A SEISMIC INVESTIGATION OF CRUST AND UPPER MANTLE STRUCTURE
BENEATH THE ZAGROS MOUNTAINS AND THE SOUTHERN AND EAST
AFRICAN PLATEAUS

A Dissertation in
Geosciences

by
Aubreya Nicole Adams

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The dissertation of Aubreya Nicole Adams was reviewed and approved* by the following:

Andrew A. Nyblade
Professor of Geosciences
Dissertation Advisor
Chair of Committee

Charles J. Ammon
Professor of Geoscience

Sridhar Anandakrishnan
Professor of Geosciences

Derek Elsworth
Professor of Energy and Geo-Environmental Engineering

Katherine H. Freeman
Professor of Geosciences
Head of the Department of Graduate Program in Geosciences

*Signatures are on file in the Graduate School
ABSTRACT

This thesis addresses crustal and upper mantle structure in three regions in and around the African Superswell: the Zagros Mountains, southern Africa, and the East African Plateau, each representing a different tectonic regime. In the Zagros Mountains of southwestern Iran, source mechanisms for six moderate-sized earthquakes are investigated using a combination of moment tensor inversion and depth phase analysis. The six earthquakes that are studied were reported in global earthquake catalogs as having lower crustal or upper mantle source depths, but upon further study, it was found that all six nucleated within the upper crust. This finding contributes to a growing number of studies indicating that seismicity in the Zagros Mountains is limited to the upper crust.

In both southern and eastern Africa, upper mantle structure was investigated to evaluate the thermal state of the upper mantle and implications for the source of uplift in each area. Rayleigh wave phase velocities were measured for these regions using a two plane wave approximation method and were then inverted for a quasi-three dimensional shear wave velocity model. In southern Africa, it was found that the lithospheric lid structure and the sublithospheric velocity reduction for the Kaapvaal Craton is comparable to the upper mantle structure beneath other Archean Cratons. Thus, little seismic evidence was found of an upper mantle thermal anomaly sufficient to support high elevations in that area. In East Africa, evidence was found for a broad thermal anomaly across the study region that extends from the base of the lithosphere into the
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Chapter 1

Introduction

In this thesis, I investigate crustal and upper mantle structure in three separate tectonic regimes within the Afro-Arabian region to advance our understanding of processes that lead to rifting, uplift, and seismicity within and surrounding the African Superswell (Figure 1-1). The depth extent of seismicity and implications for lithospheric rheology in the convergent margin of the Zagros Mountains are presented in Chapter 2. In Chapter 3, upper mantle shear wave velocity structure beneath the Kaapvaal Craton and surrounding mobile belts in southern Africa is examined to determine if there is seismic evidence for an upper mantle thermal anomaly supporting the Southern African Plateau, and in Chapter 4, upper mantle structure beneath East Africa is imaged to evaluate candidate geodynamic models for the uplift of the East African Plateau.
In Chapter 2, I use a combination of moment tensor inversion and depth phase modeling to determine source parameters of six moderate-sized earthquakes in the Zagros
Mountains. These events were reported as having lower crustal or upper mantle source depths by global earthquake catalogs. Previous studies, however, have debated whether the lower crust and upper mantle beneath the Zagros Mountains might be aseismic, deforming ductily. I find that upon detailed study, all six events appear to have nucleated in the upper 11 km of the crust, lending support to the assertion that seismicity in the area is limited to the upper crust, and suggesting that global catalogs may overestimate source depths in this region.

Chapter 3 is an investigation of the upper mantle beneath southern Africa, focusing on the upper mantle beneath the Kaapvaal Craton. Some studies have suggested that the high elevations found throughout southern Africa may be supported by buoyancy due to a thermal anomaly in the upper mantle that appears seismically as a low velocity zone (e.g., Priestley, 1999; Li and Burke, 2006; Priestley et al., 2006; Wang et al., 2008). Other studies, however, have not found evidence for a low velocity zone in the upper mantle indicative of a thermal anomaly (e.g., Zhao et al., 1999; Freybourger et al., 2001; James et al. 2001; Saltzer, 2002; Fouch et al., 2004; Larson et al., 2006; Chevrot and Zhao, 2007; Hansen et al., 2009). With a combined dataset from the Southern African Seismic Experiment, AfricaArray and Global Seismic Network stations, I build a quasi-three dimensional shear wave model by inverting phase velocities measured using a two plane wave approximation method (Forsyth and Li, 2005). I compare the shear wave velocity structure beneath the Kaapvaal Craton to velocity structure beneath other Archean Cratons, and find that the velocity reduction beneath the Kaapvaal Craton lithosphere is comparable to that beneath other Archean cratons, indicating that there is
little seismic evidence for an upper mantle thermal anomaly beneath the Kaapvaal Craton supporting the Southern African Plateau.

In Chapter 4, I apply the same methodology from Chapter 3 to eastern Africa to address the origin of the Cenozoic rifting and uplift of the East African Plateau by imaging the lateral and depth extent of low velocities beneath the plateau. In this area, the East African Rift System bifurcates around the Tanzania Craton. The entire region is characterized by anomalously high topography, which is likely supported by buoyancy from a thermal anomaly in the upper mantle. Evidence for the upper mantle thermal anomaly comes from seismic, petrological, and gravity studies (e.g., Ritsema et al., 1998; Simiyu and Keller, 1997; Chesley et al., 1999; Owens et al., 2000; Nyblade et al., 2000; Nyblade and Brazier, 2002; Weeraratne et al., 2003, Park et al., 2006). However, the depth extent, lateral extent, and source of the thermal anomaly is uncertain. I use data from the Tanzania Broadband Seismic Experiment together with new data from the AfricaArray East African Seismic Experiment to build a quasi-three dimensional shear wave model from our measured phase velocities. Because of the inclusion of new data, I am able to image structure across the entire southern East African Plateau, which has not been possible in previous studies. I find evidence of a broad low velocity zone underlying the entire region, which extends to and possibly into the mantle transition zone. This indicates that the low velocity anomaly beneath eastern Africa is broader than has been proposed in previous studies.

Chapters 2, 3, and 4 are each written in the format of independent journal articles. For this reason, there is some repetition of content in the description of methodologies and broad scale African geology in chapters 3 and 4. Chapter 2 was published in the
Bulletin of the Seismological Society of America (Adams, A., R. Brazier, A. Nyblade, A. Rodgers, and A. Al-Amri, 2009, Source Parameters for Moderate Earthquakes in the Zagros Mountains with Implications for the Depth Extent of Seismicity, Bulletin of the Seismological Society of America, 99, 2044-2049). Included in this thesis are five appendices, designed to provide additional information on each chapter that is not suited to the journal article format. Appendix A contains supplemental figures for Chapter 2, which were included as online supplements to Adams et al. (2009). Appendix B includes a list of events that were used for the surface wave tomography in southern Africa presented in Chapter 3. Appendix C includes additional figures and resolution tests for southern Africa that are not shown in Chapter 3. Appendix D is a list of the earthquakes used for the surface wave tomography in eastern Africa (Chapter 4), and Appendix E contains additional figures not included in that chapter.

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Chapter 2

Source Parameters for Moderate Earthquakes in the Zagros Mountains with Implications for the Depth Extent of Seismicity

2.1 Abstract

Six earthquakes within the Zagros Mountains with magnitudes between 4.9 and 5.7 have been studied to determine their source parameters. These events were selected for study because they were reported in open catalogs to have lower crustal or upper mantle source depths and because they occurred within an area of the Zagros Mountains where crustal velocity structure has been constrained by previous studies. Moment tensor inversion of regional broadband waveforms have been combined with forward modeling of depth phases on short period teleseismic waveforms to constrain source depths and moment tensors. Our results show that all six events nucleated within the upper crust (<11 km depth) and have thrust mechanisms. This finding supports other studies that call into question the existence of lower crustal or mantle events beneath the Zagros Mountains.

