Crustal velocity structure across the Main Ethiopian Rift: results from two-dimensional wide-angle seismic modelling

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SUMMARY

We present the results of velocity modelling of a recently acquired wide-angle seismic reflection/refraction profile across the Main Ethiopian Rift. The models show a continental type of crust with significant asymmetry between the two sides of the rift. A 2- to 5-km-thick layer of sedimentary and volcanic sequences is modelled across the entire region. This is underlain by a 40- to 45-km-thick crust with a c. 15-km-thick high-velocity lowest crustal layer beneath the western plateau. This layer is absent from the eastern side, where the crust is 35 km thick beneath the sediments. We interpret this layer as underplated material associated with the Oligocene flood basalts of the region with possible subsequent addition by recent magmatic events. Slight crustal thinning is observed beneath the rift, where Pn velocities indicate the presence of hot mantle rocks containing partial melt. Beneath the rift axis, the velocities of the upper crustal layers are 5–10 per cent higher than outside the rift, which we interpret as resulting from mafic intrusions that can be associated with magmatic centres observed in the rift valley. Variations in seismic reflectivity suggest the presence of layering in the lower crust beneath the rift, possibly indicating the presence of sills, as well as some layering in the proposed underplated body.

Key words: crustal structure, Ethiopia, refraction seismology, rifts.

INTRODUCTION

The process of continental break-up is fundamental to the development of the Earth's surface geology. The embryonic stage, the continental rift, cannot fully reveal the nature of the complex strain accommodation during break-up. In addition, the ultimate result of rifting is a volcanic or non-volcanic continental margin that are regions in which both the thermal response of the process has long since decayed away, and the structural and stratigraphic records of the separated margins are generally buried beneath thick post-rift sequences. In order to test geodynamic models of the break-up process it is essential to image magmatic modification of the crust/mantle, and to determine crust/mantle strain at the moment when there is a transfer of extension accommodation from faulting to magma supply. The Ethiopian Rift provides a unique opportunity to study the transition between continental rifting to the south in Kenya and new seafloor spreading to the north in the onshore extension of the Red Sea and Gulf of Aden in Afar (Fig. 1). While the Kenya Rift has been the subject of numerous geophysical investigations (e.g. Prodehl et al. 1994; Khan et al. 1999) and pioneering seismic studies have provided near 1-D models of the crust and upper mantle beneath Afar (Berckhemer et al. 1975; Makris & Ginzberg 1987), very little is known about the crustal structure beneath the transitional northern Main Ethiopian Rift (MER). EAGLE (the Ethiopia Afar Geoscientific Lithospheric Experiment) is a multidisciplinary project designed to investigate crust and mantle structure and the processes of strain accommodation beneath this transitional rift at the moment of continental break-up (Maguire et al. 2003). The project included a large controlled source seismic refraction and wide-angle reflection study consisting of one cross-rift and one rift-axial refraction profile as well as an areal 2-D array, the latter designed to provide a 3-D tomographic seismic velocity image of the upper crust around the intersection of the two profile lines (Fig. 2). We present results from velocity modelling of the cross-rift refraction profile, which traverses the MER immediately southwest of the splayed opening into the Afar depression. The specific objectives for this profile were: to provide estimates of total crustal thinning across this transitional sector of the rift; to assess the role of basement in the location of major faults and magmatic segments; and to determine whether significant underplating takes place during rifting.

