Velocity model estimation by full waveform inversion of time-lapse 4D passive seismic array data

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SUMMARY

In passive-source monitoring, an accurate velocity model is important to precisely estimate microseismic source locations, and to understand changes in reservoir properties. In this study, we employ frequency-domain full waveform inversion in order to obtain a high-resolution velocity model by exploiting full wavefields. We demonstrate the feasibility of the method for a surface geophone array by inverting for time-lapse 4D velocity changes in a realistic subsurface model. Our method successfully estimates the small velocity changes of a few percent within layers of 10s of meters, even for a single passive seismic source event. The analysis of wavepaths and gradients suggests that keys for the successful inversion are the use of full wavefields (both first and scattered arrivals), the vicinity of velocity changes to the source, and the wide-aperture surface array.

Key words: passive seismic, monitoring, full waveform inversion, velocity estimation.

INTRODUCTION

Passive-source monitoring of natural or induced seismicity in the subsurface traditionally focuses on determining the event location by the triangulation of picked event arrival times, or more recently by imaging techniques such as beam forming, and reverse time imaging (e.g. McMechan, 1982; Gajewski and Tessmer 2005, Lumley and Shragge, 2013). Any of these methods requires P- and/or S-wave velocity models. A 1D velocity model derived from sonic logs is currently used for the majority of field applications, but 2D/3D velocity models are required to improve the accuracy of the source estimation. Furthermore changes in the velocity model over time provide vital information about reservoir fluid flow and other time-variant subsurface phenomena.

We propose to adopt a method which utilizes full passive seismic wavefields in order to estimate a velocity model. The full waveform approach allows us to increase the spatial resolution, and also detect weak signals (Lumley and Shragge, 2013). In this study, we employ a full waveform inversion method (FWI), which explores an optimum subsurface model by fitting waveforms in the data domain (Tarantola, 1984). The method is emerging as a powerful high-resolution subsurface model building technique for active seismic data (e.g. Virieux and Operto, 2009, Kamei et al. 2012) and for regional and global earthquake data (e.g. Tape et al. 2009).

In this study, we demonstrate velocity inversion for a dense surface array deployment using a realistic subsurface model. We evaluate the feasibility of 4D FWI for passive seismic data by inverting for time-lapse velocity changes from a single microseismic event using several array configurations.

METHOD

In this study, we develop an FWI algorithm based on the P-wave scalar wave equation in the frequency domain (Pratt 1999),

\[
\left[ \nabla^2 + \frac{\omega^2}{v^2(x)} \right] P(x, \omega) = s(x, \omega),
\]

(1)

where \( v(x) \) is the velocity model, \( P(x, \omega) \) is the pressure wavefield at the angular frequency \( \omega \), \( s(x, \omega) \) is the source wavefield. The wave equations (1) can be written in compact symbolic notation as

\[
d = Fm.
\]

(2)

The adjoint operator \( F^* \) can be defined by

\[
\langle Fm, d \rangle = \langle m, F^*d \rangle,
\]

(3)

where \( \langle f, g \rangle \) is defined as the inner product between two quantities \( f \) and \( g \). When \( F \) is defined as the forward wave propagation modelling operator, \( F^* \) is interpreted as the reverse-time adjoint wave propagation operator that maps from the data space back to the model space:

\[
F^*d = u_r,
\]

(4)

where \( u_r \) is the reverse time receiver wavefield back propagated into the subsurface.

Now we consider solving equation (2) for the velocity model \( m \). It is generally impossible to directly solve for an inverse operator \( F^{-1} \), such that \( F^{-1}F = I \). Instead a practical approach is to perform a non-linear gradient-based inversion for \( m \) by minimizing an objective function that specifies a data misfit norm and constraints (Tarantola, 1984): when the L2 data misfit norm is employed,

\[
E = ||d - Fm||^2,
\]

(5)

where \( E \) is the objective function. By applying the adjoint method (Tarantola 1984), we form a gradient as
The source location is assumed known, but we estimate the source wavelet with the method described in Pratt (1999).

We first evaluate the proposed FWI method using a dense wide-aperture geophone array. A total of 180 geophones are located at 100-m interval over the entire model range. FWI converges well reducing the objective function to below 1% of that of the starting model. The inverted velocity change is displayed in Figure 3a. The two distinct velocity layers are very well delineated, though the right edges of the anomalies are not clearly delineated, due to the limited illumination geometry. We compute the acoustic synthetic wavefields from the starting and inverted velocity models, and display the residual errors in Figure 2b-c. The excellent reproduction of the observed wavefield after FWI confirms the accuracy of the inverted model.

