bo
specimen to measure the inst
and the instant when the puls
far side, we can find the time (t_p
the *P*-wave to travel the distan
the specimen. From this infor
wave speed (V_p) can be calculate

$$V_p = x/t_p$$

Next, suppose that experiment
measure the density and elasti
the specimen. We could then
P-wave speed depends on these
the following way,

$$V_p = \sqrt{\frac{k + \frac{4}{3}\mu}{\rho}}$$

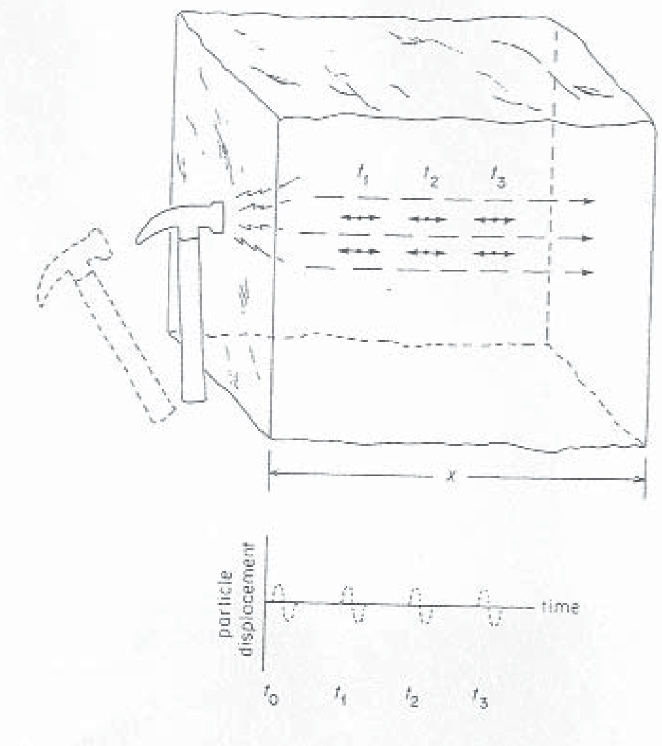

Figure 2–5
P-wave (also called a compressional
wave) pulse of vibration traveling th
specimen. A simple pulse of vibratio
on one side at time t_0 by the hamme
times, this wave pulse travels farther
specimen, causing particles to vibrat
forth in line with the direction of th
wave.

or

$$V_p = \sqrt{\frac{E}{\rho}\left[\frac{1 - \sigma}{(1 - 2\sigma)(1 + \sigma)}\right]} \quad (2\text{–}9b)$$

We can tell from these equations that *P*-waves will travel through any kind of substance: solid, liquid, or gas. They can do so because values of density and bulk modulus exist for all substances, even fluids for which the shear modulus is zero. Sound waves traveling through the air are *P*-waves.

Shear Waves

Can another kind of vibration be produced in a rock specimen? Rather than hammering directly on one side, suppose that we strike a glancing blow (Figure 2–6). At the instant of impact, the rock particles hit by the hammer will vibrate back and forth briefly in a direc-

tion parallel to this side. This p
tion will then move through
causing interior particles to vi
same direction. After a length of
pulse will have traveled the dista
other side of the specimen. In Fi
particles vibrate in a direction th
dicular, or transverse, to the path
We call this kind of vibration a *tr*
a *shear wave,* or an *S-wave.* Thes
interchangeable. The speed V_s o
through the specimen would be

$$V_s = x/t_s$$

If we had already measured the
erties of the specimen, we could v
lowing relationships between ela
sity, and *S*-wave speed,

$$V_s = \sqrt{\frac{\mu}{\rho}}$$

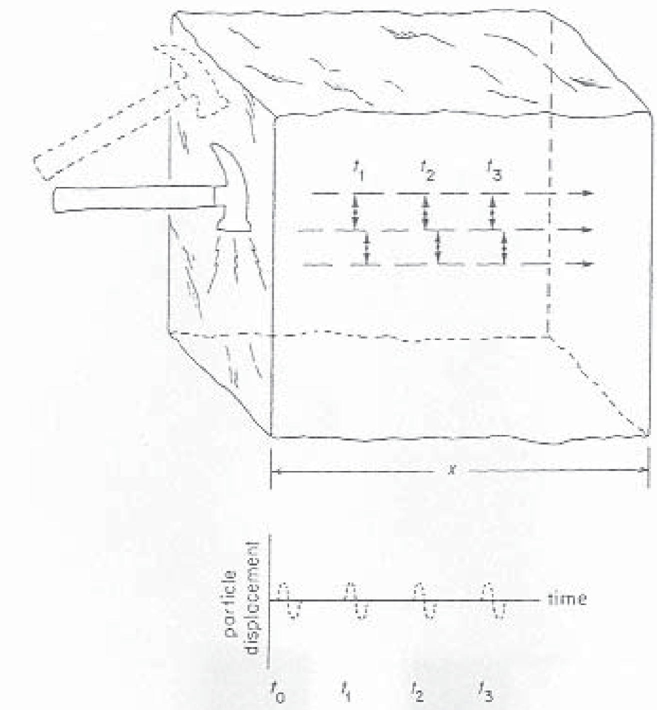

Figure 2–6
S-wave (also called a shear wave or
wave) pulse of vibration traveling t
specimen. This wave pulse causes p
vibrate back and forth in a line per
the direction of the advancing wave

or

$$V_s = \sqrt{\frac{E}{2\rho(1 + \sigma)}} \qquad (2\text{–}11\text{b})$$

We know that $\mu = 0$ for ideal liquids and gases. Therefore, Equation 2-11 clearly shows that *S*-waves do not travel in fluids. These transverse vibrations travel only in solids.

Body Waves

Both *P*- and *S*-waves are called *body waves* because they can travel directly through a mass of some substance in any direction. Other kinds of seismic waves called *surface waves* can travel only near the surface of such a mass, or close to the border between two different substances. The *P*- and *S*-waves are the most basic kinds of seismic waves. Surface waves can be explained in terms of combinations of these

body waves that interfere with c
distinctive ways when traveling
face of a mass. Surface waves wi
more fully later in this chapter.

We can learn some things a
and body wave speeds by com
tions 2–9 and 2–11 to get the ra

$$\frac{V_p}{V_s} = \sqrt{\frac{k}{\mu} + \frac{4}{3}$$

or

$$\frac{V_p}{V_s} = \sqrt{\frac{1 - \sigma}{\frac{1}{2} - \sigma$$

The fact that both k and μ are
bers indicates that $V_p/V_s > 1$. I
true, the *P*-wave must always tra
the *S*-wave through the same m

is to say that $V_p > V_s$. It also restricts the positive range of values for Poisson's ratio: $0 < \sigma < 1/2$.

