Deep Ocean 3D tomography on field data

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Summary

We have run a suite of 2D and 3D wavefield inversions to recover fine-scale velocity structure in upper oceanic crust. The data were acquired on a plateau, close to a transform fault in deep water in the Pacific ocean. At the fault, a vertical section of oceanic crust is exposed and has been mapped using submersibles, hence our inverted velocity models can be directly compared with adjacent outcrop data. Synthetic tests using the 2D inversion code suggest that the inverted velocity structure may contain artefacts caused by offline arrivals and feathering of the streamer. Synthetic tests using the 3D inversion code show that true velocity structure can be recovered, and 3D inversions of the real data suggest that there is a velocity inversion in the upper oceanic crust in the area modelled.

Introduction

Upper oceanic crust is formed from basaltic dykes, pillow lavas and lava flows, which may be covered by sediment if close to a continent. Seismic refraction data through upper oceanic crust show a rapid increase in seismic velocity, from velocities of 2-3 km/s to > 4.5 km/s, called the seismic 2A/2B boundary (Figure 1c). This velocity gradient was traditionally interpreted as the transition from extrusive lavas to dykes (e.g. Harding et al., 1993), but is also argued to be an alteration front that is associated with crack closure due to hydrothermal mineralization (e.g. Christeson et al., 2007). Variability in layer 2A thickness is often used to make inferences regarding the geology of oceanic crust, hence, identifying the true nature of the seismic 2A/2B boundary is fundamental to our understanding of the accretion and cooling of oceanic crust.

Seismic data at the Blanco Transform Fault (BTF) have been acquired on a plateau directly adjacent to the fault scarp, along which a vertical section of oceanic crustal rock is exposed and has been mapped (Fig. 1a and b). Figure 1c shows an example CDP gather and Figure 1d a reflection stack. Christeson et al. (2007) used standard processing and travel-time modelling to locate the position of the seismic 2A/2B boundary. Here, we invert for the 3D full wavefield in order to improve the spatial resolution of the final velocity model and better understand the nature of the 2A/2B transition.

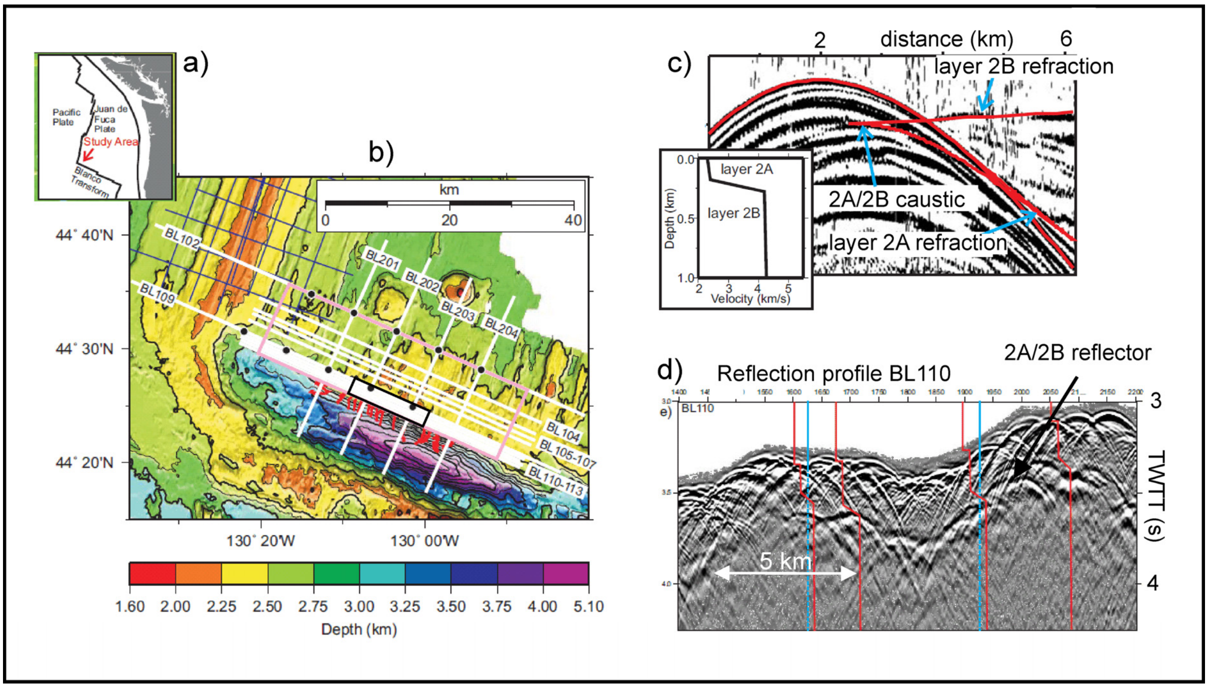

Figure 1

a) location of study area,

b) experimental geometry; white lines are reflection profiles, colour is swath bathymetry, red lines are mapped geology, black box shows area modelled here,

