Chapter 1 Introduction

Usually, it all begins with a bang generated, for example, by a dynamite explosion on land or an air gun offshore, which sends a short, sharp pulse of sound into the ground. The sound wave rushes down and down until it meets a new rock layer of hardness (hardness in the sense of the rock's resistance to being squeezed) different from the hardness of the rocks in which it is traveling. A replica of the downward-traveling sound wave echoes back toward the surface from the boundary between the two rock layers. The original pulse continues its downward journey, gradually becoming weaker, sending echoes back to the surface every time it encounters a change in rock hardness. The greater the hardness change, the stronger is the echo. Listening devices (geophones on land and hydrophones offshore) hear the echoes as they return to the surface. There are usually so many echoes that, once they start arriving, they often overlap to form a continuous stream of sound. On a typical commercial seismic survey, the geophones listen for echoes for six seconds after the initial bang. The last echoes to arrive are normally very weak, often one hundred thousandth of the strength of the early echoes; and so the geophones that detect them must be very sensitive.

The basic concept of the seismic method is illustrated in figure 1.1. Before progressing it is perhaps worthwhile to review briefly how the seismic-reflection method is applied in practice and to introduce some of the jargon that inevitably evolves with any technique.

DATA ACQUISITION

The initial bang is called the *shot;* its geographical location, the *shotpoint;* and the resulting sound, the *source pulse* or *source wavelet.* Rock hardness is called *acoustic impedance,* and is defined by the product of sound velocity in the rock and the rock's density. The echoes are called *reflections.* The stream of reflections arriving at, and recorded by, the geophone during the listening time is called a *trace.*

The boundary across which hardness changes is called an *acoustic-impedance boundary* or *seismic reflector.* The latter term is usually reserved for boundaries that

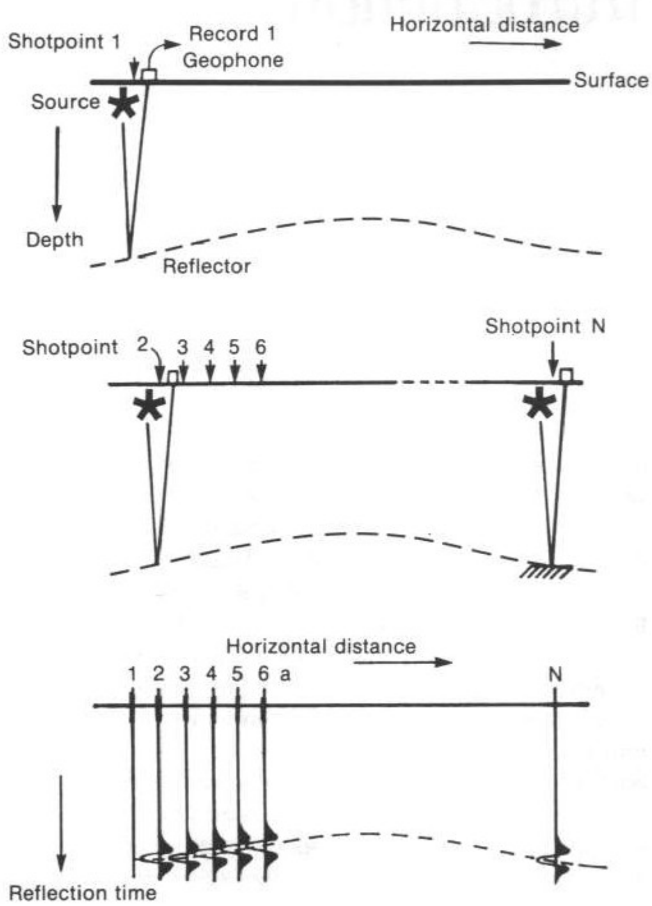


FIGURE 1.1 *The basic concept: a reflection-time measurement at regular intervals along a line, and the representation of those measurements in the form of a seismic section. Reprinted by permission of IHRDC Press from Anstey, 1982.*

produce recognizable reflection traces on seismic sections. Not all reflections produce recognizable events on seismic sections because they are too weak to be detected, are lost through interference with other reflections, etc. The type and relative size of the acoustic-impedance change is defined by the *reflection coefficient.*

Reflections resulting from sound waves that make an extra return trip, either between reflectors or down from the surface and back again, are called *multiples.* If the overlying layer is softer than the underlying layer (i.e., has lower acoustic impedance), the reflection is *positive;* if the upper layer is harder, the reflection is *negative.*

If a positive reflection on a seismic section has a certain shape (consisting of a series of wiggles about a central axis), then a negative reflection has the same shape but is reversed—every peak a trough, every trough a peak. The manner in which this (i.e., a peak or trough for a positive reflection) is displayed on a seismic section is known as *polarity.*

Sound waves travel down to a reflector and back again; therefore, the time taken from the initial bang to the recording of a reflection is called *two-way time.*

Details of the trace are analyzed in terms of: *amplitude,* a measure of reflection strength; *frequency* (measured in *hertz*), the number of oscillations per second; *bandwidth,* the range of frequencies present; and *phase,* which describes the relative shape and time position of a reflection.

In the simplest seismic system, consisting of a source and one geophone, reflections are assumed to originate from subsurface points midway between the two. However, such a system is very susceptible to *noise* (i.e., all forms of unwanted sound, such as multiples, wind noise, etc.); and it was soon found that recording reflections from the same subsurface point for different source-to-geophone spacings (*offset*) not only improved the strength of primary reflections but also resulted in a significant decrease in noise. This was termed as an improvement in the *signal-to-noise (S/N) ratio.*

A setup by which reflections are recorded from the same subsurface point with different source-to-geophone offset is known as *common-depth-point* (*CDP*) or *common-midpoint* (*CMP*) shooting. Each common midpoint consists of two or more traces, the number of which determines the *coverage* or *fold* of the seismic record. For example, two traces for a common depth point produce 2-fold or 200% coverage; 96 traces, 96-fold coverage.

Whether on land or offshore, it would be cumbersome to record all of the data for each common midpoint before proceeding to the next, and so data are acquired in the most time-efficient manner. This results in a jumble of data which has to be reordered during processing.

PROCESSING

Once all the data have been collected we arrive at the processing stage where we try to produce the perfect seismic section. The first task is to edit and reorder the data so that the series of samples corresponding to each geophone are brought together. This process is called *demultiplexing* and results in a separate trace for each shotpoint, sampled at whatever interval has been used during recording (often 4 ms). After the data have been demultiplexed there follow several processing steps before the data are assembled into an acceptable seismic section. Some of the more common processing steps are described below.

Static corrections or *statics* are corrections made to compensate for differences in land elevation (hills and valleys along the seismic line), velocity effects of the upper weathered-rock layer, depth of the hydrophones, etc.

Deconvolution: The seismic pulse starts out as a short duration burst of sound; but, as it passes through the Earth, repeated echoes eventually produce an overlapping series of reflections lasting for several seconds, a process called *convolution*. As its name suggests, deconvolution is a mathematical procedure for unscrambling the convolution effects to reveal only those reflections that stem from real reflectors.

Next, the traces are grouped together into families from common midpoints, and are then known as *CMP or CDP gathers*. Once the traces are grouped into CMP gathers, information on subsurface velocity can be obtained by a process known as *velocity analysis*, an essential processing step. Velocity analysis provides the *normal-moveout (NMO) velocities*. Since the same reflection on each CMP trace will have been recorded at progressively greater times for increasing offsets because of the ever-lengthening travel paths, the appropriate NMO velocity applied to a reflector has the effect of bringing the reflection to the same time for all traces in the CMP gather.

Following NMO correction the traces are ready for the next, and main, step in the processing sequence, in which all the values corresponding to a particular reflection time on each trace are added together. This process, called *stacking*, not only enhances the reflections from true reflectors but also usually leads to a reduction in noise.

Commonly, a second deconvolution operation is applied after stacking, in an attempt to remove unwanted reflections and perhaps also to change the shape of the pulse to some more desirable form.

By now, the seismic section is almost in its final form, and one of the last processing steps is to remove unwanted frequencies from the data. Due to attenuation by the Earth the frequency of the seismic pulse decreases with depth. It is possible to estimate the maximum frequency to be expected at a particular depth. Removal of frequencies above those that can be reasonably expected is not just a cosmetic exercise, but improves the signal-to-noise ratio. Removal of unwanted frequencies is accomplished using a filter whose passband becomes of lower frequency with increasing depth—*a time-variant filter (TVF)*. A seismic section that has undergone this processing sequence is commonly called a *filtered-stack section*.

A basic assumption of the CMP method is that reflections originate in the subsurface from the midpoint between source and geophone. However, if the reflector has sizeable dip, the assumption is no longer valid; the position of the reflection in the seismic section is then displaced downdip and the dip of the reflector is underestimated. To restore the reflections to their correct subsurface positions, the data must be *migrated*. After migration the section is, not surprisingly, called a *migrated section*.

This brief discussion has described just about the absolute minimum number of steps necessary to produce a seismic section, and additional processing procedures are often used to get the best possible results.

INTERPRETATION EQUIPMENT

Once the final processed seismic sections have landed on the interpreter's desk, the next step is interpretation. The actual requirements and equipment needed to carry out a seismic interpretation are extremely basic—namely, knowledge; some colored pencils; and, last but not least, an eraser! However, to carry out an interpretation in an efficient manner, the following requirements should be met.

Firstly, adequate space and a good, even source of lighting. Seismic interpretation requires lighting with twice the candle power usual for office work. This is essential to ensure that the sections can be viewed for normal work periods without inducing eye strain. Although, it must be said, even with the best of lighting some seismic sections induce not only eye strain, but brain strain as well.

Secondly, a table large enough to enable the sections to be unfolded and laid flat for a length of at least 130 cm. This is important because throughout the interpretation the seismic sections should be viewed both from above and obliquely along their length. Viewing the sections obliquely has a foreshortening effect on the reflector pattern and enhances lateral continuity of events. Often reflector configurations that are obscure when viewed from above become obvious when the sections are viewed in this way.

Thirdly, pencils. Marking on the seismic sections must be clear, precise, and delicate. This requires, above all, sharp pencils and a light, steady hand. The lines on seismic sections should always be thin, so as not to dominate the reflection wiggles, and should leave the section with an uncluttered appearance. Heavy, thick lines make it impossible to review the interpretation subsequently, as the eye is always drawn to the colored lines and the original reflection pattern fades into the background. It is a favorite old trick in poor data areas to put lots of thick colored lines on the sections—they can look very convincing until compared with an uninterpreted example. The other reason for thin lines is accuracy. An untidy, thick colored line can span up to 40 ms on a half-scale section (5 cm, or 2.5 inches, to 1 second two-way time), and introduce totally unnecessary inaccuracy into the interpretation. This point can be especially critical if the lines are to be digitized by someone other than the interpreter. A good practice is to use a soft (e.g., HB or No. 2) pencil to mark the horizons initially, only adding color in a uniform code later, as the interpretation progresses. For fine lines, 0.5-mm mechanical pencils have been found ideal. Colored leads are widely available; obtain the softest leads possible. If colored crayons are to be used throughout the interpretation, choose relatively soft, wax-based crayons that erase easily.

This brings us, finally, to an absolutely essential piece of equipment, the eraser. This should be as large as possible: mistakes, or changes of mind, are inevitable and the eraser will see plenty of action during an interpretation. A final refinement is an artist's horsehair brush to sweep away the erasings; you are then set up for many happy hours of seismic interpretation.

Chapter 2 Essential Theory

In this chapter we will investigate the fundamental processes of the seismic-reflection method, an understanding of which is essential for the interpreter.

THE NATURE OF REFLECTIONS

It all starts with a bang: an explosion, either in a shothole in the ground or below the surface of the water, produces an expanding compressional wavefront. Once the wavefront has left the chaos of the immediate vicinity of the explosion, it can be seen to consist of a seismic pulse with a duration of several tens of milliseconds. This seismic pulse is called the source wavelet. A simplified example of a seismic pulse and the basic elements of the seismic reflection method are shown in figure 2.1. Let us follow the course of a seismic pulse as it travels down into the ground. The seismic pulse is transmitted through the rocks as an elastic wave which transfers energy by the movement of rock particles. The dimensions of the elastic wave, or seismic wave, are very large relative to the vibration displacements of individual rock particles. Nevertheless, the seismic wave motion can be specified in terms of particle velocity and particle pressure caused by the vibrations induced by its passage. The speed in rock, typically several thousands of meters per second, at which the particle motion transports the seismic energy determines the seismic wave velocity. These high velocities contrast with those of the individual rock particles, which have velocities magnitudes of order lower, being measured in millionths of meters per second only. For each rock type, or lithology, when it is impinged by a seismic wave, there is both a particular intrinsic susceptibility to particle motion and characteristic velocity for the passage, by particle vibration, of the seismic wave through the rock.

The predictable and characteristic acoustic properties of a rock are defined as its acoustic impedance (Z), the product of density (ρ) and velocity (V).

$$Z = \rho V. \tag{2.1}$$

Velocity is usually more important than density in controlling acoustic impedance. For example, porosity varia-

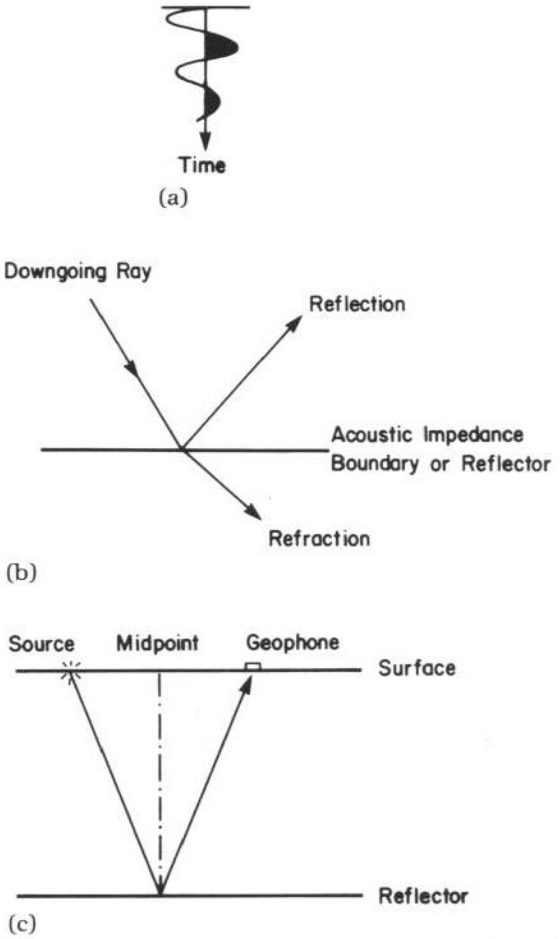


FIGURE 2.1 *The basic elements of the seismic-reflection method.* (a) *Diagrammatic source wavelet.* (b) *Reflection and refraction at an acoustic-impedance boundary.* (c) *Reflection geometry for a horizontal reflector.*

tion or the content of the pore fluids (e.g., gas in a sandstone) has a much more significant effect on velocity than on the density of the rock.

