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History and Reviews on Amplitude Variation with Offset (AVO) Technology

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ABSTRACT: Geologists and geophysicists have relied on seismic imaging for decades to identify fluid effects and precisely map hydrocarbons from brine in a target reservoir. It is clear that pore fluids have a direct role in seismic signatures since bright spots (high reflection amplitudes) were used as a signal of hydrocarbon in the early stages of exploration. Pre-stack amplitude-versus offset (AVO) analysis of reflected compressional waves shows great promise due to elastic property variations at the reflecting contact as the offset develops. When using AVO, seismic lithology may be evaluated and hydrocarbons can be discovered with high confidence. Under optimal well-control circumstances, quantitatively analyzing amplitude changes with offset is a great method for identifying fluids. Regular seismic lithologic study seldom goes beyond the detection of aberrant activity if wells are not available. Improvements in computational power and seismic gathering and processing technologies directly contribute to the development of cutting-edge methods. This study tries to put the state of the art in perspective by bringing up concepts and techniques from the last century.

KEYWORDS: Zoeppritz equation, Seismic, AVO, and Inversions

I. INTRODUCTION

Seismic exploration tool Amplitude Variation with Offset (AVO) analyzes the relationship between seismic reflection amplitude and offset. Using the reflection coefficients, incidence angle, and elastic parameters, Knott (1899) was the first to examine how seismic energy changes with distance from the contact. Zoeppritz (1919) was the next entry on the schedule, it was unfortunate that Zoeppritz's equations were so difficult to understand since they were so intricate. To analyze and model AVO anomalies, other scholars like Bortfeld (1961) and Aki and Richards (1980) developed various assumptions and came up with an approximation for AVO analysis pre-stack data. As stated by Resnick (1985), the quality of seismic data gathered at sea tends to be greater, and hence, analysis is more common. Even though land data may suffer from uneven coverage and missing offsets, AVO analysis may still be used for such data. Significant short-term multiple issues might make marine data better for AVO analysis than land data in certain circumstances (Castagnaet al., 1993).

Direct hydrocarbon detection is based on Gassmann's (1951) fundamental formulae. Using these equations, we may expect a significant decrease in P-wave velocity and a small increase in S-wave velocity after the injection of a small amount of gas into the pore space of compressible brine-saturated sand. Reduced density is the root cause of AVO anomalies, which manifest as either "bright" or "dim" spots due to changes in the P-wave reflection coefficient.

Compressional wave velocity and shear wave velocity were used by Domenico (1984) to figure out the lithology and porosity of a rock sample in his lab. Ostrander (1984), Shuey (1985), and Gassaway (1984) have demonstrated that AVO may be used with traditional P-wave data in somewhat unconsolidated sediments where the direct S-wave collection is impractical. These changes carry information about the velocity of the S-waves. There are many different types of fluids in reservoir rocks, and Sandikci (2010) demonstrated that Lambda-Mu-Rho (LMR) analysis may be utilized as a seismic attribute for identifying and determining which ones are present.

An important part of the AVO technique is detecting, modeling, and inverting data from the AVO system. Out of all the quantitative seismic methods employed in the oil business today, offset-dependent amplitude analysis (AVO analysis), acoustic and elastic impedance inversion, and forward seismic modeling are the most extensively utilized. All across the globe, explorationists are

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finding hydrocarbons utilizing AVO anomalies (Castagnaet al., 1993). When compared to other geophysical approaches, the seismic reflection method utilized in hydrocarbon exploration is highly dangerous and costly. With the help of various data processing methods, AVO is used in seismic research. Natural gas reserves may be located directly by modeling the underlying structures.

Using migration to concentrate on seismic sections and look for structural impacts that might be useful for AVO analysis in complicated structural regions, as stated by Mosher et al (1996), is possible to estimate bottom simulating reflection (BSR) characteristics from AVO results using a forward modeling method. De-risking exploration locations and better defining the size and composition of existing hydrocarbon reservoirs are two common uses for AVO among most firms' normal practices (Castagna, 1993). The study by Rocky et al. (2014) revealed that AVO can be used in risk analysis to understand the uncertainties in their interpretations. Using rock physics modeling, Eggen, (2012) conducted AVO analysis on modeled data gathered from the accessible wells and successfully identified the many scenarios that may be present in a turbidite reservoir

Using pre-stacked seismic data, which geophysicists began to look at in the 1980s, it is possible to predict a change in magnitude with offset. In the 1980s, geophysicists started to look at pre-stack seismic data and found that it was possible to model how amplitude changed with offset (Ostrander, 1984). The difference in acoustic impedance across the interface controls the zero-offset reflectivity, R(0). Koefoed (1955) was the first to show that the ratio V_P/V_S(Poisson's ratio) was important to the offset-dependent reflectivity (ODR). Ostrander showed in 1984 that a gas-filled formation would have a very low Poisson's ratio compared to the non-gaseous formations around it. This would cause the positive amplitude vs. angle to go up at the bottom of the gas layer and the negative amplitude vs. angle to go up at the top of the gas layer.

