S-WAVE VELOCITY STRUCTURE OF THE CRUST AND UPPER MANTLE BENEATH KENYA IN COMPARISON TO TANZANIA AND ETHIOPIA: IMPLICATIONS FOR THE FORMATION OF THE EAST AFRICAN AND ETHIOPIAN PLATEAUS

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ABSTRACT
New estimates of the S-wave velocity structure of the crust and upper mantle beneath the Kenya Rift and surrounding Kenya Highlands have been obtained by jointly inverting P wave receiver functions and Rayleigh wave phase and group velocities. The joint inversion was performed for seven broadband seismic stations, five that were part of the Kenya broadband seismic experiment and two belonging to the Global Seismic Network. 1-D S-wave velocity models obtained from the inversions have been compared to similar models for Tanzania and Ethiopia to identify differences in lithospheric structure beneath the East African and Ethiopian Plateaus. The S-wave velocity structure of the upper mantle to at least 100 to 150 km depth beneath the Kenya Highlands is similar to structure under the East African Plateau in Tanzania. It therefore appears that the plateau lithosphere under the Kenya Highlands has not been affected thermally to any greater extent than the plateau lithosphere beneath Tanzania. Under the Kenya Rift, however, the lithosphere has been substantially modified. S-wave velocities at depths of 80 to 150 km beneath the rift are 8 to 10% lower compared to under the Kenya Highlands, which may be a sufficiently large change to indicate the presence of partial melt.

There is little similarity between the plateau lithosphere in East Africa and Ethiopia. The lithosphere under the Ethiopian Plateau is thin, extending to a depth of no more than about 80 to 90 km depth, compared to 100 to 150 km beneath Tanzania and the Kenya Highlands. The maximum S-wave velocity in the lithosphere is also very low, reaching only to 4.2 to 4.3 km/s, compared to 4.6 to 4.7 km/s beneath Tanzania and the Kenya Highlands. These differences indicate that the buoyant support for the plateau elevation in East Africa, including the Kenya Highlands, resides at deeper depths in the mantle than beneath the Ethiopian Plateau.

Introduction
The East African and Ethiopian Plateaus form the northern portion of the African Superswell (Nyblade and Robinson, 1994), and their origin has long been debated. In order to explain the formation of these plateaus, many studies have sought to characterise the structure of crust and upper mantle beneath the plateaus, including the structure of the Cenozoic rift valleys that are associated with them (Figure 1). Much of the early work focused on Kenya because the Cenozoic rift structures are well developed there, and field studies were easier to conduct than in other parts of East Africa. For instance, crust and upper mantle structure was investigated by the Kenya Rift International Seismic Project (KRISP) between 1985 and 1990 (Fuchs et al., 1997; Prodehl et al., 1994, and references therein). More recently, a number of seismic investigations of crustal and upper mantle structure in Tanzania and Ethiopia have yielded important new information, helping to fill in gaps in our knowledge of both lithospheric and sublithospheric structure beneath the plateaus (e.g., Last et al., 1997; Nyblade et al., 2000; Owens et al., 2000; Nyblade, 2002; Dugda et al., 2005; Benoit et al., 2006; Yirgu et al., 2006 and references therein).

One approach that has been used successfully to image crustal and upper mantle structure in Tanzania and Ethiopia employs a joint inversion of P wave receiver functions and Rayleigh wave phase and group velocities (Julià et al., 2005; Dugda et al., 2007). From the 1-D S-wave velocity models obtained, estimates have been made of crustal thickness and lithospheric thermal structure. For areas of the plateaus away from the rift valleys, models show very different lithospheric structure for Tanzania (Julià et al., 2005) compared to Ethiopia (Dugda et al., 2007). The lithospheric mantle across the centre of the East African Plateau is characterised by Vs values of 4.6 to 4.7 km/s extending to depths of 100 to 150 km or deeper, suggesting that there has been limited, if any, thermal modification of the lithosphere (Julià et al., 2005). In contrast, the lithospheric mantle across the Ethiopian Plateau is characterised by Vs values of 4.2 to 4.3 km/s.
Figure 1. Map showing the topography of the study region, seismic stations locations and major tectonic features, as indicated in the key. Line AA’ shows the location of the velocity profile for Ethiopia in Figure 4a. Line BB’ shows the location of the velocity profiles for Kenya in Figures 2 and 4b. Lines CC’, DD’ and EE’ show the locations of profiles for Tanzania in Figures 4c and d.
to depths of ~80 to 90 km/s, with even lower velocities below those depths, indicating that the thermal structure of the lithosphere has been modified significantly (Dugda et al., 2007).

