LITHOSPHERIC STRUCTURE BENEATH EASTERN AFRICA
FROM JOINT INVERSION OF RECEIVER FUNCTIONS AND RAYLEIGH
WAVE VELOCITIES

A Thesis in
Geosciences

by

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Crust and upper mantle structure beneath eastern Africa has been investigated using receiver functions and surface wave dispersion measurements to understand the impact of the hotspot tectonism found there on the lithospheric structure of the region.

In the first part of this thesis, I applied H-κ stacking of receiver functions, and a joint inversion of receiver functions and Rayleigh wave group velocities to determine the crustal parameters under Djibouti. The two methods give consistent results. The crust beneath the GEOSCOPE station ATD has a thickness of 23±1.5 km and a Poisson's ratio of 0.31±0.02. Previous studies give crustal thickness beneath Djibouti to be between 8 and 10 km. I found it necessary to reinterpret refraction profiles for Djibouti from a previous study. The crustal structure obtained for ATD is similar to adjacent crustal structure in many other parts of central and eastern Afar. The high Poisson's ratio and Vp throughout most of the crust indicate a mafic composition, suggesting that the crust in Afar consists predominantly of new igneous rock emplaced during the late synrift stage where extension is accommodated within magmatic segments by diking.

In the second part of this thesis, the seismic velocity structure of the crust and upper mantle beneath Ethiopia and Djibouti has been investigated by jointly inverting receiver functions and Rayleigh wave group velocities to obtain new constraints on the thermal structure of the lithosphere. Crustal structure from the joint inversion for Ethiopia and Djibouti is similar to previously published models. Beneath the Main Ethiopian Rift (MER) and Afar, the lithospheric mantle has a maximum shear wave velocity of 4.1-4.2 km/s and extends to a depth of at most 50 km. In comparison to the lithosphere away
from the East African Rift System in Tanzania, where the lid extends to depths of ~100-125 km and has a maximum shear velocity of 4.6 km/s, the mantle lithosphere under the Ethiopian Plateau appears to have been thinned by ~30-50 km and the maximum shear wave velocity reduced by ~0.3 km/s. Results from a 1D conductive thermal model suggest that the shear velocity structure of the lithosphere beneath the Ethiopian Plateau can be explained by a plume model, if a plume rapidly thinned the lithosphere by ~30–50 km at the time of the flood basalt volcanism (c. 30 Ma), and if warm plume material has remained beneath the lithosphere since then. About 45-65% of the 1-1.5 km of plateau uplift in Ethiopia can be attributed to the thermally perturbed lithospheric structure.

In the final part of this thesis, the shear-wave velocity structure of the crust and upper mantle beneath Kenya has been obtained from a joint inversion of receiver functions, and Rayleigh wave group and phase velocities. The crustal structure from the joint inversion is consistent with crustal structure published previously by different authors. The lithospheric mantle beneath the East African Plateau in Kenya is similar to the lithosphere under the East African Plateau in Tanzania. Beneath the Kenya Rift, the lithosphere extends to a depth of at most ~75 km. The lithosphere under the Kenya Plateau is not perturbed when compared to the highly perturbed lithosphere beneath the Ethiopian Plateau. On the other hand, the lithosphere under the Kenya Rift is perturbed as compared to the Kenya Plateau or the rest of the East African Plateau, but is not as perturbed as the lithosphere beneath the Main Ethiopian Rift or the Afar. Although Kenya and Ethiopia have similar uplift and rifting histories, they have different volcanic histories. Much of Ethiopia has been affected by the Afar Flood Basalt volcanism, which may be the cause of this difference in lithospheric structure between these two regions.
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Chapter 1

Introduction

This thesis consists of three main chapters that focus on investigating the crust and upper mantle structure beneath eastern Africa using receiver functions and surface wave dispersion measurements to understand the impact of the hotspot tectonism found there on the lithospheric structure of the region. Figure 1-1 shows the study region, and the distribution of temporary broadband seismic stations and permanent seismic stations used for this thesis in Ethiopia, Kenya and northern Tanzania.

The crust under Djibouti found by this study is different from what had been published before by different authors but it is similar to adjacent crust in central and eastern Afar. The lithospheric structure beneath Ethiopia and Djibouti is highly perturbed, as compared to the relatively unperturbed lithosphere beneath northeastern Tanzania. In contrast, the lithosphere beneath the Kenya Plateau is similar to the unperturbed lithosphere in northeastern Tanzania. The lithosphere beneath the Kenya Rift, the Main Ethiopian Rift, and Afar is highly perturbed, with the lithosphere beneath the Kenya Rift being somewhat less perturbed than beneath the other areas.
In Chapter 2, crustal structure beneath the GEOSCOPE station ATD in Djibouti has been investigated using H-κ stacking of receiver functions and a joint inversion of
receiver functions and surface wave group velocities. I obtain consistent results from the two methods. The crust is characterized by a Moho depth of $23 \pm 1.5 \text{ km}$, a Poisson's ratio of $0.31 \pm 0.02$, and a mean $V_p$ of $\sim 6.2 \text{ km/s}$ but $\sim 6.9-7.0 \text{ km/s}$ below a 2 - 5 km thick low velocity layer at the surface. Some previous studies [Ruegg, 1975; Sandvol et al., 1998] of crustal structure for Djibouti placed the Moho at 8 to 10 km depth, and we attribute this difference to how the Moho is defined (an increase of $V_p$ to 7.4 km/s in this study vs. 6.9 km/s in previous studies). The crustal structure we obtained for ATD is similar to crustal structure in many other parts of central and eastern Afar [Berkhemer et al., 1975; Makris and Ginzburg, 1987; Dugda et al., 2005]. The high Poisson's ratio and $V_p$ throughout most of the crust indicate a mafic composition and are not consistent with models invoking crustal formation by stretching of pre-existing Precambrian crust. Instead, we suggest that the crust in Afar consists predominantly of new igneous rock emplaced during the late synrift stage where extension is accommodated within magmatic segments by diking [e.g., Mohr, 1989; Ebinger and Casey, 2001]. Sill formation and underplating likely accompany the diking to produce the new and largely mafic crust.

Chapter two has been published in a Geological Society Special Publication [Dugda, M. T., and A. A. Nyblade (2006), New constraints on crustal structure in eastern Afar from the analysis of receiver functions and surface wave dispersion in Djibouti, edited by Yirgu, G, C. J. Ebinger, and P. K. H. Maguire, *Geological Society Special Publications*, 259, pp.239-251.]. Regarding my specific contributions in this multi-authored article, the work is entirely mine and was supervised by the co-author Dr Andrew Nyblade.
In Chapter 3, the seismic velocity structure of the crust and upper mantle beneath Ethiopia and Djibouti has been investigated by jointly inverting receiver functions and Rayleigh wave group velocities to obtain new constraints on the thermal structure of the lithosphere. Most of the data for the study come from the Ethiopia broadband seismic experiment, conducted between 2000 and 2002 [Nyblade and Langston, 2002]. Shear velocity models obtained from the joint inversion show crustal structure that is similar to previously published models, with crustal thicknesses of 35 to 44 km beneath the Ethiopian Plateau, and 25 to 35 km beneath the Main Ethiopian Rift (MER) and the Afar [Makris and Ginzburg, 1987; Dugda et al., 2005; Stuart et al., 2006; Mackenzie et al., 2005; Dugda and Nyblade, 2006; Maguire et al., 2006]. The lithospheric mantle beneath the Ethiopian Plateau has a maximum shear wave velocity of about 4.3 km/s and extends to a depth of ~70-80 km. Beneath the MER and Afar, the lithospheric mantle has a maximum shear wave velocity of 4.1-4.2 km/s and extends to a depth of at most 50 km. In comparison to the lithosphere away from the East African Rift System in Tanzania [Julia et al., 2005; Weerartne, 2003], where the lid extends to depths of ~100-125 km and has a maximum shear velocity of 4.6 km/s, the mantle lithosphere under the Ethiopian Plateau appears to have been thinned by ~30-50 km and the maximum shear wave velocity reduced by ~0.3 km/s. Results from a 1D conductive thermal model [Carlaw and Jaeger, 1959] suggest that the shear velocity structure of the Ethiopian Plateau lithosphere can be explained by a plume model, if a plume rapidly thinned the lithosphere by ~30–50 km at the time of the flood basalt volcanism (c. 30 Ma), and if warm plume material has remained beneath the lithosphere since then. About 45-65% of the 1-1.5 km of plateau uplift in Ethiopia can be attributed to the thermally perturbed lithospheric

In Chapter 4, the shear-wave velocity structure of the crust and upper mantle beneath Kenya has been investigated using a joint inversion of receiver functions, and Rayleigh wave group and phase velocities. Most of the data for this part of the study come from the Kenya broadband seismic experiment, conducted between 2001 and 2002. Shear velocity models obtained from the joint inversion show crustal thicknesses of 37 to 42 km beneath the East African Plateau in Kenya and near the edge of the Kenya Rift, and a crustal thickness of about 30 km beneath the Kenya Rift. These crustal parameters are consistent with crustal thicknesses published previously by different authors [Dugda et al., 2005; Fuchs et al., 1997; Prodehl et al., 1994].

A comparison has been made between the lithosphere under Kenya and other parts of the East African Plateau in Tanzania. A comparison between the lithosphere under Kenya and that under Ethiopia has also been made, specifically between the lithosphere under the Ethiopian Plateau and the Kenya Plateau, and between the lithosphere beneath the Main Ethiopian Rift (MER) and the Kenya (Gregory) Rift. The lithospheric mantle beneath the East African Plateau in Kenya has a maximum shear wave velocity of about 4.6 km/s, similar to the value obtained under the East African Plateau in Tanzania [Julia et al., 2005]. Beneath the Kenya Rift, the lithosphere extends to a depth of at most ~75 km. The average velocity of the mantle lithosphere under the
East African Plateau in Kenya appears to be similar to the lithosphere under Tanzania away from the East African Rift System. The lithosphere under the Kenya Plateau is not perturbed when compared to the highly perturbed lithosphere beneath the Ethiopian Plateau. However, the lithosphere under the Kenya Rift is perturbed compared to the Kenya Plateau or the rest of East African Plateau (EAP), but is not as perturbed as under the Main Ethiopian Rift or the Afar.

Although Kenya and Ethiopia have similar uplift and rifting at the surface, the lithospheric thickness varies between the two locations. The Afar Flood Basalt volcanism may be the cause of this striking difference in the lithosphere between these two regions. The difference in the lithosphere beneath Ethiopia and Kenya has important implications for understanding the buoyant support for the plateau uplift found in the two regions. Sublithospheric processes may provide the main source of buoyancy for the Kenya Plateau, which is consistent with heat flow and gravity studies in the region [Nyblade et al., 1990; Ebinger et al., 1989], while thermal modification of the lithosphere in Ethiopia appears to have caused a considerable amount of plateau uplift in Ethiopia.

Since the thesis comprises separate papers that share similar methodology and tectonic region, the description of geodynamic settings and methodology are repeated in a number of places, although the descriptions are not exactly identical. Specially, a description of the joint inversion technique has been repeated in all the major chapters.

Appendix A provides preliminary results of a joint inversion of receiver functions and Rayleigh wave phase and group velocities for the Arabian Peninsula. The data for this work come from the Saudi Arabia National Digital Seismic Network (SANDSN) and
the inversion results give crustal structure that is consistent with previous studies [e.g.,
Al-Damegh et al., 2005].

Appendix B gives and describes the equations used in the joint inversion of
receiver functions and dispersion measurements developed by Julia et al. [2003], which is
based on Julia et al. [2000], and used for determining the shear wave velocity structure of
the region of study in this thesis.
References


Chapter 2

New Constraints on Crustal Structure in Eastern Afar

from the Analysis of Receiver Functions and Surface Wave Dispersion

in Djibouti

2.1 Introduction

Afar is a tectonically active region in between continental rifting and oceanic rifting where the Red Sea oceanic ridge, the Gulf of Aden oceanic ridge, and the continental East African rift system (EARS) meet in a rift-rift-rift triple junction (Figure 2-1). Understanding the nature and origin of the crust in Afar is important because Afar is one of the few places where it is possible to study the development of magmatic segmentation during rifting, the formation of volcanic rifted margins, and more generally, how continental rifts evolve into oceanic rifts.

There are a number of conflicting views about the thickness and composition of the crust in Afar, and how it formed. For example, some studies have reported estimates of crustal thickness between 14 and 26 km, and from these estimates inferred that the crust may be more continental than oceanic in nature (Makris and Ginzburg, 1987; Berkhemer et al., 1975). Other studies have reported estimates of crustal thickness between 8 and 10 km (Ruegg, 1975; Sandvol et al., 1998), suggesting an oceanic origin for the crust, and yet others have argued for completely new igneous crust that is much...
thicker than typical oceanic crust based on such observations as the velocity structure of the crust and Poisson's ratio (Mohr, 1989; Dugda et al., 2005).