REFRACTION AND REFLECTION OF SEISMIC BODY WAVES

Rays and Wave Fronts

Think again about the ripples made by tossing a pebble into a quiet pond. Each wave moves away from the point of impact in an ever-expanding circle. The circle at the outer edge of an advancing wave marks the *wave front.* Lines drawn outward are *rays* that show directions along which the wave is advancing. Now,

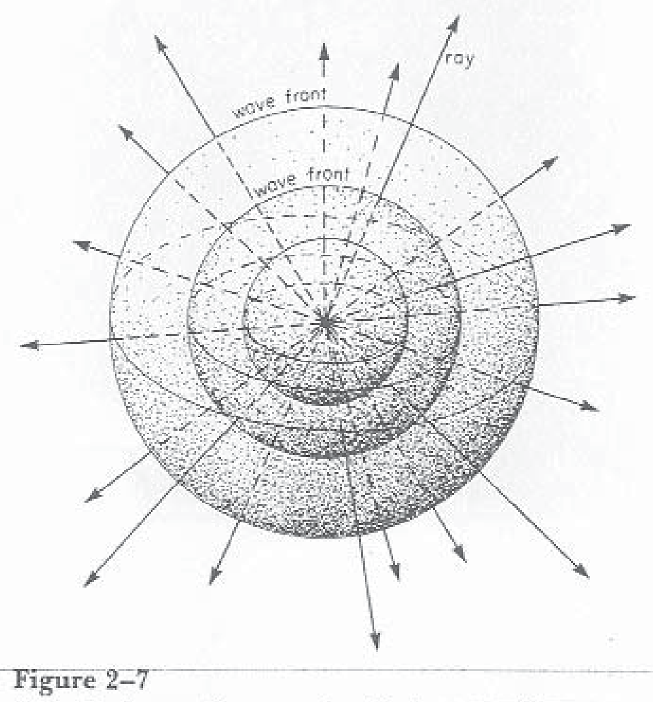

Figure 2–7
Spherical wave fronts and radiating rays illustrate the advance of waves outward in all directions from a source. The wave fronts show the position of a wave at successively later times t_1, t_2, t_3, and t_4.

rather than the surface of a pond
waves produced at a point insi
rock (Figure 2–7), perhaps by a
sion. The movement of these w
trated by spherical wave fronts
radiate outward in all directio
source. Observe that the rays ar
perpendicular to the wave fronts

Wave Conversions

What happens when a wave reac
ary between two substances in wh
speeds are different? It divides in
bounce, or *reflect,* from the bo
other waves that pass, or *refrac*
boundary. The rays in Figure 2
the movement of these different
original wave that travels to the
called the *incident* wave. At the bo
converted into *reflected* and *refra*
and *S*-waves that travel away from
ary in different directions.

Observe that an incident *P*-w
into *S*-waves as well as *P*-waves (F
An important feature of these *S*-
the transverse particle motion occ
plane that is perpendicular to th
and that contains the wave path.
zontal boundary, this will be a ve
and the shear waves causing vibra
plane are called *SV*-waves. An *SV*
incident on a boundary converts i
and refracted *SV*-waves and *P*-w
2–8b).

A shear wave that produces
transverse particle motion is ca
wave. If this kind of horizontal
shear wave is incident on a boun
verts into reflected and refracte
(Figure 2–8c) with no associated *P*

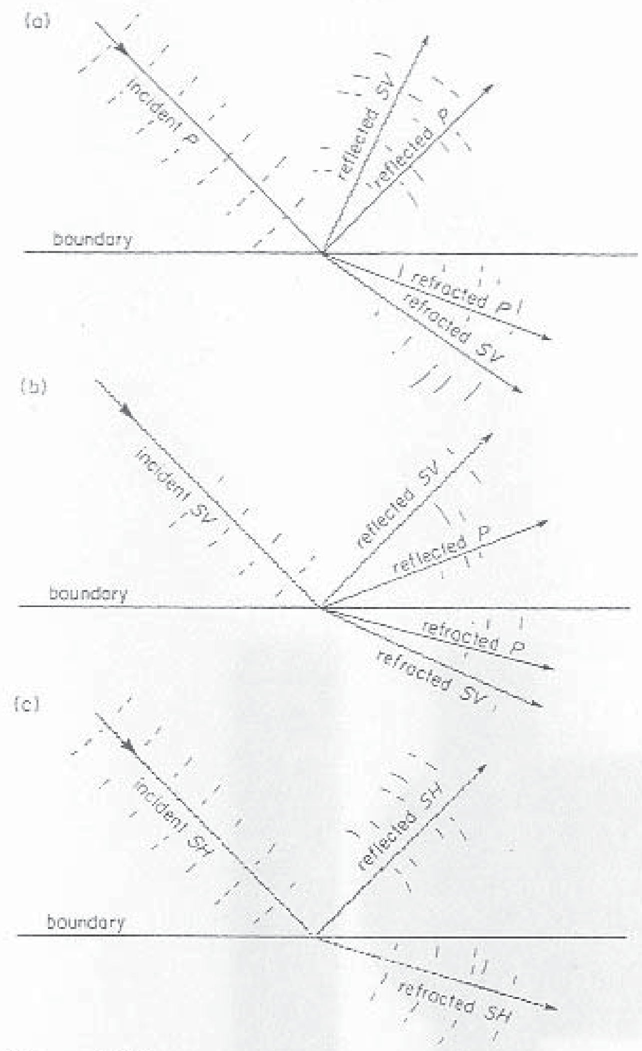

Figure 2–8

Wave conversions at the boundary between two rock layers. Each incident *P*-wave ray or *SV*-wave ray divides into four reflected and refracted *P*- and *SV*-wave rays. An incident *SH*-wave ray produces only reflected and refracted *SH*-wave rays.

Snell's Law

The directions of refracted and reflected waves traveling away from a boundary depend on the direction of the incident wave and the speeds of the waves. To explain
enon, let us suppose that the s
dent *P*-waves is situated far fro
ary. One implication is that two
together are almost parallel at th
they reach the boundary. Anoth
is that the portion of a wave f
between these two rays is so s
that it is nearly a straight line
things further, we will look on
and refracted *P*-waves, ignoring
ment other reflected and refract
waves.