The relationship between particle velocity, particle pressure, and acoustic impedance is most easily explained by analogy with electricity (see Anstey, 1977, pp. 2–7). Using the analogy we can illustrate two fundamental relationships:

1. In electricity,

   voltage = current × resistance,

   which in acoustic terms is equivalent to

   pressure = particle velocity × acoustic impedance.

2. In electricity,

   power = current × voltage,

   which in acoustic terms is equivalent to

   intensity = particle velocity × pressure,

   where acoustic intensity represents the energy flux across unit area in unit time.

To convey a more tangible meaning to the concept of acoustic impedance, Anstey (1977) likened it to acoustic hardness. "Hard" rocks, for example, limestone, granite, etc., have high acoustic impedance, whereas "soft" rocks, for example, clays, are relatively squeezable and have low acoustic impedance. Alternatively, we could say that a given pressure would produce a large particle velocity in a low-acoustic-impedance rock (e.g., clay) but a small particle velocity in a high-acoustic-impedance rock (e.g., limestone).

We can now return to the seismic pulse, which we left forming part of the expanding compressional wavefront after the initial explosion. It will continue its downward journey into the Earth with constant velocity so long as the acoustic impedance of the rocks does not change. Typically, however, the sedimentary sequence consists of successive layers of differing lithologies which also, as a rule, have differing acoustic impedance. This need not always be the case, as acoustic impedance is the product of two variables, velocity and density. It is quite common, for example, that a claystone and a relatively porous sandstone, although having quite different lithologies, have identical values for acoustic impedance. When, however, the seismic wave encounters a rock layer with different acoustic impedance from the rock in which it is traveling, the wavefront splits. Part is reflected back toward the surface and part is transmitted and refracted to continue the downward journey (figs. 2.1 and 2.2). The wavefront split occurs exactly at the boundary between the different rocks and is caused by the abrupt change in acoustic impedance.

The seismic-reflection method is based on the recording and measurement of reflections from such boundaries. It is, therefore, important to understand why reflections arise in the first instance and what information is coded in the reflection.

Using, as an example, a thick clay interval overlying a horizontal limestone, we would expect a vertically downward propagating wave to induce a large particle motion as it passes through the clay, but only a small particle velocity in the limestone. If all of the energy in the wavefront were transmitted into the limestone, we would have a situation at the interface of a large particle velocity in the

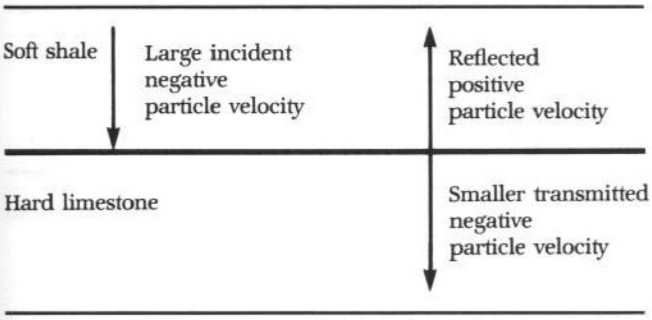


FIGURE 2.2 *Continuity of particle velocity at a soft-to-hard interface. The arrows indicate the raypath direction (i.e., direction of travel). Reprinted by permission of IHRDC Press from Robinson, 1983, fig. 3.15, p. 144.*

clay, whereas just on the limestone side of the interface the particle velocity would be small. Without there being a reflection, a difference in particle velocity across an interface can only occur if the two lithologies separate along their boundary. Separation along a boundary due to the passage of a seismic wave is impossible deep in the subsurface where overburden pressure is many orders of magnitude greater than the particle pressure of the seismic wave.

In order to balance the particle velocities on either side of the acoustic-impedance boundary, there must be a reflection (fig. 2.2). Not all of the incident energy in the clay can be transmitted into the limestone, and the amount that is reflected provides an exact balance between the particle velocities on either side of the interface. From this we can infer, as of course we would intuitively expect, that no energy is lost at the acoustic-impedance boundary and that the sum of the transmitted energy and reflected energy is equal to the incident energy.

We can also conclude that the strength of the reflection must be directly related to the contrast in acoustic impedance across the boundary. The greater the contrast, the stronger the reflection required to balance the difference in incident and transmitted energy.

The strength of a reflection generated at a boundary can be quantified in terms of the boundary's reflection coefficient (*RC*); at normal incidence this is

$$RC = \frac{Z_2 - Z_1}{Z_2 + Z_1}, \tag{2.2}$$

where

Z_1 = acoustic impedance in the upper layer.
Z_2 = acoustic impedance in the lower layer.

The reflection coefficient can be positive or negative depending upon whether "softer" rocks overlie "harder" rocks, or vice versa.

We do not actually measure directly the contrast in acoustic impedance across a boundary but deduce it from the amplitude of the recorded reflection. The greater the amplitude, the stronger the reflection and, by inference, the greater the acoustic-impedance contrast. Onland geophones respond to particle velocity amplitude; offshore hydrophones respond to acoustic pressure amplitude. In terms of amplitude the reflection coefficient is the ratio of amplitude of the reflected wave to that of the incident wave. For example, if the reflected wave has one third the amplitude of the incident wave, the reflection coefficient is 0.33. A reflection coefficient of 0.33 is relatively large; usually reflectivity is much lower. Fortunately, the energy reflected is approximately proportional to the square of the reflectivity. In the above example for a reflector with an *RC* of 0.33, only one ninth of the energy is reflected while eight ninths continues the downward journey. So, in most cases, the fraction of energy reflected is minute and, fortunately, almost all the energy is transmitted and available to generate reflections from deeper interfaces.

Recording the amplitudes of the reflections as they return to the surface enables us to assess the magnitude of the acoustic contrast causing reflection. This can have geological significance, but it would be even more useful if we could also determine whether the reflection coefficient is positive or negative and so deduce whether the change in acoustic impedance is from softer to harder or harder to softer rocks. To see if this is possible we need to examine further the nature of the measured parameters and their relationship to the reflectors. On land, geophones respond to particle velocity, while offshore hydrophones measure pressure. For a plane wave in a lossless earth, both produce an identical response to the seismic wave (fig. 2.3). A geophone located above the source shot would register an initial upward motion, or positive particle velocity; or, if the shot were in water, a hydrophone would register a positive pressure. If the same measurement were made below the shot, to record the downgoing wave, the hydrophones would again register a positive pressure but the geophone would register a negative particle velocity because the initial motion of the seismic wave is downward.

On land, where the geophone responds to particle velocity, the incident particle velocity affecting an underlying reflector is negative (i.e., the initial motion is downward). Using the example of clay overlying limestone, particle velocity is relatively large in the clay; and, to balance the lower particle velocity in the underlying limestone, the reflection must have a positive particle velocity (i.e., the initial motion must be upward). See figure 2.2. In water, the hydrophones respond to pressure. Figure 2.4 shows how the small positive pressure of the incident ray must be supplemented by a positive pressure response if the reflection is to balance the larger positive pressure in the limestone.

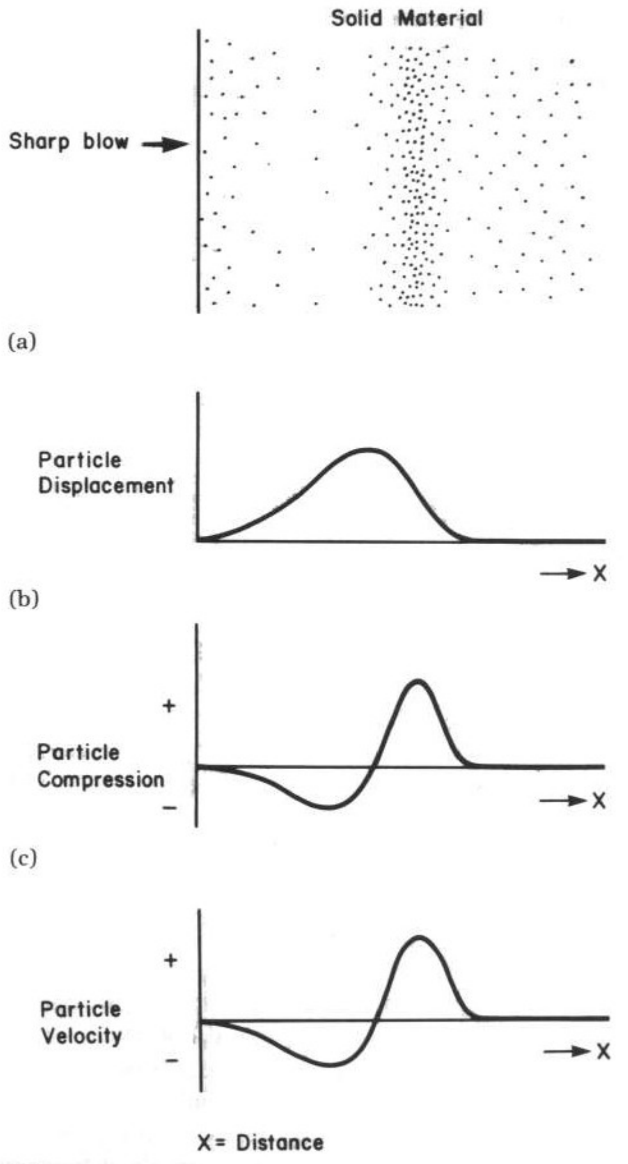


FIGURE 2.3 *Response of particle compression and particle velocity to the passage of a compressional wave.* (a) *Diagrammatic representation of particle spacing in a solid material a few milliseconds after being struck by a sharp blow.* (b) *The maximum particle displacement corresponds with the propagating compressional wave.* (c) *The particle compression is at a maximum in the propagating wave and a minimum in the rarefactional area behind the compression wave.* (d) *The particle velocity is positive (i.e., forward) in the compressional part of the wave and negative (i.e., in a backward direction) in the rarefactional part of the wave. The waveforms for particle compression and particle velocity are identical. After Anstey, 1977, by permission of IHRDC Press.*

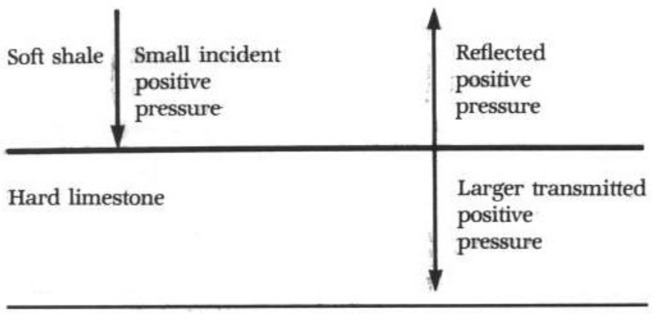


FIGURE 2.4 *Continuity of particle pressure at a soft-to-hard interface. The arrows indicate the raypath direction. Reprinted by permission of IHRDC Press from Robinson, 1983, fig. 3.16, p. 145.*

In both cases the receivers (geophones or hydrophones) would have registered a positive reflection from the contrast between clay and limestone. Note that there is a difference between the situation on land, where we start with a negative input signal (the initial motion is downward), and offshore, where the input signal is positive. The situation for an acoustic boundary with a negative reflection coefficient, for example, limestone overlying clay (harder to softer), is also shown in figure 2.5. This time both geophones and hydrophones register a negative reflection.

We can conclude that a reflection, whether measured by a geophone or a hydrophone, will always have the same response. If $Z_1 < Z_2$ (i.e., a soft rock overlying a harder rock), the reflection will be positive. If $Z_2 < Z_1$, the reflection will be negative.

