Michael E. Badley

Practical Seismic Interpretation

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Preface

This book is about the practical aspects of reflection seismic interpretation, written by a working explorationist for explorationists. It is neither mathematical nor theoretical and does not claim to give a comprehensive treatment of the seismic-reflection method. The more geophysical topics, such as acquisition, processing, modeling, and so on, are not dealt with in any detail. The book does deal, however, with many of the everyday practicalities and problems facing the working interpreter; and, as such, has a geological bias. The book is aimed especially at those new to interpretation, who, when confronted with a pile of fresh seismic sections, wonder where to start the interpretation, what to pick, how to recognize multiples, and so forth.

Reflection seismic has made great inroads into the field of exploration over the last twenty years, and in many respects technology has advanced faster than developments in interpretation techniques. From modest beginnings, seismic reflection is now an indispensable exploration technique. This is especially true in marine areas where wells are generally relatively few in number and usually widely spaced. Continued improvement in the quality of seismic data has not only increased their importance, but has also increased the volume of data being collected, both during data acquisition and in the density of the grids. The advent of 3-D surveys has brought with it the new problem of how to cope with the vast volume of data produced. Nowadays there are few people in exploration who do not have direct contact with seismic data. Increasingly, seismic interpretation is the common ground between the geologist and the geophysicist. However, this was not always the case. Over twenty years ago Dobrin (1960) in his book *Introduction to Geophysical Prospecting* felt the need to write:

The variable-density and variable-area sections give a particularly convincing illusion that they are actual pictures of geologic formations below the surface. Manufacturers of these devices (plotters producing the sections) have often suggested that such sections could be turned over to the geologist or to management as a final presentation of the geophysical data in geological terms. Such a practice would represent an abdication of the geophysicist's responsibilities. Unless the record sections are carefully interpreted by the geophysicist before being turned over to the geologist, their introduction may have to be looked upon as a backward step in the exploration process.

This was written at a time when exploration departments were polarized into two camps, namely, the geologists and the geophysicists. How relevant are such sentiments today? Certainly seismic sections have improved dramatically and look more and more akin to geological cross-sections. But some of Dobrin's words of caution still hold good: careful interpretation is still of paramount importance! What has changed most is the interpretation of sections and, indeed, the interpreters. Seismic interpretation, in common with many other modern high technologies, for example, computing, has with technological advance moved away from the realm of the specialist geophysicists into the domain of a wider geological public. No longer are seismic sections 70% noise and artefact with primary events peeping through; indeed, the reverse is now usually true. Previously the interpreter's main task was to extract the real data from the welter of noise. This required not so much geological knowledge, as detailed knowledge of the seismic system itself. Today, assuming that the interpreter has a basic understanding of the seismic method and its limitations, an interpretation is rooted more in basic geological principles, but with the details of acquisition and processing still retaining an important role.

So what does the new situation demand of the interpreter who is no longer, of necessity, a geophysicist? Firstly, as before, the interpreter must have a thorough grasp of the seismic method so that both its potential and its limitations can be appreciated. Secondly, the interpreter must have a more comprehensive geological background if the full potential of the seismic method is to be realized. To meet these requirements, this book is divided into two halves. The first half covers the areas of knowledge that every interpreter needs to have before starting an interpretation; the second half deals more with the practicalities and techniques of the interpretation itself.

There is a saying that the best geologist is the one who has seen the most rocks. Perhaps this is also true of interpreters. The best interpreter is the one who has seen the most seismic sections: it is all a question of experience. Although we should perhaps replace "most" with "greatest variety from different areas" and add the extra qualification "and has been around long enough to see the drilling results and find out if the interpretation was right, wrong, or (more usually) somewhere in between."

Good picking!

Acknowledgments

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Contents

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

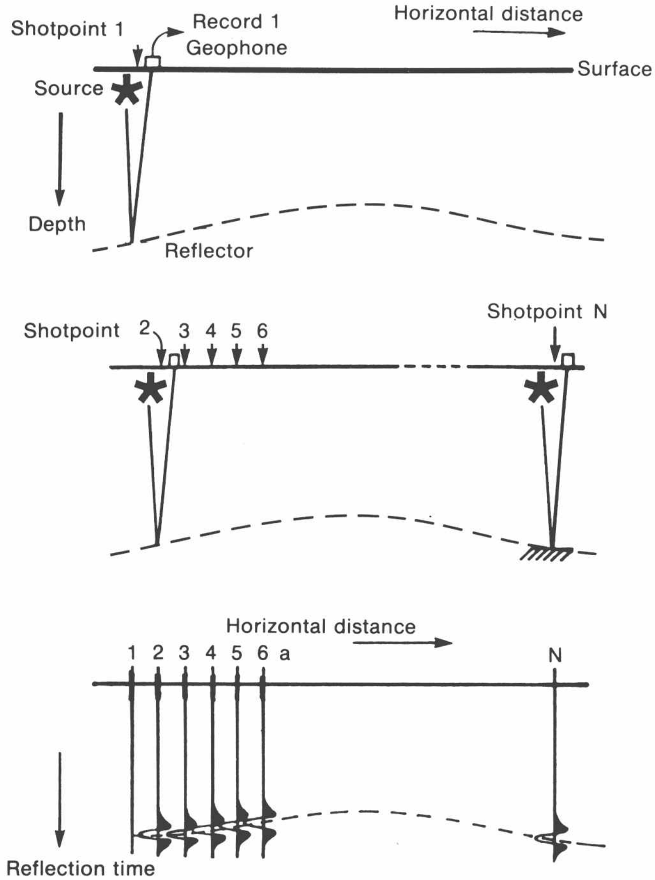

FIGURE 1.1 *The basic concept: a reflection-time measurement at regular intervals along a line, and the representation of those measurements in the form of a seismic section. Reprinted by permission of IHRDC Press from Anstey, 1982.*

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-

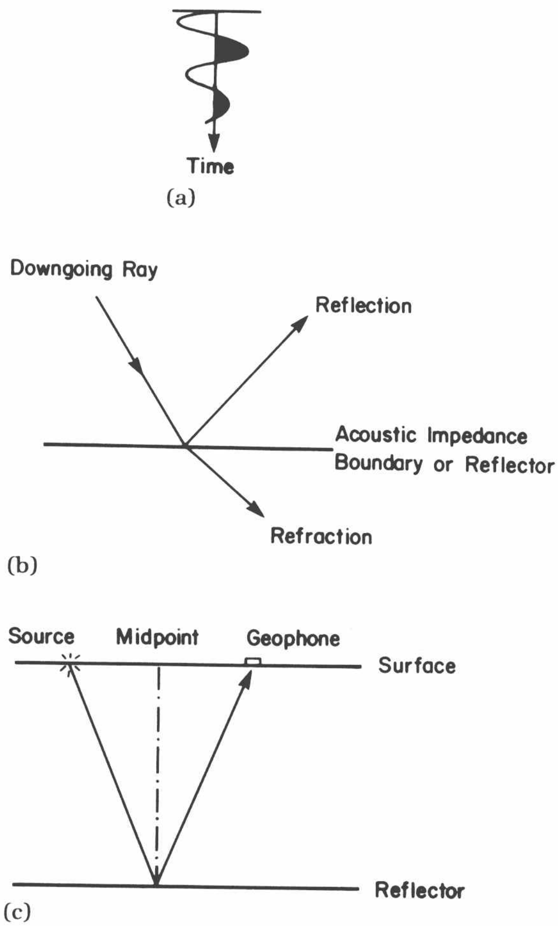

FIGURE 2.1 *The basic elements of the seismic-reflection method.* (a) *Diagrammatic source wavelet.* (b) *Reflection and refraction at an acoustic-impedance boundary.* (c) *Reflection geometry for a horizontal reflector.*

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

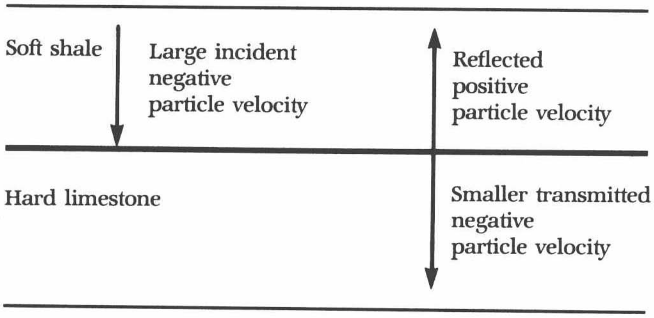

FIGURE 2.2 *Continuity of particle velocity at a soft-to-hard interface. The arrows indicate the raypath direction (i.e., direction of travel). Reprinted by permission of IHRDC Press from Robinson, 1983, fig. 3.15, p. 144.*

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.

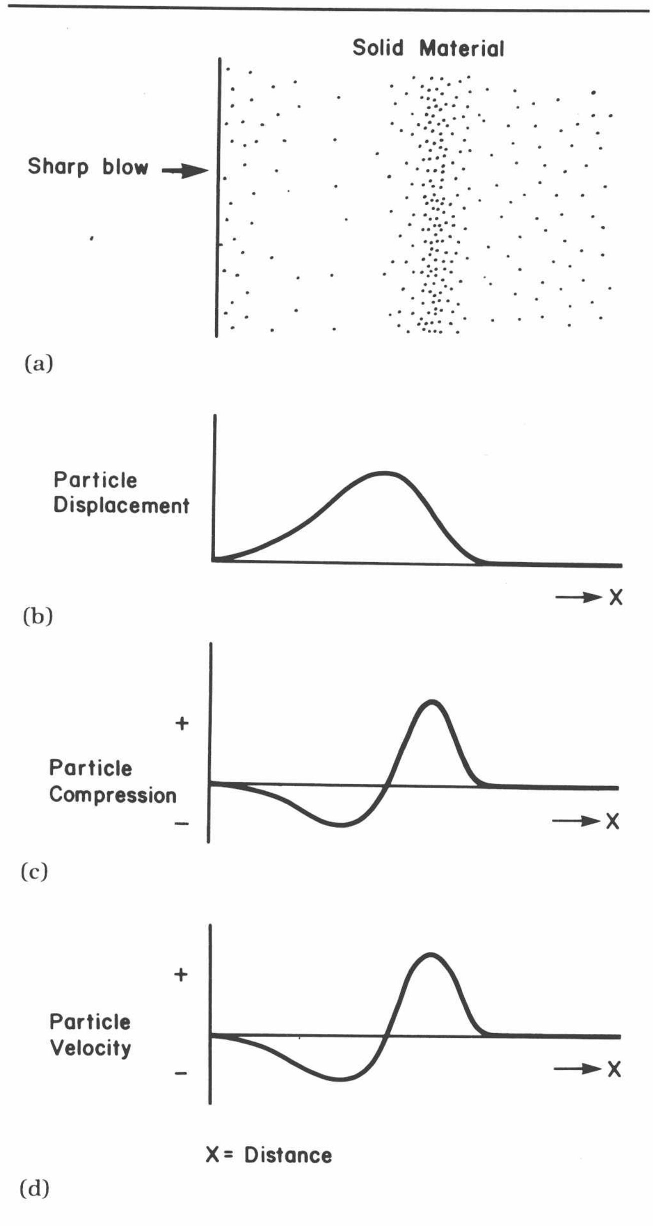

FIGURE 2.3 *Response of particle compression and particle velocity to the passage of a compressional wave.* (a) *Diagrammatic representation of particle spacing in a solid material a few milliseconds after being struck by a sharp blow.* (b) *The maximum particle displacement corresponds with the propagating 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.*

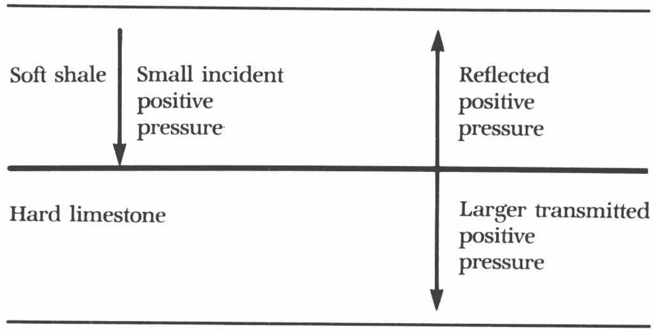

FIGURE 2.4 *Continuity of particle pressure at a soft-to-hard interface. The arrows indicate the raypath direction. Reprinted by permission of IHRDC Press from Robinson, 1983, fig. 3.16, p. 145.*

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

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

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-

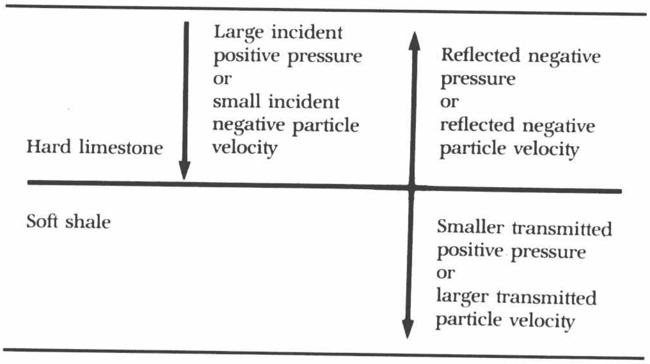

FIGURE 2.5 *Continuity of particle pressure and compression at a hard-to-soft interface. The arrows indicate the raypath direction.*

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.

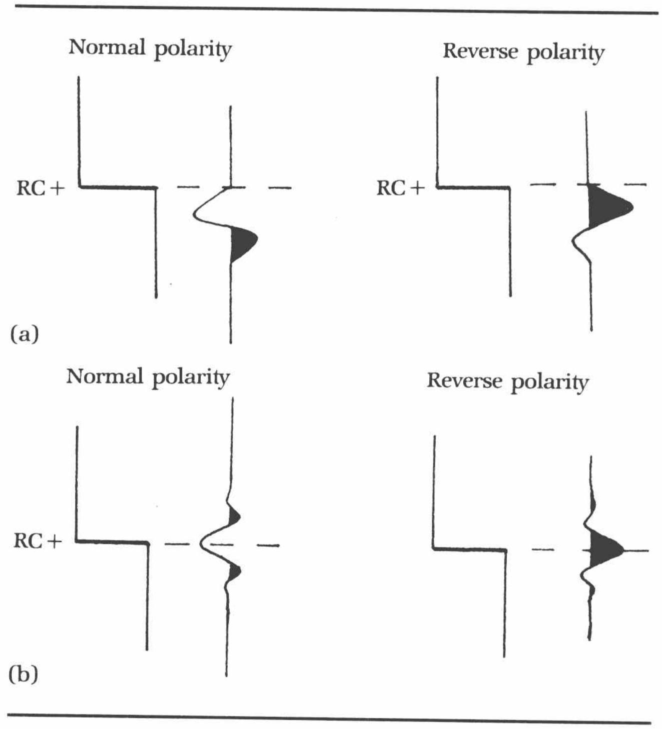

FIGURE 2.6 *Examples of idealized normal and reverse polarity for* (a) *minimum- and* (b) *zero-phase wavelets at an acoustic-impedance boundary with a positive reflection coefficient.*

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.

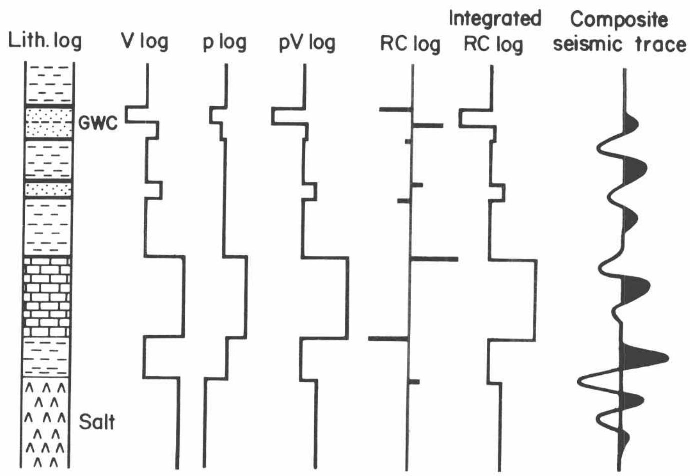

FIGURE 2.7 *The derivation of the reflection coefficient log and the resulting composite trace for a minimum-phase, normal-polarity (SEG) wavelet. The lithological sequence shown at the left consists of: a basal salt section overlain by a thick shale, massive limestone, and a shale sequence containing 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