II. FACTORS AFFECTING AVO ANALYSIS

An AVO analysis's huge promise for prospecting has not been partially achieved because of the many difficulties and complexities faced in identifying and interpreting Offset Dependent Reflectivity (ODR). This has often hampered AVO's ability to be used as a prospecting tool. As indicated in Table 1 by Castagna et al (1993), for most AVO analysis approaches, all elements that are not reflection coefficient change with offset are regarded as undesirable noise and must be dealt with or correctly accounted for in the processing of the data. The angle-dependent total system response/wavelet variation must be taken into account while attempting to isolate ODR.

Table 1: Seismic Amplitudes and the Factors That Affect Them.

A	Information That Is Sought After (Signal).	1. The relationship between the reflection coefficients and the angle of incidence.
B	Information That Maybe Available Considered noise for some methods; signal for others).	1. Reflections that are a composite of those from numerous interfaces. 2. The tuning that is brought about by NMO convergence. 3. Mode conversions.
C	Factors that are not reliant on an offset (noise).	1. Random noise. 2. Instrumentation. 3. Coupling between the source and the receiver.
D	Variables Involving an Offset Dependence (noise).	1. Directivity between the source and receiver, including ghosting, and an array of response. 2. The emergence of angles of attack. 3. Coherent noise, multiples. 4. Distribution on a spherical scale. 5. Processing artifacts such as stretch, distortion, and NMO errors.

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		6. Attenuation due to inelasticity and anisotropy. 7. Transmission coefficients and scattering above the target . 8. Structural complexity.

Source: Castagna*et al* (1993).

III. FACTORS AFFECTING AVO ANALYSIS

A study by Sherwood et al. (1983) found that "Offset-dependent-reflectivity can't be fully grasped with only the interface reflection coefficient. Layer reflections, transition zones, and complicated layered sequences are more common occurrences." The superposition of reflections from several interfaces necessitates modeling and inversion.

"Tuning" is the term used to describe the wave interference that may occur when events or reflectors are too close together to be seen separately on the seismic. When the traditional bed tuning phenomena (Widess, 1973) is viewed as a function of offset, it gets more complicated. Ostrander (1984), Hindlet and McDonald (1986), Ball (1988), and Swan (1988) were among the researchers that looked into this. Differentiating the normal moveout equation yields an approximation of the apparent change in temporal thickness of a layer, as shown by Ostrander (1984). Also

$$\frac{\Delta t_\theta}{\Delta t_0} \approx \cos\theta \qquad (1)$$

Where $t_0$is the normal incidence arrival time and t_θ is the arrival time for angle θ in the layer.

Two reflectors that are separated by less than one-quarter of the wavelength will appear as a single high-amplitude reflector. Fig.1(a) illustrates how the normal incidence time thickness-frequency product might affect the amplitude of tuning. Fig. 1(b) depicts how the amplitude and temporal thickness change with increasing offset.

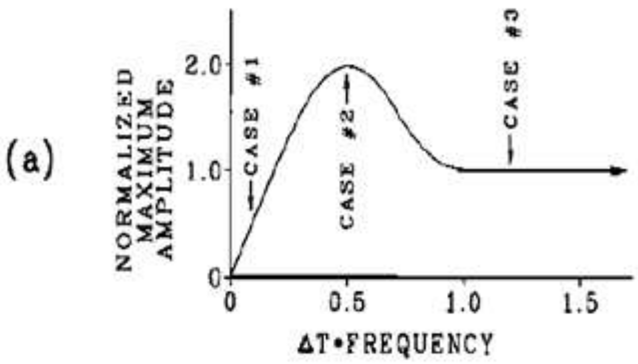

Fig. 1(a) The effect of thin-bed tuning is shown by the relationship between the normalized maximum amplitude and the frequency-time thickness product. Cases 1, 2, and 3 are shown. Castagna*et al* (1993).

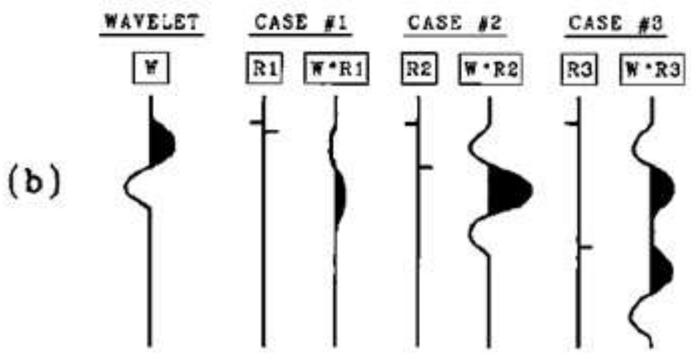

Fig.1(b) Convolution of the wavelet with three reflectivity series R1, R2, and R3 that are, in ascending order, below tuning, at tuning, and above tuning. (Castagna*et al* (1993).