Now examine in Figure 2–9
rays and straight wave fronts clos
ary. The incident wave moves a
takes the same time for one poin
front to move from A to C as i
another point to move from B t

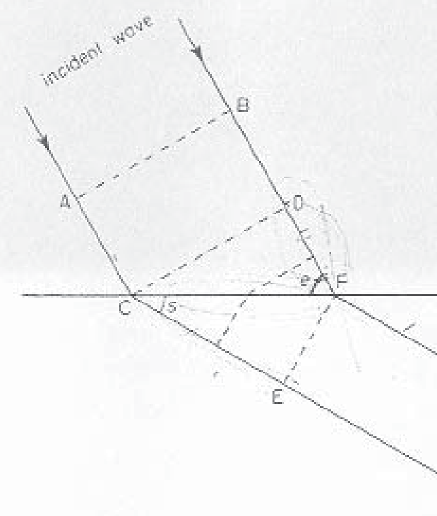

Figure 2–9

Wave refraction at the boundary bet
layers. Notice that one point on the
advances from D to F at speed V_1 du
time that another point on the same
advances from C to E at speed V_2. T
bends at the boundary as it moves al

both parts of the wave front are moving at speed V_1. After C is reached, that point on the wave front refracts into a new direction and travels at speed V_2 toward E. Let t be the interval of time needed for the wave at that point to travel from C to E. Then, by rearranging Equation 2–8, we can show that the distance CE $= V_2t$. During the same interval of time, the other point continues from D to F at speed V_1. This requires that the distance DF $= V_1t$. Observe that the wave front bends where it crosses the boundary.

The incident rays are inclined from the boundary at the angle e. Because rays are perpendicular to wave fronts, the figure CDF is a right triangle. Therefore, from trigonometry we recognize that $\cos e$ = DF/CF = V_1t/CF. This expression can be rearranged so that

$$\mathrm{CF} = V_1t/\cos e \qquad (2\text{–}13a)$$

The refracted rays are inclined at the angle s from the boundary. Since the figure CEF is a right triangle, we see that $\cos s$ = CE/CF = V_2t/CF. We can rearrange this expression to get

$$\mathrm{CF} = V_2t/\cos s \qquad (2\text{–}13b)$$

Combining and rearranging Equations 2–13a and 2–13b shows how wave speeds and directions are related:

$$\frac{\cos e}{\cos s} = \frac{V_1}{V_2} \qquad (2\text{–}14)$$

Next, let us consider the relationship between a reflected wave and the incident wave that produced it. The case of an incident *P*-wave and the reflected *SV*-wave is illustrated in Figure 2–10, where the same letters that appear in Figure 2–9 are used. Now reread the previous two paragraphs, making the following changes. Replace the word "refract" with the word "reflect" and change V_1 and V_2

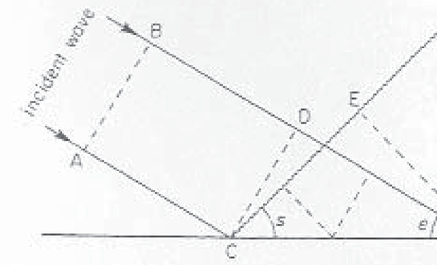

Figure 2–10
Example of wave reflection at the b
between two rock layers in which a
wave is produced by an incident *P*-
reflected *P*-wave (not shown) would
produced. Notice that a point on th
wave front travels from D to F duri
time that a point on the reflecting *S*
travels from C to E.

to V_p and V_s. We will see that E
also shows how incident and r
speeds and directions are relate
understand that an incident *P*-
produce a reflected P-wave (Fig
travels at the same speed. For th
tion 2–11 shows that the inci
flected rays are inclined from the
the same angle.

It is a more common practice
physicists to draw a line perpend
boundary and to measure the a
incident ray and the angle r of
or reflected ray from this line (
We will follow this practice and
tion 2–14 in the form

$$\frac{\sin i}{\sin r} = \frac{V_i}{V_r}$$

where V_i is the speed of the incid
V_r is the speed of the refracted
wave. This equation is called *S*
governs the directions of all refra

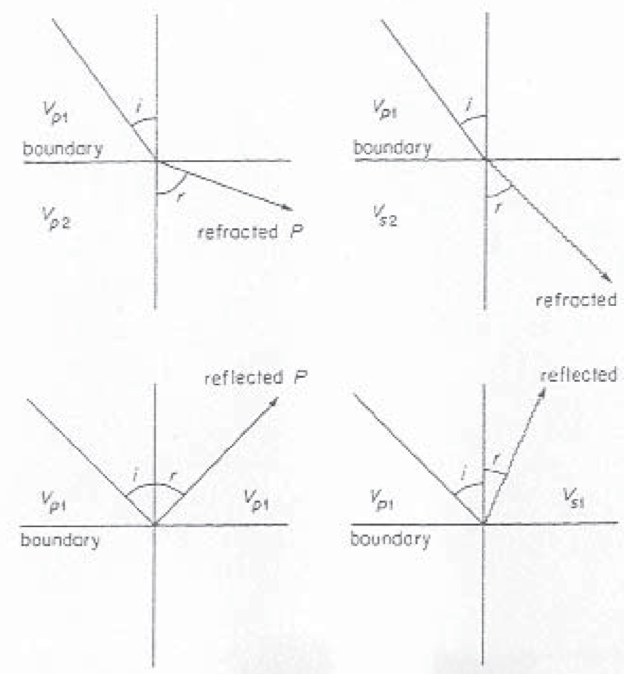

Figure 2–11
Angles of incidence (*i*) and reflection and refraction (*r*) measured from a line perpendicular to the boundary between two rock layers are shown separately for the four *P*- and *SV*-waves that are produced by an incident *P*-wave.

flected *P*- and *S*-waves produced by an incident *P*-wave or an incident *S*-wave.

Critical Refraction

Let us look further at how seismic waves cross a boundary between two different kinds of rock. To keep things as simple as possible, we will first consider only incident and refracted *P*-waves. In Figure 2–12a, rays from a source in the upper layer reach the boundary at different angles of incidence and then continue in the lower layer at different angles of refraction in accordance with Snell's law. Observe that the angle refraction is 90 degrees for one particular ray. It is the *critically* that shows how the wave travels right along the top of the lower l

The critically refracted wave is an incident wave traveling along reaches the boundary at the *critic cidence* i_c. For this case of critic note that $\sin r = \sin 90° = 1$, so ing to Snell's law we can write