In this Chapter, crustal structure is imaged beneath the GEOSCOPE station ATD in eastern Afar (Djibouti) using receiver functions and surface wave dispersion measurements. We report new estimates of Moho depth, Poisson's ratio, and shear velocity structure, and combine these estimates with seismic images of crustal structure from other parts of Afar to re-examine the nature and origin of the crust in Afar. This study differs from previous studies of receiver functions for station ATD in that 10 years of data have been used, enabling us to obtain high quality receiver function stacks and to investigate azimuthal variations in structure beneath the station. In addition, Rayleigh wave group velocities have been used to constrain crustal shear wave velocities.

2.2 Tectonic and geodynamic setting

The seismic station ATD is located on the south side of the Gulf of Tadjoura in Djibouti (Figures 2-1 and 2-2). The Gulf of Tadjoura is the western extension of the Gulf of Aden ridge, and represents the penetration of the ridge into Afar, where it joins with the East African and the Red Sea rifts (Courtillot, 1980; Cochran, 1981; Manighetti et al., 1997; Courtillot et al., 1999). Station ATD is located on 10-14 Ma rhyolites, and away from the main zones of rifting in the Gulf of Tadjoura (Figure 2-2) (Manighetti et al., 1997, 1998).

The tectonic and geodynamic setting of Afar has been studied extensively (e.g., Hayward and Ebinger, 1996; Manighetti et al., 1997, 1998; Acton et al., 2000; Ebinger
and Casey, 2001; Tesfaye et al., 2003; Audin et al., 2004; Wolfenden et al., 2005). The Red Sea and Gulf of Aden rifts began forming in the Oligocene when Arabia first separated from Africa. The western Gulf of Aden initially opened between 15 and 10 Ma.

Figure 2-1 Shaded relief map of Afar showing the location of GEOSCOPE station ATD (white triangle), political boundaries (thin solid lines), temporary broadband seismic stations operated between 2000 and 2002 (black squares), and seismic refraction lines (A = Berkhemer, 1975, solid bold lines labeled I to V; B = Ruegg, 1975, medium thickness lines in Djibouti; C = EAGLE, Maguire et al., 2003, bold dashed line). In the white boxes next to seismic stations are shown station names, Moho depth (in km), and crustal Poisson's ratio. In the white boxes next to the seismic refraction lines are shown estimates of Moho depth (in km) and mean crustal Vp (in km/s), assuming the Moho is where P wave velocities increase to >7.4 km/s.
and entered Afar c. 5 Ma (Courtillot et al., 1999; Hofmann et al., 1997). The eruption of flood basalts in Ethiopia and Yemen occurred at about 31 Ma, concurrent with or immediately prior to the opening of the Red Sea and Gulf of Aden (d’Acremont et al., 2005; Wolfenden et al., 2005; Ukstins et al., 2002; Hofmann et al., 1997). The opening of the Main Ethiopian Rift (MER) to form the third arm of the Afar triple junction commenced at ~11 Ma (Wolfenden et al., 2005; Chernet et al., 1998).

Rift models for extension in rheologically layered continental lithosphere, as found in Afar, can be grouped in two categories, 1) those invoking mechanical stretching, where strain is accommodated by large offset faults in a brittle upper crust and by ductile deformation in the lower crust, and 2) those invoking extension caused by dike intrusion within magmatic segments (e.g., Ebinger and Casey, 2001; Buck, 2004; Ebinger, 2005). The latter proposal has been supported for the MER and Afar by geodetic data indicating that magmatic segments accommodate >80% of the extension (Bilham et al., 1999), and by gravity and morpho-tectonic observations (Hayward and Ebinger, 1996; Manighetti et al., 1998). A dike intrusion model is also supported for the Afar and MER by recent shear-wave splitting studies using seismic data from the Ethiopian Afar Geoscientific Lithospheric Experiment (EAGLE) and permanent seismic stations in the region (Ayele et al., 2004; Kendall et al., 2005).

2.3 Previous studies of crustal structure in Afar

Detailed information about crustal P wave velocity structure in Djibouti comes from the deep seismic sounding experiment of Ruegg (1975) (Figure 2-1). To the south
of the Gulf of Tadjoura, Ruegg (1975) reported a velocity structure consisting of five layers with P-wave velocities, from top to bottom, of 4.0 km/s, 6.4 km/s, 6.9 km/s, 7.1 km/s, and 7.4 km/s, and layer thicknesses of 3.6 km, 5.6 km, 4.6 km, 11 km, and 10 km, respectively. Ruegg (1975) interpreted the Moho to be at ~10-11 km depth between layers with velocities of 6.4 km/s and 6.9 km/s (i.e., velocities of ~6.9 km/s and higher were considered to be velocities indicative of mantle rock). Ruegg (1975) found similar structure to the north of the Gulf of Tadjoura, with increase in velocity from 7.1 km/s to 7.4 km/s occurring at a somewhat shallower depth of about 20 km.

Crustal structure has been investigated in other parts of the Afar using seismic refraction, surface-wave dispersion, gravity and other geophysical data. Figure 2-1 summarizes Moho depths (H), crustal Vp values, and crustal Poisson’s ratios reported in previous studies. Much of the information available about crustal structure in Afar comes from seismic refraction surveys conducted in the mid-1970s (Berkhemer et al., 1975) (lines I-V, Figure 2-1). Makris and Ginzburg (1987), revising the previous interpretation of Berkhemer et al. (1975), reported Moho depths along refraction lines I and II of 30 km in the south and 26 km in the north. For profiles III and V, they found crustal thickness variations of 26 to 14 km, with a change in the middle of profile V from about 26 km to about 20 km. For refraction line IV, they reported that crustal thickness thins from 26 to 23 km toward the Red Sea coast. The values of P-wave structure reported by Ruegg (1975) for the Gulf of Tadjoura region is similar to the structure reported by Makris and Ginzburg (1987) for other parts of the Afar, but the layer thicknesses are somewhat different, and also the interpretation of what velocity indicates mantle rock is different (i.e., how the Moho is defined). Makris and Ginzburg (1987) interpret P-wave velocities
as high as 7.1 - 7.2 km/s as crustal velocities, and they concluded that the Moho is marked by an increase in velocity to 7.5 km/s.

Figure 2-2 Sketch map showing the geology around station ATD (solid triangle) (after Varet, 1975), the distribution of Ps conversion points at a depth of 25 km (dots), and the four regions used to group receiver functions.

Receiver functions from station ATD were modeled for crustal structure by Sandvol et al. (1998). They applied a grid search technique to model a receiver function stack comprised of 11 receiver functions from a limited range in backazimuth (1 event
came from the west and 10 came from the east). They obtained a Moho depth of 8 km, suggesting oceanic-like crust beneath Djibouti, similar to the interpretation of crustal structure by Ruegg (1975). More recently, Dugda et al. (2005) analyzed receiver functions from a temporary broadband station at Tendaho (TEND, Figure 2-1) in central Afar, obtaining a Moho depth of 25±3 km, in good agreement with estimates of crustal thickness along refraction lines II, III and IV (26 km, Figure 2-1) by Makris and Ginsburg (1987).

In contrast to Moho depth, there is more uniformity in the crustal Poisson's ratios reported for the Afar. Ruegg (1975) reported a high Poisson’s ratio of 0.28 to 0.33 for the Gulf of Tadjoura region, and in their global study of the continental crust, Zandt and Ammon (1995) obtained a crustal Poisson’s ratio for station ATD of 0.29±0.02. From an analysis of surface waves crossing Afar, Searle (1975) reported that Poisson’s ratio increases from 0.25 at the surface to 0.29 at the deepest portion of the crust. Dugda et al. (2005) reported a high crustal Poisson’s ratio of 0.36 for Tendaho (TEND, Figure 2-1).

2.4 New estimates of crustal structure for Djibouti

Crustal structure has been examined in this study using data from the GEOSCOPE station ATD and two complimentary methods, 1) the H-κ receiver function stacking method, and 2) a joint inversion of receiver functions and Rayleigh wave group velocities. The first method provides estimates of Moho depth and crustal Poisson's ratio, while the second method provides information about how shear wave velocities vary with depth in the crust.
2.4.1 Crustal Structure from H-κ analysis of Receiver Functions

Receiver functions have been modeled for several decades using a variety of methods (e.g., Langston, 1979; Taylor and Owens, 1984). For this study, we have used the H-κ stacking technique (H = Moho depth and κ = Vp/Vs) of Zhu and Kanamori (2000) because it provides robust estimates of crustal thickness and Poisson’s ratio. It is well known that H and κ trade off strongly (Ammon et al., 1990; Zandt et al., 1995), and in an effort to reduce the ambiguity introduced by this trade off, Zhu and Kanamori (2000) incorporated the later arriving crustal reverberations PpPs and PpSs+PsPs in a stacking procedure whereby the stacking itself transforms the time-domain receiver functions directly to objective function values in H-κ parameter space. The objective function for the stacking is

\[ s(H, κ) = \sum_{j=1}^{N} w_1 r_j(t_1) + w_2 r_j(t_2) - w_3 r_j(t_3) \] (2-1)

where \( w_1, w_2, w_3 \) are weights, \( r_j(t_i), i=1, 2, 3 \), are the receiver function amplitude values at the predicted arrival times \( t_1, t_2, \) and \( t_3 \) of the Ps, PpPs, and PpSs+PsPs phases for the jth receiver function, and \( N \) is the number of receiver functions used. The H-κ stacking algorithm is based on the premise that the weighted stack sum of the receiver function amplitudes should attain its maximum value when H and κ attain their correct values for a particular crust. By performing a grid search through H and κ parameter space, the H and κ values corresponding to the maximum value of the objective function can be determined (Zhu and Kanamori, 2000). The H-κ method provides a better estimate of
Moho depth and Vp/Vs ratio than a simple stack method because it uses the correct ray parameter in the stacking of each receiver function.

Events used in this study come from distances of $30^\circ$ – $100^\circ$ and have magnitudes greater than 5.5. Most of the events are from the east (the Indonesian and Western Pacific subduction zones), but 43 out of the 183 events come from other azimuths. For computing the receiver functions, a time-domain iterative deconvolution method (Ligorria and Ammon, 1999) was used, and to evaluate the quality of the receiver functions, a least-squares misfit criterion was applied. The misfit criteria provides a measure of the closeness of the receiver functions to an ideal case, and it is calculated by using the difference between the radial component and the convolution of the vertical component with the already determined radial receiver function. Receiver functions with a misfit of 10% and less were used in our analysis.

The receiver functions were filtered with a Gaussian pulse width of 1.6. Both radial and tangential receiver functions were examined for evidence of lateral heterogeneity in the crust and for dipping structure. Events with large amplitude tangential receiver functions were not used.

In applying the H-κ technique, it is necessary to select weights $w_1$, $w_2$, and $w_3$, and a value for Vp. More weight is typically given to the phase that is most easily picked. Given a range of plausible values for average crustal Vp values for rifted continental crust (5.8 to 6.8 km/s; e.g., Fuchs et al., 1997; Prodehl et al., 1994 and references therein), crustal thickness can vary by almost 4 km while the Vp/Vs ratio can change by 0.02, as shown in Table 2-1. Thus, when estimating errors for the H-κ
method, the uncertainty in mean crustal velocity, as well as the sensitivity of the results to
variations in the weights ($w_1, w_2, w_3$), need to be considered.

We used the H-$\kappa$ stacking together with a bootstrap algorithm and normally
distributed values of $V_p$ and phase weights to simultaneously find the best values of $H$
and $\kappa$, as well as the errors associated with these values. We began by incorporating
uncertainty in mean crustal velocity into error estimates for $H$ and $\kappa$ by specifying a
normal distribution of $V_p$ values so that 95% of the values selected fell between 5.9 and
6.5 km/s, with a mean value of 6.2 km/s, which is the mean crustal velocity in the
refraction line near ATD (Figure 2-1). For the weights ($w_1, w_2, w_3$), we also used a
normal distribution such that the sum of the weights add up to 1.00 but 95% of the values
for $w_1$ fall between 0.55 and 0.65 with a mean of 0.6, for $w_2$ they fall between 0.25 and
0.35 with a mean value of 0.3, and for $w_3$ they vary between 0.05 and 0.15 with a mean
value of 0.1.