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Juhlin and Young (1993), Lin and Phair (1993), Bakke and Ursin (1998), Dong (1998), and Sbonelo, (2008) amongst others also demonstrated the effect of tuning on AVO. The influence of tuning on AVO is more apparent on Dip Moveout (NMO) corrected data, according to Yong and Satinder (2007). However, identifying and quantifying AVO (amplitude vs. offset) behavior alone is not sufficient, according to Vaughn, (1987). An array of factors, including impedance and stiffness contrasts and AVO tuning effects, influence AVO behavior.

According to Juhlin and Young (1993), thin layers buried in homogeneous rock may cause considerably different AVO responses than simple interfaces of the same lithology, which is consistent with their findings. They showed that the AVO response of a thin bed may be approximated by modeling it as an interface phenomenon between plane P-waves from a thin layer, provided that the variations in elastic properties across layer borders are small. Thin-bed tuning has a bigger impact on the AVO response of a high-velocity layer embedded in homogeneous rock than it does on a low-velocity layer. The following expression is proposed by Lin and Phair (1993) for the amplitude variation with angle (AVA) response of a thin layer:

$$R_1(\theta) = \omega_0 \Delta T(0) \cos\theta \times R(\theta) \quad (2)$$

With ω_0 being the dominant frequency in the wavelet's spectral distribution.$\Delta T(0)$ denotes the time taken for light to travel from the top to the bottom of the thin layer, at normal incidence, and$R(\theta)$ is the top interface's reflection.

Using tuning correction factors as a function of offset, Bakke, and Ursin (1998) improved upon Lin and Phair's earlier work for a wide seismic wavelet. They showed that an earthquake may be felt up in a thick layer.

$$d(t,y) = R(y)P(t) \quad (3)$$

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 function from a thin layer is

$$d(t,y) \approx R(y)\Delta T(0)C(y)P'(t) \quad (4)$$

$P'(t)$ is the time derivative of the pulse, $\Delta T(0)$ is the travel time thickness of the thin layer at zero offsets.

$C(y)$ is the offset-dependent AVO tuning factor expressed as

$$C(y) = \frac{T(0)}{T(y)}\left[1 + \frac{V_{RMS}^2 - V^2}{2T(0)^2 V_{RMS}^4} y^2\right] \quad (5)$$

$T(0)$ and $T(y)$ are the travel times at zero offsets at a given non-zero offset respectively.

The root – mean – square velocity V_{RMS} along a ray-path is defined as

$$V_{RMS}^2 = \frac{\int_0^t V^2(t)dt}{\int_0^t dt} \quad (6)$$

For small velocity contrast, $(V_{RMS} \approx V)$, and the last term in equation 2.26 is ignored.

Thus the AVO tuning factor is approximated as

$$C(y) \approx \frac{T(0)}{T(y)} \quad (7)$$

Contributions from P-wave multiples and converted shear waves are essential when there is a considerable difference in elastic characteristics. For the AVO response of probable hydrocarbon reserves, ray-tracing modeling often ignores the locally converted shear wave. The Zoeppritz equations are used to explain the reflection amplitudes in primary-only ray-trace modeling. When using primaries-only Zoeppritz modeling, Simmons and Backus (1994) claim that the locally converted shear waves typically have a first-order influence on the seismic response.

As layer thickness decreases, interference between converted waves and primary reflections from the layer base becomes more critical. Often, the seismogram obtained in this manner differs from the seismogram obtained from the primary sources solely from the Zoeppritz hypothesis. Full elastic modeling should be applied, including the inscribed multiples and converted wave modes. There is evidence that surface-related multiples and P to SV mode converted waves may interfere with main pre-stack amplitudes and create substantial distortions of the AVO responses.

Pictured in Fig. 2 are converted S-waves and their multiples inside a layer.

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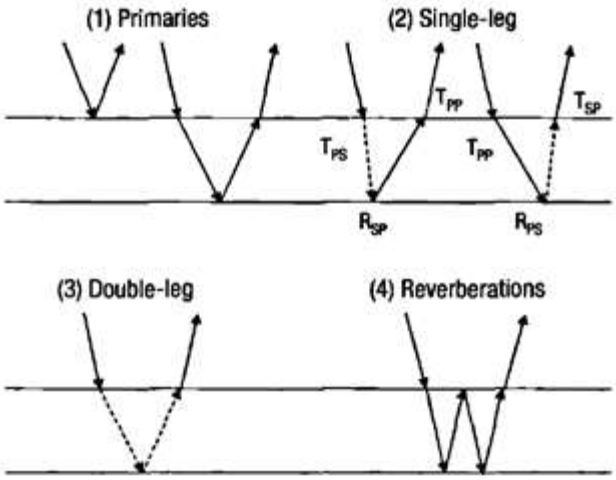

When thin layers are available, as shown in Fig. 2, it is necessary to incorporate converted S-waves and multiples in the AVO modeling, since these modes interact with the primary. (1) Primary reflections; (2) Single-leg shear waves; (3) double-leg shear waves; and (4) primary reverberations. (After Simmons and Backus, 1994).

