$$t_{\text{int}_1} = \frac{2h_1 \cos i_{c_1}}{v_1}$$

giving:

$$h_1 = \frac{v_1 t_{\text{int}_1}}{2 \cos i_{c_1}} \qquad \text{Eq. 6.4}$$

Its value is put in the equation for the intercept time for the second interface, which can be shown to be

$$t_{\text{int}_2} = \frac{2h_1 \cos i_{c_1}}{v_1} + \frac{2h_2 \cos i_{c_2}}{v_2} \qquad \text{Eq. 6.5}$$

which gives h_2. This is repeated for successive interfaces, the equation for the interface time increasing by one term each time.

6.4 Dipping interfaces

Tilting the interface does not change the value of the critical angle, but it rotates the ray diagram of Figure 6.3 by the angle of dip, α (Fig. 6.8). As a result, rays to successive receivers C', D', E', . . . not only have to travel the additional distance *CD*, *DE*, . . . along the interface, but also an extra distance up to the surface because the interface is getting deeper. Therefore the refracted line on the *t*–*x* diagram is steeper, its slope yielding a velocity less (slower) than v_2.

This presents a problem: There is no way to tell from Figure 6.8 that the refracted line is due to a dipping interface and not a horizontal interface over a layer with a slower seismic velocity equal to 1/slope. For this reason, velocities calculated from the slopes are called **apparent velocities.** However,

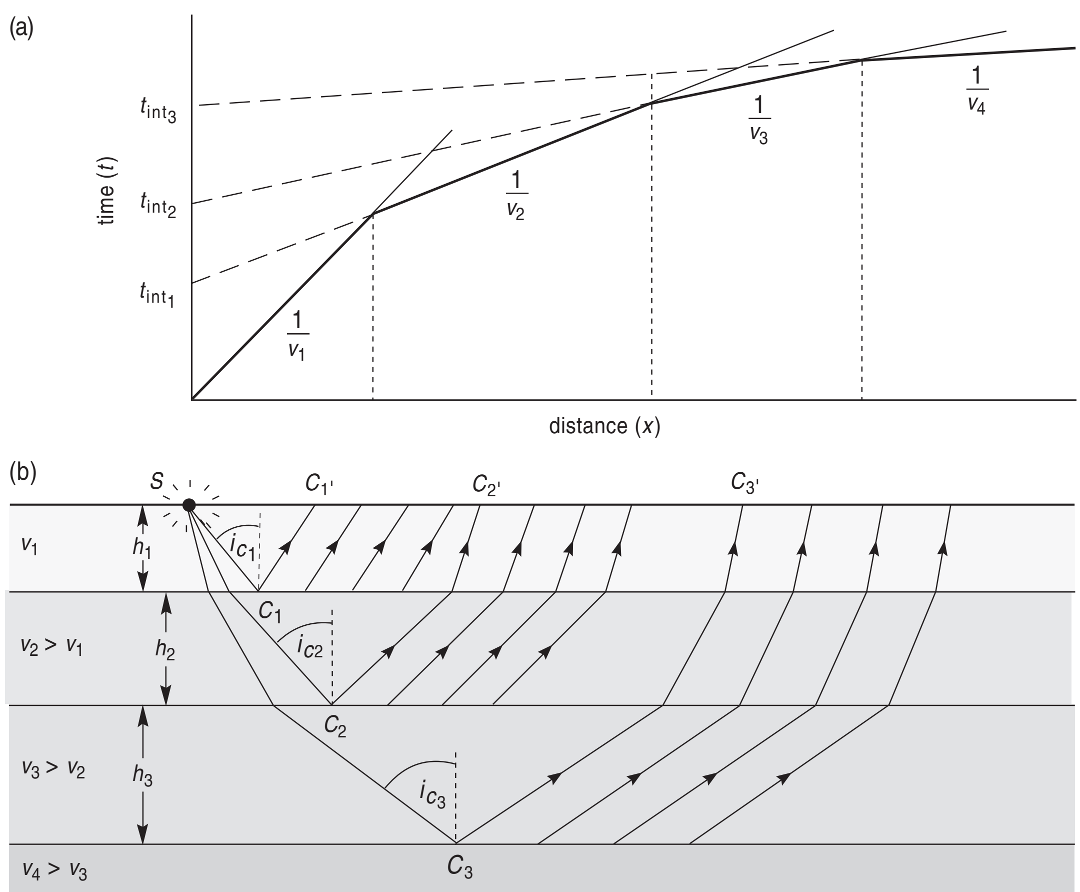

Figure 6.7 Multiple layers.

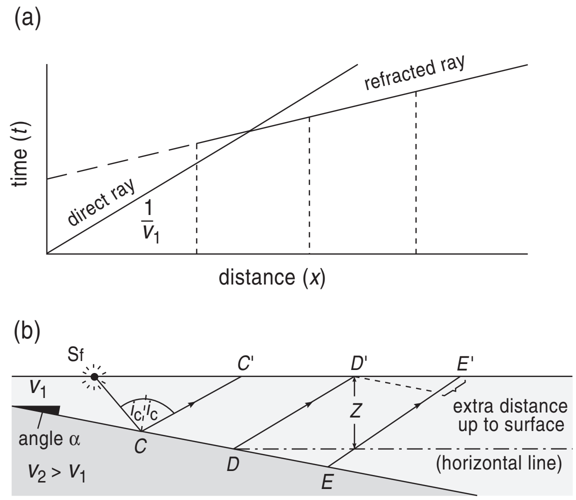

Figure 6.8 Dipping interface.

if the line is **reversed,** that is, repeated with the shot point at the right (Fig. 6.9), then successive head rays have *less* distance to travel up to the surface, and so the refracted line has a shallower slope, and appears faster. Therefore, *the presence of a dipping interface is shown by the* ***forward*** *and* ***reversed*** *refraction lines having different slopes* ('forward' and 'reverse' merely depend on which direction is shot first). If the interface has a small dip (less than about 5°), an approximate value of v_2 is found by averaging the forward and reverse slopes:

$$\frac{1}{v_2} \approx \frac{1}{2}\left(\text{slope}_{2_f} + \text{slope}_{2_r}\right) \qquad \text{Eq. 6.6}$$

v_2 can then be used to find the critical angle, as usual (Eq. 6.1).