Once values for $V_p$ and the weights were selected, we then used the bootstrap
algorithm of Efron and Tibshirani (1991), together with the H-$\kappa$ stacking, to estimate $H$
and $\kappa$ with statistical error bounds. While performing the H-$\kappa$ stacking, the contribution
of each of the receiver functions to the determination of $H$ and $\kappa$ was also weighted based
on the least squares misfit value of the receiver functions. The procedure of selecting $V_p$
and weights from the distribution described above and then performing the H-$\kappa$ stacking
with bootstrapping was repeated 200 times. After each time, new average values of $H$ and
$\kappa$ and their uncertainties were computed. It was found that after repeating the procedure
50 - 60 times (out of 200), the error values for $H$ and $\kappa$ stabilized.
H-κ stacking was performed on only the highest quality receiver functions (48 receiver functions spanning 10 years of data from 1993 to 2002). To examine azimuthal variation in crustal structure, the receiver functions were split into 4 groups from different azimuths (Figure 2-2), and the stacking was performed on the receiver functions within each group. The groups were based on the clustering of the events with backazimuth, except for group 1, which we chose to be similar to the range of backazimuths represented in the receiver functions used by Sandvol et al. (1998). The result for group 1 is shown in Figure 2-3, and the results for all the groups, as well as for an H-κ stack using all 48 receiver functions, are summarized in Table 2-2. The results are consistent between the various groups and give little indication of azimuthal variation in crustal structure.

For comparison with the results of Sandvol et al. (1998), we also computed stacks of the receiver functions by simply averaging them (Figure 2-4). Many more receiver functions were used for this (183 total), as compared to the H-κ stacking. The Ps conversion points at 25 km depth for all 183 events are shown on Figure 2-2.
\[ V_p = 6.2 \text{ km/s} \text{ w1 = 0.6, w2 = 0.3, w3 = 0.1} \]
\[ H = 22.2 \pm 0.4 \text{ km, Vp/Vs = 1.84} \pm 0.03 \]

Figure 2-3  H-k stack of receiver functions in group 1 for a mean crustal \( V_p \) of 6.2 km/s. To the left of each receiver function, the top number gives the event azimuth and the bottom number gives the event distance in degrees. Contours map out percentage values of the objective function given in the text.
2.4.2 Crustal Structure from Joint Inversion of Receiver Functions and Surface Wave Dispersion Measurements

Another approach to addressing the non-uniqueness inherent in interpreting receiver functions (besides H-κ stacking), is to invert jointly the receiver functions with observations that constrain crustal velocities, such as surface wave dispersion measurements. Joint inversions of receiver functions and surface wave dispersion measurements have been used by many authors to obtain improved models of crustal structure (e.g., Last et al., 1997; Ozalaybey et al., 1997; Du and Foulger, 1999; Julia et al., 2000, 2005). An advantage of this method compared to H-k stacking is that the inversion produces a model of shear wave velocities in the crust (in addition to Moho depth), and therefore details of crustal structure can be examined.

We used the method developed by Julia et al. (2000) to jointly invert the receiver functions from station ATD and surface wave group velocities. In the inversion we used three groups of receiver functions each corresponding to a range of ray parameters from 0.04 to 0.049 (with an average value of 0.044), from 0.05 to 0.059 (with an average value of 0.056), and 0.060 to 0.069 (with an average value of 0.065) (Figure 2-5). In addition, for each grouping of receiver functions, we computed and stacked two sets of receiver functions that have overlapping frequency bands: a low frequency band of $f \leq 0.5$ Hz and a high frequency band of $f \leq 1.25$ Hz. By inverting receiver function stacks over a range of ray parameter and frequency, details of crustal structure can often be imaged, such as sharp versus gradational seismic discontinuities (Julia et al, 2005).
Figure 2-4 (a) Receiver function stack and synthetic for a crustal thickness of 8 km for station ATD (redrawn from Sandvol et al. (1998)). Crustal reverberations not interpreted by Sandvol et al. (1998) are labeled along with the phase they interpreted to be the Moho $P_s$ conversion. (b) Simple stacks of receiver functions from this study for the different groupings shown in Figure 2-2.
Rayleigh wave group velocities between 10 to 40 sec period were used in the inversion. The dispersion measurements come from Benoit et al. (2005), who conducted a surface wave tomography study of eastern Africa by adding to the dispersion measurements of Pasyanos et al. (2001) new measurements made with data from the 2000-2002 Ethiopia broadband seismic experiment (Nyblade and Langston, 2002).

The inversion was performed using two different starting velocity models to determine how sensitive the inversion results are to the starting model. In the first case (Figure 2-5a), the starting model had a gradational velocity structure above 35 km over a half space. In the second case (Figure 2-5b), the starting model was based on Ruegg’s (1975) P-wave velocity profile converted to an S-wave model using a Poisson’s ratio of 0.31.

The results from the inversions using the different starting models are nearly identical (Figure 2-5), and therefore the inversion results do not appear to be influenced significantly by the starting model. The velocity models (Figure 2-5) show major discontinuities at depths of about 23 km and 25 km, with a change of shear-wave velocity from about 3.75 km/s to 4.2 km/s, which we interpret to be the Moho.

To estimate the uncertainties in our model results, we followed the approach of Julia et al. (2005) and repeatedly performed inversions using a range of weighting parameters, constraints, and Poisson's ratio. Similar to the results of Julia et al. (2000, 2003, 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 and uncertainties in the depth of discontinuities to be about 2-3 km.
Figure 2-5. Crustal shear wave velocity structure beneath station ATD from the joint inversion of receiver functions and Rayleigh wave group velocities for two different starting models; (a) a gradational velocity structure over a half space, and (b) a model based on Ruegg’s (1975) P-wave velocity profile converted to an S wave profile using a Poisson’s ratio of 0.31. For both (a) and (b), upper left panel displays predicted and observed receiver functions for three different ray parameters, plus high and low frequency band receiver functions in each ray parameter; lower left panel shows predicted and observed surface wave group velocities; top right panel shows the starting model for the inversion and resulting model; bottom right panel displays the joint inversion result and the PREM (Preliminary Reference Earth Model; Dziewonski and Anderson, 1981) model for comparison.
2.5 Comparison of results

The results of the H-κ stacking show little variation in crustal thickness or Poisson’s ratio with backazimuth. The crustal thickness is 23±1.5 km within each grouping and Poisson’s ratio ranges from 0.29 to 0.32 (Tables 2-1 and 2-2). Ps conversion points shown in Figure 2-2 illustrate that the receiver functions do not sample crust under the Gulf of Aden ridge. The results of the joint inversion are similar, indicating a Moho at 23 to 25 km depth.

Our results from both analyses (H-k stacking and joint inversion) are consistent with the seismic velocity structure given in Ruegg (1975) showing a velocity discontinuity at ~24-25 km depth from 7.1 km/s to 7.4 km/s. By selecting the P to s conversion on the receiver functions at ~3 secs after the P arrival as coming from the Moho, we are favoring an interpretation of crustal structure that identifies rocks with velocities as high as 7.1 km/s as crustal rock, similar to the interpretation of Makris and Ginzburg (1987) in other parts of the Afar. The Poisson’s ratio of 0.31±0.02 we obtained is consistent with the estimate from Ruegg (1975) of 0.28 to 0.33, as well as from Zandt and Ammon (1995) of 0.29±0.02. In addition, the shear velocity structure of the crust that we obtained from our joint inversion is remarkably similar to Ruegg's (1975) model at all depths.

As reviewed in section 2.3, Sandvol et al. (1998) reported a crustal thickness estimate for ATD of 8 km from analyzing receiver functions. In their analysis, they used a Poisson’s ratio of 0.25 and receiver functions from 11 events, 1 from the west and the
rest from the east (backazimuths of ~39° and ~123°). Thus, the stack of the receiver functions they modeled primarily reflected structure beneath the area of group 1 in Figure 2-2.

In comparison to Sandvol et al. (1998), our analysis of receiver functions is more comprehensive. We stacked many more receiver functions with good signal to noise ratios, and to check for heterogeneous structure, the receiver functions were examined in 4 groupings from different backazimuths, as mentioned previously. The stacked receiver function for group 1 (Figure 2-4b) show a Moho Ps phase that is much stronger than in the stack from Sandvol et al. (1998) (Figure 2-4a). We interpret the Ps phase picked by Sandvol et al. (1998) to represent a shallower discontinuity, perhaps the discontinuity at about 4 km or 10 km depth seen on the refraction line from Ruegg (1975). The Ps phase that we picked as the Moho Ps conversion is not as clear on the stack used by Sandvol et al. (1998) as in our stack (Figure 2-4b), but it is nonetheless apparent (Figure 2-4a). In addition, the two reverberation phases picked by our H-κ stacking algorithm can be seen in the stack used by Sandvol et al. (1998) at ~10 and ~12.5 seconds (Figure 2-4a). The synthetic receiver function obtained by Sandvol et al. (1998) for their preferred crustal model does not fit the arrivals of the receiver function at those times (Figure 2-4a).
2.6 Discussion

As described in section 2.3, the Moho in Afar has been interpreted either as an increase in velocity from ~6.4 km/s to ~6.9 km/s or as an increase in velocity from ~7.1 km/s to 7.4 or 7.5 km/s. We favor the latter interpretation because a Moho at ~23-25 km

Table 2-1. Results from H-κ stacking of receiver functions for the different groups shown in Figure 2-2 for a range of plausible mean crustal Vp.

<table>
<thead>
<tr>
<th>Group</th>
<th>Vp</th>
<th>Moho Depth (km)</th>
<th>Vp/Vs</th>
<th>Poisson's ratio (σ)</th>
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<td>20.7</td>
<td>1.84</td>
<td>0.29</td>
</tr>
<tr>
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</tr>
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depth produces a Ps conversion that arrives at the time of the clearest Ps conversion on the receiver function stack for each grouping (Figure 2-4b). The timing of the crustal reverberations is also well matched. In addition, the joint inversion yields a velocity structure for the crust with a clear velocity discontinuity at depths of 23-25 km (Figure 2-5). Our preferred interpretation from the H-k stacking (Table 2-2) is based on a mean crustal Vp of 6.2 km/s. Results summarized in Table 2-1 show that even for a mean crustal Vp of 5.8 km/s (i.e., similar to the mean crustal Vp from Ruegg's model if the 7.1 km/s layer is considered to be in the mantle) the Moho depth is still around 21 km.

<table>
<thead>
<tr>
<th>Group</th>
<th>Moho Depth (km)</th>
<th>Depth uncertainties (+/-)</th>
<th>Vp/Vs</th>
<th>Vp/Vs uncertainties (+/-)</th>
<th>Poisson's ratio (σ)</th>
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<td>48</td>
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</tbody>
</table>

To the north of the Gulf of Tadjoura, Ruegg's (1975) refraction profiles show an increase in velocity from 7.1 km/s to 7.4 km/s at about 20 km depth. Hence, the crust to the north of the Gulf of Tadjoura may be somewhat thinner than to the south, but there is
little indication of thin (i.e. 8-10 km thick) crust across areas of Djibouti away from the main spreading centres.

By combining our model of crustal structure beneath station ATD with the estimates of crustal structure from Makris and Ginzburg (1987), and Ruegg's (1975) velocity models for other parts of Djibouti, it appears that crustal thickness and composition may be fairly uniform across many parts of central and eastern Afar (Figure 2-1). Moho depths are between 20 - 25 km, mean Vp is ~6.2 km/s but ~6.9-7.0 km/s below a 2 - 5 km thick low velocity layer at the surface, and Poisson’s ratio is ~0.30 or higher.

What are the tectonic implications of this crustal structure for understanding the transition from continental rifting to sea-floor spreading? Mohr (1989) reviewed two plausible models for the nature of Afar crust which have different tectonic origins; 1) stretched (thinned) Precambrian continental crust modified by igneous intrusions, and 2) new igneous crust created by the addition of large volumes of mafic magma and lesser amounts of silicic magma capped by coeval flood lavas. Using estimates of crustal stretching, crustal structure, and sea-floor spreading parameters for the Red Sea and Gulf of Aden basins, together with the geology of Afar indicating a region affected predominantly by fissure volcanism, Mohr (1989) argued in favor of Model 2.

Poisson's ratio is particularly diagnostic of crustal modification and was not commented on by Mohr (1989). Below the melting point of many rocks, mineralogy is the most important factor influencing Poisson's ratio (Christensen, 1996), with the abundance of quartz and plagioclase feldspar having a dominant effect on the common igneous rocks. Granitic rocks have a Poisson's ratio of about 0.24, while intermediate
composition rocks have values of around 0.27 and mafic rocks about 0.30 (Christensen, 1996; Tarkov and Vavakin, 1982).

Crustal Poisson's ratios in Afar can be used to further argue against Model 1. The Precambrian basement surrounding Afar is mostly Neoproterozoic Mozambique Belt. Dugda et al. (2005) have reported that average Mozambique Belt crust in eastern Africa is ~40 km thick, has a Poisson’s ratio of ~0.25, and an average Vp of 6.5 km/s. It is very difficult to take such a crust and create the crust in Afar by simply stretching it. The resulting thinned crust would not have a sufficiently high Poisson's ratio to account for the observed Poisson's ratio in Afar of ~0.30, nor would it have a sufficiently high average Vp.

