4 Common techniques for quantitative seismic interpretation

4.1 Introduction

Conventional seismic interpretation implies picking and tracking laterally consistent seismic reflectors for the purpose of mapping geologic structures, stratigraphy and reservoir architecture. The ultimate goal is to detect hydrocarbon accumulations, delineate their extent, and calculate their volumes. Conventional seismic interpretation is an art that requires skill and thorough experience in geology and geophysics.

Traditionally, seismic interpretation has been essentially qualitative. The geometrical expression of seismic reflectors is thoroughly mapped in space and traveltime, but little emphasis is put on the physical understanding of seismic amplitude variations. In the last few decades, however, seismic interpreters have put increasing emphasis on more quantitative techniques for seismic interpretation, as these can validate hydrocarbon anomalies and give additional information during prospect evaluation and reservoir characterization. The most important of these techniques include post-stack amplitude analysis (bright-spot and dim-spot analysis), offset-dependent amplitude analysis (AVO analysis), acoustic and elastic impedance inversion, and forward seismic modeling.

These techniques, if used properly, open up new doors for the seismic interpreter. The seismic amplitudes, representing primarily contrasts in elastic properties between individual layers, contain information about lithology, porosity, pore-fluid type and saturation, as well as pore pressure – information that cannot be gained from conventional seismic interpretation.

4.2 Qualitative seismic amplitude interpretation

Until a few decades ago, it would be common for seismic interpreters to roll out their several-meters-long paper sections with seismic data down the hallway, go down

There are no facts, only interpretations. Friedrich Nietzsche
on their knees, and use their colored pencils to interpret the horizons of interest in order to map geologic bodies. Little attention was paid to amplitude variations and their interpretations. In the early 1970s the so-called “bright-spot” technique proved successful in areas of the Gulf of Mexico, where bright amplitudes would coincide with gas-filled sands. However, experience would show that this technique did not always work. Some of the bright spots that were interpreted as gas sands, and subsequently drilled, were found to be volcanic intrusions or other lithologies with high impedance contrast compared with embedding shales. These failures were also related to lack of wavelet phase analysis, as hard volcanic intrusions would cause opposite polarity to low-impedance gas sands. Moreover, experience showed that gas-filled sands sometimes could cause “dim spots,” not “bright spots,” if the sands had high impedance compared with surrounding shales.

With the introduction of 3D seismic data, the utilization of amplitudes in seismic interpretation became much more important. Brown (see Brown et al., 1981) was one of the pioneers in 3D seismic interpretation of lithofacies from amplitudes. The generation of time slices and horizon slices revealed 3D geologic patterns that had been impossible to discover from geometric interpretation of the wiggle traces in 2D stack sections. Today, the further advance in seismic technology has provided us with 3D visualization tools where the interpreter can step into a virtual-reality world of seismic wiggles and amplitudes, and trace these spatially (3D) and temporally (4D) in a way that one could only dream of a few decades ago. Certainly, the leap from the rolled-out paper sections down the hallways to the virtual-reality images in visualization “caves” is a giant leap with great business implications for the oil industry. In this section we review the qualitative aspects of seismic amplitude interpretation, before we dig into the more quantitative and rock-physics-based techniques such as AVO analysis, impedance inversion, and seismic modeling, in following sections.

4.2.1 Wavelet phase and polarity

The very first issue to resolve when interpreting seismic amplitudes is what kind of wavelet we have. Essential questions to ask are the following. What is the defined polarity in our case? Are we dealing with a zero-phase or a minimum-phase wavelet? Is there a phase shift in the data? These are not straightforward questions to answer, because the phase of the wavelet can change both laterally and vertically. However, there are a few pitfalls to be avoided.

First, we want to make sure what the defined standard is when processing the data. There exist two standards. The American standard defines a black peak as a “hard” or “positive” event, and a white trough as a “soft” or a “negative” event. On a near-offset stack section a “hard” event will correspond to an increase in acoustic impedance with depth, whereas a “soft” event will correspond to a decrease in acoustic impedance with depth. According to the European standard, a black peak is a “soft” event, whereas a
white trough is a “hard” event. One way to check the polarity of marine data is to look at the sea-floor reflector. This reflector should be a strong positive reflector representing the boundary between water and sediment.

**Data polarity**

- **American polarity**: An increase in impedance gives positive amplitude, normally displayed as black peak (wiggle trace) or red intensity (color display).
- **European (or Australian) polarity**: An increase in impedance gives negative amplitude, normally displayed as white trough (wiggle trace) or blue intensity (color display).

(Adapted from Brown, 2001a, 2001b)

For optimal quantitative seismic interpretations, we should ensure that our data are zero-phase. Then, the seismic pick should be on the crest of the waveform corresponding with the peak amplitudes that we desire for quantitative use (Brown, 1998). With today’s advanced seismic interpretation tools involving the use of interactive workstations, there exist various techniques for horizon picking that allow efficient interpretation of large amounts of seismic data. These techniques include manual picking, interpolation, autotracking, voxel tracking, and surface slicing (see Dorn (1998) for detailed descriptions).

For extraction of seismic horizon slices, autopicked or voxel-tracked horizons are very common. The obvious advantage of autotracking is the speed and efficiency. Furthermore, autopicking ensures that the peak amplitude is picked along a horizon. However, one pitfall is the assumption that seismic horizons are locally continuous and consistent. A lateral change in polarity within an event will not be recognized during autotracking. Also, in areas of poor signal-to-noise ratio or where a single event splits into a doublet, the autopicking may fail to track the correct horizon. Not only will important reservoir parameters be neglected, but the geometries and volumes may also be significantly off if we do not regard lateral phase shifts. It is important that the interpreter realizes this and reviews the seismic picks for quality control.

### 4.2.2 Sand/shale cross-overs with depth

Simple rock physics modeling can assist the initial phase of qualitative seismic interpretation, when we are uncertain about what polarity to expect for different lithology boundaries. In a siliciclastic environment, most seismic reflectors will be associated with sand–shale boundaries. Hence, the polarity will be related to the contrast in impedance between sand and shale. This contrast will vary with depth (Chapter 2). Usually, relatively soft sands are found at relatively shallow depths where the sands are unconsolidated. At greater depths, the sands become consolidated and cemented, whereas the
4.2 Qualitative seismic amplitude interpretation

Sand versus shale impedance depth trends and seismic polarity (schematic)

Figure 4.1 Schematic depth trends of sand and shale impedances. The depth trends can vary from basin to basin, and there can be more than one cross-over. Local depth trends should be established for different basins.

Shales are mainly affected by mechanical compaction. Hence, cemented sandstones are normally found to be relatively hard events on the seismic. There will be a corresponding cross-over in acoustic impedance of sands and shales as we go from shallow and soft sands to the deep and hard sandstones (see Figure 4.1). However, the depth trends can be much more complex than shown in Figure 4.1 (Chapter 2, see Figures 2.34 and 2.35). Shallow sands can be relatively hard compared with surrounding shales, whereas deep cemented sandstones can be relatively soft compared with surrounding shales. There is no rule of thumb for what polarity to expect for sands and shales. However, using rock physics modeling constrained by local geologic knowledge, one can improve the understanding of expected polarity of seismic reflectors.

"Hard" versus "soft" events

During seismic interpretation of a prospect or a proven reservoir sand, the following question should be one of the first to be asked: what type of event do we expect, a "hard" or a "soft"? In other words, should we pick a positive peak, or a negative trough? If we have good well control, this issue can be solved by generating synthetic seismograms and correlating these with real seismic data. If we have no well control, we may have to guess. However, a reasonable guess can be made based on rock physics modeling. Below we have listed some "rules of thumb" on what type of reflector we expect for different geologic scenarios.
Typical "hard" events

- Very shallow sands at normal pressure embedded in pelagic shales
- Cemented sandstone with brine saturation
- Carbonate rocks embedded in siliciclastics
- Mixed lithologies (heterolithics) like shaly sands, marls, volcanic ash deposits

Typical "soft" events

- Pelagic shale
- Shallow, unconsolidated sands (any pore fluid) embedded in normally compacted shales
- Hydrocarbon accumulations in clean, unconsolidated or poorly consolidated sands
- Overpressured zones

Some pitfalls in conventional interpretation

- Make sure you know the polarity of the data. Remember there are two different standards, the US standard and the European standard, which are opposite.
- A hard event can change to a soft laterally (i.e., lateral phase shift) if there are lithologic, petrographic or pore-fluid changes. Seismic autotracking will not detect these.
- A dim seismic reflector or interval may be significant, especially in the zone of sand/shale impedance cross-over. AVO analysis should be undertaken to reveal potential hydrocarbon accumulations.

4.2.3 Frequency and scale effects

Seismic resolution

Vertical seismic resolution is defined as the minimum separation between two interfaces such that we can identify two interfaces rather than one (Sherif and Geldhart, 1995). A stratigraphic layer can be resolved in seismic data if the layer thickness is larger than a quarter of a wavelength. The wavelength is given by:

\[ \lambda = \frac{V}{f} \]  \hspace{1cm} (4.1)

where \( V \) is the interval velocity of the layer, and \( f \) is the frequency of the seismic wave. If the wavelet has a peak frequency of 30 Hz, and the layer velocity is 3000 m/s, then the dominant wavelength is 100 m. In this case, a layer of 25 m can be resolved. Below this thickness, we can still gain important information via quantitative analysis of the interference amplitude. A bed only \( \lambda/30 \) in thickness may be detectable, although its thickness cannot be determined from the wave shape (Sherif and Geldhart, 1995).
4.2 Qualitative seismic amplitude interpretation

The horizontal resolution of unmigrated seismic data can be defined by the Fresnel zone. Approximately, the Fresnel zone is defined by a circle of radius, $R$, around a reflection point:

$$R \approx \sqrt{\lambda z/2}$$

where $z$ is the reflector depth. Roughly, the Fresnel zone is the zone from which all reflected contributions have a phase difference of less than $\pi$ radians. For a depth of 3 km and velocity of 3 km/s, the Fresnel zone radius will be 300–470 m for frequencies ranging from 50 to 20 Hz. When the size of the reflector is somewhat smaller than the Fresnel zone, the response is essentially that of a diffraction point. Using pre-stack migration we can collapse the diffractions to be smaller than the Fresnel zone, thus increasing the lateral seismic resolution (Sheriff and Geldhart, 1995). Depending on the migration aperture, the lateral resolution after migration is of the order of a wavelength. However, the migration only collapses the Fresnel zone in the direction of the migration, so if it is only performed along inlines of a 3D survey, the lateral resolution will still be limited by the Fresnel zone in the cross-line direction. The lateral resolution is also restricted by the lateral sampling which is governed by the spacing between individual CDP gathers, usually 12.5 or 18 meters in 3D seismic data. For typical surface seismic wavelengths (~50–100 m), lateral sampling is not the limiting factor.

**Interference and tuning effects**

A thin-layered reservoir can cause what is called event tuning, which is interference between the seismic pulse representing the top of the reservoir and the seismic pulse representing the base of the reservoir. This happens if the layer thickness is less than a quarter of a wavelength (Widess, 1973). Figure 4.2 shows the effective seismic amplitude as a function of layer thickness for a given wavelength, where a given layer has higher impedance than the surrounding sediments. We observe that the amplitude

![Figure 4.2 Seismic amplitude as a function of layer thickness for a given wavelength.](image-url)
increases and becomes larger than the real reflectivity when the layer thickness is between a half and a quarter of a wavelength. This is when we have constructive interference between the top and the base of the layer. The maximum constructive interference occurs when the bed thickness is equal to \( \lambda/4 \), and this is often referred to as the tuning thickness. Furthermore, we observe that the amplitude decreases and approaches zero for layer thicknesses between one-quarter of a wavelength and zero thickness. We refer to this as destructive interference between the top and the base. Trough-to-peak time measurements give approximately the correct gross thicknesses for thicknesses larger than a quarter of a wavelength, but no information for thicknesses less than a quarter of a wavelength. The thickness of an individual thin-bed unit can be extracted from amplitude measurements if the unit is thinner than about \( \lambda/4 \) (Sheriff and Geldhart, 1995). When the layer thickness equals \( \lambda/8 \), Widess (1973) found that the composite response approximated the derivative of the original signal. He referred to this thickness as the theoretical threshold of resolution. The amplitude–thickness curve is almost linear below \( \lambda/8 \) with decreasing amplitude as the layer gets thinner, but the composite response stays the same.

