

Amplitude Migration in Anisotropic Media

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Abstract

The effects of weak anisotropy have been investigated in a two-layer thrust sheet model. The model consists of anisotropic horizontal and dipping interfaces (Figure 1). A ray tracing technique has been used to calculate the aplanatic surfaces for the travel times. Kirchhoff migration is applied to a noiseless zero offset synthetic seismogram which includes anisotropic velocity effects. When an isotropic model is used to migrate the data, the image of the dipping reflector is misplaced significantly. However, with the use of anisotropic migration, the reflector images are correctly recovered.

Introduction

Traditionally seismologists have considered Earth models in which P -wave velocity varies with depth (Gray, 1999; Zhu et al., 1998). In recent years, however, seismic anisotropy has become of growing concern. Seismic anisotropy deals with the directional dependence of the seismic wave velocity and ignoring the effects of this phenomenon may result in the lateral and/or depth misplacement of the subsurface image. In the case of transverse isotropy (TI) all directions perpendicular to the axis of symmetry of the media are equivalent (i.e. the seismic wave velocity is symmetric with respect to the axis that is perpendicular to the surface of the anisotropic layer). Therefore, along the symmetry axis the phase and the group velocity of the seismic wave coincide. In this work we are concerned with the propagation of the P -waves in the Earth's subsurface. Therefore, in any direction θ the velocity v of the P -wave depends on the weak anisotropic parameters δ , ε and α (Thomsen 1986) where α is the P -wave velocity along the axis of symmetry.

The phase angle χ and the group angle θ of a P -wave ray in a weak anisotropic medium are related through

$$\cos(\theta) = \frac{\alpha \sin(\chi) \sin(2\chi) (\delta \cos(2\chi) + 2\varepsilon \sin^2(\chi)) + \frac{\cos(\chi)}{r(\chi)}}{\sqrt{\frac{I}{r(\chi)^2} + [\alpha \sin(2\chi) (-\delta \cos(2\chi) - 2\varepsilon \sin^2(\chi))]^2}} \quad (1)$$

(Slawinski, 1996). The radius of the phase-slowness surface $r(\chi)$ is then given as

$$r(\chi) = \frac{I}{\alpha(1 + \delta \sin^2(\chi) \cos^2(\chi) + \varepsilon \sin^4(\chi))}. \quad (2)$$

In the top layer, given the group angle θ , Π is calculated using (1). The phase and the group velocities $v(\Pi)$ and $v(\theta)$ are then found using

$$\begin{aligned} v_l(\chi) &= \alpha_l(1 + \delta \sin^2(\chi) \cos^2(\chi) + \varepsilon \sin^4(\chi)), \\ v_l(\theta) &= \alpha_l(1 + \delta \sin^2(\theta) \cos^2(\theta) + \varepsilon \sin^4(\theta)). \end{aligned} \quad (3)$$

In the equation (3) Fermat's Principle which states that for a weakly anisotropic medium $v(\theta) \equiv v(\chi)$ is used. To find the P -wave velocities in the bottom layer we proceed as follows. The slowness p in the top layer is given as $p = \sin(\chi_i) / v_l(\chi_i)$. In the bottom layer we can solve for the phase angle Π_i using:

$$\alpha_2 p (\varepsilon_2 - \delta_2) \sin^4(\chi_i) + \alpha_2 p \delta \sin^2(\chi_i) - \sin(\chi_i) + \alpha_2 p = 0 \quad (4)$$

(Slawinski, 1996) where α_2 is the P -wave velocity along the symmetry axis in the bottom layer.

The transmitted angle θ_i (the group angle for the ray propagating in the bottom layer) is then found using (1). The phase and group velocities are then given by (3) using the anisotropic parameters for the bottom layer.

Results

$$x = \sin(120 \pi t) e^{-120 \pi t^{3/2}} \quad (14)$$

In this section results are reported for a zero offset shot-records. The source-receiver pairs are placed 40 meters apart. Figure 2 shows an example of a first arrival travel time map (aplanatic surfaces) for the anisotropic model. For the shot-records a source wavelet of the form is used. A synthetic seismograph of the above is shown in Figure 3. The amplitudes are migrated, using a Kirchhoff method, and then stacked for the anisotropic model. Figure 4 shows the contour map of these amplitudes. Figure 4 shows that the reflector images are positioned correctly.

Next we set $\delta_1=\epsilon_1=\delta_2=\epsilon_2=0$ so that the model is isotropic. The amplitudes in Figure 3 are migrated, using this model, and then stacked. Figure 5 shows the contour map of these amplitudes. In Figure 5 the dipping angle of the dipping reflector has decreased slightly and x_s , the point of intersection of the horizontal and dipping top reflector, has moved to the left by about 50 meters, which would be roughly equal to the wavelength for much seismic data. The horizontal reflectors are not affected as the rays from a source to the horizontal reflector and back to the corresponding receiver must travel along the symmetry axis and along this axis the P -wave velocities are the same for the isotropic and anisotropic model.

Conclusion

The aplanatic surfaces are calculated for a two layer anisotropic model. Using zero offset source-receivers a synthetic seismograph is calculated. The amplitudes are then migrated and stacked for an anisotropic and an isotropic model. It is shown that the dipping reflectors are misplaced if a correct model is not used.