c) typical CDP gather (inset is velocity increase across seismic layer 2A and 2B)

d) reflection stack showing rough sea-bottom and 2A/2B reflector

Method

From the data acquired we have selected five ~50 km long seismic profiles, as they are reasonably closely spaced at ~500 m apart, and are located close to the Blanco transform fault where oceanic crust is exposed (Figure 1b). Shots were fired along straight lines parallel to the fault, and recorded on a 480 channel streamer that was ~6.2 km long and was feathered at an angle of ~5 degrees. We have run a suite of 2D wavefield inversions on the real data using a frequency domain based code (Pratt et al., 1996; Pratt, 1999). Wavefield inversions remain challenging, particularly in geologically complex area such as the one studied here. These particular data have some advantages, for example: arrivals from layers 2A and 2B are clear and easy to identify (Fig. 1c), the signal frequency content is broad (3-100 Hz), the signal-to-noise ratio is high, and neither the water-bottom multiples or direct wave arrivals interfere with the primaries.

2D wavefield inversions of the real data provided intriguing images of the fine-scale velocity structure in the upper crust, and suggest that there might be a velocity inversion in the upper oceanic crust above seismic layer 2A/2B (Morgan et al., 2009). However, the sea-bed is topographically rough and individual shot gathers occasionally contain more than one sea-bed reflection, and it is not clear which reflections are from directly below the seismic profile and which are offline. In addition, the streamer is feathered so that the shots and common midpoints are not aligned. Hence, we were concerned that both the offline sea-bed reflections and the experimental geometry may produce artefacts in the inverted 2D velocity structure.

In order to test whether we are likely to incorrectly map 3D structure into our 2D velocity models, we performed a series of synthetic tests. One of these tests is shown in Figure 2. Figure 2a shows a 3D synthetic velocity model; the sea-bottom has been constructed using swath bathymetry (Figure 1b) and a constant velocity profile has been hung off the sea-bottom. The velocities were chosen to be typical for the layer 2A/2B transition. We acquired 2D synthetic data across this 3D velocity model using the same geometry as for the real experiment. Thus these synthetic data will contain online and offline arrivals, and will include the effects of the feathered streamer.

Results

We have run two separate inversions of the synthetic data acquired across the 3D velocity model shown in Figure 2a and Figure 2b – the latter is a vertical velocity slice through the shots in figure 2a, and is roughly coincident with the reflection section shown in Fig. 1c. Figure 2c shows the results using the 2D inversion code. The bottom of the inverted model is masked out, as no surface arrivals have travelled through this region. The inverted velocities differ from the known velocities by up to a few hundred metres per second in places. This suggests that our 2D inversions of the real data may be unreliable, and that some of the inverted velocity structure may be an artefact of modelling a 3D world in two dimensions. The same synthetic data were them modelled with our newly-developed 3D frequency domain inversion code (Warner et al, 2008) – Figure 2d. The 3D inversion recovers the velocity structure well, and contains small artefacts only (< 50 m/s). This test suggests that a 3D inversion of the real data should be able to resolve the real velocity structure.

Figure 3 shows a preliminary inversion of the real data along BL110 using the 3D inversion code. The 3D inversion shows a similar feature to the 2D inversions - that there is a velocity inversion in the upper oceanic crust in the area modelled.