2.2 Introduction

The depth distribution of earthquakes in convergent plate boundaries and the implications it has for rheologic strength distribution in the lithosphere has been highly debated for many years (Baker et al., 1993; Bird et al., 1975; Jackson, 2002; Maggi et al.,
Much of the debate has centered on the seismically active Zagros Mountains, where plate subduction is believed to have ceased c. 5 Ma (Berberian and King, 1981). Early seismic studies of the region using the event location information in the ISC and USGS catalogs reported earthquakes in the upper crust and upper mantle, but not in the lower crust (Bird et al., 1975; Nowroozi, 1971). More recent studies, however, have reexamined earthquake depths and suggest that earthquakes occur only within the upper crust (Baker et al., 1993; Jackson, 2002; Maggi et al., 2000; Tatar et al., 2004). For example, Maggi et al. (2000) modeled teleseismic P and SH waveforms to determine earthquake source depths in several regions, including the Zagros Mountains, for events reported to have nucleated within the lower crust or upper mantle. All 13 events investigated in that study for the Zagros Mountains were found to have nucleated within the upper 20 km of the crust.

For several decades, the preferred model of lithospheric strength and earthquake depth distribution has been the so-called “Jelly-Sandwich Model” (Jackson, 2002). In this model, the lithosphere consists of a strong upper crust, a weak, ductile lower crust, and a strong upper mantle. This three-layered lithospheric model was based on the assumption that rock strength is primarily a function of composition (Brace and Byerlee, 1970; Brace and Kohlstedt, 1980; Chen and Molnar, 1983) and the thermal structure of the lithosphere (Afonso and Ranalli, 2004; Brace and Kohlstedt, 1980). Other studies, however, have argued that the depth of the brittle-ductile transition may also depend on fluid content (Hirth and Kohlstedt, 1996; Mackwell et al., 1998). Lithospheric models that account for these factors show considerable variability in the depth distribution of lithospheric strength (Brace and Kohlstedt, 1980; Hirth and Kohlstedt, 1996; Jackson,
In this paper, we contribute to the debate about the depth extent of continental seismicity and the strength of the lithosphere by studying six moderate earthquakes that occurred between 1997 and 2003 in the central Zagros Mountains for which lower crustal or upper mantle focal depths have been reported in a number of catalogs (e.g., CMT, NEIC, ISC) (Tables 2-1, 2-2). We focus on these events in the 1997-2003 time interval because they occurred where crustal structure is best constrained within the Zagros Mountains and because broadband seismic data at regional distances were provided from the Saudi Arabian National Digital Seismic Network (SANDSN). We have combined these data with other data from open stations to constrain source depth and focal mechanism for each event by inverting for moment tensors and performing a grid search over source depth. Source depths have been further constrained by forward modeling teleseismic depth phases using short period data from GSN and NORSAR stations.

Table 2-1 Depths and Magnitudes from Public Catalogs

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<th>CMT Depth(km)</th>
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</tr>
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<td>4</td>
<td>4/13/01</td>
<td>1:04:27</td>
<td>20</td>
<td>29</td>
<td>4.9</td>
<td>26</td>
<td>5.1</td>
</tr>
<tr>
<td>5</td>
<td>2/17/02</td>
<td>13:03:53</td>
<td>15</td>
<td>f15</td>
<td>5.5</td>
<td>f33</td>
<td>5.3</td>
</tr>
<tr>
<td>6</td>
<td>7/10/03</td>
<td>17:06:38</td>
<td>11</td>
<td>19</td>
<td>5.8</td>
<td>f15</td>
<td>5.7</td>
</tr>
</tbody>
</table>

EHB is Engdahl, van der Hilst, and Buland (Engdahl et al., 1998, 2006).
“f” indicates a fixed depth.
Geologic Setting

The Zagros Mountains of southern Iran, Turkey, and Iraq are part of a large tectonic region that marks the convergent boundary between the Arabian and Eurasian plates following the closure of the Neo-Tethys Sea. The Zagros Mountains are primarily located along the southwestern border of Iran, where GPS measurements indicate that oblique convergence occurs at a rate of 2.2 cm per year (Regard et al., 2004). The Zagros Mountains parallel the coast of the Persian Gulf for approximately 1200 km from Turkey in the north to the Strait of Hormoz in the south and range in width from 200 to 300 km (Tatar et al., 2004) (Figure 2-1). Seismicity rates in this region are among the highest in the world for a fold and thrust belt (Talebian and Jackson, 2004; Tatar et al., 2004).

<table>
<thead>
<tr>
<th>Evt #</th>
<th>Date</th>
<th>Time</th>
<th>Latitude, Longitude</th>
<th>MTI Mw</th>
<th>Depth Phases Depths (km)</th>
<th>MTI Depths (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>11/13/98</td>
<td>13:01:10</td>
<td>27.79N, 53.64E</td>
<td>5.5</td>
<td>6-10</td>
<td>4-7</td>
</tr>
<tr>
<td>2</td>
<td>10/31/99</td>
<td>15:09:40</td>
<td>29.41N, 51.81E</td>
<td>5.4</td>
<td>5-11</td>
<td>4-5</td>
</tr>
<tr>
<td>3</td>
<td>3/1/00</td>
<td>20:06:29</td>
<td>28.39N, 52.85E</td>
<td>5</td>
<td>7-11</td>
<td>7-10</td>
</tr>
<tr>
<td>4</td>
<td>4/13/01</td>
<td>1:04:27</td>
<td>28.28N, 54.87E</td>
<td>4.9</td>
<td>5-10</td>
<td>8-11</td>
</tr>
<tr>
<td>5</td>
<td>2/17/02</td>
<td>13:03:53</td>
<td>28.09N, 51.75E</td>
<td>5.2</td>
<td>2-10</td>
<td>2-6</td>
</tr>
<tr>
<td>6</td>
<td>7/10/03</td>
<td>17:06:38</td>
<td>28.35N, 54.17E</td>
<td>5.7</td>
<td>7-11</td>
<td>4-7</td>
</tr>
</tbody>
</table>

Moment Tensor Inversion (MTI)

2.2.1 Geologic Setting
Previous geophysical studies provide constraints on crustal structure for parts of the Zagros region. In the central Ghir region of the Zagros Mountains (see box in Figure 2-1), a combined study of local P- and S-wave traveltimes and teleseismic receiver functions indicates that the crustal thickness averages 47 km (Hatzfeld et al., 2003). This estimate of crustal thickness is consistent with thicknesses of 45±2 km determined by receiver functions (Paul et al., 2006). The crust is divided into an upper, 11 km thick, sedimentary layer and a lower, 35 km thick, crystalline basement layer (Hatzfeld et al., 2003). A Bouguer gravity anomaly study by Snyder and Barazangi (1986) found that Moho depth increased smoothly from 40 km in the south beneath the Persian Gulf to 65 km just north of the Zagros Mountains (Figure 2-1).
The Zagros Mountains are bordered to the northeast by the Iranian Plateau. This plateau has an average elevation of 1500 m (Zamani and Hashemi, 2000). A low mountain range separates the Iranian Plateau into two distinct regions, with the eastern region extending into Afghanistan (Zamani and Hashemi, 2000). Surface wave
tomography indicates the presence of a low velocity zone beneath the Iranian Plateau (Maggie and Priestly, 2005). The unusual presence of a low velocity zone in a convergent margin and the existence of Neogene volcanism within the plateau (Berberian and King, 1981) indicate a warm, buoyant upper mantle beneath the Plateau (Maggie and Priestly, 2005).

To the southeast, the Zagros Mountains are bordered by the Makran. The Oman Line, or the Minab Fault, separates these two regions. Although both regions were created by the convergence along the Eurasian plate, they differ in convergence mechanisms. The Zagros Mountains have undergone continent-continent convergence with the Arabian plate for the past 5 MY without evidence for continuing subduction (Berberian and King, 1981) while subduction of the Indian Oceanic plate continues beneath the Makran (Quittmeyer and Jacob, 1979). Thus, seismicity extends to much greater depths in the Makran relative to the Zagros Mountains.