Model 2 could have an appropriately high Poisson's ratio to account for the observed high ratio found in Afar, but for this model to be viable there needs to be a reasonable explanation for how to generate the new igneous crust. One possibility is that the early Afar crust was initially oceanic in origin but was then modified by plume-generated melts from the mantle. Mohr (1978) suggested this possibility, noting that most of the anomalous crustal thickening would have occurred in the lower crustal layer (15-20 km thick in Afar compared to 4-5 km in oceanic crust), and that this amount of crustal thickening was consistent with the excessive magmatism found in Afar.

Another possibility, also proposed by Mohr (1989), is that the new igneous crust of Afar (Model 2) was generated by the intrusion of mantle-derived magmas breaking the crust. The formation of new igneous crust in this way is supported by Ebinger and Casey (2001), who suggested that in transitional rift settings extensional strain is accommodated locally within magmatic centres instead of along rift border faults. According to them,
border faults (detachments) play an active controlling role in the continental break-up process during the early stages of rifting, but, in the late synrift stages, crustal extension results primarily from dikes intruding into the crust. Recent results from the Ethiopian Afar Geoscientific Lithosphere Experiment indicate that within the northern end of the Main Ethiopian Rift strain is indeed being accommodated within magmatic segments (Keir et al., 2006; Bendick et al., 2006). Consequently, the Afar crust could have been created in a similar way, with the seismic structure described above reflecting the product of extension via the addition of large volumes of intrusive rock, predominantly mafic in composition, as dikes, sills and underplate.

2.7 Conclusions

A crustal thickness of 23±1.5 km and a crustal Poisson’s ratio of 0.29 to 0.33 have been obtained for station ATD in Djibouti (eastern Afar) from an H-κ stacking analysis of receiver functions, and a joint inversion of receiver functions and surface wave dispersion measurements. These results are consistent with the seismic velocity structure of the crust in Djibouti obtained from seismic refraction profiles (Ruegg, 1975). By combining our results of crustal structure beneath station ATD with the estimates of crustal structure elsewhere in Afar, it appears that crustal thickness and composition may be fairly uniform across many parts of central and eastern Afar, with Moho depths between 20 - 25 km. The high Poisson's ratio and high Vp throughout most of the crust indicates a mafic composition.
The high crustal Poisson's ratio and high mean crustal Vp throughout much of Afar, as well as the crustal thickness, are not consistent with models invoking crustal formation through stretching of pre-existing Precambrian crust. Instead, we suggest that crust in Afar consists predominantly of new igneous rock emplaced as part of the extensional process. During late synrift stages in the Main Ethiopian Rift, extensional strain is accommodated within magmatic segments through dike intrusion. In addition to diking, sill formation and underplating associated with the magmatic centres likely combine to help form new igneous crust. The formation of the new igneous (mafic) crust in Afar could have also taken place through modification of oceanic crust that was subsequently altered by plume-derived magmas from the mantle.
References


Chapter 3

Thin Lithosphere Beneath the Ethiopian Plateau Revealed by a

Joint Inversion of Rayleigh Wave Group Velocities and Receiver Functions

3.1 Introduction

In this chapter, we investigate the seismic velocity structure of the crust and upper mantle beneath Ethiopia and Djibouti by jointly inverting receiver functions and Rayleigh wave group velocities to obtain new insights into the thermal structure of the lithosphere. Much of Ethiopia and Djibouti has experienced Cenozoic hotspot tectonism, including flood basalt volcanism, plateau uplift, and rifting, and the impact of these processes on the lithospheric structure of the region is poorly understood, particularly in regard to the thickness of the lithosphere beneath the Ethiopian Plateau.

The hotspot tectonism began with basaltic volcanism in southwestern Ethiopia at about 45-40 Ma [Davidson and Rex, 1980; Zanettin et al, 1980; Berhe et al, 1987; WoldeGabriel et al, 1990; Ebinger et al, 1993; George et al, 1998]. The main episode of volcanism, however, initiated during the Oligocene (c. 29-31 Ma) with the emplacement of thick (500-2000 m) flood basalts and rhyolites in the future sites of the Red Sea, the easternmost Gulf of Aden, and the central Ethiopian Plateau [Hofmann et al., 1997; Mohr and Zanettin, 1988; Baker et al., 1996; Ayalew, 2002; Coulie et al., 2003; Kieffer et al., 2004]. Less voluminous syn-rift shield volcanoes formed between 30 and 10 Ma, locally
creating an additional 1000 to 2000 m of relief on top of the flood basalts [Berhe et al., 1987; Coulie et al., 2003]. Uplift of the Ethiopian Plateau commenced between 20 and 30 Ma [Pik et al., 2003].

The formation of the Red Sea and Gulf of Aden rifts began in the Oligocene when Africa started separating from Arabia, and can be linked to the complex geometry of collision along the Alpine-Himalayan chain [Wolfenden et al., 2005]. The opening of the Eastern Branch of the East African Rift System to form the Afar triple junction occurred long after Arabia separated from Africa. Extension commenced around 11 Ma in the northern sector of the Main Ethiopian Rift (MER) [Wolfenden et al., 2004; Chernet et al., 1998; WoldeGabriel et al., 1999] and at c. 18 Ma in southwestern Ethiopia. Wolfenden et al. [2004] suggest that there was a hiatus in volcanism between 6.5 and 3.2 Ma within the MER, after which time deformation migrated towards a narrow zone in the rift center, and that by 1.8 Ma volcanism and faulting had localized to magmatic segments within the rift.

Much of the Quaternary volcanism in the MER has occurred within the magmatic segments, but some has also occurred along the rift shoulders [Ayalew et al., 2006; Furman et al., 2006; Wolfenden et al., 2004]. Historical flows [Gibson, 1967] and elevated temperatures at shallow crustal depths in geothermal fields in the MER suggest that magmatic processes within the MER have been active recently. Ebinger and Casey [2001] and Casey et al. [2006] proposed that the magmatic segments act now as the locus of extension within this transitional rift setting rather than the rift border faults.
Rooney et al. [2005] have used xenoliths from locations in and along the sides of the MER to investigate lithospheric and sublithospheric processes under Ethiopia, and
their findings indicate that the lithospheric mantle beneath Ethiopia has been modified significantly by silicate melts, forming pervasive dikes and veins. The results from Rooney et al. [2005] indicating thermal modification of the lithosphere away from the rift axis are consistent with interpretations of shear-wave splitting observations and magnetotelluric data suggesting thermal modification of the lithosphere under the northwestern part of the Ethiopian Plateau [Kendall et al., 2005, 2006; Whaler and Hautot, 2006; Kier et al., 2005].

Because the Cenozoic volcanism, plateau uplift and rifting in the Horn of Africa cannot be explained easily by simple passive rifting related to the development of the Afar triple junction, many authors have invoked one or more mantle plumes to account for the hotspot tectonism. For example, two plumes (i.e., one at c. 45 Ma in southern Ethiopia and one at c. 30 Ma in the Afar region) have been invoked by George et al. [1998] and Rogers [2006], a single plume has been argued for by Manighetti et al. [1997], Ebinger and Sleep [1998], and Courtillot et al. [1999], and a superplume has been suggested by Ritsema et al. [1999] and Furman et al. [2004, 2006].

In an attempt to evaluate the different plume models, Benoit et al. [2006a, b] imaged upper mantle structure under Ethiopia using data from the 2000-2002 Ethiopia broadband seismic experiment [Nyblade and Langston, 2002]. By tomographically imaging the P and S velocity structure of the upper mantle and by examining topography on the transition-zone discontinuities, Benoit et al. [2006a, b] concluded that the hotspot activity could be the surface manifestation of a broad mantle upwelling that is part of the African Superplume located in the lower mantle beneath southern Africa. However,
because of limited model resolution, Benoit et al. [2006a, b] were not able to image structure at depths above 100 – 150 km and therefore could not comment on the structure of the lithosphere. In a related tomographic study using teleseismic P and S travel times recorded by the Ethiopia Afar Geoscientific Lithospheric Experiment (EAGLE), Bastow et al. [2005] were able to image velocity variations starting at depths as shallow as 75 km beneath the northern part of the MER. However, the region imaged by Bastow et al. (2005) did not extend very far across the Ethiopian Plateau or include Djibouti.

Building on the results of these previous studies, in this study we investigate further the nature of the lithosphere beneath Ethiopia and Djibouti, particularly beneath the Ethiopian Plateau where little is known about lithospheric mantle structure, by simultaneously inverting receiver functions and Rayleigh wave group velocities. The shear wave velocity models obtained from the joint inversion can be used to estimate the amount of heating and thinning of the mantle lithosphere that has occurred beneath the Ethiopian Plateau related to the Cenozoic hotspot activity, and provide new constraints on how much of the plateau uplift may have been caused by thermal alteration of the mantle lithosphere.

3.2 Data and Methodology

Seismic data collected between 2000 and 2002 by the Ethiopia broadband seismic experiment (EBSE) have been used for this study together with data from permanent stations in the region (the IRIS/GSN station FURI and the GEOSCOPE station ATD;
Figure 3-1). 22 of the 27 seismic stations in the EBSE were located either on the eastern or western side of the Ethiopian Plateau (Figure 3-1), and the rest of the stations were situated in the MER or Afar Depression. Additional details of the station configuration and recording parameters used in the EBSE can be found in Nyblade and Langston [2002].

In the joint inversion, we used receiver functions and fundamental mode Rayleigh wave group velocities. Love wave group velocities were not included because there are few high quality measurements available for Ethiopia [Benoit, 2005]. The two kinds of data used are complementary. Receiver functions are a time series that represent the radial impulse response (or, in the frequency domain, radial component transfer function) of the shallow structure of the Earth in the neighborhood of the seismic station [Langston, 1979], and can be used to resolve velocity contrasts at discontinuities and relative travel times [Ammon et al, 1990; Julia et al., 2000; Julia et al., 2005]. Rayleigh wave group velocities, on the other hand, can be used to constrain the average shear wave velocity between the discontinuities [Julia et al., 2000]. Joint inversions of receiver functions and surface wave dispersion measurements have been performed by many authors using data from a variety of tectonic settings [e.g., Ozalaybey et al., 1997; Du and Foulger, 1999; Julia et al., 2000; Tkalcić et al., 2006].

3.2.1 Joint Inversion of Receiver Functions and Rayleigh Wave Group Velocities

We have used the method developed by Julia et al. [2000] for the joint inversion. This method makes use of a linearized inversion procedure that minimizes 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. For the joint inversion, the two data sets must be consistent (i.e., they should sample the same study area). To make the contribution of each data set to the joint least squares misfit comparable, a normalization of the data set is necessary, and this is done using the number of data points and variance for each of the data sets. An influence factor can then be used to control the trade-off between fitting the receiver functions and the group velocity curves.

3.2.2 Rayleigh Wave Group Velocities

Rayleigh wave group velocities between 10 to 85 sec period from the model of Pasyanos [2005] and between 90 and 175 sec period from the Harvard model [Larson and Ekström, 2001] were used for the joint inversion. The model of Pasyanos [2005] includes group velocity measurements from Benoit et al. [2006c], who conducted a surface wave tomography study of eastern Africa by adding to the dispersion measurements of Pasyanos et al. [2001] new measurements made with data from the EBSE. Typical uncertainties in group velocities range between 0.01-0.02 km/s but increase to 0.02-0.03 km/s for the shortest periods in the Pasyanos [2005] model. Uncertainties in the Harvard group velocities are not specified [Larson and Ekström, 2001]. We did not observe any systematic differences in the two group velocity models where they overlap. In order to create a smooth dispersion curve for each station, we extracted dispersion curves from the
two models, joined them, and then applied a 3-point moving (running) average to obtain a smooth composite curve.

Figure 3-2 shows Rayleigh wave group velocity curves for three different regions in Ethiopia and for the Mozambique Belt in northeastern Tanzania. The three regions in

![Figure 3-2](image-url)
Ethiopia include the Main Ethiopian Rift (MER), and the eastern and western sides of the
Ethiopian Plateau. At periods greater than ~70s, lower group velocities can be seen for
the MER and the two Ethiopian Plateau regions (western and eastern sides), indicating a
difference in lithospheric structure between Ethiopia and the Mozambique Belt in
northeastern Tanzania. We use the group velocities for the Mozambique Belt in Tanzania
as a reference since the lithosphere under most of Ethiopia prior to rifting was
unperturbed Mozambique Belt lithosphere [Burke and Sengor, 1986; Shackleton, 1986;
Berhe, 1990; Vail, 1988, 1985; Kroner et al, 1987]. Although the Mozambique Belt in
northeastern Tanzania sits on the eastern edge of the East African Plateau and may thus
be somewhat perturbed by the Cenozoic tectonism affecting most of eastern Africa, this
region provides the best possible estimate from within eastern Africa for the structure of
unperturbed Mozambique Belt lithosphere [Weeraratne et al., 2003; Julia et al., 2005].