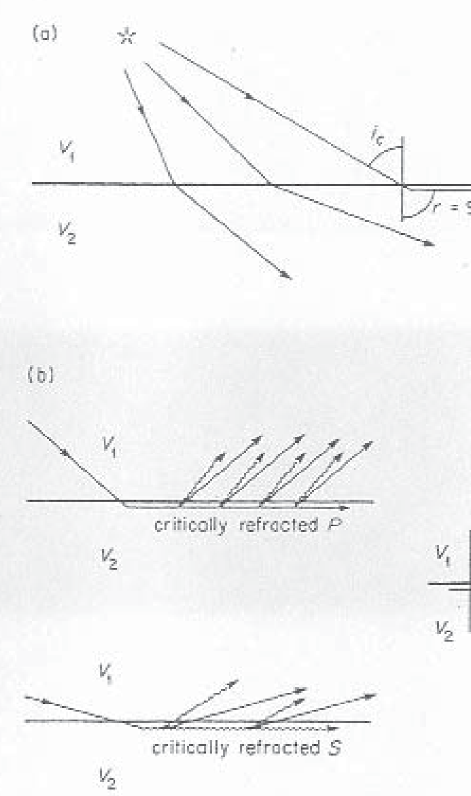

Figure 2–12
Refraction of rays at the boundary b layers includes (a) a critically refracte which the angle of refraction $r = 90$ ray is produced from an incident ray boundary at the critical angle of inci Critically refracted *P*- and *SV*-waves along the ray produce *P*- and *SV*-wa refract across the boundary at angles

$$\sin i_c = V_1/V_2 \qquad (2\text{–}16)$$

which shows how the critical angle depends on the wave speeds. It is obvious that the angle of refraction (r = 90 degrees) is larger than the critical angle of incidence. We can tell from Snell's law that this condition is possible only if the refracted wave travels faster than the incident wave. Recall from Figure 2–8 that refracted *P*- and *SV*-waves are both produced by either an incident *P*-wave or an incident *SV*-wave. Critical refraction can result from any of these combinations if the refracted wave travels faster than the incident wave.

The critically refracted wave does an interesting thing as it travels close to a boundary between two layers. Figure 2–12b shows how a critically refracted *P*-wave traveling at speed V_{p2} along the top of the lower layer continually produces *P*- and *SV*-waves that refract across the boundary. They travel at speeds V_{p1} and V_{s1} in the upper layer. Let i_p and i_s be the angles of the upward directed rays. They are related to the wave speeds according to Snell's law in the same way as Equation 2–16 relates angles of incident and critically refracted rays: $\sin i_p = V_{p1}/V_{p2}$ and $\sin i_s = V_{s1}/V_{p2}$.

A critically refracted *SV*-wave traveling along the top of the lower layer at speed V_{s2} can also produce *P*- and *SV*-waves that refract into the upper layer (Fig. 2–12b). For this case, the angles are related to wave speeds in the following way: $\sin i_p = V_{p1}/V_{s2}$ and $\sin i_s = V_{s1}/V_{s2}$. The *P*-wave is generated only if $V_{p1} < V_{s2}$.

How does a critically refracted wave continually produce the upward traveling waves illustrated in Figure 2–12? A complete answer would be very complicated. We can glean some understanding from *Huygen's principle*, a statement which asserts that every point on a wave front is a source of new waves that travel

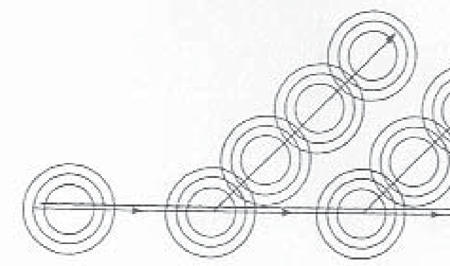

Figure 2–13
Huygen's principle asserts that each
wave front is a source of spherically
waves. This is shown by spherical w
spreading from different points alo
advancing waves.

away from it in all directions. T lustrated in Figure 2–13, whic wave fronts spreading out from old wave front situated close to between two layers.

Paths of Seismic Body Waves

The purpose of exploration seis discover what lies underground b surements made on the earth's mologists detect seismic waves p source close to the surface that ent paths into the earth, and th the surface. Rays from the source directions and refract and ref boundary between different subs cordance with Snell's law. Which will reach a detector located so away from the source?

First, consider a very simple st sisting of two horizontal layers (F in which the wave speeds are V $V_{s2} > V_{s1}$. Suppose that only *P*-wa duced at the source. We can see extends along the land surface

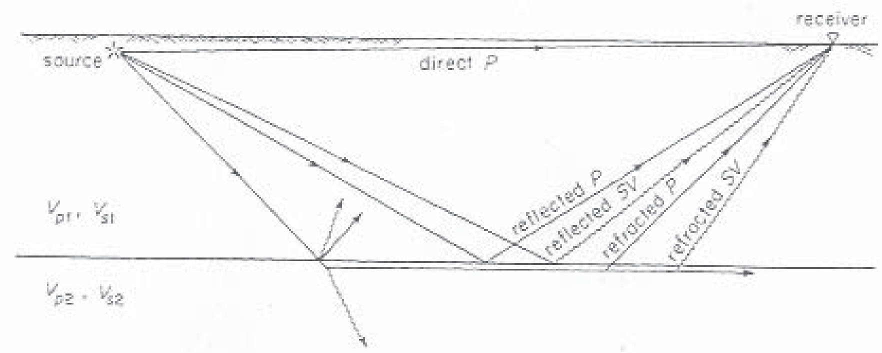

Figure 2–14
Paths followed by s
from a source to a
the idealized structu
of only two rock lay
the wave speeds V_{p}
$V_{p1} > V_{s2} > V_{s1}$. Ea
consistent with Snel

path of the *direct wave.* Other rays mark paths of the reflected *P*- and *SV*-waves produced by different incident *P*-waves. Still another incident *P*-wave produces the critically refracted ray from which *P*- and *SV*-waves refract to the surface. Altogether, there are five different paths leading to the detector.

Next, look at a structure with several horizontal layers (Figure 2–15). We know that each ray divides into four refracted and reflected *P*- and *SV*-wave rays when it reaches a boundary. Clearly, there will be a large number of paths leading to a detector. By omitting all *SV*-wave rays from the diagram and showing only refracted and reflected *P*-waves, we still have many rays converging
tor.

How much time is required
wave to follow one of these path
interval can be calculated from
if we know the length of the p
wave speed in each layer. For exa
tal time t needed for a wave to tr
ABCD in Figure 2–16 would be

$$t = t_{AB} + t_{BC} + t_{CD}$$
$$= \frac{AB}{V_{p1}} + \frac{BC}{V_{p2}} +$$

Suppose for the moment that
were not constrained by Snell's l
follow other paths as well. What i
be refracted along various paths

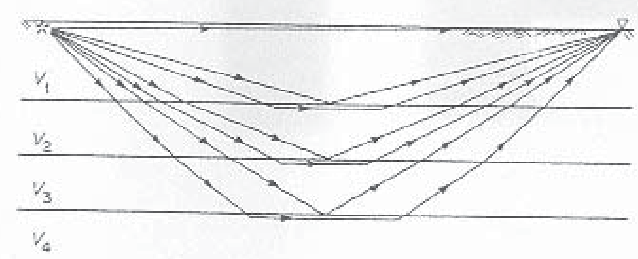

Figure 2–15
Paths followed by *P*-waves from a source to a receiver in a structure consisting of four parallel layers in which the wave speeds $V_{p1} < V_{p2} < V_{p3} < V_{p4}$. The paths of *SV*-waves that also would be produced are not shown in this example.