The Arabian Platform and the Arabian Shield comprise the Arabian Peninsula to the south of the Zagros Mountains, where the majority of the seismic stations used in this study are located. The Arabian Platform lies southwest of the Persian Gulf. Sedimentary thickness on the Platform increases towards the Persian Gulf, where it reaches a maximum thickness of nearly 10 km (Seber et al., 1997). Total crustal thickness in this region is modeled to be 40 km (Rodgers et al., 1999). The Arabian Shield is uplifted relative to the Platform to its north, in spite of having a thinner crust (36 km). This anomalous uplift and the presence of recent volcanism in the Shield indicate the existence of mantle upwelling in this region (Camp and Roobol, 1992).
2.3 Datasets & Methods

2.3.1 Datasets

The data used in this study comes from a collection of both open and closed seismic networks at regional and teleseismic distances. The majority of the data for determining moment tensors comes from the Saudi Arabia National Digital Seismic Network (SANDSN) (Figure 2-1). This network consists of 11 short-period and 27 broadband three-component seismometers located in the Arabian Shield and Plateau. Data for this study were provided by the SANDSN for the 7 years from 1997 to 2003. Complimentary broadband seismic data for the same time period was also used from open stations (e.g., CSS, EIL, RAYN) in the region belonging to GSN and international cooperative networks. The teleseismic waveforms used for modeling depth phases were obtained from GSN stations as well as from NORSAR (Norwegian Seismic Array).

Earthquakes were considered for inversion if they occurred during the time frame for which we had access to SANDSN data, if they were listed as having depths of at least 15 km in the ISC catalog, if they were located within the central Zagros region where crustal structure is best constrained by previous studies (Figure 2-1), and if they were well recorded by at least three regional stations.

2.3.2 Methodology

For each earthquake, moment tensor inversion was used to determine source mechanisms from the regional waveforms, filtered between 0.02 and 0.029 Hz, using the
method of Randall et al. (1995). Because the earthquakes studied here have moderate magnitudes ($4.9 < M_w < 5.7$), the source time function was assumed to be a delta function. Regional seismograms were selected based upon visual inspection of quality throughout the wavetrain. Moment tensor inversion was performed for each event over a depth range of 0 to 80 km in 1 to 5 km increments. RMS error and visual inspection of the fit of synthetic seismograms to the data were considered to determine the best source depth for each earthquake.

A single velocity model was used for the calculation of the Green’s functions (Table 2-3). The velocity model uses the crustal structure for the Arabian Platform from Rodgers et al. (1999) and the IASP91 mantle model (Kennett and Engdahl, 1991). This model was chosen because the largest portion of the event to station travel paths lie within the Arabian Platform.

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>P-wave Velocity (km/s)</th>
<th>S-wave Velocity (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-4</td>
<td>4.00</td>
<td>2.31</td>
</tr>
<tr>
<td>4-20</td>
<td>6.22</td>
<td>3.59</td>
</tr>
<tr>
<td>20-38</td>
<td>6.44</td>
<td>3.72</td>
</tr>
<tr>
<td>38-42</td>
<td>7.30</td>
<td>4.21</td>
</tr>
<tr>
<td>42-74.5</td>
<td>8.04</td>
<td>4.48</td>
</tr>
</tbody>
</table>

To confirm and refine source depths, arrival times for teleseismic pP and sP phases recorded between distances of $30^\circ$ and $90^\circ$ were modeled using ray theory. The goal of modeling depth phases was only to constrain the source depth, so no source time function or instrument response was included; efforts at modeling focused on matching arrival times of the depth phases. A second velocity model representative of structure in
the central Zagros was used in this modeling (Table 2-4). The model consists of crustal structure from Hatzfeld et al. (2003) over a half-space mantle.

Table 2-4 Velocity Model Used for Modeling Teleseismic Depth Phases

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>P-wave Velocity (km/s)</th>
<th>S-wave Velocity (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-11</td>
<td>4.70</td>
<td>2.71</td>
</tr>
<tr>
<td>11-19</td>
<td>5.85</td>
<td>3.38</td>
</tr>
<tr>
<td>19-46</td>
<td>6.50</td>
<td>3.75</td>
</tr>
<tr>
<td>46-</td>
<td>8.00</td>
<td>4.62</td>
</tr>
</tbody>
</table>

To find clear teleseismic depth phases for each event, data from a wide range of GSN stations were examined after filtering between 0.5 and 3 Hz, in addition to stacked waveforms from NORSAR. Because of the moderate size of the earthquakes in this study, only a small number (1-3) of short-period waveforms were found for each event that showed clear depth phases.

A simple grid search was used to find the source depth that best matched the timing of the observed depth phases. In this grid search, the best-fitting strike, dip, and rake obtained from the moment tensor inversion for a given depth were used to generate halfspace synthetics of the pP and sP arrivals. The best-fitting source depth range was determined by the match of either the pP or the sP arrival time, or both to the observed waveforms.

2.4 Discussion and Conclusions

Table 2-2 gives the depths and moments for the six events examined and Table 2-5 provides the moment tensor elements for the events. Figure 2-2 shows the results of the
moment tensor inversion and depth phase matching for one sample event. Results for additional events can be found in Appendix A. For all events, focal mechanisms indicate a thrust-faulting source, with some degree of strike-slip motion, as has been observed previously for the central Zagros by many studies. No systematic change in quality of fit with increasing distance is found. Depths from moment tensor inversion are less well constrained that those from depth phase modeling. We attribute this difference to the inherent difficulty of constraining depths from surface waves and therefore prefer the depths calculated by depth phase modeling, which is a method designed to well constrain source depths. In all cases, where RMS error from moment tensor inversion has a clear minimum, the range of optimal depths determined by moment tensor inversion and from depth phase modeling show broad agreement, allowing source depth to be constrained to within a few kilometers.

Table 2-5  Moment Tensor Elements (dyne-cm)

<table>
<thead>
<tr>
<th>Evt #</th>
<th>Mxx</th>
<th>Mxy</th>
<th>Mxz</th>
<th>Myy</th>
<th>Myz</th>
<th>Mzz</th>
<th>Mo</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.847e24</td>
<td>0.205e24</td>
<td>-0.152e25</td>
<td>0.634e23</td>
<td>0.753e24</td>
<td>-0.911e24</td>
<td>1.93e24</td>
</tr>
<tr>
<td>2</td>
<td>0.237e24</td>
<td>0.114e24</td>
<td>-0.569e24</td>
<td>0.258e24</td>
<td>-0.105e25</td>
<td>-0.495e24</td>
<td>1.34e24</td>
</tr>
<tr>
<td>3</td>
<td>0.181e24</td>
<td>-0.397e23</td>
<td>-0.202e23</td>
<td>-0.167e23</td>
<td>0.278e24</td>
<td>-0.164e24</td>
<td>3.87e23</td>
</tr>
<tr>
<td>4</td>
<td>0.179e24</td>
<td>0.432e23</td>
<td>0.118e24</td>
<td>0.234e23</td>
<td>-0.300e23</td>
<td>-0.202e24</td>
<td>2.31e23</td>
</tr>
<tr>
<td>5</td>
<td>0.380e24</td>
<td>0.199e24</td>
<td>-0.393e24</td>
<td>0.550e23</td>
<td>0.561e24</td>
<td>-0.435e24</td>
<td>8.23e23</td>
</tr>
<tr>
<td>6</td>
<td>0.271e25</td>
<td>-0.212e24</td>
<td>0.225e25</td>
<td>-0.607e24</td>
<td>0.328e25</td>
<td>-0.210e25</td>
<td>4.69e24</td>
</tr>
</tbody>
</table>
The six events have source depths between 2 km and 11 km ±2 km. Based on a priori knowledge of the crustal structure of the central Zagros Mountains (Hatzfeld et al.,

Figure 2-2  A) Fit of full waveform synthetics to observed data from moment tensor inversion. Observations are shown as a solid line, while synthetics are shown as a dashed line. The bar beneath each set of waveforms indicates a scale of 100 seconds. B) The best fitting focal mechanism at each depth is shown plotted against RMS error. C) Observed depth phases are shown bracketed by synthetics for the maximum and minimum possible source depths. The bar beneath the seismograms indicates a time scale of 5 seconds. Records from NORSAR arrays are stacked to improve signal-to-noise ratios, but where GSN stations are used, records from only a single station are shown. D) Our best double couple solution and source depth are shown along with the double couple CMT solution and source depth.
2003; Snyder and Barazangi, 1986), all the events nucleated within the sedimentary upper crust, a finding that agrees with recent studies of seismic deformation in the area (e.g., Lohman and Simons, 2005; Nissen et al., 2007). These results indicate that deformation in the upper crust is not restricted to ductile folding as suggested by some studies (e.g., Hatzfeld et al., 2003; Tatar et al., 2004).