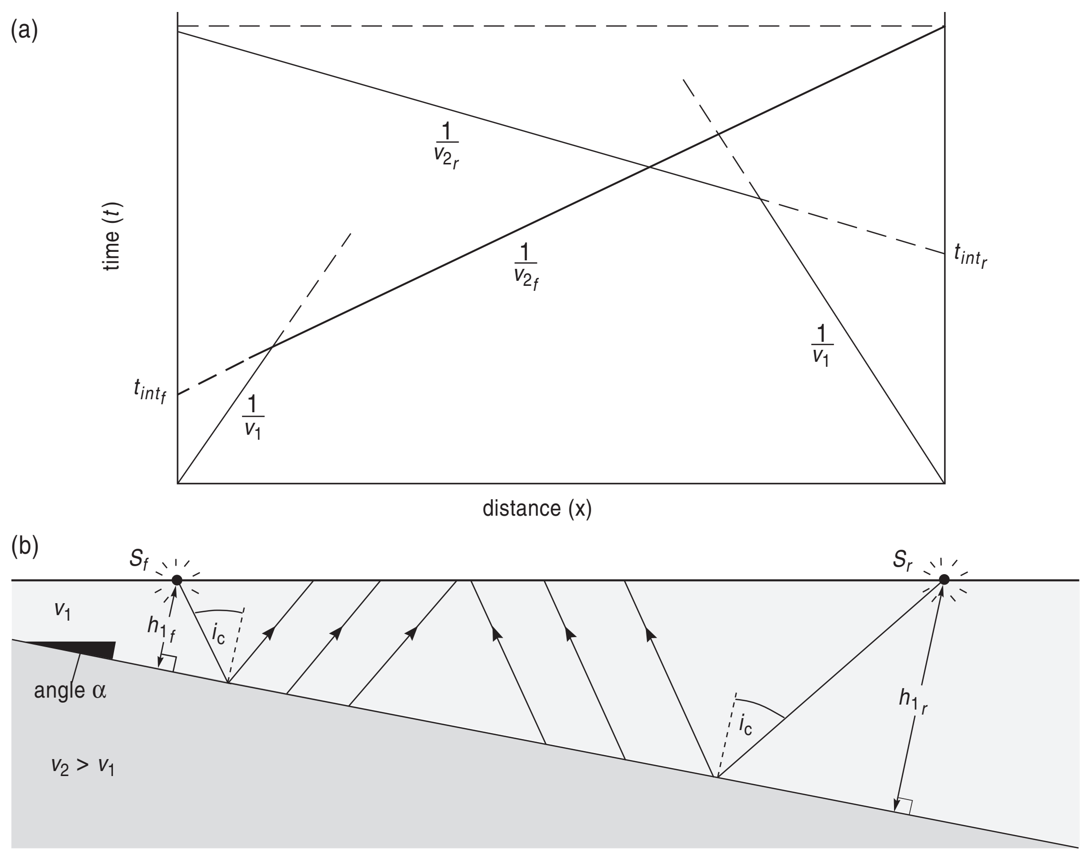

Figure 6.9 Reversed lines.

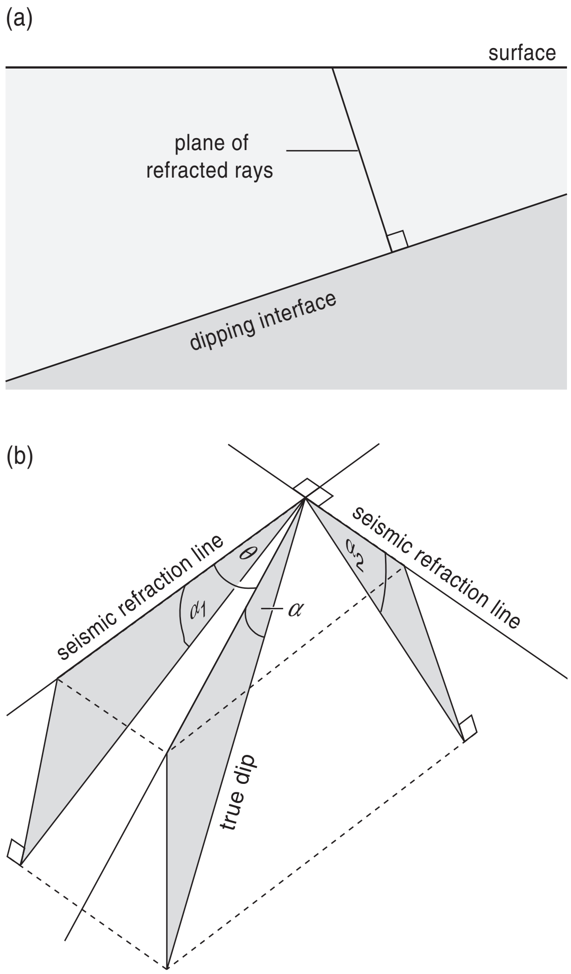

Figure 6.10 Dip in three dimensions.

The intercepts are different too, being less at the up-dip than the down-dip end. The depths to the interface are given by

$$h_{1_f} = \frac{t_{\text{int}_f} v_1}{2 \cos i_c} \qquad h_{1_r} = \frac{t_{\text{int}_r} v_1}{2 \cos i_c} \qquad \text{Eq. 6.7}$$

These are the distances perpendicular to the interface, not the vertical distances, but for shallow dips they are little different. (Exact formulas are given in some of the textbooks listed at the end of the chapter.)

In summary, t–x plots for a dipping interface compare with those for a horizontal interface as follows:

(i) The slope of the refraction line is less steep up dip, steeper down dip.
(ii) The intercept is less at the up-dip than the down-dip end.
(iii) The slopes for direct rays (first sections on *t–x* plot) are unchanged.

The true dip. It has been assumed so far that the seismic line is in the direction of the dip of the interface, but there is no reason why this should be so, unless it was known when the survey was set out. If the reversed lines are shot along strike, the rays recorded will have travelled in a plane perpendicular to the interface rather than in a vertical one (Fig. 6.10a), but the interface will appear to be horizontal. More generally, lines will be shot obliquely to the dip direction and give a value for the dip that is too low.

To find the true dip of the interface, two pairs of reversed lines at right angles can be used. Each is used to deduce a dip, α_1 and α_2 (Fig. 6.10b), from which the true dip, α, and its direction can be calculated from

$$\sin \alpha = \sqrt{\sin^2 \alpha_1 + \sin^2 \alpha_2} \qquad \text{Eq. 6.8a}$$

$$\cos \theta = \frac{\sin \alpha_1}{\sin \alpha} \qquad \text{Eq. 6.8b}$$

where θ is the angle between the dip direction and the seismic line that gave the dip component α_1.

6.5 Seismic velocities in rocks

The velocities found from travel-time diagrams give some indication of the types of rocks that form the layers, though rock types have a range of velocities. Velocities of some common rock types are given in Table 6.1.

In general, velocity increases with consolidation, so alluvium and loose sands have a very low velocity, cemented sandstones have a higher one, and crystalline rocks tend to have the highest velocities of all. Consolidation tends to increase with geological age, so a Palaeozoic sandstone, for example, may have a seismic velocity twice that of a Tertiary one.

Most rock types have a range of velocities, sometimes large, but in a particular area the range is often much less. Therefore, rock types can be identified more confidently if the velocities are measured in the area of interest – preferably in boreholes, for velocities measured on exposed rocks usually give a lower and more variable value because of weathering, opening of fractures under the reduced pressure, and pores not being fully saturated with water.