In order to evaluate the effects of the density of the geophone array, we also conduct the inversion using 90 geophones with 500-m interval, and 18 geophones with 1-km intervals, and show the inverted velocity changes in Figure 3b-c. In both cases, the inversion converged well. As the geophone interval increases, the artifacts due to spatial aliasing contaminate the velocity model, and the estimated velocity changes become blur. However even with 1-km interval, the two velocity layers are still clearly recognizable.

Finally, we limit the range of the aperture of the array, by locating the geophones only directly above the reservoir at 4 - 10 km. Instead we increase the total number of geophones to 1920, reducing the geophone interval to 3.125 m. The estimated velocity change after FWI is displayed in Figure 3d. While the inversion still converges, the results are significantly degraded from the previous examples. The top layer is only partially retrieved, and the bottom layer is not imaged.

The results suggest that velocity changes may be recovered from a single event if a sufficiently wide aperture array is deployed. We now analyze gradients from a single trace and from entire traces to provide insights into the behavior of FWI. The gradient from a single trace illustrates the part of the model to which the trace is sensitive to, and is an approximate to a wavepath (Woodward, 1992) and a sensitivity kernel (Tape et al. 2009). We refer to the single-trace gradient as a wavepath to distinguish a gradient formed by all available sources. We display the wavepaths and the gradients from the first iteration in Figure 4 and 5.

As discussed in Woodward (1992), the wavepaths in Figure 4 demonstrate the benefit of using full wavefields: the centre of the wavepath corresponds to the (first-order) Fresnel zone sensitive to the first arrival, and the surrounding high-wavenumber ripples are the part sensitive to the scattered later arrivals. The combination of these two parts contributes to increase the spatial resolution and illumination. We also observe a wavepath from a different geophone illuminates a different part of the model near the surface, but overlaps each other near the source (Figure 4). This allows to constructively form a gradient near the source, although the gradient suffers significantly from the spatial aliasing near the surface (Figure 5b-c). Furthermore, as observed from Figure 4c, wide-angle traces illuminate the reservoir very well thanks to the horizontal layering structure, even when the geophone is apparently too far from the reservoir. This additional angular coverage is important to resolve fine-scale features as indicated in Figure 6d.

**RESULT**

We use a realistic reservoir model of P-wave velocity model displayed in Figure 1a. The reservoir is at ~4 km depth and is subjected to fluid injection and withdrawal as in the case of hydrocarbon production or CO2 sequestration. We consider a hypothetical case where fluid injection and production has changed pore pressure, leading to a small velocity increase in two distinct layers within the reservoir as displayed in Figure 1b. The thickness of each layer is approximately 75 m and 50 m with velocity increase of 85 m/s and 40 m/s. These changes have created a hypothetical microseismic fault reactivation event at 5 km depth and a horizontal distance of 5.5 km. The micro-earthquake is recorded at a geophone array at the surface (Figure 2a).

In this test, we use synthetic acoustic wavefields, and estimate a velocity model after the injection by fitting the waveforms between 2-15 Hz. We precondition the gradient by weak wavenumber filtering to eliminate undesired high-wavenumber oscillations, and by masking outside a rectangular area of 12.5 x 4.5 km² around the reservoir to accelerate the convergence. We assume we accurately know the baseline velocity model before the injection, and use the velocity model in Figure 1a as our starting model for FWI.

![Figure 1](image-url)
CONCLUSIONS

We demonstrate that a time-lapse waveform inversion method successfully estimates a small 4D reservoir velocity change from a microseismic event, with surprising accuracy. We consider such fine-scale small velocity changes are successfully reproduced i) due to the exploitation of both first (transmitted) and later (scattered) arrival waveforms, ii) due to the short distances between the velocity changes and the source location, and iii) due to the wide-aperture recording. We are currently investigating the effects of noise, and velocity error on these techniques.

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REFERENCES


Figure 4: Wavepath (gradient from a single trace) of a geophone located at a horizontal distance at (top) 1 km, (middle) 8 km and (bottom) 15 km.

Figure 5: Gradient at the first iteration using a geophone array located at 0-18 km with (a) 100-m interval, (b) 500-m interval, (c) 100-m interval, and using (d) a geophone array at 4-10 km with 3.125-m interval.