Mar Geophys Res
DOI 10.1007/s11001-011-9143-z

ORIGINAL RESEARCH PAPER

Using overlapping sonobuoy data from the Ross Sea to construct a 2D deep crustal velocity model

M. M. Selvans · R. W. Clayton · J. M. Stock · R. Granot

Received: 13 June 2011 / Accepted: 1 December 2011

Abstract Sonobuoys provide an alternative to using long streamers while conducting multi-channel seismic (MCS) studies, in order to provide deeper velocity control. We present analysis and modeling techniques for interpreting the sonobuoy data and illustrate the method with ten overlapping sonobuoys collected in the Ross Sea, offshore from Antarctica. We demonstrate the importance of using the MCS data to correct for ocean currents and changes in ship navigation, which is required before using standard methods for obtaining a 1D velocity profile from each sonobuoy. We verify our 1D velocity models using acoustic finite-difference (FD) modeling and by performing depth migration on the data, and demonstrate the usefulness of FD modeling for tying interval velocities to the shallow crust imaged using MCS data. Finally, we show how overlapping sonobuoys along an MCS line can be used to construct a 2D velocity model of the crust. The velocity model reveals a thin crust (5.5 ± 0.4 km) at the boundary between the Adare and Northern Basins, and implies that the crustal structure of the Northern Basin may be more similar to that of the oceanic crust in the Adare Basin than to the stretched continental crust further south in the Ross Sea.

M. M. Selvans · R. W. Clayton · J. M. Stock
Seismological Laboratory, California Institute of Technology, 1200 E. California Blvd., MC 252-21, Pasadena, CA 91125, USA

M. M. Selvans (✉)
Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, MRC 315, Washington, DC 20560, USA
e-mail: selvans@gps.caltech.edu

R. Granot
Department of Geological and Environmental Sciences, Ben-Gurion University of the Negev, Beer Sheva, Israel

Keywords Sonobuoy · Multi-channel seismic · Ross Sea · Finite-difference · 2D velocity model · Crustal structure

Introduction

Using sonobuoys to investigate deep crustal structure

Sonobuoys provide a means of obtaining long offsets for deep velocity analysis while conducting multi-channel seismic (MCS) studies. They are preferable to using a long streamer in locations with difficult open water conditions, such as in the Ross Sea and Southern Ocean, where research cruises regularly encounter sea ice, icebergs, and intense storms. Ocean-bottom seismometers (OBS) provide another method for collecting active seismic data with large offsets, but are more expensive than sonobuoys and are challenging to recover in the conditions described above. Sonobuoys are also an ideal method for collecting large offset seismic data in other settings where ship navigation is constrained, such as locations with ship traffic and narrow bodies of water such as fjords.

Marine refraction seismology has long been recognized as an essential technique for mapping the sedimentary rock and basement structure of the Antarctic margins (e.g., Ewing and Heezen 1956; see also Anderson 1999 and references therein). We present techniques for interpreting deep crustal structure from overlapping sonobuoy data collected in the northwestern Ross Sea. We first demonstrate the necessity and method for correcting for the effect of currents and changes in ship navigation, prior to using standard processing techniques to obtain 1D velocity models from the sonobuoys. Second, we explore the benefits of using finite-difference (FD) method modeling of sonobuoys to tie interval velocities to the shallow crustal

Published online: 20 December 2011

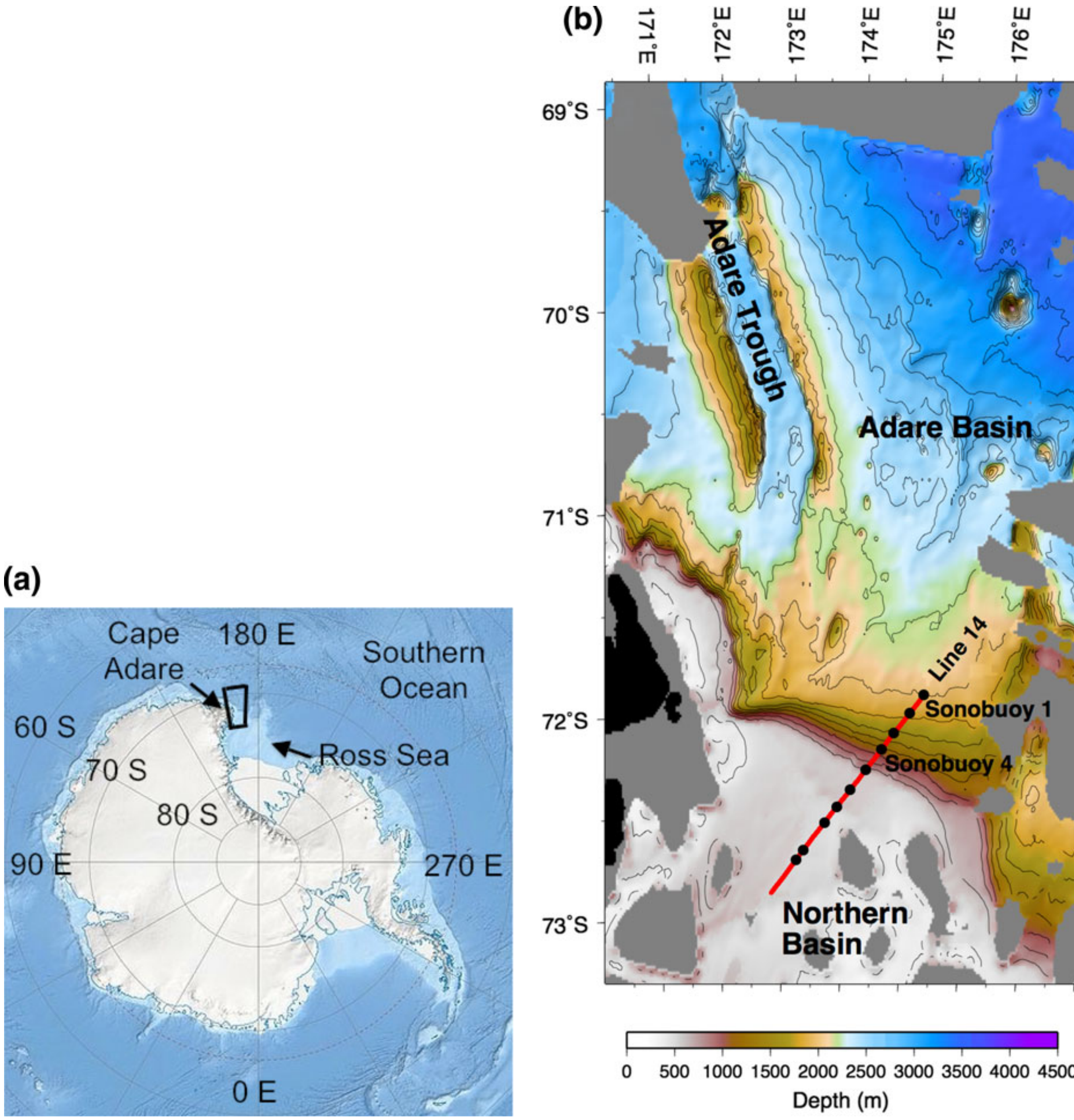

Fig. 1 **a** Sonobuoy processing methods are developed for data collected in the northwestern Ross Sea (*black box*), offshore from Antarctica. **b** We interpret a subset of the active seismic data collected in the northwestern Ross Sea during research cruise NBP0701, shown on top of multibeam bathymetry (NBP0701 Data Report 2007). Ten sonobuoys are closely spaced along multi-channel seismic (MCS) Line 14, which runs from deep water in the Adare Basin onto the continental margin in the Northern Basin

structure imaged with MCS data. Finally, we apply these methods to a set of ten overlapping sonobuoys in order to construct a 2D velocity model of the deeper crustal structure along an MCS line.

Modeling methods for seismic refraction data

The FD method is useful for modeling propagation of seismic waves, including turning waves and their reflections and conversions at interfaces, for complex subsurface structure. Ray tracing is similarly useful, but has difficulty predicting the amplitudes of refracted energy, with the consequence that it is often difficult to judge the importance of some of the predicted arrivals. One advantage of using the FD method to model seismic refraction data is the ability to model the entire wave front from source to receiver, which allows us to both tie the shallow structure detected with MCS data to particular velocity horizons and to determine the deeper crustal velocity structure.

Shipp and Singh (2002) used a 2D elastic finite-difference forward model and an iterative full wavefield inversion scheme to best fit their streamer data (with a 12 km offset), solving for velocity structure down to 4 km below the seafloor. This approach is computationally demanding, with residuals from the comparison of data and the model being back propagated for every time step, which is why they were constrained to analysis of relatively shallow crust. Jones et al. (2007) performed a similar analysis on streamer data (with offsets of 15 and 18 km); they additionally derived 1D velocity models from the intercept-time-slowness (τ-p) domain, and downward continued the data in order to create an image of the layers at depth. They were able to image the velocity structure to a depth of 6 km.