6.6 Hidden layers

There are two situations where a seismic interface is not revealed by a *t*–*x* refraction plot.

6.6.1 Hidden layer proper

It was explained in Section 6.2 that a refracted ray overtakes the direct ray, provided it travels a sufficient distance, but before that can happen it may be overtaken in turn by the refracted ray from the interface below. Figure 6.11a shows a 3-layer case where the second layer is revealed only by the short length of the line 'refracted ray 1' that is a first arrival. Suppose layer 2 were thinner: Then head rays from layer 3 would arrive earlier, displacing line 'refracted ray 2' downwards to the left, and no part of line 'refracted ray 1' would be a first arrival (Fig. 6.11b). Layer 2 is then **'hidden'.** Layer 2 would also be hidden if v_2 were decreased, or v_3 increased, for the first would displace crossover 1 to the right, while the second would displace crossover 2 to the left. If the second layer is hidden, the *t*–*x* plot will be interpreted as two layers, one with velocity v_1 over a layer with velocity v_3. The depth to the top of the third layer, calculated in ignorance of the existence of layer 2, will be intermediate between the true depths to the tops of the second and third layers.

6.6.2 Low-velocity layer

If a layer has a *lower* velocity than the one above there can be no critical refraction, the rays being refracted *towards* the normal (Fig. 6.12). There is no refracted segment corresponding to the layer, so the *t*–*x* plot will be interpreted as a two-layer case. The calculated depth to the top of layer 3 will be exaggerated, because the slower velocity of layer 2 means that a ray takes longer to reach it, than if there really are two layers, v_1 over v_3.

The possibility of a hidden layer of either kind can be recognised only from independent information, particularly geological sections or borehole logs. Common low-velocity situations are sand below clay, sometimes sandstones below limestones, and most sedimentary rocks beneath a lava or sill (see velocities in Table 6.1). A hidden layer proper may be inferred, for instance, by comparing the seismic models with geological sections. Depending upon the information available, some correction may be possible.

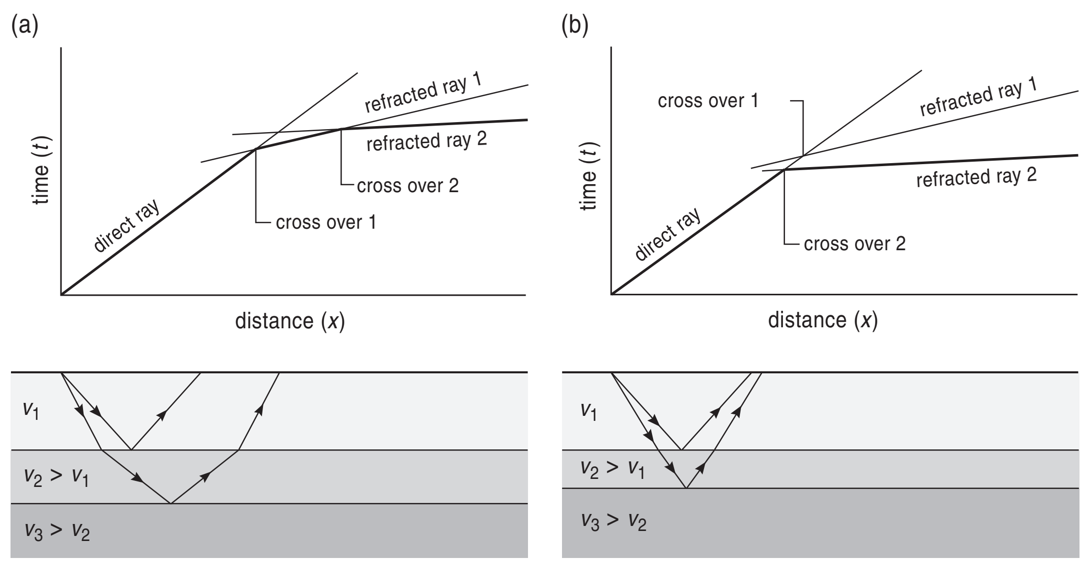

Figure 6.11 Hidden layer.

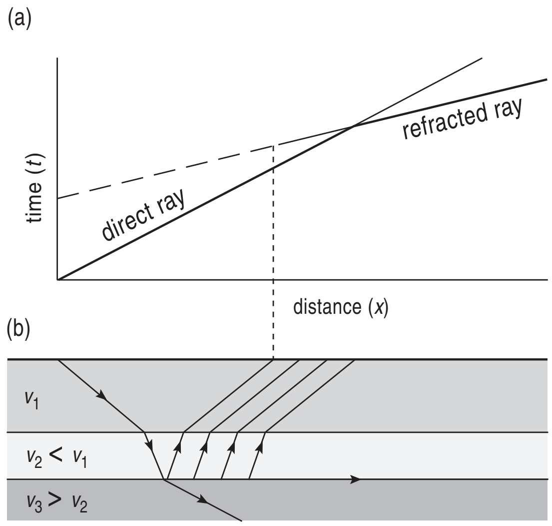

Figure 6.12 Low-velocity layer.

A final point to note is that a seismic interface is not necessarily a geological boundary and vice versa. For refracted arrivals to occur there must be an abrupt increase in velocity, that is, a seismic discontinuity. The water table may be such a interface, though it is not usually a geological boundary; conversely, Table 6.1 shows that two quite different rocks types – such as shale and sandstone – may chance to have similar velocities and so would not form a seismic interface.

6.7 Carrying out a seismic-refraction survey

Refraction surveys can be carried out on all scales from the shallow investigation of a building site to deep studies of the lithosphere that extend over hundreds of kilometres (e.g., Section 21.3.3). All require a source, a line of receivers, and a way of timing arrivals. Most modern recording is in digital form, where the seismogram is recorded as a series of binary numbers, as in a computer (described in Section 7.7.1), rather than analog form (wiggle trace) because the former gives better records and permits sophisticated processing of data to make corrections and extract the maximum amount of information, as outlined in Chapter 3.

Rather than discuss the various kinds of sources and receiver separately, it is better to describe complete systems.

Hammer seismics. Most sources send a pulse of waves into the ground. The simplest way to do this is to strike a plate on the ground with a sledgehammer. When the hammer strikes the plate a switch closes, starting the recorder, and the time of the first arrival, in milliseconds, is easily read from a screen (the duration of the flat part of the trace) or displayed as a number. Because it is so simple to strike the plate only a single geophone is needed, and it is moved progressively along a line.