4.2.4 Amplitude and reflectivity strength

"Bright spots" and “dim spots”

The first use of amplitude information as hydrocarbon indicators was in the early 1970s when it was found that bright-spot amplitude anomalies could be associated with hydrocarbon traps (Hammond, 1974). This discovery increased interest in the physical properties of rocks and how amplitudes changed with different types of rocks and pore fluids (Gardner et al., 1974). In a relatively soft sand, the presence of gas and/or light oil will increase the compressibility of the rock dramatically, the velocity will drop accordingly, and the amplitude will decrease to a negative “bright spot.” However, if the sand is relatively hard (compared with cap-rock), the sand saturated with brine may induce a “bright-spot” anomaly, while a gas-filled sand may be transparent, causing a so-called dim spot, that is, a very weak reflector. It is very important before starting to interpret seismic data to find out what change in amplitude we expect for different pore fluids, and whether hydrocarbons will cause a relative dimming or brightening compared with brine saturation. Brown (1999) states that “the most important seismic property of a reservoir is whether it is bright spot regime or dim spot regime.”

One obvious problem in the identification of dim spots is that they are dim – they are hard to see. This issue can be dealt with by investigating limited-range stack sections. A very weak near-offset reflector may have a corresponding strong far-offset reflector. However, some sands, although they are significant, produce a weak positive near-offset reflection as well as a weak negative far-offset reflection. Only a quantitative analysis of the change in near- to far-offset amplitude, a gradient analysis, will be able
to reveal the sand with any considerable degree of confidence. This is explained in Section 4.3.

**Pitfalls: False “bright spots”**

During seismic exploration of hydrocarbons, “bright spots” are usually the first type of DHI (direct hydrocarbon indicators) one looks for. However, there have been several cases where bright-spot anomalies have been drilled, and turned out not to be hydrocarbons.

Some common “false bright spots” include:

- Volcanic intrusions and volcanic ash layers
- Highly cemented sands, often calcite cement in thin pinch-out zones
- Low-porosity heterolithic sands
- Overpressured sands or shales
- Coal beds
- Top of salt diapirs

Only the last three on the list above will cause the same polarity as a gas sand. The first three will cause so-called “hard-kick” amplitudes. Therefore, if one knows the polarity of the data one should be able to discriminate hydrocarbon-associated bright spots from the “hard-kick” anomalies. AVO analysis should permit discrimination of hydrocarbons from coal, salt or overpressured sands/shales.

A very common seismic amplitude attribute used among seismic interpreters is reflection intensity, which is root-mean-square amplitudes calculated over a given time window. This attribute does not distinguish between negative and positive amplitudes; therefore geologic interpretation of this attribute should be made with great caution.

**“Flat spots”**

Flat spots occur at the reflective boundary between different fluids, either gas–oil, gas–water, or water–oil contacts. These can be easy to detect in areas where the background stratigraphy is tilted, so the flat spot will stick out. However, if the stratigraphy is more or less flat, the fluid-related flat spot can be difficult to discover. Then, quantitative methods like AVO analysis can help to discriminate the fluid-related flat spot from the flat-lying lithostratigraphy.

One should be aware of several pitfalls when using flat spots as hydrocarbon indicators. Flat spots can be related to diagenetic events that are depth-dependent. The boundary between opal-A and opal-CT represents an impedance increase in the same way as for a fluid contact, and dry wells have been drilled on diagenetic flat spots. Clinoforms can appear as flat features even if the larger-scale stratigraphy is tilted. Other “false” flat spots include volcanic sills, paleo-contacts, sheet-flood deposits and flat bases of lobes and channels.
Pitfalls: False “flat spots”

One of the best DHIs to look for is a flat spot, the contact between gas and water, gas and oil, or oil and water. However, there are other causes that can give rise to flat spots:
- Ocean bottom multiples
- Flat stratigraphy. The bases of sand lobes especially tend to be flat.
- Opal-A to opal-CT diagenetic boundary
- Paleo-contacts, either related to diagenesis or residual hydrocarbon saturation
- Volcanic sills

Rigorous flat-spot analysis should include detailed rock physics analysis, and forward seismic modeling, as well as AVO analysis of real data (see Section 4.3.8).

Lithology, porosity and fluid ambiguities

The ultimate goal in seismic exploration is to discover and delineate hydrocarbon reservoirs. Seismic amplitude maps from 3D seismic data are often qualitatively interpreted in terms of lithology and fluids. However, rigorous rock physics modeling and analysis of available well-log data is required to discriminate fluid effects quantitatively from lithology effects (Chapters 1 and 2).

The “bright-spot” analysis method has often been unsuccessful because lithology effects rather than fluid effects set up the bright spot. The consequence is the drilling of dry holes. In order to reveal “pitfall” amplitude anomalies it is essential to investigate the rock physics properties from well-log data. However, in new frontier areas well-log data are sparse or lacking. This requires rock physics modeling constrained by reasonable geologic assumptions and/or knowledge about local compactional and depositional trends.

A common way to extract porosity from seismic data is to do acoustic impedance inversion. Increasing porosity can imply reduced acoustic impedance, and by extracting empirical porosity–impedance trends from well-log data, one can estimate porosity from the inverted impedance. However, this methodology suffers from several ambiguities. Firstly, a clay-rich shale can have very high porosities, even if the permeability is close to zero. Hence, a high-porosity zone identified by this technique may be shale. Moreover, the porosity may be constant while fluid saturation varies, and one simple impedance–porosity model may not be adequate for seismic porosity mapping.

In addition to lithology–fluid ambiguities, lithology–porosity ambiguities, and porosity–fluid ambiguities, we may have lithology–lithology ambiguities and fluid–fluid ambiguities. Sand and shale can have the same acoustic impedance, causing no reflectivity on a near-offset seismic section. This has been reported in several areas of the world (e.g. Zeng et al., 1996; Avseth et al., 2001b). It is often reported that fluvial channels or turbidite channels are dim on seismic amplitude maps, and the
Plate 1.1 Seismic P-P amplitude map over a submarine fan. The amplitudes are sensitive to lithofacies and pore fluids, but the relation varies across the image because of the interplay of sedimentologic and diagenetic influences. Blue indicates low amplitudes, yellow and red high amplitudes.

Plate 1.30 Top left, logs penetrating a sandy turbidite sequence; top right, normal-incidence synthetics with a 50 Hz Ricker wavelet. Bottom: increasing water saturation $S_w$ from 10% to 90% (oil API 35, GOR 200) increases density and $V_p$ (left), giving both amplitude and travelttime changes (right).
interpretation is usually that the channel is shale-filled. However, a clean sand filling in the channel can be transparent as well. A geological assessment of geometries indicating differential compaction above the channel may reveal the presence of sand. More advanced geophysical techniques such as offset-dependent reflectivity analysis may also reveal the sands. During conventional interpretation, one should interpret top reservoir horizons from limited-range stack sections, avoiding the pitfall of missing a dim sand on a near- or full-stack seismic section.

Facies interpretation

Lithology influence on amplitudes can often be recognized by the pattern of amplitudes as observed on horizon slices and by understanding how different lithologies occur within a depositional system. By relating lithologies to depositional systems we often refer to these as lithofacies or facies. The link between amplitude characteristics and depositional patterns makes it easier to distinguish lithofacies variations and fluid changes in amplitude maps.

Traditional seismic facies interpretation has been predominantly qualitative, based on seismic traveltimes. The traditional methodology consisted of purely visual inspection of geometric patterns in the seismic reflections (e.g., Mitchum et al., 1977; Weimer and Link, 1991). Brown et al. (1981), by recognizing buried river channels from amplitude information, were amongst the first to interpret depositional facies from 3D seismic amplitudes. More recent and increasingly quantitative work includes that of Ryseth et al. (1998) who used acoustic impedance inversions to guide the interpretation of sand channels, and Zeng et al. (1996) who used forward modeling to improve the understanding of shallow marine facies from seismic amplitudes. Neri (1997) used neural networks to map facies from seismic pulse shape. Reliable quantitative lithofacies prediction from seismic amplitudes depends on establishing a link between rock physics properties and sedimentary facies. Sections 2.4 and 2.5 demonstrated how such links might be established. The case studies in Chapter 5 show how these links allow us to predict lithofacies from seismic amplitudes.

Stratigraphic interpretation

The subsurface is by nature a layered medium, where different lithologies or facies have been superimposed during geologic deposition. Seismic stratigraphic interpretation seeks to map geologic stratigraphy from geometric expression of seismic reflections in traveltime and space. Stratigraphic boundaries can be defined by different lithologies (facies boundaries) or by time (time boundaries). These often coincide, but not always. Examples where facies boundaries and time boundaries do not coincide are erosional surfaces cutting across lithostratigraphy, or the prograding front of a delta almost perpendicular to the lithologic surfaces within the delta.

There are several pitfalls when interpreting stratigraphy from traveltime information. First, the interpretation is based on layer boundaries or interfaces, that is, the contrasts
between different strata or layers, and not the properties of the layers themselves. Two layers with different lithology can have the same seismic properties; hence, a lithostratigraphic boundary may not be observed. Second, a seismic reflection may occur without a lithology change (e.g., Hardage, 1985). For instance, a hiatus with no deposition within a shale interval can give a strong seismic signature because the shales above and below the hiatus have different characteristics. Similarly, amalgamated sands can yield internal stratigraphy within sandy intervals, reflecting different texture of sands from different depositional events. Third, seismic resolution can be a pitfall in seismic interpretation, especially when interpreting stratigraphic onlaps or downlaps. These are essential characteristics in seismic interpretation, as they can give information about the coastal development related to relative sea level changes (e.g., Vail et al., 1977). However, pseudo-onlaps can occur if the thickness of individual layers reduces beneath the seismic resolution. The layer can still exist, even if the seismic expression yields an onlap.

**Pitfalls**

There are several pitfalls in conventional seismic stratigraphic interpretation that can be avoided if we use complementary quantitative techniques:

- Important lithostratigraphic boundaries between layers with very weak contrasts in seismic properties can easily be missed. However, if different lithologies are transparent in post-stack seismic data, they are normally visible in pre-stack seismic data. AVO analysis is a useful tool to reveal sands with impedances similar to capping shales (see Section 4.3).

- It is commonly believed that seismic events are time boundaries, and not necessarily lithostratigraphic boundaries. For instance, a hiatus within a shale may cause a strong seismic reflection if the shale above the hiatus is less compacted than the one below, even if the lithology is the same. Rock physics diagnostics of well-log data may reveal nonlithologic seismic events (see Chapter 2).

- Because of limited seismic resolution, false seismic onlaps can occur. The layer may still exist beneath resolution. Impedance inversion can improve the resolution, and reveal subtle stratigraphic features not observed in the original seismic data (see Section 4.4).

