Crustal S-wave velocity structure of the Main Ethiopian Rift from ambient noise tomography

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SUMMARY
Ambient noise tomography has been used to construct Rayleigh-wave group velocity maps covering the northern (NMER), central (CMER) and southern (SMER) parts of the Main Ethiopian Rift (MER). In addition, dispersion curves, extracted from the group velocity maps, have been inverted to obtain a quasi-3-D model of crustal shear wave velocities. In comparison to crustal structure on the Ethiopian Plateau, we find (1) lower shear wave velocities at all crustal depths beneath the Yerer-Tullu Welwel Volcanotectonic Lineament, (2) lower shear wave velocities throughout the MER at upper crustal depths (<10 km), (3) regions of lower shear wave velocities at mid- (10–20 km) crustal depths beneath the Wonji Fault Belt (WFB), in the transition between the NMER and CMER, and beneath the Silti-Debre zeit Fault Zone (SDFZ) on the western side of the CMER, (4) an offset in the velocity pattern at mid-crustal depths between the NMER and CMER coincident with the Boru-Toru Structural High (BTSH) and (5) little evidence for lower shear wave velocities at mid- or lower-crustal depths beneath the SMER. We attribute these findings primarily to along-strike changes in crustal composition, melt content and thermal structure resulting from the Cenozoic to recent magmatism, and also, at upper crustal depths (<10 km), to basin structure and fill. Our findings corroborate a magmatic plumbing model for the MER that shows two major zones of magmatic activity, one beneath the WFB and the other beneath the SDFZ. The shear wave velocity patterns in our model show good correlation with the depth extent of seismicity, upper-mantle seismic anomalies and seismic anisotropy, as would be expected if the along-strike changes in shear wave velocity reflect the thermal and compositional structure of the crust.

Key words: Surface waves and free oscillations; Seismic tomography; Continental tectonics: extensional; Crustal structure; Africa.

1 INTRODUCTION
The Main Ethiopian Rift (MER) has long been recognized as one of the best places to investigate continental rifting, and, in particular, rift processes leading to seafloor spreading (e.g. Makris & Ginzburg 1987; Ebinger & Casey 2001; Bastow & Keir 2011; Fig. 1). In this paper, we use ambient noise tomography to construct Rayleigh-wave group velocity maps covering most of the MER and some of the surrounding Ethiopian Plateau. Dispersion curves, extracted from the group velocity maps, have been inverted to obtain a quasi-3-D model of crustal shear wave velocities that provides new constraints on the along-strike variations in crustal structure.

The primary motivation for this study comes from a need to better understand the along-strike variability in MER crustal structure. Variations in magmatism and faulting along rift axes have been identified as potentially influential factors in rift development (e.g. Shillington et al. 2009). In the parts of the MER that have been well studied (i.e. the northern section and parts of the central section), faulting, magmatic segmentation and the magmatic alteration of the rifted continental crust is well documented (e.g. Hayward & Ebinger 1996; Ebinger & Casey 2001; Kendall et al. 2005; Casey et al. 2006; Beutel et al. 2010). Crustal structure beneath the southern section of the MER, as well as the southern part of the central section, in contrast, has not been as well studied (e.g. Bonini et al. 2005; Rooney et al. 2007; Keranen & Klemperer 2008), and therefore the along-strike variability of MER crustal structure is not fully known. This limits our understanding of the deeper rifting processes within the MER and the role those processes may play in the formation of spreading centres (e.g. Kendall et al. 2005; Keir et al. 2006; Rooney et al. 2007; Bastow et al. 2008).

The results presented in this paper characterize the shear wave velocity structure of the MER crust over a much larger area of the
and material of low Red Sea and Gulf of Aden at about the same time as the eruption of Africa commenced of Aden–EARS triple junction (Fig. 1). Separation between Arabia western parts, and forms the southern arm of the Red Sea–Gulf System (EARS), divides the Ethiopian Plateau into eastern and

The temporal development of faulting within the MER is debated. Wolfinden et al. (2004) suggested that rifting began ca. 25 Ma in the Lake Turkana region in Kenya and propagated northwards, while Buck (2006) and Rogers (2006) suggested that the rifting propagated to the south, beginning in the Afar. Bonini et al. (2005) proposed that rifting was heterogeneous in time and space beginning in the Afar ca. 20–25 Ma and the SMER ca. 20–21 Ma, however there may have been renewed extension in the SMER ca. 11 Ma (Pik et al. 2008). Bonini et al. (2005) suggested a southward migration of rifting within the NMER from the Afar Depression at ca. 10–11 Ma, consistent with Wolfinden et al. (2004), and the commencement of rifting in the CMER ca. 5 Ma. Keranen & Klemperer (2008) support the Bonini et al. (2005) scenario of rift development, but others have suggested that rifting within the SMER and CMER initiated between 18 and 15 Ma (Wolfgang et al. 1991; Ebinger & Casey 2001).