In order to conduct a marine seismic refraction experiment with an offset larger than tens of kilometers, Ritzmann et al. (2004) used OBS instruments. They used a ray tracing model to reproduce the data, and found that crustal thickness varied widely, from 32 km for the continental crust of Svalbard to as little as 2 km (excluding the ~2 km of sediment on top of the basement) near the margin between continental and oceanic crust. Mantle velocities are generally >8.0 km/s, but were found to be as low as 7.7 km/s west of Molloy transform fault, where the mantle is likely serpentinized (Ritzmann et al. 2004). Also offshore from Svalbard, Geissler and Jokat (2004)

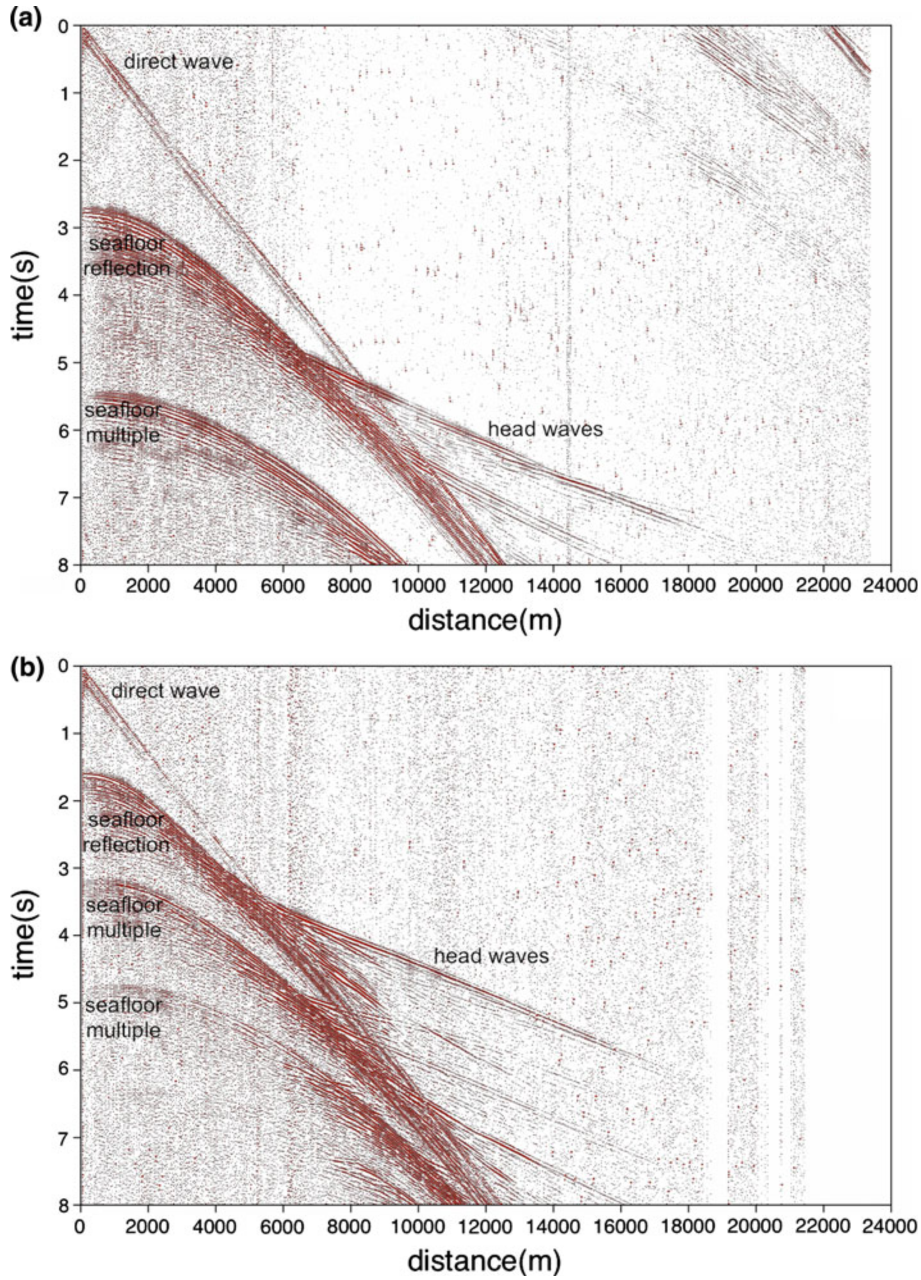

Fig. 2 Raw data for Sonobuoys 1 (**a**) and 4 (**b**) on Line 14 show key features used to determine the 1D velocity profile at these locations. Time is equivalent to two-way travel time, and distance is relative to the airgun source. The direct arrival comes in first in time for smaller offsets, while energy from head waves traveling along layer interfaces at depth comes in first for larger offsets; the slopes of these linear features are determined by the interval velocities in the crust. The hyperbolic reflection from the seafloor comes in second for smaller offsets and provides a time constraint used to determine the water layer thickness. Later hyperbolic features are multiples of the seafloor reflection. A bandpass filter of 5, 15, 35, and 40 Hz is applied to the data

used sonobuoy data as 1D velocity profile "pseudo-boreholes" along an MCS line to obtain deeper crustal structure.

Collecting sparse sonobuoy data along MCS lines and modeling the resulting profiles with ray tracing methods has been used to investigate deep crustal structure along Antarctic continental margins as well, for example offshore from Wilkes Land (Close et al. 2009) and Enderby and Mac. Robertson Lands (Stagg et al. 2004). In this study, we combine the analysis of sonobuoy data in the τ-p domain with FD modeling of the waveform to derive a sequence of overlapping 1D velocity models, which are then interpreted into a 2D deep crustal velocity model.

Seismic data in the Ross Sea

We will illustrate the analysis procedure for sonobuoy data collected in the Adare Trough region of the Ross Sea, offshore from Antarctica (Fig. 1a). The Adare Trough, a dead mid-ocean spreading ridge, lies in the deep water of the Adare Basin and trends southward toward the Northern Basin, which is located up on the continental shelf (Fig. 1b). How seafloor spreading was linked to extension in the West Antarctic Rift System to the south is of major interest, yet still poorly resolved. Shallow structure is well imaged by MCS data (Granot et al. 2010). However, understanding the tectonics in the area requires knowledge

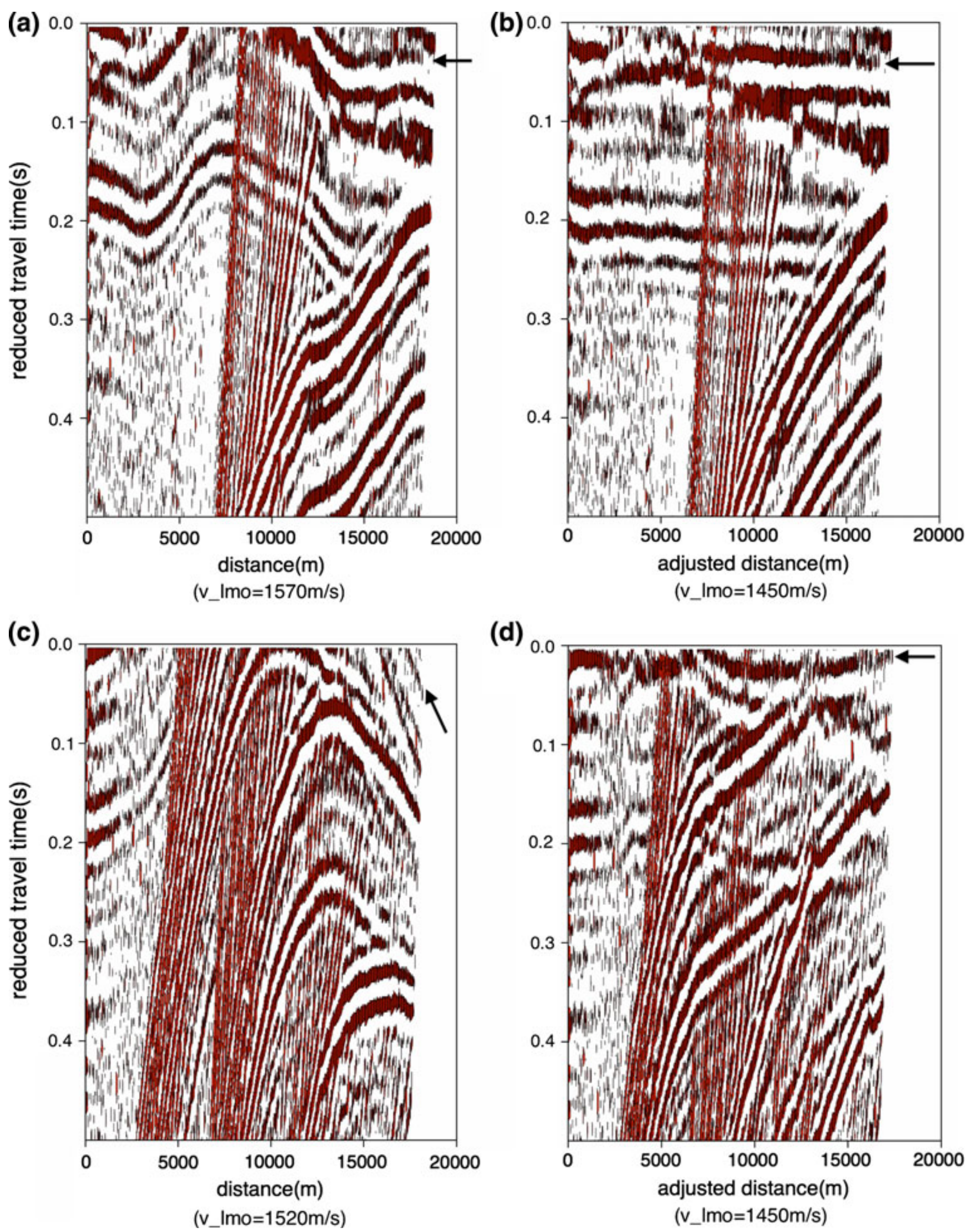

Fig. 3 We correct the sonobuoy shot spacing to account for ocean currents and changes in ship navigation. A best possible linear moveout (l mo) of 1570 m/s is applied to the direct wave in the raw data for Sonobuoy 1 (**a**), and displayed in terms of the resulting reduced travel time. The distances between shots are corrected (**b**), based on picks made along the direct wave, producing a linear direct wave with the correct water velocity of 1450 m/s (as determined from multi-channel seismic data). Sonobuoy 4 has a linear moveout of 1520 m/s before the correction (**c**), and 1450 m/s after (**d**). See Table 1 for direct wave velocities prior to the correction

of the deep velocity structure, for example the Moho depth along the continental shelf break (see Selvans et al. (submitted) for an analysis of all sonobuoy data, and implications for tectonics).

