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Logarithmic Phase-only waveform inversion on two different shallow marine geologic settings

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Summary

Two-dimensional frequency domain acoustic waveform tomography is applied to two different marine seismic reflection sections contaminated with strong guided waves which are highly dispersive in nature. We show that, it is possible to obtain a reasonable P-wave velocity model for both the sections. For the first section where the basement is deep, we describe the problem in terms of the average amplitude variation of the first arriving wave train of both observed and synthetic data with offset bins (AVO), and present a strategy to recover a subsurface velocity model that reproduces the observed field data. And for the second section where the basement is shallow, we design a strategy for inversion, by careful preconditioning of the observed data and choosing appropriate parameters and boundary conditions for the forward modeling.

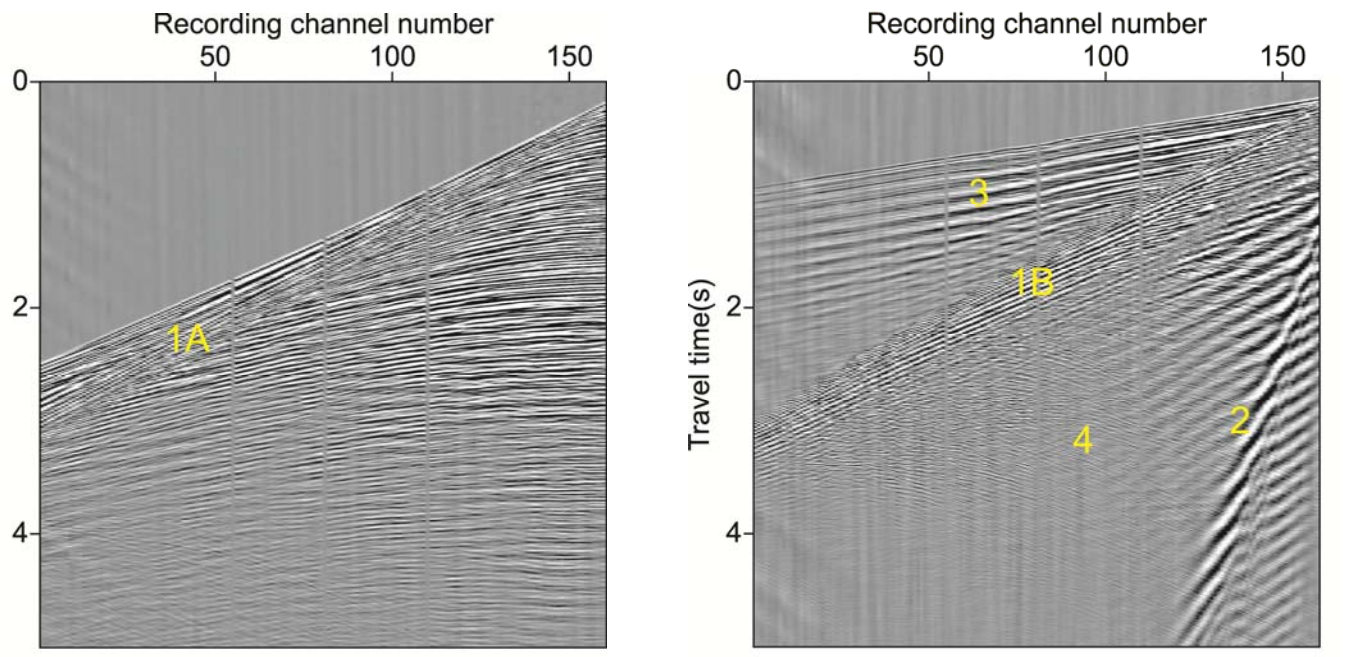
Keywords: Guided waves, Scholte waves, Travel-time tomography, Waveform tomography

Introduction

In this study, we used a multi-channel seismic reflection profile (line 1242) from south western alaska offshore, which were acquired as part of the project EW9409 Pacific to Bering Sea deep seismic experiment. The seismic section was acquired with a maximum offset of 4230m and near offset of 255m. The shot interval is approximately 50m and receiver interval is 25m. At this location, the water bottom is shallow, i.e. less than 100 m. The streamer data showed distinct difference in respect to the observed wave types at each location. At the southern end of the seismic line i.e section A, where a thick pile of sediments overlaid on the deep basement, acoustic guided waves were measured with the similar velocity as first arrivals shown in Fig.1a. But towards the northern end of the seismic line i.e section B, where a hard basement is found at shallow depth, Scholte interface waves in addition to acoustic guided waves were measured as shown in Fig.1b.