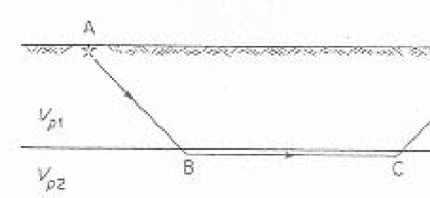

Figure 2–16
Segments of the path of the refracted
traveling from a source at A to a rece
Wave speed is V_{p1} along segments AB
Along the segment BC, the wave trav
V_{p2}.

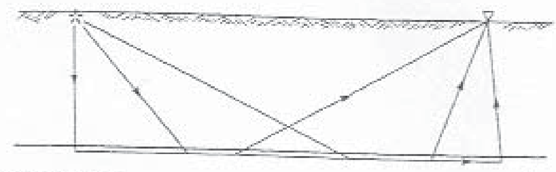

Figure 2–17
Alternate wave paths for testing Fermat's principle. By comparing the travel times along these different paths, we can find the smallest value for the path that is consistent with Snell's law.

ones illustrated in Figure 2–17? Calculating the travel times for these different routes reveals that the particular path requiring the least time is the path predicted from Snell's law. But we know that seismic waves are governed by Snell's law, which means that they do follow paths of minimum travel time. A statement of this fact is called *Fermat's principle,* which asserts that elastic waves travel between two points along paths requiring the least time.

SEISMIC SURFACE WAVES

Recall our discussion of the nature of vibrations produced by passing seismic body waves. As a *P*-wave pulse moves through a rock layer, particles vibrate back and forth in the direction of the ray that marks the path of the wave. The vibrations produced by an *S*-wave are perpendicular to the ray pointing out its path. Material close to the earth's surface experiences these kinds of *P*- and *S*-wave vibrations and other more complicated patterns of vibration as well. These additional kinds of vibration can be measured only at locations close to the surface. Instruments placed away from the surface in boreholes
them. Such vibrations must resu
that follow paths close to the e
We call them seismic surface wa

Rayleigh Waves

Vibrations produced by seismic
are separated into two types.
ground movement in a vertical
with the path of the wave (Figur
type of vibration is produced by
surface wave bearing the name o
J. W. S. Rayleigh (1842–1919), w
portant contributions to our und
elasticity. Unlike body waves t
simple movements back and fo
straight line, passing Rayleigh w
point on the ground to move a
shaped like an ellipse.

Another important difference

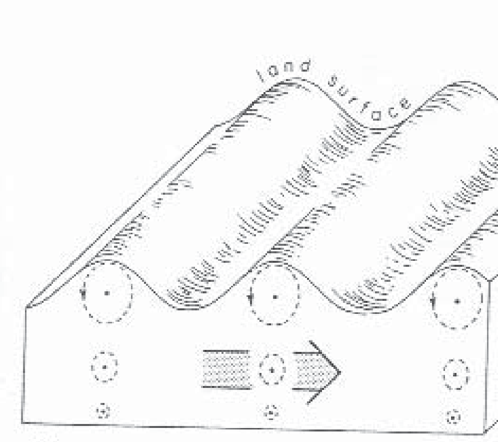

Figure 2–18
Ground vibration produced by Rayleig
Notice that the amplitude of vibration
with depth. As the waves pass, a point
surface moves through an elliptical orb
in a vertical plane.

tion of the Rayleigh wave vibration. Ordinarily, a passing body wave pulse will cause a point on the ground to move back and forth a few times before coming to rest. But the Rayleigh wave vibration tends to persist much longer. The point on the ground will cycle around its elliptical path many times.

Rayleigh wave vibrations change with time in an interesting way. As the wave travels past, the ground vibrates slowly at first and then more and more rapidly. The wave frequency increases with time. As we have noted previously, this kind of vibration can be detected close to the earth's surface but becomes weaker with depth below the surface, rapidly diminishing to a level that cannot be measured.

The Rayleigh wave travels at a speed that is slower than the direct *S*-wave. It follows a path along the earth's surface.

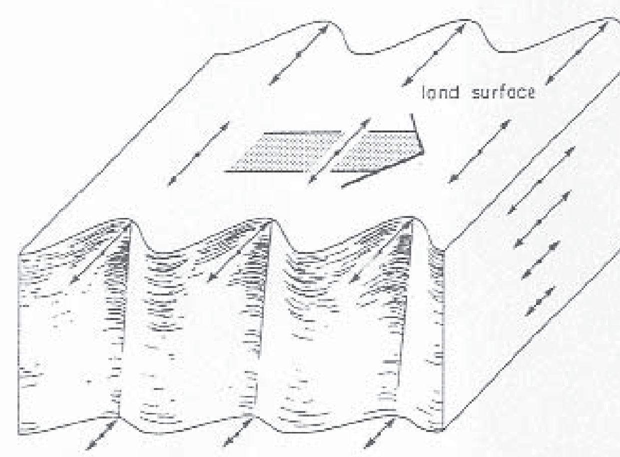

Figure 2–19
Ground vibration produced by Love waves. A point on the land surface cycles back and forth on a horizontal line perpendicular to the direction in which the wave is moving. Amplitude of this horizontal ground movement diminishes with depth.

Love Waves

The other type of seismic su
named after A. E. H. Love (186
oneer geophysicist. Like the R
the Love Wave travels at a slow
the direct *S*-wave. But the direc
bration is different. A passin
causes horizontal ground move
dicular to the path of the wave (
And, unlike an *S*-wave, wh
ground movement in the same
Love wave vibration persists mu
lowing many cycles of oscillation
of this persistent Love wave vib
lar to Rayleigh wave ground mo
beginning slowly, the oscillatio
grows faster as the wave pas
words, frequency increases with

Wave Guides

In this discussion, we will not gi
theoretical explanation of se
waves. But we can gain some u
by looking at *P*- and *S*-waves
trapped in layers close to the e

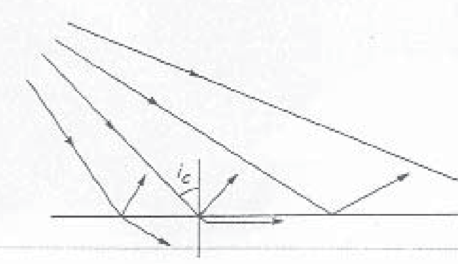

Figure 2–20
Rays reaching a boundary at angles
the critical angle of incidence are to
No refracted waves are produced.