3.2.3 Receiver Functions

Receiver functions were computed using seismograms from teleseismic events
between distances of 30° and 95° with magnitudes greater than 5.5. A list of events used
in this study can be found in Dugda et al. [2005]. Most of the events are from the east
(the Indonesian and Western Pacific subduction zones) or the northeast (Hindu Kush –
Pamir region).

The time-domain iterative deconvolution method of Ligorria and Ammon [1999]
was employed to compute the receiver functions, and the quality of the receiver functions
was evaluated using a least-squares misfit criterion. This misfit criterion provides a
measure of the closeness of a receiver function to an ideal case, and it is calculated by using the difference between the radial component seismogram and the convolution of the vertical component seismogram with the already determined radial receiver function. Usually, receiver functions with a fit of 90% and above were used in the inversion. However, in a few cases, when it was difficult to get a reasonable number of receiver functions for a station, receiver functions with fits of 70-90% were included. The receiver functions were filtered with Gaussian pulse widths of 1.0 and 2.5 to obtain low and high frequency band receiver functions, respectively. Radial and tangential receiver functions were examined for evidence of lateral heterogeneity and for dipping structure. Events with large amplitude tangential receiver functions were not used.

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, from 0.05 to 0.059, and from 0.060 to 0.069 (Figure 3-3b). 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 \( f \leq 0.5 \) Hz (Gaussian of 1.0), and a higher frequency band of \( f \leq 1.25 \) Hz (Gaussian of 2.5). By inverting receiver function stacks over a range of ray parameter and frequency, details of lithospheric structure can be imaged, such as sharp versus gradational discontinuities [Julia et al, 2005; Cassidy, 1992].

The initial model 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 results do not depend on the starting model but that velocities at lithospheric depths (≤ ~150-100 km) trade-off with velocities below about 190 km depth. To minimize this trade-off, we forward modeled the structure below 190 km depth by a priori fixing the S-velocities in this depth range. The velocity structure below 190 km depth was determined through a trial-and-error process by finding models that best fit the 140 – 175 s group velocities for several stations in each tectonic region. This was done by fixing velocities below 190 km depth between a range
of −15 to +10% of PREM velocities [Dziewonski and Anderson, 1981], while at the same time inverting for the velocity structure above 190 km depth.

Figure 3-3. Figure illustrating, for station ARBA, the procedure used to determine structure below 190 km depth. The four columns show different models tested for structure below 190 km depth using velocities 15%, 10%, 7%, and 5% less than PREM [Dziewonski and Anderson, 1981].

(a) Observed (black line) and predicted (gray line) group velocity dispersion curves. The highlighted part in the dispersion curves shows the fitting of the longest period group velocities for the four different models tested.

(b) Observed (black line) and predicted (gray line) receiver functions. See text for explanation of the different receiver functions.

(c) The shear wave velocity models obtained from the joint inversion (black line) and the PREM shear wave velocity model (gray line) for reference. The 10% less than PREM model for shear velocities below 190 km depth gives the best fit to the longest period group velocities. The difference in the fit of the receiver functions is not significant between the models.
Figure 3-3 shows models with velocities of 5, 7, 10 and 15% less than PREM below 190 km depth for one station. For each station tested, the model with velocities of 10% less than PREM (Figure 3-3b) 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 190 km depth to values 10% less than PREM for all stations and inverted for velocity structure above 190 km depth. The Poisson’s ratios in the velocity models were held constant during the inversion. Poisson’s ratios for the crust were taken from Dugda et al. [2005], and for the mantle we used Poisson’s ratio from the PREM model. The Poisson’s ratios used for the crust are average values obtained for the crust beneath each station. Further testing of our models indicated that our results are not sensitive to reasonable variations in Poisson’s ratio.

To estimate the uncertainties in our model results, we followed the approach of Julia et al. [2005] and repeatedly performed inversions using a range of weighting parameters, constraints, and Poisson's ratio. Similar to the results of Julia 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, and 0.2 km/s in the lower part of the upper mantle, and uncertainties in the depth of discontinuities to be about 2-3 km in the crust and uppermost mantle.

3.3 Results

Results from the joint inversion for all stations are shown in Figure 3-4, and in Figure 3-5 we show results for selected stations along two E-W profiles highlighting the
structure above 100 km depth. A summary of our results plus a comparison with results from other studies follows.

### 3.3.1 Crust

The crustal thickness beneath both the eastern and western sides of the Ethiopian Plateau ranges between 35 and 40 km, and for the MER, the crustal thickness ranges between about 30 and 35 km. For the two stations in Afar (ATD and TEND), the crustal thickness is about 25 km. These crustal thickness estimates agree closely with the structure obtained by Dugda et al. [2005], Stuart et al. [2006], and Dugda and Nyblade [2006] using the H-κ receiver function stacking technique of Zhu and Kanamori [2000] (Figure 3-5), where $H = \text{Moho depth}$ and $κ = \frac{V_p}{V_s}$. The H-κ stacking technique provides robust estimates of crustal thickness and Poisson’s ratio by incorporating the P-to-S converted phase from the Moho and two later arriving crustal reverberations in a stacking procedure.

The average crustal structure (e.g., Moho depth and average crustal velocity) obtained from the joint inversion also agrees closely with the structure obtained from seismic refraction profiles in places where seismic stations are within about 50 km of the refraction profiles [Makris and Ginzburg, 1987; Mackenzie et al., 2005; Maguire et al., 2006] (Figure 3-5). Only a few stations lie closer than about 20 km from the refraction profiles and thus a more detailed comparison of crustal structure, such as intercrustal layering, between the ensemble of our results and the refraction profiles is not warranted.
3.3.2 Upper Mantle

For the uppermost mantle, if we assume a Poisson’s ratio between 0.28 and 0.30 and compare our S-wave velocity structure to the P-wave velocity structure from several refraction profiles [Makris and Ginzburg, 1987; Mackenzie et al., 2005; Maguire et al., 2006], we also find good agreement. The average Sn velocity beneath the Ethiopian Plateau is about 4.3 km/s, while for the MER it is about 4.0 to 4.1 km/s, corresponding to Pn velocities found by Mackenzie et al. [2005] of ~7.9-8.0 km/s for the Ethiopian Plateau and ~7.6-7.7 km/s for the MER. The assumed Poisson’s ratios are not unreasonable considering the there may be partial melts in some areas of the Western Plateau [Whaler and Hautot, 2006] and that very high Poisson’s ratios are characteristic of the Main Ethiopian Rift and Afar.
Figure 3-6 shows two profiles, one crossing the western side of the Ethiopian Plateau, the MER, and the eastern side of the Ethiopian Plateau (Profile A-A’), and the other crossing the western side of the Ethiopian Plateau, the MER, and the Afar (Profile B-B’). It can be seen from this figure that the lithosphere beneath the plateau has a thin “lid” structure that extends from the Moho to about 60 or 80 km depth, while the lid beneath the MER and Afar is either much thinner or nonexistent. The maximum shear
wave velocity in the upper mantle lid is about 4.3 km/s for both sides of the Ethiopian Plateau, and is underlain by a low velocity zone characterized by velocities of about 4.1 km/s. When there is a lid beneath the MER/Afar, the maximum velocity in the lid reaches a value of about 4.1-4.2 km/s, with velocities of about 4.0 km/s under the lid.

Figure 3-5. Graph showing the correlation between the Moho depth estimates from the joint inversion technique and previous studies. A comparison with Moho depths from refraction profiles is only made where seismic stations are within about 50 km of a refraction profile. See text for further explanation.

In the tomographic image of Bastow et al. [2005], at a depth of 75 km there is a
maximum difference of about 5% in shear velocities beneath the Ethiopian Plateau and the MER. The difference in shear wave velocities beneath the two sides of the Ethiopian Plateau and the MER at 75 km depth in our models is ~0.2 km/s, or ~5%, consistent with the results of Bastow et al. [2005].

3.4 Discussion

The main finding of this study is a thin seismic lid (between the Moho and ~60 to ~80 km depth) under the Ethiopian Plateau. We also find essentially little or no lid under the MER and Afar, in accord with the results of the EAGLE project [Bastow et al., 2005; Mackenzie et al., 2005; Maguire et al., 2006]. As is common, we interpret the seismic lid to represent the lithospheric mantle. To examine the extent to which the Ethiopian lithosphere has been perturbed, we use the velocity structure from relatively unperturbed Mozambique Belt in northeastern Tanzania for comparison, as discussed previously.
In Figure 3-7, we compare the upper mantle structure obtained in this study for each station with the result obtained by Julia et al. [2005] for Mozambique Belt upper
mantle structure beneath northeastern Tanzania, also obtained using a joint inversion of receiver functions and surface wave dispersion measurements. By comparing the profiles, it can be seen that the lithosphere beneath Ethiopia and Afar, including Djibouti, relative to northeastern Tanzania, has been pervasively modified. The maximum velocities beneath the Ethiopian Plateau are 0.3 km/s to 0.4 km/s less than under the Mozambique Belt in northeastern Tanzania to depths of ~100-120 km. The difference in the lid structure is also illustrated in Figure 3-8, which shows the average shear wave velocity structure beneath the Ethiopian Plateau and the Mozambique Belt. The lid under the Ethiopian Plateau has an average maximum velocity of about 4.3 km/s, whereas the lid under the Mozambique Belt in Tanzania has an average maximum velocity of about 4.6 km/s [Julia et al., 2005; Weeraratne et al., 2003]. The lid structure under the MER and Afar is even more perturbed.
What is the cause of this large perturbation in lithospheric mantle structure beneath Ethiopia? Possible causes of seismic velocity variations in the mantle include changes in temperature, grain size, fluid content (melt), dislocations and composition, but
temperature variations are most often invoked [Faul and Jackson, 2005]. The derivative of the shear wave velocity with respect to temperature in the upper mantle depends mainly on grain size and the particular temperature at which this derivative is considered. The derivative also varies with seismic frequency. Jackson et al. [2002] and Faul and Jackson [2005] report a derivative for shear wave velocity with respect to temperature of ~1.2 m/s/K for an upper mantle grain size of 10 mm at an average temperature of 1250°C. Thus, to reduce the shear wave velocity in the mantle lithosphere by ~0.3 km/s (Figure 3-8), rock temperatures must increase by ~250K.

Figure 3-8. Average lithospheric structure of the Ethiopian Plateau (black line) and the Mozambique Belt under Tanzania (grey line). A pronounced reduction in the shear wave velocity of the seismic lid under Ethiopia is seen compared to the Mozambique Belt under Tanzania. The average lithospheric structure for Ethiopia is obtained by averaging shear velocities from both the eastern and western sides of the Ethiopian Plateau, as they are similar. The velocity profile for the Mozambique Belt in Tanzania is for station KIBE [Julia et al., 2005].
Because of the strong indications from previous geochemical and geophysical studies for thermally perturbed upper mantle structure beneath Ethiopia, as reviewed in section 3.1, to explain the modified shear velocity structure of the Ethiopian mantle lithosphere we consider the effect of a thermal plume instantaneously eroding the bottom of the lithosphere at the time of the flood basalt volcanism (c. 30 Ma) and then remaining under the thinned lithosphere to the present day. While instantaneous thinning of the mantle lithosphere does not capture the complexity of geodynamic (hotspot) processes in the region, it nonetheless provides a reasonable approach for assessing first-order changes in the thermal structure of the mantle lithosphere resulting from a plume impinging on the lithosphere.