We collected seismic reflection and refraction data during research cruise NBP0701 on board the *R/VIB Nathaniel B. Palmer*. Sonobuoy data were collected in the Adare and Northern Basins, with maximum offsets from the ship of 20–30 km. Sonobuoys presented here were deployed with a regular spacing of ∼15 km (Fig. 1b inset) in order to obtain overlapping records, since data was generally returned from as far as 20–30 km from the ship.

Most sonobuoy studies in the western Ross Sea have focused on basins south of the Northern Basin, and have used linear moveout and ray tracing methods to determine sediment velocity gradients and basement depth (Houtz and Davey 1973; Cooper et al. 1987; Cochrane et al. 1995). Ocean bottom seismic refraction studies have investigated the deeper crustal structure in the central and southern Ross Sea, using ray tracing and amplitude modeling methods (Tréhu et al. 1993; Trey et al. 1999). Crustal thickness, defined as the depth to velocities of >8000 m/s, is as little as 16 km beneath sedimentary basins underlain by thinned crust and magmatic intrusions, and 21 km beneath intervening basement highs (Trey et al. 1999).

MCS data from NBP0701 contain resolvable primaries from up to 2.3 s travel time (i.e., “two-way travel time”) below the seafloor in the Adare and Northern Basins, revealing variable sediment thickness (Granot et al. 2010). Migrated MCS data delineate the structure and deformation of the sedimentary units in the Adare and Northern Basin, but cannot tie velocity values to the subsurface layers, and cannot detect crustal structure below the basement rock.

Methods for sonobuoy analysis and interpretation

Trace spacing adjustment and construction of 1D velocity profiles

MCS data were collected using a 1.2 km, 48-channel streamer; this short streamer was required due to regular interactions with sea ice. Data presented here were obtained using a 6-element Bolt-gun array with a total capacity of 34.8 liters, with a typical source spacing of approximately 40 m. We present data analysis methods, and results, for ten overlapping sonobuoy profiles along MCS Line 14, which ran from the deep water of the Adare Basin south onto the shallow-water shelf of the Northern Basin (see Fig. 1b).

These sonobuoys were deployed approximately every 15 km, and returned data for 20–30 km of offset. While the source moved away from the sonobuoy launch position at a nearly uniform speed and direction, sonobuoy drift after launch (due to ocean currents) and slight variations in ship navigation led to varying shot spacing in the refraction data set. We correct for these effects before analysis, modeling, and interpretation of the data. The sonobuoy data provide details of the crustal structure to greater depth than the MCS data along the same line, and directly detect interval velocities; MCS data are useful in determining the sound speed (seismic velocity) of the water layer and the shallowest of the rock layers. Consistent with the generally flat horizons observed in the MCS data, we assume plane layers for all velocity horizons.

We show the processing steps needed for accurate interpretation of the sonobuoy data, starting from the raw data for the deep-water Sonobuoy 1 (Fig. 2a) and the shallower-water Sonobuoy 4 (Fig. 2b). We highlight the low-frequency reflected and refracted energy from the seafloor and within the crust by applying a tapered band-pass filter to the data, set to 5, 15, 35, and 40 Hz (i.e., data with frequencies between 15 and 35 Hz are passed through with their amplitudes unaltered, and data in the ranges 5–15 and 35–45 Hz are passed through with amplitudes that decrease to zero at the edges of the filter). Several key features of the data are obvious in these sonobuoy images: (1) the direct wave, from source to receiver, is the first arrival at smaller offsets (e.g., 0–8500 m in Fig. 2a), (2) the reflection from the seafloor comes in next for smaller offsets (e.g., 0–6500 m in Fig. 2a), and due to its strength as well as noise in the data no reflections from deeper layers are observed at these offsets (subsequent reflections are seafloor multiples), (3) head waves, refracted energy from layer interfaces at depth, come in as first arrivals for larger

Table 1 Direct wave velocities, as determined using a best possible linear moveout fit to the feature, are listed for Sonobuoys 1–10 (S1–S10) on Line 14

	Direct wave velocity before correction (m/s)
S1	1570
S2	1570
S3	1480
S4	1520
S5	1330
S6	1050
S7	1180
S8	1200
S9	1500
S10	1580

The slope of the direct wave is corrected by adjusting the distance between traces, such that the sonobuoy data has the correct water velocity value of 1450 m/s

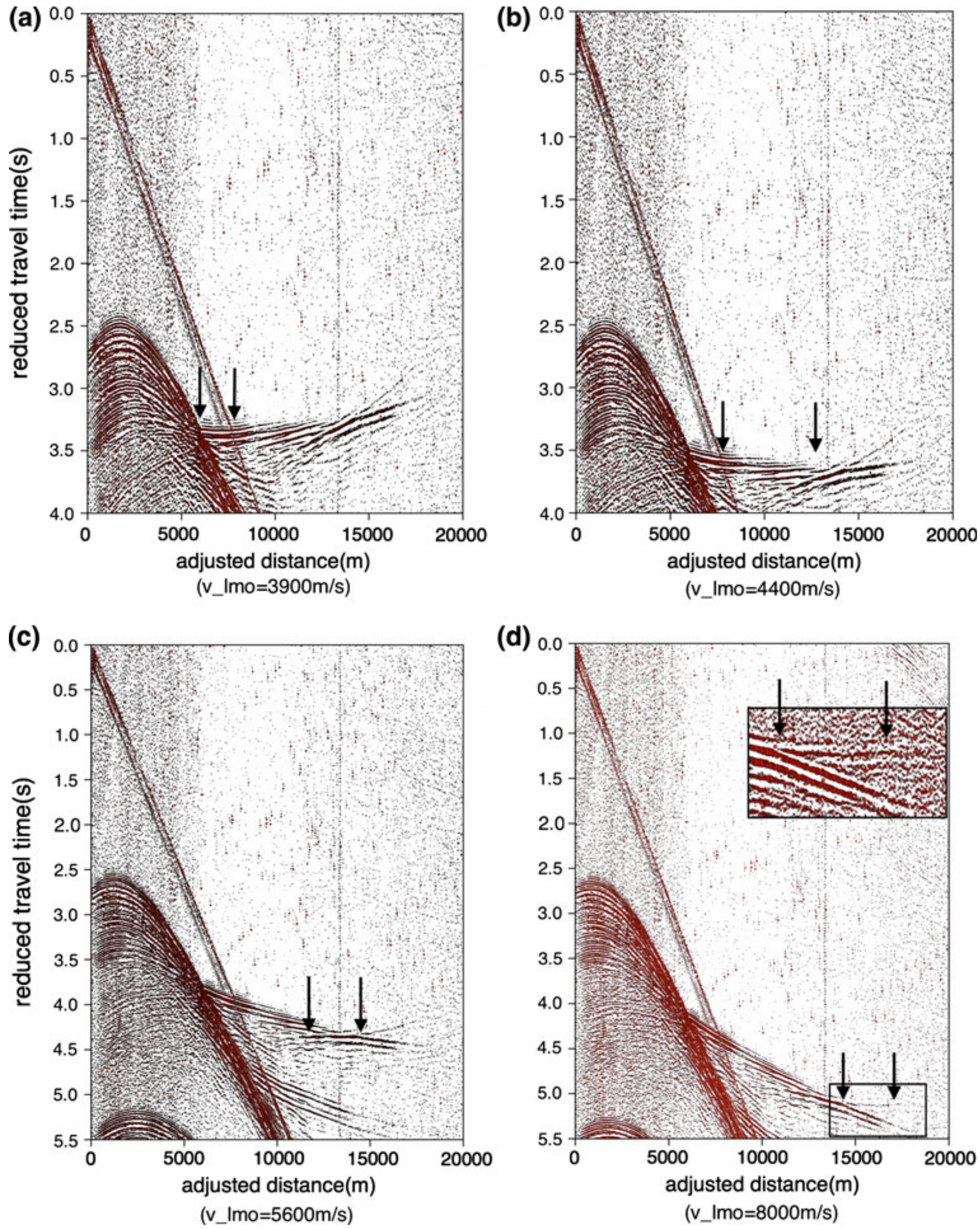

Fig. 4 Linear moveout (l mo) is applied to the refracted energy in the distance-adjusted version of Sonobuoy 1, to directly measure interval velocities and associated times (τ). We detect layers with velocities of 3900 m/s (**a**), 4400 m/s (**b**), 5600 m/s (**c**), and 8000 m/s (**d**), with head waves lying within the arrows; the 8000 m/s layer is also shown as an inset. The latter value is the only instance of a velocity that can be interpreted as the Moho. Refracted energy from the seafloor is not observed (an interval velocity of 2200 m/s for the uppermost sediment layer is determined from multi-channel seismic data)

offsets (e.g., 8,500–19,000 m in Fig. 2a). The offset range of each arrival type varies with water depth and subsurface layer thicknesses (see Fig. 2b).

We correct sonobuoy shot spacing so that the direct wave has a slope of 1450 m/s, as determined by applying a linear moveout to the direct wave in the MCS data (accurate to ±50 m/s). This value for the seismic velocity of the water layer is consistent along the MCS line, and is therefore taken as the default value for all sonobuoys; it is also consistent with measurements of sound speed in water taken during the cruise. Data from the one expendable sound velocimeter (XSV) show water velocities within 1441–1451 m/s (to 600 m depth), while expendable bathythermographs (XBTs, which measure temperature with depth and use surface measurements of salinity to compute sound velocity) show water velocities of 1444–1480 m/s at the northern end of MCS Line 14, and 1446–1448 m/s at the southern end of Line 14. These sensor measurements of sound speed in water are accurate to ±0.25 m/s.