Quantitative interpretation of amplitudes can add information about stratigraphic patterns, and help us avoid some of the pitfalls mentioned above. First, relating lithology to seismic properties (Chapter 2) can help us understand the nature of reflections, and improve the geologic understanding of the seismic stratigraphy. Gutierrez (2001) showed how stratigraphy-guided rock physics analysis of well-log data improved the sequence stratigraphic interpretation of a fluvial system in Colombia using impedance inversion of 3D seismic data. Conducting impedance inversion of the seismic data will
give us layer properties from interface properties, and an impedance cross-section can reveal stratigraphic features not observed on the original seismic section. Impedance inversion has the potential to guide the stratigraphic interpretation, because it is less oscillatory than the original seismic data, it is more readily correlated to well-log data, and it tends to average out random noise, thereby improving the detectability of laterally weak reflections (Gluck et al., 1997). Moreover, frequency-band-limited impedance inversion can improve on the stratigraphic resolution, and the seismic interpretation can be significantly modified if the inversion results are included in the interpretation procedure. For brief explanations of different types of impedance inversions, see Section 4.4. Forward seismic modeling is also an excellent tool to study the seismic signatures of geologic stratigraphy (see Section 4.5).

Layer thickness and net-to-gross from seismic amplitude

As mentioned in the previous section, we can extract layer thickness from seismic amplitudes. As depicted in Figure 4.2, the relationship is only linear for thin layers in pinch-out zones or in sheet-like deposits, so one should avoid correlating layer thickness to seismic amplitudes in areas where the top and base of sands are resolved as separate reflectors in the seismic data.

Meckel and Nath (1911) found that, for sands embedded in shale, the amplitude would depend on the net sand present, given that the thickness of the entire sequence is less than $\lambda/4$. Brown (1996) extended this principle to include beds thicker than the tuning thickness, assuming that individual sand layers are below tuning and that the entire interval of interbedded sands has a uniform layering. Brown introduced the "composite amplitude" defined as the absolute value summation of the top reflection amplitude and the base reflection amplitude of a reservoir interval. The summation of the absolute values of the top and the base emphasizes the effect of the reservoir and reduces the effect of the embedding material.

Zeng et al. (1996) studied the influence of reservoir thickness on seismic signal and introduced what they referred to as effective reflection strength, applicable to layers thinner than the tuning thickness:

$$ R_e = \frac{Z_{sh} - Z_{sh}}{Z_{sh}} \cdot h $$  (4.3)

where $Z_{sh}$ is the sandstone impedance, $Z_{sh}$ is the average shale impedance and $h$ is the layer thickness. A more common way to extract layer thickness from seismic amplitudes is by linear regression of relative amplitude versus net sand thickness as observed at wells that are available. A recent case study showing the application to seismic reservoir mapping was provided by Hill and Halvatis (2001).

Vernik et al. (2002) demonstrated how to estimate net-to-gross from P- and S-impedances for a turbidite system. From acoustic impedance (AI) versus shear impedance (SI) cross-plots, the net-to-gross can be calculated with the following formulas:
\[
N/G = \frac{\int_{z_{\text{top}}}^{z_{\text{base}}} V_{\text{sand}} \, dZ}{\Delta Z} \tag{4.4}
\]

where \( V_{\text{sand}} \) is the oil-sand fraction given by:

\[
V_{\text{sand}} = \frac{SI - bAI - a_0}{a_1 - a_0} \tag{4.5}
\]

where \( b \) is the average slope of the shale slope \((b_0)\) and oil-sand slope \((b_1)\), whereas \( a_0 \)
and \( a_1 \) are the respective intercepts in the AI-SI cross-plot.

Calculation of reservoir thickness from seismic amplitude should be done only in areas where sands are expected to be thinner than the tuning thickness, that is a quarter of a wavelength, and where well-log data show evidence of good correlation between net sand thickness and relative amplitude.

It can be difficult to discriminate layer thickness changes from lithology and fluid changes. In relatively soft sands, the impact of increasing porosity and hydrocarbon saturation tends to increase the seismic amplitude, and therefore works in the same “direction” to layer thickness. However, in relatively hard sands, increasing porosity and hydrocarbon saturation tend to decrease the relative amplitude and therefore work in the opposite “direction” to layer thickness.

### 4.3 AVO analysis

In 1984, 12 years after the bright-spot technology became a commercial tool for hydrocarbon prediction, Ostrander published a breakthrough paper in *Geophysics* (Ostrander, 1984). He showed that the presence of gas in a sand capped by a shale would cause an amplitude variation with offset in pre-stack seismic data. He also found that this change was related to the reduced Poisson’s ratio caused by the presence of gas. Then, the year after, Shuey (1985) confirmed mathematically via approximations of the Zoeppritz equations that Poisson’s ratio was the elastic constant most directly related to the offset-dependent reflectivity for incident angles up to 30°. AVO technology, a commercial tool for the oil industry, was born.

The AVO technique became very popular in the oil industry, as one could physically explain the seismic amplitudes in terms of rock properties. Now, bright-spot anomalies could be investigated before stack, to see if they also had AVO anomalies (Figure 4.3). The technique proved successful in certain areas of the world, but in many cases it was not successful. The technique suffered from ambiguities caused by lithology effects,
tuning effects, and overburden effects. Even processing and acquisition effects could cause false AVO anomalies. But in many of the failures, it was not the technique itself that failed, but the use of the technique that was incorrect. Lack of shear-wave velocity information and the use of too simple geologic models were common reasons for failure. Processing techniques that affected near-offset traces in CDP gathers in a different way from far-offset traces could also create false AVO anomalies. Nevertheless, in the last decade we have observed a revival of the AVO technique. This is due to the improvement of 3D seismic technology, better pre-processing routines, more frequent shear-wave logging and improved understanding of rock physics properties, larger data capacity, more focus on cross-disciplinary aspects of AVO, and last but not least, more awareness among the users of the potential pitfalls. The technique provides the seismic interpreter with more data, but also new physical dimensions that add information to the conventional interpretation of seismic facies, stratigraphy and geomorphology.

In this section we describe the practical aspects of AVO technology, the potential of this technique as a direct hydrocarbon indicator, and the pitfalls associated with this technique. Without going into the theoretical details of wave theory, we address issues related to acquisition, processing and interpretation of AVO data. For an excellent overview of the history of AVO and the theory behind this technology, we refer the reader to Castagna (1993). We expect the future application of AVO to
expand on today's common AVO cross-plot analysis and hence we include overviews of
important contributions from the literature, include tuning, attenuation and anisotropy
effects on AVO. Finally, we elaborate on probabilistic AVO analysis constrained by rock
physics models. These comprise the methodologies applied in case studies 1, 3 and 4 in
Chapter 5.

4.3.1 The reflection coefficient

Analysis of AVO, or amplitude variation with offset, seeks to extract rock parameters
by analyzing seismic amplitude as a function of offset, or more correctly as a function
of reflection angle. The reflection coefficient for plane elastic waves as a function of
reflection angle at a single interface is described by the complicated Zoeppritz equations
(Zoeppritz, 1919). For analysis of P-wave reflections, a well-known approximation is
given by Aki and Richards (1980), assuming weak layer contrasts:

\[
R(\theta_1) \approx \frac{1}{2} \left(1 - 4 \rho^2 V_s^2 \right) \frac{\Delta \rho}{\rho} + \frac{1}{2 \cos^2 \theta} \frac{\Delta V_p}{V_p} - 4 \rho^2 V_s^2 \frac{\Delta V_s}{V_s} \tag{4.6}
\]

where:

\[
\rho = \frac{\sin \theta_1}{V_{p1}} \quad \theta = (\theta_1 + \theta_2)/2 \approx \theta_1
\]

\[
\Delta \rho = \rho_2 - \rho_1 \quad \rho = (\rho_2 + \rho_1)/2
\]

\[
\Delta V_p = V_{p2} - V_{p1} \quad \Delta V_s = (V_{s2} + V_{s1})/2
\]

\[
\Delta V_s = V_{s2} - V_{s1}
\]

In the formulas above, \( \rho \) is the ray parameter, \( \theta_1 \) is the angle of incidence, and \( \theta_2 \) is
the transmission angle; \( V_{p1} \) and \( V_{p2} \) are the P-wave velocities above and below a given
interface, respectively. Similarly, \( V_{s1} \) and \( V_{s2} \) are the S-wave velocities, while \( \rho_1 \) and
\( \rho_2 \) are densities above and below this interface (Figure 4.4).

The approximation given by Aki and Richards can be further approximated (Shuey,
1985):

\[
R(\theta) \approx R(0) + G \sin^2 \theta + F(\tan^2 \theta - \sin^2 \theta) \tag{4.7}
\]

where

\[
R(0) = \frac{1}{2} \left(\frac{\Delta V_p}{V_p} + \frac{\Delta \rho}{\rho}\right)
\]

\[
G = \frac{1}{2} \frac{\Delta V_p}{V_p} - 2 \frac{V_s^2}{V_p^2} \left(\frac{\Delta \rho}{\rho} + 2 \frac{\Delta V_s}{V_s}\right)
\]

\[
= R(0) - \frac{\Delta \rho}{\rho} \left(\frac{1}{2} + \frac{2 V_s^2}{V_p^2}\right) - 4 V_s^2 \frac{\Delta V_s}{V_s}
\]
4.3 AVO analysis

Medium 1
\( (V_p, V_s, \rho) \)

Medium 2
\( (V_p', V_s', \rho') \)

Figure 4.4 Reflections and transmissions at a single interface between two elastic half-space media for an incident plane P-wave, \( PP(i) \). There will be both a reflected P-wave, \( PP(r) \), and a transmitted P-wave, \( PP(t) \). Note that there are wave mode conversions at the reflection point occurring at nonzero incidence angles. In addition to the P-waves, a reflected S-wave, \( PS(r) \), and a transmitted S-wave, \( PS(t) \), will be produced.

\[ R(0) + G \sin^2 \theta \]  

(4.8)

The zero-offset reflectivity, \( R(0) \), is controlled by the contrast in acoustic impedance across an interface. The gradient, \( G \), is more complex in terms of rock properties, but from the expression given above we see that not only the contrasts in \( V_p \) and density affect the gradient, but also \( V_s \). The importance of the \( V_p/V_s \) ratio (or equivalently the Poisson’s ratio) on the offset-dependent reflectivity was first indicated by Koefoed (1955). Ostrander (1984) showed that a gas-filled formation would have a very low Poisson’s ratio compared with the Poisson’s ratios in the surrounding nongaseous formations. This would cause a significant increase in positive amplitude versus angle at the bottom of the gas layer, and a significant increase in negative amplitude versus angle at the top of the gas layer.

4.3.2 The effect of anisotropy

Velocity anisotropy ought to be taken into account when analyzing the amplitude variation with offset (AVO) response of gas sands encased in shales. Although it is generally
thought that the anisotropy is weak (10–20%) in most geological settings (Thomsen, 1986), some effects of anisotropy on AVO have been shown to be dramatic using shale/sand models (Wright, 1987). In some cases, the sign of the AVO slope or rate of change of amplitude with offset can be reversed because of anisotropy in the overlying shales (Kim et al., 1993; Blangy, 1994).

The elastic stiffness tensor C in transversely isotropic (TI) media can be expressed in compact form as follows:

\[
C = \begin{pmatrix}
    C_{11} & (C_{11} - 2C_{66}) & C_{13} & 0 & 0 & 0 \\
    (C_{11} - 2C_{66}) & C_{11} & C_{13} & 0 & 0 & 0 \\
    C_{13} & C_{13} & C_{33} & 0 & 0 & 0 \\
    0 & 0 & 0 & C_{44} & 0 & 0 \\
    0 & 0 & 0 & 0 & C_{44} & 0 \\
    0 & 0 & 0 & 0 & 0 & C_{66}
\end{pmatrix}
\]

where \( C_{66} = \frac{1}{2}(C_{11} - C_{12}) \) (4.9)

and where the 3-axis (z-axis) lies along the axis of symmetry.