We model the thermal effect of the plume on the instantaneously-thinned lithosphere by considering the transient thermal response within a slab whose lower boundary temperature is increased by $\Delta T_L$, according to the equation

$$\Delta T(z,t) = \Delta T_L \left\{ \frac{z}{L} + \frac{2}{\pi} \sum_{n=1}^{\infty} \frac{1}{n} \cos(n\pi) \sin \left[ \frac{n\pi}{L} \right] \exp \left[ -\frac{n^2 \pi^2 \alpha t}{L^2} \right] \right\}$$  \hspace{1cm} (3-1)$$

where $\Delta T(z,t)$ is the anomalous temperature at depth $z$ and after time $t$, $L$ is slab thickness, and $\alpha$ is thermal diffusivity [Carslaw and Jaeger, 1959]. Thus, a specific type of plume (i.e., single plume, multiple plumes, superplume) is not modeled, but rather the consequence of simply thinning and heating the lithosphere by any kind of plume. For a thermal diffusivity of 32 km$^2$/m.y. [Turcotte and Schubert, 2002], and different thicknesses of the instantaneously thinned lithosphere, temperatures within the
lithosphere at depths of 30 and 60 km are shown in Figure 3-9. If the Ethiopian lithosphere was eroded by the plume from some initial thickness to a thickness of 70 or 80 km, and if the plume material sitting beneath the lithosphere remained some 300 to 400K hotter than the surrounding mantle [Schilling, 1991; Wyllie, 1988; McKenzie and Bickle, 1988], then the temperature at ~60 km depth in the lithosphere would increase by ~250K or more after 30 Ma (Figure 3-9), sufficient to account for the reduction of the maximum shear wave velocity in the lid beneath the Ethiopian Plateau of about 0.3 km/s. From this analysis, we conclude that thermal alteration of the Ethiopian lithospheric mantle can, to a first order, account for the low shear wave velocities in the lithosphere, provided that the lithosphere was thinned to a thickness of ~70-80 km at c. 30 Ma and that the warm mantle material remained under the thinned lithosphere since that time.
The amount of uplift of the Ethiopian Plateau due to the thermally perturbed mantle lithosphere can be estimated from $\alpha l \Delta T$, where $\alpha$ is the coefficient of thermal expansion, $l$ is the thickness of rock being heated, and $\Delta T$ is the temperature perturbation. If we consider the altered mantle lithosphere to be 40 km thick with a temperature perturbation of 250K, then ~0.30 km of uplift is created isostatically, assuming $\alpha = 3 \times 10^{-5} \text{K}^{-1}$. [Turcotte and Schubert, 2002]. And if we assume that the lithosphere was 120 km thick prior to 30 Ma, an additional 0.36 km of uplift would result from the warm plume.
material with a $\Delta T = 300\text{K}$ replacing the $\sim 40$ km thick bottom section of the lithosphere. Thus, a significant portion (0.66 km) of the observed plateau uplift in Ethiopia of 1–1.5 km (Figure 3-1) can be attributed to the thermal alteration of the lithosphere.

To explain the full 1–1.5 km of uplift across the Ethiopian Plateau, other sources of buoyancy may be required, which could come from thermally perturbed structure deeper in the upper mantle and/or modifications to crustal structure. Regional and global tomographic models [e.g., Benoit et al., 2006a, b; Bastow et al., 2005; Debayle et al., 2001; Ritsema et al., 1999; Grand, 2002] show low velocity anomalies beneath Ethiopia and Afar in the upper mantle extending to depths of $\geq 400$ km and these models are consistent with the lower-than-PREM velocities in our models (Figure 3-4) extending to $\sim 400$ km depth. All of the models suggest that the upper mantle beneath Ethiopia is thermally perturbed to great depths, providing possible sources of thermal buoyancy that could play a role in creating the Ethiopian Plateau, both isostatically and dynamically [e.g., Daradich et al., 2003]. In addition, crustal structure across the Ethiopian Plateau is variable, with underplating and thermal perturbations providing yet other possible sources of buoyancy [Mackenzie et al., 2005; Maguire et al., 2006; Stuart et al., 2006; Whaler and Hautot, 2006].

Although the focus of this discussion has been on the lithosphere beneath the Ethiopian Plateau, the thin or non-existent mantle lithosphere under the MER and Afar is also noteworthy. The seismic velocity structure of the uppermost mantle under the MER and Afar is similar to that found beneath the Kenya Rift where it has been commonly attributed to extension of the lithosphere together with upwelling warm plume material
that has thermally modified the thinned lithosphere [Fuchs et al., 1997; Prodehl et al., 1994, and references therein]. Such a combined plume-rift explanation can also account for the nature of the uppermost mantle structure under the MER and Afar, as suggested by many authors [e.g., Maguire et al., 2006; Stuart et al, 2006; Mackenzie et al., 2005; Bastow et al., 2005; Makris and Ginzburg, 1987].

Finally, the thinned and thermally perturbed mantle lithosphere under the MER, Afar, and Ethiopian Plateau has important consequences for interpreting seismic estimates of mantle anisotropy derived from shear-wave splitting. Shear wave splitting studies for Ethiopia using teleseismic data, showed delay times of ~0.5 to 1.7s for the MER, Afar, and the Ethiopian Plateau, with fast-polarization directions generally parallel to the orientation of the rift system [Gashawbeza et al., 2004; Ayele et al., 2005; Kendall et al., 2005, 2006]. Different interpretations for the source of the anisotropy have been published. Gashabeza et al.[2004] suggest that the anisotropy could be controlled by fossil anisotropy in the Mozambique Belt lithosphere, while the studies by Ayele et al., [2005] and Kendall et al. [2005, 2006] attribute the anisotropy to melt-filled microcracks and dikes in the lithosphere. And Kier et al. [2005] provide for a similar interpretation using estimates of seismic anisotropy obtained from regional seismicity. The average SKS splitting time is about ~1.0 second for the region, comparable to the splitting times observed in and around the Kenya Rift [Walker et al., 2004]. Walker et al. [2004] show that for the hot, thin mantle beneath and surrounding the Kenya Rift, fossil anisotropy in the lithosphere is not likely a dominant source of anisotropy. A similar argument can be used for Ethiopia, thus calling into question interpretations that attribute the shear-wave
splitting observations to fossil anisotropy in the lithosphere.

3.5 Summary

The seismic velocity structure of the crust and upper mantle beneath Ethiopia and Djibouti has been investigated in this study using a joint inversion of receiver functions and Rayleigh wave group velocities. Crustal structure obtained from the joint inversion is similar to crustal structure reported in previous studies, with crustal thicknesses of 35 to 44 km beneath the Ethiopian Plateau, and 25 to 35 km beneath the Main Ethiopian Rift and the Afar. The lithospheric mantle beneath the Ethiopian Plateau has a maximum shear wave velocity of about 4.3 km/s and extends to a depth of ~70-80 km. Beneath the MER and Afar, the lithospheric mantle has a maximum shear wave velocity of 4.1-4.2 km/s and extends to a depth of at most 50 km.

Prior to the Cenozoic hotspot tectonism, Ethiopia and Djibouti were underlain with unperturbed Mozambique Belt lithosphere. In comparison to Mozambique Belt lithosphere in Tanzania along the edge of the East African Plateau where the lid extends to depths of ~100-125 km and has a maximum shear velocity of 4.6 km/s, the mantle lithosphere under the Ethiopian Plateau appears to have been thinned by ~30-50 km and the maximum shear wave velocity reduced by ~0.3 km/s. The results of a 1D conductive thermal model suggest that the shear velocity structure of the Ethiopian lithosphere can be explained by a mantle plume, if the plume instantaneously thinned the lithosphere by at least 30–50 km at the time of the Afar flood basalt volcanism (c. 30 Ma), and if the warm plume material has remained beneath the lithosphere since then. About 45-65% of
the 1-1.5 km of plateau uplift in Ethiopia can be attributed to the thermally perturbed lithospheric mantle structure.
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Chapter 4

Comparison of the Lithosphere Beneath Kenya and Ethiopia

From Joint Inversion of Receiver Functions and Rayleigh Wave Dispersion Velocities

4.1 Introduction

In this chapter, I examine the shear wave velocity structure of the crust and upper mantle beneath Kenya by simultaneously inverting receiver functions and Rayleigh wave group and phase velocities. Much of Kenya has experienced Cenozoic volcanism, plateau uplift, and rifting. The impact of these processes on the lithospheric structure of the region has been examined by many authors, but the shear-wave velocity structure of the lithosphere beneath Kenya is not well known. Insights into the nature of the lithosphere beneath Kenya are obtained by comparing the lithospheric structure beneath Kenya to other parts of the East African Plateau in Tanzania and also to the Ethiopian Plateau. In this comparison, the shear wave velocity models obtained from the joint inversion are used to estimate the amount of heating and thinning of the lithospheric mantle that has occurred beneath and surrounding the Kenya Rift related to the Cenozoic hotspot activity found there.

The models of lithospheric structure presented in this chapter provide important constraints on the thermal structure of the lithosphere and the extent to which the rifting
and magmatism has altered and/or replaced the pre-existing lithosphere beneath Kenya. The results can in turn be used to advance our understanding of the processes responsible for the rifting, magmatism, and uplift found in the region.

### 4.2 Geologic Background

The Kenya Rift is part of the Eastern Branch of the East African Rift System (EARS). The Mozambique Belt, within which the Eastern Branch of the EARS has developed, extends from Ethiopia south through Kenya, Tanzania and Mozambique and it is often interpreted as a Himalayan-type continental collision zone [Burke and Sengor, 1986; Shackleton, 1986]. Volcanism occurred in northern Kenya and southwestern Ethiopia before the occurrence of the Afar Flood Volcanism at around 30 Ma. The volcanism that began about 45-40 Ma in southwestern Ethiopia propagated to the south through Kenya and into northern Tanzania [Davidson and Rex, 1980; Zanettin et al., 1980; Berhe et al., 1987; WoldeGabriel et al., 1990; Ebinger et al., 1993; George et al., 1998]. The earliest volcanism in Kenya started in the Turkana region of northern Kenya around ~35 – 40 Ma [Furman et al., 2006a; MacDonalds et al., 2001]. Two additional episodes of volcanism also occurred in the Turkana region, one between about 26 and 16 Ma, and the other between 3.5 Ma to the present [Furman et al., 2006a]. Magmatic activity in other parts of northern Kenya began in the Oligocene (c. 30 Ma) [Morley et al., 1992; Ritter and Kaspar, 1997], while the 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]. The earliest Cenozoic rifting occurred c. 25
Ma in northern Kenya [Morley et al., 1992; Hendrie et al., 1994], and propagated southward reaching central Tanzania by c. 1 Ma [Baker, 1986; Foster et al., 1997]. Timing of plateau uplift in this part of east Africa remains poorly constrained, although there is some evidence for Neogene rift flank uplift [Noble, 1997; van der Beek et al., 1997].

The volcanism, rifting and plateau uplift in eastern Africa cannot be explained easily by simple passive rifting related to the development of the Afar triple junction, and thus many authors have used a plume model to account for the Cenozoic tectonism, for example, a single plume [e.g., Manighetti et al., 1997; Ebinger and Sleep, 1998; Courtillot et al., 1999], two plumes [Ebinger et al., 1989; George et al., 1998; Rogers, 2006], three plumes [Burke, 1996], a chemically heterogeneous plume [Furman et al., 2004, 2006b], or a superplume [e.g., Ritsema et al., 1999; Furman et al., 2004, 2006b; Rogers, 2006]. Nyblade [2002] summarized the plume and nonplume models that have been forwarded for east Africa and evaluated the geodynamic processes in light of the crust and upper mantle structure of the region. Benoit et al. [2006], for the mantle beneath Ethiopia, reported a broad thermal anomaly at depth, consistent with global and continental-scale seismic models [Ritsema et al., 1998; Grand, 2002; Ritsema and Allen, 2003]. Furman [2006b] proposed a modified single plume model in which multiple plume stems are rising from a superplume at depth, which is consistent with both the studies by Nyblade [2002] and Benoit et al.[2006].
4.3 Data and Method

Seismic data collected between 2001 and 2002 by the Kenya broadband seismic experiment (KBSE) have been used for this study together with data from permanent stations in the region (the IRIS/GSN stations NAI and KMBO, Figure 4-1). Six of the eight seismic stations analyzed here were located either on the eastern or western side of the Kenya (Gregory) Rift (Figure 4-1), and the other 2 stations were situated within and at the edge of the Kenya Rift. Nyblade and Langston [2002] provide additional details on the station configuration and recording parameters used in the KBSE.

In the joint inversion scheme applied, we used receiver functions and fundamental mode Rayleigh wave phase and group velocities. The two kinds of data, i.e. receiver functions and surface wave dispersion measurements, give complementary information. Receiver functions are time series that represent the radial impulse response of the shallow structure of the Earth in the neighborhood of the seismic station [Langston, 1979], and can be used to resolve velocity contrasts at discontinuities and relative travel times [Ammon et al, 1990; Julia et al., 2000; Julia et al., 2005]. Rayleigh wave phase and group velocities, on the other hand, can be used to constrain the average shear wave velocity between the discontinuities [Julia et al., 2000].

In this study, we have applied the method developed by Julia et al. [2000] for the joint inversion. This method makes use of a linearized inversion procedure that minimizes 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.