In order to correct the shot spacing, we adjust the distance coordinate of each trace, such that the direct wave is linear and has the correct slope in the distance versus time plot. The direct wave of Sonobuoy 1 is shown with a linear moveout applied before (Fig. 3a, with a direct wave velocity of 1570 m/s) and after (Fig. 3b, with a direct wave

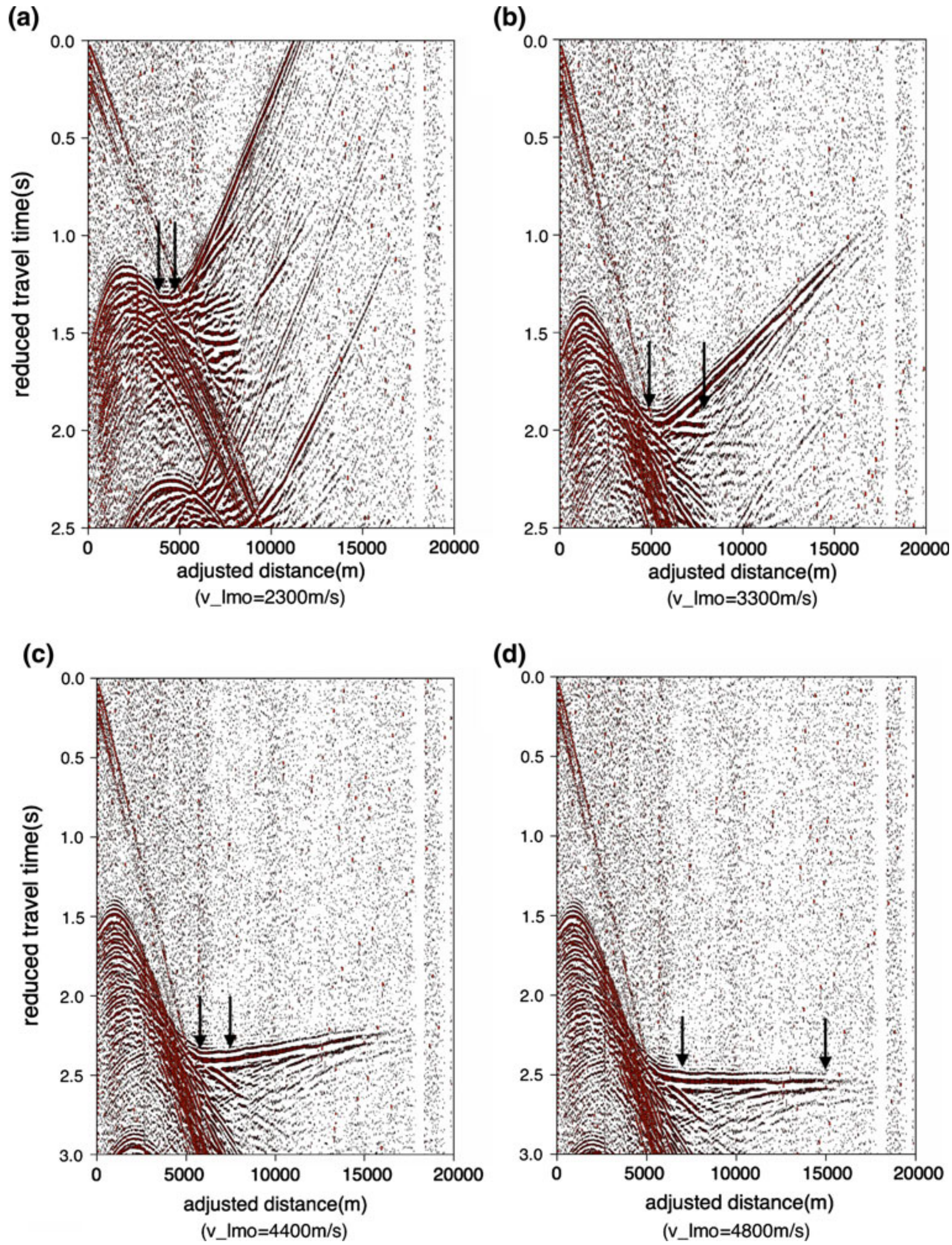

Fig. 5 Linear moveout (l mo) is applied to the refracted energy in the stretch-corrected version of Sonobuoy 4, directly measuring interval velocities and their associated times. We detect layers with velocities of 2300 m/s (**a**), 3300 m/s (**b**), 4400 m/s (**c**), and 4800 m/s (**d**), with head waves lying within the *arrows*. Refracted energy from the seafloor is not observed (an interval velocity of 2000 m/s for the uppermost sediment layer is determined from multi-channel seismic data)

velocity of 1450 m/s) this correction is made; the adjustment for Sonobuoy 4 makes a similar direct wave velocity adjustment (Fig. 3c, d). Other sonobuoys have larger discrepancies between the direct wave velocities before and after the correction (see Table 1). Sonobuoy 3 requires the smallest correction based on water velocity (average original shot spacing of 38.4 $\pm$ 1.6 m, and average corrected shot spacing of 37.5 $\pm$ 2.1 m), while Sonobuoy 6 requires the largest (similarly, 32.1 $\pm$ 2.6 m, and when corrected 43.7 $\pm$ 3.5 m). This correction will be more straightforward for sonobuoys with reliable GPS; in that case, the sonobuoy location will be at least as well known as that of the MCS streamer.

We directly detect layer velocities, and their associated times, by applying linear moveout to head waves in the corrected sonobuoy data. We detect between three and six distinct layers at depth for each sonobuoy with this method, as shown in detail for Sonobuoys 1 and 4 (Figs. 4 and 5, respectively). This same approach yielded a consistent velocity of 2000 m/s for the shallowest rock layer in seven out of the ten sonobuoys; in other cases this seafloor head wave is not visible. Even when it is visible, it is usually brief ($\sim$10 traces) and so less distinct than most other refractors we measure. For this reason, we verify this layer velocity by applying a normal moveout to the hyperbolic arrival of the seafloor reflection in the raw MCS data, and

Table 2 Velocity model values for Sonobuoys 1–10 (S1–S10) on Line 14 (also see Fig. 10b), and sonobuoy distance along the line

	1 (H_2O)	2	3	4	5	6	7	8	Distance along the line (km)
S1: *d*	0	1960	2950	4090	5850	7500			0
v	1450	2200	3900	4400	5600	8000			
S2: *d*	0	1780	2320	2520	2910	3660	5000		13
v	1450	2000	2200	2800	3700	4400	5000		
S3: *d*	0	1510	2130	2620	3060	3680	5130		27
v	1450	2000	2900	3500	4200	4800	5600		
S4: *d*	0	1160	1460	1800	2600	2990			38
v	1450	2000	2300	3300	4400	4800			
S5: *d*	0	520	1200	1680	2440	2930	3690		52
v	1450	2000	2800	3300	4100	4400	4700		
S6: *d*	0	500	1010	1270	1800	2500	3040		66
v	1450	2000	2500	2800	3500	4100	4500		
S7: *d*	0	470	1320	2320	2820				78
v	1450	2000	3000	4300	4500				
S8: *d*	0	440	1030	1390	2420	3040	3560		88
v	1450	2000	2500	3000	4300	4700	4900		
S9: *d*	0	470	1020	1250	2020	2700			107
v	1450	2000	2400	2900	4000	5000			
S10: *d*	0	460	1030	1230	1790	2260	2660	3350	114
v	1450	2000	2500	2900	3500	4300	4500	4900	

Sonobuoy 1 lies in the deep water of the Adare Basin, while Sonobuoys 5–10 lie in the shallow water of the Northern Basin. Columns contain the depth (*d*, in m) and velocity (*v*, in m/s) for each layer, as derived from the sonobuoy data. Note that the first rock layer velocity for Sonobuoy 1 is 2200 m/s, while the rest have a velocity of 2000 m/s for this layer; this value is obtained by performing a normal moveout on multi-channel seismic data at the sonobuoy location. Velocities are resolved to ±100 m/s, while depths below the seafloor have resolutions of ±0.1–0.4 km, increasing with depth

use it as the first rock layer velocity for all ten sonobuoys; one exception is Sonobuoy 1, where analysis of MCS data provides a first rock layer velocity of 2200 m/s. Layer velocities for all sonobuoys are listed in Table 2.

The velocities and their associated times are used to calculate the 1D velocity profile for each sonobuoy (e.g., Fowler 1990, p. 119–123). Firstly, the water depth and first rock layer thickness are calculated using velocities (v) and reflection times (t) from the MCS data (i.e. $h_1 = 0.5(t_1)(v_1)$). Secondly, the thicknesses of deeper layers ($i = 2, 3$, etc. are rock layers) are determined using standard seismic refraction analysis:

$$h_i = \frac{v_i}{2\,\cos(\theta_{i,i+1})}\left[\tau_{i-1} - \sum_{j=1}^{i-1}\frac{2h_j}{v_j}\cos(\theta_{j,i+1})\right]$$

where $\theta_{a,b} = \arcsin(v_a/v_b)$ and τ is the reduced travel time associated with the layer velocity, obtained when linear moveout is applied to the sonobuoy data. Additionally, a direct image of the τ-p curve can be obtained by a radon transform of the data (McMechan and Ottolini 1980); an example is shown in Fig. 6.

Layer thicknesses are summed to determine layer depths, resulting in a 1D velocity profile for the sonobuoy. Velocity profiles for Sonobuoys 1 and 4 are plotted with respect to depth below the seafloor for ease of comparison (Fig. 7), revealing the unusually high-velocity detail in Sonobuoy 1, with a maximum velocity of 8000 ± 100 m/s at 5.5 ± 0.4 km crustal depth. Table 2 lists the details of the velocity models for all ten sonobuoys.