The above 6 \times 6 matrix is symmetric, and has five independent components, \( C_{11}, C_{13}, C_{33}, C_{44}, \) and \( C_{66} \). For weak anisotropy, Thomsen (1986) expressed three anisotropic parameters, \( \varepsilon, \gamma \) and \( \delta \), as a function of the five elastic components, where

\[
\varepsilon = \frac{C_{11} - C_{33}}{2C_{33}} \quad (4.10)
\]

\[
\gamma = \frac{C_{66} - C_{44}}{2C_{44}} \quad (4.11)
\]

\[
\delta = \frac{(C_{13} + C_{44})^2 - (C_{33} - C_{44})^2}{2C_{33}(C_{33} - C_{44})} \quad (4.12)
\]

The constant \( \varepsilon \) can be seen to describe the fractional difference of the P-wave velocities in the vertical and horizontal directions:

\[
\varepsilon = \frac{V_p(90^\circ) - V_p(0^\circ)}{V_p(0^\circ)} \quad (4.13)
\]

and therefore best describes what is usually referred to as “P-wave anisotropy.”

In the same manner, the constant \( \gamma \) can be seen to describe the fractional difference of SH-wave velocities between vertical and horizontal directions, which is equivalent to the difference between the vertical and horizontal polarizations of the horizontally propagating S-waves:
\[ \gamma = \frac{V_{SH}(90^\circ) - V_{SV}(90^\circ)}{V_{SV}(90^\circ)} = \frac{V_{SH}(90^\circ) - V_{SH}(0^\circ)}{V_{SH}(0^\circ)} \] (4.14)

The physical meaning of \( \delta \) is not as clear as \( \varepsilon \) and \( \gamma \), but \( \delta \) is the most important parameter for normal moveout velocity and reflection amplitude.

Under the plane wave assumption, Daley and Hron (1977) derived theoretical formulas for reflection and transmission coefficients in TI media. The P-P reflectivity in the equation can be decomposed into isotropic and anisotropic terms as follows:

\[ R_{pp}(\theta) = R_{ip}(\theta) + R_{app}(\theta) \] (4.15)

Assuming weak anisotropy and small offsets, Banik (1987) showed that the anisotropic term can be simply expressed as follows:

\[ R_{app}(\theta) \approx \frac{\sin^2 \theta}{2} \Delta \delta \] (4.16)

Blangy (1994) showed the effect of a transversely isotropic shale overlying an isotropic gas sand on offset-dependent reflectivity, for the three different types of gas sands. He found that hard gas sands overlain by a soft TI shale exhibited a larger decrease in positive amplitude with offset than if the shale had been isotropic. Similarly, soft gas sands overlain by a relatively hard TI shale exhibited a larger increase in negative amplitude with offset than if the shale had been isotropic. Furthermore, it is possible for a soft isotropic water sand to exhibit an “unexpectedly” large AVO effect if the overlying shale is sufficiently anisotropic.

### 4.3.3 The effect of tuning on AVO

As mentioned in the previous section, seismic interference or event tuning can occur as closely spaced reflectors interfere with each other. The relative change in travelt ime between the reflectors decreases with increased travelt ime and offset. The travelt ime hyperbolas of the closely spaced reflectors will therefore become even closer at larger offsets. In fact, the amplitudes may interfere at large offsets even if they do not at small offsets. The effect of tuning on AVO has been demonstrated by Juhlin and Young (1993), Lin and Phair (1993), Bakke and Ursin (1998), and Dong (1998), among others.

Juhlin and Young (1993) showed that thin layers embedded in a homogeneous rock can produce a significantly different AVO response from that of a simple interface of the same lithology. They showed that, for weak contrasts in elastic properties across the layer boundaries, the AVO response of a thin bed may be approximated by modeling it as an interference phenomenon between plane P-waves from a thin layer. In this case thin-bed tuning affects the AVO response of a high-velocity layer embedded in a homogeneous rock more than it affects the response of a low-velocity layer.
Lin and Phair (1993) suggested the following expression for the amplitude variation with angle (AVA) response of a thin layer:

$$ R_t(\theta) = \omega_0 \Delta T(0) \cos \theta \cdot R'(\theta) \tag{4.17} $$

where $\omega_0$ is the dominant frequency of the wavelet, $\Delta T(0)$ is the two-way traveltime at normal incidence from the top to the base of the thin layer, and $R'(\theta)$ is the reflection coefficient from the top interface.

Bakke and Ursin (1998) extended the work by Lin and Phair by introducing tuning correction factors for a general seismic wavelet as a function of offset. If the seismic response from the top of a thick layer is:

$$ d(t, y) = R(y)p(t) \tag{4.18} $$

where $R(y)$ is the primary reflection as a function of offset $y$, and $p(t)$ is the seismic pulse as a function of time $t$, then the response from a thin layer is

$$ d(t, y) \approx R(y) \Delta T(0) C(y)p'(t) \tag{4.19} $$

where $p'(t)$ is the time derivative of the pulse, $\Delta T(0)$ is the traveltime thickness of the thin layer at zero offset, and $C(y)$ is the offset-dependent AVO tuning factor given by

$$ C(y) = \frac{T(0)}{T(y)} \left[ 1 + \frac{V_{RMS}^2 - V^2}{2T(0)^2V_{RMS}^4} \right] \tag{4.20} $$

where $T(0)$ and $T(y)$ are the traveltimes at zero offset and at a given nonzero offset, respectively. The root-mean-square velocity $V_{RMS}$ is defined along a ray path:

$$ V_{RMS}^2 = \frac{\int_0^T V^2(t) \, dt}{\int_0^T dt} \tag{4.21} $$

For small velocity contrasts ($V_{RMS} \approx V$), the last term in equation (4.20) can be ignored, and the AVO tuning factor can be approximated as

$$ C(y) \approx \frac{T(0)}{T(y)} \tag{4.22} $$

For large contrast in elastic properties, one ought to include contributions from P-wave multiples and converted shear waves. The locally converted shear wave is often neglected in ray-tracing modeling when reproduction of the AVO response of potential hydrocarbon reservoirs is attempted. Primaries-only ray-trace modeling in which the Zoeppritz equations describe the reflection amplitudes is most common. But primaries-only Zoeppritz modeling can be very misleading, because the locally converted shear waves often have a first-order effect on the seismic response (Simmons and Backus, 1994). Interference between the converted waves and the primary reflections from the
Figure 4.5 Converted S-waves and multiples that must be included in AVO modeling when we have thin layers, causing these modes to interfere with the primaries. (1) Primary reflections; (2) single-leg shear waves; (3) double-leg shear wave; and (4) primary reverberations. (After Simmons and Backus, 1994.) $T_{PS} =$ transmitted S-wave converted from P-wave, $R_{SP} =$ reflected P-wave converted from S-wave, etc.

Base of the layers becomes increasingly important as the layer thicknesses decrease. This often produces a seismogram that is different from one produced under the primaries-only Zoeppritz assumption. In this case, one should use full elastic modeling including the converted wave modes and the interbed multiples. Martinez (1993) showed that surface-related multiples and P-to-SV-mode converted waves can interfere with primary pre-stack amplitudes and cause large distortions in the AVO responses. Figure 4.5 shows the ray images of converted S-waves and multiples within a layer.

4.3.4 Structural complexity, overburden and wave propagation effects on AVO

Structural complexity and heterogeneities at the target level as well as in the overburden can have a great impact on the wave propagation. These effects include focusing and defocusing of the wave field, geometric spreading, transmission losses, interbed and surface multiples, P-wave to vertically polarized S-wave mode conversions, and anelastic attenuation. The offset-dependent reflectivity should be corrected for these wave propagation effects, via robust processing techniques (see Section 4.3.6). Alternatively, these effects should be included in the AVO modeling (see Sections 4.3.7 and 4.5). Chibiris (1993) provided a simple but robust methodology to correct for overburden effects as well as certain acquisition effects (see Section 4.3.5) by normalizing a target horizon amplitude to a reference horizon amplitude. However, in more recent years there have been several more extensive contributions in the literature on amplitude-preserved imaging in complex areas and AVO corrections due to overburden effects, some of which we will summarize below.
AVO in structurally complex areas

The Zoeppritz equations assume a single interface between two semi-infinite layers with infinite lateral extent. In continuously subsiding basins with relatively flat stratigraphy (such as Tertiary sediments in the North Sea), the use of Zoeppritz equations should be valid. However, complex reservoir geology due to thin beds, vertical heterogeneities, faulting and tilting will violate the Zoeppritz assumptions. Resnick et al. (1987) discuss the effects of geologic dip on AVO signatures, whereas MacLeod and Martin (1988) discuss the effects of reflector curvature. Structural complexity can be accounted for by doing pre-stack depth migration (PSDM). However, one should be aware that several PSDM routines obtain reliable structural images without preserving the amplitudes. Grubb et al. (2001) examined the sensitivity both in structure and amplitude related to velocity uncertainties in PSDM migrated images. They performed an amplitude-preserving (weighted Kirchhoff) PSDM followed by AVO inversion. For the AVO signatures they evaluated both the uncertainty in AVO cross-plots and uncertainty of AVO attribute values along given structural horizons.

AVO effects due to scattering attenuation in heterogeneous overburden

Widmaier et al. (1996) showed how to correct a target AVO response for a thinly layered overburden. A thin-bedded overburden will generate velocity anisotropy and transmission losses due to scattering attenuation, and these effects should be taken into account when analyzing a target seismic reflector. They combined the generalized O’Doherty–Anstey formula (Shapiro et al., 1994a) with amplitude-preserving migration/inversion algorithms and AVO analysis to compensate for the influence of thin-bedded layers on traveltimes and amplitudes of seismic data. In particular, they demonstrated how the estimation of zero-offset amplitude and AVO gradient can be improved by correcting for scattering attenuation due to thin-bed effects. Sick et al. (2003) extended Widmaier’s work and provided a method of compensating for the scattering attenuation effects of randomly distributed heterogeneities above a target reflector. The generalized O’Doherty–Anstey formula is an approximation of the angle-dependent, time-harmonic effective transmissivity \( T \) for scalar waves (P-waves in acoustic 1D medium or SH-waves in elastic 1D medium) and is given by

\[
T(f) \propto T_0 e^{-[\alpha(f, \theta) + i\beta(f, \theta)]L}
\]

(4.23)

where \( f \) is the frequency and \( \alpha \) and \( \beta \) are the angle- and frequency-dependent scattering attenuation and phase shift coefficients, respectively. The angle \( \theta \) is the initial angle of an incident plane wave at the top surface of a thinly layered composite stack; \( L \) is the thickness of the thinly layered stack; \( T_0 \) denotes the transmissivity for a homogeneous isotropic reference medium that causes a phase shift. Hence, the equation above represents the relative amplitude and phase distortions caused by the thin layers with regard to the reference medium. Neglecting the quantity \( T_0 \) which describes the transmission
response for a homogeneous isotropic reference medium (that is, a pure phase shift), a phase-reduced transmissivity is defined:

$$\bar{T}(f) \propto e^{-\alpha(f,\theta) + i\beta(f,\theta)t}.$$  \hspace{1cm} (4.24)

For a P-wave in an acoustic 1D medium, the scattering attenuation, $\alpha$, and the phase coefficient, $\beta$, were derived from Shapiro et al. (1994b) by Widmaier et al. (1996):

$$\alpha(f, \theta) = \frac{4\pi^2 a_2 \sigma^2 f^2}{V_0^2 + 16\pi^2 a_2^2 f^2 \cos^2 \theta}$$  \hspace{1cm} (4.25)

and

$$\beta(f, \theta) = \frac{\pi f \sigma^2}{V_0 \cos \theta} \left[ 1 - \frac{8\pi^2 a_2^2 f^2}{V_0^2 + 16\pi^2 a_2^2 f^2 \cos^2 \theta} \right]$$  \hspace{1cm} (4.26)

where the statistical parameters of the reference medium include spatial correlation length $a_2$, standard deviation $\sigma$, and mean velocity $V_0$. The medium is modeled as a 1D random medium with fluctuating P-wave velocities that are characterized by an exponential correlation function. The transmissivity (absolute value) of the P-wave decreases with increasing angle of incidence.