Figure 4-1 Map showing the topography of the study region, seismic station locations from the 2001-2002 Kenya Broadband Seismic Experiment, and permanent seismic stations (NAI and KMBO). Lines AA' and BB' show the location of the velocity profiles given in Figures 4-2 and 4-3.
Rayleigh wave group velocities between 10 to 45 sec period from the model of Pasyanos [2005], and Rayleigh wave phase velocities between 60 and 140 sec period from Weeraratne et al. [2003] were used in the joint inversion. The Rayleigh wave group velocities were extracted from a regional surface wave tomography study by Pasyanos [2005], while the Rayleigh wave phase velocities were obtained from a regionalized inversion.

Typical uncertainties in the group velocities range between 0.01-0.02 km/s but increase to 0.02-0.03 km/s for the shortest periods in the Pasyanos [2005] model. These uncertainties in the group velocities represent total uncertainties and contain the information on how well the dispersion curves are fitting at each period. The total uncertainties are obtained by taking the square root of the sum of model uncertainty and measurement uncertainty. A modified bootstrapping technique is implemented in the inversions to calculate model uncertainty while measurement uncertainty is determined by observing how much the velocities are deviating from similar paths. The measurement uncertainties encompass such effects as source mislocation.

Uncertainties in the phase velocities are below 0.01 km/s for the background region where the Kenya stations outside the rift are located, but for the Kenya Rift the uncertainties are higher, about 0.03-0.05 km/s [Weeraratne et al., 2003]. To create smooth dispersion curves for each station, we selected phase velocity dispersion curves for the region of study and extracted group velocity dispersion curves and then applied a 3-point moving average.

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 Ligorria and Ammon [1999] was employed to compute the receiver functions, and the quality of the receiver functions was evaluated using a least-squares misfit criterion. The receiver functions were filtered using lowpass filters with Gaussian pulse widths of 1.0 and 2.5 to obtain lower and higher frequency band receiver functions, respectively. Radial and tangential receiver functions were examined for evidence of lateral heterogeneity and for dipping structure. Events with large amplitude tangential receiver functions were not used.

In the joint inversion, we tried to use three groups of receiver functions, each corresponding to a range of ray parameters from 0.04 to 0.049, from 0.05 to 0.059, and from 0.060 to 0.069. 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 \( f \leq 0.5 \) Hz (Gaussian of 1.0), and a higher frequency band of \( f \leq 1.25 \) Hz (Gaussian of 2.5).

The initial model 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 results do not depend on the starting model but that velocities at lithospheric depths (\( \leq \sim 150-100 \) km) trade-off with velocities below about 200 km depth. To minimize this trade-off, we forward modeled the structure below 205 km depth by a priori fixing the S-velocities in this depth range. The velocity structure below 205 km depth was determined through a trial-and-error process by finding models that best fit the 120-140 s phase
velocities for each station. This was done by fixing velocities below 205 km depth between 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 5-7% 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 5-7% less than PREM and inverted for velocity structure above 205 km depth.

Similar to the results of Julia 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, and 0.2 km/s in the lower part of the upper mantle, and uncertainties in the depth of discontinuities to be about 2-3 km in the crust and uppermost mantle.

4.4 Results

Results from the joint inversion for all stations are shown in Figure 4-2. The Mozambique Belt in northeastern Tanzania provides the best possible estimate from within eastern Africa for the structure of unperturbed Mozambique Belt lithosphere [Weeraratne et al., 2003; Julia et al., 2005]. Therefore, an average velocity structure from northeastern Tanzania is used as a reference model in Figure 4-2. In Figure 4-3 we show results for the stations along a NW-SE profile considering the structure above 100 km
depth and make a comparison to structure beneath Ethiopia. A summary of our results plus a comparison with results from previous studies follows.

The crustal thickness to the eastern side of the Kenya Rift lies between 38 and 42 km and that to the western side of the Kenya Rift ranges between 37 and 38 km. For the central Kenya Rift, a crustal thickness of 30 km has been obtained. These crustal thickness estimates agree closely with the structure obtained by Dugda et al. [2005] using the H-κ receiver function stacking technique of Zhu and Kanamori [2000], where H = Moho depth and κ = Vp/Vs. The H-κ stacking technique provides robust estimates of crustal thickness and Poisson’s ratio by incorporating the P-to-S converted phase at the Moho and two later arriving crustal reverberations in a stacking procedure. The average crustal structure (e.g., Moho depth) obtained from the joint inversion also agrees closely with the structure obtained from KRISP seismic refraction profiles in places where seismic stations are within about 50 km of the refraction profiles [Fuchs et al., 1997; Prodehl et al., 1994, and references therein] (Figure 4-4).

For the uppermost mantle, if we assume Poisson’s ratios 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 EAP in Kenya is about 4.5 km/s, while for the Kenya Rift it is about 4.3 km/s, corresponding to Pn velocities found by the KRISP group of ~8.1-8.3 km/s beneath the EAP in Kenya and about ~7.5-7.8 km/s beneath the Kenya Rift.
Figure 4-2 Shear wave velocity models for all the eight stations analyzed in this study, along with the receiver functions and surface wave velocities used. Shear wave velocity models for Kenya stations (black lines) are compared to an average shear wave velocity model for the Mozambique Belt lithosphere in Tanzania (gray line). Station names are indicated at the top of each receiver function, surface wave velocity and the shear wave velocity model. The average model for the Mozambique Belt in Tanzania is calculated using stations KIBA, KIBE, and HALE from the study of Julia et al. [2005].
4.5 Discussion

In Figure 4-2, we compare the upper mantle structure obtained in this study for each station with an average model obtained from the study of Julia et al. [2005] for Mozambique Belt upper mantle structure beneath northeastern Tanzania, also obtained

Figure 4-3 (a) A profile of shear wave velocity along line A-A’ (given in Figure 4-1) is used to show the seismic velocity structure and its variation beneath the Kenya Rift and the East African Plateau in Kenya. The lid in each profile is defined as the lithospheric mantle and is shown with a gray shading. The profile shows a thin lid left beneath the Kenya Rift.
(b) A profile of the lithosphere beneath Ethiopia, as indicated along B-B’ on Figure 4-1, is provided for a comparison with the lithosphere beneath Kenya. Station names are indicated at the top of each velocity model. Vertical line indicating the 4.0 km/s velocity is superimposed on each velocity model.
using a joint inversion of receiver functions and surface wave dispersion velocities. For this discussion we define the seismic lid as the lithospheric mantle. By comparing the profiles, it can be seen that the lithosphere beneath the EAP in Kenya, relative to northeastern Tanzania, has not been affected significantly by the hotspot tectonism except beneath the Kenya Rift. Under the Kenya Rift, the lithosphere is highly perturbed and thinned, with a thickness of only about 75 km, when compared to 100 – 125 km thick lithosphere under northeastern Tanzania. The percentage reduction in shear velocity when crossing the Kenya Rift is similar to the P-wave velocity reduction at depths of 50 and 200 km from previous studies [Park and Nyblade, 2006; Prodehl et al., 1997] using body wave tomography. However, at 100 km depth, the shear wave speed is slower compared to the reduction in the P-wave velocity found by Park and Nyblade [2006]. This difference could be due to the presence of some partial melt at that depth, as proposed by Green et al. [1991], which may affect the shear waves more than the P-waves. On the other hand, our results for the percentage S-wave velocity reduction is close to the P-wave velocity reductions found by some other studies at the 100 km depth level [Achauer et al., 1994; Slack et al., 1994; Green et al., 1991].
Table 4-1 summarizes and compares the percentage reductions in S-wave velocity from this study to percentage reduction in the P-wave velocity from previous studies at different depths. The percentage reductions from this study are similar to percentage reductions found in previous studies at 50 km, 150, and 200 km depths [Park and Nyblade, 2006; Prodehl et al., 1997; Achauer et al., 1994; Slack et al., 1994; Green et al., 1991].

Figure 4-4 Graph showing correlation between crustal thickness estimates from the joint inversion technique and previous studies. A comparison of crustal thickness from KRISP refraction profiles is made only where seismic stations are within about 50 km of a refraction line. See text for further explanation.
When compared to the lithosphere in Ethiopia, which has a similar geologic history of uplift and rifting as in Kenya, we observe a clear difference in both the lithospheric thickness and maximum velocity of the lid (Figure 4-3). The lithospheric lid under the Ethiopian Plateau is more similar to the lithospheric lid under the Kenya Rift, both in thickness and with respect to the maximum velocity of the lid (~4.3 km/s). The seismic velocity structure of the uppermost mantle under the Kenya Rift is similar to that found beneath the Main Ethiopian Rift (MER) and Afar [Fuchs et al., 1997; Prodehl et al., 1994, and references therein; Dugda et al., 2007], except that the remaining lid under the Kenya Rift is not as perturbed as that under the MER and Afar, both in terms of thickness and maximum velocity. The average velocity of the lid is higher by about 0.2 km/s for the Kenya rift.

Why are the lid thickness and velocity so different beneath the Kenya and Ethiopian Plateaus? We know that the volcanism and rifting in central and southern

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Kenya started no earlier than c. 15 Ma but commenced around 30 Ma in central Ethiopia. Is this difference of 15 million years sufficient enough for thermal conduction to create the observed variations in the velocity structure of the lithosphere? A thermal calculation assuming an initial 120 km thick lithosphere c. 15 Ma under Kenya and c. 30 Ma under Ethiopia cannot create a 0.2 km/s velocity difference and a 30 - 40 km thickness variation in lithosphere, as observed.

Although it is interesting to note that the two regions under comparison happened to be geologically similar in uplift and rift ing, there is a major geologic event that may explain why the lithosphere under these two areas is so different. The Afar Flood Basalt (AFB) volcanism, which covered most of Ethiopia, has no parallel in the geologic history of Kenya. This huge flood basalt volcanism event, which is often attributed to a mantle plume, might have eroded the lithosphere beneath Ethiopia c. 30 Ma within a very short period of time, as suggested by Dugda et al. [2007]. Thus, much of the buoyancy for the lithosphere under the Ethiopian Plateau may come from thermal alteration and erosion of the lithosphere, while much of the buoyant support for the uplift in Kenya may come from sublithospheric processes [Ebinger et al., 1989]. The average elastic plate thickness for the East African Plateau away from the rift was found by Ebinger et al. [1989] to be 64 km, while the elastic plate thickness beneath the Ethiopian Plateau was found to be only 49 km, which is consistent with our finding of thicker lithosphere beneath the Kenya Plateau and thinner lithosphere beneath the Ethiopian Plateau.

Surface heat flow measurements for the Mozambique Belt in Kenya are low and similar to heat flow measurements for the Mozambique Belt in Tanzania [Nyblade et al.,
1990], while the heat flow measurements for the Kenya Rift are high [Nyblade et al., 1990; Morgan, 1973; Williamson, 1975; Evans, 1975; Wheildon et al., 1994]. The low heat flow measurements at the surface of the Mozambique Belt lithosphere in Kenya indicate that thermal anomaly has not yet reached the surface. This observation is in agreement with what has been found in this study, because the thick lithosphere beneath the EAP in Kenya will require a long time for a thermal anomaly to reach the surface.

4.6 Summary

This chapter compares the upper mantle structure in Kenya with an average model of the Mozambique Belt upper mantle structure beneath northeastern Tanzania [Julia et al., 2005], and also with the lithospheric structure in Ethiopia. It is observed that the lithosphere beneath the EAP in Kenya has not been affected significantly by the hotspot tectonism found there relative to northeastern Tanzania. However, beneath the Kenya Rift, the lithosphere appears to be highly perturbed and thinned. Generally, the percentage reduction in shear wave velocity when crossing the Kenya Rift is similar to the P-wave velocity reductions reported in previous studies [Park and Nyblade, 2006; Prodehl et al., 1997; Achauer et al., 1994; Slack et al., 1994; Green et al., 1991].

The comparison of the lithosphere beneath Ethiopia to that under Kenya shows a clear difference in lithospheric thickness and maximum velocity of the lid, although both regions have similar geologic features of uplift, volcanism, and rifting. The lithospheric lid under the Ethiopian Plateau is more similar to the lithospheric lid under the Kenya Rift, both in terms of thickness and maximum velocity of the lid. Although the seismic
velocity structure of the uppermost mantle under the Kenya Rift is similar to that found beneath the Main Ethiopian Rift (MER) and Afar [Fuchs et al., 1997; Prodehl et al., 1994, and references therein; Dugda et al., 2007], the lid under the Kenya Rift is not as perturbed as that under MER and Afar. The lid under the Kenya Rift is thicker and the average velocity of the lid is higher by about 0.2 km/s.