Finite-difference method modeling and depth migration

We verify our 1D velocity models via finite-difference (FD) method modeling of the sonobuoys, and by imaging the subsurface through depth migration of the data. Our FD model solves the acoustic wave equation in an elastic medium (Vireaux 1986), using the velocity models calculated above to reproduce the sonobuoy data (Fig. 8). The model solutions are 2nd order in time and 8th order in space; the resulting lack of dispersion of the wave front through the medium allows for accurate timing. The high spatial accuracy is required because of the wide range of velocities in the model (1500–8000 m/s). We assume layer densities of 2600 kg/m^3 for velocities from 2000 to 4000 m/s; 2700 kg/m^3 for velocities from 4000 to 6000 m/s; and 2800 kg/m^3 for velocities of 6000 m/s and greater;

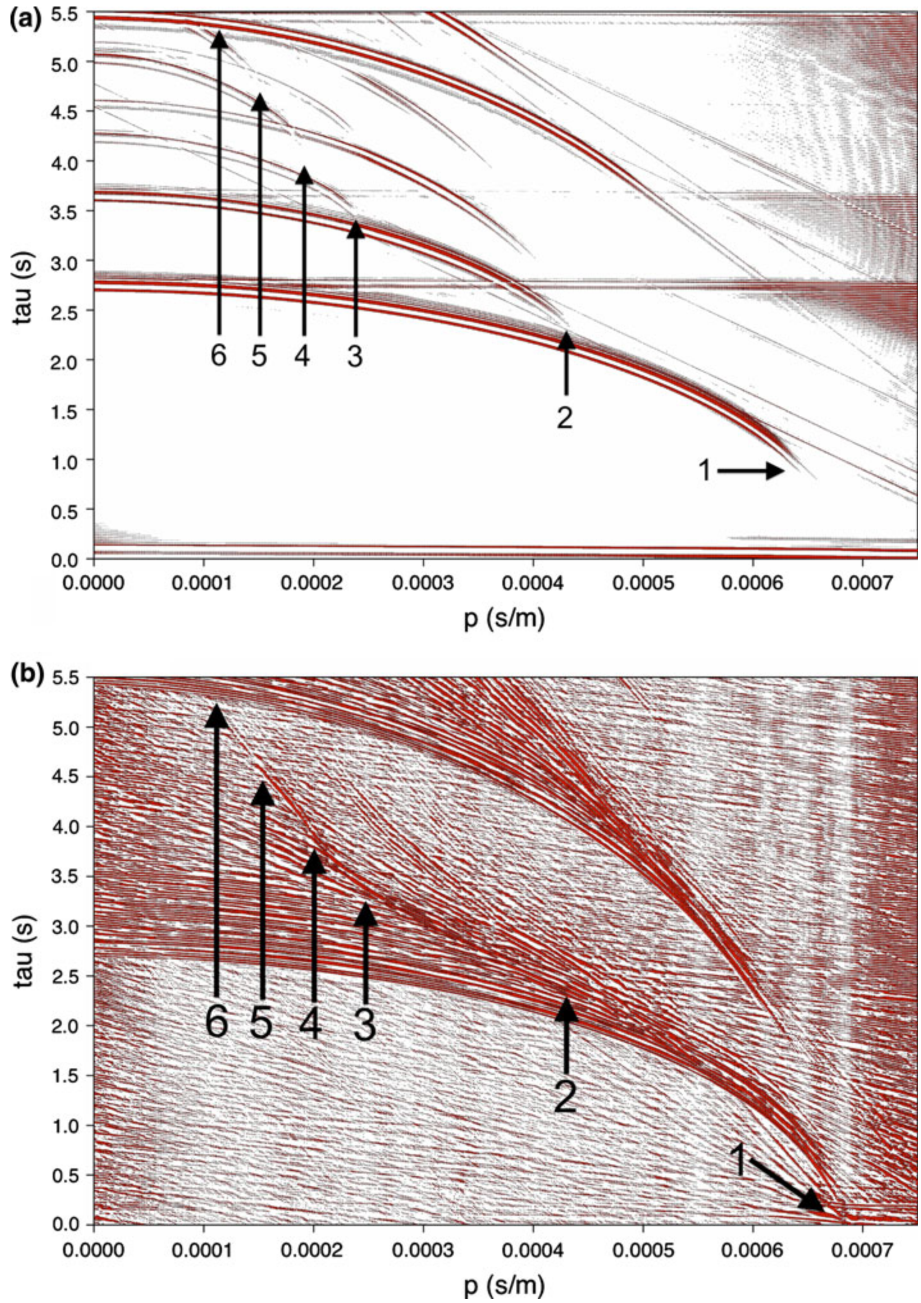

Fig. 6 Applying a radon transform to the model (**a**) and data (**b**) for Sonobuoy 1 reveals the ellipses that image each velocity layer in τ-p space. The cusp at the beginning of each ellipse is defined in terms of the slowness (p) of the layer (or $1/v$, where v is the velocity), where layers 1–6 are labeled (1450, 2200, 3900, 4400, 5600, and 8000 m/s respectively). Note that cusps of layers 1, 4, and 5 are relatively low amplitude, and that layer 6 is hidden by the multiple of layer 1

these densities are consistent with sedimentary, upper crust (e.g., basement rock), and lower crustal rock respectively (Jones 1999, p. 64).

By reproducing the sonobuoy data without noise, we obtain reflection times associated with each head wave. These allow us to tie the shallow structure observed in the stacked and migrated MCS data to particular velocity horizons (Fig. 9). We are therefore able to interpret the shallow structure revealed by MCS data in terms of directly observed interval velocities. We additionally extend the depth to which the crustal velocity structure is delineated, below the acoustic basement that limits MCS penetration.

We also verify our 1D velocity models by depth migrating the sonobuoy data (Fig. 10). First a radon transform is applied to the distance-adjusted sonobuoy data (e.g., Jones 1999, p. 74–75), which maps the data into slowness ($1/v$)—intercept time (τ) space (Fig. 6). The transformed data is then migrated in depth, using the finite-difference version of the wave-equation method (Clayton and McMechan 1981), based on the 1D velocity model.

The subsurface image is in terms of velocity versus depth, and for Sonobuoy 1 (Fig. 10a) shows flat shallow layers below the water layer, as observed in the stacked and

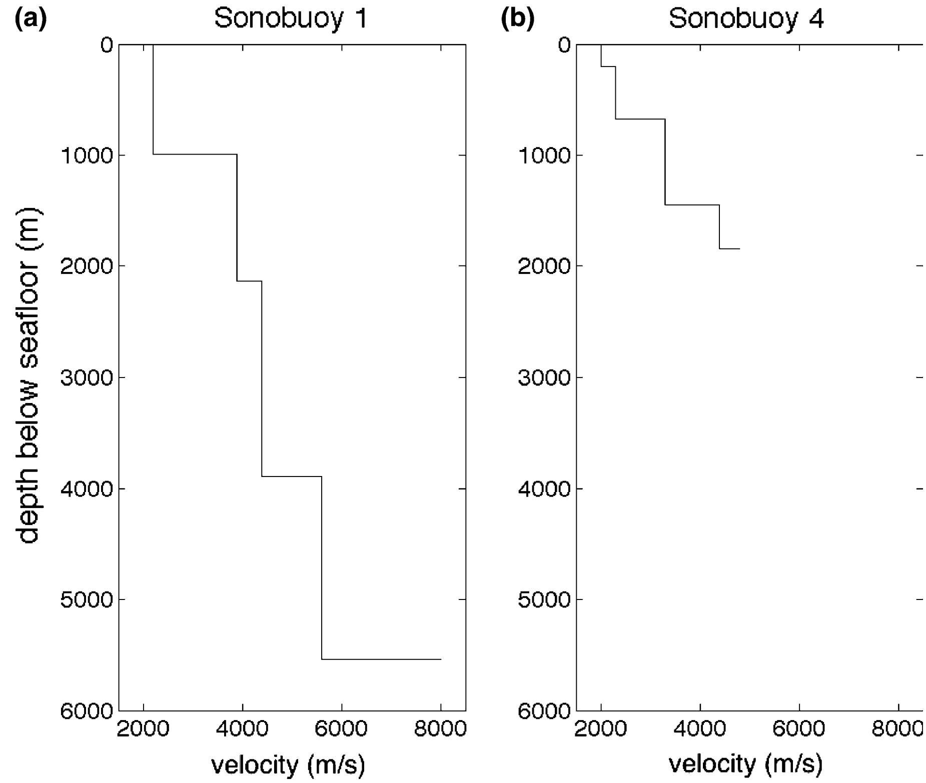

Fig. 7 Velocity models are constructed for Sonobuoys 1 (**a**) and 4 (**b**) using the velocities and associated times obtained through standard linear moveout and normal moveout methods (see text for detail). The velocity of 8000 m/s suggests that the Moho is only 5.5 km below the seafloor beneath Sonobuoy 1. See Table 2 for velocity model details

migrated MCS data (Fig. 9a); in general, flat interfaces in depth-migrated sonobuoy images indicate use of an accurate velocity profile. Flat shallow interfaces in the migration data for Sonobuoy 4 (Fig. 10b) validate its respective velocity model. Slight tilting in the deeper layers indicates small lateral variations in velocity that cannot be properly modeled with 1D analysis.

Constructing a 2D velocity model

We interpolate velocity values between the 1D velocity models from overlapping sonobuoys, using them as “pseudo-borehole” data, to construct a 2D velocity model along Line 14 (Fig. 11). Since the sonobuoys record a greater offset than their spacing and are reveal similar 1D velocity profiles, we interpolate between sonobuoys to construct a 2D velocity profile across the continental shelf break. We use a Delaunay triangulation algorithm to perform the interpolation, the vertices being location (depth below seafloor and distance along the line) and velocity. The 1D models include maximum velocities of 4500–8000 m/s, to maximum depths of 2.7–7.5 km below sea level (i.e., 1.8–5.5 km below seafloor; see Table 2 and Fig. 11b, respectively). Velocities are resolved to ±100 m/s, while depths below the seafloor have resolutions of ±0.1–0.4 km, increasing with depth.