Studies using data from two seismic experiments, the Ethiopia Broadband Seismic Experiment (EBSE; Nyblade & Langston 2002) and the Ethiopia Afar Geoscientific Lithosphere Experiment (EAGLE; Maguire et al. 2003; Bastow et al. 2011), have greatly increased our knowledge of crust and upper-mantle structure beneath the MER. Seismic studies using EAGLE controlled-source seismic data revealed the existence of high P-wave velocity regions ($V_p > 6.6 \text{ km s}^{-1}$) beneath the Quaternary magmatic segments, which were attributed to cooled mafic bodies (Keranen et al. 2004; Mackenzie et al. 2005; Maguire et al. 2006). The ∼5 km s$^{-1}$ of low $V_p$ material imaged above the mafic bodies in these studies were attributed to sediments and volcanic rock (Keranen et al. 2004; Mackenzie et al. 2005).

Using travel times from local earthquakes recorded on the EAGLE network, Daly et al. (2008) inverted for models of $V_p$ and $V_p/V_s$ ratios beneath the MER to mid-crustal depths. Their result shows

2 BACKGROUND

The MER is located in the northern part of the East African Rift System (EARS), divides the Ethiopian Plateau into eastern and western parts, and forms the southern arm of the Red Sea–Gulf of Aden–EARS triple junction (Fig. 1). Separation between Arabia and Africa commenced ca. 30 Ma (Hofmann et al. 1997) along the Red Sea and Gulf of Aden at about the same time as the eruption of flood basalts ca. 31–29 Ma (Baker et al. 1996; Hofmann et al. 1997; Kieffer et al. 2004; Furman et al. 2006) and the initiation of plateau uplift ca. 30 Ma (Pik et al. 2003). The MER developed somewhat later, with the initiation of faulting no earlier than ca. 20–25 Ma (Wolfgang et al. 1991; Ebinger et al. 1993; Bonini et al. 2005).

The MER, based on its surface morphology, can be divided into three sections, northern (NMER), central (CMER) and southern (SMER; Fig. 1). The NMER and CMER are defined by major

border fault (BF) systems, namely the Ankober and Arboye BFs for the NMER, and the Fonko-Gurage and Asela BFs for the CMER. The Boru-Toru Structural High (BTHS, Fig. 1), together with a rotation in BF orientation from NE–SW to N–S, defines the boundary between the NMER and CMER.

Mmagmatic activity in the NMER and CMER began at ca. 12–10 Ma (Chernet et al. 1998; Wolfenden et al. 2004), which corresponds to a second phase of flood basalt and silicic volcanism found on the rift shoulders and surrounding plateaus (Kieffer et al. 2004; Wolfenden et al. 2004; Ayalew & Gibson 2009). From the Late Miocene (∼7 Ma), volcanic activity became more dominant on the floor of the MER basin compared to the rift shoulders (Wolfgang et al. 1990; Chernet et al. 1998; Furman 2007). The timing of this shift in volcanic activity away from the plateau and rift margin correlates with the relocation of extension from the BF to magmatic segments in the NMER and CMER, such as the Aluto–Gedemsa (A–G), Boset–Kone (B–K) and Fentale–Dofen (F–D) magmatic segments, and the Wonji Fault Belt (WFB) and Silti-Debre zeit Fault Zone (SDFZ; Fig. 1; Ebinger & Casey 2001; Casey et al. 2006).