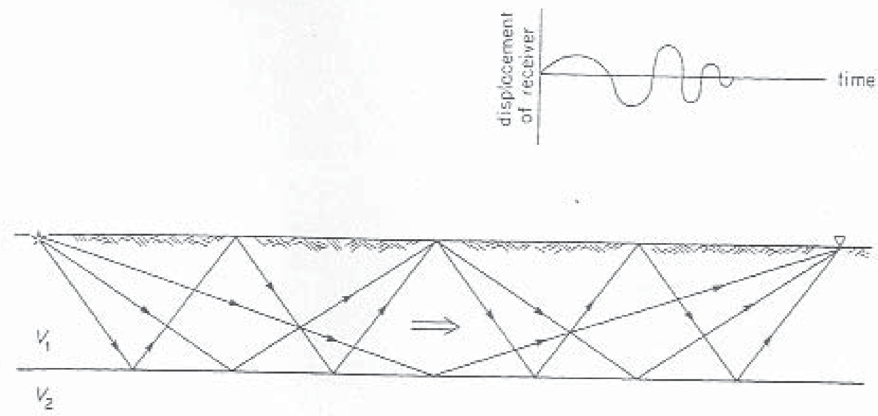

Figure 2–21
Body waves that a[r]
reflected at the up[p]
boundaries of a "w[aveguide"]
layer can interact t[o]
surface waves.

We have already seen what happens to body wave rays that are incident on a boundary at angles less than or equal to the critical angle i_c (Figure 2–11). These divide into refracted and reflected *P*- and *SV*-waves. But what about a ray with an angle of incidence larger than i_c? It produces only a reflected wave! No wave will refract across the boundary (Figure 2–20). What does this have to do with surface waves?

Typically, layers very close to the earth's surface are "low-speed" layers. Body waves travel more slowly in these layers of soil and weathered rock than in the underlying harder rock. Figure 2–21 shows what happens to *P*- and *S*-waves that are incident on the base of a single "low-speed" layer at angles larger than the critical angle i_c. These waves remain trapped in the low-speed layer, repeatedly reflecting from its base and its upper surface. Because the waves are guided along paths completely within the layer and cannot escape from it, the low-speed layer is called a *waveguide*.

Look at the individual paths in Figure 2–21 of waves traveling in the waveguide from a source to a seismometer. The length of a path depends on the number of times it is reflected. Shorter paths have fewer reflections. Because more time is needed to travel a l
succession of waves will reach t
ter. Each successively later wave
lowed a path with one more re
the base of the waveguide.

Now suppose that each wave a
of oscillation to the ground mov
seismometer. The succession of r
flected waves will then produc
quence of oscillations. *P*-waves
this fashion contribute to the osc
we recognize as Rayleigh waves
reflected *SH*-waves produce th
sequence of vibrations that w
waves.

SEISMOGRAMS

Seismic waves follow many differe
tween a source and a receiver l
distance away. At the receiver, the
tions are recorded on a chart th
seismogram. The pattern of vib
played on a seismogram depends
ture of the waves produced at the
on how the size of a wave chang
path.

The Source Wavelet

Suppose that an explosive is detonated at some point in the earth. This is a particularly effective way to produce *P*-waves. Severe deformation occurs in the zone very close to the charge. Rock is shattered and partly melted in this small *source zone.* Outside the source zone, however, there is only momentary elastic deformation indicated by vibrations.

Let us examine the vibration just outside the source zone. In Figure 2–22, *P*-wave rays extend outward in all directions. An individual particle vibrates back and forth in the direction of a ray. This motion is described on the

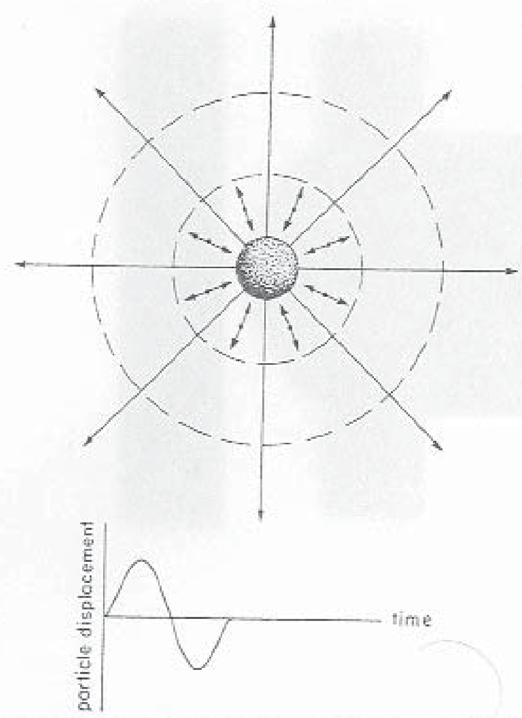

Figure 2–22
Wavelet produced in the source zone is the pulse of vibration that travels outward in all directions. In this idealized example, the source wavelet consists of a single cycle of oscillation.

accompanying graph, which sh
displacement of the particle fro
position changes with time. The
this brief pulse of vibration is ca
wavelet. It advances outward fro
zone in all directions.

We can think of the graph in
as the seismogram that would b
a location very close to the source
it shows the "shape" of the sourc
the example, it has a simple f
only one cycle of oscillation. So
of more complicated shape are
Chapter 5.

Geometrical Spreading and A

As a source wavelet travels farthe
from the source, its amplitude
grows smaller. The ground vibrat
by the energy of the wave, which
the source. After the wavelet leave
zone, it receives no more energy
vancing wavelet has a continuall
spherical wave front (Figure 2–7).
front expands, the same quantit
must be distributed over a larg
surface. This means that less ene
able for each particle on that surfa
available to other particles closer t
where the wave front was smaller.
particles farther from the source
smaller amplitudes than particles c
source. This is the result of the ef
geometrical spreading.

How much change in wave am
we expect from geometrical sprea
surface area of a sphere with radiu
The energy available to a particle
dividing the total energy Ω by th
$4\pi x^2$. It is possible to prove that the
of the particle vibration is prop

$\sqrt{\Omega}$. This means that amplitude decreases directly with distance from the source,

$$H = H_0/x \qquad (2\text{–}18)$$

where H_0 is the amplitude of the wavelet leaving the source zone.

When something begins to vibrate, it usually begins to grow warmer at the same time. Heat is produced by the internal friction of particles rubbing together during the process of vibration. In this way, some of the energy of a seismic wave is converted into heat. We say that this wave energy has been absorbed by the material through which the wavelet is traveling. Energy used in this way is not available to cause vibration. So the effect we call *absorption* causes the amplitude of vibration to decrease with distance from the source.

Although we do not understand the process of absorption very well, we have learned from experiments that its effect on wave amplitude can be predicted from the formula

$$H = H_0 e^{-\alpha x} \qquad (2\text{–}19)$$

where α is called the *absorption coefficient*, and $e = 2.71828$ is the Napierian constant. The value of α is different for different materials.

We can combine the equations for geometrical spreading and absorption to predict how the amplitude of a seismic wave should change as it travels away from the source:

$$H = H_0 e^{-\alpha x}/x \qquad (2\text{–}20)$$

Transmission and Reflection Coefficients

The amplitude of a seismic wave is changed by reflection and refraction. Recall from Figure 2–8 that a wave incident on a boundary divides into reflected and refracted *P*- and *SV*-waves. Because the energy must be shared between these reflected and r
they will have smaller amplitu
cident wave.