If the uncorrected seismic amplitude (i.e., the analytical P-wave particle displacement) is defined according to ray theory by:

$$U(S, G, t) = R_C \frac{1}{\gamma} W(t - \tau_M)$$  \hspace{1cm} (4.27)

where $U$ is the seismic trace, $S$ denotes the source, $G$ denotes the receiver, $t$ is the varying traveltime along the ray path, $R_C$ is the reflection coefficient at the reflection point $M$, $\gamma$ is the spherical divergence factor, $W$ is the source wavelet, and $\tau_M$ is the traveltime for the ray between source $S$, via reflection point $M$, and back to the receiver $G$.

A reflector beneath a thin-bedded overburden will have the following compensated seismic amplitude:

$$U^T(S, G, t) = \tilde{T}_{\omega}(t) * R_C \frac{1}{\gamma} W(t - \tau_M)$$  \hspace{1cm} (4.28)

where the two-way, time-reduced transmissivity is given by:

$$\tilde{T}_{\omega}(t) = \tilde{T}_{MG}(t) * \tilde{T}_{SM}(t)$$  \hspace{1cm} (4.29)

The superscript $T$ of $U^T(S, G, t)$ indicates that thin-bed effects have been accounted for. Moreover, equation (4.28) indicates that the source wavelet, $W(t)$, is convolved with the transient transmissivity both for the downgoing ($\tilde{T}_{SM}$) and the upgoing raypaths ($\tilde{T}_{MG}$) between source ($S$), reflection point ($M$), and receiver ($G$).
In conclusion, equation (4.28) represents the angle-dependent time shift caused by transverse isotropic velocity behavior of the thinly layered overburden. Furthermore, it describes the decrease of the AVO response resulting from multiple scattering additional to the amplitude decay related to spherical divergence.

Widmaier et al. (1995) presented similar formulations for elastic P-wave AVO, where the elastic correction formula depends not only on variances and covariances of P-wave velocity, but also on S-wave velocity and density, and their correlation and cross-correlation functions.

Ursin and Stovas (2002) further extended on the O’Doherty–Anstey formula and calculated scattering attenuation for a thin-bedded, viscoelastic medium. They found that in the seismic frequency range, the intrinsic attenuation dominates over the scattering attenuation.

**AVO and intrinsic attenuation (absorption)**

Intrinsic attenuation, also referred to as anelastic absorption, is caused by the fact that even homogeneous sedimentary rocks are not perfectly elastic. This effect can complicate the AVO response (e.g., Martinez, 1993). Intrinsic attenuation can be described in terms of a transfer function \( \hat{G}(\omega, t) \) for a plane wave of angular frequency \( \omega \) and propagation time \( t \) (Luh, 1993):

\[
\hat{G}(\omega, t) = \exp(\omega t / 2Q_e + i(\omega t / Q_e) \ln \omega / \omega_0) \tag{4.30}
\]

where \( Q_e \) is the effective quality factor of the overburden along the wave propagation path and \( \omega_0 \) is an angular reference frequency.

Luh demonstrated how to correct for horizontal, vertical and offset-dependent wavelet attenuation. He suggested an approximate, “rule of thumb” equation to calculate the relative change in AVO gradient, \( \delta G \), due to absorption in the overburden:

\[
\delta G \approx \frac{f_1 \tau}{Q_e} \tag{4.31}
\]

where \( f_1 \) is the peak frequency of the wavelet, and \( \tau \) is the zero-offset two-way travel time at the studied reflector.

Carcione et al. (1998) showed that the presence of intrinsic attenuation affects the P-wave reflection coefficient near the critical angle and beyond it. They also found that the combined effect of attenuation and anisotropy affects the reflection coefficients at non-normal incidence, but that the intrinsic attenuation in some cases can actually compensate the anisotropic effects. In most cases, however, anisotropic effects are dominant over attenuation effects. Carcione (1999) furthermore showed that the unconsolidated sediments near the sea bottom in offshore environments can be highly attenuating, and that these waves will for any incidence angle have a vector attenuation perpendicular
to the sea-floor interface. This vector attenuation will affect AVO responses of deeper reflectors.

4.3.5 Acquisition effects on AVO

The most important acquisition effects on AVO measurements include source directivity, and source and receiver coupling (Martinez, 1993). In particular, acquisition footprint is a large problem for 3D AVO (Castagna, 2001). Irregular coverage at the surface will cause uneven illumination of the subsurface. These effects can be corrected for using inverse operations. Different methods for this have been presented in the literature (e.g., Gassaway et al., 1986; Krail and Shin, 1990; Chemingui and Biondi, 2002). Chiburis' (1993) method for normalization of target amplitudes with a reference amplitude provided a fast and simple way of correcting for certain data collection factors including source and receiver characteristics and instrumentation.

4.3.6 Pre-processing of seismic data for AVO analysis

AVO processing should preserve or restore relative trace amplitudes within CMP gathers. This implies two goals: (1) reflections must be correctly positioned in the subsurface; and (2) data quality should be sufficient to ensure that reflection amplitudes contain information about reflection coefficients.

**AVO processing**

Even though the unique goal in AVO processing is to preserve the true relative amplitudes, there is no unique processing sequence. It depends on the complexity of the geology, whether it is land or marine seismic data, and whether the data will be used to extract regression-based AVO attributes or more sophisticated elastic inversion attributes.

Cambois (2001) defines AVO processing as any processing sequence that makes the data compatible with Shuey's equation, if that is the model used for the AVO inversion. Cambois emphasizes that this can be a very complicated task.

Factors that change the amplitudes of seismic traces can be grouped into Earth effects, acquisition-related effects, and noise (Dey-Sarkar and Suatok, 1993). Earth effects include spherical divergence, absorption, transmission losses, interbed multiples, converted phases, tuning, anisotropy, and structure. Acquisition-related effects include source and receiver arrays and receiver sensitivity. Noise can be ambient or source-generated, coherent or random. Processing attempts to compensate for or remove these effects, but can in the process change or distort relative trace amplitudes. This is an important trade-off we need to consider in pre-processing for AVO. We therefore need
to select a basic but robust processing scheme (e.g., Ostrander, 1984; Chiburis, 1984; Fouquet, 1990; Castagna and Backus, 1993; Yilmaz, 2001).

**Common pre-processing steps before AVO analysis**

**Spiking deconvolution and wavelet processing**

In AVO analysis we normally want zero-phase data. However, the original seismic pulse is causal, usually some sort of minimum phase wavelet with noise. Deconvolution is defined as convolving the seismic trace with an inverse filter in order to extract the impulse response from the seismic trace. This procedure will restore high frequencies and therefore improve the vertical resolution and recognition of events. One can make two-sided, non-causal filters, or shaping filters, to produce a zero-phase wavelet (e.g., Leinbach, 1995; Berkhout, 1977).

The wavelet shape can vary vertically (with time), laterally (spatially), and with offset. The vertical variations can be handled with deterministic Q-compensation (see Section 4.3.4). However, AVO analysis is normally carried out within a limited time window where one can assume stationarity. Lateral changes in the wavelet shape can be handled with surface-consistent amplitude balancing (e.g., Cambois and Magesan, 1997). Offset-dependent variations are often more complicated to correct for, and are attributed to both offset-dependent absorption (see Section 4.3.4), tuning effects (see Section 4.3.3), and NMO stretching. NMO stretching acts like a low-pass, mixed-phase, nonstationary filter, and the effects are very difficult to eliminate fully (Cambois, 2001). Dong (1999) examined how AVO detectability of lithology and fluids was affected by tuning and NMO stretching, and suggested a procedure for removing the tuning and stretching effects in order to improve AVO detectability. Cambois recommended picking the reflections of interest prior to NMO corrections, and flattening them for AVO analysis.

**Spherical divergence correction**

Spherical divergence, or geometrical spreading, causes the intensity and energy of spherical waves to decrease inversely as the square of the distance from the source (Newman, 1973). For a comprehensive review on offset-dependent geometrical spreading, see the study by Ursin (1990).

**Surface-consistent amplitude balancing**

Source and receiver effects as well as water depth variation can produce large deviations in amplitude that do not correspond to target reflector properties. Commonly, statistical amplitude balancing is carried out both for time and offset. However, this procedure can have a dramatic effect on the AVO parameters. It easily contributes to intercept leakage and consequently erroneous gradient estimates (Cambois, 2000). Cambois (2001) suggested modeling the expected average amplitude variation with
offset following Shuey’s equation, and then using this behavior as a reference for the statistical amplitude balancing.

**Multiple removal**

One of the most deteriorating effects on pre-stack amplitudes is the presence of multiples. There are several methods of filtering away multiple energy, but not all of these are adequate for AVO pre-processing. The method known as $f-k$ multiple filtering, done in the frequency–wavenumber domain, is very efficient at removing multiples, but the dip in the $f-k$ domain is very similar for near-offset primary energy and near-offset multiple energy. Hence, primary energy can easily be removed from near traces and not from far traces, resulting in an artificial AVO effect. More robust demultiple techniques include linear and parabolic Radon transform multiple removal (Hampson, 1986; Herrmann et al., 2000).

**NMO (normal moveout) correction**

A potential problem during AVO analysis is error in the velocity moveout correction (Spratt, 1987). When extracting AVO attributes, one assumes that primaries have been completely flattened to a constant traveltime. This is rarely the case, as there will always be residual moveout. The reason for residual moveout is almost always associated with erroneous velocity picking, and great efforts should be put into optimizing the estimated velocity field (e.g., Adler, 1999; Le Meur and Magneron, 2000). However, anisotropy and nonhyperbolic moveouts due to complex overburden may also cause misalignments between near and far offsets (an excellent practical example on AVO and nonhyperbolic moveout was published by Ross, 1997). Ursin and Ekren (1994) presented a method for analyzing AVO effects in the offset domain using time windows. This technique reduces moveout errors and creates improved estimates of AVO parameters. One should be aware of AVO anomalies with polarity shifts (class IIP, see definition below) during NMO corrections, as these can easily be misinterpreted as residual moveouts (Ratcliffe and Adler, 2000).

**DMO correction**

DMO (dip moveout) processing generates common-reflection-point gathers. It moves the reflection observed on an offset trace to the location of the coincident source–receiver trace that would have the same reflecting point. Therefore, it involves shifting both time and location. As a result, the reflection moveout no longer depends on dip, reflection-point smear of dipping reflections is eliminated, and events with various dips have the same stacking velocity (Sheriff and Geldhart, 1995). Shang et al. (1993) published a technique on how to extract reliable AVA (amplitude variation with angle) gathers in the presence of dip, using partial pre-stack migration.
Pre-stack migration

Pre-stack migration might be thought to be unnecessary in areas where the sedimentary section is relatively flat, but it is an important component of all AVO processing. Pre-stack migration should be used on data for AVO analysis whenever possible, because it will collapse the diffractions at the target depth to be smaller than the Fresnel zone and therefore increase the lateral resolution (see Section 4.2.3; Berkhout, 1985; Mosher et al., 1996). Normally, pre-stack time migration (PSTM) is preferred to pre-stack depth migration (PSDM), because the former tends to preserve amplitudes better. However, in areas with highly structured geology, PSDM will be the most accurate tool (Cambois, 2001). An amplitude-preserving PSDM routine should then be applied (Bleistein, 1987; Schleicher et al., 1993; Hanitzsch, 1997).

