Estimation of Thomsen's anisotropy parameters by moveout velocity analysis

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Summary

Anisotropic NMO analysis is used to estimate anisotropy parameters in anisotropic media in combination with well-log data. Analysis of four reflection-traveltime inversions in weakly anisotropic media shows that inversion accruacy is related to the spread length and subsurface anisotropic parameters. Under their own offset range, the accuracy of estimated δ decreases with their offsets and the accuracy of estimated ε increases with their offsets, and the accuracy of the estimated Thomsen anisotropy parameter δ depends not only on the accuracy of the picked NMO velocity but also on the subsurface anisotropy parameters. The smaller the value of $(\varepsilon - \delta)$, the higher the accuracy of the estimated δ value. The results of the four reflection-traveltime inversions by semblance analysis for synthetic seismic examples demonstrate that in estimating δ, the nonhyperbolic and the shifted-hyperbolic estimations are better than the three-term Taylor-series method. Only the nonhyperbolic approximation can be used to estimate the anisotropy parameter ε accurately. Hyperbolic estimation is only suitable for estimation of elliptical anisotropy which is rarely happened in practice. The use of the method above to Blackfoot seismic data shows more challenge.

Introduction

One of the common assumptions in conventional seismic exploration is that the subsurface consists of a series of elastically homogeneous isotropic layers. Under this assumption, the velocities of elastic waves in one such medium are independent of the direction of propagation and wavefronts of elastic waves are spherical. Also, for a stack of several such layers, the moveout velocity derived from short-offset surface-seismic data is equivalent to the vertical root-mean-square (RMS) velocity (Taner and Koehler, 1969). Using these moveout velocities, we can convert reflection times to depth according to the Dix formulation (Dix, 1955) and then carry out normal-moveout (NMO) or dip-moveout (DMO) correction, seismic data imaging and AVO analysis. However, it has long been recognized that elastic anisotropy is intrinsic to the structure of most rock (Thomsen, 1986). Ignoring these anisotropic effects can adversely influence the results of most basic seismic data processing and interpretation steps, such as NMO correction, velocity analysis, stacking, migration, DMO correction, time-to-depth conversion, and AVO analysis (Banik, 1984; Thomsen, 1986; Alkhalifah and Larner, 1994; Tsvankin, 1995).

There are various methods to estimate anisotropy parameters (Alkhalifah and Tsvankin, 1995; Brown et al., 2000; Elapavuluri and Bancroft, 2002; Gaiser, 1990; Isaac and Lawton, 2004; White et al., 1983). In this paper, we carry out four inversions on synthetic seismic data examples using hyperbolic, shifted hyperbolic, modified three-term Taylor and Alkhalifa's moveout equations and try to determine the relations between the estimated anisotropy parameters and the true anisotropy parameters. Then we estimate anisotropic parameters from Blackfoot seismic data. Finally, we formulate some conclusions for guiding the application of these approximations.

Theory and Method

In order to make qualitative estimates of the influence of anisotropy on seismic reflection-traveltime and to develop inversion algorithms for anisotropic media, it is very important to understand the relationships between the medium parameters and seismic signatures.

The P-wave traveltime approximations for four reflection-traveltime inversion methods are given as follows.

1) The hyperbolic reflection-traveltime approximation:

$$t^2(x) = t_0^2 + \frac{x^2}{V_{\mathrm{NMO}}^2} \ , \tag{1}$$

2) The modified three-term Taylor-series approximations (Tsvankin and Thomsen, 1994) in the limit of weak anisotropy:

$$t^2(x) = t_0^2 + \frac{x^2}{V_{\mathrm{NMO}}^2} + \frac{A_4 x^4}{1 + Ax^2} \ , \tag{2}$$

3) The shifted-hyperbolic approximation (Castle, 1994):

$$t(x) = \tau_s + \sqrt{\tau_x^2 + \frac{x^2}{SV_{\mathrm{NMO}}^2}} \ , \tag{3}$$

4) The nonhyperbolic approximation (Tsvankin and Thomsen, 1994):

$$t^2(x) = t_0^2 + \frac{x^2}{V_{\mathrm{NMO}}^2} - \frac{[V_{\mathrm{h}}^2 - V_{\mathrm{NMO}}^2]x^4}{V_{\mathrm{NMO}}^2[t_0^2 V_{\mathrm{NMO}}^4 + V_{\mathrm{h}}^2 x^2]} \ . \tag{4}$$