Results

By interpolating between 1D velocity profiles for ten overlapping sonobuoys, we obtain a 2D velocity model along 115 km of a seismic reflection line that crosses the continental shelf break, with a maximum depth of 5.5 km below the seafloor under Sonobuoy 1 (Fig. 11). We determine the crustal structure to a greater depth than is possible with the MCS data alone, and provide interval velocities for interpretation of the subsurface. We obtain accurate velocities by correcting the sonobuoy data before linear moveout analysis, since changing currents and ship navigation can alter the apparent velocities of head waves coming in from interfaces at depth. Other sonobuoy studies in the Ross Sea have detected sound speeds in water of 1430–1540 m/s, close to the true value (compare to our spread of 1050–1580 m/s, see Table 1), and so have not needed to make this distance adjustment before performing data analysis (e.g., Houtz and Davey 1973; Cooper et al. 1987; Cochrane et al. 1995).

We are able to tie our 1D velocity models to layers imaged in the shallow crust using MCS because we use an FD model to reproduce the obvious features in the sonobuoy data (Fig. 8), which allows us to determine the reflection time associated with each head wave. These reflection times are directly compared to MCS images of the subsurface, to place accurate constraints on layer

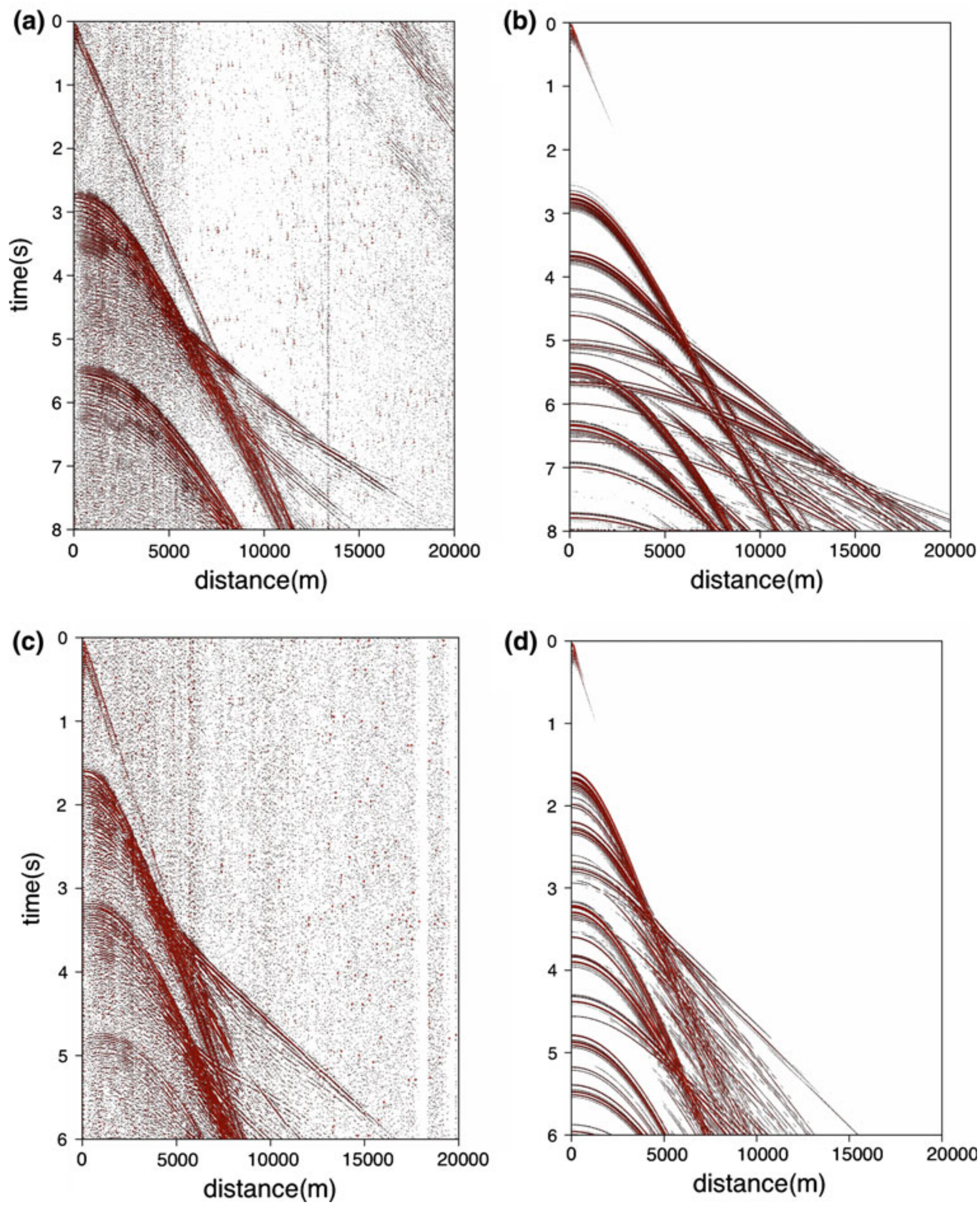

Fig. 8 Using the velocity model derived from the distance-adjusted data for Sonobuoy 1 (**a**), we reproduce the key features in the data using a finite difference model (**b**). The finite difference model allows head wave features to be tied to their associated reflections, which come in after the seafloor reflection and are masked in the data. The corrected data for Sonobuoy 4 (**c**) and resulting finite difference model (**d**) highlight the usefulness of modeling in shallow-water locations for the identification of multiples

velocities. This provides a method for determining sediment thickness, whose structure is otherwise interpreted (from MCS data) entirely in time space (Granot et al. 2010).

Due to the experimental design and processing methods we employ, we are able to accurately determine crustal structure in the Adare and Northern Basins to a greater depth than is possible with the MCS data. Overlapping sonobuoys allow us to construct a 2D velocity profile of the crust more cheaply than is possible with ocean-bottom seismometers. Since the NBP0701 cruise deployed similarly spaced sonobuoys along other MCS lines in the southern Adare Basin and northern part of the Northern Basin, these methods can be used to construct a pseudo-3D interpretation of the crustal structure in locations where MCS lines cross (see Selvans et al. (submitted)).

Discussion and conclusions

Obtaining sonobuoy FD models that accurately reproduce the main features being analyzed allows for further interpretation of MCS data than is otherwise possible. MCS data image shallow structure in great detail, but cannot directly measure layer velocities at depth. Matching the reflection times from each layer of known velocity in the FD model to the MCS image provides the velocity model details needed for depth migration of MCS data.

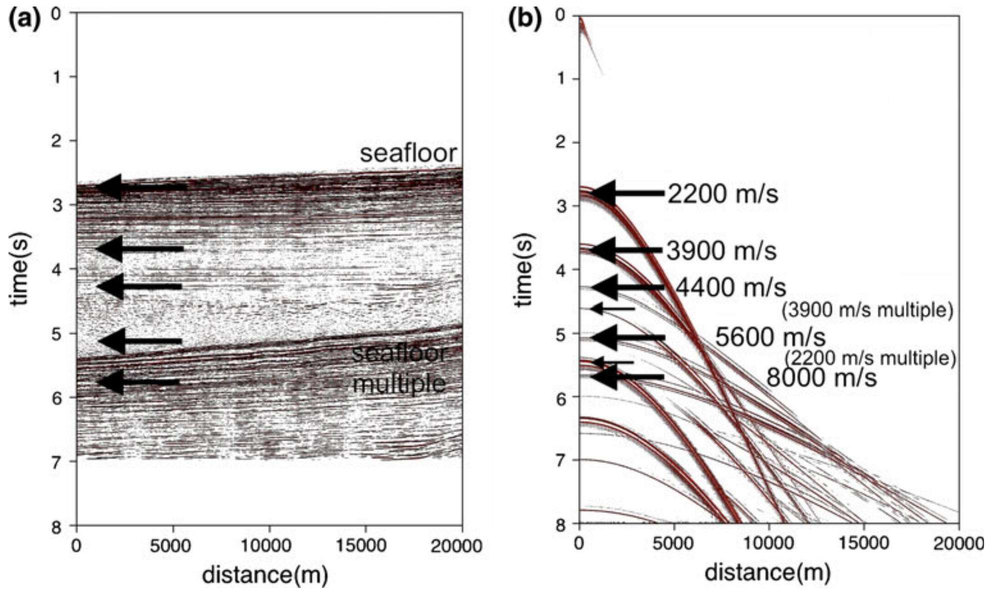

Fig. 9 Multi-channel seismic (MCS) data from the same location as Sonobuoy 1 (**a**) can be compared to the finite difference model (**b**) in order to determine the layer velocities within the reflection data. In the model, reflections from which head waves originate are labeled with large arrows, while multiples are labeled with small arrows. Note that the first layer at depth (3900 m/s) corresponds to sediment layers imaged in the MCS data, the 4400 m/s seems to be the top of the basement rock, 5600 m/s is within the basement rock, and the 8000 m/s layer is deep in the basement rock (below the seafloor multiple). Sonobuoy data also determine layer velocities below the acoustic basement, which is the limit to which reflection data can delineate subsurface structure

Modeling sonobuoys using FD has further potential for sonobuoy data analysis, since refinements in the model can reveal more details about the crustal structure. For instance, models could be altered to include head waves that occasionally come in late as a result of interbed multiples, further refining the 1D velocity model at that sonobuoy location. An elastic version of the model could be used to identify the converted phases in the data; an elastic FD model was run for Sonobuoys 1 and 4 to confirm that none of the head waves used in this analysis are converted phases.

Using an elastic FD model would also constrain the allowable range of shear wave velocities, based on the amplitude of those phases. Elastic FD models with both $v_s = (1/\sqrt{3})v_p$ and a $v_s = (1/2\sqrt{3})v_p$ have converted phases with amplitudes similar to those of the main head wave features of the sonobuoy; these phases are not observed in the data, indicating that shear wave velocities are small. Modeling the full range of possible shear wave velocities would help in identifying the crustal lithology, below the layers that can be mapped in MCS data over long distances and then tied to borehole results (e.g., Granot et al. 2010). Modeling of overlapping, and particularly reversed sonobuoys could additionally include lateral variations in layer velocity and layer thickness and a smoothly varying profile with depth, in order to determine whether these secondary effects are observed in the sonobuoy data.