Another prominent feature that may be related to the development of the MER is the Yerer-Tullu Wellel Volcanotectonic Lineament (YTML), an east–west trending structure to the west of the CMER that includes the Ambo fault (Fig. 1). Tectono-magmatic activity along the YTML since ∼12 Ma has been associated with the development of the MER (Abebe et al. 1998; Keranen & Klemperer 2008). The SMER extends south from Lake Awasa (∼7°N) into a broad zone of rifting in southern Ethiopia (Ebinger et al. 2000). There have been two main phases of basaltic magmatism within the SMER at ca. 45–35 Ma and ca. 19–11 Ma (Ebinger et al. 1993).

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high values of $V_p$ and $V_p/V_s$ ratios along the axis of the MER and the Debre Zeit chain of Quaternary volcanoes, suggestive of pervasive dyking and some partial melt beneath these areas.

Tomographic studies using teleseismic body waves have resolved low $P$- and $S$-wave velocity anomalies in the upper mantle beneath the MER and the Afar (Bastow et al. 2005; Benoit et al. 2006a,b; Bastow et al. 2008). Bastow et al. (2008) imaged zones of particularly low velocities beneath the CMER–NMER boundary (39° E, 9° N) that extend westwards beneath the Ethiopian Plateau. In addition, they showed abrupt increases in upper-mantle velocities from the low-velocity zones into the Afar and the SMER.

In receiver function studies using EBSE and EAGLE data, thinned crust (∼30 km) and high $V_p/V_s$ ratios (>2.0) were found beneath much of the CMER and NMER (Dugda et al. 2005; Stuart et al. 2005; Keranen et al. 2009; Cornwell et al. 2010). SKS splitting studies using the same data sets obtained axis-parallel fast directions of seismic anisotropy, which have been attributed primarily to fluid-filled cracks in the lithosphere (Kendall et al. 2005; Hammond et al. 2010). Analyses of shear wave anisotropy in the MER crust also support the presence of magmatically modified crust (Keir et al. 2005; Kendall et al. 2006; Bastow et al. 2010).

Seismicity recorded by the EBSE and the EAGLE experiments is mainly localized beneath the F–D magmatic segment, the Ankober BF, the eastern part of the CMER and the YTVL (Keir et al. 2006; Brazier et al. 2008). Results from other geophysical studies are consistent with results from seismic images of crustal structure in the MER. A magnetotelluric study (Whaler & Hautot 2006) imaged two high-conductivity bodies beneath the centre and western flank of the NMER at upper and mid-crustal depths, which correspond to the B–K magmatic segment and northern end of the SDFZ, respectively (Corti 2009). In a number of gravity studies (Mahatsente et al. 1999; Tiberi et al. 2005; Mickus et al. 2007), thinned and high-density crust in the MER has been modelled, which is consistent with seismic images showing thinned crust modified by magmatic intrusions.

3 DATA AND GROUP VELOCITY MEASUREMENT USING CROSS-CORRELATION METHOD

3.1 Data

The data used in this study come from the two temporary seismic networks mentioned previously (Fig. 1). The EBSE was conducted between 2000 March and 2002 February, and consisted of 27 broadband seismic stations deployed within the MER and the surrounding Ethiopian Plateau. The EAGLE phase I network was comprised of 30 broad-band stations, many of which were deployed in the MER and operated during 17 months starting in 2001 October 2001. The operation of the two seismic networks overlapped in time between 2001 October and 2002 February. In addition to data from the temporary stations, data from two permanent stations, FURI and ATD, were also used.

In general, cross-correlation functions (CFs) calculated between station pairs with a longer time-series length yield higher signal-to-noise ratios (SNRs; Bensen et al. 2007). Therefore, where possible, 1 yr of data were used for computing CFs. However, for obtaining CFs between the EBSE and EAGLE stations, only 4 months of data were available. Nonetheless, relatively good consistency in the timing of the maximum peaks was observed between the CFs obtained with different time lengths of data (Fig 2).

3.2 Calculation of CFs

CFs were obtained for all possible combinations of station pairs. To increase the SNR, data from two stations were cross-correlated using a 6 hr moving window with a 1 hr time-shift,
and then the resulting CFs were averaged. Before calculating CFs, windowed traces with obvious large glitches or many zeros (>40 per cent of entire trace) were removed. Following the processing procedure from previous studies (e.g. Kang & Shin 2006; Bensen et al. 2007), one bit normalization and spectral whitening were performed on the remaining data to reduce the effects of transient signals and broaden the available frequency band. To measure Rayleigh-wave group velocities, we used only the averaged CFs generated using more than 800 time windows (Fig. 3a).