We are now in a position to conclude that by recording reflections it is possible, theoretically, to relate the amplitude of the reflection to the size of the acoustic-impedance change and determine whether the reflection originates from an interface with a positive or negative reflection coefficient. Both the magnitude of the reflection coefficient and its sign are influenced mainly by geological factors, that is, lithological change (with some notable exceptions), and form the basis for providing a predictable link between reflections and geology. Last, but not least, we are able to measure the traveltime of a seismic pulse to the interface and back to the surface again (i.e., the two-way traveltime). If the velocity of the seismic wave through the rocks is known, or can be estimated, the two-way time can be converted into depth using

$$\text{depth} = \frac{\text{two-way time} \times \text{velocity}}{2}. \qquad (2.3)$$

At this stage we should introduce the concept of polarity. Instead of describing reflections as positive or negative, it is more usual to use the term *polarity*. Use of the word polarity is merely a recording and display conven-

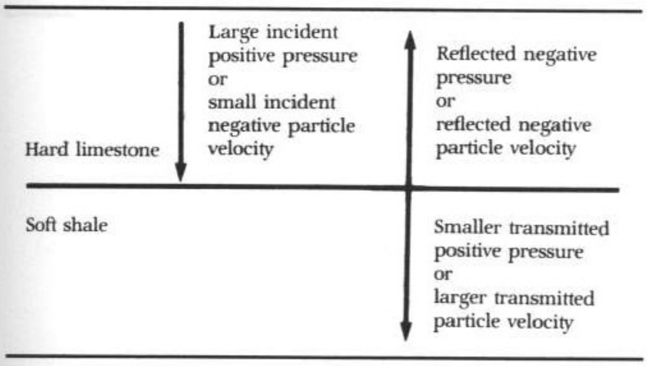


FIGURE 2.5 *Continuity of particle pressure and compression at a hard-to-soft interface. The arrows indicate the raypath direction.*

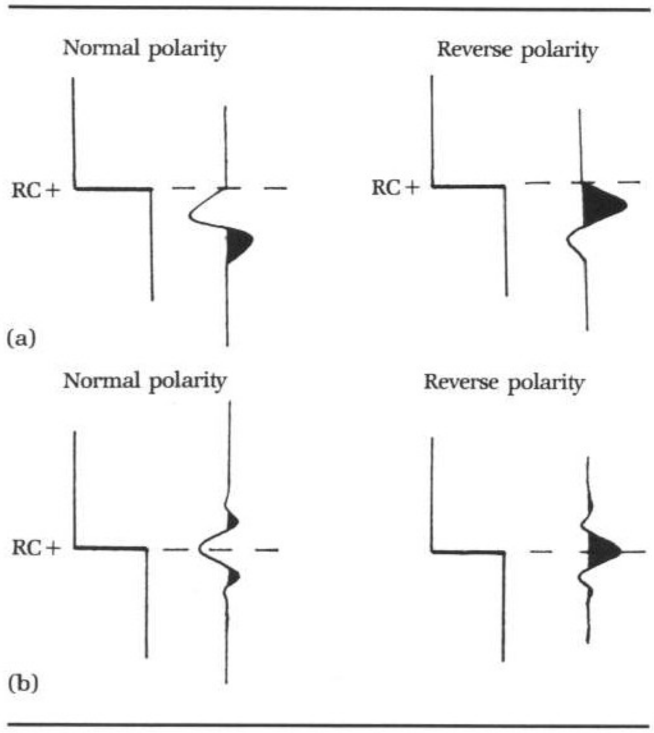


FIGURE 2.6 *Examples of idealized normal and reverse polarity for* (a) *minimum- and* (b) *zero-phase wavelets at an acoustic-impedance boundary with a positive reflection coefficient.*

tion and has no special significance in its own right. The Society of Exploration Geophysicists (SEG) defines normal polarity in this way:

1. A positive seismic signal produces a positive acoustic pressure on a hydrophone in water or an upward initial motion on a geophone on land.
2. A positive seismic signal is recorded as a negative number on a tape, a negative deflection (downswing) on a monitor record, and a trough (white) on a seismic section.

Using this convention, in a seismic section displayed with SEG normal polarity we would expect:

A reflecting boundary to appear as a trough in the seismic trace if $Z_2 > Z_1$
A reflecting boundary to appear as a peak in the seismic trace if $Z_2 < Z_1$

Figure 2.6 shows normal and reverse polarity displays for minimum- and zero-phase pulses, two common seismic-pulse types.

REFLECTIONS: SOME LIMITATIONS AND PROBLEMS

In a typical sequence of sedimentary rocks, seismic reflections will arise at each lithological boundary across which the acoustic impedance changes. These boundaries are called seismic reflectors. All acoustic-impedance changes have the potential to produce reflections. However, whether or not these changes are significant enough for their reflections to be recognized and recorded will depend upon the sensitivity of the seismic recording and processing system. Invariably, many reflections that arise from the acoustic-impedance changes present in sedimentary sequences are too small to be recorded by the methods currently available.

Figure 2.7 shows a layered sedimentary sequence and corresponding logs of velocity, density, and acoustic impedance. The rock sequence includes common sedimentary rocks and typical acoustic-impedance contrasts. Two of the boundaries are especially instructive in showing the relationship between velocity, density, and acoustic impedance. At the boundary between the gas-filled and water-bearing sands there is a sharp acoustic-impedance change which is an example of a nonlithologic acoustic-impedance change. The presence of gas in porous sand greatly reduces the seismic velocity and thereby the acoustic impedance. The other interesting boundary is between claystone and salt. The velocity log indicates a significant velocity increase at the top of the salt. If we were to rely solely on velocity as an indicator of acoustic impedance, we would expect the boundary to generate a strong reflection. However, the density shows a significant decrease from the claystones into the salt—a change in the opposite direction to that of the velocity. As acoustic impedance is the product of velocity and density, the changes in velocity and density largely cancel each other to produce only a small change in acoustic impedance at the top of the salt. A reflection from the top salt will be much weaker than we would have expected had we based our expectations of reflection strength on the velocity increase.

On the right side of figure 2.7 is the simplified seismic trace that would be produced by the acoustic-impedance changes. This could represent just one of many traces on a seismic section.

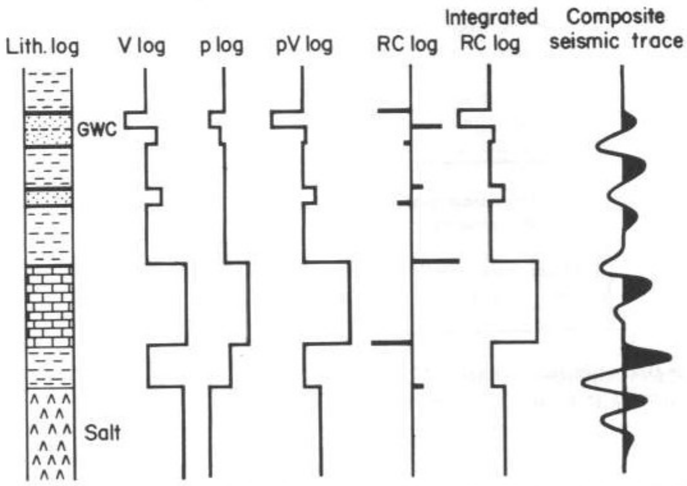


FIGURE 2.7 *The derivation of the reflection coefficient log and the resulting composite trace for a minimum-phase, normal-polarity (SEG) wavelet. The lithological sequence shown at the left consists of: a basal salt section overlain by a thick shale, massive limestone, and a shale sequence containing two sands. The lower sand is water saturated, while the upper sand contains gas overlying water-saturated sand. The velocity is shown under* V *log; with high velocity in the salt and limestone, a velocity in the water-wet sands slightly higher than in the shales, and a major depression of velocity in the gas sand. The density log is shown under* ρ *log. The salt density is very low, and the porosity in the lower sand causes the density to be coincident with that of the shales. The density in the gas sand is depressed. The acoustic-impedance log, shown under* ρV *log, is the product of velocity and density. For most lithologies it has similar form to the* V *log, excepting cases where velocity and density change in opposite directions. This occurs in the upper water sand, and is not significant; but in the salt the changes in velocity and density almost cancel. The reflection coefficients of the acoustic-impedance boundaries are shown under* RC *log, which shows the sign and expected strength of reflections. A composite seismic trace that would be produced by convolving a minimum-phase, normal-polarity (SEG) wavelet with the* RC *log is shown. The integrated* RC *log shows the effect of making a running sum of all values in a moving window down the* RC *log; this restores the* ρV *log. After Anstey, 1980a, and Robinson, 1983, by permission of IHRDC Press.*

Before going further, it is worthwhile to look at an example that shows many of the features discussed above. It is taken from a normal-polarity seismic section which passes across the Troll Field, a giant gas/oil accumulation offshore Norway (fig. 2.8). The reservoir is formed by a sequence of heterogeneous clastics some 250 ms (about 400 m) thick and overlain by shales. Several very interesting features related to acoustic-impedance changes are present. The changes in acoustic impedance between the water-filled sands, gas-filled sands, and shales are all large enough to produce strong reflections. Acoustic impedance is highest in the water-bearing sands and lowest in the gas-bearing sands. The shales have an intermediate value of acoustic impedance. From these relationships it is possible to predict the anticipated reflections on the normal-polarity section. In the water-bearing interval the reservoir top is marked by an increase in acoustic impedance and should produce a trough on the seismic trace (arrow *A*). However, in the gas zone the impedance decrease across the boundary and the reservoir top should produce a black peak (arrow *B*). At the gas-water contact the polarity of the reservoir top reflector should change from a peak to a trough (arrow *C*).

Within the sands the gas-water contact should produce a strong trough defining the contact (arrow *D*). This gas-water contact is horizontal, but its reflection shows gentle westward dip across the structure. This is a velocity effect. The dip of the "flat spot" is caused by lower velocities in the gas sands. The greater the thickness of overlying gas sand, the longer the traveltime of the seismic pulse through it, and so the deeper the flat spot appears to lie.

The seismic section shows a response to the acoustic-impedance changes in line with our predictions, but there is a further interesting effect in the seismic section, which gives some measure of the dependence of the system on the magnitude of acoustic-impedance change. The water-bearing reservoir interval appears to be quite homogeneous. No internal reflections are present—the interval is seismically transparent. However, well results show the interval to consist of massive sands separated by more shaly intervals. Although changes in small acoustic impedance occur at lithological boundaries, evidently they are too small to be detected by this seismic system. By contrast, the gas-filled reservoir interval does show some internal reflections, which dip into the flat spot and disappear. The ability of the seismic system to see internal structure in the gas-filled part of the reservoir is a direct result of the gas. The gas-filled sands have significantly

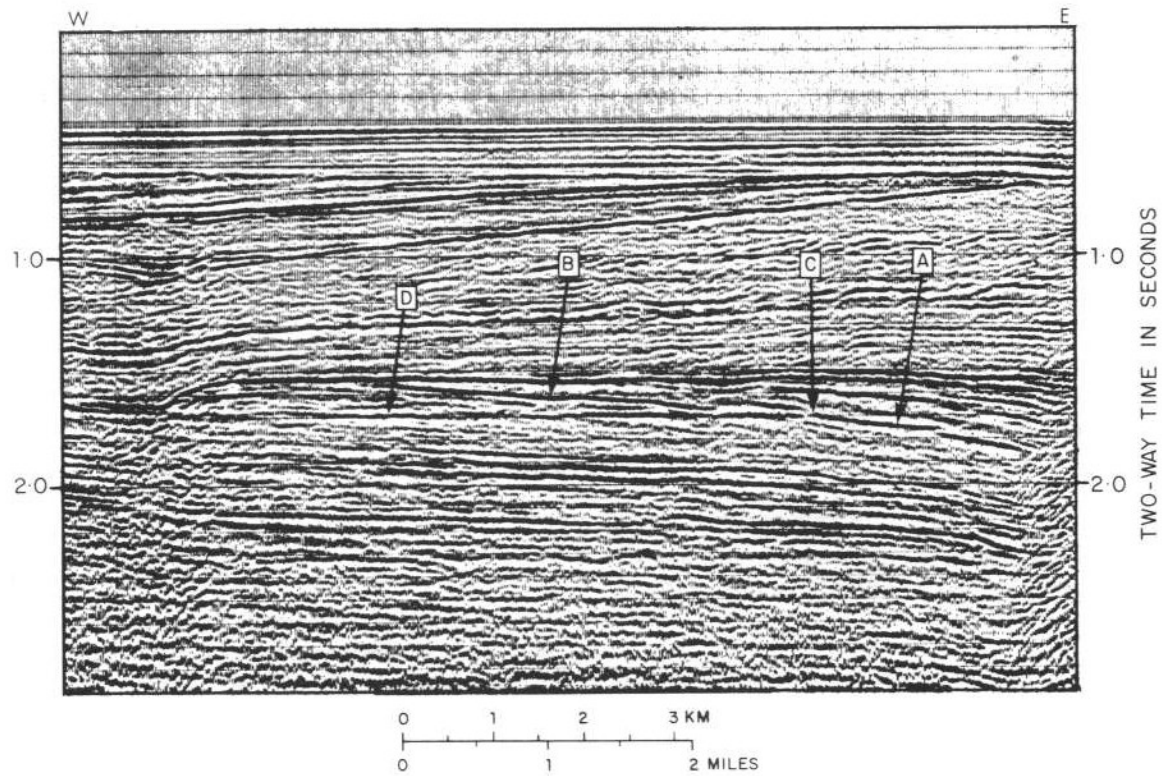


FIGURE 2.8 *Seismic section across the Troll Field, offshore Norway. Minimum-phase normal-polarity (SEG). See text for explanation. Courtesy Norsk Hydro.*

lower acoustic impedance than the argillaceous zones, which have a much lower gas saturation. The changes in acoustic impedance at the lithological boundaries are now sufficiently large to be detected by the seismic system and so reveal the reservoir's internal structure.

Figure 2.8 demonstrates that the seismic system does respond in a predictable manner. We need now to investigate further the capabilities of the method when it is used to investigate rock units in the subsurface. The example in figure 2.8 will be used to illustrate these problems. We have to contend with the selective manner in which the seismic method recognizes boundaries in the subsurface. They are visible to the seismic system only if there is an acoustic-impedance contrast *and* if the contrast is great enough to produce a reflection that can be detected. This can be a serious shortcoming of the method if, for example, there is no acoustic-impedance contrast between a reservoir sandstone and an overlying claystone seal. Lithologically, such a boundary would be clear cut and sharp to the geologist examining either a core of rock from a well or wireline logs. But to the seismic method the reservoir top would be quite invisible: no acoustic-impedance contrast, no reflection. The seismic system "sees" in a different way. What may be obvious to the naked eye is invisible to seismic system if there is no acoustic-impedance contrast. The absence of reflectors within the water-bearing sandstones of figure 2.8 demonstrates that even if there are acoustic-impedance contrasts, in order to appear on a seismic trace as reflections these contrasts must be above a certain threshold level.

Even when readable reflections are generated, there may still be problems. Figure 2.7 shows the seismic response to a typical rock sequence. No vertical scale is indicated. If a scale is added, we will perhaps be in for an unpleasant surprise. In figure 2.9 the trace of a sonic-log curve* has been added. The sonic-log curve gives a very detailed impression of the interbedded nature of the rock sequence. In most cases the sequence can be subdivided into lithological units using the sonic curve, and further subdivision internally of some thicker units may be possible. Assuming that for most boundaries the change in velocity is a good guide to likely change in acoustic impedance we can compare sonic-log and seismic responses

* A sonic log is made with an instrument, lowered on a cable into a borehole. It measures the compressional wave velocity of the rock units adjacent to the borehole. In favorable circumstances it is able to detect and display beds with thickness down to about 20 cm. The sonic log curve shows the varying velocities of the individual beds and if we assume this is primarily due to lithological changes (except the gas sand), then the curves can be considered as a crude depiction of the bedded nature and vertical variability of the rock sequences.