From equation (1) to equation(4),

$$V_{\mathrm{NMO}}^2(P) = \alpha_0^2(1 + 2\delta) \, , V_{\mathrm{h}}^2 = \alpha_0^2(1 + 2\varepsilon) \ , \tag{5}$$

$$A = 1/(\alpha_0 t_0)^2 \ , A_4(P) = -\frac{2(\varepsilon - \delta)}{t_0^2 \alpha_0^4} \, , \tag{6}$$

$$\tau_s = t_0(1 - \frac{1}{S}) \, , \tau_x = \frac{t_0}{S} \, , S = 1 + \frac{8(\varepsilon - \delta)}{(1 + 2\delta)} \ . \tag{7}$$

where α_0 is vertical velocity for P waves, β_0 is vertical SV-wave velocity, δ and ε are Thomsen's anisotropy parameter; V_{NMO} is NMO velocity, V_{h} is horizontal velocity for P-waves, t_0 and t are the two-way traveltimes for zero-offset and offset x, respectively, and S is the shift parameter. Xiao et al (2004) have demonstrated that these traveltime approximations have their own ranges of the offset. The ranges of the offset (ratio of offsets and depth) are about 0.5, 1.0, 1.2 and 1.5 for hyperbolic, Tayler, the shifted hyperbolic and nonhyperboic approximations, respectively.

Using equations (1) to (4), we can pick up effective coefficients A_4, V_{NMO}, V_{h} and S, and then obtain anisotropic parameters ε and δ by using equations (5) to (7) through a Dix-type differentiation procedure (here vertical P-wave velocity α_0 is known from well log). Semblance scanning is employed to estimate effective coefficients.

Estimated anisotropy parameters

For simplicity, we consider a series of single-layer case in order to determine how both actual anisotropy parameters and spread length affect the estimation of anisotropy parameters. The input CMP gather for anisotropy-parameter estimation contains a single reflection from a flat interface. The depth of this interface is 500 m. Vertical P- and S-wave velocities above the reflector are 3000 m/s and 1500 m/s, respectively. The values of ε are fixed at 0.2, 0.1 and 0.0, respectively, and those of δ range from –0.2 to 0.2 at increments of 0.02.

Our research shows that under their own offset range, the accuracy of estimated delta decreases with their offsets and the accuracy of estimated epsilon increases with their offsets. Figures 1 shows the errors in estimated δ, plotted versus δ when offset/depth = 1.0, for ε values of 0.2, 0.1 and 0.0. and δ values ranging from –0.2 to 0.2 at increments of 0.02 (Blue solid line: the exact anisotropic parameters; purple dotted line: hyperbolic traveltime inversion; green dash-dot line: the modified three-term Taylor-series inversion; red solid line: the shifted hyperbolic inversion; cyan dashed line: nonhyperbolic inversion). From Figure 1 it appears that i) the smaller the value of $(\varepsilon - \delta)$, the higher the accuracy of the estimated δ value; ii) the estimated values deviate greatly from the true values when $|\varepsilon - \delta| > 0.2$, and iii) the nonhyperbolic and the shifted-hyperbolic estimations are better than the three-term Taylor-series method while hyperbolic

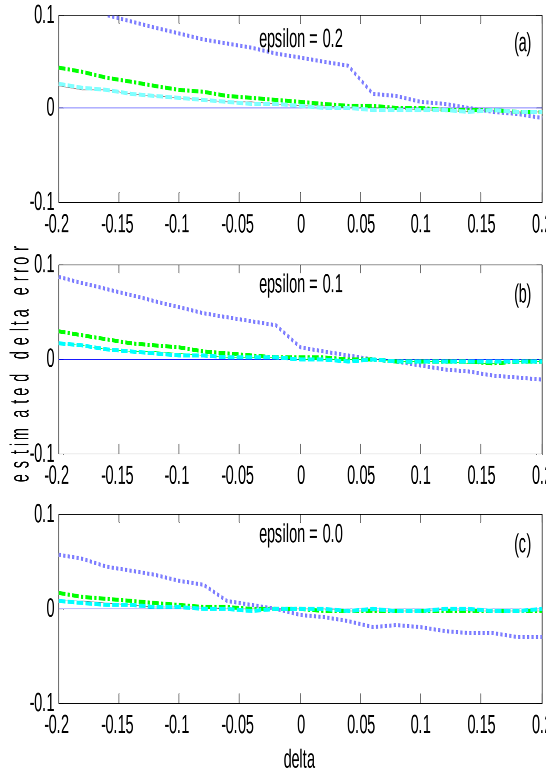

Figure 1. The error in estimated δ plotted vs. true δ when offset/depth = 1.0.