### 3.3 Group velocity measurements

Interstation Rayleigh-wave group velocity measurements were made using the multiple filter technique (Dziewonski et al. 1969) in the period range of 6–30 s for each of the averaged CFs. To further increase SNR and reduce the effect of an uneven distribution of noise sources, averaged CFs were folded in half at zero time before making the dispersion measurement. A continuous ridge was tracked manually in the FTAN (Frequency Time ANalysis) diagrams (Levshin et al. 1989) to make group velocity measurements.

**Figure 3.** (a) Record sections showing ‘folded’ cross-correlation functions for all station pairs used in this study. The shaded area in grey indicates coherent arrivals of Rayleigh waves. (b) Frequency–time diagram of the Rayleigh wave obtained between stations DIYA and NEKE. Picked group velocities are superimposed with circles between 6 and 30 s period. (c) Map shows locations of stations and ray paths for the station pairs in (d). (d) Examples of dispersion curves for station pairs crossing major tectonic regions. The average dispersion curve for all station pairs is shown with the grey line.

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Figure 4. Ray-path distribution for selected periods (7, 14, 21 and 28 s). Total number of rays used is presented at the bottom of each map. The stations are shown by red triangles. The green lines indicate major border faults shown in Fig. 1 and the grey polygon encompasses the region used in modelling crustal Vs.

We rejected data during the picking of group velocities when a clear dispersion curve was not evident. For further analysis, we only used the group velocity measurements for station pairs where their interstation distances were greater than 12 times the corresponding period to satisfy the far-field assumption of the cross-correlation method (Bensen et al. 2007). Figs 3(c) and (d) give examples of dispersion measurements for six different ray paths. Approximately, the period ranges 6–10, 10–15 and 15–25 s have maximum sensitivities to shear wave velocity structure in the depth ranges of 5–15, 5–25 and 10–40 km, respectively. Paths through the Afar have higher group velocities at periods >16 s compared to other regions. A maximum number of 980 rays were obtained at 7 s period, and the number of rays at each period gradually decreases with increasing period (Fig. 4).

4 GROUP VELOCITY TOMOGRAPHY

Using a non-linear 2-D tomographic inversion technique (Rawlinson et al. 2008; Saygin & Kennett 2010), the measured interstation group velocities were inverted to obtain maps of Rayleigh-wave group velocities for periods between 6 and 30 s. This inversion method uses a fast marching method (FMM; Rawlinson & Sambridge 2004a,b) that tracks rays for the initial and updated velocity distributions before every inversion step. This allows for wave propagation effects in regions of rapidly varying velocities to be accounted for. In the inversion process, a discretized grid with 0.2° × 0.2° cells was used, and uncertainties in the group velocity maps obtained were assessed by a bootstrap resampling technique (Efron & Tibshirani 1991). A series of 200 tomographic inversions were performed for each wave period, and in each one three-quarters of the total number of rays available for each period were randomly selected and used. The maximum standard error for each group velocity map in the period range of 6–30 s was found to be within the range of 0.046–0.078 km s⁻¹. Optimal values of damping and smoothing of 5.0 and 1.0, respectively, were selected for our final model after testing several combinations of values ranging between 0 and 1000.

To investigate the resolution of our path geometries for each period, checkerboard tests were performed. Each checker was assigned a constant velocity of +0.2 km s⁻¹ or −0.2 km s⁻¹ from the observed mean group velocity of the corresponding period. Two different checker sizes (0.6° × 0.6° and 1° × 1°) were used to test the sensitivity to different size structures. Synthetic arrival times for all station pairs represented in the real data set were calculated and then inverted using the same procedure and parameters as for the real data. Fig. 5 shows selected results for periods of 7, 14, 21 and 28 s. In spite of the smoothing and damping that were applied during the inversion, the recovered group velocity maps show good resolution for the NMER and CMER, and the northern part of the SMER for all periods and checker sizes. The region with best resolution for all periods is outlined by the black polygon in Fig. 5.
Fig. 5 shows inversion results of group velocity for selected periods. At a period of 7 s, regions of lower velocities are observed over most of the CMER and parts of the NMER and SMER. At a period of 14 s, higher velocities are observed beneath the eastern side of the CMER, and lower velocities are located beneath the western side. Some regions of lower velocity are also seen in the NMER and SMER. At periods of 21 and 28 s, lower velocities can be seen widely distributed across the NMER, CMER and SMER, and the YTFL.