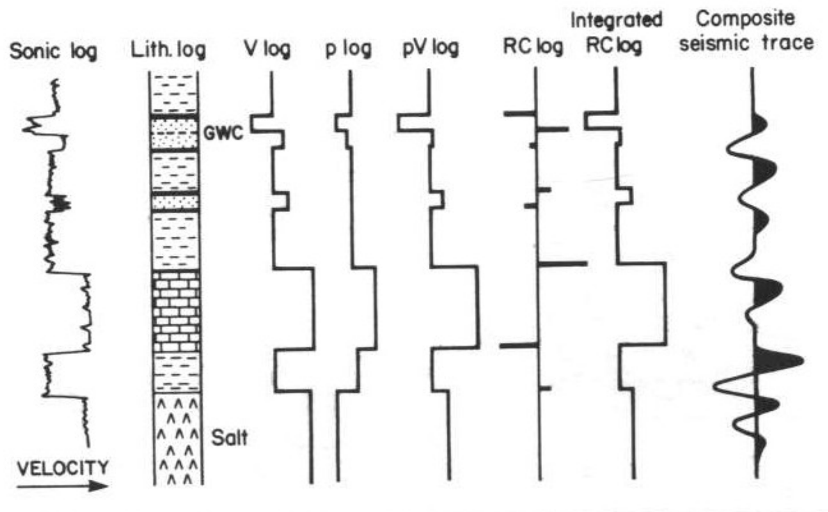


FIGURE 2.9 *Diagram showing the difference in resolution between a wireline sonic log and a seismic trace. After Anstey, 1980a, by permission of IHRDC Press.*

for the same sequences. The comparison is not too favorable. There is an incredible contrast in scale for information displayed in well logs and seismic traces. Geologists are familiar with using well logs at scales of 1:200, 1:500, 1:1000, etc., but the scales for seismic records are in the order of 1:10000 and upwards. Bridging the gap between the level of detail seen in well logs and the relative lack of detail in seismic records is a major problem. There is no tool or method that falls between the two. Well logs give great detail at isolated locations but provide no information about the geology between wells. Seismic traces show what happens laterally, but in a very limited and selective way.

INTERFERENCE

The above examples show clearly that one main problem facing the seismic method is interference between the seismic responses from closely spaced acoustic-impedance boundaries. The seismic trace in figure 2.9 is a composite of the seismic response at each acoustic-impedance contrast. The composite trace cannot be interpreted simply, as there is not a straightforward one-to-one relationship between the seismic trace and the acoustic boundaries. Because the seismic pulse is longer than the separation between some of these contrasts, the reflections interfere.

Interference always occurs when the reflections from different reflectors overlap. Interference is controlled by the length of seismic pulse in milliseconds and the spacing of acoustic-impedance boundaries in time, which is a function of the interval velocity.

Interference can be constructive or destructive (fig. 2.10). Obviously there are many possible combinations of acoustic-impedance spacing that produce interference. The length of the seismic pulse is critical. Ideally, the pulse would be a spike whose reflections would be similar spikes of lesser amplitude (fig. 2.11a); particle motion would be instantaneous and wavelength infinitesimal; and we would have almost perfect resolution. Typically, however, the input pulse consists of one or two peaks and one or two troughs and has a duration (length in the seismic section) of 20–100 ms (fig. 2.11b). The fact that the wavelet often comprises several cycles (follow half-cycles) rather than the desired spike, means that a single reflecting horizon can generate a reflection consisting of the primary event followed immediately by one or more follow half-cycles (fig. 2.11c). This can give a false impression of interbedding.

For the interpreter to evaluate the pulse shape it is necessary to know at which basic pulse shape the processing was aimed. This is not necessarily the same as the input pulse, since transformations to a new pulse shape are often made during processing. In an extremely simplified way seismic pulses displayed on seismic sections can be grouped into two main types, minimum phase and zero phase (fig. 2.6). (Although this is a reasonable simplification for the interpreter it may not be for the processing geophysicist or the modeler).

A minimum-phase pulse has its energy concentrated at its front, and is thought to be typical of many seismic signals. The pulse is said to be "front loaded," with its onset at the acoustic-impedance boundary. However, detailed analysis of pulses of the minimum-phase type reveals many different varieties. The first peak or trough need not have the greatest amplitude, and processing usually results in the first follow half-cycle having a comparable amplitude followed by a strongly attenuated tail. Quite often the lead event can be much weaker than the

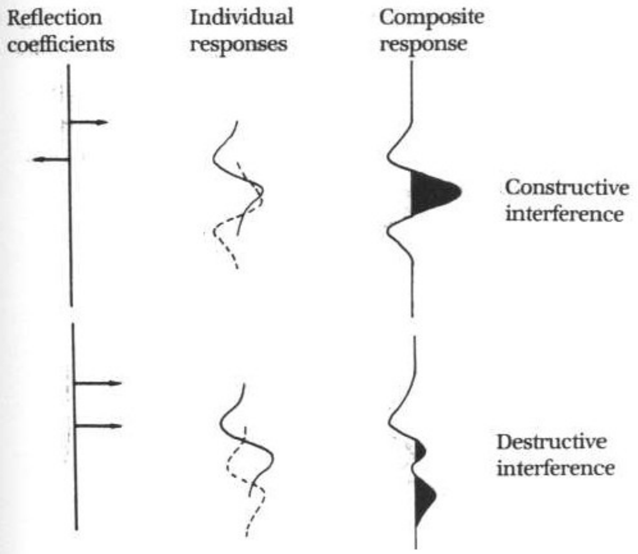


FIGURE 2.10 *Constructive and destructive interference affecting a minimum-phase normal-polarity wavelet.*

first follow half-cycle in data processed and displayed to minimum phase.

Zero-phase pulses, a product of wavelet processing and land Vibroseis data, have become more common in recent years. Zero-phase pulses consist of a central peak and two side lobes of opposite sign and lesser amplitude. Here the boundary is located at the central peak and not at the wavelet onset as is the case for minimum-phase pulses. Although a zero-phase pulse is only theoretical and is not physically realizable—since it requires that particle motion begin before the wavefront reaches the surface of the impedance contrast—this type of pulse offers the following advantages:

1. Given the same amplitude spectrum, a zero-phase signal is always shorter and always has greater amplitude than the equivalent minimum-phase signal; it therefore has a greater signal/noise ratio.
2. The maximum amplitude of zero-phase signals always coincides with the theoretical reflectivity spike. The maximum amplitude of a minimum-phase signal is delayed with reference to the reflectivity spike.

Figures 2.12 and 2.13 illustrate the contrasting interference effects for a variety of bed configurations on minimum- and zero-phase wavelets. The figures not only show the dramatic effect of interference on the composite response but also the marked differences between the responses of minimum- and zero-phase pulses to the same geological configuration. The differences are due to interference effects caused by the different shapes of the zero-

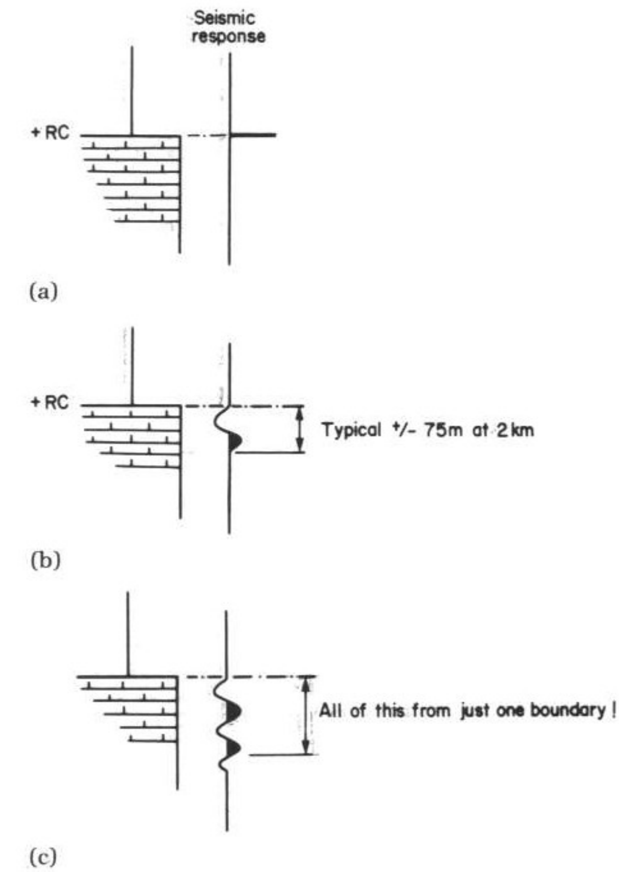


FIGURE 2.11 *Seismic response at a positive acoustic-impedance boundary.* (a) *Idealized situation where the seismic response is a spike at each acoustic-impedance boundary.* (b) *Idealized normal-polarity minimum-phase response of a good seismic system.* (c) *Response of a poor seismic system that produces a reflection consisting of several cycles from a single boundary.*

and minimum-phase wavelets. The zero-phase wavelet is centered around each acoustic-impedance boundary, and so has scope for interference with wavelets from nearby boundaries both below and above. The minimum-phase wavelet, however, extends down from the boundaries and so can only interfere with wavelets from underlying boundaries.

VERTICAL RESOLUTION

Having encountered the problem of interference we can now ask:

1. How thick must a bed or unit be before there is no interference between the reflections from acoustic-impedance contrasts at the unit's top and base?
2. How thin can a bed or unit be before its top and base are no longer resolvable?

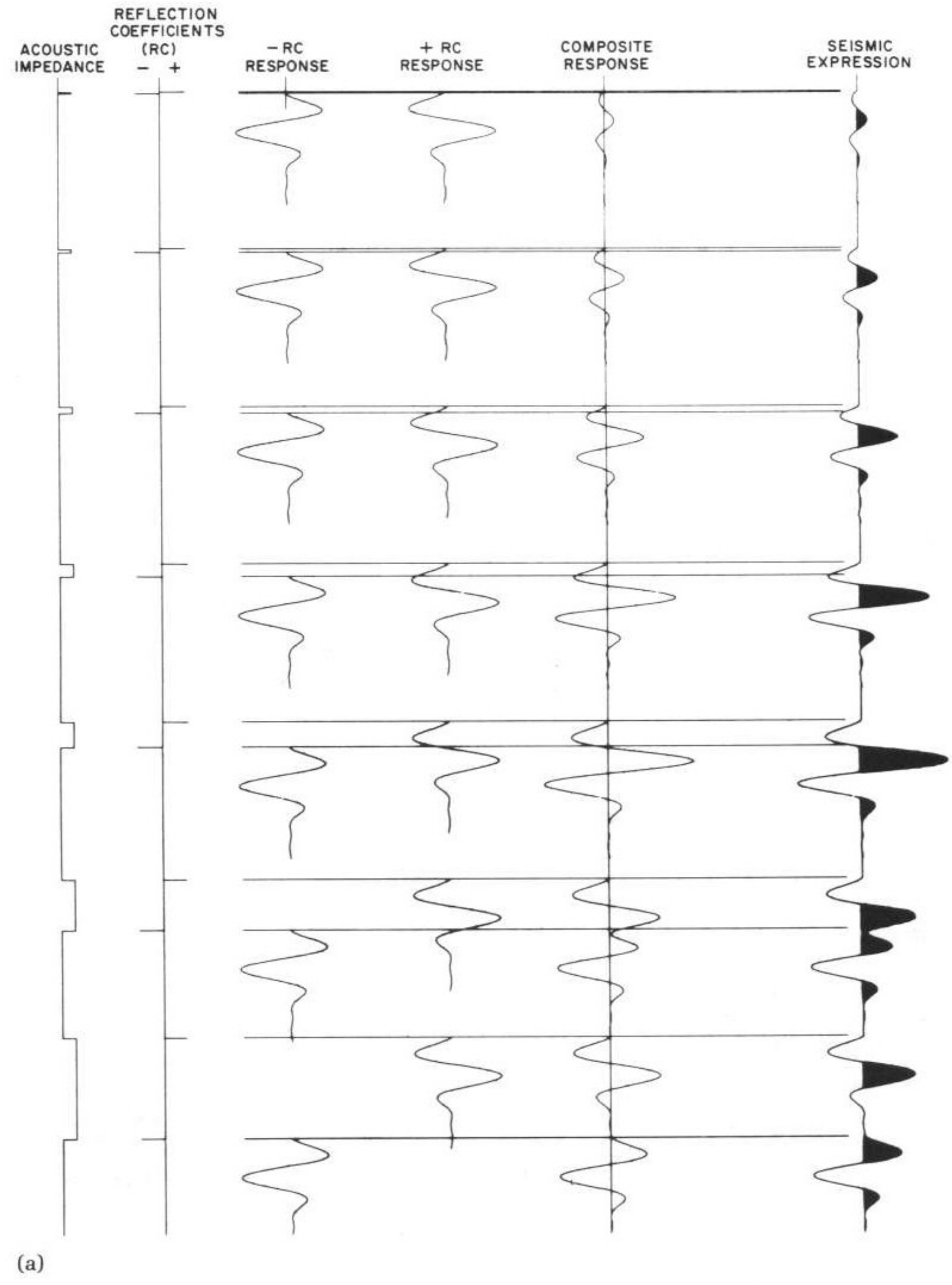
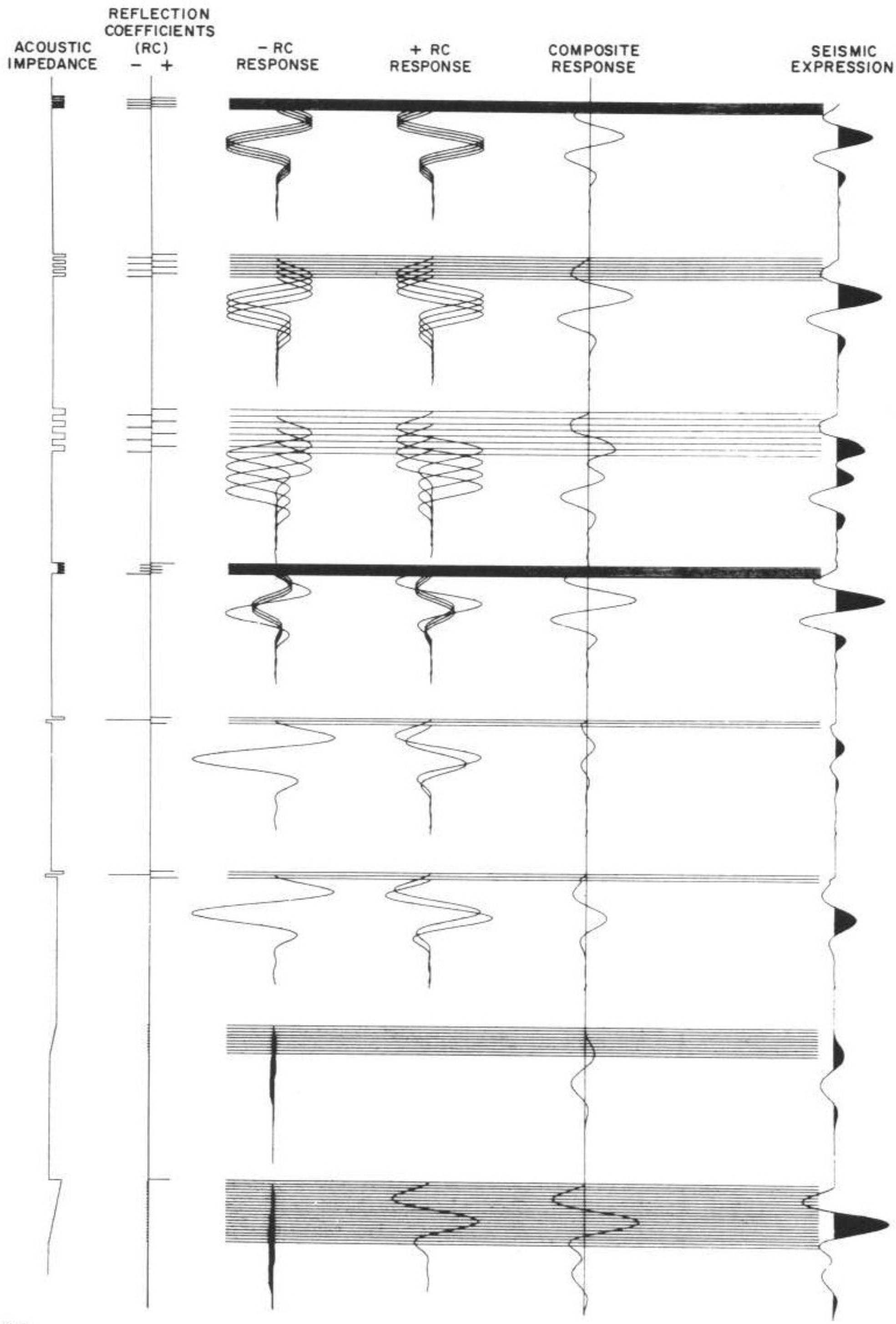