The maximum workable distance depends upon the sensitivity of the system, the size of the hammer blows, the subsurface lithologies, and noise. Noise can be due to traffic and machinery, and to wind, which disturbs the ground by shaking trees, buildings, fences, and so on, or by directly shaking the geophones and cables. This last can be reduced by burying the geophones and laying the cables flat along the ground. Poor arrivals can also be improved by stacking: The plate is struck repeatedly

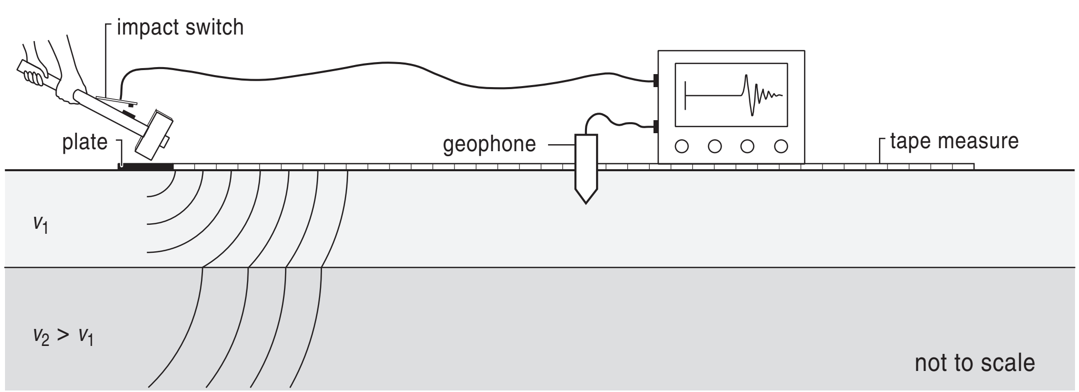

Figure 6.13 Hammer system.

with the geophone in the same position, and the arrivals are added together to average out chance noise but not the arrivals and so increase the signal-to-noise ratio. In practice, the maximum distance is, roughly, 100 m. Using a rule of thumb that the geophone line should be about ten times the depth to the interface gives a detectable depth of about 10 m, sufficient for many small-scale surveys. The range can be extended by using bigger sources such as a large dropped weight or mechanical hammers.

Explosion seismics. For larger-scale surveys a more powerful source is needed; on land a charge of explosives is commonly used. Since the object is to send seismic energy into the ground and not to produce a spectacular but wasteful blowout, the charge is buried to a sufficient depth, usually in a drilled hole. As boreholes for large charges (which can be tonnes in weight for lithospheric studies) are expensive – and the results poor if the charge is not below the water table – charges are often fired in water, making use of ponds, lakes, or the sea, as available. Firing in water is so much cheaper and generally more effective that a survey line is often chosen to take advantage of suitable bodies of water.

Because of the effort and expense of using explosives, a single receiver ceases to be practical, for it would entail a separate explosion for each receiver position. Instead, a line of geophones or seismometers is laid out. Lines up to a few kilometres in length can be connected to a central recorder by cable, but for surveys extending tens or hundreds of kilometres, cables are impracticable and each receiver has its own accurate clock, often synchronised to a master clock by radio, to time arrivals, or signals are radioed to a central recorder.

Far from the source the signal can be very weak, and high-sensitivity seismometers are essential. These are set up carefully, often in a pit to reduce noise and make contact with more solid ground. Three seismometers are often used, two horizontal instruments at right angles plus a vertical one, to record all components of the ground motion (an example is shown in Figure 5.22).

Though a large-scale survey – designed to investigate structure down to the Moho or deeper – is, in essence, no more than the hammer seismics survey greatly magnified, the logistical effort required may need years of planning and preparation.

General remarks about land surveying. Often there are near-surface layers with low velocities; in small-scale surveys these are often soil, subsoil, and the weathered top of the rock below. Though these layers may be of little interest in themselves, they often have low velocities, so the *time* spent in them may be significant even if they are thin. If receivers are spaced too far apart, these interfaces are not recognised (compare Fig. 6.14b with 6.14a), and then layers may be lumped together and the time spent in them attributed to a single layer with a faster velocity and greater thickness, and so produce an incorrect interpretation. To ensure that they are detected, receivers may need to be spaced more closely, particularly near the source (ideally, receivers would be close all along the profile, but this would be too expensive).

Surveys at sea. These generally use different sources and receivers from those used on land. As well as explosives, sources include a range of devices, which use compressed air, high-voltage discharges, or other types of energy sources. As these have their most important use in reflection seismology, they are described in Section 7.7.1. Receivers are usually **hydrophones**, immersed in the water to respond to *pressure* changes rather than water motion caused by the passage of P-waves (S-waves, of course, cannot propagate through water, Section 4.5.2). As shooting and receiving are carried out on the move, long lines, or a set of parallel lines, can be easily surveyed, an advantage offsetting the high cost of ships.

For the largest-scale surveys explosives are often used. These are dropped overboard from a moving ship and fired when safely astern, with seismic energy into the water maximised by firing at a depth that increases with the size of the charge. Two ships may be used to reverse lines. Sonobuoys may be left floating to record while the source ship sails away. Ocean-bottom seismometers that sink to the ocean floor and record like land seismometers (before rising again to the surface for retrieval), are increasingly used to improve the quality of records and provide all three components of motion.

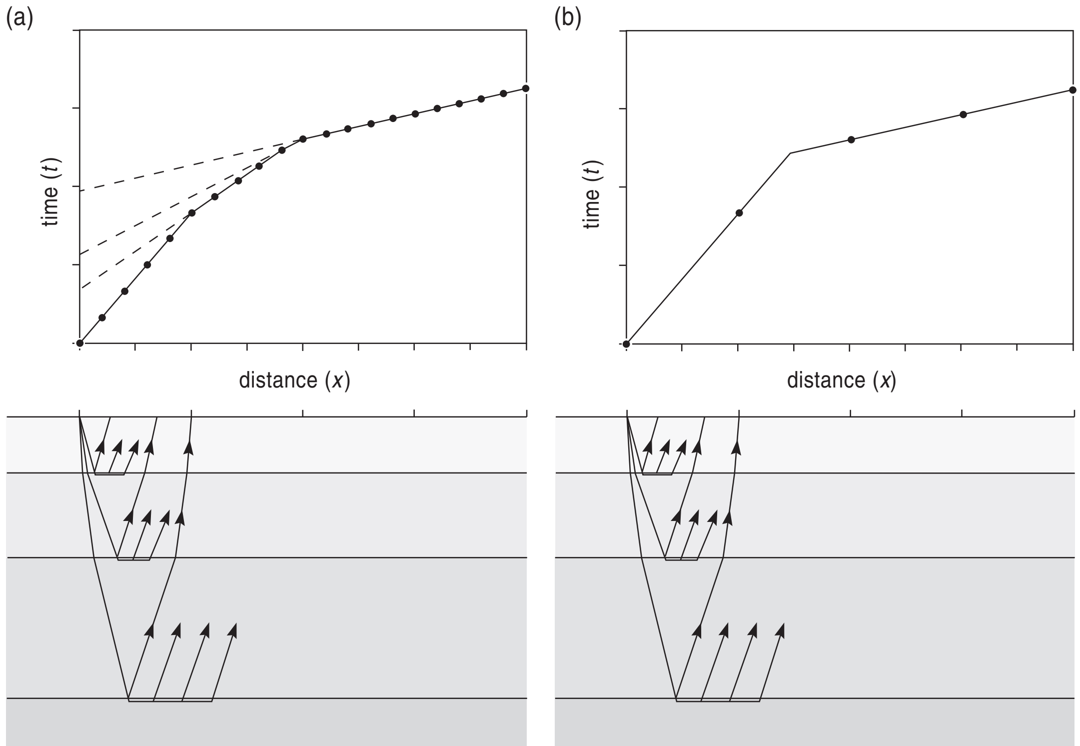

Figure 6.14 Effect of slow surface layers.