5 3-D SHEAR WAVE VELOCITY MODEL

5.1 Setup of 1-D depth inversion

To obtain a model showing the 3-D distribution of S-wave velocities in the crust, we first inverted for a 1-D S-wave velocity profile for each node shown in Fig. 6 similar with other studies (e.g. Bensen et al. 2009; Choi et al. 2009). Dispersion curves of Rayleigh-wave group velocity used in the inversion for each node were constructed.
Crustal structure of the Main Ethiopian Rift

Figure 6. Rayleigh-wave group velocity maps for selected periods of 7, 14, 21 and 28 s. Dots in 7 s map indicate inversion nodes to obtain the 3-D shear wave velocity model described in the text. Major faults (black lines) are the same as in Fig. 1, and the grey polygon encompasses the region used in modelling crustal V$_s$.

by combining group velocities between 35 and 60 s period from the revised Pasyanos & Nyblade (2007) model with group velocities between 6 to 30 s period from our maps (Fig. 6).

The method of Julia et al. (2000) was used to invert the group velocity curves for 1-D velocity profiles. This method uses a linearized iterative inversion technique to minimize weighted least-square norms. The norms consist of a model roughness norm and a matrix difference norm between inverted and preset model parameters. The method was developed for joint inversion of receiver functions and surface wave dispersion measurements. The influence factor that controls the trade-off between the two different data sets was set to zero for the receiver functions.

The inversion using only surface wave group velocities tends to produce smoothly varying velocity models with depth because the depth sensitivity of the group velocity for a specific period is fairly broad. Thus, velocities above and below sharp boundaries (e.g. Moho discontinuity) tend to be averaged across the boundary.

To better estimate velocities above and below the Moho, we have constrained Moho depths in the inversion using crustal thickness estimates from Dugda et al. (2005) and Keranen et al. (2009). Moho depth was constrained by imposing a 2 per cent increase and reduction in velocity above and below the Moho, respectively, compared to velocities reported just below and above the Moho by Keranen et al. (2009). The velocity constraints were imposed on 1-km-thick layers below and above the Moho only.

The 1-D models were parametrized using 1-km-thick layers for the top 5 km of the crust, 4-km-thick layers from 5 km depth to 1 km above the Moho, 1-km-thick layers just above and below the Moho and then 8-km-thick layers for the mantle to a maximum model depth of 70 km. Velocities for the starting model are similar to average crustal velocities for the region reported by Dugda et al. (2005) and Keranen et al. (2009). The $V_p/V_s$ ratio for the inversions was fixed to the values reported by Dugda et al. (2005) and Stuart et al. (2006) for the crust and mantle. Further details about the starting model are presented in Fig. 7.

To determine the influence of the starting velocities on the 1-D inversions, the initial values were randomly perturbed with a maximum ±10 per cent variation of each value, and a standard deviation of each layer was measured using inverted models from 200 runs of the inversion with the perturbed initial models. To determine the sensitivity of the constrained Moho depths on the inversion results, we additionally performed inversions constraining the Moho to be either 5 km shallower or 5 km deeper than the depths reported by Dugda et al. (2005) and Keranen et al. (2009).

In Fig. 8, modelling results from the 1-D inversions are presented for selected nodes. All of the synthetic dispersion curves
show a good fit to the observations with low mean value (0.02 km s$^{-1}$) of the rms misfit for all inversions. The maximum values of error from the perturbed initial models for each layer do not exceed 0.05 km s$^{-1}$. The maximum errors were obtained for the deepest model layers around depths of 70 km. Models obtained by perturbing the Moho by 5 km depth are also shown for comparison. Velocities above the Moho remain essentially the same whether or not the Moho depth is altered.