These acoustic guided waves are reinforcement fronts formed by multiple reflections of sound in water with some penetration into the underlain sediment (Burg et al., 1951). These waves are sensitive to both shear and compressional wave velocity variation with depth where as Scholte waves are interface waves, which are highly sensitive to shear wave velocity variations with depth propagate at distinct layer interfaces and are characterized by high amplitude, low velocity and frequency (Klein et al., 2005). The dispersive nature of acoustic guided and Scholte waves poses a serious problem in high resolution acoustic waveform tomography imaging of the subsurface because it jeopardizes the identification of desired refractions i.e. first arrivals which is a key component in the success of acoustic waveform tomography.

Velocity discrimination method such as frequency-wavenumber (F-K) domain filtering is of little use, because these guided waves are highly dispersive in nature. Since these guided waves are coupled with Scholte waves which are shear in nature, they cannot be modeled in acoustic waveform tomography.



(a) (b)

Fig.1 : a) Section A: Raw shot gather from Deep basement b) Section B : Raw shot gather from Shallow basement settings. Various seismic events are numbered from 1-4 on the raw shot gathers.1A-not so strong guided wave, 1B-strong dispersive guided wave coupled with Scholte waves, 2- strong dispersive Scholte waves 3- strong refraction multiples, 4-point scatterers.

Preconditioning of data:

For the seismic data from both sections , same data preconditioning scheme was adapted to ensure the data input to waveform tomography are as consistent as possible with the (acoustic) assumptions of the method. This implies that any aspect of the data that is not predicted by the 2D acoustic propagation scheme, e.g. shear waves, coherent noise, shot to shot energy variations, amplitude discrepancy, and bad traces should be removed or corrected (Takam Takougang and Calvert, 2011).

We first resampled the data to 4ms and killed some of the bad traces. Since method assumes 2D wave propagation, the 3D seismic data are corrected for geometrical spreading, i.e. the data were multiplied by t^0.5. Shot gathers were then subjected to amplitude balancing to avoid any shot to shot variations which can bias the model update during the inversion. The data were then subjected to sharp low pass filtering with minimum phase using corner frequencies (0-0-11-13Hz) of an Ormsby filter with 0.25% prewhitening. Time windowing is applied to the shot gathers to exclude late arrivals, multiples and include first arrivals, direct and refracted energy and early secondary arrivals. The seismic data were finally arranged in reduced time to ensure the inclusion of all arrivals during inversion, and reduce the computational cost. A scaling factor was applied to the amplitudes of processed field data as described by Brenders and Pratt (2007) to minimize effects arising from complex geometrical spreading, 3D scattering effects etc.

Starting velocity model

We used first arrival tomography code of Aldridge and Oldenburg (1993), which is based on a finite difference solution of the eikonal equation to derive a starting velocity model for waveform tomography from picked first arrivals for both the sections. The generated starting velocity models from both sections as shown in Fig.2a and Fig.3a satisfies convergence criterion, i.e. both models able to predict the first arrival to within half a cycle (Sirgue 2003). Inorder to avoid numerical artefacts in the constructed model during waveform inversion, the finite difference cell size was chosen to approximately satisfy

$$\Delta_s \leq \frac{\lambda_{min}}{4} = \frac{v_{min}}{4 f_{max}}$$

Where λ_{min} is the minimun wavelength, v_{min} is the minimum velocity, and the f_{max} is maximum frequency used in the modelling and Δ_s is spatial sampling interval. In the present case the minimum velocity was chosen to be that of water velocity i.e 1480m/s and maximum frequency as 12Hz, so the corresponding Δ_s would be approximately equal to 30m. However, we used grid cell size of 25 m which satisfies at all frequencies for both sections.

Inversion:

We know that full-waveform inversion is often nonlinear and ill-posed. In addition to it, the presence of strong dispersive guided wave noise in the data even complicates the problem of nonlinearity. So, we designed a careful strategy to recover a subsurface velocity model that reproduces the observed field data. For both sections, we applied Laplace-Fourier domain logarithmic phase only waveform inversion described by (Kamei et al., 2014). The main advantage of this approach is that, it will utilizes only phase of the residuals which are less sensitive to the noise present in the data and provide deeper gradient illumination.

Deep Basement:

In this case, since the first arrivals are contaminated with strong dispersive guided waves as shown in Fig.1a, we used a very low frequency sampling interval i.e 0.16s.We followed a layer stripping approach i.e we first used a time damping term of τ= 0.2s for a 2s data window and inverted the preconditioned waveforms using frequencies from 5 to 10 Hz in sets of three frequencies. After that, tau value is increased to 0.4 s and inverted for all the frequencies again. And then the data window is increased to 3s and tau value is set to 0.6s and inverted for all the frequencies. The final velocity model obtained after performing logarithmic phase only inversion for final set of frequencies (9, 9.5, 10 Hz) for τ= 0.6s is shown in Fig. 2b.