T_{PS}= the transformation of the P-wave into the transmitted S-wave
T_{SP}= The P-wave transmission, which was transformed from S-wave.
R_{PS}= A transformed version of the P-wave into a reflected S-wave
R_{SP}= reflected P-wave converted from S-wave
T_{PP}= transmitted P-wave from the original P-wave

IV. NEAR SURFACE CONSIDERATIONS

Seismic amplitudes can be perturbed by a variety of near-surface effects. It is possible to alter AVO by altering the strength of the source and the coupling between the source and receiver. Surface-consistent processing may be used to verify this. The pattern of radiation from the source, the reaction of the geophone, and the response of the array are all clear emerging effects that are angle-dependant. All of these side effects are well-documented and easy to fix. Free surface mode conversion energy partitioning may also be rectified (Cerveny and Ravindra, 1971). Angle-dependent ghosting is viewed as a second-order issue in aquatic environments (Castagna et al 1993). As explained below, near-surface transmission impacts may still have a role.

V. GENERAL DATA QUALITY ISSUES

Seismic data quality is the most important prerequisite for a wonderful AVO analysis. A good signal-to-noise ratio (SNR), wide dynamic range (WDR), large aperture (offset), and fold are all essential. If amplitude calibration is not possible, at least the channels should be constant throughout the survey. It's possible that maintaining a proper balance between the channels will be a challenge. Frasier (1988) noted that individual traces have lower S/N than stacks, hence channel balancing measures may just equalize the noise level. In particular, coherent noise created by the source in the form of multiple converted waves and diffractions is a major issue Incoherent noise is less of an issue, as shown by Pan et al. (1990).

F-k filtering is often used to reduce coherent noise. AVO estimates may be skewed by coherent noise, such as multiples with variable spatial frequency (moveout) inside a collection. Ray parameter filtering and predictive deconvolution may also be used to combat multiples in a variety of contexts. It's very impossible to eliminate misleading amplitude variations caused by multiples with tiny residual moveout (half to one cycle). For marine AVO, this is arguably the most critical concern. Full-waveform inversion may be achieved when specific source-generated noise is transformed into a signal.

VI. CURVATURE AND STRUCTURAL COMPLEXITY

Using the optics model, we predict that curved contacts will either focus or scatter seismic energy. Hilterman (1975) established the curvature effect (CE) as the ratio of reflection amplitudes out of a curved interface to those of a flat interface for a reflecting interface.
According to Hilterman (1975), the following constitutes normal incidence:

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$$CE = (1 + Z/A)^{-1/2} \qquad (8)$$

The bed's radius of curvature, A, and the depth, Z, are both used here in equation (8).
This equation was generalized to the non-normal incidence situation by Shuey et al. (1984).

$$CE(\theta_1) = \left(1 + \frac{Z}{A_x \cos^2\theta_1}\right)^{-1/2}\left(1 + \frac{Z}{A_y}\right)^{-1/2} (9)$$

Where
A_x is the radius of curvature in the x-direction, A_y is the radius of curvature in the y-direction, and Z is the depth of crest of anticline or trough of a syncline.

It is worthy to note that for buried focus synclines, the radii are negative, and that, at normal incidence, Shuey's equation reduces to Hilterman's equation. According to Castagnaet al (1993), for anticlines, the curvature effect decreases with an offset according to equation (9), whereas for synclines with a focus above the surface, it increases, and for synclines with a focus below the surface, it decreases according to the same equation.
Curvature may be corrected by DMO, as shown by Macleod and Martin (1988).In the general situation of three-dimensional curvature, Bernitsas's (1990) equation was developed:

$$CE(\theta_1) = (1 + Z/A_x)^{-1/2}(1 + Z/A_y)^{-1/2} \times \left[\frac{X^2}{Z(A_x+Z)} + \frac{Y^2}{Z(A_Y+Z)} + 1\right]^{-1/2} \qquad (10)$$

Where X, Y, and Z are the spatial coordinates.

Complex surface roughness isn't well understood. Herman and Bionk (1990) say a transition layer may mimic surface roughness. According to Resnick et al. (1987), a dip causes errors in an angle of incidence, it mixes information from distinct subsurface sites, and dip also causes incorrect normal moveout and mispositioning events. Prestack migration that preserves amplitude is optimal for correcting the lensing impact of curved layers above the target. Because of the intricate design, lateral overburden velocity gradients and accompanying AVO problems are not uncommon.

VII. GEOMETRICAL SPREADING

The equations needed to correct for geometrical spreading were given by Newman (1973). At normal incidence

$$D_0 = tV_a^2/V_1, \qquad (11)$$

$$t_0 = \sum t_i, \qquad (12)$$

And

$$V_a^2 = \sum t_i V_i^2 / t_0 \qquad (14)$$

Where, D_0= normal incidence geometric divergence, V_1= velocity in the first layer, t_0= zero-offset two-way reflection time, t_i= interval two-way transit time of the ith layer, V_a = time-weighted RMS velocity, and n = the number of layers i.