5.2 3-D shear wave velocity model

A model showing a 3-D distribution of shear wave velocities was generated from the inverted 1-D models for all 400 nodes (Fig. 6). The 3-D model was constructed by using a linear interpolation to obtain velocities at 1 km depth intervals for each 1-D model, and then smoothing the velocities at each 1 km depth interval by applying a minimum curvature spline (Smith & Wessel 1990) across all of the nodes. Fig. 9 shows horizontal slices through the model at depths of 5, 10, 15, 20 and 27 km. The slice at 5 km depth shows lower shear wave velocities throughout most of the SMER and CMER, as well as the eastern side of the NMER and the YTFL. The lowest velocities are found in the CMER. At 10 km and 15 km depths, low velocities are found beneath the western part of the CMER and eastern part of the NMER, with an offset in the velocity pattern coinciding with the BTSH. In this depth interval, the lowest velocities are found beneath the western side of the CMER coinciding with the location of the SDFZ and extending to the west under the YTFL. At 20 and 27 km depths, lower velocities are found beneath most of the NMER, and the region of low velocities beneath the western part of the CMER and SDFZ remains prominent. These velocity patterns are also apparent in vertical slices through the 3-D model (Fig. 10). Prominent features seen on the vertical sections include the region of lower velocities beneath the western side of the CMER (profile A–A’), at all depths, a region beneath the eastern side of the CMER/NMER transition with low velocities extending to mid-crustal depths (profile B–B’), a region beneath the centre part of the NMER with low velocities extending to mid-crustal depths (profiles C–C’ and D–D’), and a region of low velocities extending to mid-crustal depths beneath the YTFL and SDFZ (profile E–E’).

6 DISCUSSION

To summarize, the main findings of this study are (1) a region of lower shear wave velocities at all crustal depths beneath the YTFL; (2) lower shear wave velocities throughout the MER at upper crustal depths (<10 km); (3) prominent regions of lower shear wave velocities at mid- (10–20 km) crustal depths beneath the WFB on the eastern side of the MER, in the transition between the NMER and CMER and beneath the SDFZ on the western side of the CMER; (4) an offset in the velocity pattern at mid-crustal depths between the NMER and CMER coincident with the BTSH; and (5) little evidence for lower shear wave velocities at mid- or lower-crustal depths beneath the SMER. In this section, we discuss the significance of these findings for understanding along-strike variations in the thermal and compositional structure of the MER crust.

At upper crustal depths (<10 km), there appears to be little resolvable along-strike variations in our model. We attribute the lower velocities throughout the MER to magmatically perturbed upper crust and basin fill. Moving to the mid-crust, as described in Section 2, both Daly et al. (2008) and Keranen et al. (2004) found regions of high $V_p$ (6.3–6.5 km s$^{-1}$) beneath the NMER and CMER that correlate well with the locations of the Quaternary magmatic segments. Mackenzie et al. (2005) also imaged high $P$-wave velocities beneath the B–K magmatic segment (Fig. 1), and Daly et al. (2008) reported high $V_p/V_s$ ratios (1.80–1.84) associated with the high $P$-wave velocity regions. Our model shows relatively low shear wave velocities in the regions of high $V_p$ in the model by Daly et al. (2008), and thus may appear to be discrepant with the $P$-wave models. However, given the high $V_p/V_s$ ratios reported by Daly et al. (2008), the low $V_s$ and high $V_p$ values are expected. For example, in the B–K magmatic segment, converting a $V_p$ of around 6.5 km s$^{-1}$ in the Daly et al. (2008) model using their reported $V_p/V_s$ ratios yields a $V_s$ of 3.3–3.4 km s$^{-1}$, which is similar to the velocities at 10 and 15 km depths in Fig. 9.

Keranen et al. (2004) and Daly et al. (2008) attributed the high $V_p$ anomalies, together with positive gravity anomalies, to solidified mafic intrusions at mid- to upper-crustal depths. Furthermore, Daly et al. (2008) and Beutel et al. (2010) noted that if the somewhat higher $V_p/V_s$ ratios (>2) for the entire rifted crust (Dugda et al. 2005; Stuart et al. 2006) are considered together with the presence of conductive bodies at upper- and mid-crustal depths (Whaler & Hautot 2006) and increased crustal seismic anisotropy from fluid-filled cracks (Keir et al. 2005; Kendall et al. 2006; Bastow et al. 2010), then the mafic intrusions could possibly be associated with some partial melt in the mid-crust, or else that partial melt could exist below the intrusions in the lower crust. Such an interpretation can account for the high $V_p/V_s$ ratios, the relatively low $V_s$ values, and the high conductivity in the regions of high $V_p$. It has been often suggested that small variations in the volume of melt and...
temperature have greater effects on $V_s$ and electrical conductivity than $V_p$ and density (e.g. Mavko 1980; Roberts & Tyburczy 1999; Goes et al. 2000; Priestley & McKenzie 2006), which are primarily sensitive to compositional variations (e.g. Kern 1982; Christensen & Mooney 1995; Christensen 1996). A small increase in melt fraction and temperature within or just below the mafic intrusions may thus significantly decrease $V_s$ and increase conductivity while not greatly affecting $V_p$. Consequently, the along-strike variability in our model showing relatively low shear wave velocities at mid-crustal depths in regions of higher $P$-wave velocities can be understood in terms of the magmatic segmentation within the MER.