estimation is accurate only for elliptical anisotropy $(\varepsilon = \delta)$. Figures 2 shows the errors in estimated ε, plotted versus δ when offset/depth = 2.0, for ε values of 0.2, 0.1 and 0.0. and δ values ranging from –0.2 to 0.2 at increments of 0.02. We can see that only nonhyperbolic inversion is able to estimate parameters ε with any accuracy.

Table 1 demonstrates the model parameters and estimated anisotropy parameters for a four-layer model. Note that all $(\varepsilon - \delta)$ values in model 1 are less than 0.2. The only difference between model 2 and model 1 is that the value of $(\varepsilon - \delta)$ in the second layer is larger than 0.2. Figure 3 and 4 show estimated anisotropy-parameter values and actual values. These estimation results from multilayer VTI media also demonstrate that the estimated interval anisotropy parameters are very close to the true parameter values. Only when $(\varepsilon - \delta)$ is larger than 0.2 do the estimated interval parameter values depart significantly from the true value.

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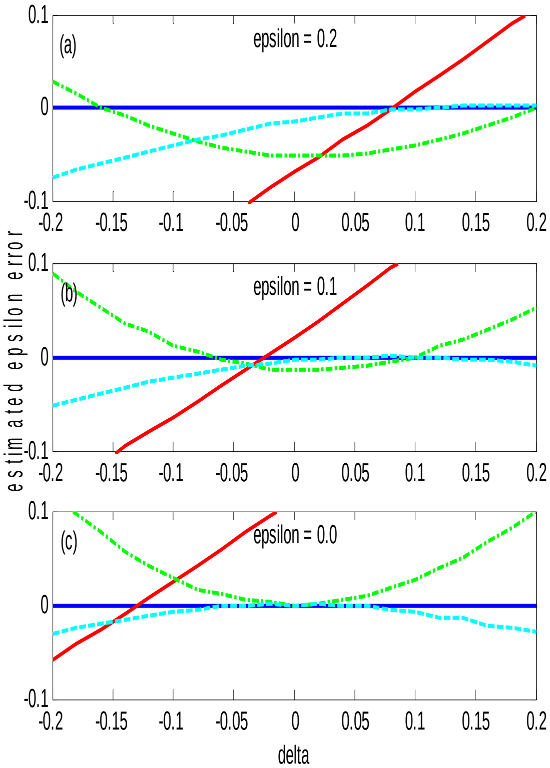


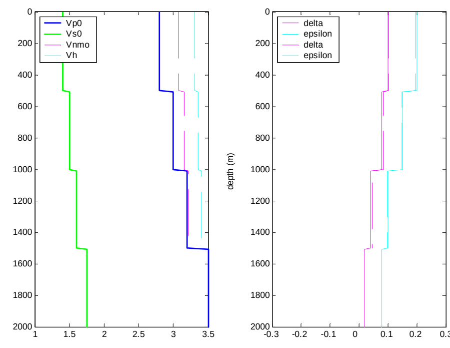
Figure 2. The error in estimated ε plotted vs. true δ when offset/depth = 2.0

Table 1. Model parameters and estimated anisotropic parameters

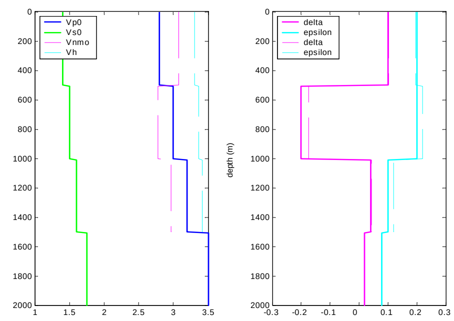
Layer	α_0 (m/s)	β_0 (m/s)	Model 1 ε , δ	Model 2 ε , δ
1	2800	1400	0.20, 0.10	0.20, 0.10
2	3000	1500	0.15, 0.08	0.20, - 0.20
3	3200	1600	0.10, 0.04	0.10, 0.04
4	3500	1750	0.08, 0.02	0.08, 0.02

Application of Blackfoot seismic data

The use of the method above to Blackfoot seismic data shows more challenge. We correlate the synthetic data and field seismic data in order to get vertical velocities. Figure 4 (a) shows the correlation of well logs, synthetic data and Blackfoot seismic data as well as interpreted tops of formations. The four seismic interfaces shown in Figure 4 (b) are choused to estimate Thomsen anisotropy parameters. Figure 5 is the CDP gather after weathering static and residual static for estimating effective coefficients. Table 2 lists the vertical velocities from well data and estimated effective coefficients and anisotropy parameters.