At non-normal incidence

$$D(\theta_1) = \frac{\left(X^2 + 2X\sum d_i \tan^3\theta_1\right)^{1/2}}{\tan\theta_i} \qquad (15)$$

and

$$X = 2\sum d_i \tan\theta_i \qquad (16)$$

Where, $D(\theta_1)$= geometric divergence versus angle of incidence, X = offset, θ_i= incidence angle in the ith layer, and d_i= thickness of the ith layer.

VIII. NORMAL MOVEOUT ERRORS

Sample-by-sample measurements of the AVO need correcting the common depth point (CDP) gather for normal moveout (NMO) (Spratt, 1987; Swan, 1988). Special attention must be paid to conventional velocity analysis to get the task done (Chiburis, 1984). Stacking velocity and ODR have a basic ambiguity, as seen in Figure 3. (Spratt, 1987). AVO and approaches that measure the energy inside a window around the event of interest (Mazzotti, 1990) have fewer NMO-related issues. Another option for resolving the issue is to use seismic modeling and waveform inversion.

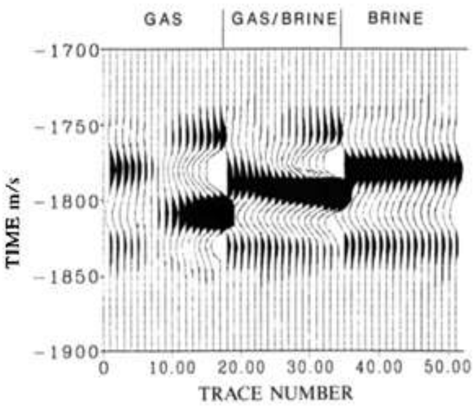

Fig. 3. Illustrates that all three models (A, B, and C) have the same P-wave velocity structure but

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have different S-wave velocity structures, and this is reflected in the results of the three synthetic CDP ensembles that have had their NMO errors addressed. (After Castagnaet al, 1993).

IX. OVERBURDEN EFFECTS

Depending on the angle at which transmission coefficients are calculated and applied, the P-amplitude waves may also vary. When the reflectivity above the target is particularly high, this issue is more pronounced. Both the P-wave and S-wave velocity structures influence angle-dependent transmission losses (Castagna et al, 1993).

Fig. 4 shows the effects of varying V_p and V_s over a particular target. Models A and B have the same V_p structure but different V_s structures. At the near offset, the reflection amplitudes are the same but diverge significantly at far offsets. Similarly, models A and C have the same V_s structure but different V_p structures.

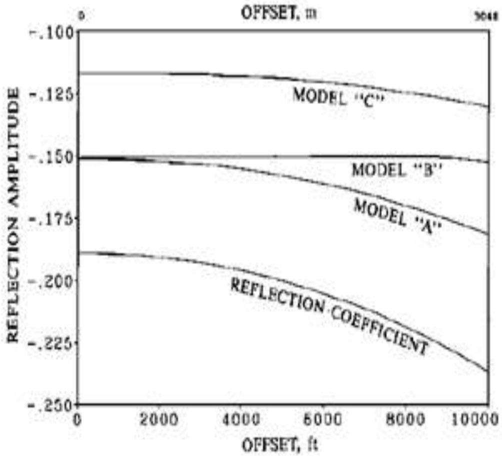


Fig. 4. Illustrates that different overburden transmission losses result in the same target reflection coefficient versus offset response for three models (A, B, and C). (After Castagnaet al, 1993).

Both models have a comparable AVO, but their near-offset reflection amplitudes are substantially different. Figure 5 shows that V_P/V_S variations in the very near surface can also significantly affect the AVO response.

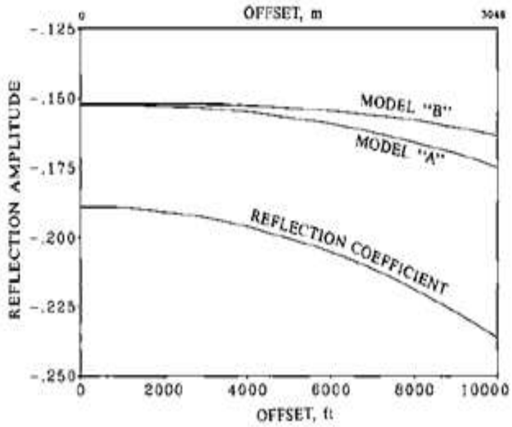
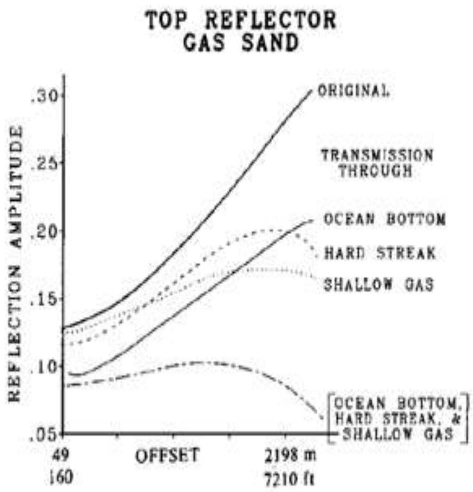


Fig. 5. Variation in amplitude with offset for two models (A and B) with the same target reflection coefficient versus offset response but differing overburden transmission losses. Near-surface Vp/Vs (wet sand) is high in Model A, whereas it is low in Model B. (dry sand). (After Castagnaet al, 1993).