Another significant along-strike variation at mid-crustal depths in our model is found between the NMER and CMER. Here our model shows a region of higher velocities beneath the BTSH that coincides with an offset in the velocity pattern between the NMER and CMER, most prominently seen on Figs 9(b) and (c) at depths of 10 and 15 km. The tectonic importance of the BTSH to the temporal development of the MER is uncertain. The BTSH has been interpreted as a transfer zone formed by north-to-south propagation within the MER (Bonini et al. 2005), and as a breached structure by the joining of the southern Red Sea rift with the EARS (Rooney et al. 2007). Regardless of its origin, our model indicates that it is a prominent crustal feature extending to at least mid-crustal levels.

Other along-strike variations at mid- and lower-crustal depths within the CMER include (1) higher velocities to the south of the southernmost Quaternary magmatic centre (A–G, Fig. 1) along the WFB, and (2) the persistence of low velocities through the crust beneath the SDFZ. These variations can be explained by a magmatic plumbing model for the MER based on geochemical studies (Pecceillo et al. 2003, 2007; Rooney et al. 2005, 2007, 2011). According to this model, along the WFB magmas are stored for less time and fractionate dominantly in the upper $\sim$5 km of the crust before their eruption. This is consistent with the lower velocities observed in the upper crust south of the A–G magmatic segment (Fig. 9), which are absent at greater depths. Conversely, in the SDFZ, magmas fractionate over a broader depth range, and the magmatic bodies are imaged as lower velocity anomalies extending throughout the crust (Fig. 9).
Figure 9. Shear wave velocity maps for selected depths. Depth is indicated in the lower right corner of each map. Dashed lines show locations of vertical profiles in Fig. 10. The white squares in panel (a) indicate node points of 1-D models shown in Fig. 8. Earthquake epicentres from Keir et al. (2006) within ±2.5 km of the depth of each model slice are shown with open circles.
Moving further to the south, there also appears to be along-strike variability between the SMER and other sections of the MER. The higher shear wave velocities at mid- and lower-crustal depths in the SMER compared to the NMER and CMER indicates that the SMER crust is probably not being modified by magmatism to the extent that is occurring beneath the CMER and NMER. In particular, shear wave velocities throughout the mid- and lower crust within the SMER are significantly higher than beneath the perturbed regions of the SDFZ and WFB, consistent with the lesser amounts of Quaternary-Recent volcanism in the SMER, and especially the absence of basaltic fissure eruptions since \(\sim 650 \text{ Ka}\) (Woldegabriel et al. 1991; Ebinger et al. 1993; Abebe et al. 2007). Nevertheless, recent volcano deformation observed within the SMER suggests the presence of some melt within the upper crust (Biggs et al. 2011). Therefore, part of the lower velocity anomalies within the SMER at upper crustal depths (Fig. 9a) may reflect the presence of shallow magma reservoirs, but the relatively high velocities in the middle crust indicate that the magmatic feeder systems are either sufficiently narrow or that the conduits are sufficiently separated such that they cannot be resolved.

Much of the along-strike variability in our model has been attributed to thermally modified crust. As a check on the extent to which this is the case, we examine the depth distribution of seismicity in relation to crustal shear wave velocities. It is well established that the depth extent of seismicity is strongly influenced by crustal temperatures (Byerlee 1968; Brace & Kohlstedt 1980; Ranalli 1987; McKenzie et al. 2005). The cross-sections in Fig. 10 reveal that to a first order there is good correlation between the depth extent of seismicity and the shear wave velocity pattern. For example, on profile D–D’, lower velocities at mid-crustal depths are found beneath the B–K magmatic segment compared to the A–G and F–D segments on either side. There is significantly less seismicity beneath
the B–K segment than beneath the neighbouring segments, and the seismicity extends only to a depth of about 10 km, compared to 15–20 km beneath the adjacent segments. As mentioned, the increase in crustal anisotropy beneath the B–K segment and the overall higher crustal \( \frac{V_p}{V_s} \) ratios in this region of the MER are consistent with the lower shear wave velocities seen in our model.