TECTONIC AND GEOLOGICAL SETTING

The NNE–SSW trending MER transects the uplifted Ethiopian plateau, which is believed to have developed in response to the impact of a Palaeogene mantle plume head on the base of the lithosphere (Schilling et al. 1992; Ebinger & Sleep 1998). This plateau is
covered by extensive flood basalts that have been dated at 31–29 Ma (e.g. Baker et al. 1996; Hofmann et al. 1997) almost co-eval with opening of the Red Sea and Gulf of Aden rifts by c. 30 and 28 Ma, respectively (Wolfenden et al. 2004). Rifting within the southern and central MER is believed to have occurred later, between 18–15 Ma (WoldeGabriel et al. 1990) and in the northern MER at c. 11 Ma with formation of the Arboye border fault (Wolfenden et al. 2004) (see Fig. 2). To the north, within the structurally complex Afar depression, the MER intersects the Gulf of Aden and Red Sea rifts to form a triple junction between the Nubian, Arabian and Somalian plates. Structural and stratigraphic patterns indicate a migration of extensional strain after 2.5 Ma from the border faults to a narrow 20-km-wide subaxial zone of narrow near north–south trending zones of aligned eruptive centres cut by small offset faults and dykes arranged in a right stepping en echelon pattern (Boccaletti et al. 1998; Wolfenden et al. 2004) (see Fig. 2). Bilham et al. (1999) have shown from GPS measurements that 80 per cent of the current rate of extension of 4 mm yr$^{-1}$ is concentrated in these magmatic segments which have been interpreted by Ebinger & Casey (2001) as zones of intensive dyke injection.

The Pan-African metamorphic crustal basement with a strong N–S fabric is overlain by Carboniferous—Cretaceous sediments followed by the dominantly Oligocene flood basalts, based on rare outcrops in the central MER and good exposures in the Southern MER (Davidson 1983; Ebinger et al. 1993). Within the rift valley are a series of asymmetric sedimentary basins, for example the Adama basin, bounded on one side by steep border faults (Ebinger & Casey 2001) and containing Pleistocene volcanic, volcaniclastic and lacustrine strata which overlie the Miocene—Pliocene felsic and mafic sequences of the Kesem and Balchi formations (Wolfenden et al. 2004). These sequences also extend beyond the western rift escarpment into the Bishoftu embayment around 9°N and onto the eastern escarpment (Fig. 3).

**Previous geophysical work**

Previous wide-angle seismic surveys within the Ethiopian Rift consist of six refraction profiles (Berckhemer et al. 1975; Makris & Ginzberg 1987); five within the Afar depression and a sixth trending east–west on the plateau, extending into the rift south of the current profile (Line 1 in Fig. 2, inset). Within the Afar region a highly thinned crust was modelled with low sub-Moho velocities of 7.4–7.5 km s$^{-1}$ interpreted as high-temperature upper mantle material. The sixth profile beneath the plateau indicated a c. 38-km-thick crust thinning to the west and underlain by normal mantle with a velocity of 8.1 km s$^{-1}$ although Berckhemer et al. (1975) noted that available gravity data required the presence of a high-density lower crustal body beneath the rift escarpment.

Mahatsente et al. (1999) derived crustal structure beneath the rift between 6.5° and 8.5°N from 3-D modelling of gravity data. This indicated crustal thinning from 38–51 km beneath the plateaux to ≤31 km beneath the rift axis with a large dense intrusion in both the lower-and upper crust following the magmatic segments in the centre of the rift. The intrusion includes several lateral discontinuities in an east–west direction and is suggested to originate from partial melting in the lithosphere.

Recent receiver function work by Ayele et al. (2004) from data recorded at the FURI broad-band station near Addis Ababa suggests a c. 40-km-thick crust on the margin of the rift near Addis Ababa with anomalously low mantle lithospheric velocities and a possible lithosphere/asthenosphere boundary at a depth of 90 km.

**Data acquisition**

A 400-km-long wide-angle reflection/refraction profile (Fig. 2) was acquired in January 2003 as part of project EAGLE (Maguire et al. 2003). The NW–SE striking profile traversed the MER between the Blue Nile gorge in the Ethiopian Highlands and the Bale Mountains on the eastern part of the plateau. Within the rift the profile crossed the Boset magmatic centre within the asymmetric Adama basin, bounded by the Arboye border fault.

Eight in-line borehole shots of between 1 and 2 tonnes of TNT with a nominal shot spacing of 50 km were recorded on 431 seismic recorders with an average station spacing of 1 km. The recorders included 92 three-component Guralp CMG-6TD seismometers (bandwidth 30–100 s) which had been deployed every 4–5 km to collect earthquake data over a 1- to 2-month period leading up to the project. These were augmented with 339 single-component Retek ‘Texan’ recorders deployed only for the period of the controlled source survey. Data were recorded with a sample rate of 100 sps for both instrument types. GPS timing was used for the shots with shot and instrument locations measured using multiple GPS readings.