Having described how refraction surveys are carried out, we return to other layer geometries.

6.8 Undulating interfaces and delay times

If interfaces are not flat, a more sophisticated method of analysis is needed. Figure 6.15 compares the actual, undulating interface with a flat reference interface joining C_f to C_r. The forward and reverse refraction lines for this reference interface are shown dashed in the *t–x* diagram. As *M*, for instance, is closer to the surface than the reference interface, the actual travel time to *M′* plots below the reference line; conversely, that for *N′* is above it. The same argument applies to the reversed line, so the shapes of the refraction lines on the *t–x* plot show qualitatively how the interface differs from a flat surface. Note that the vertical separation of the forward and reversed refraction lines is about the same for the actual and reference interfaces.

These observations can be made more precise using the concept of delay times.

6.8.1 Delay times

The travel-time of a refracted ray is made up of three parts: travelling obliquely down to the interface, along the interface, and up to the receiver (Fig. 6.16a). But we can think of it as being made up in a different way (Fig. 6.16b): the time it takes to travel the distance between source and receiver, S_vR_v, just below the interface at velocity v_2, plus a term δ_S at the source end to equal the *extra* time it takes to go SC at velocity v_1 compared to going S_vC at v_2, and similarly δ_R at the receiver end. In effect, we are pretending that the ray travels all the way at v_2, but there is a delay between the shot time and when the ray starts on its way, and another after it finishes before the signal is recorded by the receiver:

$$t_{SR} = \delta_S + \frac{SR}{v_2} + \delta_R \qquad \text{Eq. 6.9}$$

δ_S and δ_R are called **delay times** (or sometimes time terms).

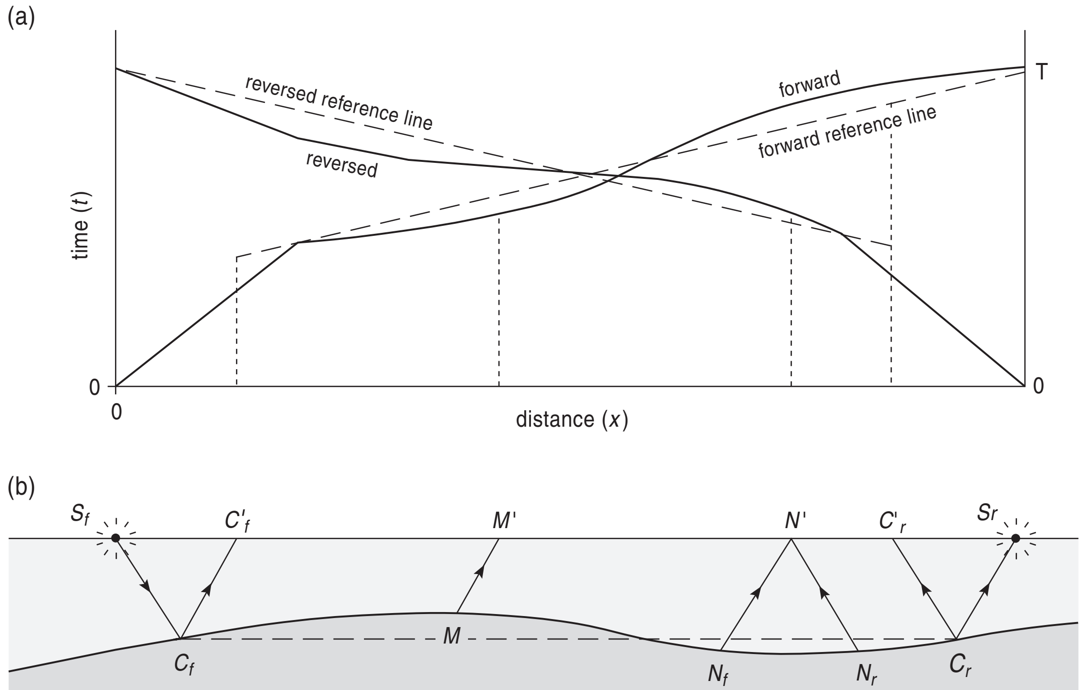

Figure 6.15 Undulating interface.

The delay time for a receiver is easily measured (Fig. 6.17a). The time, t_f, to go from one end to a receiver (path S_fCDR), and then on to the other end, t_r (path $REFS_r$), is longer than the time, t_{total}, to go from end to end (along S_fCDEFS_r) because of the extra times taken to travel from intercept to receiver, along DR and ER. As each of these extra times is just the delay time, we have

$$t_f + t_r = t_{\text{total}} + 2\delta_R \qquad \text{Eq. 6.10}$$

$$\delta_R = \frac{1}{2}\left(t_f + t_r - t_{\text{total}}\right) \qquad \text{Eq. 6.11}$$

As t_f, t_r, and t_{total} can all be read off the t–x diagram, δ_R can be calculated. This can be done for each receiver.

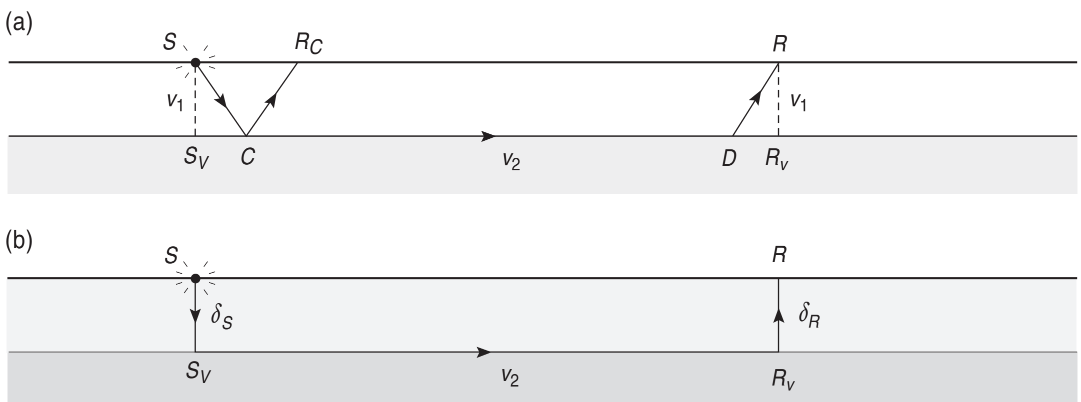

Figure 6.16 Two ways of treating travel-times.