Another good example is seen in profile E–E', where there is a significant change in the shear wave velocities going from the NW Plateau across the Ankober BF and into the Afar. Beneath the NW Plateau, where velocities at mid-crustal depths are about 3.5–3.7 km s\(^{-1}\), seismicity extends to about 25 km depth, whereas just to the east of the Ankober BF, velocities at the same depths are 3.3–3.5 km s\(^{-1}\) and the seismicity extends only to a depth of about 15–20 km.

The aforementioned regions of lower velocities within the NMER and the CMER crust, and higher velocities within the SMER crust, show a parallel pattern to uppermost mantle seismic velocities, where lower velocities have been found at 75 km depth beneath the CMER and NMER compared to the SMER and Afar (Bastow et al. 2008, 2010). The similar patterns in crustal and uppermost mantle velocities can be explained by the initiation ages of the various sections of the MER, with rift initiation in the Afar and the SMER beginning at least 10 Ma earlier than in the CMER and NMER. It has been suggested by Bastow et al. (2008, 2010) that the difference in the initiation of rifting along strike could lead to larger melt volumes beneath the CMER and NMER than in southern Afar and the SMER due to more recent decompression melting. In particular, the prominent crustal low velocities beneath the B–K segment and the northern part of the SDFZ coincide with the most intense zones of melt, as well as the lowest velocities in the uppermost mantle (Bastow et al. 2005, 2008) and the maximum SKS delay times (Kendall et al. 2005; Hammond et al. 2010). The B–K segment and SDFZ also exhibit volumetrically significant Quaternary magmatism and less faulting compared to the adjacent magmatic segments (Woldegabriel et al. 1990; Rooney et al. 2005, 2011; Abebe et al. 2007). Thus, the along-strike variations in crustal shear wave velocities in our model not only reflect the along-strike loci of magmatism and faulting within the crust, but they also appear to be consistent with regions of greater melt extraction from the mantle.

7 SUMMARY

We have used ambient noise tomography to construct Rayleigh-wave group velocity maps covering most of the MER and some of the surrounding Ethiopian Plateau, and from these maps, dispersion curves have been extracted and inverted to obtain a quasi-3-D model of shear wave velocities for the crust. Given the good correlation between our model results and those from previous studies, we adopt a similar approach and argue that at mid- and lower-crustal depths, velocity variations in our model primarily reflect modifications to the crust by magmatism through changes in composition, melt content and thermal structure. At upper crustal depths (<10 km), velocity variations in our model also likely reflect basin structure and fill.

Our model provides new information on crustal structure over a significant portion of the MER. At mid- and lower-crustal depths in the CMER, the main feature in our model is higher velocities to the south of the southernmost Quaternary magmatic centre along the WFB, and the persistence of lower velocities through the crust beneath the SDFZ. Given the similar velocities at mid-crustal depths beneath the SDFZ compared to the B–K magmatic segment, we interpret the region of lower velocities beneath the SDFZ as an area modified by mafic intrusions, elevated temperatures and possibly some partial melt. This interpretation is consistent with the magmatic plumbing model for the MER proposed by Rooney et al. (2007, 2011), which shows two major zones of magmatic activity, one beneath the WFB and the other beneath the SDFZ.

Shear wave velocities throughout the mid- and lower crust under the SMER are significantly higher than beneath the perturbed regions of the SDFZ and WFB. This is consistent with the paucity of Quaternary volcanism in this part of the MER. Another important feature in our model is the region of higher velocities beneath the BTH that coincides with an offset in the velocity pattern at mid-crustal depths between the NMER and CMER, most prominently seen at depths of 10 and 15 km. The shear wave velocity pattern in our model shows good correlation with the depth extent of seismicity, upper-mantle seismic anomalies and seismic anisotropy, corroborating our interpretation that shear wave velocity variations in the MER crust reflect to large extent along-axis variations in the thermal and compositional structure of the crust.

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**REFERENCES**


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