The difference in lid thickness and velocity beneath the Kenya and Ethiopian Plateaus does not appear to be the result of the 15 million year difference in the occurrence of volcanism and rifting in Kenya and Ethiopia. The difference may be attributed to the major geologic event of the Afar Flood Basalt (AFB) volcanism. An important implication of this difference in lithospheric structure for the similar plateau uplifts in Kenya and Ethiopia is that much of the buoyancy for the Ethiopian Plateau may come from thermal alteration of the lithosphere, while much of the buoyant support for the uplift in Kenya may come from sublithospheric processes. Heat flow measurements and gravity studies in previous studies give results consistent with the findings of this study.
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Chapter 5

Conclusion

Crust and upper mantle structure beneath Ethiopia, Kenya, and Djibouti has been investigated in this thesis using receiver functions and surface wave dispersion measurements to advance our understanding of the impact of the hotspot tectonism found there on the lithospheric structure of the region. In the first part of the thesis, a study on the crustal structure of Djibouti has been presented. From this study I obtained new crustal parameters which are contradictory to two previously published results, including a crustal thickness of 23±1.5 km and a crustal Poisson’s ratio of 0.29 to 0.33 for station ATD in Djibouti (eastern Afar) using an H-κ stacking analysis of receiver functions and a joint inversion of receiver functions and Rayleigh wave dispersion measurements. These results are consistent with the seismic velocity structure of the crust in Djibouti obtained from seismic refraction profiles [Ruegg, 1975], but I interpreted the Moho in the profiles differently. By combining my results of crustal structure beneath station ATD with the estimates of crustal structure elsewhere in Afar, it appears that crustal thickness and composition may be fairly uniform across many parts of central and eastern Afar, with Moho depths between 20 - 25 km [Berkhemer et al., 1975; Makris and Ginzburg, 1987; Dugda et al., 2005]. The high Poisson's ratio and high Vp throughout most of the crust indicates a mafic composition.
The high crustal Poisson's ratio and high mean crustal Vp throughout much of Afar, as well as the crustal thickness, are not consistent with models invoking crustal formation through stretching of pre-existing Precambrian crust. Instead, I suggest that crust in Afar consists predominantly of new igneous rock emplaced as part of the extensional process [Mohr, 1989]. During late synrift stages in the Main Ethiopian Rift, extensional strain has been accommodated within magmatic segments through dike intrusion [Ebinger and Casey, 2001]. In addition to diking, sill formation and underplating associated with the magmatic centers have likely combined to help form new igneous crust. The formation of the new igneous (mafic) crust in Afar could have also taken place through modification of oceanic crust that was subsequently altered by plume-derived magmas from the mantle.

In the second part of the thesis, I investigated the seismic velocity structure of the crust and upper mantle beneath Ethiopia and Djibouti using a joint inversion of receiver functions and Rayleigh wave group velocities. Crustal structure obtained from the joint inversion is similar to crustal structure reported in previous studies, with crustal thicknesses of 35 to 44 km beneath the Ethiopian Plateau, and 25 to 35 km beneath the Main Ethiopian Rift and the Afar [Makris and Ginzburg, 1987; Dugda et al., 2005; Stuart et al., 2006; Mackenzie et al., 2005; Dugda and Nyblade, 2006; Maguire et al., 2006]. The lithospheric mantle beneath the Ethiopian Plateau has a maximum shear wave velocity of about 4.3 km/s and extends to a depth of ~70-80 km. Beneath the MER and Afar, the lithospheric mantle has a maximum shear wave velocity of 4.1-4.2 km/s and extends to a depth of at most 50 km.
Prior to the Cenozoic hotspot tectonism, Ethiopia and Djibouti were underlain with unperturbed Mozambique Belt lithosphere. In comparison to Mozambique Belt lithosphere in Tanzania along the edge of the East African Plateau where the lid extends to depths of ~100-125 km and has a maximum shear velocity of 4.6 km/s [Julia et al., 2005], the mantle lithosphere under the Ethiopian Plateau appears to have been thinned by ~30-50 km and the maximum shear wave velocity reduced by ~0.3 km/s. The finding of thin lithosphere beneath the Ethiopian Plateau and thinner or nonexistent lithospheric mantle beneath the Main Ethiopian Rift and Afar is a new result for the region. The results of a 1D conductive thermal model suggest that the shear velocity structure of the Ethiopian lithosphere can be explained by a mantle plume, if the plume rapidly thinned the lithosphere by at least 30–50 km at the time of the Afar flood basalt volcanism (c. 30 Ma), and if the warm plume material has remained beneath the lithosphere since then. About 45-65% of the 1-1.5 km of plateau uplift in Ethiopia can be attributed to the thermally perturbed lithospheric mantle structure.

In order to test further the findings of this study, it would be useful to obtain heat flow measurements both from the rift and plateau areas of Ethiopia. Heat flow data could be used to confirm if there is a substantial difference in the surface heat flow from the plateau in Ethiopia compared to unperturbed Mozambique Belt lithosphere (East African Plateau away from rifting and volcanism) in Tanzania.

In the last part of the thesis, I examined the crust and upper mantle structure beneath Kenya and compared the upper mantle structure beneath the EAP in Kenya to an average model of Mozambique Belt upper mantle structure beneath northeastern
Tanzania [Julia et al., 2005]. A comparison of the lithosphere beneath Kenya to the lithosphere beneath Ethiopia has also been made. It is observed that the lithosphere beneath the EAP in Kenya has not been affected significantly by the hotspot tectonism found there relative to northeastern Tanzania, but only beneath the Kenya Rift, which is actually highly perturbed and thinned. Generally, the percentage reduction in shear wave velocity when crossing the Kenya Rift is similar to the P-wave velocity reduction from previous studies at different depths [Park and Nyblade, 2006; Prodehl et al., 1997; Achauer et al., 1994; Slack et al., 1994; Green et al., 1991].

A comparison of the lithosphere beneath Ethiopia to that under Kenya shows a clear difference in lithospheric thickness and maximum velocity of the lid, although both regions have similar geologic features of uplift, volcanism, and rifting. The lithospheric lid under the Ethiopian Plateau is more similar to the lithospheric lid under the Kenya Rift, both in thickness and maximum shear wave velocity. Even though the seismic velocity structure of the uppermost mantle under the Kenya Rift is similar to that found beneath the Main Ethiopian Rift (MER) and Afar [Fuchs et al., 1997; Prodehl et al., 1994, and references therein; Dugda et al., 2007], the lid under the Kenya Rift is not as perturbed as that under MER and Afar. The lid under the Kenya Rift is thicker and the average velocity of the lid is higher by about 0.2 km/s.

The difference in lid thickness and velocity beneath the Kenya and Ethiopian Plateaus does not seem to result from the 15 million year difference in the timing of volcanism and rifting in southern and central Kenya versus Ethiopia. This difference may be attributed to the major geologic event of the Afar Flood Basalt (AFB) volcanism. An
important implication of this difference in lithospheric structure for the similar plateau uplifts in Kenya and Ethiopia is that much of the buoyancy for the lithosphere under the Ethiopian Plateau likely comes from thermal alteration of the lithosphere, whereas much of the buoyant support for the uplift in Kenya probably comes from sublithospheric processes [Ebinger et al., 1989]. Surface heat flow for the Mozambique Belt in Kenya and Tanzania is not elevated [Nyblade et al., 1990], indicating that the thermal anomaly at depth in the upper mantle has not yet reached the surface. This is in agreement with the thick lithosphere beneath the EAP in Kenya found in this study, which requires a long time for a thermal anomaly to reach the surface. The finding that the lithosphere beneath Kenya is unperturbed is also new for the region.

In order to better understand the lithosphere in northern Kenya compared to central and southern Kenya and Ethiopia, it would be advantageous to conduct a broadband seismic experiment covering the area around Lake Turkana to determine detailed lithospheric structure. Such an experiment would be important to determine whether or not the lithosphere under the Turkana region, which was disturbed by volcanism earlier than the Afar Flood Basalt volcanism by up to ~10 M.y., has any similarity to the lithosphere under Kenya rift.
References


Appendix A: Preliminary Results of Joint Inversion of Receiver Functions and Surface Wave Velocities for the Arabian Peninsula

The shear-wave velocity structure beneath the Arabian Shield has been investigated using a joint inversion of receiver functions and fundamental-mode Rayleigh wave group and phase velocities [Julia et al., 2000] to image the upper mantle structure down to ~200 km depth. This particular study may enable us to answer the question to what extent the lithosphere under the Arabian Shield has been modified by the hotspot tectonism found there. Three different data sources have been used for this study: 10 - 50 sec Rayleigh wave group velocities are extracted from the tomographic studies of Pasyanos [2005]; receiver functions are calculated from seismic data collected by the Saudi Arabia National Digital Seismic Network (SANDSN) [Al-Amri and Al-Amri, 1999]; 60-140 sec Rayleigh wave phase velocities are extracted from the tomographic study of Park [2007], who used different data sets including data from SANDSN. Preliminary results for the Arabian Shield from the joint inversion approach show crustal structure consistent with previous studies [e.g., Al-Damegh et al., 2005]. Figure A1 shows the SANDSN stations analyzed in this appendix and Figure A2 gives the preliminary results of the joint inversion.
Figure A1. Location map of the Broad Band (BB) seismic stations from Saudi Arabia National Digital Seismic Network (SANDSN) that have been analyzed in this study. Triangles represent the BB stations analyzed and thin lines show country boundaries.
Figure A2. Results of the joint inversion for SANDSN network stations. For each station, three panels are shown: receiver functions (top), Rayleigh wave dispersion curves (middle), and shear wave velocity models (bottom). PREM model [Dziewonski and Anderson, 1981] is shown in the bottom panel for comparison purpose. The percentage value given with station names at the top of each inversion result is for how much percent of PREM value is subtracted from PREM below 190 km depth to fit the long period phase velocities.
Figure A2. Continued.
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References


Appendix B

The Linearized Joint Inversion Scheme of Julia et al. [2003]

Because surface wave group velocities and receiver functions show sensitivity to the shear wave velocity of the subsurface structure from which they are obtained, Julia et al. [2000] and Julia et al. [2003] developed a joint inversion technique that makes a simultaneous inversion of the two data sets. In this appendix, the joint inversion of receiver functions and dispersion measurements approach developed by Julia et al. [2003] and applied in this thesis to determine the shear wave velocity structure of my region of study is presented.

To develop the scheme of joint inversion of receiver functions and surface wave measurements, Julia et al. [2000] defined a joint prediction error for surface wave dispersion measurements and receiver functions as follows:

\[
\varepsilon_{jr} = \frac{p}{N_s} \sum_{i=1}^{N_s} \left( \frac{r_i - \sum_{j=1}^{L} D_{s,i} m_j}{\sigma_{s,i}} \right) + \frac{1}{N_r} \sum_{i=1}^{N_r} \left( \frac{r_i - \sum_{j=1}^{L} D_{r,i} m_j}{\sigma_{r,i}} \right) \quad (1)
\]

where differences in physical units and number of data points for the two different data sets has been taken care of by dividing each data set using the number of data points \(N_s\) and \(N_r\) and standard deviations \(\sigma_s\) and \(\sigma_r\) of surface wave dispersion measurements and receiver functions, respectively, \(m\) is the shear wave velocity model with certain number of layers and fixed thicknesses, \(D_s\) and \(D_r\) are partial derivative matrices for the
dispersion curve and receiver functions with the indices $ij$ indicating the ith row and jth column element of each matrix, and $r_s$ and $r_r$ are the residuals for the dispersion data and receiver functions. The influence factor $p$ (with $q = 1 - p$) controls the relative influence of each data set so that $p=0$ inverts receiver functions alone while $p=1$ inverts dispersion measurements alone.

The joint inversion scheme has been implemented using parameters from the joint prediction error by the following system of equations [Julia et al., 2003]:

$$
\begin{pmatrix}
\sqrt{\frac{p}{w_s^2}} D_s \\
\sqrt{\frac{q}{w_s^2}} D_r \\
\eta \Delta \\
W
\end{pmatrix} \rightarrow m = \begin{pmatrix}
\sqrt{\frac{p}{w_s^2}} r_s \\
\sqrt{\frac{q}{w_s^2}} r_r \\
\eta \Delta \\
W
\end{pmatrix} \rightarrow m_0 + \begin{pmatrix}
0 \\
0 \\
0 \\
W
\end{pmatrix} \rightarrow m_a
$$

(2)

where $\rightarrow r_s$ and $\rightarrow r_r$ are residual vectors of surface wave dispersion and receiver functions, respectively, $w_s^2$ and $w_r^2$ are equalizing weights obtained from $N_s \sigma_s^2$ and $N_r \sigma_r^2$ for dispersion measurements and receiver functions, respectively, as in equation (1), $\rightarrow m$ is a vector of shear wave velocity layers, $\rightarrow m_0$ is a vector of initial shear velocities, $\rightarrow m_a$ is a vector of a priori shear velocities and $W$ holds constraint weights for those a priori velocities. Matrix $\Delta$ constructs a second difference model to make the variation of velocity profiles smooth.
References:


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