Figure 2

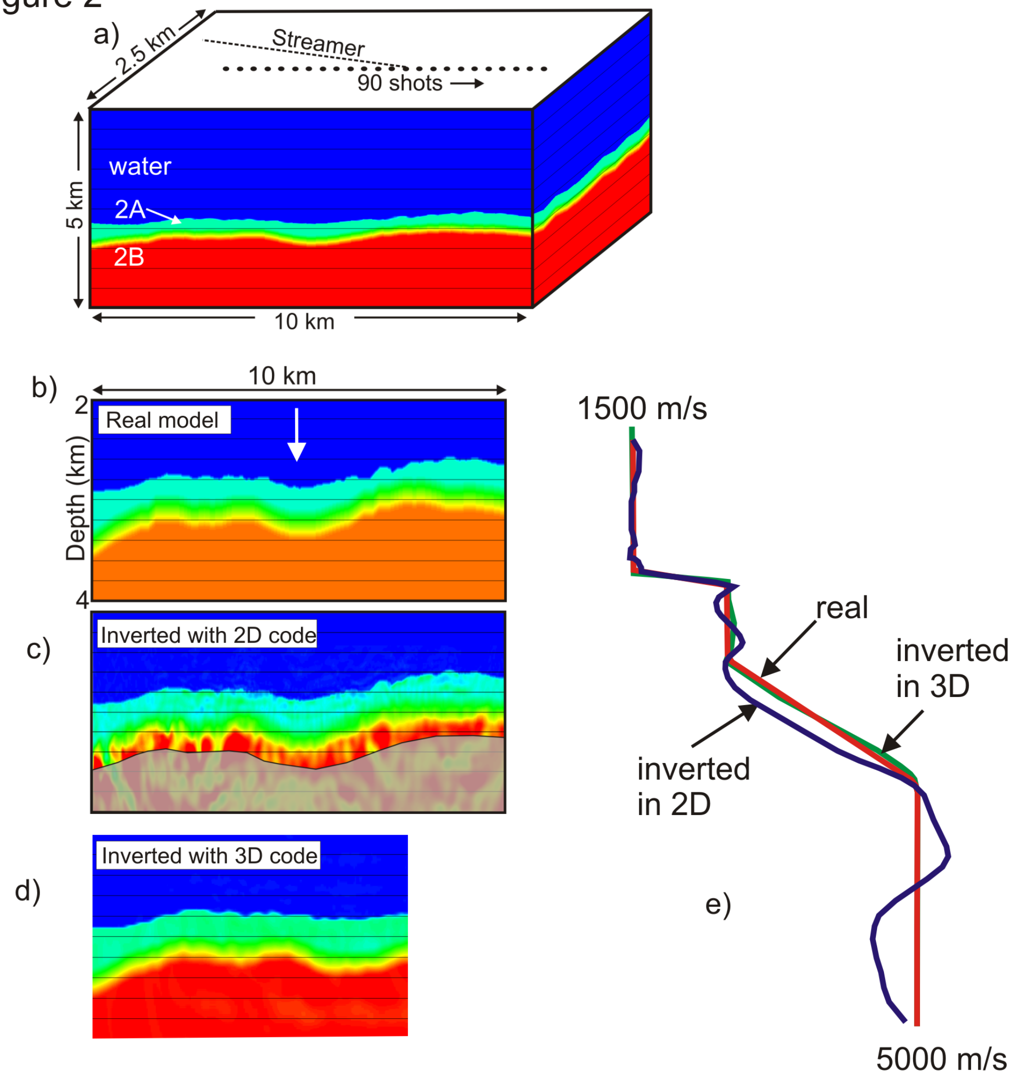

Figure 3

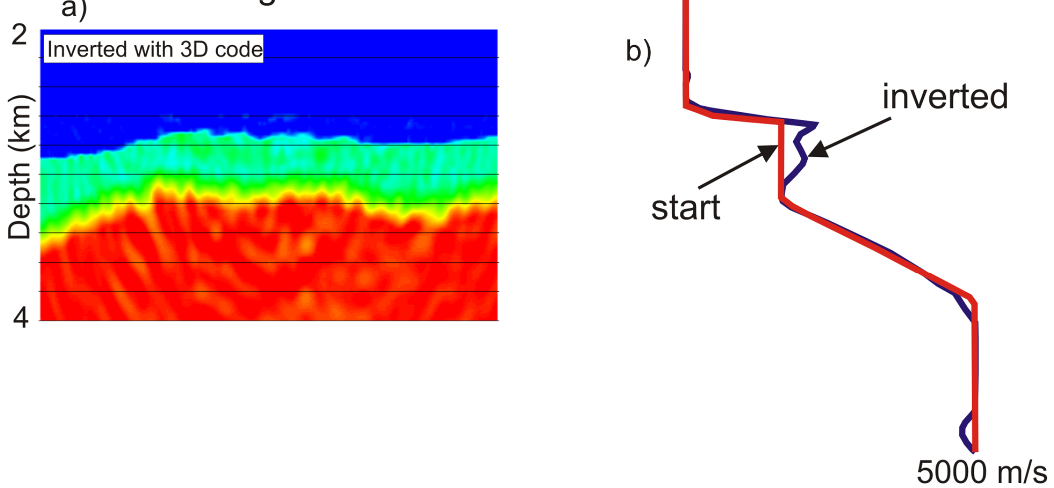

Figure 2

a) 3D synthetic (real) velocity model,

b) synthetic velocity model below shots (see Fig 2a),

c) inverted velocity model using 2D inversion code,

d) inverted velocity model using 3D inversion code,

e) velocity profiles from 2b (red), 2c (blue) and 2d (green); white arrow indicates location.

Figure 3

a) 3D inversion of real data

b) velocity profiles through starting and inverted velocity models.

Conclusions

We have run 3D full wavefield inversions of real seismic data acquired in deep water in the Pacific ocean. The sea-bottom is topographically rough, and the uppermost kilometre of crustal rocks are comprised of basaltic dykes, pillow lavas and lava flows. Shots were recorded on a 480 channel that was ~6.2 km long and was feathered at an angle of ~5 degrees. 2D wavefield inversions of 5 parallel profiles provide intriguing images of the fine-scale velocity structure in the upper crust, but synthetic tests suggest that some of the velocity structure is an artefact of modelling a 3D world in two dimensions. Synthetic tests using the 3D inversion code show that true velocity structure can be recovered for a wide range of plausible velocity models. Preliminary 3D inversions of the real data suggest that there is a velocity inversion in the upper oceanic crust.

References

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