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Pseudo 2D elastic waveform inversion for attenuation in the near surface

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ABSTRACT

Seismic waveform propagation could be significantly affected by heterogeneities in the near surface zone (0 m–500 m depth). As a result, it is important to obtain as much near surface information as possible. Seismic attenuation, characterized by Q_P and Q_S factors, may affect seismic waveform in both phase and amplitude; however, it is rarely estimated and applied to the near surface zone for seismic data processing. Applying a 1D elastic full waveform modelling program, we demonstrate that such effects cannot be overlooked in the waveform computation if the value of the Q factor is lower than approximately 100. Further, we develop a pseudo 2D elastic waveform inversion method in the common midpoint (CMP) domain that jointly inverts early arrivals for Q_P and surface waves for Q_S. In this method, although the forward problem is in 1D, by applying 2D model regularization, we obtain 2D Q_P and Q_S models through simultaneous inversion. A cross-gradient constraint between the Q_P and Qs models is applied to ensure structural consistency of the 2D inversion results. We present synthetic examples and a real case study from an oil field in China.

1. Introduction

As a seismic wave travels through the earth, its energy is partially converted into heat due to anelasticity and heterogeneity of the earth. The loss of energy corresponds to attenuation and dispersion (Futterman, 1962). An accurate Q structure is required to both simulate a seismic wavefield and perform any seismic analysis or imaging method that utilizes waveform data, such as amplitude variation with offset (AVO) (Carcione et al., 1998; Chapman et al., 2008) and full waveform inversion (FWI) (Tarantola, 1988; Liao and McMechan, 1996; Pratt, 1999; Hicks and Pratt, 2001; Kamei and Pratt, 2008; Virieux and Operto, 2009; Malinowski et al., 2011; Kurzmann et al., 2013; Groos et al., 2014; Xue et al., 2016). Q estimation is also a potential diagnostic tool for reservoir characterization and hydrocarbon detection (Xie et al., 2009) as the rock attenuation property is more sensitive to the change in rock conditions than the seismic velocity (Toksoz et al., 1979; Winkler and Nur, 1979). Currently, the Q factor is routinely estimated and applied for imaging deep earth structures (Bennington et al., 2008). However, near-surface seismic imaging can be very challenging because of the heterogeneity of the near surface, such as rugged topography, large velocity variations, or hidden low-velocity layers (Zhang, 2009). We assume that for the near surface modelling problem, it is not necessary to resolve a high-resolution Q structure and that smooth Q_P and Q_S models should be sufficient. Technologies for Q inversion have been previously developed. For example, in the time domain, the Q factor is usually estimated by the pulse amplitude decay (Brzostowski and McMechan, 1992), pulse rising time (Kjartansson, 1979), or pulse broadening method (Wright and Hoy, 1981). In the frequency domain, the Q factor can be estimated using a spectral ratio method (Sams and Goldberg, 1990) or a centroid frequency-shift method (Quan and Harris, 1997). Zhu and Harris (2015) developed a method to exploit crosswell traveltime data to estimate the Q factor. As the seismic waveforms are substantially affected by attenuation and waveform inversion has been a widespread approach for constructing high-resolution subsurface images (Tarantola, 1984, 1988; Sambridge et al., 1991; Hicks and Pratt, 2001), we conduct full waveform inversion for Q. Ribodetti and Virieux (1998) developed a fast inversion technique based on both the Born approximation and asymptotic Green's functions for recovering elastic and attenuation parameters. Ribodetti et al. (2000) derived the formulae for an asymptotic viscoacoustic diffraction tomography and applied this method to ultrasonic laboratory data to estimate the velocity and attenuation factor Q. Smithyman et al. (2009) applied acoustic 2D waveform inversion to obtain the P-wave velocity and attenuation. To mitigate the problem of cycle-skipping, a windowed-amplitude waveform inversion method (Pérez Solano et al., 2014) and ω-p and ω-k domains inversion method (Brossier et al., 2014) for surface wave analysis are developed. Bai et al. (2014) performed visco-acoustic full waveform inversion on synthetic data and field data in the time domain to determine the velocity and attenuation. Krohn and Routh (2017a, 2017b) developed the surface-wave impulse estimation and removal (SWIPER) method to estimate the velocity dispersion and attenuation of surface waves in the frequency domain and applied this method to 2D and 3D datasets.

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Although performing 2D or 3D elastic waveform inversion for attenuation is an accurate approach, it is time consuming (Zhou et al., 1993, 1995; Pratt and Shipp, 1999). Note that it is also feasible to obtain 2D or 3D images by interpolating multiple 1D profiles. Bickel and Natarajan (1985) and Wang (2002) developed a 1D inversion method to determine the 1D Q structure and applied the Q filter to seismic traces. Xia et al. (2012) estimated 1D near-surface quality factors by performing constrained inversion of the Rayleigh-wave attenuation coefficients. Yang et al. (2009) employed surface seismic and VSP data jointly to perform 1D viscoelastic waveform inversion for Q factors. However, the near-surface area is complex, and thus, the 1D model cannot illustrate the accurate structure. Auken and Christiansen (2004) developed a pseudo 2D inversion for resistivity data by applying a lateral constraint. The lateral constraint inversion (LCI) method has been applied to surface wave imaging (Socco et al., 2009; Boiero and Socco, 2010). We develop a method to produce a 2D model based on waveform inversion. In this study, we apply a 1D elastic waveform forward modelling method and conduct joint 2D Q_P and Q_S model inversion by applying 2D Tikhonov regularization (Tikhonov and Arsenin, 1977) to the model parameters. The regularization term strengthens the relation between the target grid and its four neighbouring grids. This technique is extremely efficient, even when it is extended to 3D. This pseudo 2D inversion method should be applicable to small or moderate lateral variations in Q models, similar to the common midpoint (CMP) assumption (Mayne, 1962, 1967). When the wave propagates in a 1D medium, the waveform can be used in the inversion process to determine the structure beneath the midpoint between the source and receiver. If the lateral variations of the velocities and Q are significant, then the CMP assumption in the pseudo 2D inversion method may fail. In this approach, we assume that the velocity models are sufficiently accurate and we only invert for Q factors. Elastic full waveform data include a variety of wave types: P waves, converted waves, and surface waves. The early arrivals primarily contain information of P waves, and the surface waves primarily contain information of S waves. Therefore, we apply the early arrival waveform to invert Q_P and the surface wave to invert Q_S and then establish a joint inversion scheme by minimizing the cross gradients (Gallardo and Meju, 2003, 2004) between the Q_P and Q_S models so that both Q_P and Q_S are associated with each other in 2D structures. The application of cross gradients improves the structural similarity of models.