A layer stripping approach may be used to account for the overburden, as illustrated in Fig. 6, and transmission loss is the major difficulty faced in AVO analysis, as stated by Gassaway (1984).

Fig.6. is the intended area of influence (AVO) may be altered by overburden transmission losses. In this particular scenario, transmission via the ocean floor, a hard streak, and shallow gas sand all work together to suppress the target gas sand's amplitude rise with offset (from Gassaway, 1984).



The statistical correction of global transmission loss is fairly straightforward. Deep gas, on the other hand, would cause significant lateral shifts in overburden effects, which would be difficult to remediate. Since it is impossible to precisely quantify the overburden, deterministic

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correction presents a challenge. AVO measurements may be normalized by comparing them to a reference event (Chiburis, 1984 and 1987).
The normalized AVO is given by:

$$\frac{a(x)^{tar}/a_0^{tar}}{a(x)^{ref}/a_0^{ref}} \quad (17)$$

where
$a(x)^{tar}$ is the AVO for the target event.
$a(x)^{ref}$ is the AVO for the reference event.
a_0^{tar} is the normal incidence amplitude for the target
a_0^{ref} is the normal incidence amplitude for the reference

Huston and Backus (1986) claim that checking ties at line may occasionally reveal problematic differences in the overburden. Errors in ray-tracing calculations of the local angle of incidence due to spatial changes in overburden velocities are common. This is what we get if we can express the overburden as a linear velocity gradient:

$$V_i = V_0 + K_z \quad (18)$$

Where,
V_i = interval velocity
Z = depth,
V_0= velocity when z = 0, and
K is a constant, then (Ostrander, 1984)

$$\theta_i = tan^{-1}\left(zX + V_0^{x/k}\right)/(z^2 + 2V_0 z/K - X^2/4) \quad (19)$$

AVO errors happen when vertical or horizontal speeds are different from those used to calculate the angle of incidence (Xu and McDonald, 1988). When the speed goes up or down vertically or horizontally, Poisson's ratio will be bigger. It also works the other way around. Chiburis (1993) gave a way to fix overburden effects and some acquisition effects by comparing the amplitude of the target horizon to the amplitude of the reference horizon. On amplitude-preserved imaging in complex areas and AVO corrections due to overburden effects, there have been a number of longer and more in-depth contributions. These include AVO effects caused by scattering attenuation in heterogeneous overburden and AVO effects in places where the structure is complicated.

X. AVO IN STRUCTURALLY COMPLEX AREAS

Zoeppritz equations work in basins where the layers are mostly flat and the basin is always sinking. This is because the equations are based on the idea that there is only one boundary between two semi-infinite layers that go on forever. Zoeppritz's assumptions will be broken by thin beds that slope downward, vertical heterogeneities, faulting, and tilting. Resnick et al. (1987) look at how geologic dip affects the AVO signature, while MacLeod and Martin (1988) talk about how reflector curvature affects the signature. Pre-stack depth migration takes into account the complexity of the structure (PSDM).

But without keeping the amplitudes, you can get good structural images from several PSDM routines. Grubb et al. (2001) did an amplitude-preserving PSDM followed by an AVO inversion to find out how sensitive PSDM-migrated images are to changes in both structure and amplitude due to changes in velocity. Resnick et al. (1987) showed that problems caused by dip can be fixed by either partial migration before stack or full migration before stack. They stressed how important it is to make sure that these prestack processes handle amplitude correctly to get rid of dip as a problem when doing AVO analysis on data from areas with complex structures.