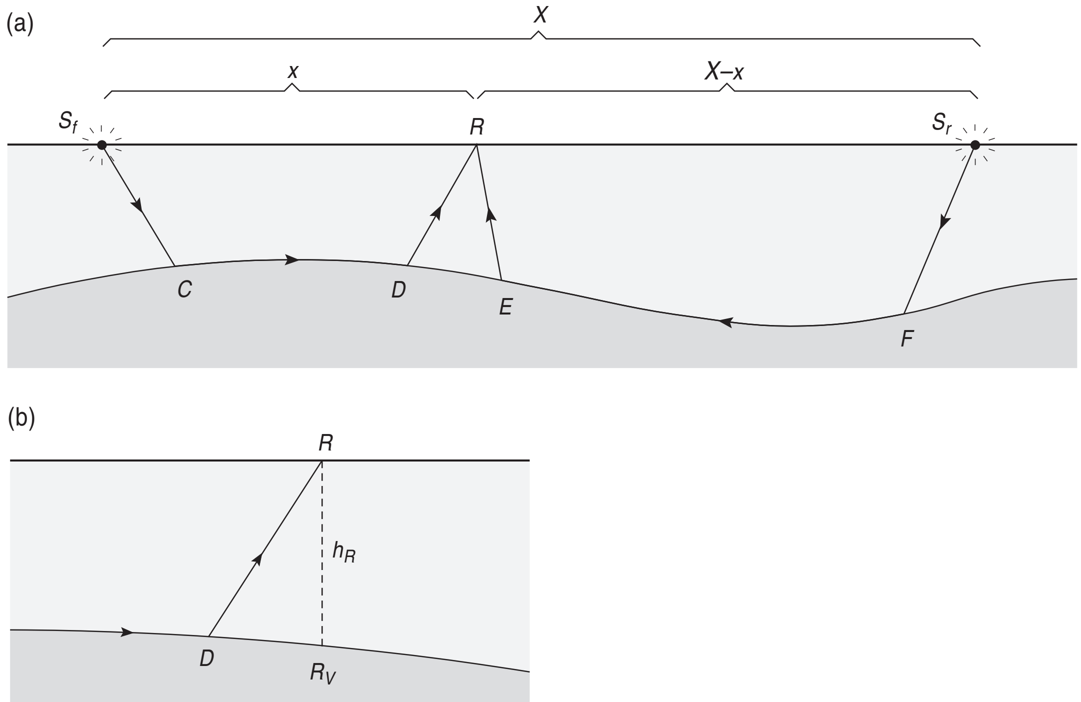

Figure 6.17 Finding delay times.

The depth to the interface can be deduced because obviously the deeper it is the larger is the delay time. The delay time is the extra time to travel from D to R, compared to D to R_v, that is ($DR/v_1 - DR_V/v_2$) (Fig. 6.17b), and it can be shown that

$$h_R = \delta_R \frac{v_1 v_2}{\sqrt{v_2^2 - v_1^2}} \qquad \text{Eq. 6.12}$$

This can be used to find the depth to the intercept, once v_1 and v_2 have been found.

6.8.2 The plus–minus method

One method that uses the above theory is the **plus–minus method.** Forward and reversed profiles are measured in the usual way, though it is essential that forward and reverse lines are exactly the same lengths, to provide t_{total}. As in Figure 6.18, t–x plots are drawn. Two further lines are plotted, one the sum of the forward and reverse times for each receiver, ($t_f + t_r$), and the other their difference, ($t_f - t_r$), as in Figure 6.18a; this is why it called the *'plus–minus' method* (it is also known as Hagedoorn's method). The minus line should be straight; this is a test that the method is applicable. Its slope is $2/v_2$, from which v_2 is found. v_1 is found from the slope of the direct lines, in the usual way.

For each receiver, the end-to-end time, t_{total}, is subtracted from the added times of forward and reverse arrivals, and the result is halved to give the delay time (Eq. 6.11). This value is put into Eq. 6.12 to give the depth to the interface below the receiver. Finally, below each receiver an arc is drawn proportional to this depth (Fig. 6.18b) and a line is drawn through them to show the interface.

There are some points to note. The method applies only to the part of the profile where refracted rays are the first arrival, so, to be useful, the profile should be long compared to the crossover distances. As the method is usually applied only to the first interface, first arrivals from the second interface may limit the useful length. Lastly, it loses accuracy if the dip anywhere is more than about 10°; if, after drawing the interface, the dips are found to be steeper than 10°, the general shape will still be correct but the depths will not be accurate.

There are other methods for dealing with uneven interfaces, but only a very general one, ray tracing, will be described.

6.9 Ray tracing and synthetic seismograms

The methods described so far for interpreting refraction data have all used inverse analysis; that is, starting with arrival times, the subsurface structure has been deduced by drawing travel-time diagrams, measuring slopes, and so on, and putting the values in equations. This works well for interfaces with simple geometries but becomes more difficult as the geometry becomes more complex. In addition, the methods use only the times of first arrivals, which means no account has been taken whether the amplitude is large or small, or of later arrivals. The power of modern computers offers another approach, forward modelling by **ray tracing.** In essence, a structure is guessed (based on a model deduced using the methods described above) and the paths of rays leaving the source at different angles are traced through the structure by using Snell's Law at each interface. The velocities can also be allowed to vary between interfaces, which causes the rays to curve. The time for each ray to return to the surface is calculated and used to construct a travel-time diagram, which is compared with the observed travel-times and thence modified as necessary. Figure 6.19 shows an example of ray tracing, with rays in the lowermost layer curving because of a significant downward increase in velocity. This example is discussed more fully in Section 21.3.3.

More sophisticated computer programs also calculate the amplitudes, taking into account loss of energy at interfaces and due to spreading out of the wave front as it travels further from the source, and may also allow for absorption. Comparing the resulting **synthetic seismogram** with the observed one provides a better test of the model than travel-times alone, for arrivals have to match the observations in amplitude as well as time.