In the following, we first present the effect of the Q_P and Q_S factors on waveforms. We test a three-layered model with different Q_P and Q_S structures and find that early arrivals are sensitive to the Q_P factor and surface waves are sensitive to the Q_S factor. Combining these two wave modes to perform inversion for attenuation is feasible for near-surface imaging. Next, we describe the objective function of this joint inversion method. The technique employs multiple traces at different locations and forms a pseudo 2D structural inversion problem by applying 2D Tikhonov regularization. Next, we apply this method to test a synthetic model and discuss the effects of density error, P-wave velocity error, and S-wave velocity error. Finally, we apply this method to test a real dataset and obtain reasonable results.

2. Method

2.1. Forward problem

Seismic waveforms contain information regarding the subsurface, such as the velocity, density, and attenuation. The attenuation factor Q can be assumed to be independent of frequency in the interior of the solid parts of the earth (Knopoff, 1964). In wave propagation, the elastic velocity can be written as (Aki and Richards, 1980):

$$v(\omega) = v_1\left[1+\frac{1}{\pi Q}\ \ln\left(\frac{\omega}{2\pi}\right)-\frac{i}{2Q}\right], \tag{1}$$

where v_1 is the velocity of an elastic wave at a reference frequency and $v(\omega)$ is the velocity of an elastic wave at frequency ω. When v is the P-wave velocity, the attenuation factor Q represents Q_P. When v is the S-wave velocity, the Q represents Q_S. We assume that Q is a constant for the frequency band from 0.2 Hz to 100 Hz (Liu et al., 1976). The nonlinear forward problem can be expressed as:

$$\mathbf{d} = G\,(\mathbf{m}), \tag{2}$$

where $\mathbf{d}$ represents the seismic data vector and $\mathbf{m}$ represents the model parameter vector:

$$\mathbf{m} = (Q_1, Q_2, \ldots, Q_k, Q_{k+1}, Q_{k+2}, \ldots, Q_{2k})^T, \tag{3}$$

where the range from Q_1 to Q_k represents the Q_P values; the range from Q_{k+1} to Q_{2k} represents the Q_S values. G is the nonlinear forward function. Next, we design a three-layered model to test the effects of Q_P and Q_S on waveforms. The thicknesses of the first layer and second layer are both 300 m. The third layer is a homogeneous half-space. The P-wave velocity values, S-wave velocity values, and density values are shown in Table 1. The Q_P and Q_S structures are homogeneous. For the near-surface area, due to the existence of the weathered layer, the Q values might be much less than 100. We test three Q structures as shown in Fig. 1. The Q_P values are set to 50, 100, and 200. The Q_S values are set to 25, 50, and 100. The source is buried 10 m beneath the surface, and the offset is 1.0 km. We take a Ricker wavelet as the source, and the central frequency for forward modelling is 10 Hz. We apply the discrete wavenumber method (Bouchon and Aki, 1977) to calculate the waveforms. The amplitudes of the simulated waveforms are normalized for comparison. Fig. 1(a) shows a comparison of the early arrivals. Fig. 1(b) shows a comparison of the surface waves. We observe significant differences in traveltimes, amplitudes, and the phases of surface waves. Early arrivals are notably affected by a lower Q value. Q affects waveforms both in amplitudes and phases, especially if the Q values are lower than 50.0. For example, in an alluvial plain, silt is deposited in the near-surface area and has a loose physical structure and low quality factor, which lead to the substantial attenuation of seismic wave energy (Cui et al., 2013). We focus on the Q values in the near-surface area, where Q significantly affects waveforms.

2.2. Inverse problem

Commonly, in seismic waveform inversion research, one applies 1D forward modelling to invert 1D structures and applies 2D forward modelling to invert 2D structures. However, for large production data, 2D or 3D elastic full waveform inversion might be much more time-consuming than 1D inversion. As a result, we develop a method that applies 1D forward modelling to invert the 2D structure; this method is called pseudo 2D elastic waveform inversion for Q factors. This method may provide an initial model for true 2D elastic full waveform inversion. For each location, we perform 1D forward modelling, and the received data include information regarding the subsurface medium. Along the 2D survey, multiple traces over the surface range can be processed simultaneously, and through applying 2D Tikhonov regularization to the model parameters, the method outputs 2D models. The application of Tikhonov regularization strengthens the relationship between one grid and its four neighbouring grids.

Table 1
The model parameters include the thickness, P-wave velocity, S-wave velocity, and density of each layer.

Layer	Thickness (m)	V_P (m/s)	V_S (m/s)	Density (g/cm^3)
1	300.0	1800.0	900.0	2.019
2	300.0	2200.0	1100.0	2.123
3	∞	2400.0	1200.0	2.170