(a)



(b)

Figure3. (a) Model 1 parameters (solid lines) with estimated coefficients (dashed lines); (b) Model 2 parameters (solid lines) with estimated coefficients (dashed lines).

Conclusions

The accuracy of the estimated anisotropic parameter δ depends not only on the accuracy of the picked NMO velocity but also on the value of $(\varepsilon - \delta)$. The smaller the value of $(\varepsilon - \delta)$ and the value of ε , the higher the accuracy of estimated δ. The results of the four traveltime inversions by semblance analysis for the seismic examples demonstrate that the nonhyperbolic and shifted-hyperbolic estimations are better than the three-term Taylor-series method. Only nonhyperbolic inversion can be used to estimate accurately the anisotropy parameter ε . Hyperbolic estimation is only suitable for estimation

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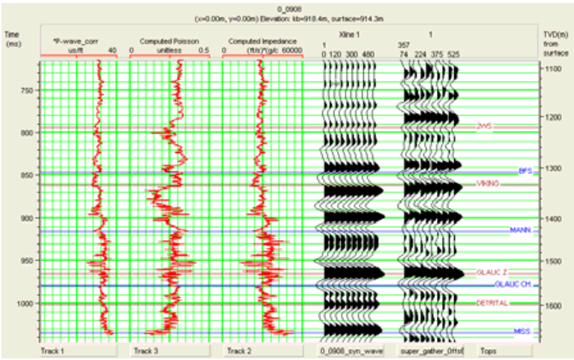
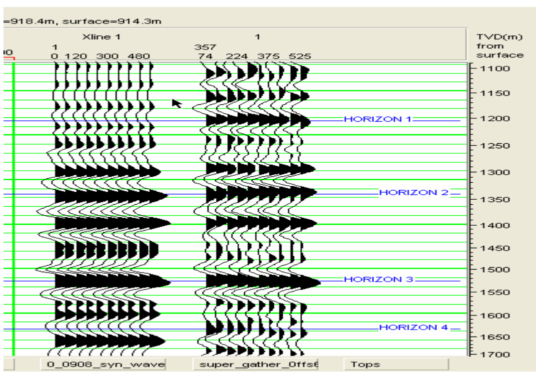


(a)



(b)

Figure4. Correlation of synthetic data and real seismic data (a) with the tops of formation and (b) with the horizons

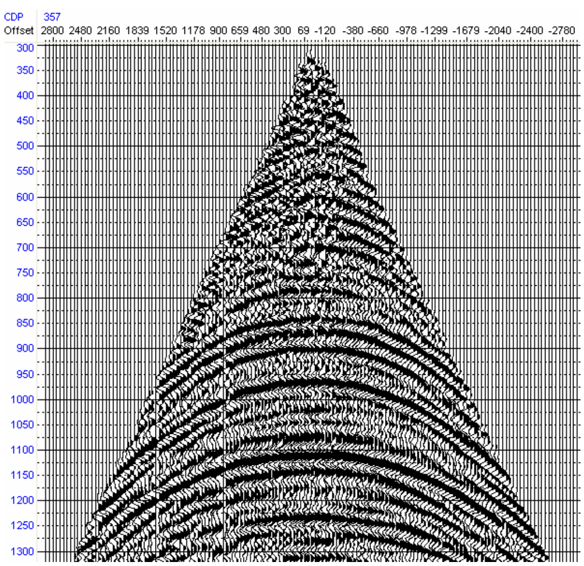


Figure5. The CDP gather for estimating effective coefficients.

Table 2. Estimated effective coefficients and anisotropy parameters.

Layer	α_0 (m/s)	V_{NMO} (m/s)	V_h (m/s)	ε , δ
1	3099	2919	3185	0.0281, -0.0564
2	3299	3266	3257	0.5587, +0.9353
3	3823	3279	3637	0.6902, -0.1063
4	3882	3315	3782	0.3880, -0.0286

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Acknowledgments

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