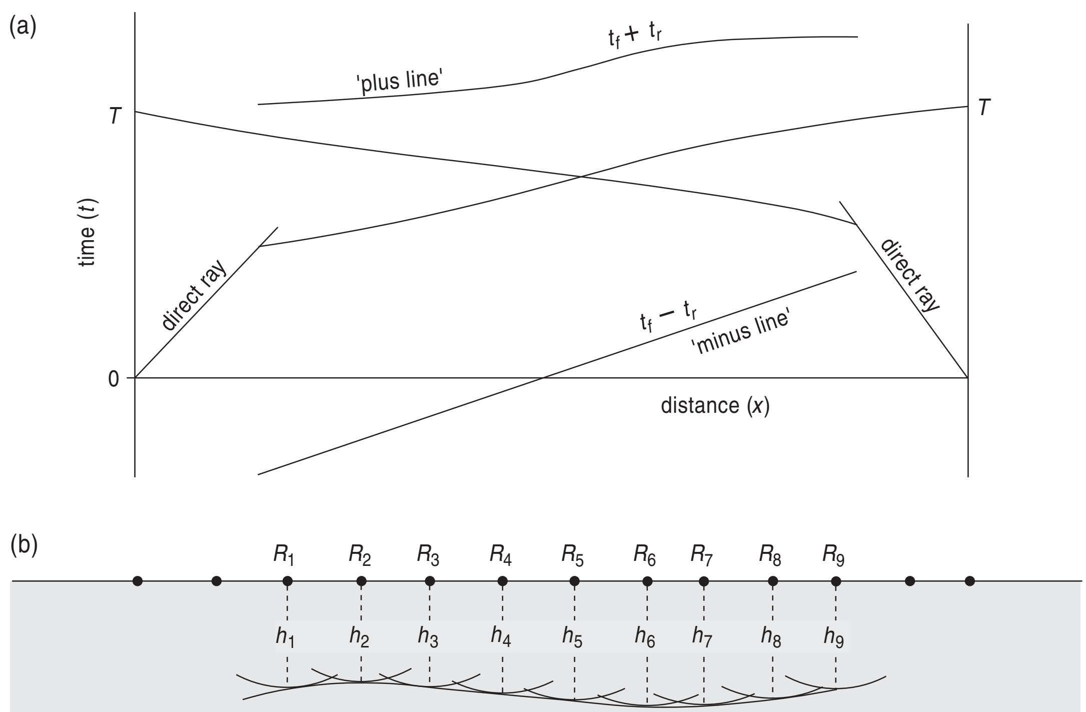

Figure 6.18 Plus–minus plots.

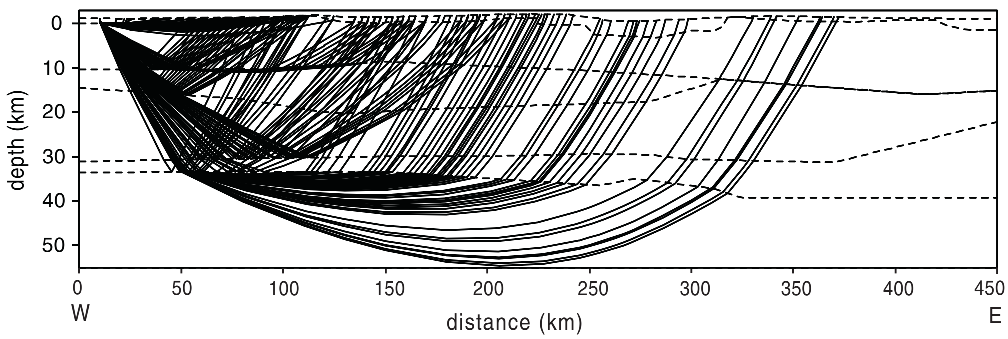

Figure 6.19 Example of ray tracing.

6.10 Detecting offsets in interfaces

Refraction seismology is mostly about measuring the depths to continuous, roughly horizontal interfaces, but it may be used to detect a fault if it offsets an interface. In Figure 6.20b a critically refracted ray travels just below the interface, producing head waves as usual along the interface from *C* to *D*. But what happens beyond the offset: Does the corner *E* cast a shadow preventing seismic energy reaching *EF* . . . ? This is best answered by considering waves. *F* is not in shadow because waves diffract into it, as explained in Section 6.1.1, and therefore head waves originate along *EF*. . . . The *t*–*x* plot (Fig. 6.20a) has an offset of the refraction line: Little energy arrives in the interval *D′* to *E′*, and when arrivals resume at *E′* they are later than those at *D′* not just because it is further from the source but also because *E* is further from the surface.

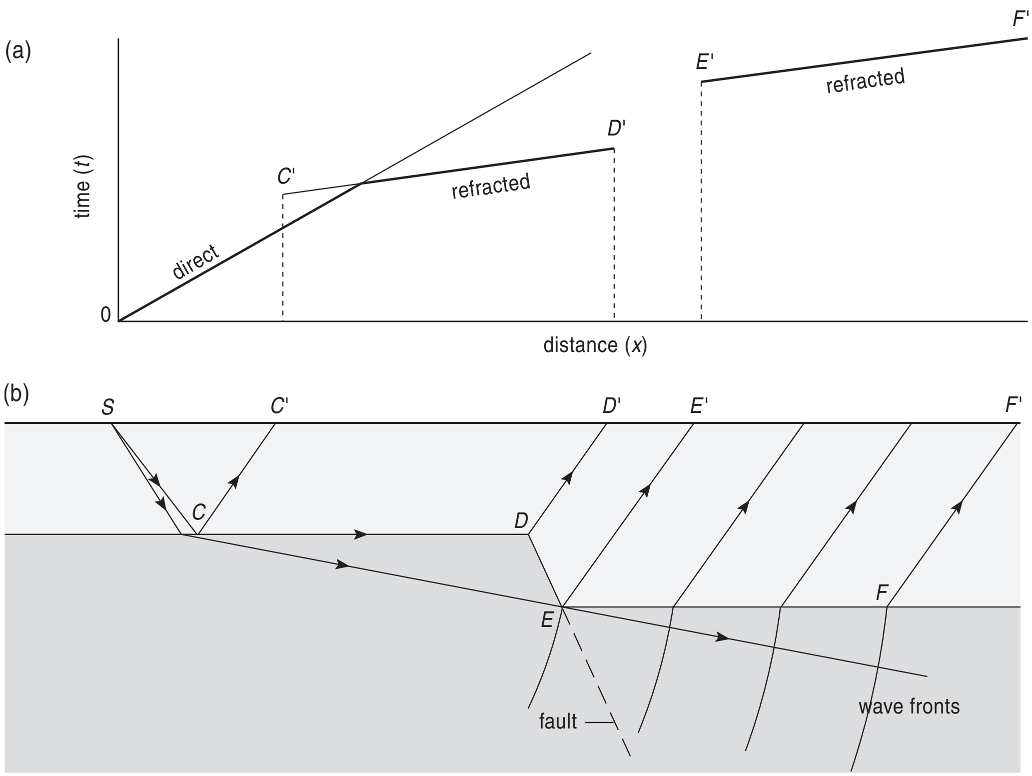

Figure 6.20 Faulted interface.