FIGURE 2.12 *Examples of interference on a minimum-phase normal-polarity wavelet for a range of bed thicknesses and bed spacings. Courtesy Norsk Hydro.*


ACOUSTIC IMPEDANCE
REFLECTION COEFFICIENTS (RC)
− +
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+ RC RESPONSE
COMPOSITE RESPONSE
SEISMIC EXPRESSION

(b)

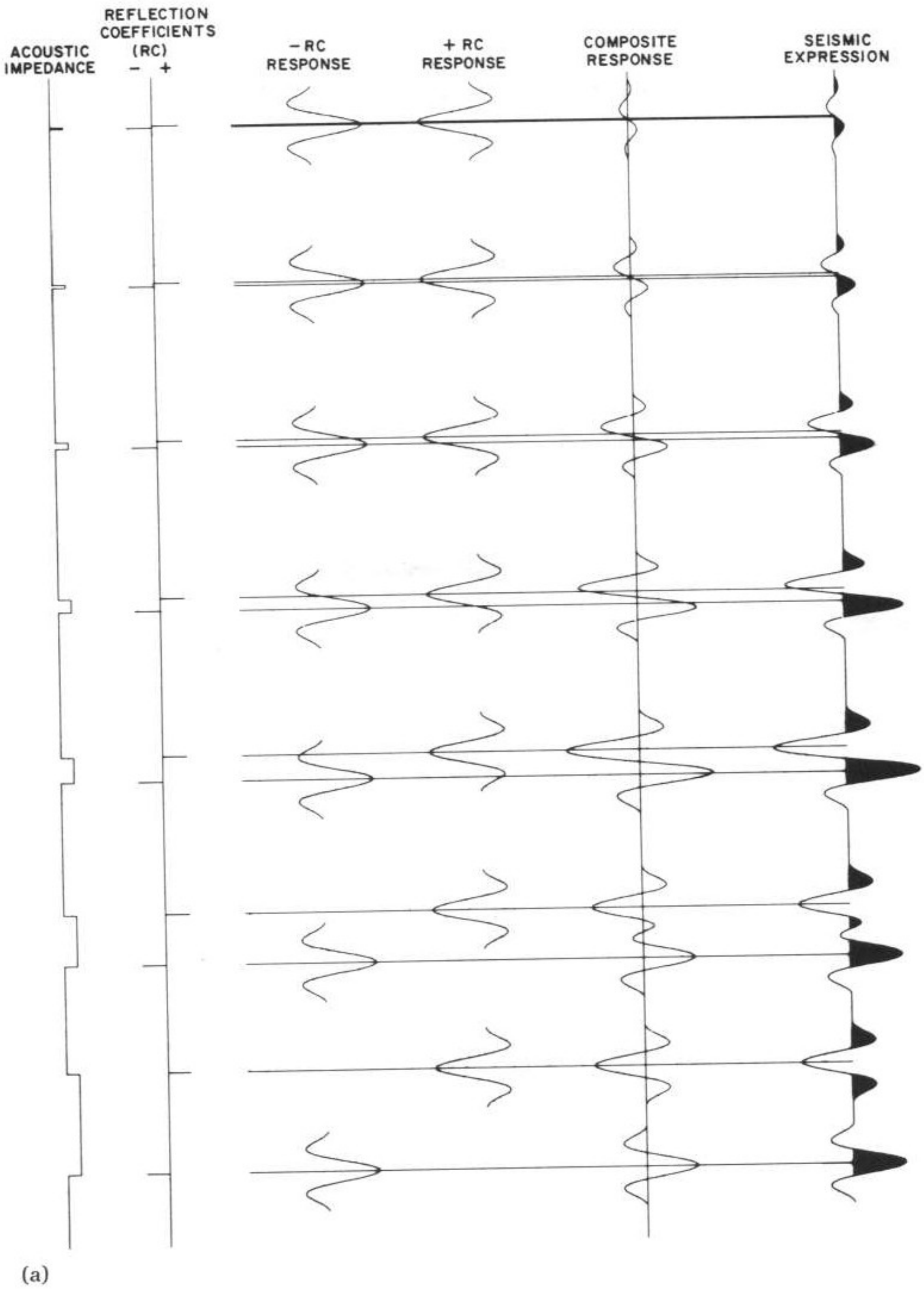
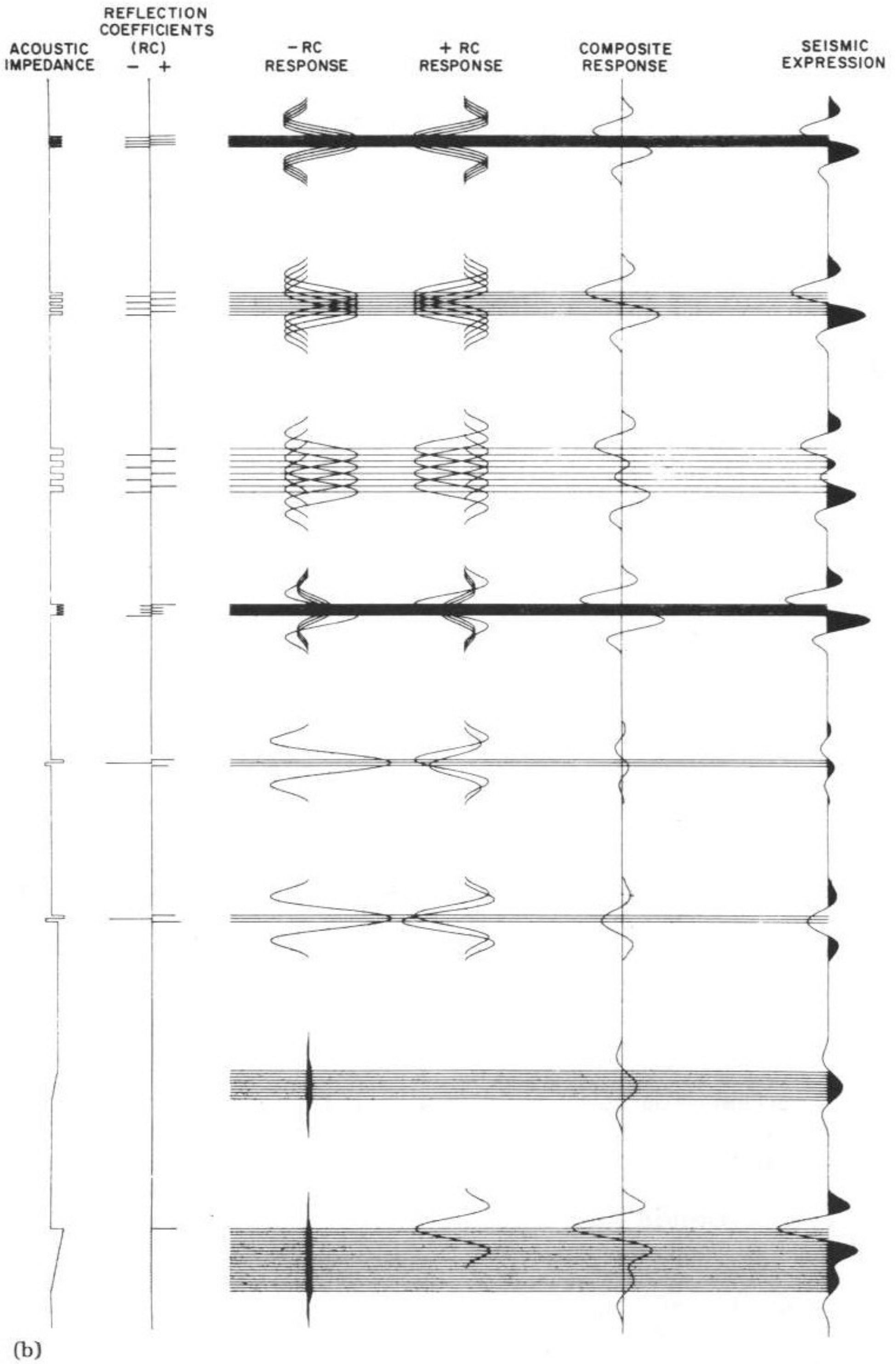


FIGURE 2.13 *Examples of interference on a zero-phase normal-polarity wavelet for a range of bed thicknesses and spacings. Courtesy Norsk Hydro.*


ACOUSTIC IMPEDANCE
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SEISMIC EXPRESSION

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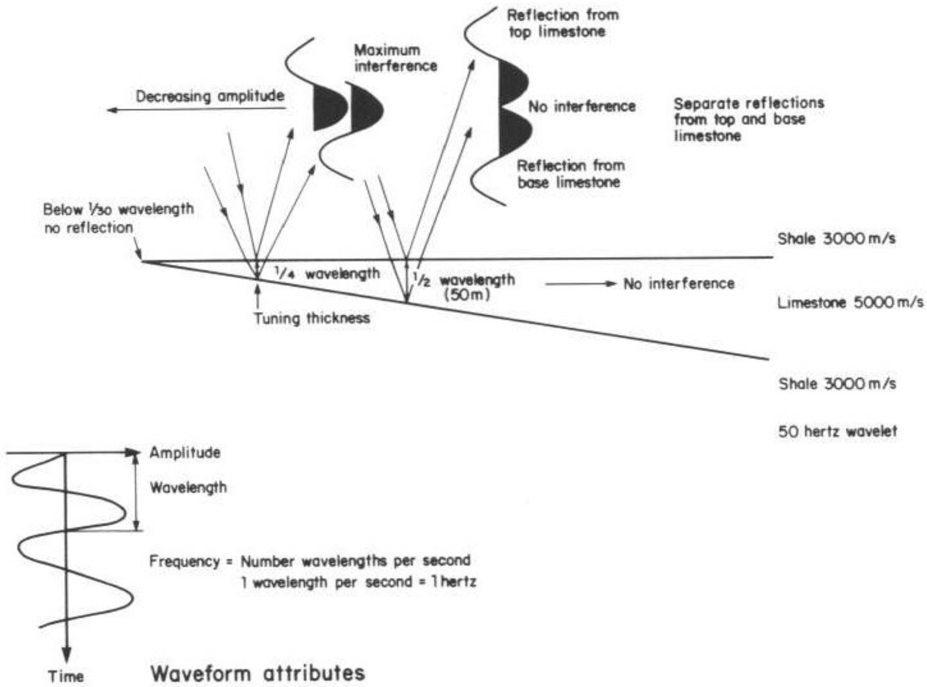


FIGURE 2.14 *Interference effects associated with a high acoustic-impedance wedge encased in lower acoustic-impedance shale. The limestone must be thicker than half the seismic wavelength for no interference between reflections from its top and base. Maximum interference and amplitude of the resulting reflection occurs at a limestone thickness equivalent to one quarter of the seismic wavelength—the tuning thickness. For limestone thicknesses below one-quarter wavelength, the reflection remains the same shape but decreases in amplitude. Once the limestone is about one-thirtieth wavelength or less, reflections from the top and base effectively cancel and there is no detectable seismic response.*

Figure 2.14 is an example of a wedge-shaped limestone unit (high velocity) encased in claystones (lower velocity). An expanding wavefront, in this case a minimum-phase wavelet, is reflected from the top of the limestone. The polarity of the incident wavelet is preserved on the reflected wavelet. The reflection from the base limestone, however, has its polarity reversed, as we would expect from a change from higher to lower acoustic impedance.