We show the necessity of adjusting distance between traces for sonobuoy studies, since ocean currents and changing speed and direction of the ship can alter the direct wave slope such that it does not accurately reflect the sound speed in water of 1450 m/s (see Table 1). Prior to the correction, direct wave velocities were as much as 130 m/s higher (Sonobuoy 10) and 400 m/s lower (Sonobuoy 6) than the true sound speed in water. Directly measuring layer velocities at depth by applying a linear moveout to head waves in the uncorrected sonobuoy data will be off from the true velocities of those layers by similar amounts, resulting in errors larger than the measurement uncertainty of ±100 m/s in most cases (seven out of ten sonobuoys).

In order to use the reflected energy features in the sonobuoy analysis, this error cannot be corrected through a simple stretch in the time dimension, based on the best fit direct wave velocity, since a hyperbolic feature will not retain its proper shape if a linear stretch is applied. Generally, in cases where direct wave velocities differed from the true sound speed in water by more than the uncertainty of the velocity measurement, sonobuoy studies have reported inaccurate layer velocities.

Finally, our analysis is greatly aided by the experiment design. Collecting overlapping sonobuoy data along several MCS lines during research cruise NBP0701 allows us to construct 2D velocity models deep within the crust, to a greatest depth of 5.5 km below the seafloor. This depth is

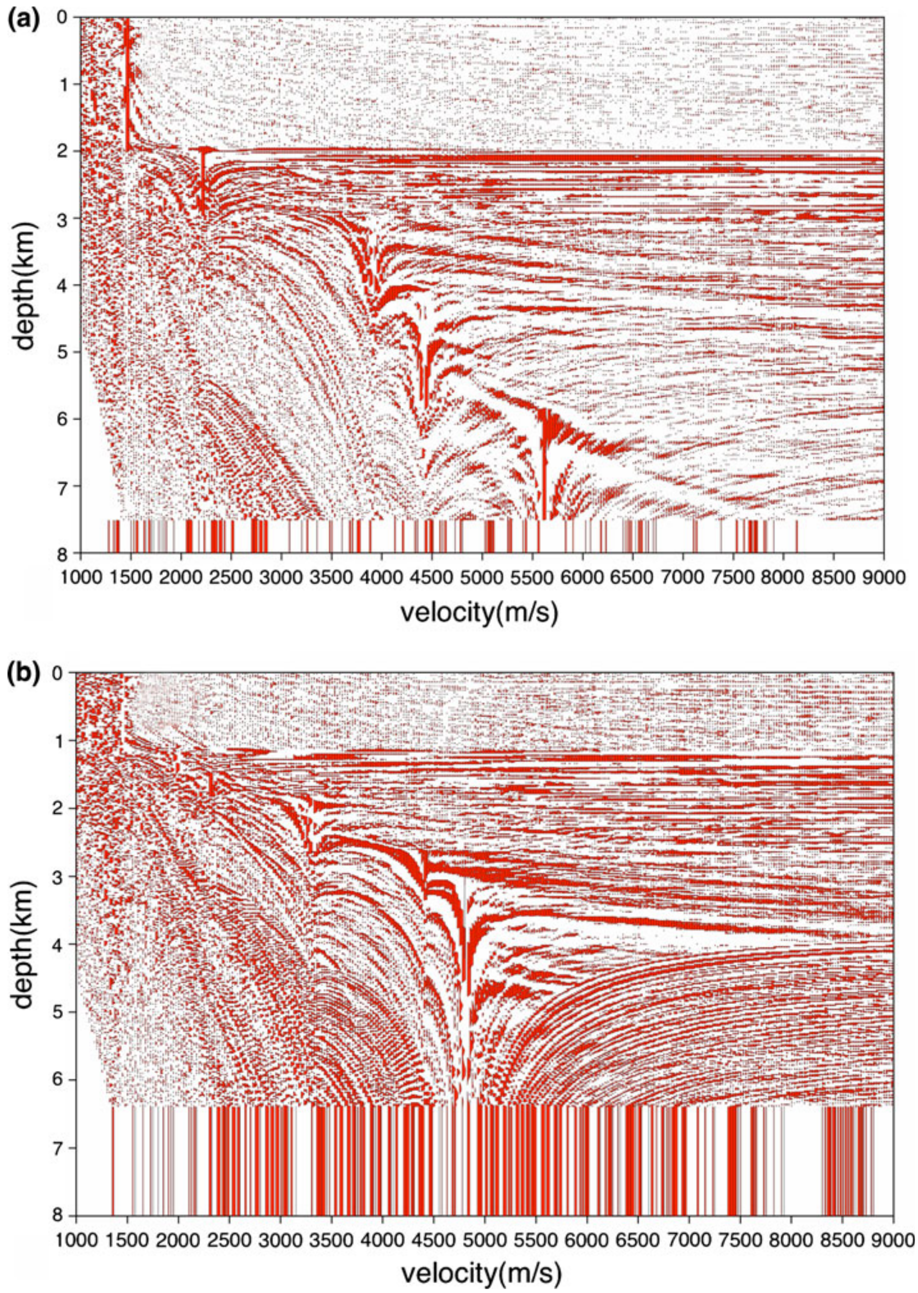

Fig. 10 Depth migration is applied to the distance-adjusted data for Sonobuoy 1 (**a**) and Sonobuoy 4 (**b**), using their respective 1D velocity models. Parallel bedding, as observed in the multi-channel seismic data (e.g., Fig. 8a), indicates a good model fit to the data; slight tilting in deeper layers indicates small lateral variations in velocity that cannot be properly modeled with 1D analysis

equivalent to 5.0 s travel time below the seafloor, deeper than the maximum depth of 2.3 s below the seafloor at which basement rock is imaged with MCS data (Granot et al. 2010). We regularly determine layer velocities of 4500–5000 m/s, at 2.0–3.0 km below the seafloor, and are able to see that the velocity contours fit to these data are at consistent depths along the 115 km of MCS Line 14. Being able to use the depth penetration and ability to directly measure layer velocities that sonobuoys provide in order to construct a 2D velocity model is a particularly useful analytical technique for locations such as the Ross Sea, where sea ice conditions preclude the use of a longer MCS streamer.

Having a deep 2D velocity profile along Line 14 of the NBP0701 MCS data allows us to begin examining the crustal structure along the margin between the Adare and Northern Basins (Fig. 11a). Two distinct features stand out. One is the unusually thin crust under Sonobuoy 1: the data reveal a maximum velocity of 8000 m/s at 5.5 km below the seafloor, which is interpreted as the Moho. Since there is 2.1 km of overlying sediment at this location (as indicated by comparing reflection time of the 4400 m/s layer in the FD model with that of acoustic basement as imaged with MCS data (see Fig. 9), and referring to the 1D velocity model of Sonobuoy 1 for the depth of that layer), the igneous crustal thickness is 3.4 km.

Although unusual, similarly thin crustal thicknesses are observed in the Ross Sea (4–27 km; see Anderson (1999) and references therein), and in continental margins of East Antarctica (<4 km in continental crust offshore from