Migration for AVO analysis can be implemented in many different ways. Resnick et al. (1987) and Allen and Peddy (1993) among others have recommended Kirchhoff migration together with AVO analysis. An alternative approach is to apply wave-equation-based migration algorithms. Mosher et al. (1996) derived a wave equation for common-angle time migration and used inverse scattering theory (see also Weglein, 1992) for integration of migration and AVO analysis (i.e., migration-inversion). Mosher et al. (1996) used a finite-difference approach for the pre-stack migrations and illustrated the value of pre-stack migration for improving the stratigraphic resolution, data quality, and location accuracy of AVO targets.

Example of pre-processing scheme for AVO analysis of a 2D seismic line

(Yilmaz, 2001.)

1. Pre-stack signal processing (source signature processing, geometric scaling, spiking deconvolution and spectral whitening).
2. Sort to CMP and do sparse interval velocity analysis.
3. NMO using velocity field from step 2.
5. Sort to common-offset and do DMO correction.
7. Sort data to common-reflection-point (CRP) and do inverse NMO using the velocity field from step 2.
8. Detailed velocity analysis associated with the migrated data.
9. NMO correction using velocity field from step 8.
10. Stack CRP gathers to obtain image of pre-stack migrated data. Remove residual multiples revealed by the stacking.
11. Unmigrate using same velocity field as in step 6.
13. Remigrate using migration velocity field from step 8.
Some pitfalls in AVO interpretation due to processing effects

- Wavelet phase. The phase of a seismic section can be significantly altered during processing. If the phase of a section is not established by the interpreter, then AVO anomalies that would be interpreted as indicative of decreasing impedance, for example, can be produced at interfaces where the impedance increases (e.g., Allen and Peddy, 1993).

- Multiple filtering. Not all demultiple techniques are adequate for AVO preprocessing. Multiple filtering, done in the frequency–wavenumber domain, is very efficient at removing multiples, but the dip in the $f-k$ domain is very similar for near-offset primary energy and near-offset multiple energy. Hence, primary energy can easily be removed from the near-offset traces, resulting in an artificial AVO effect.

- NMO correction. A potential problem during AVO analysis is errors in the velocity moveout correction (Spratt, 1987). When extracting AVO attributes, one assumes that primaries have been completely flattened to a constant traveltime. This is rarely the case, as there will always be residual moveout. Ursin and Ekren (1994) presented a method for analyzing AVO effects in the offset domain using time windows. This technique reduces moveout errors and creates improved estimates of AVO parameters. NMO stretch is another problem in AVO analysis. Because the amount of normal moveout varies with arrival time, frequencies are lowered at large offsets compared with short offsets. Large offsets, where the stretching effect is significant, should be muted before AVO analysis. Swan (1991), Dong (1998) and Dong (1999) examine the effect of NMO stretch on offset-dependent reflectivity.

- AGC amplitude correction. Automatic gain control must be avoided in preprocessing of pre-stack data before doing AVO analysis.

Pre-processing for elastic impedance inversion

Several of the pre-processing steps necessary for AVO analysis are not required when preparing data for elastic impedance inversion (see Section 4.4 for details on the methodology). First of all, the elastic impedance approach allows for wavelet variations with offset (Cambois, 2000). NMO stretch corrections can be skipped, because each limited-range sub-stack (in which the wavelet can be assumed to be stationary) is matched to its associated synthetic seismogram, and this will remove the wavelet variations with angle. It is, however, desirable to obtain similar bandwidth for each inverted sub-stack cube, since these should be comparable. Furthermore, the data used for elastic impedance inversion are calibrated to well logs before stack, which means that average amplitude variations with offset are automatically accounted for. Hence, the complicated procedure of reliable amplitude corrections becomes much less labor-intensive than for
standard AVO analysis. Finally, residual NMO and multiples still must be accounted for (Cambois, 2001). Misalignments do not cause intercept leakage as for standard AVO analysis, but near- and far-angle reflections must still be in phase.

4.3.7 AVO modeling and seismic detectability

AVO analysis is normally carried out in a deterministic way to predict lithology and fluids from seismic data (e.g., Smith and Gidlow, 1987; Rutherford and Williams, 1989; Hilterman, 1990; Castagna and Smith, 1994; Castagna et al., 1998).

Forward modeling of AVO responses is normally the best way to start an AVO analysis, as a feasibility study before pre-processing, inversion and interpretation of real pre-stack data. We show an example in Figure 4.6 where we do AVO modeling of different lithofacies defined in Section 2.5. The figure shows the AVO curves for different half-space models, where a silty shale is taken as the cap-rock with different underlying lithofacies. For each facies, $V_p$, $V_s$, and $\rho$ are extracted from well-log data and used in the modeling. We observe a clean sand/pure shale ambiguity (facies IIb and facies V) at near offsets, whereas clean sands and shales are distinguishable at far offsets. This example depicts how AVO is necessary to discriminate different lithofacies in this case.
Figure 4.7 Schematic AVO curves for cemented sandstone and unconsolidated sands capped by shale, for brine-saturated and oil-saturated cases.

Figure 4.7 shows another example, where we consider two types of clean sands, cemented and unconsolidated, with brine versus hydrocarbon saturation. We see that a cemented sandstone with hydrocarbon saturation can have similar AVO response to a brine-saturated, unconsolidated sand.

The examples in Figures 4.6 and 4.7 indicate how important it is to understand the local geology during AVO analysis. It is necessary to know what type of sand is expected for a given prospect, and how much one expects the sands to change locally owing to textural changes, before interpreting fluid content. It is therefore equally important to conduct realistic lithology substitutions in addition to fluid substitution during AVO modeling studies. The examples in Figures 4.6 and 4.7 also demonstrate the importance of the link between rock physics and geology (Chapter 2) during AVO analysis.

When is AVO analysis the appropriate technique?

It is well known that AVO analysis does not always work. Owing to the many cases where AVO has been applied without success, the technique has received a bad reputation as an unreliable tool. However, part of the AVO analysis is to find out if the technique is appropriate in the first place. It will work only if the rock physics and fluid characteristics of the target reservoir are expected to give a good AVO response. This must be clarified before the AVO analysis of real data. Without a proper feasibility study, one can easily misinterpret AVO signatures in the real data. A good feasibility study could include both simple reflectivity modeling and more advanced forward seismic modeling (see Section 4.5). Both these techniques should be founded on a thorough understanding of local geology and petrophysical properties. Realistic lithology substitution is as important as fluid substitution during this exercise.
Often, one will find that there is a certain depth interval where AVO will work, often referred to as the “AVO window.” Outside this, AVO will not work well. That is why analysis of rock physics depth trends should be an integral part of AVO analysis (see Sections 2.6 and 4.3.16). However, the “AVO window” is also constrained by data quality. With increasing depth, absorption of primary energy reduces the signal-to-noise ratio, higher frequencies are gradually more attenuated than lower frequencies, the geology usually becomes more complex causing more complex wave propagations, and the angle range reduces for a given streamer length. All these factors make AVO less applicable with increasing depth.

4.3.8 Deterministic AVO analysis of CDP gathers

After simple half-space AVO modeling, the next step in AVO analysis should be deterministic AVO analysis of selected CDP (common-depth-point) gathers, preferably at well locations where synthetic gathers can be generated and compared with the real CDP gathers. In this section, we show an example of how the method can be applied to discriminate lithofacies in real seismic data, by analyzing CDP gathers at well locations in a deterministic way. Figure 4.8 shows the real and synthetic CDP gathers at three adjacent well locations in a North Sea field (the well logs are shown in Figure 5.1, case study 1). The figure also includes the picked amplitudes at a top target horizon superimposed on exact Zoeppritz calculated reflectivity curves derived from the well-log data.

In Well 2, the reservoir sands are unconsolidated, represent oil-saturated sands, and are capped by silty shales. According to the saturation curves derived from deep resistivity measurements, the oil saturation in the reservoir varies from 20–80%, with an average of about 60%. The sonic and density logs are found to measure the mud filtrate invaded zone (0–10% oil). Hence, we do fluid substitution to calculate the seismic properties of the reservoir from the Biot–Gassmann theory assuming a uniform saturation model (the process of fluid substitution is described in Chapter 1). Before we do the fluid substitution, we need to know the acoustic properties of the oil and the mud filtrate. These are calculated from Batzle and Wang’s relations (see Chapter 1). For this case, the input parameters for the fluid substitution are as follows.

<table>
<thead>
<tr>
<th>Property</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oil GOR</td>
<td>64 l/l</td>
</tr>
<tr>
<td>Oil relative density</td>
<td>32 API</td>
</tr>
<tr>
<td>Mud-filtrate density</td>
<td>1.09 g/cm³</td>
</tr>
<tr>
<td>Pore pressure at reservoir level</td>
<td>20 MPa</td>
</tr>
<tr>
<td>Temperature at reservoir level</td>
<td>77.2 °C</td>
</tr>
</tbody>
</table>
Figure 4.8 Real CDP gathers (upper), synthetic CDP gathers (middle), and AVO curves for Wells 1–3 (lower).
The corresponding AVO response shows a negative zero-offset reflectivity and a negative AVO gradient. In Well 1, we have a water-saturated cemented sand below a silty shale. The corresponding AVO response in this well shows a strong positive zero-offset reflectivity and a relatively strong negative gradient. Finally, in Well 3 we observe a strong positive zero-offset reflectivity and a moderate negative gradient, corresponding to interbedded sand/shale facies capped by silty shales. Hence, we observe three distinct AVO responses in the three different wells. The changes are related to both lithology and pore-fluid variations within the turbidite system. For more detailed information about this system, see case study 1 in Chapter 5.

Avseth et al. (2000) demonstrated the effect of cementation on the AVO response in real CDP gathers around two wells, one where the reservoir sands are friable, and the other where the reservoir sands are cemented. They found that if the textural effects of the sands were ignored, the corresponding changes in AVO response could be interpreted as pore-fluid changes, just as depicted in the reflectivity modeling example in Figure 4.7.

**Importance of AVO analysis of individual CDP gathers**

Investigations of CDP gathers are important in order to confirm AVO anomalies seen in weighted stack sections (Shuey’s intercept and gradient, Smith and Gidlow’s fluid factor, etc.). The weighted stacks can contain anomalies not related to true offset-dependent amplitude variations.

### 4.3.9 Estimation of AVO parameters

**Estimating intercept and gradient**

The next step in an AVO analysis should be to extract AVO attributes and do multivariate analysis of these. Several different AVO attributes can be extracted, mapped and analyzed. The two most important ones are zero-offset reflectivity \( R(0) \) and AVO gradient \( G \) based on Shuey’s approximation. These seismic parameters can be extracted, via a least-squares seismic inversion, for each sample in a CDP gather over a selected portion of a 3D seismic volume.

For a given NMO-corrected CDP gather, \( R(t, x) \), it is assumed that for each time sample, \( t \), the reflectivity data can be expressed as Shuey’s formula (equation (4.8)):

\[
R(t, x) = R(t, 0) + G(t) \sin^2 \theta(t, x)
\]

where \( \theta(t, x) \) is the incident angle corresponding to the data sample recorded at \((t, x)\).
For a layered Earth, the relationship between offset (x) and angle (θ) is given approximately by:

\[
\sin \theta(t, x) \approx \frac{x}{(t_0^2 + x^2/V_{\text{RMS}}^2)^{1/2}} V_{\text{INT}} V_{\text{RMS}}^2
\]  

where \( V_{\text{INT}} \) is the interval velocity and \( V_{\text{RMS}} \) is the average root-mean-square velocity, as calculated from an input velocity profile (for example obtained from sonic log).