Tables 2-1 and 2-2 can be used to compare source depths from this study in relation to depths reported for the same events in other catalogs. For all six events, the depths reported in the catalogs were deeper than those obtained in our analysis. Consequently, our results are consistent with other seismological studies of the region which have found that seismicity is limited to the upper crust within the Zagros Mountains (Maggi et al., 2000). In contrast with the discrepancy of earthquake depths, we find no systematic increase or decrease between the moment magnitudes we obtained and those in other catalogs, and in all cases our magnitudes differ from those published in the CMT catalog by .2 magnitude units or less (Tables 2-1, 2-2).

The moment tensor solutions calculated for these six events include some degree of non-double couple motion, which is proportionally similar to the non-double couple components listed in the CMT catalog. Part D of Figure 2-2 shows the double couple component of our best solution and for the CMT solution. Although the solutions are broadly similar, there are some differences in nodal plane orientations.

Comparison of the source depths calculated via moment tensor inversion of the regional surface waves and from modeling of teleseismic pP and sP depth phase arrival times shows good agreement between the two methods. The smaller of the events presented here approach the minimum magnitude threshold for which teleseismic
waveforms can be used for moment tensor inversion or for depth phase modeling. The good agreement between depths obtained from moment tensor inversion and depth phase modeling achieved for these six moderate events indicates that our method of modeling regional waveforms bandpass filtered between .02 and .029 Hz to accurately estimate source depths may be applied to other moderate events in this area for which teleseismic depth phases are not available for constraining source depth.

We acknowledge that the focus on matching surface wave amplitude and timing may limit our ability to model potentially deeper events, as these events typically excite surface waves with smaller amplitudes, which may not be well recorded with high signal-to-noise ratios at regional distances, and therefore cannot be modeled the methodology in this study. This and the limited number of earthquakes modeled prevent us of from eliminating the possibility that some deeper earthquakes may occur in the Zagros Mountains.

In summary, we find no evidence from this study for lower crustal or mantle events in the central Zagros Mountains, but find a systematic over-estimation of depths in global catalogs. This finding supports previous work that calls into question the existence of lower crustal and mantle earthquakes beneath the Zagros Mountains and contributes to recent papers that study small to moderate sized earthquakes, which may not be accurately represented in global catalogs (e.g., Lohman and Simons, 2005).

2.5 References


Chapter 3  
Phase Velocities and Shear Wave Velocity Structure of the Southern African Upper Mantle

3.1 Abstract

In this study, we use seismic data from the Southern African Seismic Experiment and the AfricaArray network to investigate the seismic structure of the upper mantle beneath southern Africa, and in particular beneath the Kaapvaal Craton. We employ a two plane approximation method to measure Rayleigh wave phase velocities, accounting for the finite frequency effects of surface waves, as well as for average azimuthal anisotropy. We find phase velocities for the Kaapvaal Craton and surrounding mobile belts that are comparable to those reported by previous studies, and we find little evidence for variation from east to west across the Namaqua-Natal Belt. We invert these phase velocity measurements for a quasi-three dimensional shear wave velocity model of the upper mantle. A high velocity upper mantle lid is found beneath the Kaapvaal Craton and most of southern Africa. For the Kaapvaal Craton, the thickness of the lid (~150 km) is consistent with the lid thicknesses reported in many previous studies, and also with the lid thickness reported for other Archean cratons. The lid of the Kaapvaal Craton is underlain by a ~100 km thick low velocity zone with a 3.9 percent maximum velocity reduction. By comparing these results to those reported for other Archean cratons, we find few differences, and therefore we infer that high elevations in southern Africa must be due to buoyancy from a lower mantle source or from a thermal anomaly within the lithosphere that is small enough that it cannot be resolved seismically.
3.2 Introduction

A first-order topographic feature of the African continent is the ~1km high elevations across the southern and eastern African plateaus. This feature, termed the African Superswell (Nyblade and Robinson, 1994), also extends into the southeastern Atlantic Ocean basin where bathymetry is shallower than expected. In eastern Africa, anomalous topography is supported in large part by buoyancy from thermally perturbed upper mantle associated with the East African Rift System (Nyblade 2002). The upper mantle thermal anomaly there is manifest as a low velocity zone (LVZ) with a large decrease in velocity (i.e., ~10-12% in the S wave velocity at depths of 150-350 km) (Ritsema et al., 1998; Weeraratne et al., 2003; Chapter 4). In southern Africa, however, it is unclear whether support for the anomalous topography is caused by a thermal perturbation in the lithospheric mantle (Nyblade and Sleep, 2002), the sublithospheric upper mantle (e.g., Deen et al., 2006; Li and Burke, 2006; Burke and Gunnell, 2008), or the lower mantle (e.g., Lithgow-Bertelloni and Silver, 1998; Gurnis et al., 2000; Conrad and Gurnis, 2003; Simmons et al., 2007).

Previous studies have employed a range of seismic imaging methods to characterize upper mantle structure beneath southern Africa and determine if seismic velocities indicate the presence of a thermal anomaly sufficiently large to support the high plateau topography observed there. Many studies find only a small decrease in sublithospheric velocities, providing little support for the suggestion that the topography of southern Africa results from an upper mantle thermal anomaly (e.g., Zhao et al., 1999; Freybourger et al., 2001; James et al. 2001; Saltzer, 2002; Fouch et al., 2004; Larson et al., 2006; Chevrot and Zhao, 2007; Hansen et al., 2009). Other studies, however, suggest
that southern Africa is characterized by a sublithospheric mantle with a pronounced LVZ (e.g., Priestley, 1999; Li and Burke, 2006; Priestley et al., 2006; Wang et al., 2008), indicative of a substantial thermal anomaly in the upper mantle.

In this study, we investigate further the seismic velocity structure of the upper mantle under southern Africa and reevaluate the evidence for a thermal anomaly. We focus our investigation on the Kaapvaal Craton, where data coverage is best, but also present results for some surrounding geologic areas, including the Namaqua-Natal Belt and the Kheis Province (Figure 3-1). We use a two plane wave approximation method (Forsyth and Li, 2005) to measure phase velocities for periods between 17 and 167 seconds, which we invert for shear wave velocity structure following the method of Julia et al. (2000). We employ a dataset combining new AfricaArray data together with data from the Southern Africa Seismic Experiment (SASE) and Global Seismic Network (GSN) stations (Li and Burke, 2006). In addition to using an expanded dataset, our study differs from previous studies in that in our modeling we account for finite frequency effects on Rayleigh wave propagation and use new estimates of crustal structure as a priori model constraints.

3.3 Background

3.3.1 The Geology of Southern Africa

The Southern African Shield, consisting of Archean cratons and Proterozoic mobile belts, is characterized topographically by anomalously high elevations averaging more than 500m higher than the global continental average elevation (Nyblade and
Robinson, 1994). This study focuses on upper mantle structure beneath the center portion of the Shield, which is comprised of the Archean Kaapvaal Craton, the Paleoproterozoic Kheis Province, and the Neoproterozoic Namaqua-Natal Belt (Figure 3-1).

The Kaapvaal Craton, which covers an area of 1.2 million square kilometers and formed between 2.7 and 3.7 Ga (de Wit et al., 1992), can be divided into the Pietersburg, Witwatersrand, Kimberley, and Swaziland terrains (de Wit et al., 1992; Eglinton and
Armstrong, 2004). The Swaziland terrain is the oldest part of the Kaapvaal Craton and consists of the Ancient Gneiss complex (3.55 – 3.68 Ga) and the southern Barberton Greenstone Belts (3.4 – 3.55 Ga) (Figure 3-1) (Eglington and Armstrong, 2004). The basement of the younger Witwatersrand terrain to the north and west includes granitoid plutons and the northern Barberton Greenstone Belts. The Swaziland and Witwatersrand terrains are separated by the Inyoka Fault, marking the suture between the two terrains, which occurred at approximately 3.2 Ga (de Wit et al., 1992; Eglington and Armstrong, 2004).

The basement of the Kimberley terrain (Figure 3-1) includes greenstone belts, gneisses, and deformed volcanosedimentary sequences (e.g., Anhaeusser, 1999; Schmitz et al., 2004). The Kimberley terrain was accreted to the western margin of the combined Swaziland and Witwatersrand terrains approximately 2.88 - 2.93 Ga ago. The Colesberg Magnetic Lineament demarcates the location of the suture (Schmitz et al., 2004). To the north, the Pietersburg terrain was accreted between 2.8 and 3.1 Ga, and the Thabazimbi-Murchison Lineament forms the southern boundary of this terrain (Figure 3-1) (Eglington and Armstrong, 2004).