XI. HETEROGENEOUS OVERBURDEN SCATTERING ATTENUATION AS AVO EFFECT.

Widmaier et al. (1996) demonstrated how to stabilize the AVO response for a thin-layered overburden. These effects must be taken into consideration while examining a seismic reflector as a target. Transmission losses owing to scattering attenuation and velocity anisotropy will result from a thin-bedded overburden. To correct the impact of thin-bedded strata on seismic travel durations and amplitudes, they combined the generalized O'Doherty Anstey Formula (Shapiro et al., 1994a) with amplitude-preserving migration/inversion algorithms and AVO analysis. They demonstrated how accounting for the impact of thin-bedded scattering improves the accuracy with which the zero-offset amplitude and the AVO gradient are estimated. Based on Widmaier's research, Sick et al. (2003) devised a method to compensate for the scattering attenuation generated by randomly distributed heterogeneities above a target reflector. The generalized O' Doherty Anstey Formula is given as

$$T(f)\,\alpha\,T_0 e^{-(\alpha(f,\theta)+i\beta(f,\theta))L} \quad (20)$$

Where;
f is the frequency and $\alpha\beta$ represent the angle- and frequency-dependent scattering attenuation and

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phase shift coefficients, respectively.
θ is the initial angle of an incident plane wave at the top surface of a thinly layered composite stack
L represents the thickness of the thinly layered stack
T_0 represent the transmissivity for a homogeneous isotropic medium that causes a phase shift.
To put it another way, the angle-dependent time-harmonic effective transmissivity T for the scalar wave (p-waves in an acoustical 1-dimensional (1D) media or SH-waves in an elastic 1-dimensional medium) may be approximated by equation 20. The equation also demonstrates how the thin layer affects the amplitude and phase of the reflections from 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:

$$\tilde{T}(f)\ \alpha\ e^{-(\alpha(f,\theta)+i\beta(f,\theta))L} \qquad (21)$$

For a p-wave in an acoustic 1D medium, Widmaieret al (1996) derived the scattering attenuation, $\alpha$and the phase coefficient, $\beta$from Shapiro et al (1994b) as

$$\alpha(f,\theta) = \frac{1}{\cos^2\theta}\frac{4\pi^2 a\sigma^2 f^2}{V_0^2+16\pi a^2 f^2\cos^2\theta} \qquad (22)$$

and

$$\beta(f,\theta) = \frac{\pi f\sigma^2}{V_0\cos\theta}\left[1-\frac{8\pi^2 a^2 f^2}{V_0^2+16\pi a^2 f^2\cos^2\theta}\right] \qquad (23)$$

Where the statistical parameters of the reference medium include:
Spatial correlation length$a$, standard deviationσ, and mean velocity V_0.

To represent the 1D random media, we use an exponential correlation function to describe the changing P-wave velocity. The absolute value of the p-transmissivity wave diminishes as the angle of incidence increases. (Avseth et al, 2005). If the uncorrected seismic amplitude (i.e., the analytical P-wave particle displacement) is defined according to ray theory by:

$$U(S,G,t) = R_C\frac{1}{\gamma}W(t-\tau_M) \qquad (24)$$

Where
U is the seismic trace
S denotes the source
G denotes the receiver
t is the varying travel time along the ray path
$R_C$is the reflection coefficient of the reflection point M
γ is the spherical divergence factor
W is the source wavelet, and
τ_M is the travel time for the ray between source S, through reflector point M, and back to the receiver G.

Then, the expression for compensated seismic amplitude derived from a reflector beneath a thin-bedded overburden is given as

$$U^T(S,G,t) = \tilde{T}_{tw}(t) * R_C\frac{1}{\gamma}W(t-\tau_M) \qquad (25)$$

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

$$\tilde{T}_{tw}(t) = \tilde{T}_{MG}(t) * \tilde{T}_{SM}(t) \qquad (26)$$

The superscript T of $U^T(S,G,t)$ indicates that thin-bed effects have been accounted for.
Equation 25 indicates that the source wavelet,$W(t)$, is convolved with the transient transmissivity both for the downgoing ($\tilde{T}_{SM}$) and the upgoing ray paths ($\tilde{T}_{MG}$) between the source S, reflection point (M), and receiver (G). Overburden's transverse isotropic velocity behavior results in a time shift that varies with angle, as seen by the preceding equation. In addition to the amplitude decay associated with spherical divergence, it also represents the reduction of the AVO response due to multiple scattering.

It has been proposed by Widmaier et al (1995) for elastic P-wave AVO that the elastic correlation formula relies on variances and covariances of P-wave velocity, but also the S-wave velocity and density, as well as their cross-correlation functions. It was Ursin and Stovas (2002) that expanded the O'Deherty-Anstey formula further. It was observed that in the seismic frequency range, the intrinsic attenuation is more important than the scattering attenuation in a thin-bedded viscoelastic material

XII. ANISOTROPY EFFECT

It is critical to include velocity anisotropy when examining the amplitude variation with offset (AVO) response of gas sands embedded in shales. According to Thomsen (1986), most geological settings have mild anisotropy (10–20 percent).
In TI media, the elastic stiffness tensor C may be compactly written as follows:

$$C = \begin{vmatrix} C_{11} & (C_{11}-2C_{66}) & C_{13} & 0 & 0 & 0 \\ (C_{11}-2C_{66}) & C_{11} & C_{13} & 0 & 0 & 0 \\ C_{13} & C_{13} & C_{33} & 0 & 0 & 0 \\ 0 & 0 & 0 & C_{44} & 0 & 0 \\ 0 & 0 & 0 & 0 & C_{44} & 0 \\ 0 & 0 & 0 & 0 & 0 & C_{66} \end{vmatrix}$$

Where$C_{66} = \frac{1}{2}(C_{11}-C_{12}) \qquad (27)$
and where the 3-axis (z-axis) lies along the axis of symmetry.