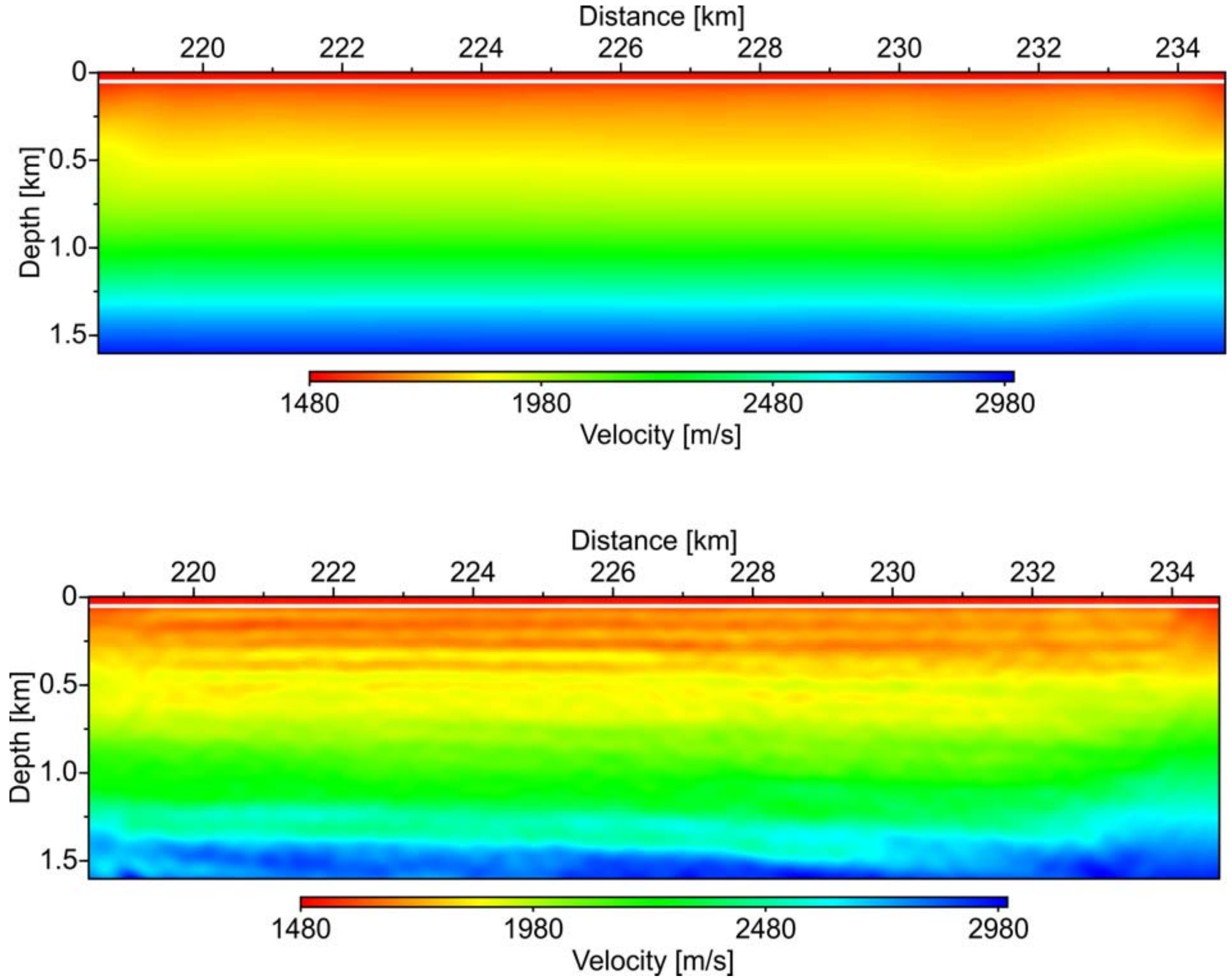
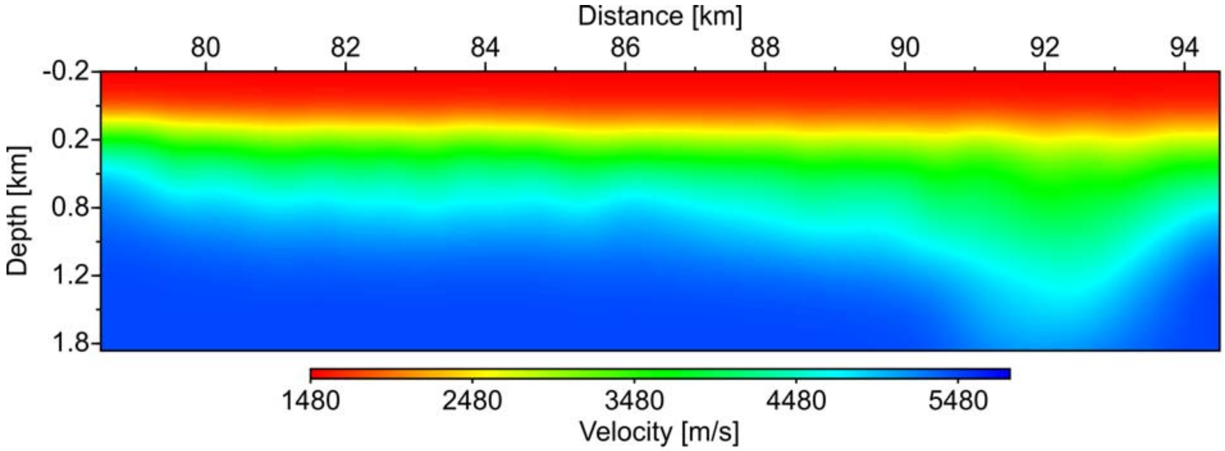