In this paper, we provide comparable information about the S-wave velocity structure of the lithosphere in Kenya using the same joint inversion method as used in Tanzania and Ethiopia. We report new 1-D S-wave velocity models for several locations across central and southern Kenya, letting us compare crust and upper mantle structure in this portion of the East African Plateau with structure beneath the central part of the plateau in Tanzania and beneath the Ethiopian Plateau. Our results show that the lithospheric mantle in Kenya away from the rift valley is similar to that found beneath Tanzania. This finding, when combined with the previously published results from Tanzania and Ethiopia, provides important constraints on the depth extent and nature of the mantle processes that have created the East African and Ethiopian Plateaus.

**Geologic Background**

The Precambrian tectonic framework of the East African Plateau consists of the Archean Tanzania Craton surrounded by the Proterozoic Kibaran, Ubendian, Usagaran and Mozambique Belts (Figure 1). The Cenozoic East African Rift System (EARS) has developed primarily within the mobile belts and forms two branches (eastern and western; Figure 1). The Eastern Branch in Ethiopia divides the Ethiopian Plateau into eastern and western sections, and continues southward through west-central Kenya and into northern Tanzania, transecting the northeastern corner of the East African Plateau. In Kenya, the Eastern Branch is locally referred to as the Kenya or Gregory Rift, and the uplifted and volcano-capped regions away from the Kenya Rift we refer to as the Kenya Highlands. The Eastern Branch of the EARS formed within the Mozambique Belt, which runs from Ethiopia south through Kenya, Tanzania and Mozambique, and is often considered as a Himalayan-type continental collision zone (Burke and Sengor, 1986; Shackleton, 1986). The Western Branch of the EARS defines the western side of the East African Plateau (Figure 1).

Volcanism and rifting in Kenya are thought to have occurred at about the same time (Ebinger, 1989). The earliest volcanism in Kenya started in the Turkana region of northern Kenya around ~35 to 40 Ma (Furman et al., 2006a; MacDonald et al., 2001). Magmatic activity in other parts of northern Kenya began c. 30 Ma (Morley et al., 1992; Ritter and Kaspar, 1997), while volcanism started around 15 Ma in the central portion of the Kenya rift, at c. 12 Ma in southern Kenya (Morley et al., 1992; Hendrie et al., 1994; Mechie et al., 1997), and at about 8 Ma in northern Tanzania (Dawson, 1992; Foster et al., 1997). Timing of plateau formation in East Africa remains poorly constrained, although there is some evidence for localised Neogene uplift along the flanks of some rift valleys (Noble, 1997; Van der Beek et al., 1997).

To the north, widespread volcanic activity started in the central Ethiopian Plateau during the Oligocene (c. 31 to 29 Ma) and resulted in the emplacement of thick (500 to 2000 m) flood basalts and rhyolites within 1 to 2 Ma (Hofmann et al., 1997; Mohr and Zanettin, 1988; Mohr, 1983; Berhe et al., 1987; Baker et al., 1996; Ayalew, 2002; Coudie et al., 2003). Less voluminous syn-rift shield volcanoes formed between 30 and 10 Ma and sit on top of the flood basalts to create additional relief of 1000 to 2000 m (Berhe et al., 1987; Coudie et al., 2003). The Afar triple junction formed long after the eruption of the Afar flood basalt volcanism as a result of the opening of the Eastern Branch of the EARS. In the central and northern part of the Main Ethiopian Rift (MER), extension started c. 11 Ma (Wolfenden et al., 2004; Chernet et al., 1998; WoldieGabriel et al., 1999), while in southwestern Ethiopia extension started c. 18 Ma (Ebinger et al., 2000). Uplift of the Ethiopian Plateau occurred between 20 and 30 Ma, soon after the major flood basalt eruption (Pik et al., 2003).

The P wave velocity structure of the crust and upper mantle beneath Kenya has been studied extensively (e.g., Prodehl et al., 1994; Fuchs et al., 1997 and references therein; Green et al., 1991; Achauer et al., 1994; Slack et al., 1994; Achauer and Masson, 2002; Davis and Slack, 2002; Park and Nyblade, 2006), but the thermal structure of the bottom part of the lithosphere under the Kenya Highlands is not easily inferred from these studies. Absolute P wave velocities are provided to depths of about 80 km by seismic refraction profiles, which is not sufficiently deep to infer the thermal state of the base of the lithosphere beneath the Kenya Highlands (Prodehl et al., 1994; Fuchs et al., 1997 and references therein). P wave tomography models show relative variations in P wave speeds to depths of 200 to 300 km, but it is not clear how much of the velocity variation in these models results from thermally perturbed lithosphere under the rift valley as opposed to thermally perturbed lithosphere away from the rift valley (Green et al., 1991; Achauer et al., 1994; Slack et al., 1994; Achauer and Masson, 2002; Davis and Slack, 2002; Park and Nyblade, 2006).

**Data and Modeling**

Seismic data collected between 2001 and 2002 at five stations (TALE, KAKA, BARI, ANGA, and KITU) of the Kenya broadband seismic experiment (KBSE) have been used for this study together with data from two permanent stations in the region (the IRIS/GSN stations NAi and KMBO) (Figure 1). Six of the seven seismic stations were located either on the eastern or western side of the Kenya Rift (Figure 1), and one station was situated within the Kenya Rift. Data from another five stations of the KBSE were not of sufficient quality to be modeled using the joint inversion method. Nyblade and Langston (2002) provide details on the station configuration and recording parameters used in the KBSE.

In the joint inversion scheme applied, we used P wave receiver functions and fundamental mode
Figure 2. Results of the joint inversion of receiver functions and surface wave dispersion measurements for stations in Kenya. The top panels show P wave receiver functions in different ray parameter bins and for different frequencies (see text for details). The middle panels show Rayleigh wave group (15 to 45 seconds) and phase (50 to 140 seconds) velocities, and the bottom panels show the velocity models obtained from the inversion (thick black line) and the average S-wave model for stations in the Mozambique Belt in northern Tanzania (thin black line). A reference line at 4 km/s is also shown, and the number next to each receiver function stack gives the number of receiver functions included in the stack.
Rayleigh wave phase and group velocities. The two kinds of data provide complementary information on earth structure. Receiver functions are time series obtained by deconvolving the vertical component of the teleseismic P-wave coda from the corresponding radial component and can be used to resolve velocity contrasts and relative travel times in the neighborhood of a seismic station (Langston, 1979; Ammon et al., 1990; Julià et al., 2000). Rayleigh wave phase and group velocities, on the other hand, can be used to constrain the average shear wave velocity with depth (Julià et al., 2000; Takeuchi and Saito, 1972).

We have applied the method developed by Julià et al. (2003) for the joint inversion. This method makes use of a linearised inversion procedure that minimises a weighted combination of least-squares norms for each data set, a model roughness norm, and a vector-difference norm between inverted and preset model parameters. The velocity models obtained are, consequently, a compromise between fitting the observations, model simplicity and a priori constraints.

Rayleigh wave group velocities between 10 to 45 sec period from Pasyanos (2005) have been used in the inversion of receiver functions and dispersion velocities, together with Rayleigh wave phase velocities between 50 and 140 sec period from the “background” and “eastern branch” regions defined by Weeraratne et al. (2003). The background region in the Weeraratne et al. (2003) study encompasses the Mozambique Belt in northern Tanzania and southern Kenya away from the Kenya Rift. The region, as defined by Weeraratne et al. (2003), is representative of the Kenya Highlands on either side of the Kenya Rift between −1 degree south latitude and the Tanzania border, in addition to the Mozambique Belt in northern Tanzania. Phase velocities for the background region were used in the joint inversion for stations TALE, KAKA, which are located within the Mozambique Belt just to the north of the background region, and stations NAII, KMBO, ANGA and KITU, which are located within the background region. For station BARI, which is located within the Kenya Rift, the phase velocities from the eastern branch region in the Weeraratne et al. (2003) study were used.

Uncertainties in the group velocities from Pasyanos (2005) are between 0.01-0.03 km/s throughout the study region. For the phase velocities, uncertainties are ~0.01 km/s for the Kenya Highlands and 0.03 to 0.05 km/s for the Kenya Rift (Weeraratne et al., 2003). To create smooth dispersion curves for each station, we applied a 3-point moving average to the dispersion measurements reported by Pasyanos (2005) and Weeraratne et al. (2003).

Receiver functions were computed using seismograms from teleseismic events between distances of 30° and 95° with magnitudes greater than 5.5. The time-domain iterative deconvolution method of Ligoria and Ammon (1999) was employed to compute the receiver functions with 200 iterations, and the quality of the receiver functions was evaluated by convolving the receiver function back with the vertical component and using a least-squares misfit criterion to assess the percentage of recovery of the original radial component. Those receiver functions not recovering at least 85% of the original waveform were rejected. Tangential receiver functions were examined for evidence of lateral heterogeneity and for dipping structure. Events with large amplitude tangential receiver functions were not used. We obtained between 5 and 15 high quality receiver functions per station that were used for further processing.

In the joint inversion, we used three groups of receiver functions, each corresponding to a range of ray parameters from 0.04 to 0.049 s/km, from 0.05 to 0.059 s/km, and from 0.060 to 0.069 s/km, to properly account for the phase moveout. In addition, for each grouping of receiver functions, we computed and stacked two sets of receiver functions that have overlapping frequency bands: a lower frequency band of f ≤0.5 Hz (Gaussian of 1.0), and a higher frequency band of f ≤1.25 Hz (Gaussian of 2.5). However, there was not sufficient data to obtain two sets of stacked receiver functions for each ray parameter grouping at any of the stations.

The model parameterisation used for the inversion consisted of constant velocity layers that increase in thickness with depth. Layer thicknesses were 1 and 2 km at the top of the model, 2.5 km between 3 and 60.5 km depth, 5 km between 60.5 and 260.5 km depth, and 10 km below a depth of 260.5 km. Our initial inversions showed that the model result have little dependence on the starting model but that velocities at lithospheric depths (≥150 to 100 km) trade-off with velocities below about 200 km depth. Therefore, the velocity structure below 205 km depth was determined through a trial-and-error process by finding models that best fit the 120 to 140 s phase velocities for each station. This was done by fixing velocities below 205 km depth within a range of −7 and +5% of PREM velocities (Dziewonski and Anderson, 1981), while at the same time inverting for the velocity structure above 205 km depth. For all stations tested, the model with velocities of 5 to 7% less than PREM fit the long period surface wave observations the best, while all the models fit the rest of the dispersion measurements equally well. Thus, for our final inversion we fixed velocities below 205 km depth to values 5 to 7% less than PREM and inverted for velocity structure above 205 km depth.

Similar to the results of Julià et al. (2005), by repeating the inversions for many combinations of model parameters and data, we found the uncertainties in the shear wave velocities to be about 0.1 km/s in the crust and uppermost mantle, 0.2 km/s in the lower part of the upper mantle, and uncertainties in the depth of discontinuities to be about 2 to 3 km in the crust and uppermost mantle.

Results

Results of the joint inversion are shown in Figure 2 for each station. Also shown on the same figure
for comparison purposes is an average S velocity profile
for unperturbed Mozambique Belt lithosphere obtained
by averaging the velocity profiles from stations KIBE,
HALE and KIBA in northeastern Tanzania obtained by
Julià et al. (2005) (Figure 1). These stations are located
away from the region in northern Tanzania where the
Eastern Branch of the rift system impinges on the margin
of the Tanzania Craton (Figure 1). The stations in Kenya
lie within the Mozambique Belt, and thus by showing
the average S profile from the three Mozambique Belt
stations in Tanzania, we can compare structure beneath
portions of the East African Plateau in Kenya and
Tanzania with lithosphere of similar age away from the
major zones of rifting.

Crustal thickness beneath stations on the eastern side
of the Kenya Rift is between 38 and 42 km, and on the
western side, 37 and 38 km. For the station in the Kenya
Rift, a crustal thickness of 30 km is obtained. The crustal
thickness estimates agree closely with the structure
obtained by Dugda et al. (2005), using the H-k receiver
function stacking technique of Zhu and Kanamori
(2000), where \( H = \) Moho depth and \( k = \frac{V_p}{V_s} \) (Figure 3),
and also with structure from the KRISP seismic refraction
profiles in places where the seismic stations are near to
the refraction profiles (Fuchs et al., 1997; Prodehl et al.,
1994, and references therein) (Figure 3).

Figure 3. Graph showing correlation between crustal thickness estimates from the joint inversion technique and results from previous
studies. A comparison of crustal thickness from KRISP refraction profiles is made only where seismic stations are near to a refraction line. See text for further explanation. The dashed line shows a one-to-one correlation.

Crustal thickness in the Mozambique Belt of the
Kenya Highlands is similar to crustal thickness in
the Mozambique Belt in Tanzania, which ranges from
36 to 44 km (Last et al., 1997; Julià et al., 2005).
Also within the Kenya Rift (station BARI), we observe a
high-velocity mid-crustal layer that is several kilometers
thick. This layer is consistent with previous reports of
intrusive bodies within the mid to upper crust under the
rift inferred from gravity anomalies (i.e., Swain, 1992).

For the uppermost mantle, if we assume a Poisson’s
ratio between 0.26 and 0.27 and compare our S-wave
velocity structure to the P-wave velocity structure from
refraction profiles of KRISP studies (Fuchs et al., 1997;
Prodehl et al., 1994, and references therein), we also
find good agreement. The average Sn velocity beneath
the Kenya Highlands is about 4.5 km/s, while for the
Kenya Rift it is about 4.2 to 4.3 km/s, corresponding
to Pn velocities found by the KRISP group of ~8.1 to
8.2 km/s beneath the Kenya Highlands and about
~7.6 to 7.8 km/s beneath the Kenya Rift.

In comparison to the average S velocity structure of
the upper mantle for the Mozambique Belt in Tanzania,
there is a suggestion in Figure 2 of some variability in
the velocities. For example, stations to the east of
the Kenya Rift (NAI, KMBO, ANGA, KITU) exhibit slightly
lower velocities to depths of ~90 to 100 km, while the
Figure 4. S-wave velocity profiles for selected stations in Ethiopia (a), Kenya (b), and Tanzania (c and d) showing crustal and upper mantle structure to a depth of 100 km. The results for Ethiopia are from Dugda et al., (2007) and for Tanzania from Julià et al., (2005). The velocity profiles have been obtained using the same joint inversion method. The locations of the profiles and stations are given in Figure 1. A reference line at 4 km/s is shown on each profile to aid in comparing the profiles, and upper mantle structure with a S-wave velocity of greater than 4 km/s has been shaded.
two stations to the west of the Kenya Rift (TALE and KAKA) exhibit almost identical velocities over that depth interval. In addition, below ~100 km depth, the stations to the east of the rift exhibit higher velocities than those found in Tanzania, while the stations to the west of the rift show either slightly faster velocities or similar velocities to those in Tanzania. However, these variations fall within the ± 0.1 km uncertainty in our velocity estimates, and therefore may not be significant.

Discussion
In this section, we interpret the S velocity profiles to infer, qualitatively, the degree to which the lithospheric mantle beneath the East African and Ethiopian Plateaus has been thermally modified. To assist with the interpretation, in Figure 4 we show S velocity profiles to 100 km depth for selected stations in Ethiopia, Kenya and Tanzania. Structure below 100 km depth is not shown because of the more limited depth resolution of the models for Ethiopia compared to Kenya and Tanzania. Our interpretation is based on the assumption that lithospheric structure everywhere beneath the Mozambique Belt in Ethiopia, Kenya and Tanzania was similar prior to the Cenozoic uplift, rifting and volcanism.