Ratios that compare the am
flected and refracted waves to
of the incident wave can be c
pose that an incident wave of
produces reflected *P*- and *SV*-
plitudes of H_{1p} and H_{1s}. The ra

$$R_{1p} = H_{1p}/H_0 \quad \text{and} \quad R_{1s} = H$$

are called the *reflection coefficients*
cident wave also produces ref
SV-waves with amplitudes H_{2p}
ratios

$$T_{2p} = H_{2p}/H_0 \quad \text{and} \quad T_{2s} = H_2$$

are called either *transmission co*
fraction coefficients.

Values of transmission and r
ficients depend on the angle
densities of the two layers, a
speeds in these layers. For most
cidence, the values of R_{1p} and
larger than R_{1s} and T_{2s} for an inc
The opposite is true for an inci
where R_{1s} and T_{2s} tend to be lar
and T_{2p}. If the angle of inciden
grees for a *P*-wave, then $R_{1s} = R$

$$R_{1p} = \frac{\rho_2 V_{2p} - \rho_1 V_{1p}}{\rho_2 V_{2p} + \rho_1 V_{1p}}, \qquad T_{2p} = \text{–}$$

More complicated formulas are
compute the transmission and re
ficients when the angle of incid
zero.

Vibrations at a Receiver

What pattern of vibration do se
from a source some distance away
a receiver? Suppose that a wavel

form (Figure 2–22) is produced at the source. Look again at Figures 2–14 and 2–21, which show that many different paths lead this simple pulse of vibration from the source to a receiver. Because of the different distances and wave speeds along these various paths, pulses of vibration reach the receiver at different times. The effects of geometrical spreading, absorption, refraction, and reflection are different along each of these paths. Therefore, the pulses of vibration will have different amplitudes when they reach the receiver.

Now suppose that we plot a graph of the ground vibration at the receiver that would be produced by waves following the paths in Figures 2–14 and 2–21. The result is the *seismogram* seen in Figure 2–23. It consists of several discrete pulses followed by a continuous series of oscillations. Each pulse has the same shape as the source wavelet, but a different amplitude. These pulses indicate waves that followed the paths shown in Figure 2–14. The continuous oscillations that follow are surface waves related to paths seen in Figure 2–21.

The example in Figure 2–23 shows several

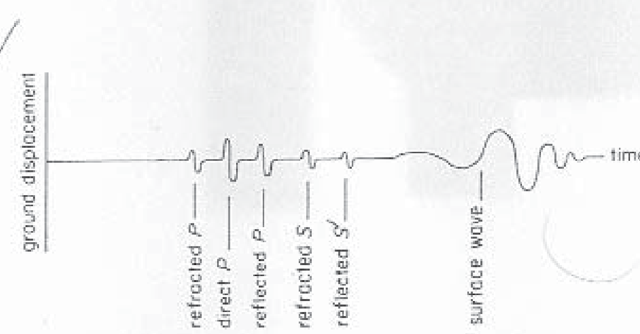

Figure 2–23
Seismogram consisting of body wave pulses, each having the same shape as the source wavelet, and surface wave oscillations with frequency that increases with time. The pulses represent the ground vibration produced by body waves that followed the paths in Figure 2–14.

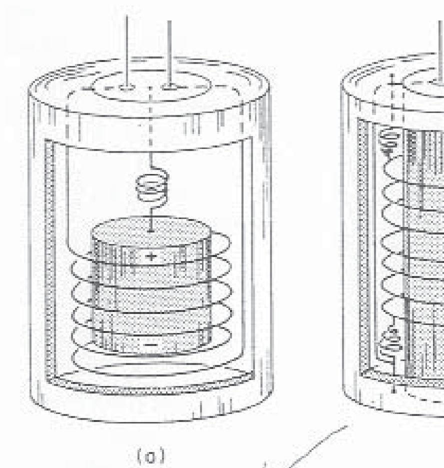

Figure 2–24
A typical geophone (a) has a magnet
spring that can move inside a coil wh
ground vibrates. Movement of a mag
causes an electric current to flow in t
An alternate design consists of a coil
springs that encloses a magnet fixed
container.

pulses that are clearly separated
other. The exploration seismolog
presented with a somewhat more
pattern of vibrations. In nature,
lets are more irregular, and they
through many more layers. Like
however, these more complic
grams consist of discrete pulses
one for each different path throu
together with surface wave oscill

Recording Seismic Waves

The exploration seismologist de
vibrations with a device that is c
seismometer or a *geophone,* an instr
usually about the size of a fist. T
signs are illustrated in Figure 2-
sign consists of a coil of wire r

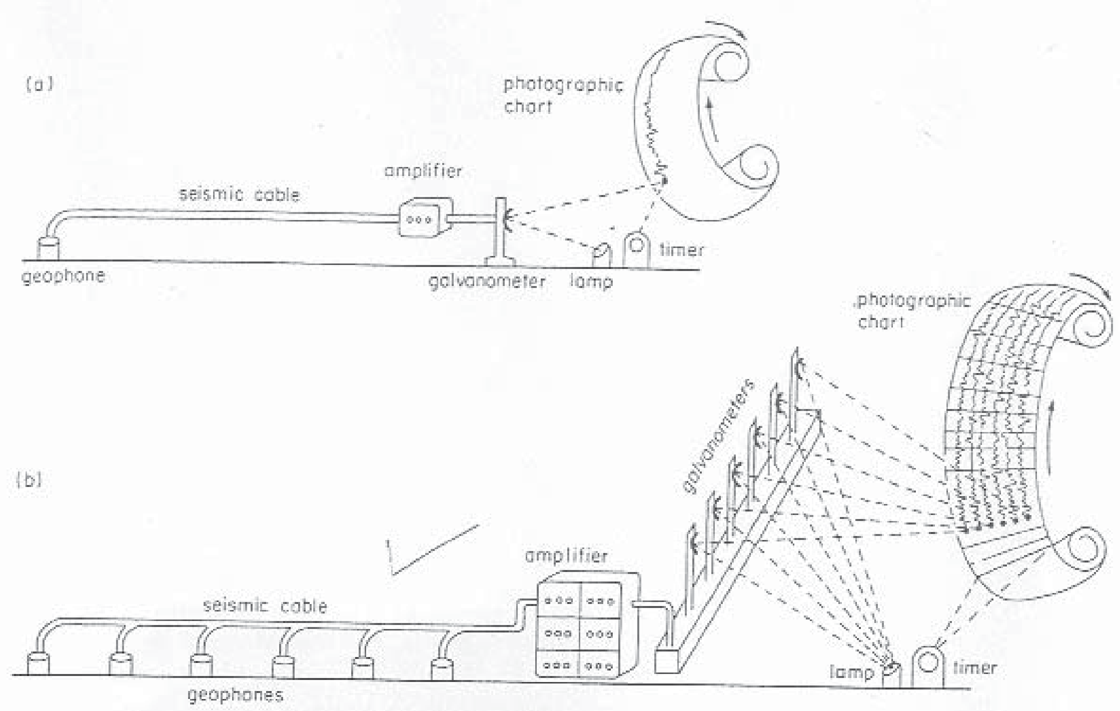

Figure 2–25

Optical–electronic recording of seismic waves. (a) Electric current from the geophone is transmitted through the seismic cable to an amplifier. Then the amplified current activates the galvanometer mirror, which reflects a beam of light onto an advancing photographic chart. (b) A six-channel exploration seismic system with a separate amplifier and galvanometer circuit for recording the electric signal from each geophone.