The assembly of the Kaapvaal Craton was completed by 2.75 Ga (e.g., Tinker et al., 2002; Eglington and Armstrong, 2004). Following its formation, additional crustal material was added to the craton through sedimentation and magmatism. Between 2.06 and 2.1 Ga, one of the largest layered mafic intrusion on Earth, the Bushveld Complex (Figure 3-1), was emplaced into the northern part of the craton. And across the craton, basement rocks have been covered by a series of sedimentary basins ranging in age from 1.8 – 3.1 Ga.
The Kaapvaal Craton is bordered on three sides by mobile belts. To the north, the Archean Limpopo Belt joins the Kaapvaal Craton to the Archean Zimbabwe Craton (Figure 3-1). The Limpopo Belt consist of highly metamorphosed granite-greenstone and granulite terrains, which underwent a series of orogenic events between 2.0 and 3.0 Ga during the suturing of the Kaapvaal and Zimbabwe Cratons (McCourt and Armstrong, 1998; Kramers et al., 2006).

To the south, the Kaapvaal Craton is bordered by the Namaqua-Natal Belt (Figure 3-1). The Namaqua-Natal Belt is a band of igneous and metamorphic rocks from the Namaqua or Kibaran Orogeny (1.0 – 1.2 Ga). Much of the belt has been covered by phanerozoic sediments, but surficial exposures are found in the eastern Natal sector and the western Namaqua sector (Thomas et al., 1994; Cornell et al., 2006). The Namaqua sector is separated on its northeastern boundary from the Kaapvaal Craton by the Kheis Province (Figure 3-1), a fold-thrust belt of low-grade metamorphosed supracrustal rocks, with structures thought to pre-date the Namaqua Orogeny (approx. 2.0 Ga) (Moen, 1999; Eglington and Armstrong, 2004; Cornell et al., 2006). The boundaries between the Namaqua-Natal Belt and the Kheis Province and between the Kheis Province and the Kaapvaal Craton are not well defined due to a limited number of dates and difficulties with defining boundaries based on tectonic structures alone (Moen, 1999).

### 3.3.2 Previous Studies of Southern Africa

Southern Africa has been the subject of numerous seismic studies over the past several decades. Some of the earliest seismic studies used body wave travel-times from tremors in the Witwatersrand gold mines to determine average crustal velocities and
thickness (e.g., Gane et al, 1949, 1956; Willmore et al., 1952). These studies concluded that there was little variation in crustal structure, and that the crust was underlain by high velocity mantle rock. Bloch et al. (1969) used seismograms from regional events recorded on World Wide Standard Seismograph Network and other local seismic stations to measure phase and group velocities for surface waves. From their measurements, they obtained shear wave models of the crust and upper mantle that confirmed the existence of a high velocity upper mantle. But their results suggested greater variability in crustal structure than reported in previous studies, as well as the presence of several thin and slow layers in the upper mantle (Bloch et al., 1969).

The deployment of the Southern African Seismic Experiment (SASE) stations (Carlson et al., 1996) and, more recently, AfricaArray stations, has allowed for renewed and more detailed studies of the crustal and upper mantle beneath southern Africa (e.g., Nguuri et al., 2001; Niu and James et al., 2002; Li and Burke, 2006; Nair et al., 2006; Kgaswane et al., 2009).

Several studies have used SASE data and receiver function analysis to estimate crustal thickness and Vp/Vs ratios in southern Africa (e.g., Nguuri et al., 2001; James et al., 2003; Nair et al., 2006). These studies found that crustal thickness beneath the Kaapvaal Craton varies from 35 to 45 km, and is slightly thicker (40-53 km thick) beneath the Bushveld Complex. For the Kheis Province and the Namaqua-Natal Belt, they obtained crustal thickness estimates of 40 km and 40-50 km, respectively. Nguuri et al. (2001) and Nair et al. (2006) also report estimates of Vp/Vs ratios greater than 1.78 for these two terrains, values which indicate the presence of mafic lithologies within the crust.
More recently, Kgaswane et al. (2009) conducted a study of crustal and uppermost mantle structure in southern Africa by jointly inverting receiver functions and Rayleigh wave group velocities for both SASE and AfricaArray stations. They found that the average crustal thickness within the four main terrains of the Kaapvaal Craton varied between 37 and 39 km, with some thickening (41 km) beneath the Bushveld Complex. They report a similar crustal thickness of 39 km for the Kheis Province, and within the Namaqua-Natal Belt, crustal thickness that ranges from 26 km near the Cape Fold Belt to 46 km near the Kaapvaal Craton. These results are generally consistent with those reported by previous studies using SASE data.

As mentioned in the introduction, many seismic studies have attempted to characterize the seismic structure of the lithospheric mantle (i.e., seismic lid) in southern Africa and determine whether or not there is a significant low velocity zone (LVZ) in the upper mantle beneath it. Every study has found that southern Africa is underlain by a high velocity upper mantle lid that is at least 150 km thick, but the amount of velocity reduction beneath the lid varies considerably between studies. A brief review of these studies follows in chronological order, and where the studies present results relevant to just the Kaapvaal Craton, we highlight the results for that terrain.

Priestley (1999), building on the model of Qui et al. (1996), developed a shear wave model for southern Africa by matching regional S waveforms and previously reported phase velocity measurements (Bloch and Hales, 1968; Bloch et al., 1969). The Priestley (1999) model included a high velocity lid extending to a depth of 160 km, underlain by a low velocity zone with minimum velocities between 4.35 and 4.53 km/s and a thickness of ~180 km. These finding are consistent with the more recent studies by
Preistley et al. (2006) and Priestley et al. (2008). Priestley et al. (2006) constructed an SV model for southern Africa using fundamental and higher mode surface waves, and found a lithospheric thickness of 175 km. They also used kimberlite nodules to estimate a geotherm for southern Africa, and from it estimated a lithospheric thickness of 204 km. Priestley et al. (2008) constructed an SV model for the entire African continent using fundamental and higher mode Rayleigh waves and report high velocities to a depth of 170 km beneath the Kalahari Craton, consistent with Priestley et al. (2006).

Zhao et al. (1999) used P waveforms and traveltimes from a South African mine event recorded on a network of stations in Tanzania to model P and SV velocity structure beneath southern Africa. They found that P and SV velocities in the upper 300 km smoothly increase with depth, providing little evidence for a low velocity zone beneath the lithosphere.

Freybourger et al. (2001) measured Rayleigh and Love wave phase delays for ten earthquakes recorded on SASE stations, and then inverted for S velocities in the mantle. They did not find any evidence for a significant velocity decrease within the upper 200 km of the mantle, but did find that the upper 100 km of the mantle was radially anisotropic (~4%). Saltzer (2002) also obtained similar results using phase delays from Rayleigh and Love waves measured at SASE stations.

Simon et al. (2002) used regional P wave travel times recorded on SASE stations to estimate the seismic structure of the upper 800 km of the mantle beneath southern Africa. They did not observe any significant decrease in velocities within the upper 400 km of the mantle, and they also found little evidence for thinning of the transition zone, indicating the absence of a thermal anomaly at those depths.
A P and S wave tomography study by Fouch et al. (2004) using SASE data found a high velocity upper mantle lid (~0.6% faster than reference model for P and S velocities) extending to depths between 250 and 300 km, with no evidence for a low velocity asthenosphere. They found reduced velocities in the upper mantle beneath the Bushveld Complex and suggested that the reduction in velocity may be due to alteration during the emplacement of the Bushveld Complex or during the Karoo Magmatic Event (~180 Ma). They also found reduced velocities in the upper mantle beneath the Cape Fold Belt and the Namaqua-Natal Belt.

Silver et al. (2004) analyzed azimuthal anisotropy in southern Africa obtained from S wave splitting measurements (Silver et al., 2001; Fouch et al., 2004b). The fast direction of anisotropy is primarily in an east to northeast direction, and is most well developed along the western and northern boundaries of the Kaapvaal Craton. Silver et al. (2004) suggested that most of the anisotropy is within the lithospheric mantle and that it developed during orogenic events prior to 1.8 Ga.

Using a combination of xenolith data and Rayleigh wave phase velocities measured between SASE station pairs in the Kaapvaal Craton and the Kheis Province, Larson et al. (2006) developed a shear wave model for southern Africa that does not include a significant low velocity zone in the upper mantle. Furthermore, Larson et al. (2006) argue that their measured phase velocities cannot be matched by a shear wave model with a low velocity zone such as that proposed by Priestley (1999).

Li and Burke (2006) used a two plane-wave approximation method (Forsyth and Li, 2005) to measure Rayleigh wave phase velocities for the SASE network, and then inverted them for a model of mantle SV velocities. Their models show a high velocity
upper mantle lid, with thickness varying between 80 and 180 km between the mobile belts and the cratons, underlain by a low velocity zone between the depths of 160 and 260 km, where velocities are on average four percent slower than lid velocities. In the southern Kaapvaal Craton, Li and Burke (2006) report that the minimum LVZ velocity is 5 percent slower than the maximum lid velocity. They interpreted this velocity reduction to be indicative of a thermally perturbed upper mantle that distinguishes the sublithospheric mantle under the Kaapvaal Craton from the sublithospheric mantle under Archean cratons in other parts of the world. Their models also show that at depths less than 100 km, the northern Kaapvaal Craton is slow relative to the southern Kaapvaal Craton.

Deen et al. (2006) used a combination of geochemical data from xenoliths and seismic velocities from a global velocity model (Grand, 2002) to model density and geotherms beneath Africa and areas surrounding other cratons. From these models, they draw inferences about the composition of the lithospheric mantle. They report that the composition of the lithospheric mantle beneath the Kaapvaal Craton is a depleted “Archon-type”, indicating that it has not experienced any tectonic or thermal alteration in the past 2.5 Ga.

Using phase delays from fundamental mode Rayleigh waves recorded on SASE stations, Chevrot and Zhao (2007) inverted for a three dimensional shear wave model of southern Africa that included finite frequency effects on wave propagation. Their results show little evidence for a LVZ under the high velocity lid of the Kaapvaal Craton, with fast velocities extending to a depth of at least 250 km beneath most of the Kaapvaal Craton. They also suggest that the Colesberg Magnetic Lineament coincides with a
velocity contrast between the Witwatersrand and Kimberley terrains at depths between 100 and 200 km, and that the lower velocities they image in the southeastern portion of the Kaapvaal Craton could have resulted from metasomatism and lithospheric heating during the initial formation of the Atlantic Ocean c. 180 Ma.

Pasyanos and Nyblade (2007) conducted a study of crustal and upper mantle S wave structure across Arabia and Africa using Rayleigh and Love wave group velocity measurements. However, because their group velocity measurements include periods of 100 seconds or less, they are unable to constrain sublithospheric mantle velocity structure.

Wang et al. (2008) used triplicated P and S phases from a regional event recorded on the SASE stations to model P and SH velocities in the upper mantle beneath southern Africa. They find evidence for a 150 km thick, high velocity upper mantle lid underlain by a 200 km thick low velocity zone, where SH velocities are reduced by five percent and P velocities are reduced by 2 percent.

In a recent study of S-wave receiver functions and Rayleigh wave phase velocities for SASE, AfricaArray, and GSN stations, Hansen et al. (2009) estimated that the lithosphere under the Kaapvaal Craton extends to a depth of 160 km, and that a low velocity zone of no more than 90 km thick with an approximate 4 percent decrease in velocity exists under the lithosphere. They argued that the depth and thickness of the LVZ is not necessarily anomalous when compared to the upper mantle structure beneath other Archean cratons.

Begg et al. (2009) used a combination of compositional measurements from xenoliths combined with the seismic tomography model from Grand (2002) to characterize the upper mantle beneath Africa. Their model shows a high velocity lid
extending to at least 250 km, with no evidence for reduced velocities beneath this lid when compared to the mantle beneath the rest of Africa at comparable depths. Based on xenolith geochemistry, they estimate that the subcontinental lithospheric mantle extends to a depth of approximately 200 km beneath the Kaapvaal Craton.

In addition to the growing number of seismic studies of the upper mantle beneath the Kaapvaal Craton, a number of studies have also employed xenolith composition and thermo-barometry measurements to characterize the cratonic upper mantle of the region (e.g., Rudnick and Nyblade, 1999; Eaton et al., 2009 and references therein). For example, Easton et al. (2009) study xenolith and xenocryst thermo-barometry and report that the lithospheric thickness beneath the Kimberley region of southern Africa is between 195 and 215 km thick, and do not find evidence of asthenospheric material. James (2004) used xenolith composition to estimate P and S wave velocities and densities, and showed that although there is a decrease in velocity with depth in the lithospheric mantle, the minimum velocity obtained is 4.55 km/s, which is greater than what is typically required to be considered an asthenospheric velocity.

3.4 Phase Velocity Inversion

3.4.1 Data Processing

Our study area is 1650 km by 1442 km and includes the Kaapvaal Craton and the surrounding Kheis Province and Namaqua-Natal Belt, but not the Limpopo Belt, Cape-Fold Belt, or Zimbabwe Craton (Figure 3-1). This area was selected to allow for investigation of mantle structure beneath the interior portion of the southern African
shield, where our dataset provides the best ray coverage, while simultaneously minimizing the total aperture between stations. The additional constraint of limiting the aperture between stations is necessary because of the assumption of planar waves on a spherical Earth when employing the 2 plane wave method to measure phase velocities (Donald Forsyth, personal communication).

In this study, we use data from a large number of broadband stations spread across our study area (Figure 3-2). The stations belong to the Global Seismic Network (stations LBTB, BOSA, and SUR), the AfricaArray network (www.africaarray.org), and the IRIS/PASSCAL Southern Africa Seismic Experiment (SASE) (Carlson et al., 1996).

The SASE experiment consisted of 78 broadband stations deployed from April 1997 to July 1999. The stations were installed in a linear swath extending from the Cape Fold Belt in the southwest to the Zimbabwe Craton in the northeast (Figure 3-2). Thirty-two stations remained in the same position during the entire experiment, while 23 stations were redeployed after one year. Streckeisen STS-2 seismometers were used in these stations; other details about this experiment can be found in Carlson et al. (1996) or at (www.dtm.ciw.edu/mantle/kaapvaal). For this study, we use data from 69 SASE stations, located south of 22° S latitude.
The AfricaArray permanent network consists of broadband stations throughout the continent. We use data from 9 stations in South Africa in operation since 2006. Seismometers at these stations include KS2000, Gulralp 3T, and Gulrap 40T models. The locations of the AfricaArray stations improve data coverage in the southeastern part of the southern African shield where no SASE stations were located.

For events recorded by SASE stations within our study area, we used windowed and filtered data that was processed by Li and Burke (2006), and have followed a similar processing technique for events recorded on AfricaArray and GSN stations. Events were selected to add to the dataset only if they had a high signal-to-noise ratio, and were well...
recorded on at least 5 stations. Individual records were selected for inclusion in the
inversion if the unfiltered data had no clear interference patterns or timing problems prior
to filtering. Because it is important to model both constructive and destructive
interference of the two plane waves within our study area, low signal-to-noise ratio at a
single station for a generally well-recorded event was not a sufficient reason to eliminate
the record. By including data from both SASE stations (Li and Burke, 2006) and
AfricaArray and GSN stations, we use a total of 101 events with ms > 5.8, providing
good azimuthal coverage (Figure 3-3). To illustrate the data coverage, in Figure 3-4 we
show raypaths for event station pairs for 50 second Rayleigh waves.

Each record was processed separately for 20 periods ranging from 17 to 167
seconds. At each period, records were filtered using a zero-phase-shift 10-mHz wide
bandpass filter centered around each frequency of interest. A single window length was
manually selected for each period-event combination such that it contained the Rayleigh
wave arrivals at all stations. For each period, the same window length was used for all
stations recording a particular event. This windowing process limits noise and
interference from body wave arrivals and the surface wave coda. The phase and
amplitude of the Rayleigh wave recorded in each seismogram were then calculated using
Fourier analysis.
Figure 3-3: Event locations. Gray circles indicate events recorded by the SASE array and used in Li and Burke (2006). Black circles indicate events recorded by AfricaArray stations. Large gray circles indicate distance intervals of 30 degrees from our study area.
3.4.2 Inversion Methodology

Many surface tomography studies assume that the incoming surface wave field from an earthquake can be represented by a single plane wave propagating along a great-circle path from a source to a receiver. Although this assumption is accurate for a radially symmetric Earth, it neglects the multipathing effects caused by heterogeneities a wave field may encounter along a real path. To better account for the non-great circle propagation of an incident wave, we model the wave-field for each event in this study using the method of Forsyth and Li (2005), by estimating the phase, amplitude and
azimuth of two hypothetical incoming plane waves before solving for velocity parameters. In the initial approximation of the incoming wave-field, velocity is held constant and iterative simulated downhill simplex annealing is used to approximate the wave parameters (Press et al., 1992). After solving for initial estimates of wave parameters, a linearized inversion (Tarantola and Valette, 1982) is used to solve simultaneously for corrections to the modeled wave parameters and current velocity model by minimizing errors, in a least-squares sense.

To implement this methodology, the study region was divided into a set of 667 nodes, with an interior node spacing of 0.5° and an exterior node spacing of 1° (Figure 3-5). An a priori error estimate of 0.15 km/s was assigned to each of the nodes to determine damping values, while the error estimate for the exterior nodes was assigned a value of 1.5 km/s. This differential assignment of estimated error allows any abnormalities of the incoming wave field, which are not accounted for by the two-plane wave approximation, to be preferentially mapped into the exterior nodes, outside of our region of primary interest.

For obtaining phase velocity maps for each period, a series of inversions were preformed, starting with a simple inversion for an average one-dimensional velocity dispersion curve for all of the nodes in the study area. We chose to use the average southern Africa phase velocities from Li and Burke (2006) as our starting model for the average one dimensional inversion, but tests we performed using different starting models indicated that the inversion is sensitive to the starting model only for periods of 20 seconds or less. The average one dimensional curve we obtained from this initial inversion was then used as a starting model to invert for average one-dimensional
velocity curves within the craton and mobile belts, as shown in Figure 3-5. Velocity curves for the craton and mobile belts were in turn used as starting models for an inversion for two-dimensional velocity maps. This approach minimizes the bias imposed by the starting model on the two dimensional inversions for phase velocity maps.

The one-dimensional models in our study used a Gaussian sensitivity function to represent sensitivity to structures off of the great-circle paths of the two plane waves. Many studies, however, have shown that, because the frequencies of surface waves are finite, this Gaussian approximation is only valid for structures larger than the characteristic wavelength of the period being measured (e.g., Zhou et al., 2004; Yang and Forsyth, 2006a). Zhou et al. (2004) developed a method of calculating finite frequency sensitivity kernels for both the phase and amplitudes of surface waves from source to receiver. They demonstrated that the sensitivity of surface waves to off path structure varies along the assumed great-circle travel path, as well as away from the path in both polarity and amplitude. Yang and Forsyth (2006a) applied the kernels developed by Zhou et al. (2004) to regional surface wave tomography studies using both one and two plane waves, and demonstrated that the resolution of small features can be improved by accounting for the finite frequency effects of surface wave propagation. Therefore, in our two-dimensional models, we account for the finite frequencies effects of surface wave propagation for our two plane wave tomography following the method described and applied by Yang and Forsyth (2006a; 2006b). We also solve for an average azimuthal anisotropy at each period.
3.4.3 Phase Velocity Results

Average one dimensional phase velocities for the entire study area are shown in Figure 3-6, and are compared with phase velocities reported by Li and Burke (2006). Good agreement between the dispersion curves is found. Figure 3-7 shows the one dimensional phase velocity dispersion curve, with error estimates, for each of the areas.

Figure 3-5: Inversion node locations. Small circles indicate background nodes. Stars represent nodes in the northern Kaapvaal Craton. Nodes marked by squares are in the southern Kaapvaal Craton, and triangles indicate the Kheis Province. Large circles represent nodes in the Namaqua-Natal Belt. Terrain boundaries are the same as in Figure 3-1.
within the study region. The Namaqua-Natal Belt has significantly lower velocities than the other geologic regions at all periods. The velocities of the Kheis Belt are comparable to the Kaapvaal Craton at periods of 100 seconds and less. The phase velocities of the northern and southern Kaapvaal Craton are similar, but at periods less than 100 seconds, the southern part of the craton is slightly faster, while the northern part of the craton is faster at periods greater than 100 seconds.

Figure 3-8 shows 2 dimensional velocity maps for selected periods. (Velocity maps for all periods can be found in Appendix C.) At all periods, the Kaapvaal Craton is faster than the Namaqua-Natal Belt and the Kheis Belt. Velocities are highest in the central region of the Kaapvaal Craton, and we do not observe systematic differences between the Kimberley, Witwatersrand, and Swaziland terrains across the southern Kaapvaal Craton at any period. In the northern Kaapvaal Craton, low velocities are observed centered around the surficial exposures of the Bushveld Complex for periods shorter than 50 seconds. At longer periods, however, velocities are comparable throughout the Kaapvaal Craton. When compared to the Kaapvaal Craton, velocities are lower at all periods for the Namaqua-Natal Belt, but there are no significant differences in velocity between the Namaqua and Natal Provinces.
Figure 3-6: Average one dimensional phase velocity dispersion curve for southern Africa. For comparison, results from Li and Burke (2006) are shown by a dashed line. Error bars indicate one standard deviation. The inset figure shows the average one dimensional shear wave model for southern Africa.
Figure 3-7: Average phase velocity dispersion curves for areas shown in Figure 3-5. Error bars indicate one standard deviation.
As mentioned above, the two dimensional velocity models not only include finite-frequency wave propagation, but also include a solution for an average azimuthal anisotropy at each period. Figure 3-9 shows the fast direction and percent velocity change at each period. The fast direction is predominantly east-northeast, which is consistent with findings from previous studies of shear-wave splitting (e.g., Silver et al., 2004). Variations in the fast direction and percent velocity change at different periods are small, indicating similar anisotropic characteristics at different depths.
Figure 3-9: Plot showing the fast direction for azimuthal anisotropy as a function of period and percent change in velocity.
The error bars in Figures 3-6 and 3-7 indicate a formal error estimate of one standard deviation, as calculated during the one dimensional inversions. These error estimates are probably unrealistically small, and a more reasonable estimate of error is obtained for the two dimensional inversions based upon the diagonals of the covariance matrices (Figure 3-10 and Appendix C). We choose to crop our two dimensional maps of phase velocity to show only areas with uncertainty less than 0.06 km/s, an intermediate value compared to similar studies (e.g., Weeraratne et al., 2003; Li and Burke, 2006). Figure 3-10 shows models uncertainty at 50 seconds for this study and using only the data from Li and Burke (2006) and without accounting for anisotropy or finite frequency effects, and illustrates the improved resolution achieved by the addition of these features in this study. Further comparisons are shown in Appendix C.

To further examine the resolution of the phase velocity maps, we have performed checkerboard tests. At each period, a checkerboard pattern was set-up with velocities...
equal to ±5 percent of the average velocity at that period. Resolution was tested for square checkers of a variety of sizes. Recovery of shape, location, and amplitudes of the checkers is good for the 2 degree checkers for periods up to 125 seconds and for 3 degree squares at greater periods (Figure 3-11). As expected, resolution is best where we have the highest concentration of stations in the center of our study area. There is slight smearing of the checkerboard structure to the northwest and southeast, but the general shape of the checkers is still recovered. Resolution is best for moderate periods (25-87 seconds). Results from the checkerboard resolution tests for other periods are shown in Appendix C.
3.5 Inversion for Shear Velocity Structure

Figure 3-11: Checkerboard resolution tests for selected periods. Input checkers for a-d) are two degree squares, whose locations are shown by the dotted lines in d). For e) and f), input checkers are three degree squares.
3.5.1 Method of Inversion for Shear Wave Velocity

We invert our phase velocity measurements for shear wave velocities to obtain a quasi-three-dimensional velocity structure of the upper mantle beneath southern Africa following the method described by Park et al. (2008). In this method, at each of the interior nodes (Figure 3-5), we first invert for a one dimensional shear wave profile, with crustal structure constrained by \textit{a priori} information, and then apply a lateral smoothing operator to the one dimensional models. Lateral smoothing is applied using a Gaussian weight with a characteristic length of 150 km away from each node. Appendix C shows results from tests using a range of lateral smoothing lengths.

To invert phase velocity dispersion curves for shear wave velocity profiles at each grid node, we use the inversion methodology of Julia et al. (2000). This method was designed to jointly invert dispersion and receiver function data using an iterative damped generalized linearized inversion by minimizing a weighted least-squares norm of multiple datasets. To apply the method of Julia et al. (2000) to our inversion of only phase velocity curves, we set the weight given to receiver functions equal to zero. This technique inverts observed data and model smoothness and weighting parameters for changes to a starting model. For surface waves, this is accomplished by minimizing a function from Russell (1987).

For obtaining a model of the mantle below 410 km, we inverted the one dimensional curve for southern Africa (Figure 3-6), using a starting model that consisted of the average crust and uppermost mantle structure from Kgaswane et al. (2009) as described in Table 3-1. Below 50 km, velocities from IASP91 were used, with layer thicknesses set to 6 km to a depth of 260 km, and 10 km to a depth of 810 km. This is
underlain by three 30 km thick layers, one 50 km thick layer, and a halfspace. The resulting model is shown in the inset diagram in Figure 3-6.

To account for the sensitivity of phase velocities to crustal thickness and velocities, we employ a different starting model for the upper 50 km (crust and top mantle layer) for the nodes in each geologic region, taken from Kgaswane et al., (2009) (Table 3-1). Below 50 km, velocities for each starting model were tapered to the average southern African velocity at a depth of 410km (inset in Figure 3-6). During the inversion for each node, a minimal smoothing parameter was applied to the crustal layers to prevent large and unrealistic variations in velocity between layers. Also, to constrain Moho depth during the inversion to those reported by Kgaswane et al. (2009), while simultaneously allowing for a large velocity change across the boundary, a 1km thick layer was included above and below the Moho. Velocities in the 1 km thick layers were weighted, but not fixed or smoothed.

Table 3-1: Starting Crustal Models for Shear Wave Velocity Inversion

<table>
<thead>
<tr>
<th>Geologic Terrain</th>
<th>Crustal Thickness</th>
<th>Vs &lt; 20km</th>
<th>Vs 20-30km</th>
<th>Vs &gt;30km</th>
<th>Uppermost Mantle S Velocity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average for S. Africa</td>
<td>38 km</td>
<td>3.47 km/s</td>
<td>3.89 km/s</td>
<td>4.06 km/s</td>
<td>4.57 km/s</td>
</tr>
<tr>
<td>Witwatersrand Terrain</td>
<td>37 km</td>
<td>3.53 km/s</td>
<td>3.83 km/s</td>
<td>4.00 km/s</td>
<td>4.60 km/s</td>
</tr>
<tr>
<td>Swaziland Terrain</td>
<td>39 km</td>
<td>3.41 km/s</td>
<td>3.91 km/s</td>
<td>4.10 km/s</td>
<td>4.60 km/s</td>
</tr>
<tr>
<td>Kimberley Terrain</td>
<td>37 km</td>
<td>3.62 km/s</td>
<td>3.73 km/s</td>
<td>3.90 km/s</td>
<td>4.60 km/s</td>
</tr>
<tr>
<td>Bushveld Complex</td>
<td>41 km</td>
<td>3.38 km/s</td>
<td>4.00 km/s</td>
<td>4.00 km/s</td>
<td>4.50 km/s</td>
</tr>
<tr>
<td>Kheis Province</td>
<td>42 km</td>
<td>3.66 km/s</td>
<td>3.77 km/s</td>
<td>4.10 km/s</td>
<td>4.65 km/s</td>
</tr>
<tr>
<td>Namaqua-Natal Belt</td>
<td>45 km</td>
<td>3.56 km/s</td>
<td>4.03 km/s</td>
<td>4.30 km/s</td>
<td>4.60 km/s</td>
</tr>
</tbody>
</table>
3.5.2 Shear Wave Velocity Results

Our quasi-three dimensional shear wave model is illustrated with several depth slices and cross-sections in Figure 3-12. In the upper 200 km of the model, similar velocities are found within the Witwatersrand, Swaziland, and Kimberley terrains, which comprise the southern portion of the Kaapvaal Craton, and there is little evidence for systematic variations between terrains. Beneath this part of the Kaapvaal Craton, the uppermost mantle has maximum velocities of approximately 4.65 km/s. At depths between 100 and 250 km, shear wave velocities are reduced and reach a minimum velocity of approximately 4.5 km/s at 175 km depth (i.e., a 3.9 percent reduction from the highest lid velocity).

Velocities in the northern Kaapvaal Craton are similar beneath the Bushveld Complex and the Pietersburg Terrain. In these regions, velocities do not vary significantly with depth throughout the uppermost mantle, and remain between 4.5 and 4.6 km/s to a depth of at least 150 km. Velocities reach a minimum of 4.45 km/s at a depth of 175 km, which is comparable, within our uncertainties, to velocities at that depth beneath the southern part of the Kaapvaal Craton.

Velocities in the upper 150 km of the Kheis Province are as high as 4.65 km/s. At depths greater than 150 km, velocities are reduced, reaching a minimum velocity of 4.4 km/s at 250 km depth. This represents a five percent reduction in velocity from the highest velocities in the lid.
Velocities in the uppermost mantle beneath the Namaqua-Natal Belt vary between 4.5 and 4.6 km/s, but these variations do not appear to be correlated from east to west. At depths between 125 and 300 km, velocities are reduced by up to five percent, reaching a

Figure 3-12: Shear wave velocities for selected depths (a-d) and cross-sections (e-h). Locations for the cross-sections are shown on the 300km depth slice (d).

Velocities in the uppermost mantle beneath the Namaqua-Natal Belt vary between 4.5 and 4.6 km/s, but these variations do not appear to be correlated from east to west. At depths between 125 and 300 km, velocities are reduced by up to five percent, reaching a
minimum velocity of 4.35 km/s at 200 km depth. Velocities at all depths are comparable in the eastern Natal and western Namaqua regions.

Figure 3-13 shows average one dimensional models for each geologic region, along with one standard deviation variability at each depth. For comparison, the AK135 model is also shown (Kennett et al., 1995). All regions exhibit a high velocity lid at shallow mantle depths, underlain by a reduction in velocity, but the absolute velocities and the depth at which velocity reduction occurs varies between the craton and mobile belts. At depths less than 125 km, all regions within our study area are faster than AK135, but at greater depths, all regions are comparable to or slightly slower than the global model, within our uncertainties. Phase velocities for the models from this study and Li and Burke (2006) shown in Figure 3-13 are shown in Figure 3-14.
Figure 3-13: Comparison of shear wave velocity profiles and phase velocity curves from this study (solid lines), Li and Burke (2006) (dotted lines), and AK135 (dashed line). Standard deviation for shear velocities from this study are shown with a gray swath and represent model variability. Observations for this study are shown by a solid black line, a solid gray line shows the curve predicted by our shear wave model, and the dotted line shows observations from Li and Burke (2006).
Figure 3-14: Phase velocity curves corresponding to the areas shown in Figure 3-13. Observations for this study are shown by a solid black line, a solid gray line shows the curve predicted by our shear wave model, and the dotted line shows observations from Li and Burke (2006).
3.6 Discussion

In this study, our inversion for phase velocity uses the two plane-wave method of Forsyth and Li (2005), and therefore shares similarities with the study by Li and Burke (2006). However, there are several differences in our studies. In this study, we use a smaller study area, and add additional data from the AfricaArray stations. Our inversions also include a solutions for average azimuthal anisotropy and accounts for finite frequency effects, which were not considered in the study by Li and Burke (2006). Given these differences, we first briefly summarize the main findings of our study in this section, then compare our models to comparable models published in previous studies, and end by addressing to what extent the upper mantle structure beneath southern Africa may be thermally perturbed by examining sublithospheric mantle velocities.

3.6.1 Review of Shear Wave Velocities and Models

Beneath the northern Kaapvaal Craton, velocities are similar throughout the upper 150 km of the mantle and range between 4.4 and 4.6 km/s. The velocities to 150 km depth beneath the southern Kaapvaal Craton are somewhat higher (~4.5-4.7 km/s). