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Excerpt
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Chapter 1

Overview

1.1 Introduction

This book is about the physical interpretation of seismic amplitude principally for the purpose of finding and exploiting hydrocarbons. In appropriate geological scenarios, interpretations of seismic amplitude can have a significant impact on the 'bottom line'. At all stages in the upstream oil and gas business techniques based on the analysis of seismic amplitude are a fundamental component of technical evaluation and decision making. For example, an understanding of seismic amplitude signatures can be critical to the recognition of direct hydrocarbon indicators (DHIs) in the exploration phase as well as the evaluation of reservoir connectivity or flood front monitoring in the field development phase. Given the importance of seismic amplitude information in prospect evaluation and risking, all technical disciplines and exploration/asset managers need to have a familiarity with the subject.

1.2 Philosophy, definitions and scope

The central philosophy is that the seismic interpreter working in exploration and appraisal needs to make physical models to aid the perception of what to look for and what to expect from seismic amplitude responses in specific geological settings. This usually involves the creation of synthetic seismic models for various rock and fluid scenarios based on available well log data. In rank exploration areas the uncertainties are generally such that only broad concepts, assumptions and analogies can be used. By contrast, in field development settings where data are readily available, physical modelling can lead to a quantification of reservoir properties from seismic (with associated error bars!).

Fundamental to this process of applying models in seismic interpretation is the integration of data from a variety of disciplines including geology, geophysics, petrophysics and reservoir engineering. The core aspect of the data integration is rock physics, which can be defined as the study of 'the relationships between measurements of elastic parameters (made from surface, well, and laboratory equipment), the intrinsic properties of rocks (such as mineralogy, porosity, pore shapes, pore fluids, pore pressures, permeability, viscosity, and stress sensitivity) and overall rock architecture (such as laminations and fractures)' (Sayers and Chopra, 2009). Rock physics effectively provides the rock and fluid parameters for seismic models.

Pennington (1997) describes 'the careful and purposeful use of rock physics data and theory in the interpretation of seismic observations' and calls this approach 'Seismic Petrophysics'. Others commonly refer to it simply as 'Rock Physics' (*sensu lato*), 'Seismic Rock Physics' (Wang, 2001b) or Quantitative Interpretation (QI). The mind-set which drives the approach is not new of course but modern data have provided a new context. More than ever before there is an opportunity, paraphrasing Sheriff (1980), to 'reveal the meaning of the wiggles'. There are numerous workers, past and present, to whom the authors are indebted and whose names occur frequently in the following pages.

The book describes the theory of seismic reflectivity (Chapter 2) and addresses the key issues that underpin a seismic interpretation such as phase, polarity and seismic to well ties (Chapters 3 and 4). On these foundations are built a view of how contrasts in rock properties give rise to variations in seismic reflectivity (Chapter 5). Seismic data quality is an all important issue, controlling to a large extent the confidence in an interpretation, and this is addressed, from an interpreter's perspective, in Chapter 6. Examples of fluid and rock interpretations using Amplitude Versus Offset (AVO) techniques are presented in Chapter 7 for a variety of geological contexts. The key rock physics components that drive seismic models are documented in Chapter 8, whilst the concept of seismic trace inversion is introduced in Chapter 9. Chapter 10 outlines some key applications of seismic amplitudes such as the description of reservoir properties from seismic and the use of amplitude information in prospect evaluation and reserves determination.

1.3 The practice of seismic rock physics

The practice of seismic rock physics depends to a large extent on the application. In some cases, simply fluid substituting the logs in a dry well and generating synthetic gathers for various fluid fill scenarios may be all that is needed to identify seismic responses diagnostic of hydrocarbon presence. On the other hand, generating stochastic inversions for reservoir prediction and uncertainty assessment will require a complete rock physics database in which the elastic properties of various lithofacies and their distributions are defined in an effective pressure context. Either way, the amount of knowledge required to master the art of seismic rock physics is a daunting prospect for the seismic interpreter.

The broad scope of the subject inevitably means that geophysicists need to work closely with petrophysicists, geologists and engineers. Often this is easier said than done. To quote Ross Crain (2013): 'Geophysicists engaged in seismic interpretation seldom use logs to their full advantage. This sad state is caused, of course, by the fact that most geophysicists are not experts in log analysis. They rely heavily on others to edit the logs and do the analysis for them. But, many petrophysicists and log analysts have no idea what geophysicists need from logs, or even how to obtain the desired results'.

Effectively, the use of rock physics in seismic interpretation blurs the distinctions between subsurface disciplines. This book introduces the subject from a practical viewpoint with a description of how it works and how connections are made between the various disciplines. Whilst there is good practice, there is no single workflow to follow. It is hoped that the perspective presented here will be a source of encouragement to those eager to learn the trade as well as providing ideas for creative hydrocarbon exploration and development.

Chapter 2

Fundamentals

2.1 Introduction

Interpreting seismic amplitudes requires an understanding of seismic acquisition and processing as well as modelling for describing and evaluating acoustic behaviour. Separate books have been written about each of these subjects and there is certainly more to say on these issues than can be presented here. The aim of this chapter is to provide a framework of basic information which the interpreter requires in order to start the process of seismic amplitude interpretation.

2.2 Seismic basics

2.2.1 Seismic geometry

Seismic data are acquired with acoustic sources and receivers. There are numerous types of seismic geometry depending on the requirements of the survey and the environment of operation. Whether it is on land or at sea the data needed for seismic amplitude analysis typically require a number of traces for each subsurface point, effectively providing measurements across a range of *angles of incidence*. The marine environment provides an ideal setting for acquiring such data and a typical towed gun and streamer arrangement is illustrated in Fig. 2.1a. Each shot sends a wave of sound energy into the subsurface, and each receiver on the cable records energy that has been reflected from contrasts in acoustic hardness (or impedance) associated with geological interfaces. It is convenient to describe the path of the sound energy by *rays* drawn perpendicular to the seismic *wavefront*; this in turn clarifies the notion of the angle of incidence (θ in Fig. 2.1a). Usually, the reflections recorded on the near receivers have lower angles of incidence than those recorded on the far receivers.

Figure 2.1b illustrates the recorded signal from the blue and red raypaths shown in Fig. 2.1a. The signal recorded at each receiver is plotted against time (i.e. the travel time from source to receiver), and the receiver traces are ordered by increasing source–receiver distance, usually referred to as *offset*. Plotting the traces for all receivers for one particular source position provides a *shot gather* display. In Fig. 2.1 the reflected energy is shown as a wiggle display and the shape of the reflection signal from the isolated boundary describes the shape of the seismic pulse (the *wavelet*) at the boundary. Owing to the difference in travel path, the arrival time of the reflection from the geological boundary increases with offset and, usually, the relation between travel time and offset is approximately hyperbolic. The amplitude of the reflection from the boundary is related to the contrast in acoustic parameters across the boundary, but is also affected by distance travelled, mainly because the energy becomes spread out over a larger area of wavefront. This phenomenon has commonly been referred to as *spherical divergence*, although it is now evident that wavefronts have shapes between spherical and elliptical. An objective of seismic processing is to produce traces where the amplitudes are related only to the contrasts at the reflecting boundary, and all other effects along the propagation path are removed (this is often referred to as *true amplitude* processing). This can be difficult for land data, where there may be large differences from one trace to the next, related to the effectiveness of the coupling of sources and receivers to the surface, as well as rapid lateral variation in the properties of the shallow zone immediately below the surface.

2.2.2 Gathers and stacks

During seismic acquisition, each shot is recorded by many receivers. Figure 2.2 illustrates that each receiver is recording reflections from different subsurface locations for any given shot. The shot gather therefore mixes together energy from different subsurface locations, and is of little direct use for interpretation. If the Earth is made up of relatively flat-lying layers then the various traces relating to source–receiver pairs

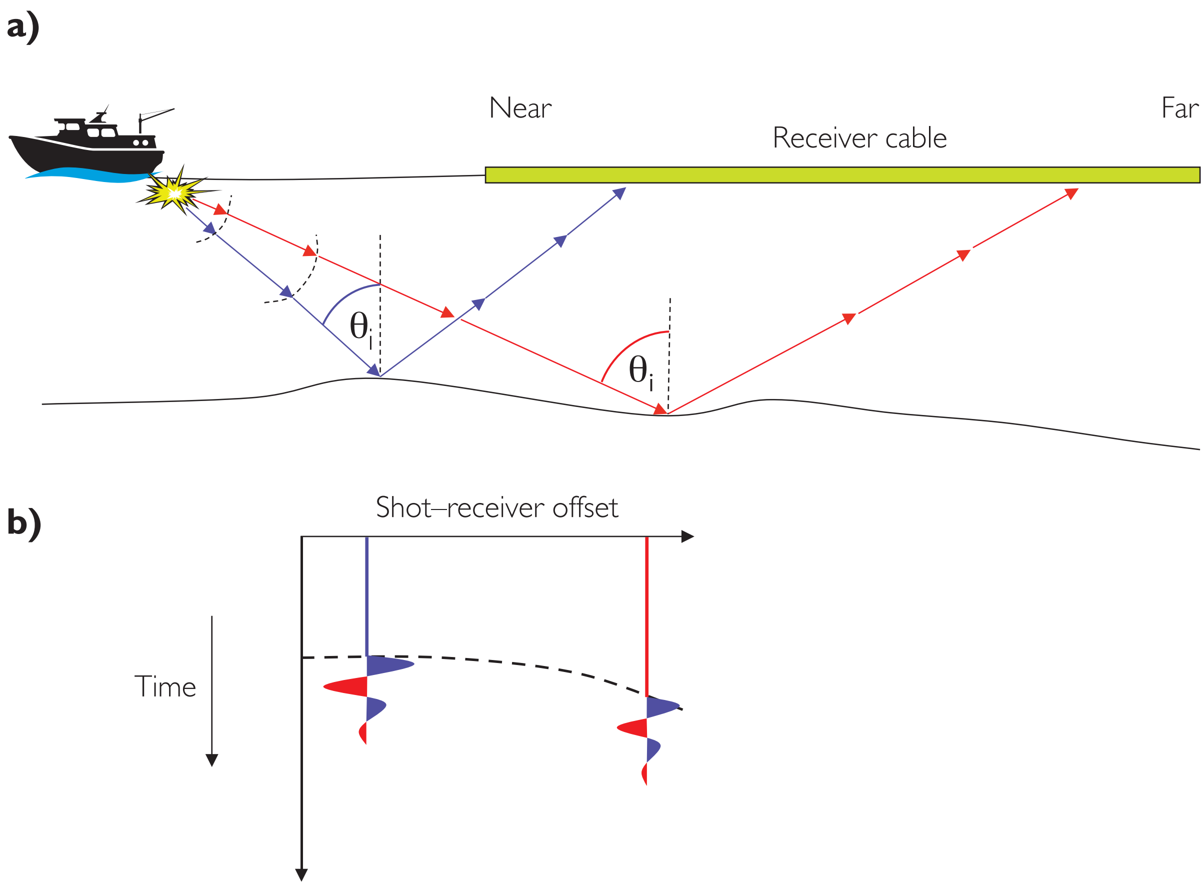

Figure 2.1 Marine seismic geometry 1; (a) source and receiver configuration showing wavefronts, rays (perpendicular to wavefronts) and angle of incidence increasing with offset (b) a shot gather representation of the recorded energy. For a horizontal layered case the reflections form a hyperbola on the gather. Note that the amplitude of a reflection on the gather is related to the rock contrast across the boundary and the decrease in amplitude due to wavefront divergence.

which share a common midpoint (CMP) will also share common subsurface reflection points. These are typically brought together to form a *CMP gather* (Fig. 2.2) and form the basis for further analysis. If the subsurface is not a simple stack of plane layers, it is still possible to create a gather for a common reflection point provided that the subsurface geometry and seismic velocities can be determined reasonably accurately from the data. This is an aspect of seismic *migration*, which attempts to position subsurface reflectors in their true spatial location. There are several different approaches to migration, and a vast literature exists on the subject. Jones (2010) gives a useful overview. For the purpose of this narrative it is assumed that a gather has been produced in which all the traces are related to the same subsurface point at any given reflection time.

In order for the gathers to be interpretable, they need to be processed. Figure 2.3 gives a generalised overview of some of the steps involved. A time-varying gain is applied to remove the effects of wavefront divergence, a mute is applied to remove unwanted signal (typically high-amplitude near-surface direct and refracted arrivals visible on the further offsets at any given time), and pre-stack migration is applied to bring traces into the correct geometrical subsurface location. As shown on the left-hand side of Fig. 2.3, the reflection time of any particular interface on the raw gather becomes later with increasing offset, due to the increased path length. An important step is the application of time-varying time shifts to each trace so as to line up each reflection horizontally across the gather, as shown on the right-hand side of Fig. 2.3. This is needed in conventional processing

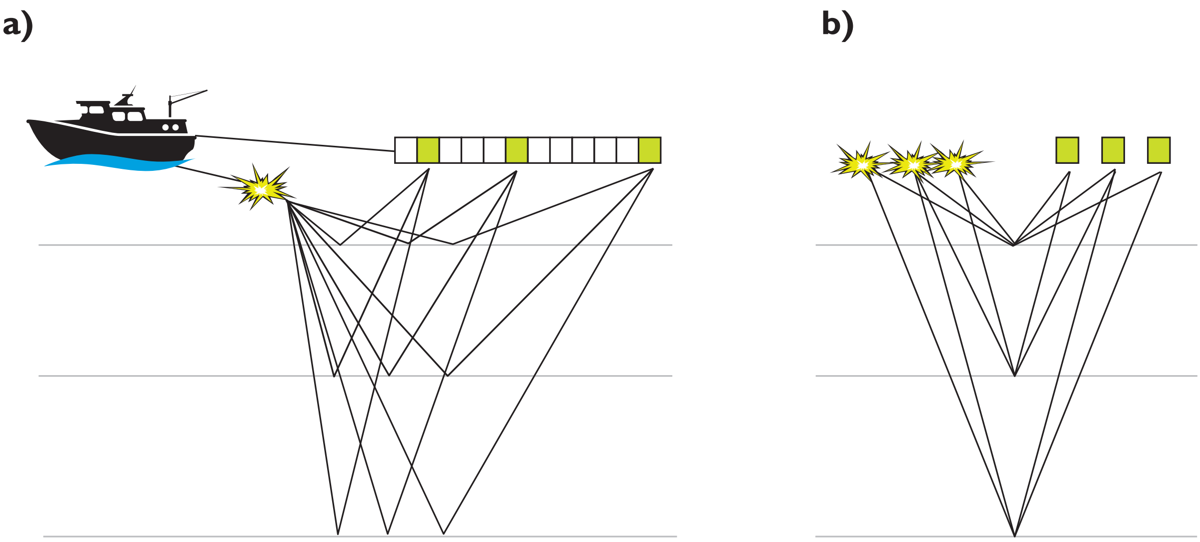

Figure 2.2 Marine seismic geometry 2; (a) acquisition; each shot is recorded at a variety of receivers depending on the depth and the angle of reflection and (b) the common midpoint (CMP); if it is assumed that the Earth is flat the data can be arranged according to reflection location, i.e. different source–receiver pairs sampling the same position in the subsurface. For more complex velocity overburden sophisticated imaging solutions are required.

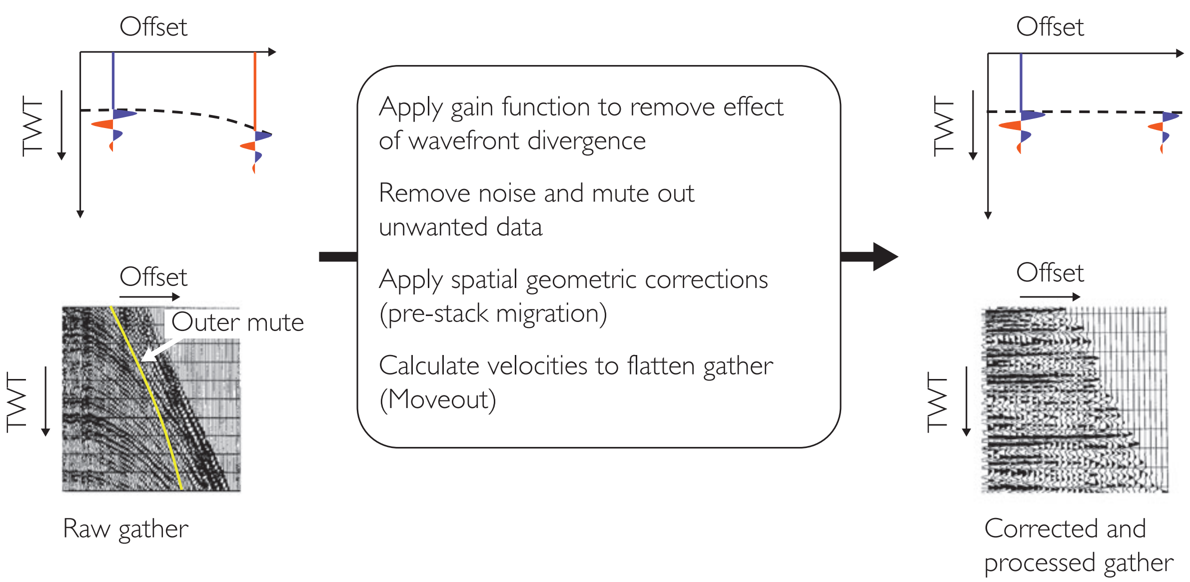

Figure 2.3 Key steps in the processing of seismic gathers.

because the next step will be to *stack* the data by summing the traces of the gather along lines of constant time, i.e. along horizontal lines in the display of Fig. 2.3. This has the important effect of enhancing signal and suppressing noise. Accurate horizontal alignment across the gather is also important for the study of amplitude variation with offset (AVO) described later in this chapter and in Chapter 5. The process of time-shifting to flatten the reflections is called *moveout correction*. A commonly used term is *normal moveout (NMO)*, which refers to the specific case where there is no dip on the reflector.

Figure 2.4 illustrates a stacking methodology that is popular for seismic AVO analysis. Seismic sections have been created by stacking the near-offset data and the far-offset data separately, giving an immediate visual impression of AVO effects but also providing information that can be analysed quantitatively (see Chapters 5 and 7). In this particular case, the difference between near and far trace data is unusually large but it illustrates the point that where there is only a 'full stack' section available (i.e. created by adding together all the traces of the gather) the interpreter may be missing valuable information.

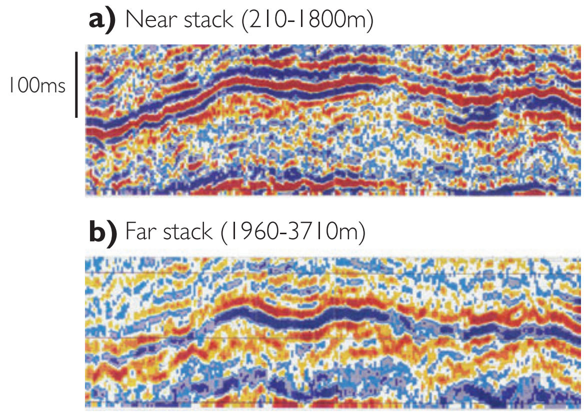

Figure 2.4 Offset stacking example; (a) near stack, 210–1800 m offset, (b) far stack, 1960–3710 m offset. Note that the choice of offsets is dependent on the offset range available and the depth of the target. In this case the offsets have been selected to avoid stacking across a zone of phase reversal.

2.3 Modelling for seismic interpretation

Propagation of seismic energy in the Earth is a complex phenomenon. Figure 2.5 shows some of the numerous factors related to geology and acquisition. The goal of course is to relate seismic amplitude to rock property contrasts across reflecting boundaries but there are several other factors besides geology that also have an influence on amplitude. Some of these are associated with the equipment used for the survey; these include variability of source strength and coupling from shot to shot, variability of sensitivity and coupling from one receiver to another, the directivity of the receiver array (i.e. more sensitive at some

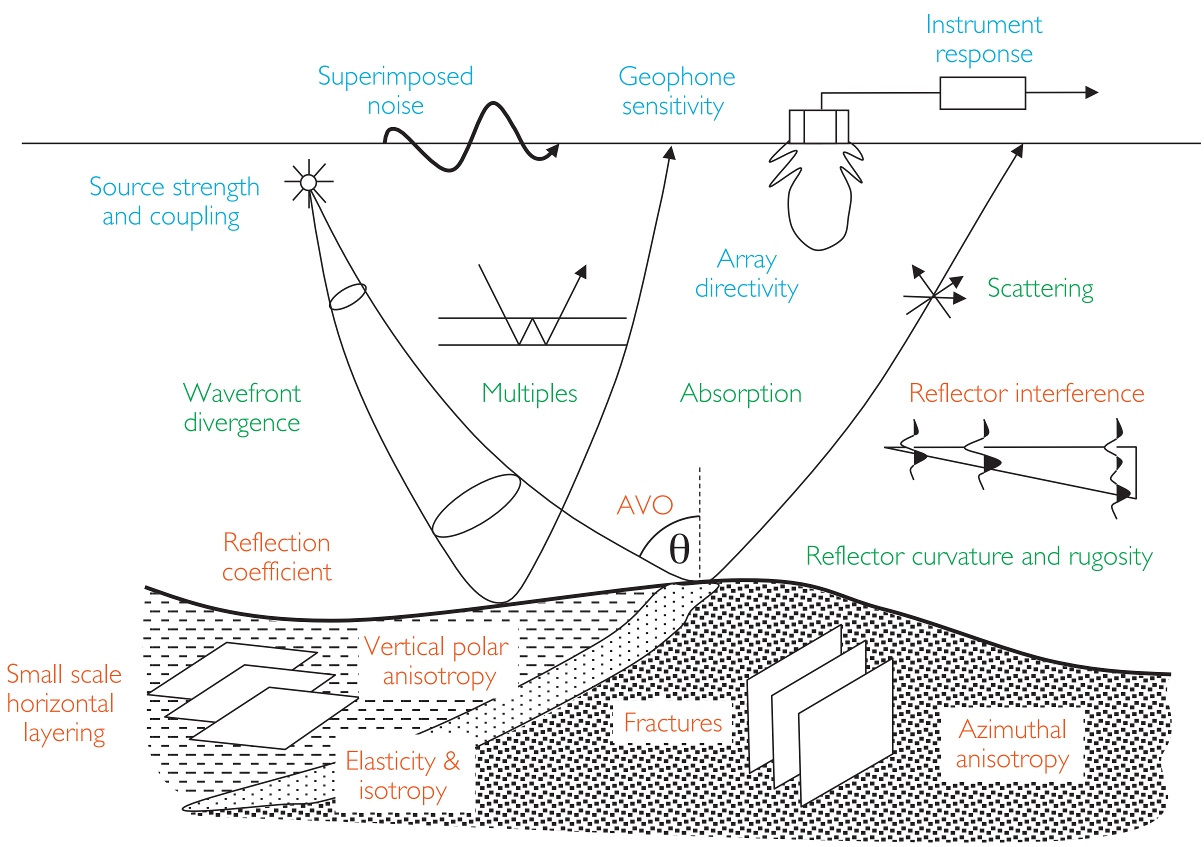

Figure 2.5 Factors affecting seismic amplitude (modified after Sheriff, 1975).

angles of incidence than others) and the imperfect fidelity of the recording equipment. Marine seismic has the advantage that sources and receivers are very repeatable in their characteristics. This is not true for land data, where the coupling of source and receivers to the ground may be quite variable from one shot to another, depending on surface conditions. However, these effects can be estimated and allowed for by the seismic processor.

Some amplitude effects are features of the subsurface that are of little direct interest and ideally would be removed from the data during processing if possible; these include divergence effects, multiples, scattering, reflection curvature and rugosity, and general superimposed noise. Depending on the individual data set, it may be quite difficult to remove these without damaging the amplitude response of interest. For example, processors often have difficulty in attenuating multiple energy whilst preserving the fidelity of the geological signal. Other effects on seismic amplitude, for example related to absorption and anisotropy, might be a useful signal if their origin were better understood. The processor clearly faces a tough challenge to mitigate the effects of unwanted acquisition and transmission factors and enhance the geological content of the data.

Interpretation of seismic amplitudes requires a model. A first order aspect of the seismic model is that the seismic trace can be regarded as the convolution of a seismic pulse with a reflection coefficient related to contrasting rock properties across rock boundaries. This idea is an essential element in seismic processing as well as seismic modelling. Given that the seismic processor attempts to remove unwanted acquisition and propagation effects and provide a dataset in which the amplitudes have 'correct' relative scaling, the interpreter's approach to modelling tends (at least initially) to focus on primary geological signal in a target zone of interest. Of course, one eye should be on the look out for 'noise' remaining in the section that has not been removed (such as multiple energy and other forms of imaging effects). The presence of such effects might dictate more complex (and more time consuming) modelling solutions and can often negate the usefulness of the seismic amplitude information.

From a physical point of view the geological component of seismic reflectivity can be regarded as having various levels of complexity. For the most part, the geological component can be described in simple physical terms. In the context of the small stresses and strains related to the passage of seismic waves, rocks can be considered perfectly elastic (i.e. they recover their initial size and shape completely when external forces are removed) and obey Hooke's Law (i.e. the strain or deformation is directly proportional to the stress producing it). An additional assumption is that rocks are to first order *isotropic* (i.e. rocks have the same properties irrespective of the direction in which the properties are measured). Experience has shown that in areas with relatively simple layered geology this isotropic/elastic model is very useful, being the basis of well-to-seismic ties (Chapter 4) and seismic inversion (Chapter 9).

There are, however, complexities that should not be ignored. These complexities can broadly be characterised as (a) signal-attenuating processes such as absorption and scattering and (b) anisotropic effects, related to horizontal sedimentary layering (vertical polar isotropy) and vertical fracture effects (azimuthal anisotropy) (see Section 5.3.7). One effect of absorption is to attenuate the seismic signal causing changes in wavelet shape with increasing depth and this is usually taken into account. Attenuation is difficult to measure directly from seismic but at least theoretically this information could have a role in identifying the presence of hydrocarbons (e.g. Chapman *et al*., 2006; Chapman, 2008). Whilst there is a good deal of theoretical understanding about anisotropy (Thomsen, 1986; Lynn, 2004), there is currently limited knowledge of how to exploit it for practical exploration purposes. One problem is the availability of data with which to parameterise anisotropic models. Practical seismic analysis in which anisotropic phenomena are exploited has so far been restricted to removing horizontal layering effects on seismic velocities and moveout in seismic processing (i.e. flattening gathers particularly at far offsets) (Chapter 6) and defining vertical fracture presence and orientation (Chapter 7).

2.3.1 The convolutional model, wavelets and polarity

The cornerstone of seismic modelling is the convolutional model, which is the idea that the seismic trace can be modelled as the convolution of a seismic pulse with a reflection coefficient series. In its simplest form the reflection coefficient is related to change in acoustic impedance, where the acoustic impedance (AI) is the product of velocity (V) and bulk density (ρ) (Fig. 2.6):

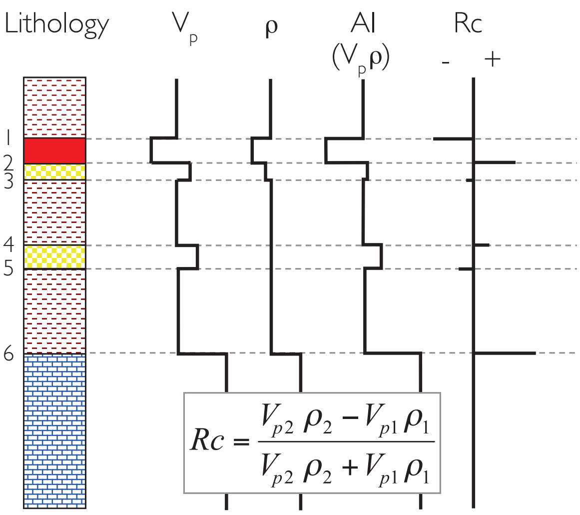

Figure 2.6 The reflection coefficient as defined by the differentiation of the acoustic impedance log (re-drawn and modified after Anstey, 1982).

$$R = \frac{AI_2 - AI_1}{AI_2 + AI_1}, \tag{2.1}$$

where AI_1 is the acoustic impedance on the incident ray side of the boundary and AI_2 is the acoustic impedance on the side of the transmitted ray. Reflectivity can be either positive or negative. In the model of Fig. 2.6 the top of the limestone (interface 6) is characterised by a *positive reflection* whilst the top of the gas sand (interface 1) is characterised by a *negative reflection*.

It should be noted that the equation above is relevant for a ray which is normally incident on a boundary. The change in reflectivity with incident angle (i.e. offset dependent reflectivity) will be discussed in more detail later in this chapter. A useful approximation that derives from the reflectivity equation and which describes the relationship between reflectivity and impedance is:

$$R \approx 0.5 \Delta \ln AI. \tag{2.2}$$

Effectively, the amount of reflected energy determines how much energy can be transmitted through the section. Following the normal incidence model described above, the transmission coefficient (at normal incidence) is defined by

$$T = 2AI_1/(AI_2 + AI_1). \tag{2.3}$$

Given the boundary conditions of pressure continuity and conservation of energy it can be shown that the amount of reflected energy is proportional to R^2, whereas the transmitted energy is proportional to $(AI_2/AI_1)\ T^2$. Thus, less energy is transmitted through boundaries with high AI contrasts (e.g. such as a hard sea floor or the top of a basalt layer). In extreme cases, lateral variations in AI contrasts can result in uneven amplitude scaling deeper in the section.

Transmission effects are important in understanding the general nature of recorded seismic energy. O'Doherty and Anstey (1971) noted that seismic amplitudes at depth appear to be higher than can be accounted for with a simple (normal incidence) model of reflection and transmission at individual boundaries. It was concluded that seismic reflection energy is being reinforced by reflections from thin layers for which the top and base reflections have opposite sign (Anstey and O'Doherty, 2002). The cumulative effects of many cyclical layers can be significant and this may provide an explanation for the observation that reflections tend to parallel chronostratigraphic boundaries.

To generate a synthetic seismogram requires knowledge of the shape of the seismic pulse as well as a calculated reflection coefficient series. A recorded seismic pulse typically has three dominant loops, the relative amplitudes of which can vary according to the nature of the source, the geology and the processes applied to the data (Chapter 3). Assuming that no attempt has been made to shape the wavelet or change its timing, a time series representation of the recorded wavelet will start at time zero (i.e. the wavelet is causal). Figure 2.7 shows the reflection coefficient series from Fig. 2.6 convolved with a recorded seismic pulse, illustrating how the synthetic trace is the addition of the individual reflections.

With regard to the polarity representation of the wavelet shown in Fig. 2.7, reference is made to the recommendation of a SEG committee on polarity published by Thigpen *et al.* (1975). It states that an upward geophone movement or increase in pressure on a hydrophone should be recorded as a negative number and displayed as a trough ('SEG standard polarity') (Fig. 2.8). This definition is almost universally adhered to in seismic recording. The implication is that a reflection from a positive reflection coefficient (a *positive or 'hard' reflection*), will start with a trough. Note that a positive reflection is the interpreter's reference for describing polarity. Figure 2.7 conforms, therefore, to the SEG standard polarity convention with positive reflections such as the top of the limestone

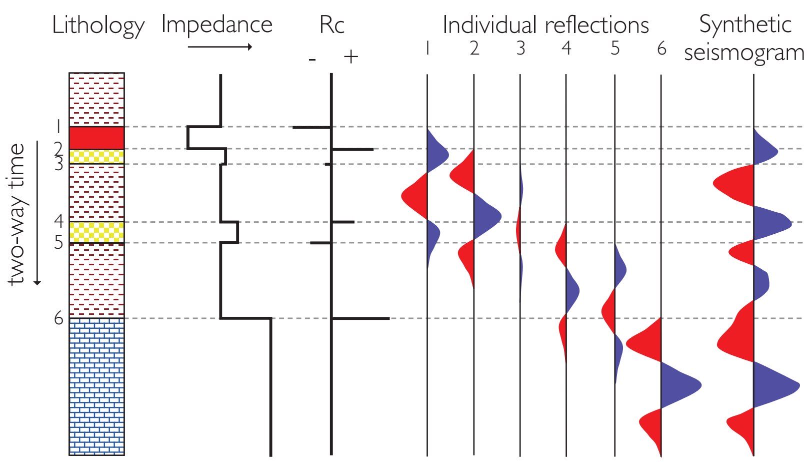

Figure 2.7 Synthetic seismogram using a causal (i.e. recorded) wavelet with SEG standard polarity (re-drawn and modified after Anstey, 1982).

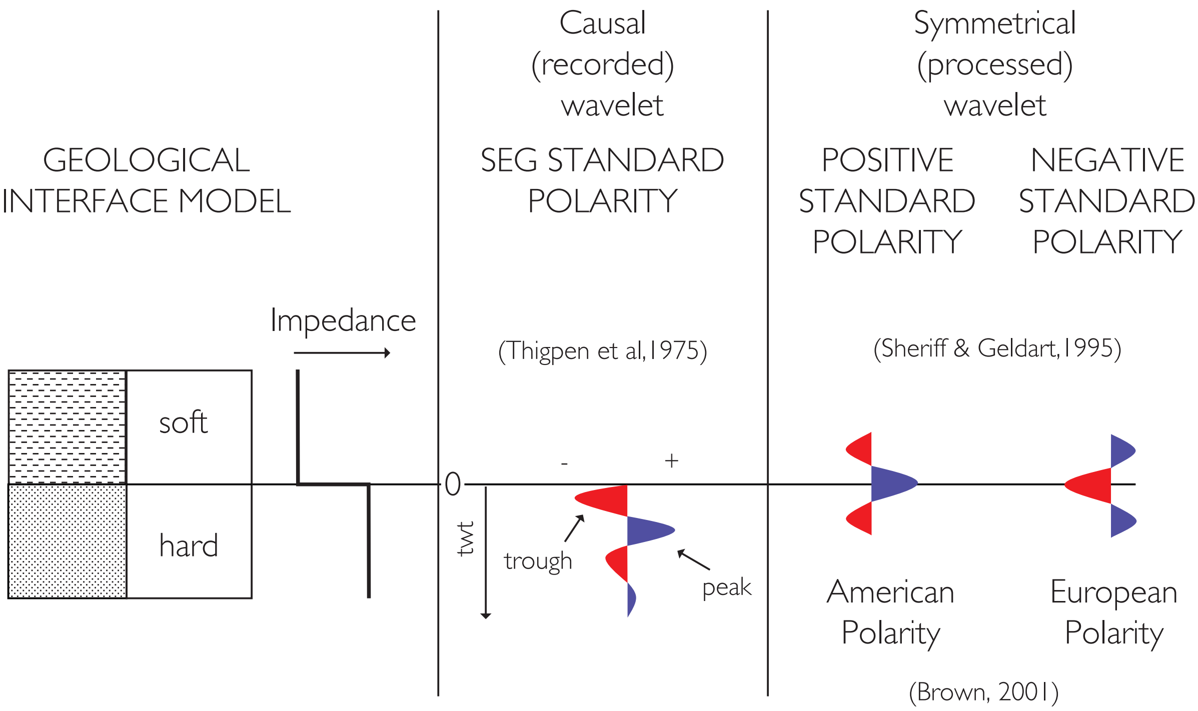

Figure 2.8 Seismic polarity conventions.

starting with a trough and negative reflections (such as the top of the gas sand) starting with a peak.

One problem with causal wavelets (amongst others) is that there is a time lag between the position of the boundary and the energy associated with a reflection from the boundary, making it difficult to correlate the geology with the seismic. Thus, there is a requirement for processing the seismic wavelet to a symmetrical form which concentrates and correctly aligns the energy with the position of geological boundaries. Figure 2.9 shows the same synthetic but now with a symmetrical wavelet. It is now much clearer which loops in the seismic the interpreter needs to pick for the various geological boundaries.

The polarity conventions in common usage that apply to symmetrical wavelets have been defined by Sheriff and Geldart (1995), again with reference to a positive reflection. If a positive reflection is represented

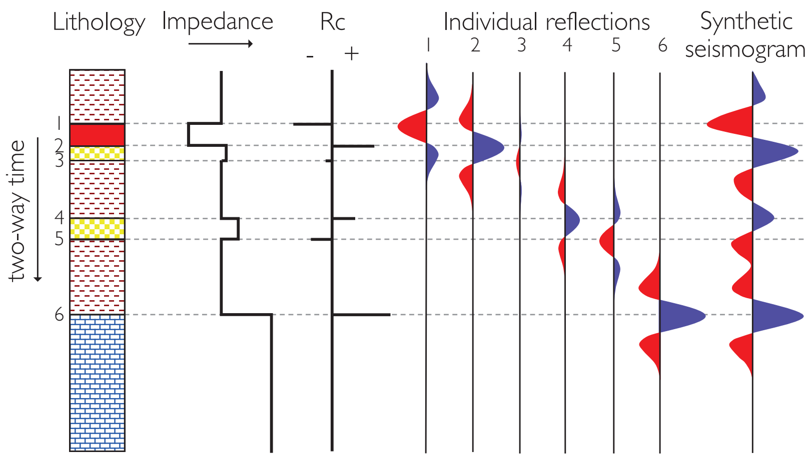

Figure 2.9 Synthetic seismogram using a symmetrical wavelet with positive standard polarity (re-drawn and modified after Anstey, 1982). This illustrates the importance of zero phase to the interpreter; the main layer boundaries are more easily identified with the processed symmetrical wavelet compared to the recorded asymmetrical wavelet shown in Fig 2.7.

as a peak (i.e. positive number) then this is referred to as 'positive standard polarity', whereas if it is represented as a trough (i.e. negative number) it is referred to as 'negative standard polarity' (Fig. 2.8). Historically the usage of these two conventions has been broadly geographical and they have been referred to as 'American' and 'European' (Brown, 2001, 2004). In addition, these conventions are sometimes informally referred to respectively as 'SEG normal' and 'SEG reverse' polarity. It is evident, however, that the use of 'normal' and 'reverse' terms can easily lead to confusion and it is recommended that their use should be avoided.

From the point of view of seismic amplitude and AVO studies it is recommended that positive standard polarity be used as it lessens the potential confusion in the representation of amplitude data. Following this convention will mean that AVO plots will be constructed with positive numbers representing positive reflections and integration type processes (such as coloured inversion (Chapter 5)) will produce the correct sense of change in bandlimited impedance traces (i.e. negative to positive for a boundary with a positive reflection) (see Section 5.5.3).

With respect to colour, it is common for seismic troughs to be coloured red and peaks to be coloured blue or black. There are some notable exceptions, however. For example, in South Africa the tendency has been to adhere to positive standard polarity but colouring the troughs blue and the peaks red. With modern software the interpreter is not restricted to blue and red and can choose from a whole range of colour options. For more discussion of the role of colour in seismic interpretation the reader is referred to Brown (2004) and Froner *et al.* (2013).

For the interpreter who inherits a seismic project it is evident that polarity and colour coding issues (as well as uncertainties concerning the processing of the data) introduce significant potential for misunderstanding and error. It is critical that the interpreter develops a good idea of the shape of the seismic wavelet prior to detailed horizon picking (see Chapters 3 and 4).

2.3.2 Isotropic and elastic rock properties

2.3.2.1 P and S velocities and bulk density

Seismic models for exploration purposes are constructed using velocities and densities, principally from wireline log data (Chapter 8). As will be shown in the following section the calculation of offset reflectivity requires two types of velocity (compressional (P) and shear (S)) as well as the bulk density (ρ). Figure 2.10 illustrates the P and S waves of interest in 3D exploration. The P wave is characterised by particle motion in the direction of wave propagation. The S wave travels in the same direction as the P wave (and at approximately half the speed of the compressional wave) but the particle motion is perpendicular to the direction of wave propagation. Strictly speaking this shear wave is the vertically polarised shear wave. Whilst it is only compressional waves that are recorded in marine seismic, the reason that shear information is important to the interpreter is because changes in amplitude with angle are related to the contrast in the velocities of P waves and the vertically polarised S wave. By contrast, in land exploration a comparison of vertically and horizontally polarised shear waves (recorded with three component sensors)