Fig.2: Section A (deep basement): (a) Starting velocity model derived from Travel time tomography (b) Final velocity model obtained after inverting final set of frequencies (9, 9.5,10) Hz .White line indicating the water depth.

Shallow Basement:

For section B, we developed a methodology that coped with a starting model for the inversion from travel-time tomography that itself generated guided waves from the shallow water layer in the synthetic data.We used starting velocity model with absorbing boundary condition which satisfies half cycle convergence criterion. In this case, we have chosen a frequency sampling interval of 0.5s. We used a time damping term of τ= 0.2s and inverted the preconditioned waveforms using frequencies from 5 to 10 Hz in sets of three frequencies After that, tau value is increased to 0.4 s and inverted for all the frequencies again. The final velocity model obtained after performing logarithmic phase only inversion for frequencies 5-10Hz for τ= 0.4s is shown in Fig. 3b. We found anomalously low zones in the final velocity model produced by waveform inversion. These anomalies may well be the expression of sub-vertical strike-slip faults that cannot be seen in the conventional reflection section due to strong water layer multiples and guided waves.



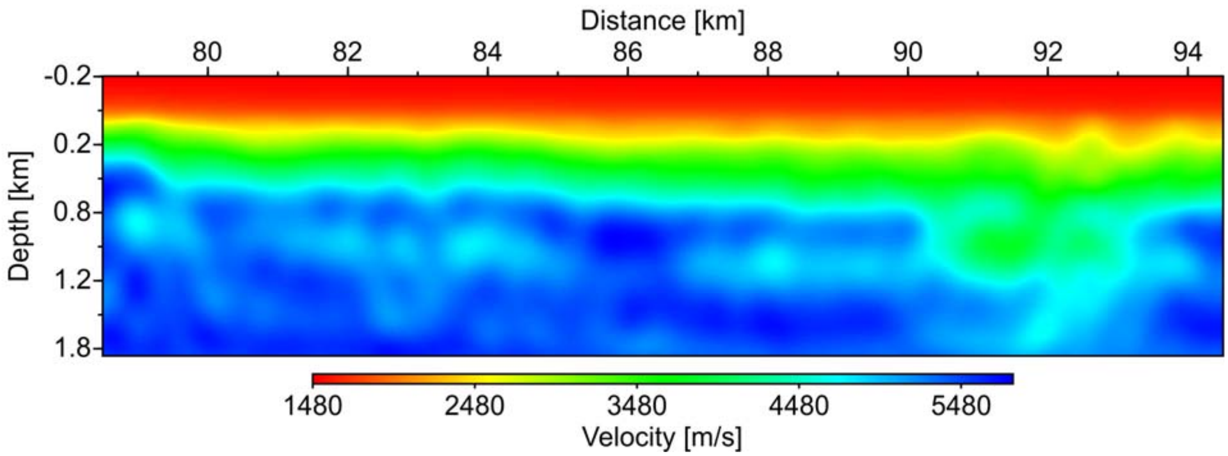


Fig.3: Section B (Shallow basement): (a) Starting velocity model derived from Travel time tomography (b) Final velocity model obtained after inverting final set of frequencies (9, 9.5,10) Hz .

Conclusions

We successfully applied acoustic waveform tomography to two different geologic settings. one setting where deep basement is overlaid by sediments and in second setting where the basement is shallow. We conclude here that, it is possible to obtain reasonable p-wave velocity model by following a careful methodology. In order to evaluate the waveform inversion results obtained above, we generated synthetic waveforms from final velocity model by using acoustic finite-difference code and compared the results with observed waveforms. The synthetic offset gather waveforms seems to be in good match with the observed offset gather waveforms at most of the places. We also examined the frequency domain logarithmic phase only residuals at 6Hz for both starting velocity model and final velocity model from waveform tomography for both sections. We observed there is significant reduction in the residuals in most of the source-receiver pairs.

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