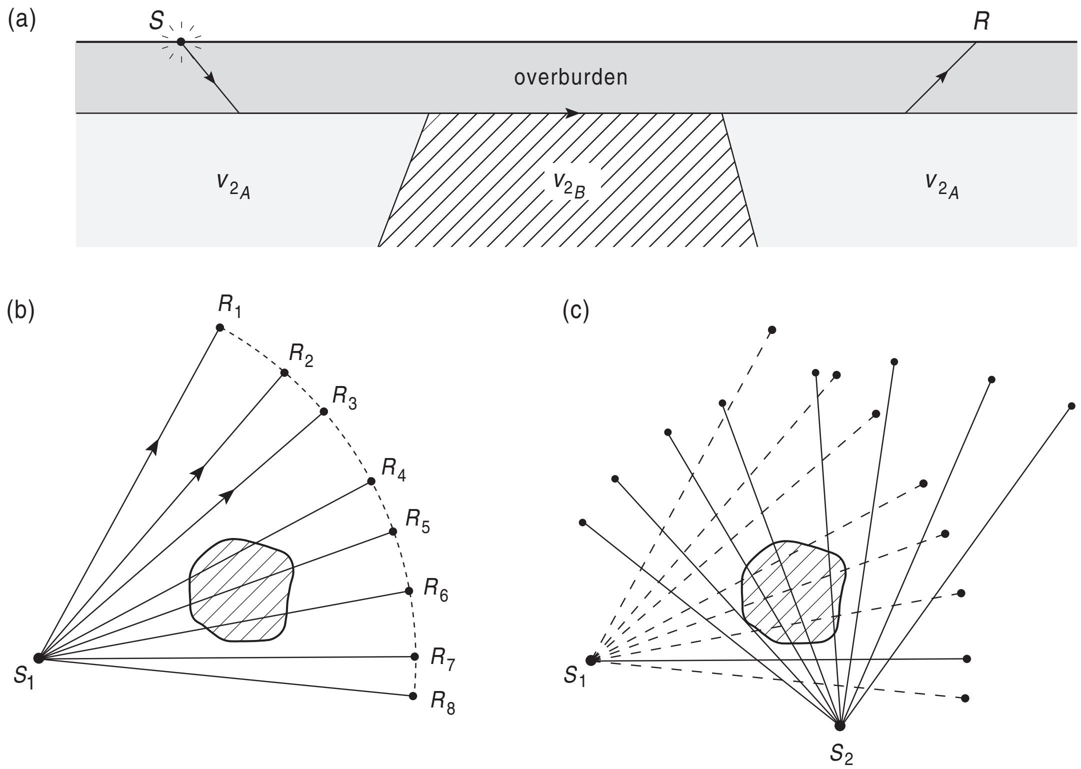

Figure 6.21 Fan shooting.

6.11 Fan shooting: Simple seismic tomography

This is a method that can detect steep-sided features even if they do not have a sharp boundary. Suppose the structure beneath the interface is as shown in Figure 6.21a, perhaps an intrusion that was truncated by erosion and then covered by sediments. Critically refracted rays travel just below the interface as usual, but those passing through the intrusion go faster and so have shorter travel-times. If receivers are arranged in an arc about the source, called **fan shooting** (Fig. 6.21b), the travel-times to R_4, R_5, and R_6 will be less than to the other receivers. This will reveal that *somewhere* along these ray paths is a body with higher seismic velocity. A set of travel-times using a different shot point will locate its position closely (6.21c). In practice, the receivers need not be all the same distance from the shot points and can be placed to record both shots without needing moving, though analysis is somewhat more complicated.

Fan shooting is an simple example of seismic tomography (Section 4.6). In the early days of oil prospecting in Texas it was used successfully to find salt domes because these were often associated with oil; the high seismic velocity of salt compared to most near-surface rocks made them fairly easy to locate.

Summary

1. Refraction seismic surveys are mainly used to detect roughly horizontal interfaces separating layers with different seismic velocities.
2. The seismic refraction method utilises head waves, which are generated by a critical ray that travels just below the interface. Head waves leave the interface at the critical angle.
3. Refracted rays arrive only beyond the critical distance but give first arrivals only beyond the crossover distance; up to the crossover distance the first arrivals are direct rays.

4. Seismic refraction results are usually interpreted by first plotting a travel–time (t–x) diagram. Refracted arrivals are recognised by a decrease of slope, corresponding to their faster apparent velocity. Multiple interfaces give rise to a succession of line segments with decreasing slopes. The velocities of the layers are found from the slopes of these line segments, thicknesses from their intercepts, using appropriate formulas.
5. The value of the velocities can help identify the rock type of a layer, but not uniquely because rocks types have a range of velocities. Ambiguity can be reduced by comparing values with the velocities of known rocks in the area.
6. Survey lines are reversed to allow the dips of interfaces to be detected and measured. The down-dip refraction line has a smaller intercept but greater slope than the up-dip line.
7. An interface is not detectable if:
   (i) The layer below has a lower velocity than the layer above (low-velocity layer).
   (ii) Though it has a higher velocity, its velocity and thickness are such that no refracted rays from the interface arrive before those from the interface below (hidden layer).
8. An undulating interface may be investigated using delay times. One method is the plus–minus method.
9. A general and powerful method for interpreting the data from complex structures is modelled by ray tracing, which uses a computer to adjust models until calculated travel-times and traces agree with those observed.
10. An offset in an interface produces an offset in the refraction line. Steeply sided bodies – such as an intrusion – can be found using fan shooting, a simple tomographic method.
11. Seismic interfaces need not be geological boundaries, and vice versa.
12. Seismic refraction surveys require some kind of source and one or more receivers, often laid out in a line. Sources range from a hammer blow to a ton or more of explosive detonated in a borehole or underwater. Receivers are geophones or seismometers on land, hydrophones in water.
13. You should understand these terms: shot point, hydrophone; direct ray, refracted ray, head ray, reflected ray; critical angle and distance, crossover distance; time–distance diagram, forward and reversed profile, apparent velocity; low-velocity layer, hidden layer; delay time, plus–minus method; fan shooting; Huygens's wavelets, diffraction; ray tracing, synthetic seismogram.

Further reading

The basic theory of seismic refraction is covered in standard textbooks on applied geophysics, such as Doyle (1995), Kearey and Brooks (1991), and Reynolds (1997) – which also gives case histories of environmental interest, Robinson and Coruh (1988), and – at a more advanced level – Telford et al. (1990). Milsom (1996) gives advice for carrying out surveys.

Problems

1. Explain, with the aid of a ray diagram, how direct, refracted, and reflected rays can reach a receiver on the surface, given that the subsurface consists of a layer overlying higher-velocity material. Explain why not all the rays can be received at all distances from the shot point.
2. Compare the relative arrival times of direct, refracted, and reflected rays (i) close to the shot point, (ii) at the critical distance, (iii) at the crossover distance, (iv) a long way from the shot.
3. Explain why it is important to determine the structure and velocities of thin, near-surface layers even though the interest is in deeper layers.
4. Sketch the t–x diagram for reversed profiles for an interface that is horizontal. Show how the diagram changes when the interface is tilted about a point midway between shot points.
5. A seismic survey is to be carried out to measure the approximate thickness of overburden, which is thought to be about 4 m thick. Deduce approximate distances from the shot point at which geophones should be placed to measure the depth of the overburden, given that it has a velocity of 1.4 km/sec and overlies sandstone with a velocity of 2.4 km/sec.
6. Explain why a hammer refraction seismic survey is often carried out using only one geophone, whereas in explosion seismics it is usual to use a line of them.