cylindrical container. A magnetized mass attached to a spring is suspended inside the coil. Ground vibration causes movement of the mass within the coil. Because the mass is a magnet, an electric current that is related to the amplitude of ground vibration is generated in the coil. In the second basic design, the magnet can be fixed to the container and the coil mounted on the spring.

The electric current from a geophone is carried through a line called the *seismic cable* to a recorder. In older seismic recorders, this current was first amplified electronically and then transmitted to a galvanometer (Figure 2–25). This device contains a coil that rotates on a suspension in response to an rent. A concave mirror attached to cuses a point of light from a nearb a chart of photographic paper. As mirror rotate back and forth and t per advances, usually at a speed centimeters per second, an irregul posed on the chart. This irregular ing all the pulses of ground vibr seismogram. Marks are also pla chart by a timing device.

The most common practice in seismology is to record simultaneo same chart the vibrations detecte 24, 48, or 96 separate geophones

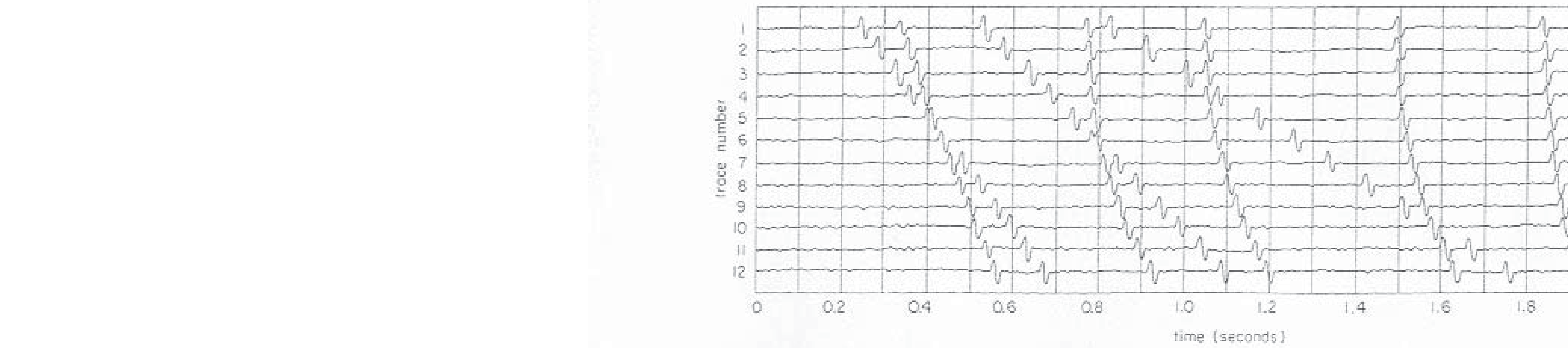

Figure 2–26
A 12-trace exploration seismogram showing pulses of seismic waves refracted and reflected from several rock layers. Vertical time lines on the chart provide the reference for determ which these various pulses arrive geophones.

25). The galvanometers in the recording system are mounted in a row. Lines crossing the chart are exposed every 0.01 second by the timing device. The result is an exploration seismogram (Figure 2–26).

Each geophone–amplifier–galvanometer unit of the system functions independently of the others. Such a unit is one *channel* of an exploration seismic system. The ground vibrations detected by this channel are displayed by one of the irregular lines on the exploration seismogram. Each irregular line is called a *trace*. A 12-channel seismic system, which is larger but otherwise similar to the 6-channel system in Figure 2–25, would produce the 12-trace seismogram shown in Figure 2–26. Modern seismic systems with 6, 12, 24, 48, and 96 channels are described more fully in Chapter 5. These systems are equipped with digital magnetic tape recorders as w recorders we have described. tal magnetic tape are more data processing by compute cussed in Chapter 6.

How does a geophysicis pulses of ground vibration th exploration seismogram? By from a line of geophones on we can recognize different pulses. Observe in Figure 2 pulses lie along straight lines record at different angles. T from refracted waves. Metho them are presented in Cha analysis of reflected waves is Chapter 4. In Figure 2–26, ref indicated by the pulses that lines.

STUDY EXERCISES

1. Suppose that a *P*-wave ray produced by an explosion detonated 5 m beneath the ocean surface travels down through the water and reaches solid rock on the ocean bottom at an angle of incidence i = 30 degrees. *P*-wave speed is 1500 m/s in the water and 5000 m/s in the rock.

   a. How many reflected and refracted rays are generated by this incident ray?

   b. What is the angle of refraction of the refracted *P*-wave ray?

2. An *SH*-wave originating in a solid layer where the shear modulus is μ_1 and the density is ρ_1 refracts across a boundary into another solid layer where the shear modulus μ_2 is equal to μ_1, but where the density ρ_2 is greater than ρ_1. This implies that the angle of incidence is (a) larger than or (b) smaller than the angle of incidence. Explain!

3. Consider two solid layers, one resting on the other. In the top layer, the *P*-wave speed V_{p1} = 4000 m/s and the *S*-wave speed V_{s1} = 2200 m/s, and in the deeper layer the *P*-wave speed V_{p2} = 4000 m/s and Poisson's ratio σ = 0.25. Suppose that an *SH*-wave originating in t fracts into the deeper layer.

   a. Is the angle of refraction la than the angle of incidence? Exp

   b. If the density in the deeper la g/cm^3, what are the values of shea bulk modulus in that layer?

4. A *P*-wave traveling at a speed of 3 layer refracts across a boundary layer where its speed increases t the frequency of vibration is 30 H length of the refracted *P*-wave lon than the wavelength of the inciden plain!

5. An incident *P*-wave traveling at a s m/s is critically refracted at a bou critical angle is 30 degrees, what is the refracted *P*-wave?

6. At a distance of 100 m from a sou plitude of a *P*-wave is 0.1000 mm, tance of 150 m the amplitude d 0.0665 mm. What is the absorptio of the rock through which the wave

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