For any given value of zero-offset time, \( t_0 \), we assume that \( R \) is measured at \( N \) offsets \( (x_i, i = 1, N) \). Hence, we can rewrite the defining equation for this time as (Hampson and Russell, 1995):

\[
\begin{bmatrix}
R(x_1) \\
R(x_2) \\
\vdots \\
R(x_N)
\end{bmatrix} =
\begin{bmatrix}
1 & \sin^2 \theta(t, x_1) \\
1 & \sin^2 \theta(t, x_2) \\
\vdots & \vdots \\
1 & \sin^2 \theta(t, x_N)
\end{bmatrix}
\begin{bmatrix}
R(t, 0) \\
G(t)
\end{bmatrix}
\]  

This matrix equation is in the form of \( \mathbf{b} = \mathbf{A} \mathbf{c} \) and represents \( N \) equations in the two unknowns, \( R(t, 0) \) and \( G(t) \). The least-squares solution to this equation is obtained by solving the so-called "normal equation":

\[
\mathbf{c} = (\mathbf{A}^T \mathbf{A})^{-1} (\mathbf{A}^T \mathbf{b})
\]  

This gives us the least-squares solution for \( R(0) \) and \( G \) at time \( t \).

**Inversion for elastic parameters**

Going beyond the estimation of intercept and gradient, one can invert pre-stack seismic amplitudes for elastic parameters, including \( V_p, V_s \) and density. This is commonly referred to as AVO inversion, and can be performed via nonlinear methods (e.g., Dahl and Ursin, 1992; Buland et al., 1996; Gouveia and Scales, 1998) or linearized inversion methods (e.g., Smith and Gidlow, 1987; Loertzer and Berkhout, 1993). Gouveia and Scales (1998) defined a Bayesian nonlinear model and estimated, via a nonlinear conjugate gradient method, the maximum a-posteriori (MAP) distributions of the elastic parameters. However, the nonlinearity of the inversion problem makes their method very computer intensive. Loertzer and Berkhout (1993) performed linearized Bayesian inversion based on single interface theory on a sample-by-sample basis. Buland and Omre (2003) extended the work of Loertzer and Berkhout and developed a linearized Bayesian AVO inversion method where the wavelet is accounted for by convolution. The inversion is performed simultaneously for all times in a given time window, which
makes it possible to obtain temporal correlation between model parameters close in time. Furthermore, they solved the AVO inversion problem via Gaussian priors and obtained an explicit analytical form for the posterior density, providing a computationally fast estimation of the elastic parameters.

**Pitfalls of AVO inversion**

- A linear approximation of the Zoeppritz equations is commonly used in the calculation of $R(0)$ and $G$. The two-term Shuey approximation is known to be accurate for angles of incidence up to approximately 30°. Make sure that the data inverted do not exceed this range, so the approximation is valid.
- The Zoeppritz equations are only valid for single interfaces. Inversion algorithms that are based on these equations will not be valid for thin-bedded geology.
- The linear AVO inversion is sensitive to uncharacteristic amplitudes caused by noise (including multiples) or processing and acquisition effects. A few outlying values present in the pre-stack amplitudes are enough to cause erroneous estimates of $R(0)$ and $G$. Most commercial software packages for estimation of $R(0)$ and $G$ apply robust estimation techniques (e.g., Walden, 1991) to limit the damage of outlying amplitudes.
- Another potential problem during sample-by-sample AVO inversion is errors in the moveout correction (Spratt, 1987). Ursin and Ekren (1994) presented a method for analyzing AVO effects in the offset domain using time windows. This technique reduces moveout errors and creates improved estimates of AVO parameters.

### 4.3.10 AVO cross-plot analysis

A very helpful way to interpret AVO attributes is to make cross-plots of intercept ($R(0)$) versus gradient ($G$). These plots are a very helpful and intuitive way of presenting AVO data, and can give a better understanding of the rock properties than by analyzing the standard AVO curves.

**AVO classes**

Rutherford and Williams (1989) suggested a classification scheme of AVO responses for different types of gas sands (see Figure 4.9). They defined three AVO classes based on where the top of the gas sands will be located in an $R(0)$ versus $G$ cross-plot. The cross-plot is split up into four quadrants. In a cross-plot with $R(0)$ along x-axis and $G$ along y-axis, the 1st quadrant is where $R(0)$ and $G$ are both positive values (upper right quadrant). The 2nd is where $R(0)$ is negative and $G$ is positive (upper left quadrant). The 3rd is where both $R(0)$ and $G$ are negative (lower left quadrant). Finally, the 4th quadrant is where $R(0)$ is positive and $G$ is negative (lower right quadrant). The AVO classes
4.3 AVO analysis

Table 4.1 AVO classes, after Rutherford and Williams (1989), extended by Castagna and Smith (1994), and Ross and Kinman (1995)

<table>
<thead>
<tr>
<th>Class</th>
<th>Relative impedance</th>
<th>Quadrant</th>
<th>R(0)</th>
<th>G</th>
<th>AVO product</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>High-impedance sand</td>
<td>4th</td>
<td>+</td>
<td>-</td>
<td>Negative</td>
</tr>
<tr>
<td>II</td>
<td>No or low contrast</td>
<td>4th</td>
<td>+</td>
<td>-</td>
<td>Negative</td>
</tr>
<tr>
<td>IIp</td>
<td>Low impedance</td>
<td>3rd</td>
<td>-</td>
<td>-</td>
<td>Positive</td>
</tr>
<tr>
<td>III</td>
<td>Low impedance</td>
<td>3rd</td>
<td>-</td>
<td>-</td>
<td>Positive</td>
</tr>
<tr>
<td>IV</td>
<td>Low impedance</td>
<td>2nd</td>
<td>-</td>
<td>+</td>
<td>Negative</td>
</tr>
</tbody>
</table>

Figure 4.9 Rutherford and Williams AVO classes, originally defined for gas sands (classes I, II and III), along with the added classes IV (Castagna and Smith, 1994) and IIp (Ross and Kinman, 1995). Figure is adapted from Castagna et al. (1998).

must not be confused with the quadrant numbers. Class I plots in the 4th quadrant with positive R(0) and negative gradients. These represent hard events with relatively high impedance and low \( V_p/V_s \) ratio compared with the cap-rock. Class II represents sands with weak intercept but strong negative gradient. These can be hard to see on the seismic data, because they often yield dim spots on stacked sections. Class III is the AVO category that is normally associated with bright spots. These plot in the 3rd quadrant in R(0)–G cross-plots, and are associated with soft sands saturated with hydrocarbons (see Plate 4.10).

Ross and Kinman (1995) distinguished between a class IIp and class II anomaly. Class IIp has a weak but positive intercept and a negative gradient, causing a polarity change with offset. This class will disappear on full stack sections. Class II has a weak but negative intercept and negative gradient, hence no polarity change. This class may be observed as a negative amplitude on a full-offset stack.

Castagna and Swan (1997) extended the classification scheme of Rutherford and Williams to include a 4th class, plotting in the 2nd quadrant. These are relatively rare, but occur when soft sands with gas are capped by relatively stiff shales characterized by \( V_p/V_s \) ratios slightly higher than in the sands (i.e., very compacted or silty shales).
Summary of AVO classes

- AVO class I represents relatively hard sands with hydrocarbons. These sands tend to plot along the background trend in intercept–gradient cross-plots. Moreover, very hard sands can have little sensitivity to fluids, so there may not be an associated flat spot. Hence, these sands can be hard to discover from seismic data.
- AVO class II, representing transparent sands with hydrocarbons, often show up as dim spots or weak negative reflectors on the seismic. However, because of relatively large gradients, they should show up as anomalies in an $R(0)$–$G$ cross-plot, and plot off the background trend.
- AVO class III is the “classical” AVO anomaly with negative intercept and negative gradient. This class represents relatively soft sands with high fluid sensitivity, located far away from the background trend. Hence, they should be easy to detect on seismic data.
- AVO class IV are sands with negative intercept but positive gradient. The reflection coefficient becomes less negative with increasing offset, and amplitude decreases versus offset, even though these sands may be bright spots (Castagna and Swan, 1997). Class IV anomalies are relatively rare, but occur when soft sands with gas are capped by relatively stiff cap-rock shales characterized by $V_p/V_s$ ratios slightly higher than in the sands (i.e., very compacted or silty shales).

The AVO classes were originally defined for gas sands. However, today the AVO class system is used for descriptive classification of observed anomalies that are not necessarily gas sands. An AVO class II that is drilled can turn out to be brine sands. It does not mean that the AVO anomaly was not a class II anomaly. We therefore suggest applying the classification only as descriptive terms for observed AVO anomalies, without automatically inferring that we are dealing with gas sands.

AVO trends and the effects of porosity, lithology and compaction

When we plot $R(0)$ and $G$ as cross-plots, we can analyze the trends that occur in terms of changes in rock physics properties, including fluid trends, porosity trends and lithology trends, as these will have different directions in the cross-plot (Figure 4.11). Using rock physics models and then calculating the corresponding intercept and gradients, we can study various “What if” scenarios, and then compare the modeled trends with the inverted data.

Brine-saturated sands interbedded with shales, situated within a limited depth range and at a particular locality, normally follow a well-defined “background trend” in AVO cross-plot (Castagna and Swan, 1997). A common and recommended approach in qualitative AVO cross-plot analysis is to recognize the “background” trend and then look for data points that deviate from this trend.
Castagna et al. (1998) presented an excellent overview and a framework for AVO gradient and intercept interpretation. The top of the sands will normally plot in the 4th quadrant, with positive \( R(0) \) and negative \( G \). The base of the sands will normally plot in the 2nd quadrant, with negative \( R(0) \) and positive \( G \). The top and base of sands, together with shale-shale interfaces, will create a nice trend or ellipse with center in the origin of the \( R(0)-G \) coordinate system. This trend will rotate with contrast in \( V_p/V_s \) ratio between a shaly cap-rock and a sandy reservoir (Castagna et al., 1998; Sams, 1998).

We can extract the relationship between \( V_p/V_s \) ratio and the slope of the background trend (\( a_b \)) by dividing the gradient, \( G \), by the intercept, \( R(0): \)

\[
a_b = \frac{G}{R(0)}
\] (4.36)

Assuming the density contrast between shale and wet sand to be zero, we can study how changing \( V_p/V_s \) ratio affects the background trend. The density contrast between sand and shale at a given depth is normally relatively small compared with the velocity contrasts (Foster et al., 1997). Then the background slope is given by:

\[
a_b = 1 - 8 \left[ \frac{(V_{s1} + V_{s2}) \Delta V_S}{(V_{p1} + V_{p2}) \Delta V_P} \right]
\] (4.37)

where \( V_{p1} \) and \( V_{p2} \) are the P-wave velocities in the cap-rock and in the reservoir, respectively; \( V_{s1} \) and \( V_{s2} \) are the corresponding S-wave velocities, whereas \( \Delta V_P \) and \( \Delta V_S \) are the velocity differences between reservoir and cap-rock. If the \( V_p/V_s \) ratio is 2 in the cap-rock and 2 in the reservoir, the slope of the background trend is \(-1\), that is a 45° slope diagonal to the gradient and intercept axes. Figure 4.12 shows different lines corresponding to varying \( V_p/V_s \) ratio in the reservoir and the cap-rock.

The rotation of the line denoting the background trend will be an implicit function of rock physics properties such as clay content and porosity. Increasing clay content...
at a reservoir level will cause a counter-clockwise rotation, as the $V_p/V_s$ ratio will increase. Increasing porosity related to less compaction will also cause a counter-clockwise rotation, as less-compacted sediments tend to have relatively high $V_p/V_s$ ratio. However, increasing porosity related to less clay content or improved sorting will normally cause a clockwise rotation, as clean sands tend to have lower $V_p/V_s$ ratio than shaly sands. Hence, it can be a pitfall to relate porosity to AVO response without identifying the cause of the porosity change.