**Data modelling**

Phase identification and picking

Data quality is extremely good with substantial energy propagation along the entire length of the profile. Fig. 4 shows three examples of seismic sections from SP12 on the northwest plateau, SP14 on the western edge of the rift and SP17 at the southeast end of the profile. Two seismic phases ($P_s$ and $P_{st}$) with turning points in the
sedimentary and volcanic layers are observed in the seismic sections for SP14 near the rift escarpment, whereas only a single such phase is observed on the sections for SP12 and SP18 on the plateaux. On a few of the sections it was possible to pick a reflection within and at the base of the sediments but in most cases such phases could not be identified. The sedimentary phases are followed by a $P_g$ phase with turning points in the upper crust, which can be clearly observed with an apparent velocity of c. 6 km $s^{-1}$ on all the sections. A series of intracrustal reflected phases ($P_{i1}P$, $P_{i2}P$) and a generally strong, reflected phase from the Moho ($PmP$) are correlated on all record sections. On record sections for shots on the western side of the rift zone, $PmP$ is preceded by a weak reflected phase ($P_xP$) from the top of a high-velocity lower crustal (HVLC) layer, which gives rise to a diving wave ($P_x$) with an apparent velocity of c. 7.4 km $s^{-1}$. A weak, low-amplitude mantle diving wave ($P_n$) is observed on only a few of the sections. Phase identification was confirmed through the

Figure 2. Topographic map of the Main Ethiopian Rift showing the location of the EAGLE controlled source experiment. Major border faults are shown by the white lines and rift magmatic segments are indicated by the red hatch regions. BE = Bishoftu Embayment. EAGLE shotpoints are shown by yellow stars and the locations of seismic recorders are shown by triangles. Addis Ababa is in the vicinity of SP42. Inset shows the location of previous wide-angle seismic profiles (Berckhemer et al. 1975) in relation to the EAGLE profiles (grey lines).
checking of reciprocal traveltimes, which proved especially useful in identifying the $P_xP$ and $P_mP$ phases. A weak reflected phase from the upper mantle has been identified behind the $P_n$ phase on a few sections. Traveltimes were picked on shot gathers using ZP, an interactive plotting and picking software package (B. Zelt personal communication 2003). Picks were assigned a picking error based on the relative signal to noise ratio (Table 1).

Traveltime modelling

Traveltime modelling was conducted using a combination of forward modelling and inversion with the RAYINVR code of Zelt & Smith (1992). In order to reduce the errors associated with seismic modelling assuming a flat earth, which can become noticeable for long profiles (Gorman 2002), a new local Cartesian coordinate system was defined for the project area and model offsets calculated based upon this system. Therefore, the curvature of the earth appears as a slight topographic bulge in the model. An initial model was obtained through trial and error forward modelling to fit the picked traveltimes using a top-down layer stripping approach. This initial model was then refined via damped least-squares inversion to minimize the traveltime residuals. The resultant model was subsequently modified slightly, mainly regarding the vertical velocity gradients, based on qualitative comparison of the record sections with synthetic seismograms (Fig. 5) calculated using the TRAMP code of Zelt & Forsyth (1994).

The final velocity model (Fig. 6) shows a thin low-velocity layer of $3.3 \, \text{km s}^{-1}$ within the rift valley which extends westward to near SP14. The layer thickens eastwards from c. 1 to c. 2.5 km in the rift valley. It is underlain by a c. 2- to 5-km-thick layer of $4.5–5.5 \, \text{km s}^{-1}$ which extends the length of the profile. Beneath this layer, the top of the crystalline basement (defined as the $>6.0 \, \text{km s}^{-1}$ layer) shows considerable topography, in particular beneath the rift escarpments. This topography resembles the surface expression of the rift with a small throw beneath the western margin of the rift and a larger throw beneath the Arboye border fault on the eastern margin.

Normal continental crustal velocities of $6.0–6.8 \, \text{km s}^{-1}$ are modelled to 35- to 40-km depth with the exception of a narrow 20- to 30-km-wide region in the upper crust beneath the Boset magmatic centre in the rift valley, where velocities are approximately $6.5 \, \text{km s}^{-1}$ (c. 5 per cent higher than in the surrounding parts of the profile). An intracrustal reflector at c. 21-km depth beneath the surface shallows eastwards beneath the western plateau and is modelled at c. 12-km depth beneath the eastern plateau. Another intracrustal reflector is also modelled beneath the rift axis but it is unclear whether these reflectors are continuous along the whole profile or separated with possible different origins (e.g. those beneath the rift valley may result from the presence of magma chambers or...
Figure 4. Example seismic sections from (a) SP12, (b) SP14 and (c) SP17. Sections are reduced at 6 km s$^{-1}$ with trace normalized amplitudes and bandpass filtered from 4–16 Hz. Calculated travelt ime curves from the velocity model shown in Fig. 5 are overlain. Phase labelling: $P_S$ and $P_S^1$—sedimentary diving waves; $P_g$—basement diving wave; $P_1$—diving wave in the mid crust; $P_x$—diving wave in high-velocity lower crustal (HVLC) layer; $P_n$—mantle diving wave; $P_{i1}$ and $P_{i2}$—intracrustal reflectors; $P_xP$—reflector off top of HVLC layer; $P_mP$—Moho reflector; $P_1P$—mantle reflector. Note the delay in the $P_g$ phase as a result of the deeper basement and thicker sediments within the rift e.g. between c. 50- and 100-km offset from SP14.

sills). A second intracrustal reflector is modelled at c. 28-km depth below the surface beneath the eastern and western plateaux. This reflector appears to be more continuous than the upper reflector. Its depth of 23 km below the surface beneath the rift shows that the upper crust is considerably thinned with the maximum thinning occurring slightly to the west of the rift axis. Due to the uncertainty in the continuity of the first intracrustal reflector we will in the following discussions use the terms upper and lower crust for the regions above or below this second reflector.

Beneath the eastern plateau the depth to the Moho is 40 km below the surface and the $P_n$ mantle velocity is 8.0–8.1 km s$^{-1}$ as expected in stable areas with normal heat flow. Beneath the rift axis the Moho is shallower at c. 35-km depth below the surface and the $P_n$ mantle velocity is 7.7 km s$^{-1}$. Beneath the western plateau a
Table 1. Modelling results.

<table>
<thead>
<tr>
<th>Phase</th>
<th>N</th>
<th>Per cent</th>
<th>Error(s)</th>
<th>$T_{RMS}$</th>
<th>$\chi^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$P_s$</td>
<td>70</td>
<td>94</td>
<td>0.06</td>
<td>0.105</td>
<td>3.359</td>
</tr>
<tr>
<td>$P_{s}P$</td>
<td>11</td>
<td>100</td>
<td>0.11</td>
<td>0.191</td>
<td>5.553</td>
</tr>
<tr>
<td>$P_{s}P$</td>
<td>187</td>
<td>97</td>
<td>0.06</td>
<td>0.085</td>
<td>2.647</td>
</tr>
<tr>
<td>$P_{s}P$</td>
<td>24</td>
<td>100</td>
<td>0.10</td>
<td>0.110</td>
<td>1.925</td>
</tr>
<tr>
<td>$P_g$</td>
<td>895</td>
<td>95</td>
<td>0.066</td>
<td>0.159</td>
<td>3.724</td>
</tr>
<tr>
<td>$P_{1}$</td>
<td>334</td>
<td>97</td>
<td>0.146</td>
<td>0.112</td>
<td>0.647</td>
</tr>
<tr>
<td>$P_{1}$</td>
<td>287</td>
<td>92</td>
<td>0.093</td>
<td>0.147</td>
<td>2.772</td>
</tr>
<tr>
<td>$P_{1}P$</td>
<td>334</td>
<td>99</td>
<td>0.151</td>
<td>0.205</td>
<td>2.431</td>
</tr>
<tr>
<td>$P_{1}P$</td>
<td>418</td>
<td>94</td>
<td>0.149</td>
<td>0.199</td>
<td>1.894</td>
</tr>
<tr>
<td>$P_{1}P$</td>
<td>287</td>
<td>98</td>
<td>0.155</td>
<td>0.176</td>
<td>1.901</td>
</tr>
<tr>
<td>$P_{1}$</td>
<td>90</td>
<td>93</td>
<td>0.131</td>
<td>0.126</td>
<td>2.049</td>
</tr>
<tr>
<td>$P_{1}$</td>
<td>32</td>
<td>69</td>
<td>0.160</td>
<td>0.204</td>
<td>1.490</td>
</tr>
</tbody>
</table>

$N$ is the number of picks, per cent is the percentage of data points used, error is the average assigned traveltime error for that phase, $T_{RMS}$ and $\chi^2$ after Zelt & Smith (1992).

A 15-km-thick layer of $7.4 \text{ km s}^{-1}$ velocity is modelled between c. 33- and 48-km depth. This may only extend to the rift and is definitely absent from the eastern part of the profile. The precise location of its eastward termination is uncertain, but it is likely that the layer terminates beneath the western rift flank. Mantle velocities beneath this body are not constrained by the refraction data but overlaying seismic sections with theoretical traveltime curves for a number of different possible velocities suggests a preferred mantle velocity of $<8.0 \text{ km s}^{-1}$ for this region. A possible mantle reflector at c. 60 km below the surface is modelled with relatively large depth uncertainty due to the lack of velocity information immediately above it.

**MODEL RESOLUTION**

Model resolution was assessed by monitoring the number of rays traced, the RMS traveltime residual, and the normalized $\chi^2$ value (Zelt & Forsyth 1994). These values are summarized in Table 1 for...
each phase and the observed and calculated traveltimes for each shot are shown in Fig. 7. In most cases the percentage of picks that were modelled is >90 per cent of the total number of traveltime picks, and the final RMS is nearly equal to the average phase uncertainty indicating that the data has not been overmodelled. The high $\chi^2$ value for the $P_g$ phase is due to a poor fit near the eastern edge of the rift where the steep near vertical boundaries make ray tracing difficult. Ray paths for the different phases (Fig. 8) give an indication of the resolution of the various parts of the model. In addition the diagonal elements of the resolution matrix for the depth and velocity nodes are shown in Fig. 9. Ideally the diagonal should equal 1 but typically values of $>0.5$ are considered to represent reasonably well resolved model parameters (Zelt 1999). Velocity resolution within most of the crust corresponds to a value of $>0.7$, decreasing towards the model edges. The reduced resolution in the upper crust in the centre of the model is due to a combination of the increased number of nodes required to model the narrow high-velocity region beneath the rift axis and the difficulty in tracing $P_g$ rays from SP16 westward through the rift border. However, this high-velocity feature is consistent with gravity modelling (Mahatsente et al. 1999) and seismic tomography (Keranen et al. 2004).

Errors associated with the interfaces and velocities have been assessed through perturbation of model parameters and are ±0.1 km s$^{-1}$ for upper crustal velocities, ±0.2 km s$^{-1}$ for lower crustal velocities, ±1 km for the intracrustal layers and ±2 km for the Moho.

**INTERPRETATION**

**Sedimentary structure**

A 2- to 5-km-thick layer with a velocity of less than 6.0 km s$^{-1}$ extends across the length of the profile. It most likely represents a combination of pre-rift Jurassic sediments and Oligocene flood basalts. Above this, within the central part of the model a thin layer with a lower velocity is modelled, thickening to the east into the Arboye border fault. This is not limited to within the rift but extends further to the west. It correlates with surface outcrops of the Late Miocene—Early Pliocene ignimbrites and rhyolites of the Kessem and Balchi formations and the Quaternary sediments and volcanics of the Bishoftu embayment and rift valley (Wolfenden et al. 2004). The thickening into the Arboye fault would be expected for syn-rift sedimentation.

The model suggests a thickness of 5 km for the Adama basin, not inconsistent with the 4 km proposed by Wolfenden et al. (2004) from the projection of stratal dips (Fig. 3b). The stepwise deepening of the basement beneath the rift is more pronounced on the eastern side where it correlates with the uplifted rift flank. The large throw on the Arboye border fault on this flank is likely a result of basement faulting. Pronounced basement topography is also modelled below SP16. This is outside of the rift and does not correlate with topography. It may be the result of an earlier period of Mesozoic rifting (e.g. Korme et al. 2004).

**Upper crust**

Normal upper crustal velocities (6.1–6.2 and 6.3–6.4 km s$^{-1}$ in the two sublayers) are modelled across the profile with the exception of beneath the rift where $P$-wave velocities are 5–10 per cent higher ($>6.5$ km s$^{-1}$). These high velocities correlate with the Boset volcanic segment suggesting the presence of a gabbrico magmatic intrusion based on the high velocities and a modelled density of 3000 kg m$^{-3}$ (Mahatsente et al. 1999). This is consistent with upper crustal tomographic models based on data from the EAGLE areal array (Keranen et al. 2004) and with the limited available gravity data, which shows a short wavelength Bouguer anomaly peak within...
the rift. The upper crust (including the surficial volcanic/sediment layer) thins from c. 28 km–c. 23 km thick beneath the rift with the region of maximum thinning offset to the west of the axis of the surface expression of the rift. Similar crustal velocities are modelled under both the western and eastern plateaux. This is different to findings in Kenya, where significantly higher upper crustal velocities were modelled on the eastern than on the western side of the rift which has led to the proposal that the Kenya rift formed along a pre-existing Proterozoic–Archean suture zone (Birt et al. 1997). The finding of similar velocities on the two sides of the MER suggests that this part of the East African Rift system dissects crust of similar origin.

Lower crust and upper mantle

Normal continental lower crust and mantle is modelled beneath the eastern plateau. The crust is thinner beneath the rift axis than beneath the eastern side of the rift zone. A sub-Moho $Pn$ mantle velocity of 7.7 km s$^{-1}$ is modelled beneath the rift axis, similar to the $Pn$ mantle velocity beneath the Kenya rift. There such a low velocity was attributed to the presence of high-temperature mantle material with 3–5 per cent partial melt (Mechie et al. 1994) and we suggest a similar origin for the low mantle velocity modelled here. The crustal thinning and low mantle velocity agree with preliminary results from the modelling of data along the axial profile (Maguire et al. 2005).

Beneath the western plateau a 15-km-thick layer with a velocity of 7.4 km s$^{-1}$ is modelled which, although uncertain, is believed to terminate beneath or near the western rift escarpment. The resolution test (Fig. 9) shows that the velocity is reasonably well resolved, that its upper interface is well constrained by the data, and that a reflection from its base is resolved in the central part of this body. We only identify this layer on the western side of the MER. The layer exhibits strong reflectivity in the seismic sections and its velocity of c. 7.4 km s$^{-1}$ is between those expected of normal lower crust and mantle. The depth to the base of this layer is greater than the Moho.

Figure 7. Observed and calculated traveltimes for all shot points along EAGLE profile 1. Grey bars indicate the traveltime picks with a length corresponding to the assessed uncertainty, black dots the calculated traveltimes.
depth on the eastern side of the rift and to the south (Berckhemer et al. 1975; Makris & Ginzberg 1987). Teleseismic receiver functions also suggest a thicker crust exists under the northwestern than under the southeastern plateau and to the south, with crustal thinning beneath the rift axis (Dugda et al. 2005; Stuart et al. 2003). The available gravity data also exhibits a general decrease in Bouguer anomaly values from east to west consistent with a thick crust including a dense lower crustal layer beneath the western plateau and a thinner crust beneath the eastern plateau (Tiberi et al. 2005; Cornwell et al. 2005). We see three possible origins for this layer which are discussed below: (1) a relic pre-rift Proterozoic crust; (2) a low velocity mantle layer; or (3) magmatic underplating of Oligocene and/or recent origin.

(1) A 7-km-thick lower crustal layer with velocities of 7.0–7.3 km s\(^{-1}\) is commonly observed in Proterozoic crust worldwide (Christensen & Mooney 1995) and was modelled at the base of the Proterozoic crust on the eastern side of the Kenyan rift (Birt et al. 1997). However, given the relatively high geothermal gradient and thereby temperature of the region, the velocities in our model of the
crust in Ethiopia are higher than would be expected for Proterozoic crust. Further, we do not observe any significant change in upper crustal velocities across the rift as was modelled by Birt et al. (1997) in Kenya, as might be expected here if this layer indicates that the crust under the northwestern plateau is of a different origin to that to the southeast. We, therefore, find it unlikely that this layer has a Proterozoic origin.

(2) Mantle velocities of 7.4 km s\(^{-1}\) have been modelled beneath the Ethiopian Rift proper and interpreted by Berckhemer et al. (1975) and Makris & Ginzberg (1987) to represent hot, partially molten mantle rocks. However, these velocities were modelled farther to the north beneath the rift in Afar where surface wave studies indicate the existence of a low-velocity anomaly of possible thermal origin, consistent with a mantle plume extending to at least 660-km depth (e.g. Debayle et al. 2001). Similar velocities of 7.4 km s\(^{-1}\) are modelled here beneath the plateau. It is, therefore, unlikely that the layer represents partially molten mantle material which has undergone purely decompressional melting, especially considering that the velocities are lower than those modelled beneath the rift axis and that a strong reflector is observed from the base of this layer. However, Bastow et al. (2005) show the existence of a mantle anomaly offset to the west of the rift axis at 75-km depth close to the rift escarpment but slightly to the south of the profile. Whilst this suggests the presence of a thermal anomaly at this depth which could produce partial melt one might expect that such a layer would also have been modelled by Berckhemer et al. (1975) and Makris & Ginzberg (1987) along the profile further to the south. Although there is no direct evidence of this layer to the south, it is not certain that it is not present owing to the layer not being observable due to a combination of the relatively sparse density of stations and relative small thickness of the layer. However, receiver function studies (Dugda et al. 2005) indicate that the crust to the south is thinner than under the plateau, consistent with Berckhemer et al. (1975) and Makris & Ginzberg (1987) and our results if this layer is crustal rather than mantle in origin.

(3) The HVLC due to crustal underplating is a characteristic feature of all continental volcanic margins (e.g. Kelemen & Holbrook 1995). Typically velocities of such layers are approximately 7.2–7.5 km s\(^{-1}\) (e.g. Holbrook & Kelemen 1993) but have been modelled as high as 7.7 km s\(^{-1}\) (e.g. Caress et al. 1995). The layers may reach thicknesses of 10–15 km, consistent with the velocities and geometry modelled here, even though the high geothermal gradient in Ethiopia may suggest that the observed velocity is slightly higher than expected. The existence of the layer beneath the northwestern plateau where the maximum thickness of flood basalts are observed, suggests a connection to the Oligocene flood basalts. The observation of riftward-dipping volcanic wedges in the southern Red Sea rift further to the north, analogous to seaward-dipping reflector (SDR) sequences imaged on magmatic margins worldwide (Wolfenden et al. 2005), which are normally formed above, or seaward of a region of HVLC would also suggest that a HVLC layer could be expected to exist under the plateau.

Evidence for the generation of magmatic underplate associated with the Oligocene flood basalts has also been suggested from thermochronological and morphological studies of the Upper Nile drainage network. These indicate that erosion initiated in the Blue Nile canyon after the Oligocene uplift of the plateau at 25–29 Ma, which is proposed to be the result of Afar plume impingement and/or massive underplating and associated with Oligocene flood basin differentiation (Pik et al. 2003). Studies of the generation of major sequences of plateau rhyolitic ignimbrites of mainly Oligocene age, which are believed to represent 20 per cent of the flood basalts suggest that their derivation through fractional crystallisation requires the creation of at least 3 km of gabbric underplate assuming an area equal to the surface area of these ignimbrites on the plateau (Ayalew et al. 2002). Trua et al. (1999) have also proposed that recent (<1.8 Ma) felsic magmatism may have been generated by partial melting of existing gabbric underplate followed by low-pressure fractionation.

Given the present evidence we find it most likely that the 7.4 km s\(^{-1}\) layer came into existence as a magmatic addition at the base of the crust, although we cannot uniquely interpret its origin. It is most likely to have been emplaced during the Oligocene but a component of it may be a relatively recent feature. The finding of a narrow vertical zone of low velocity in the mantle beneath the layer (Bastow et al. 2005) could indicate a conduit of hot material presently rising under Ethiopia, which may have added to existing underplated material. Emplacement of underplated magma at the Moho would cause substantial local heating of the lowermost crust, which probably already was close to solidus temperatures due to the relatively high strain rates. It appears likely that some original lowermost crustal rocks have melted and dissolved in the melts that
Figure 10. True amplitude seismic section from SP12, SP17 and SP18 illustrating the difference in lower crustal reflectivity characteristics. SP18 has bounce points under the eastern plateau, SP17 bounce points directly beneath the rift axis and SP12 bounce points in the high-velocity lower crustal layer. Arrowed sections indicate the range over which $P_{m}P$ reflectivity is discussed in the text.

were added to the base of the crust. This may explain the slight shallowing of the upper interface of the anomalous layer compared to the depth to Moho on the eastern side of the rift. The upper interface of this anomalous layer may represent an upper limit of a relict Moho.

The existence of thicker crust under the northwestern plateau may also have influenced the evolution of the MER on a local scale and partly explain the kink in the rift at around $9^\circ$ N where the rift axis changes from NNE to a more NE direction (Figs 2 and 3). The rift’s northward propagation may have followed the pre-existing boundary.
between thicker underplated crust to the northwest and thinner crust to the southeast.

Lower crustal reflectivity

Comparison of seismic sections with bounce points beneath the southeastern plateau from SP18, the rift centre from SP17, and the northwestern plateau from SP12 show different lower crustal reflectivity characteristics (Fig. 10). Beneath the southeastern plateau the distinct PmP phase (seen clearly from SP18 between 90 and 130 km offset) has a constant amplitude with low frequency content. Beneath the rift centre the PmP phase (seen clearly from SP17 between 90- and 150-km offset) has a more variable higher amplitude with a high frequency reverberative content. This reverberatory nature is similar to that observed on seismic sections from the Kenya Rift where reflectivity modelling suggests the most likely cause of such reverberation results from sills within the lowest crust (Thybo et al. 2001). Beneath the northwestern plateau the lower crustal phases also show a high amplitude and frequency content (seen clearly from SP12 between 110 and 150-km offset) which may be consistent with layering within underplated material (e.g. Deemer & Hurich 1994). These observations indicate the presence of magmatic intrusions into the lower crust beneath the MER Zone. The reflectivity on the northwestern side is consistent with the presence of a large underplated layer.

CONCLUSIONS

Velocity modelling of data from a new wide-angle profile across the MER has provided new insights into the likely geological structure of the region (Figs 6 and 11). Sedimentary and volcanic sequences of 4- to 5-km thick are modelled across the length of the profile, thickest within the rift and correlating with geological observations at the surface. Basement topography shows that the rift bounding faults may extend into crystalline basement, strongly expressed by the modelling of the subsurface continuation of the Arboye border fault.

The sediments are underlain by a crust which exhibits significant structural asymmetry across the region. A 35-km-thick crystalline layer with normal continental velocity structure is modelled beneath sediments and volcanics of the eastern plateau. A 45-km-thick crystalline layer including a c. 15-km-thick high-velocity (7.4 km s$^{-1}$) lower crust, believed to be associated with underplating processes during the Oligocene and possibly more recent magmatic activity, is modelled beneath the western plateau. This proposed pre-rift structural difference may have influenced the evolution of rifting within the MER. Beneath the rift, higher than normal upper crustal velocities are interpreted to result from mafic intrusions beneath the Boset magmatic centre. We suggest that the reflectivity characteristics of lower crustal phases on the controlled source seismic record sections result from magmatic layering within the lower crust.

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REFERENCES