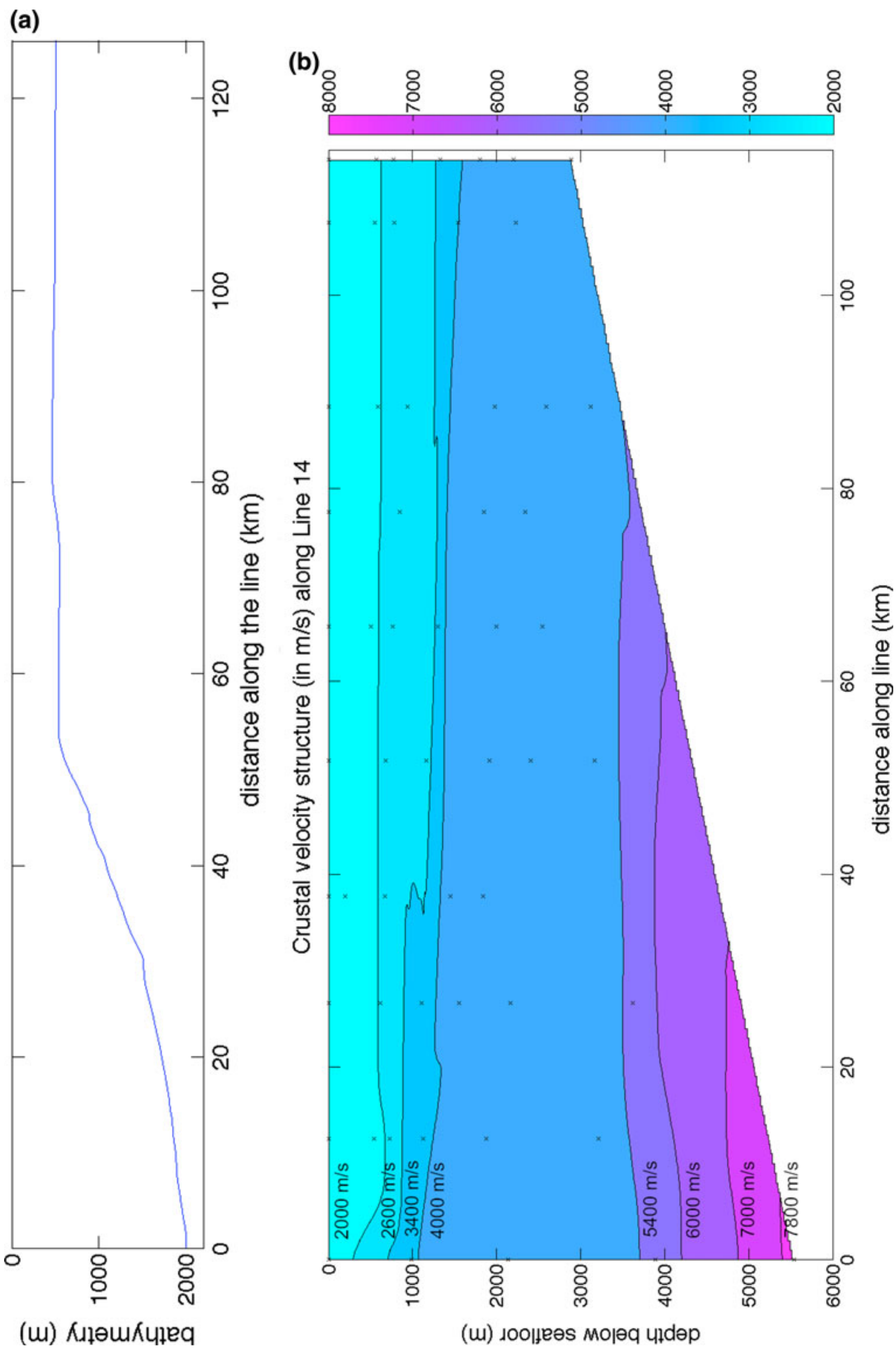

Fig. 11 The 2D velocity structure along Line 14 is determined to a depth of 5.5 km below the seafloor. Seafloor bathymetry (**a**) provides context for the 2D velocity model (**b**) interpolated from velocity models for ten overlapping sonobuoys (**c**). Layer interfaces observed in the data are indicated with an "x" (see Table 2 for details); vertical exaggeration is 1:8. Sediment layers (~2000–4000 m/s) thicken slightly from the Adare Basin (Sonobuoy 1) to the Northern Basin (Sonobuoys 5–10), as expected for moving up a shelf break onto a sediment-filled basin. Interestingly, deeper velocity contours remain approximately flat along the line, whereas if the shelf break were the transition from oceanic to continental crust, we would expect these contours to deflect down under the Northern Basin. Uncertainties on 2D velocity contours are ±200 m/s and ±100–500 m (increasing with depth)

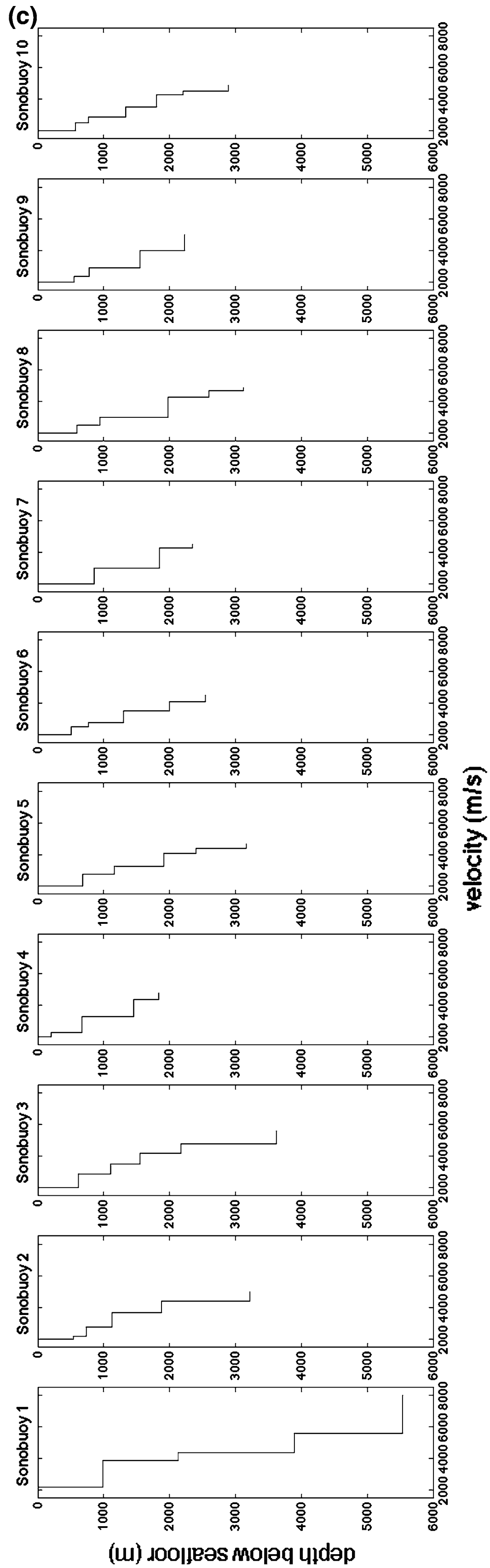

Fig. 11 continued

Wilkes Land (Close et al. 2009), and 4 km in oceanic crust offshore from Enderby Land (Stagg et al. 2004)). Results from processing all NBP0701 sonobuoys, including the interpolation of additional 2D velocity models, are discussed in the context of other tectonic settings similar to that of the Adare and Northern Basin region by Selvans et al. (submitted).

The other obvious feature is that all well-constrained velocity contours (2000–5400 m/s) are approximately flat, considering the 8:1 vertical exaggeration (Fig. 11), indicating that it is unlikely for there to be significant local relief on layer interfaces within the crust. Velocity contours at a transition between oceanic and continental crust would be expected to deflect downward under the thicker continental crust, suggesting that the continental shelf break between the Adare and Northern Basins (obvious in bathymetry) may not be the location of the transition between crustal types. As also suggested by Cande and Stock (2006), who discussed the possibility of the Northern Basin crust being ‘transitional,’ based on analysis of magnetic anomaly data, we find that the crustal structure may be continuous across the shelf break.

Acknowledgments We would like to thank Captain Mike Watson, the crew, and the Raytheon Polar Services Corporation technical staff on board the *Nathaniel B. Palmer*. This study was supported by National Science Foundation grants OPP04-40959 (S. Cande) and OPP-0440923 and OPP-0944711 (J. Stock and R. Clayton).

References

Anderson JB (1999) Antarctic marine geology. Camb University Press, Cambridge

Cande SC, Stock JM (2006) Constraints on the timing of extension in the Northern Basin, Ross Sea. 9th International symposium on Antarctic earth science proceeding

Clayton R, McMechan G (1981) Inversion of refraction data by wave field continuation. Geophysics 46:860–868

Close DI, Watts A, Stagg H (2009) A marine geophysical study of the Wilkes Land rifted continental margin, Antarctica. Geophysical J Int 177(2):430–450

Cochrane GR, De Santis L, Cooper AK (1995) Seismic velocity expression of glacial sedimentary rocks beneath the Ross Sea from sonobuoy seismic-refraction data, geology and seismic stratigraphy of the Antarctic margin. Antarctic Res Ser 68: 261–270

Cooper A, Davey F, Cochrane G (1987) Structure of extensionally rifted crust beneath the western Ross Sea and Iselin Bank, Antarctica, from sonobuoy seismic data. The Antarctic continental margin: geology and geophysics of the Western Ross Sea. Published by the Circum-Pacific Council of Energy and Mineral Resources, Earth Science Ser 5B:93–117

Ewing M, Heezen BC (1956) Some problems of Antarctic submarine geology. In: Carey AP, Gould LM, Hulbert EO, Odishaw H, Smith WE (eds) Antarctica in the IGY. Geophys Monogr 1, pp 75–81

Fowler CMR (1990) The solid earth: an introduction to global geophysics. Camb University Press, Cambridge

Geissler WH, Jokat W (2004) A geophysical study of the northern Svalbard continental margin. Geophysical J Int 158(1):50–66

Granot R, Cande S, Stock J, Davey F, Clayton R (2010) Postspreading rifting in the Adare Basin, Antarctica: regional tectonic consequences. Geochem Geophys Geosyst 11(8):Q08005. doi:10.1029/2010GC003105

Houtz R, Davey F (1973) Seismic profiler and sonobuoy measurements in Ross Sea, Antarctica. J Geophys Res 78(17):3448–3468

Jones EJW (1999) Marine geophysics. Wiley, London

Jones GD, Barton PJ, Singh SC (2007) Velocity images from stacking depth-slowness seismic wavefields. Geophys J Int 168(2):583–592

McMechan GA, Ottolini R (1980) Direct observation of a p-τ curve in a slant stacked wave field. Bull Seismol Soc Am 70(3):775

NBP0701 Data Report (2007) prepared by Ayers J. Available at http://www.marine-geo.org/tools/search/data/field/NBPalmer/NBP0701/docs/NBP0701Report.htm

Ritzmann O, Jokat W, Czuba W, Guterch A, Mjelde R, Nishimura Y (2004) A deep seismic transect from Hovgärd Ridge to northwestern Svalbard across the continental ocean transition: a sheared margin study. Geophys J Int 157(2):683–702

Selvans MM, Stock JM, Clayton RW, Cande S, Davey F, Granot R (submitted Sept. 30, 2011) Deep crustal structure of the Adare and Northern Basins, Ross Sea, Antarctica, from sonobuoy data. J Geophys Res

Shipp RM, Singh SC (2002) Two dimensional full wavefield inversion of wide aperture marine seismic streamer data. Geophys J Int 151(2):325–344

Stagg HMJ, Colwel J, Direen N, OíBrien P, Bernardel G, Borissova I, Brown B, Ishirara T (2004) Geology of the continental margin of Enderby and Mac. Robertson Lands, East Antarctica: insights from a regional data set. Marine Geophys Res 25(3):183–219

Tréhu A, Behrendt JC, Fritsch J (1993) Generalized crustal structure of the Central basin, Ross Sea, Antarctica. In: Damaske D, Fritsch J (eds) German Antarctic North Victoria Land Expedition 1988/1989. Bundesanstalt fur Geowissenschaften und Rohstoffe, Hannover, pp 291–311

Trey H, Cooper AK, Pellis G, della Vedova B, Cochrane G, Brancolini G, Makris J (1999) Transect across the West Antarctic rift system in the Ross Sea, Antarctica. Tectonophysics 301(1–2):61–74

Vireaux J (1986) P-SV wave propagation in heterogeneous media: velocity-stress finite difference method. Geophysics 51:889–901