The two reflected wavelets of opposite polarity will be separated in time so long as the time thickness of the limestone is equal to, or greater than, half the wavelength of the seismic wavelet. Potentially, we can resolve top and base limestone, so long as the limestone thickness is greater than half the wavelength. If the limestone is thinner than half the wavelength, the two opposite polarity reflections begin to overlap and interfere. When the two-way transit time of the limestone reaches half the wavelet width (i.e., the limestone thickness equals one-quarter of the wavelength), as shown in figure 2.14, the two wavelets constructively interfere to form a single wavelet of anomalously high amplitude. This thickness is known as the tuning thickness. For a limestone velocity of 5000 m/s and seismic-wave frequency of 50 Hz, maximum constructive interference or tuning would occur at a limestone thickness of 25 m. When the limestone thins to less than the tuning thickness, the unusually high amplitude of the combined top and base limestone reflections decreases, and the combined reflections appear nearly the same as for a single interface (fig. 2.14). This single reflection does not thin as the bed thins.

The relationship between frequency, velocity, and wavelength is straightforward:

$$\text{wavelength} = \frac{\text{velocity}}{\text{frequency}}.$$

For example, if the frequency of our seismic wave was 50 Hz, or alternately expressed, had a period of 20 ms, at the depth where the limestone with a velocity of 5000 m/s occurred, then the limestone would have to be at least 50 m thick (half the wavelength of 100 m) to enable resolution of its top and base by the seismic trace. If, on the other hand, the frequency of the seismic wave was only 20 Hz, then the limestone would have to be at least 125 m thick for its top and base to be resolved. In the shallow

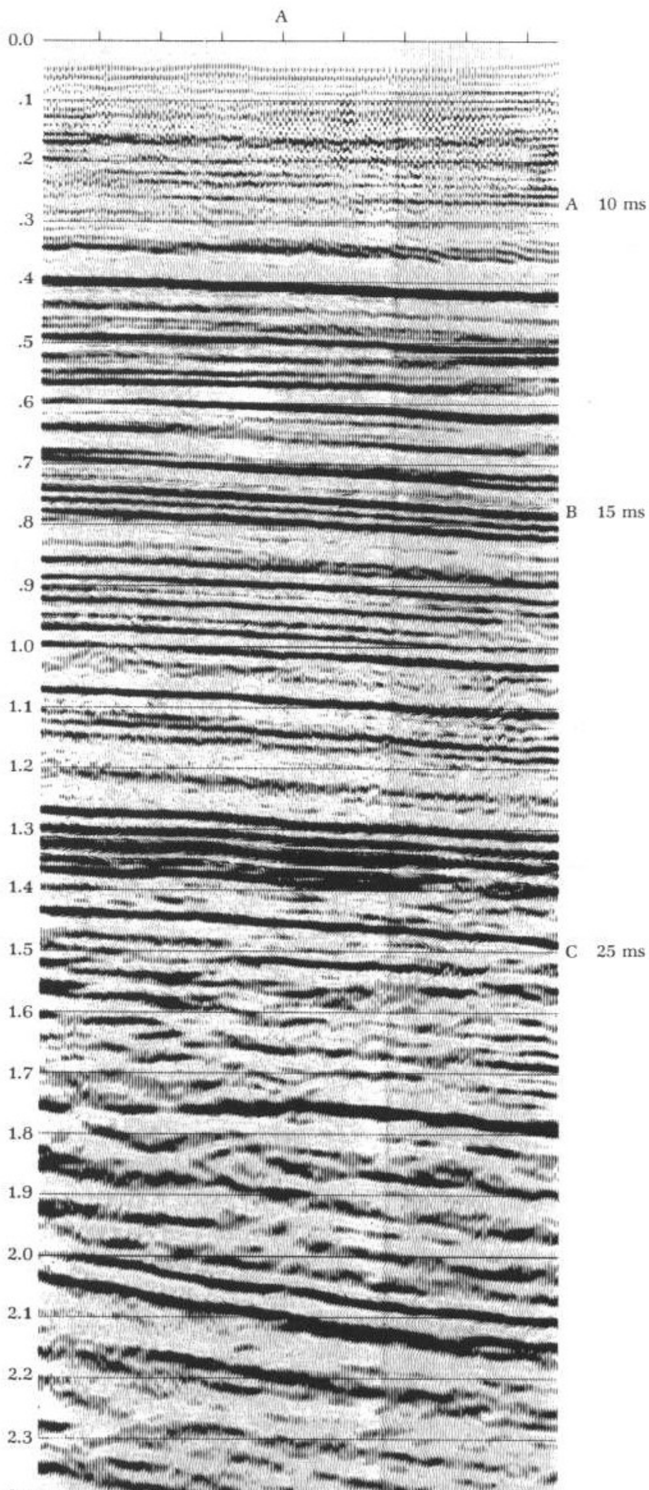


FIGURE 2.15 *Estimating the dominant frequency or period of the seismic wavelet from a seismic section. Isolated high-amplitude, continuous reflections should be chosen to reduce the risk of interference effects. It is assumed that on minimum-phase processed sections, individual reflections consist of a lead and follow half-cycles (the first and second half-cycles). Zero-phase data is assumed to have a waveform consisting of lead and follow half-cycles symmetric about a higher-amplitude central peak or trough. The dominant frequency is Hertz = 1/period(s). The seismic section here is processed to produce a minimum-phase waveform and is displayed with reverse polarity (SEG).*

Reflection A: duration 12 ms; frequency 83 Hz
Reflection B: duration 15 ms; frequency 67 Hz
Reflection C: duration 25 ms; frequency 40 Hz

The estimated frequencies should always be calculated for several reflections in a particular time range and the estimates compared with the time-variant filter (TVF) limits given in the seismic section label. The estimated value must fall within the range of the time-variant filter.

section where velocities are usually low and frequencies are high, wavelengths of around 40 m are common, with a corresponding resolvable thickness of 10 m and a detectable limit of bed thickness to produce a reflection of about 1.3 m. Deeper in the section, where velocities are higher and frequencies lower, these resolvable and detectable limits are higher. In the example above, for example, with a velocity of 5000 m/s and a frequency of 20 Hz, the resolvable and detectable limits are 62.5 m and 8.3 m, respectively. It is difficult to determine directly from the seismic trace the wavelength of the seismic pulse in meters. However, if the wavelet's period and the interval velocity of a unit can be estimated, wavelength is readily determined from the relationship.

$$\text{wavelength} = \frac{\text{velocity}}{\text{period}}. \tag{2.4}$$

The dominant wavelength or period of the seismic pulse at a particular depth can be estimated by first finding a reflection that is likely to be free from interference effects and then by measuring the duration in milliseconds of the lead and follow cycles (fig. 2.15). The tuning effect is most marked if the reflection coefficients have opposite signs at the top and base of a unit. If they are the same, for example from a wedge that has an acoustic impedance intermediate between the acoustic impedances above and below, the waveforms tend to a minimum at the quarter-wavelength thickness; and this produces a dimming of the reflection at the top of the wedge (fig. 2.16). In this case there is still a reflection after the wedge has pinched out due to the contrast in acoustic impedance between the overlying and underlying units. The detectable limit, the minimum thickness for

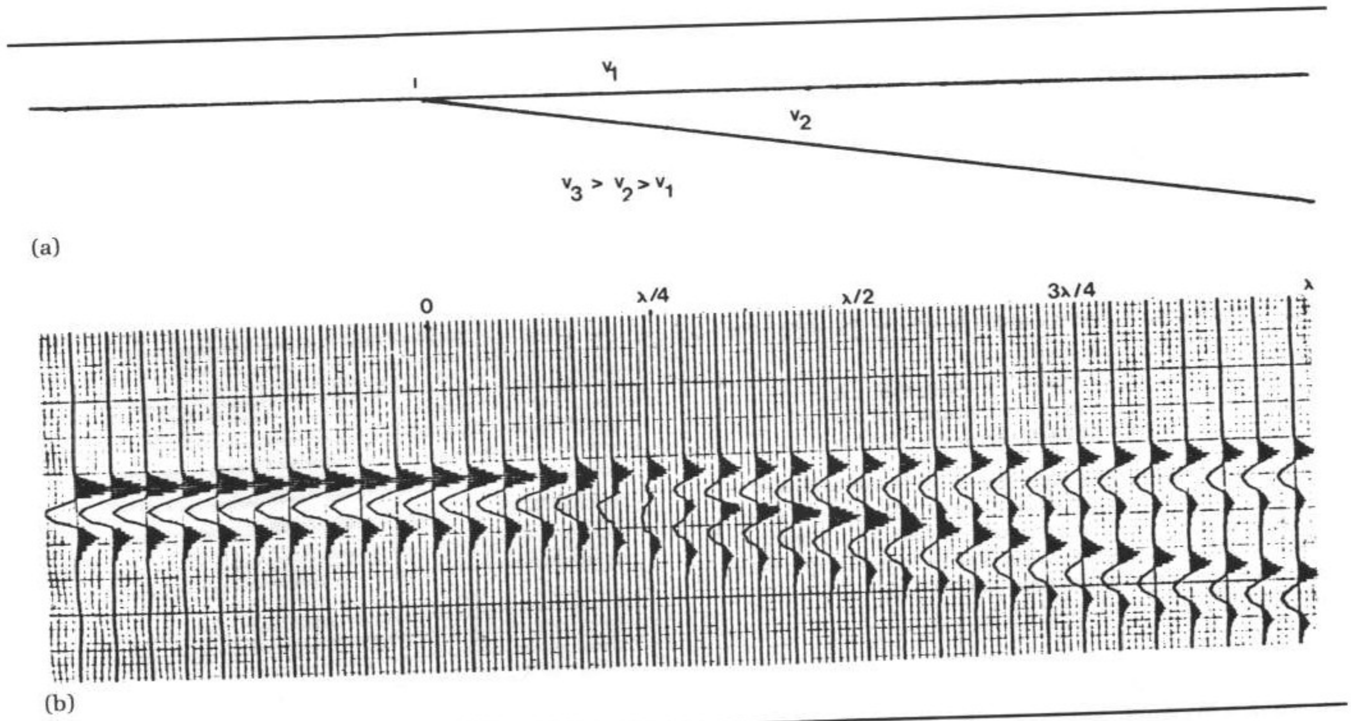


FIGURE 2.16 *Reflection from a wedge of acoustic impedance intermediate in magnitude between that of the over- and underlying units. The thickness of the wedge is indicated as a fraction of the dominant wavelength. Note that there is still a reflection beyond the limit of the wedge due to the contrast in acoustic impedance between the over- and underlying layers.* (a) *Model.* (b) *Seismic section. Reprinted by permission of the EAEG from Sheriff, 1975, fig. 6.9, p. 128.*

a layer to give a reflection, is of the order of 1/30 of the wavelength.

In situations where it is necessary to estimate bed thickness for beds thinner than one-quarter wavelength, modeling techniques can be used to relate amplitude and thickness. Theoretically, the thickness of thin beds below the quarter-wavelength threshold can be estimated from variations in reflection amplitude. Figure 2.17 shows the build-up in amplitude at the quarter-wavelength thickness and the decrease in amplitude as the wedge thins. The main problem with the technique is the need for an amplitude reference (usually from well data) and a detailed control on the amplitude. This usually limits the technique to marine seismic data. The technique requires modification for situations where the reflection coefficients at the top and base of the thin interval have the same sign (see fig. 2.16). In addition to these specialized applications, tuning effects are commonly observed as they affect reflections truncating beneath an unconformity or as they affect onlapping reflections. With knowledge of the likely seismic frequency and interval velocity it may be possible to attempt an estimate of bed thickness at the one-quarter wavelength or tuning thickness in such situations.

HORIZONTAL RESOLUTION

Although it is often convenient to visualize seismic reflections as single rays emanating from a point (e.g., fig. 2.1) actual reflections result from the interaction of a reflecting boundary and a seismic wavefront. The wavefront impinges not just upon a single point, but upon a considerable area of the reflector surface. The resulting reflection is actually produced from a circular zone of quite large diameter. The extent of the area producing the reflection is known as the *Fresnel zone*. This is the portion of the reflector from which energy returns to the geophone or hydrophone within a half-cycle (i.e., one-quarter wavelength) after the onset of the reflection. Contributions from this zone sum constructively to produce a reflection (fig. 2.18). On an unmigrated section, horizontal resolution is determined by the size of the Fresnel zone.