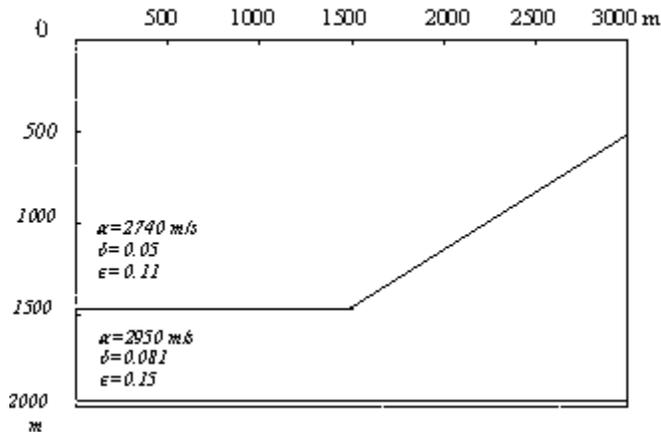


Figure 1. The two layer model used in this work.

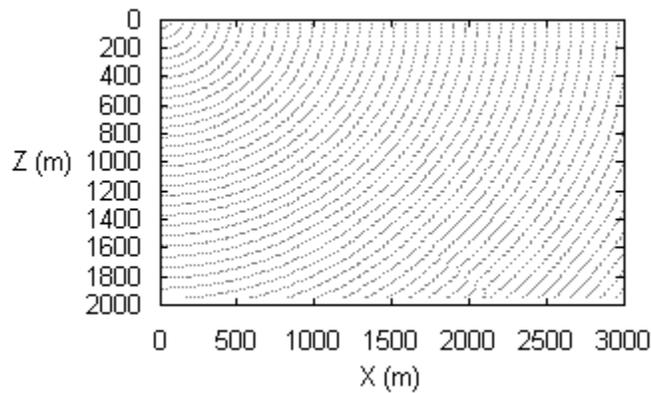


Figure 2. Example of travel-time map for the anisotropic model.

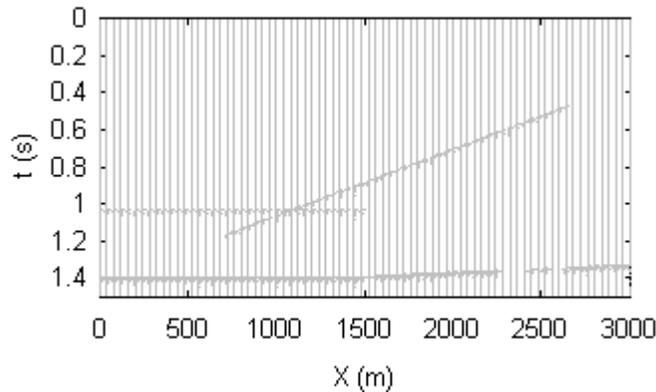


Figure 3. Noise free synthetic seismograph.

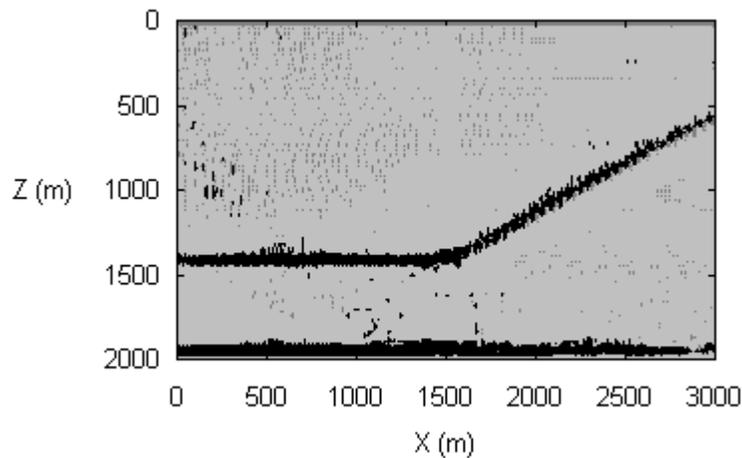


Figure 4. Contour plot of the amplitude migration for the anisotropic model; the dipping and horizontal layers meet at $x=1500$ m.

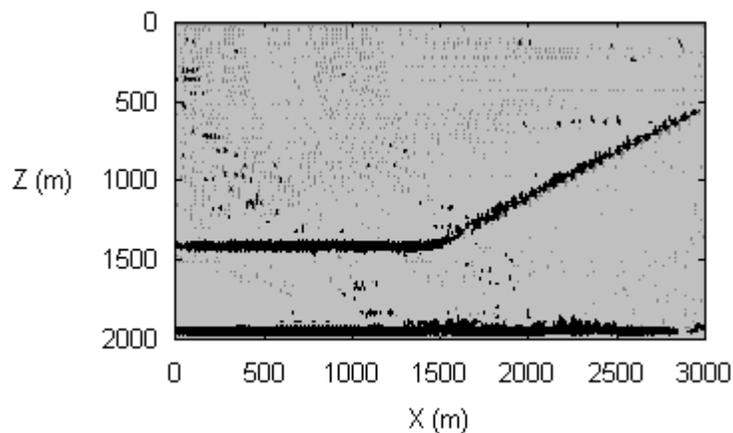


Figure 5. Contour plot of the amplitude migration for the isotropic model; the dipping and horizontal layers meet at about $x=1450$ m.

References

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