The background trend will change with depth, but the way it changes can be complex. Intrinsic attenuation, discussed in Section 4.3.4 (Luh, 1993), will affect the background trend as a function of depth, but correction should be made for this before rock physics analysis of the AVO cross-plot (see Section 4.3.6). Nevertheless, the rotation due to depth trends in the elastic contrasts between sands and shales is not straightforward, because the $V_p/V_s$ in the cap-rock as well as the reservoir will decrease with depth. These two effects will counteract each other in terms of rotational direction, as seen in Figure 4.12. Thus, the rotation with depth must be analyzed locally. Also, the contrasts between cap-rock and reservoir will change as a function of lithology, clay content, sorting, and diagenesis, all geologic factors that can be unrelated to depth. That being said, we should not include too large a depth interval when we extract background trends (Castagna and Swan, 1997). That would cause several slopes to be superimposed and result in a less defined background trend. For instance, note that the top of a soft sand will plot in the 3rd quadrant, while the base of a soft sand will plot in the 1st quadrant, giving a background trend rotated in the opposite direction to the trend for hard sands.
Fluid effects and AVO anomalies

As mentioned above, deviations from the background trend may be indicative of hydrocarbons, or some local lithology or diagenesis effect with anomalous elastic properties (Castagna et al., 1998). Foster et al. (1997) mathematically derived hydrocarbon trends that would be nearly parallel to the background trend, but would not pass through the origin in $R(0)$ versus $G$ cross-plots. For both hard and soft sands we expect the top of hydrocarbon-filled rocks to plot to the left of the background trend, with lower $R(0)$ and $G$ values compared with the brine-saturated case. However, Castagna et al. (1998) found that, in particular, gas-saturated sands could exhibit a variety of AVO behaviors.

As listed in Table 4.1, AVO class III anomalies (Rutherford and Williams, 1989), representing soft sands with gas, will fall in the 3rd quadrant (the lower left quadrant) and have negative $R(0)$ and $G$. These anomalies are the easiest to detect from seismic data (see Section 4.3.1.1).

Hard sands with gas, representing AVO class I anomalies, will plot in the 4th quadrant (lower right) and have positive $R(0)$ and negative $G$. Consequently, these sands tend to show polarity reversals at some offset. If the sands are very stiff (i.e., cemented), they will not show a large change in seismic response when we go from brine to gas (cf. Chapter 1). This type of AVO anomaly will not show up as an anomaly in a product stack. In fact, they can plot on top of the background trend of some softer, brine-saturated sands. Hence, very stiff sands with hydrocarbons can be hard to discriminate with AVO analysis.

AVO class II anomalies, representing sands saturated with hydrocarbons that have very weak zero-offset contrast compared with the cap-rock, can show great overlap with the background trend, especially if the sands are relatively deep. However, class II type oil sands can occur very shallow, causing dim spots that stick out compared with a bright background response (i.e., when heterolithics and brine-saturated sands are relatively stiff compared with overlying shales). However, because they are dim they are easy to miss in near- or full-stack seismic sections, and AVO analysis can therefore be a very helpful tool in areas with class II anomalies.

Castagna and Swan (1997) discovered a different type of AVO response for some gas sands, which they referred to as class IV AVO anomalies (see Table 4.1), or a "false negative." They found that in some rare cases, gas sands could have negative $R(0)$ and positive $G$, hence plotting in the 2nd quadrant (upper left quadrant). They showed that this may occur if the gas-sand shear-wave velocity is lower than that of the overlying formation. The most likely geologic scenario for such an AVO anomaly is in unconsolidated sands with relatively large $V_p/V_s$ ratio (Foster et al., 1997). That means that if the cap-rock is a shale, it must be a relatively stiff and rigid shale, normally a very silt-rich shale. This AVO response can confuse the interpreter. First, the gradients of sands plotting in the 2nd quadrant tend to be relatively small, and less sensitive to fluid type than the gradients for sands plotting in the 3rd quadrant. Second, these AVO anomalies will actually show up as dim spots in a gradient stack. However, they should
stand out in an $R(0) - G$ cross-plot, with lower $R(0)$ values than the background trend. Seismically, they should stand out as negative bright spots.

**Pitfalls**

- Different rock physics trends in AVO cross-plots can be ambiguous. The interpretation of AVO trends should be based on locally constrained rock physics modeling, not on naive rules of thumb.
- Trends within individual clusters that do not project through the origin on an AVO cross-plot cannot always be interpreted as a hydrocarbon indicator or unusual lithology. Sams (1998) showed that it is possible for trends to have large offsets from the origin even when no hydrocarbons are present and the lithology is not unusual. Only where the rocks on either side of the reflecting surface have the same $V_p/V_s$ ratio will the trends (not to be confused with background trends as shown in Figure 4.12) project through the origin. Sams showed an example of a brine sand that appeared more anomalous than a less porous hydrocarbon-bearing sand.
- Residual gas saturation can cause similar AVO effects to high saturations of gas or light oil. Three-term AVO where reliable estimates of density are obtained, or attenuation attributes, can potentially discriminate residual gas saturations from commercial amounts of hydrocarbons (see Sections 4.3.12 and 4.3.15 for further discussions).

**Noise trends**

A cross-plot between $R(0)$ and $G$ will also include a noise trend, because of the correlation between $R(0)$ and $G$. Because $R(0)$ and $G$ are obtained from least-square fitting, there is a negative correlation between $R(0)$ and $G$. Larger intercepts are correlated with smaller slopes for a given data set. Hence, uncorrelated random noise will show an oval, correlated distribution in the cross-plot as seen in Figure 4.13 (Cambois, 2000).

Furthermore, Cambois (2001) formulated the influence of noise on $R(0)$, $G$ and a range-limited stack (i.e., sub-stack) in terms of approximate equations of standard deviations:

\[
\sigma_{R(0)} = \frac{3}{2} \sigma_s,
\]

\[
\sigma_G = \frac{3\sqrt{5}}{2} \frac{\sigma_s}{\sin^2 \theta_{\text{max}}}
\]

\[
\sigma_{G} = \sqrt{5} \frac{\sigma_{R(0)}}{\sin^2 \theta_{\text{max}}}
\]

(4.38)

(4.39)

(4.40)
Figure 4.13 Random noise has a trend in $R(0)$ versus $G$ (after Cambois, 2000).

and

$$
\sigma_n = \sqrt{n} \cdot \sigma_s
$$

(4.41)

where $\sigma_s$ is the standard deviation of the full-stack response, $\sigma_n$ is the standard deviation of the sub-stack, and $n$ is the number of sub-stacks of the full fold data. As we see, the stack reduces the noise in proportion to the square root of the fold. These equations indicate that the intercept is less robust than a half-fold sub-stack, but more robust than a third-fold sub-stack. The gradient is much more unreliable, since the standard deviation of the gradient is inversely proportional to the sine squared of the maximum angle of incidence. Eventually, the intercept uncertainty related to noise is more or less insensitive to the maximum incidence angle, whereas the gradient uncertainty will decrease with increasing aperture (Cambois, 2001).

Simm et al. (2000) claimed that while rock property information is contained in AVO cross-plots, it is not usually detectable in terms of distinct trends, owing to the effect of noise. The fact that the slope estimation is more uncertain than the intercept during a least-square inversion makes the AVO gradient more uncertain than the zero-offset reflectivity (e.g., Houck, 2002). Hence, the extension of a trend parallel to the gradient axis is an indication of the amount of noise in the data.
Fluid versus noise trends

In areas where fluid changes in sands cause large impedance changes, we tend to see a right-to-left lateral shift along the intercept direction. This direction is almost opposite to the noise direction, which is predominantly in the vertical/gradient direction. In these cases there should be a fair chance of discriminating hydrocarbon-saturated sands from brine-saturated sands, even in relatively noisy data.

Simm et al. (2000) furthermore stressed that one should create AVO cross-plots around horizons, not from time windows. Horizon cross-plot clearly targets the reservoir of interest and helps determine the noise trend while revealing the more subtle AVO responses. Moreover, only samples of the maximum amplitudes should be included. Sampling parts of the waveforms other than the maxima will infill the area between separate clusters, and a lot of samples with no physical significance would scatter around the origin in an $R(0)-G$ cross-plot. However, picking only peaks and troughs raises a delicate question: what about transparent sands with low or no impedance contrast with overlying shales? These are significant reflections with very small $R(0)$ values that could be missed if we invert the waveform only at absolute maxima (in commercial software packages for AVO inversion, the absolute maxima are commonly defined from $R(0)$ sections). Another issue is shale-shale interfaces. These are usually very weak reflections that would be located close to the origin in an AVO cross-plot, but they are still important for assessment of a local background trend.

There are also other types of noise affecting the AVO cross-plot data, such as residual moveout. It is essential to try to reduce the noise trend in the data before analyzing the cross-plot in terms of rock physics properties. A good pre-processing scheme is essential in order to achieve this (see Section 4.3.6).

Cambois (2000) is doubtful that AVO cross-plots can be used quantitatively, because of the noise effect. With that in mind, it should still be possible to separate the real rock physics trends from the noise trends. One way to distinguish the noise trend is to cross-plot a limited number of samples from the same horizon from a seismic section. The extension of the trend along the gradient axis indicates the amount of noise in the data (Simm et al., 2000). Another way to investigate noise versus rock physics trends is to plot the anomaly cluster seen in the AVO cross-plot as color-coded samples onto the seismic section. If the cluster is mainly due to random noise, it should be scattered randomly around in a seismic section. However, if the anomaly corresponds with a geologic structure and closure, it may represent hydrocarbons (see Plate 4.10).

Finally, we claim that via statistical rock physics we can estimate the most likely fluid and lithology from AVO cross-plots even in the presence of some noise. This is referred to as probabilistic AVO analysis, and was first introduced by Avseth et al. (1998b). This method works by estimating probability distribution functions of $R(0)$
and $G$ that include the variability and background trends. Houck (2002) presented a methodology for quantifying and combining the geologic or rock physics uncertainties with uncertainties related to noise and measurement, to obtain a full characterization of the uncertainty associated with an AVO-based lithologic interpretation. These methodologies for quantification of AVO uncertainties are explained in Section 4.3.12.

**How to assess the noise content in AVO cross-plots**

- Make cross-plots of full stack versus gradient, in addition to $R(0)$ versus $G$. The stack should have no correlation with the gradient, so if trends in $R(0)-G$ plots are still observed in stack vs. $G$, these trends should be real and not random noise (Cambois, 2000).
- Identify the location of AVO anomalies in seismic sections. Color-code AVO anomalies in $R(0)-G$ plots and then superimpose them onto your seismic sections. Do the anomalies make geologic sense (shape, location), or do they spread out randomly?
- Plot the regression coefficient of $R(0)$ and $G$ inversion onto the seismic to identify the areas where $R(0)$ and $G$ are less reliable.
- Cross-plot a limited number of samples from the same horizon from a seismic section. The extension of the trend along the gradient axis indicates the amount of noise in the data (Simm et al., 2000).

**4.3.11 AVO attributes for hydrocarbon detection**

The information in the AVO cross-plots can be reduced to one-dimensional parameters based on linear combinations of AVO parameters. This will make the AVO information easier to interpret. Various attributes have been suggested in the literature, and we summarize the most common below (AVO inversion-based attributes are discussed in Section 4.4).

**Far- versus near-stack attributes**

One can create several AVO attributes from limited-range stack sections. The far stack minus the near stack (FN) is a “rough” estimate of an AVO gradient, and in particular it is found to be a good attribute from which to detect class II AVO anomalies (Ross and Kinman, 1995). For class II type prospects, the far stack alone can be a good attribute for improved delineation. However, for class IIp anomalies, both the near and the far stack can be relatively dim, but with opposite polarities. Then the difference between far and near will manifest the significant negative gradient that is present. In contrast, a conventional full stack will completely zero-out a class IIp anomaly. Ross and Kinman (1995) suggested the following equation for the FN attribute depending on whether