Figure 2.19 shows a model of Fresnel effects. The model shows a continuous bed to the left and three isolated units with lateral extents expressed in terms of Fresnel zones. Each unit thins and pinches out over a shorter distance. The model indicates the following predicted responses:

1. A diffraction is associated with each bed termination.
2. The amplitude of each diffraction decays rapidly away from its apex.
3. The diffractions show polarity reversed on opposing limbs.
4. The gaps between the units are largely masked by the diffractions.
5. A unit of only ½ Fresnel zone produces a seismic re-

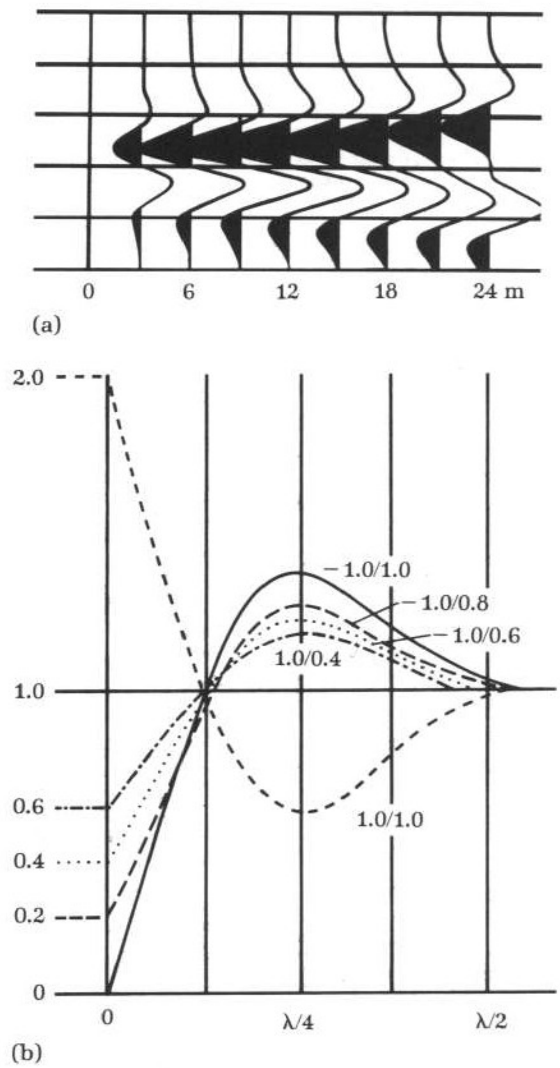


FIGURE 2.17 *Reflections from a wedge, enclosed in material of constant acoustic impedance.* (a) *Reflection waveshape; the wedge thicknesses are given in meters for a velocity of 1525 m/s.* (b) *Maximum peak-to-trough amplitudes where reflectivity at top and base of the wedge has ratios 1.0/1.0 [conditions for dim spot at one-quarter wavelength], −1.0/1.0, −1.0/0.8, −1.0/0.6, −1.0/0.4. The −1.0/1.0 curve applies to the wedge in (a). Thickness is given in wavelengths. Reprinted by permission of IHRDC Press from Sheriff, 1980a, fig. 8.6, p. 169; after Neidell and Poggiagliolmi, 1977.*

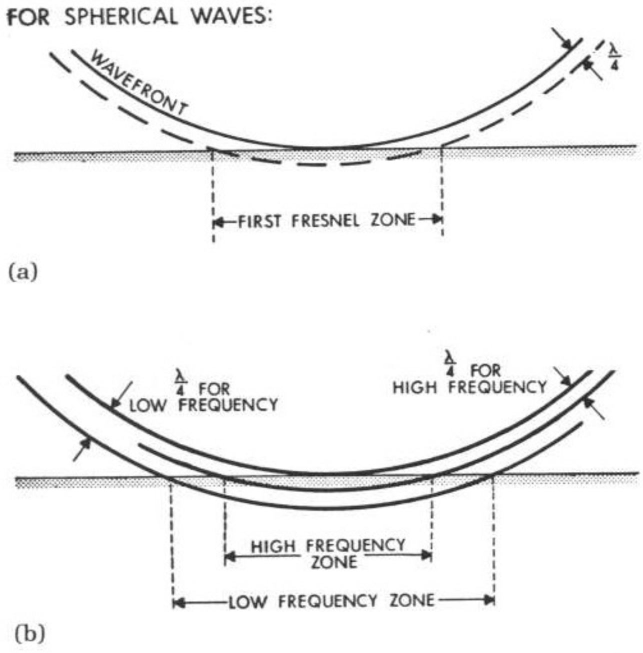


FIGURE 2.18 *Fresnel zone.* (a) *The first energy to reach a geophone from a plane reflector is from the point where the reflector is first tangent to the wavefront. The area of the reflector that produces the reflection is limited by the area that the wavefront one-quarter wavelength later makes with the reflector. The energy arriving within this interval sums constructively to produce the reflection.* (b) *The Fresnel zone is larger for low-frequency components than for high-frequency ones. Reprinted by permission of the AAPG from Sheriff, 1977, fig. 7, p. 11.*

sponse that is indistinguishable from a diffraction from a point source. Even with a width of one Fresnel zone the seismic response can only be distinguished with difficulty from that of a simple diffraction.

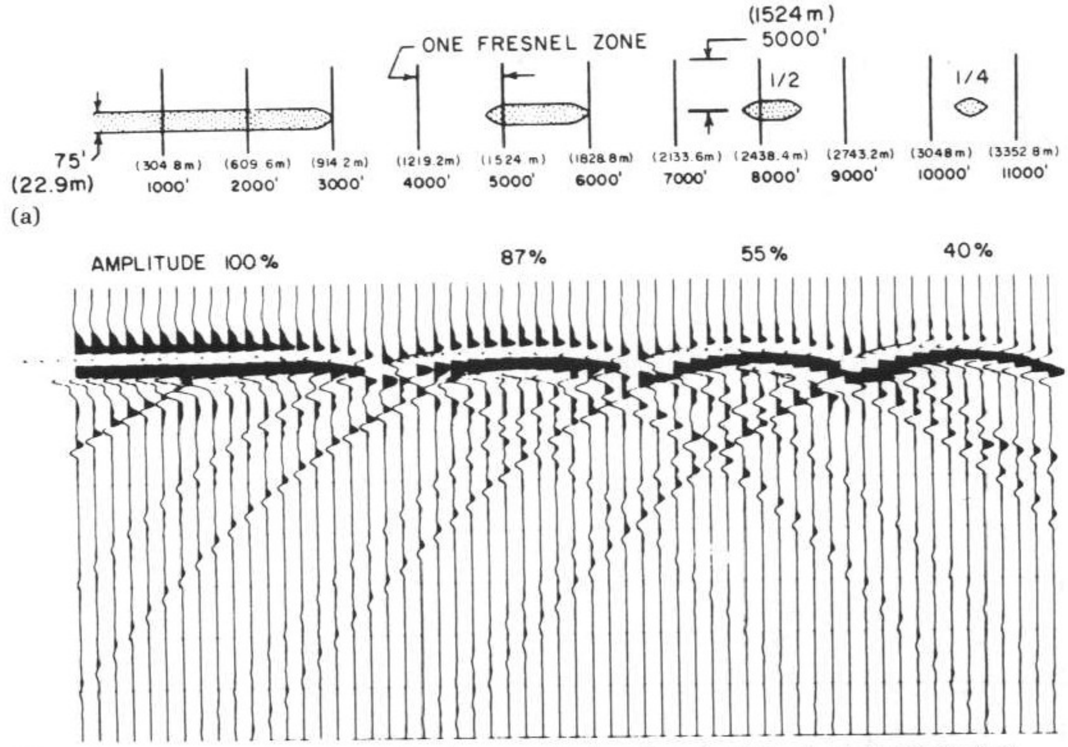
The magnitude of Fresnel zones can be approximated from the relationship

$$rf = \frac{V}{2}\sqrt{\frac{t}{f}} \tag{2.5}$$

where

rf = radius of the Fresnel zone.
V = average velocity.
t = two-way time in seconds.
f = dominant frequency in hertz.

For example, a reflection at 1.7 seconds with a 35-Hz component corresponds to a Fresnel-zone radius of 275 m (equivalent to an area of 0.237 km^2), for an average veloc-



(b) Wave-theory model response for sandstone bodies of varying lateral extents illustrating significance of Fresnel zone size.

FIGURE 2.19 *Reflections from reflectors of varying limited extents. (a) Cross-section of model: vertical lines are spaced at the Fresnel-zone size. (b) Seismic section over the model. The peak amplitudes of the four reflections are, respectively, 100%, 87%, 55%, and 40%. Reprinted by permission of the AAPG from Meckel and Nath, 1977, fig. 5, p. 421; after Neidell and Poggiagliolmi, 1977.*

ity of 2500 m/s; and a reflection at 3.5 seconds with a 20-Hz component corresponds to a Fresnel-zone radius of 732 m (equivalent to an area of 1.68 km^2), for an average velocity of 3500 m/s.

Horizontal resolution decreases with depth, increasing velocity, and lower frequency. A deep feature has to have a larger areal extent to produce the same effect as a smaller, shallower feature. If, for example, there is a hole in a reflector smaller than the Fresnel zone, the Fresnel zone extends onto the reflector surrounding the hole so that the reflection appears to be continuous across the hole. Such a hole might represent a channel cut into horizontal beds. There are, however, changes (such as in the amplitude) that can be used to distinguish the situation.

Another illustration of the effect of the Fresnel zone is shown by the box model of figure 2.20. The box could represent, for example, a pinnacle reef. A seismic line to the side of the box shows a reflection from the box (as well as the continuous reflection from the main reflector), because part of the Fresnel zone extends onto the box. The reflection from the top of the box produces a spurious event on section 4, a sideswipe reflection. Sideswipe is the most detrimental of the Fresnel zone effects. Conventional 2-D migration, which repairs much of the horizontal resolution loss due to the Fresnel zone effect, is unable to reposition correctly sideswipe reflections.

Fresnel-zone effects are often seen in the vicinity of faults. Figure 2.21 shows how the amplitude of a reflector decreases in the vicinity of a fault because only part of the Fresnel zone is available to produce a reflector.

THE EFFECT OF DEPTH

The above discussions on resolution and interference have highlighted the influence of frequency and velocity on our ability to interpret features in the seismic section. Unfortunately, both velocity and frequency show an overall change with depth that causes a gradual decrease in resolution. This is rather unwelcome, for if any of the factors vary, so will our perception of the reflection pattern. This can have a major effect on our ability to interpret the seismic section.

Velocity tends to increase with depth due to compac-

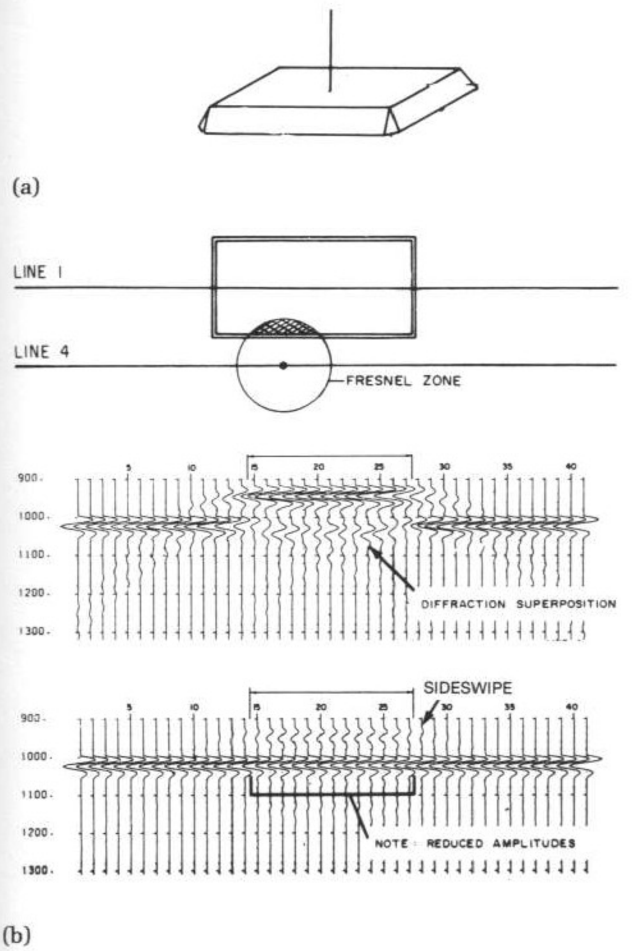
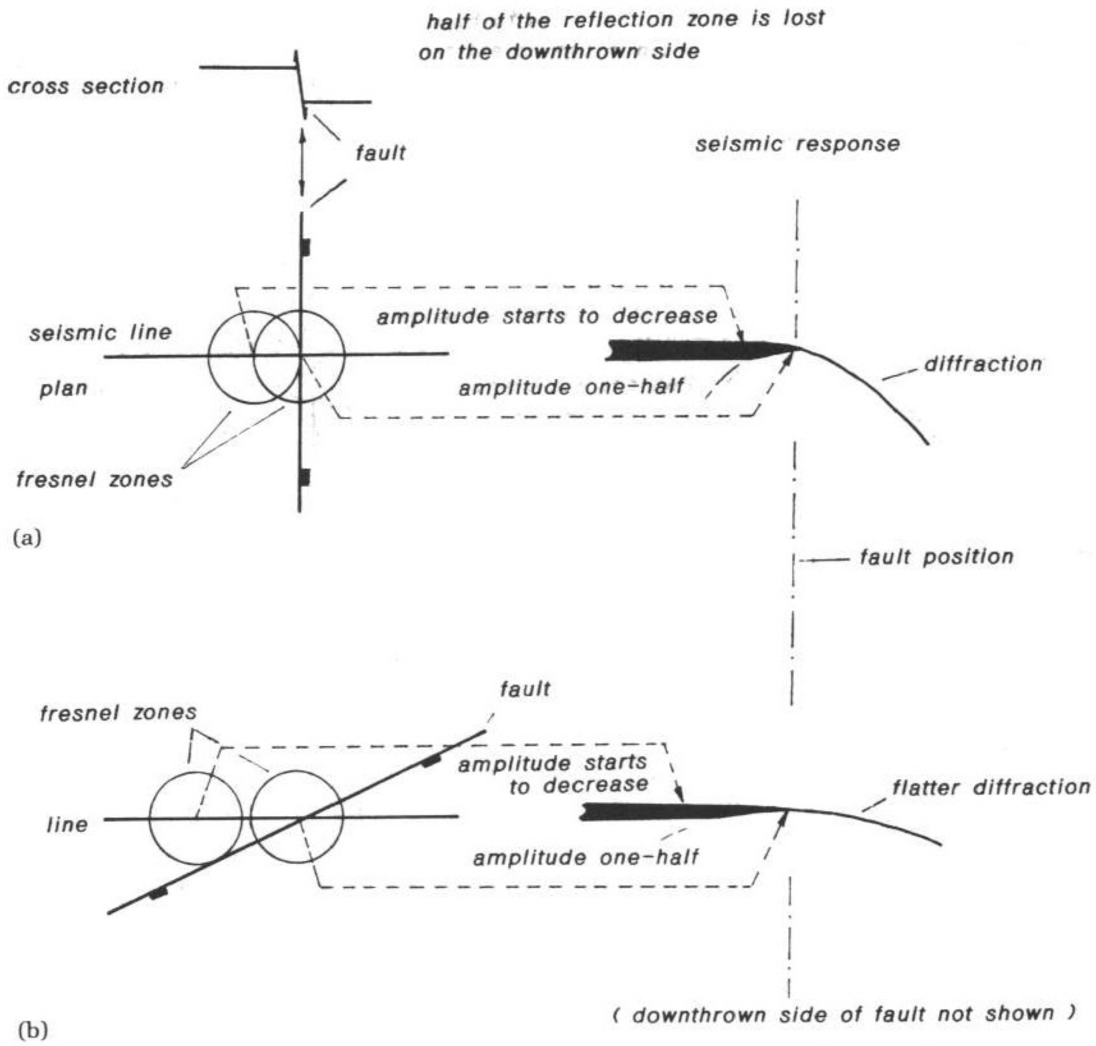


FIGURE 2.20 *Reflections from a box model.* (a) *Isometric diagram of the model; the length : width : height : depth ratios are 10 : 5 : 1 : 10.* (b) *Plan showing Fresnel-zone dimensions relative to box dimensions. Line 1 passes over the box. Line 4 passes to the side of the box, but close enough for part of the Fresnel zone to impinge on the box. Reprinted by permission of IHRDC Press from Sheriff, 1980a; after Neidell and Poggiagliolmi, 1977.*

tion and diagenetic effects. Frequency decreases due to attenuation of the seismic wave—there is an almost constant fractional energy loss per cycle of the seismic wave; and higher frequencies are attenuated more than lower frequencies for a particular path length. From this we can conclude that with increasing depth vertical and lateral resolution decreases, and interference effects become more pronounced as the pulse length increases (due to lower frequency). The loss of frequency with depth makes it less likely that a bed's thickness will exceed the minimum one-quarter wavelength for potential resolution of reflections from its top and base, or even the half-wavelength required for no interference between reflections from a unit's top and base.

The fact that there is an overall decrease in strength of contrasts in acoustic impedance as burial depth increases in a further problem that results in weaker reflections. This comes about through the gradual convergence of acoustic-impedance values for different lithologies with increasing compaction. Although acoustic impedance increases as both velocity and density increase, not all lithologies are affected to the same degree. Whereas limestone may show only modest increases in acoustic impedance with burial, a clay, as it compacts and converts into a claystone, will show a significant increase in acoustic impedance. The net effect of this process is to produce generally lower acoustic-impedance contrasts and thereby reduced reflection coefficients. Lower reflection coefficients, combined with the ever-decreasing energy and lower frequency of the seismic wave, ultimately limits the ability of the seismic reflection to detect and resolve boundaries. Figure 2.22 illustrates how the seismic response varies with depth, even for the same pair of lithologies. Limestone is overlain by clay, producing a large positive reflection coefficient in the shallow subsurface. The seismic wavelet here is also of relatively high frequency and the seismic response is a sharp, high-amplitude reflection. As the lithologies become more deeply buried, the clay compacts and converts to shale and the limestone gradually reduces in porosity. Acoustic impedance increases for both lithologies—but at a greater rate in the shale. This results in a real decrease in acoustic-impedance contrast with increasing burial. By the time the shale and limestone are buried to 5000 m or more, the difference in acoustic impedance may not be especially large. The decrease in acoustic-impedance contrast results in gradually weaker reflections from the same lithology pair with increasing depth. The reflections not only become weaker; they also change their shape. The Earth preferentially attenuates the higher frequency part of the seismic signal with increasing traveltime, resulting in a gradual increase in wavelength with depth. This wavelength increase changes the shape of the reflection (fig. 2.22).

Figure 2.23 shows the effect of different wavelet length (frequency) on interference effects. Looked at in another way, the four seismic traces shown in figure 2.23 could be used to model the seismic response for the lithological column *a* at different burial depths. At the shallowest depth, *f*, in figure 2.23 the seismic response has the highest frequency, but this decreases with depth resulting in poorer resolution. Responses *e*, *d*, and *c* would represent the changing waveform as frequency decreases with increasing burial.



Finally, figure 2.24 shows part of a real seismic section. In the interval down to 2 s the reflections are strong and lots of geological detail is indicated. The interval below 3 s is quite different. There are fewer strong continuous reflectors and the overall appearance suggests a much simpler geology. But beware, this could be an unwarranted conclusion. The loss of detail and change of reflection character of the deeper section is unlikely to be due to simpler geology, but rather to effects of lower seismic pulse frequency and lower reflection coefficients in the deeper geology.

FIGURE 2.21 *Effect of Fresnel zones on reflection amplitude near to a fault.* (a) *Fault perpendicular to the seismic line that juxtaposes a reflector against material of constant acoustic impedance. Reflection amplitude is constant until the Fresnel zone passes over the fault. Reflection amplitude decreases and reaches half of its former value when the center of the Fresnel zone is coincident with the fault.* (b) *Fault oblique to the seismic line. The decrease in reflection amplitude is spread over a larger portion of the seismic line. After Anstey, 1980a, by permission of IHRDC Press.*

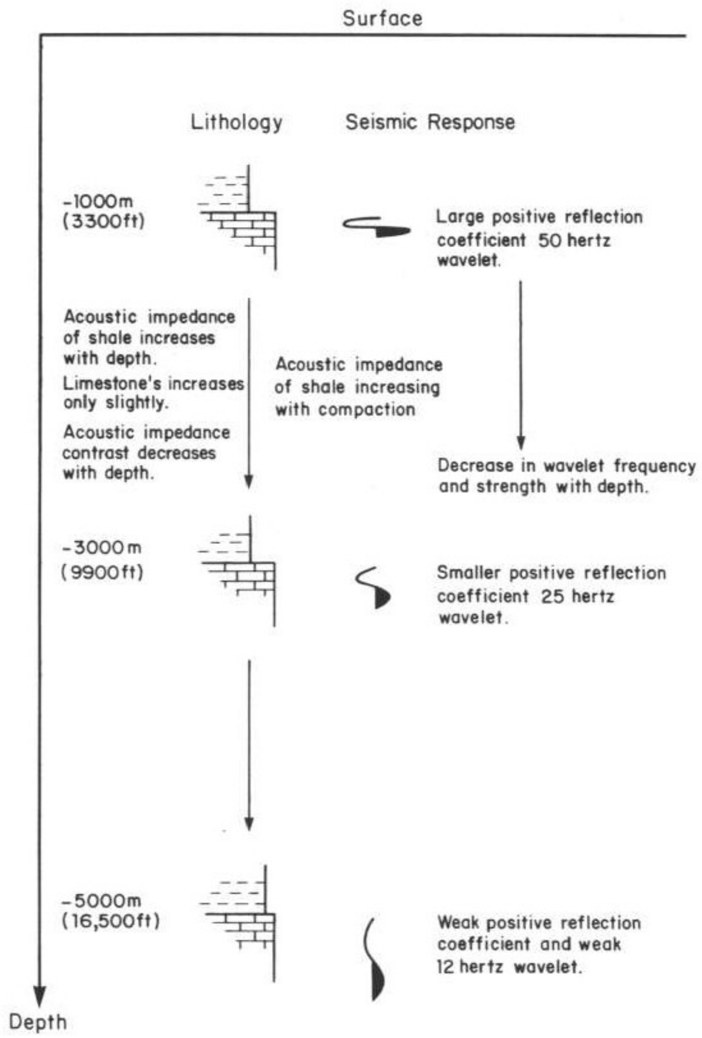


FIGURE 2.22 *Sketch showing the effect of increasing depth, through changes in acoustic impedance and seismic-wavelet frequency, on the seismic response from an acoustic-impedance contrast between clay and limestone. The effects of burial, especially compaction, affect clays more than limestone. Although both show an overall increase in acoustic impedance with depth, the rate of increase is greater in the claystones. This results in weaker reflections with depth. The Earth attenuates the seismic signal, preferentially removing the higher frequency components. The gradual loss of frequency with depth changes the wavelength and shape of the seismic wavelet. The combined effect of decreasing frequency and acoustic-impedance contrasts produces a change from shallow, strong, sharp reflections to weak, long reflections at depth.*

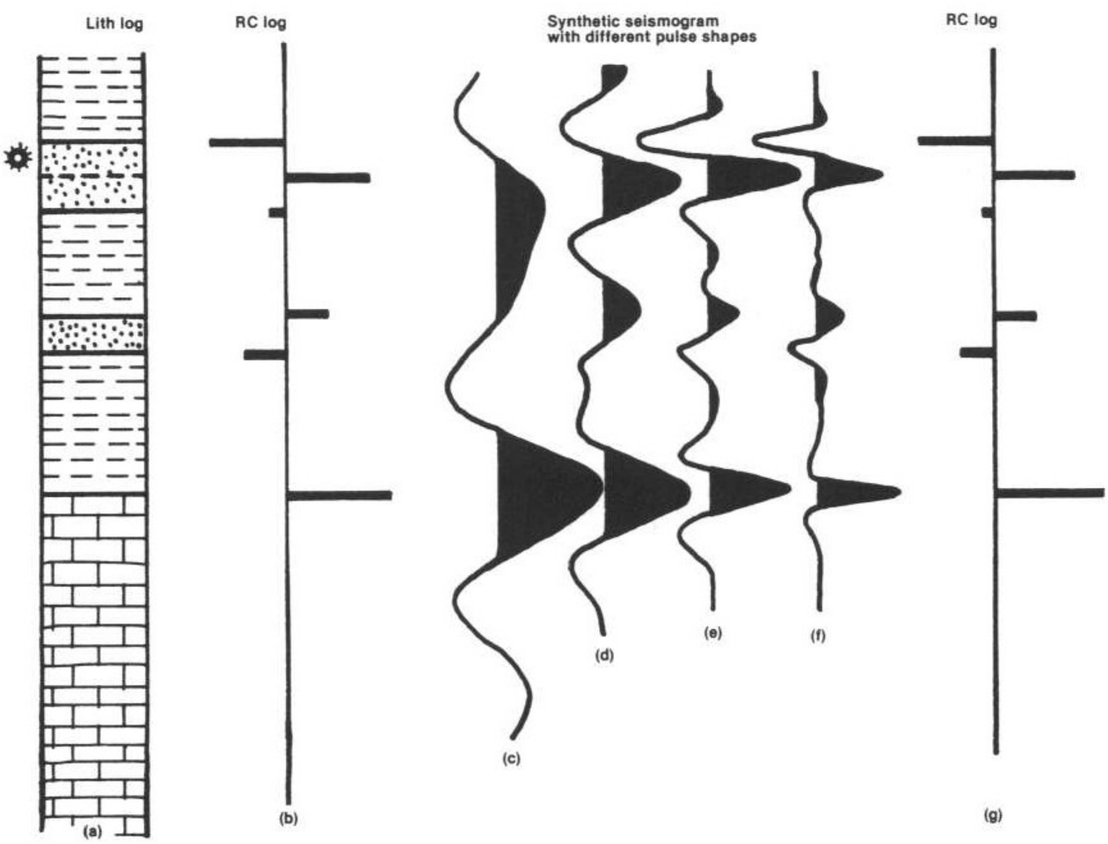


FIGURE 2.23 *The effect of wavelet frequency on the seismic response. Reprinted by permission of IHRDC Press from Anstey, 1980a.*

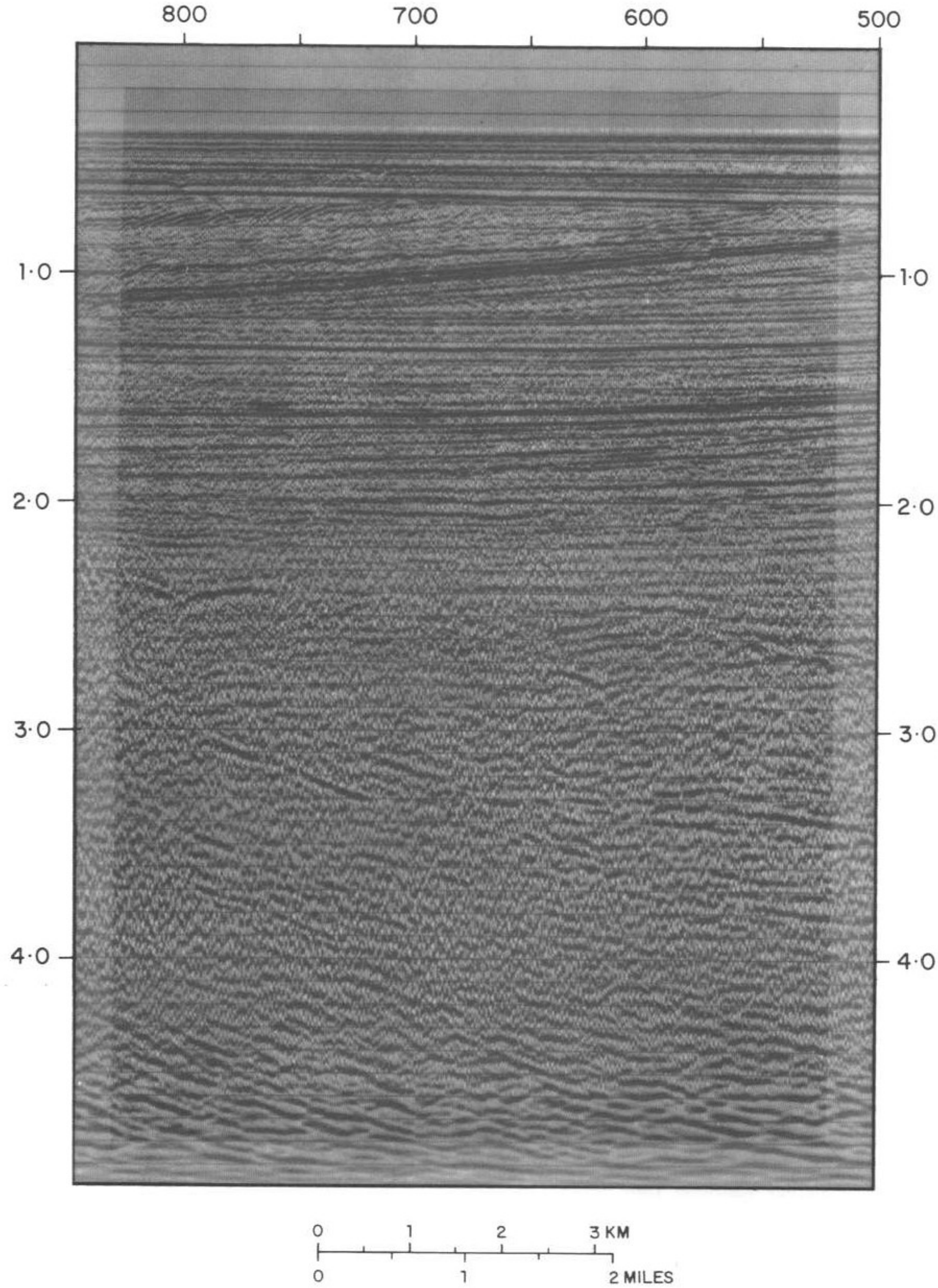


FIGURE 2.24 *Seismic section showing decreasing frequency (increasing reflector spacing) and reflection strength with depth. Courtesy Norsk-Hydro.*