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As can be seen in the above example, the matrix is symmetric and is comprised of five distinct submatrices: C11, C13, C33, C44, and C66. Thomsen (1986) defines three anisotropic parameters,ε, $\gamma$and δ, as a function of the five elastic components for weak anisotropy, where

$$\varepsilon = \frac{C_{11} - C_{33}}{2C_{33}} \quad (28)$$

$$\gamma = \frac{C_{66} - C_{44}}{2C_{44}} \quad (29)$$

$$\delta = \frac{(C_{13} + C_{44})^2 - (C_{33} - C_{44})^2}{2C_{33}(C_{33} - C_{44})} \quad (30)$$

ε is a constant which describes the fractional difference of the P-wave velocities in the vertical and horizontal directions:

$$\varepsilon = \frac{V_P(90°) - V_P(0°)}{V_P(\theta°)} \quad (31)$$

Equation 31 above describes the P-wave anisotropy.Likewise, the constant γ characterizes the proportionate differences in the vertical and horizontal velocities of SH-waves, which is equal to the difference in the vertical and horizontal polarizations of the horizontally propagating S-waves:

$$\gamma = \frac{V_{SH}(90°) - V_{SV}(90°)}{V_{SV}(90°)} = \frac{V_{SH}(90°) - V_{SH}(0°)}{V_{SH}(0°)} \quad (32)$$

By comparingδ, ε andγ, δis the most important parameter for normal moveout velocity and reflection amplitude.
For transversely isotropic (TI) media, Daley and Hron (1977) developed theoretical formulations for the reflection and transmission coefficients. The isotropic and anisotropic components of the P-P reflectivity in the equation are as follows

$$R_{PP}(\theta) = R_{1PP}(\theta) + RA_{PP}(\theta) \quad (33)$$

Banik (1987) assumed weak anisotropy and small offsets and showed that the anisotropic term can be expressed as

$$R_{APP}(\theta) \approx \frac{\sin^2\theta}{2}\Delta\delta \quad (34)$$

XIII. AVO AND INTRINSIC ATTENUATION

Although even homogenous sedimentary rocks are not fully elastic, intrinsic attenuation is produced by this feature. This phenomenon, which is also known as elastic absorption, may impede an accurate AVO response (e.g., Martinez, 1993). Intrinsic attenuation can be described in terms of a transfer function $\hat{G}(w,t)$ for a plane wave of angular frequency ωand propagation time t (Luh, 1993):

$$\hat{G}(w,t) = \exp(wt/2Qe + i(wt/\pi Qe)\ln w/w_o) \quad (35)$$

Where Qe = The overburden along the route of wave propagation that affects the effective quality. and $\omega_0$is an angular reference frequency.

Luh (1993) showed that wavelet attenuation may be fixed by compensating for it in three dimensions. To determine the fractional change in AVO gradient, δG, owing to absorption in the overburden, he proposed a rough, "rule of thumb" equation.

$$\delta G \approx \frac{f_1\tau}{Q_e} \quad (36)$$

where f_1 is the peak frequency of the wavelet, and τ is the zero-offset two-way travel time at the studied reflector.
How inherent attenuation may alter the P-wave reflection coefficient towards the critical angle and beyond was shown by Carcioneet al (1998). Reflection coefficients with non-normal incidence are similarly affected by the combination of attenuation, however, the inherent attenuation in certain situations may compensate for anisotropy. They discovered that anisotropic effects predominate over attenuation effects in the majority of situations. For any incidence angle, Carcione (1999) showed that the unconsolidated sediments along the sea bottom in offshore settings may be very attenuating and that these waves would have a vector attenuation perpendicular to the seafloor interface that affects AVO responses of deeper reflectors.

XIV. CONCLUSION

Before stacking, geophysicists should leverage and incorporate all available data using the entire dataset, including information that is often suppressed in the usual approach, and should bring in additional independent information by utilizing as much of the full seismic wavefield as is practically possible and collecting time-lapse observations to achieve improved quantitative parameter estimate. As a result of its complexity, erroneous assumptions, and lack of precision, AVO analysis is often misused. Unfortunately, this has led many expert explorers to believe that the method is useless and that divining rods may be used without risk. In contrast, AVO is based on extremely sound physical concepts; yet, for the reasons explained below, it is not suitable for use by anyone who is unwilling to invest the time and effort necessary to fully grasp the technology. Those who are able to make sense of AVO analysis's complexities will have a great advantage over their rivals. Anomalies are what we anticipate them to be, thus geophysics is just the study of anomalies. The preciseness of our anticipations and the exactness of any possible solution are both lacking. On the other hand, with the right application, AVO may help reduce danger and "illuminate" new opportunities that would have been missed before.

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