In Figures 2 and 4, it can be seen that the upper mantle beneath the Kenya Highlands is similar to upper mantle structure under the Mozambique Belt in Tanzania, as well as beneath the Tanzania Craton. This is not an unexpected result because upper mantle velocities in the inversion procedure are controlled primarily by the Rayleigh wave phase velocities, and the phase velocities used in the inversions for stations in the Mozambique Belt in Tanzania and Kenya are the same.

Under the Kenya Rift (station BARI), however, the lithosphere has been substantially modified, as reflected in the lower S velocities compared to the average Mozambique Belt profile and to the S velocities found under the Kenya Highlands. S velocities at depths of 80 to 150 km beneath the rift are 8 to 10% lower compared to the Kenya Highlands, which may be a sufficiently large decrease to indicate the presence of partial melt (Faul and Jackson, 2005). The existence of partial melt in the upper mantle beneath the Kenya Rift is also supported by results from several P wave tomography studies (Achauer et al., 1994; Slack et al., 1994; Green et al., 1991).

Figure 4 illustrates clearly that there are significant differences in upper mantle structure between the East African and Ethiopian Plateaus. The seismic lid (i.e., shaded regions on profiles in Figure 4 with \( V_s > 4 \) km/s) under the Ethiopian Plateau, which we equate with the lithosphere, is thin, extending to depths of no more than about 80 to 90 km. The maximum S velocity is also very low, reaching only to 4.2 to 4.3 km/s, compared to 4.6 to 4.7 km/s beneath the Mozambique Belt in northern Tanzania and the Kenya Highlands. There is little similarity between the lithosphere under the East African and Ethiopian Plateaus.

Dudga et al. (2007) show with a simple conductive thermal model that the thin, low velocity seismic lid beneath the Ethiopian Plateau could be attributed to a mantle plume at 30 Ma that thinned the lithosphere by about 50 km and heated the remaining portion of the lithosphere, raising lithospheric temperatures by a few hundred degrees. Dudga et al. (2007) also show that much of the buoyant support for the plateau elevation in Ethiopia could result from the thermal modification of the lithosphere. The contrast between the upper mantle S velocity structure between the Ethiopian and East African Plateaus indicate that a similar large thermal event has not pervasively modified the East African lithosphere, and that the buoyant support for the plateau elevation in East Africa, including the Kenya Highlands, resides deeper in the mantle than beneath the Ethiopian Plateau.

Summary
New estimates of the S-wave velocity structure of the crust and upper mantle beneath the Kenya Rift and surrounding Kenya Highlands have been obtained by jointly inverting P wave receiver functions and Rayleigh wave phase and group velocities. The joint inversion was performed for seven broadband seismic stations in the Kenya Rift and surrounding Kenya Highlands. The inversion results yield Moho depths beneath the rift and highlands that are consistent with previously reported estimates of crustal thickness.

The S-wave velocity structure of the upper mantle to at least 200 km depth beneath the Kenya Highlands away from the Kenya Rift is similar to structure under the East African Plateau in Tanzania. Under the Kenya Rift, however, the lithosphere has been substantially modified. S-wave velocities at depths of 80 to 150 km beneath the rift are 8 to 10% lower compared to under the Kenya Highlands, which may be a sufficiently large decrease to indicate the presence of partial melt.

There is little similarity between the lithosphere under the East African and Ethiopian Plateaus. The lithosphere under the Ethiopian Plateau is thin, extending to a depth of no more than about 80 to 90 km depth. The maximum S-wave velocity in the lithosphere is also very low, reaching only to 4.2 to 4.3 km/s, compared to 4.6 to 4.7 km/s beneath the Mozambique Belt in northern Tanzania and the Kenya Highlands. There is little similarity between the upper mantle S-wave velocity structure between the Ethiopian and East African Plateaus indicate that the buoyant support for the plateau elevation in East Africa, including the Kenya Highlands, resides deeper in the mantle than beneath the Ethiopian Plateau.

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References


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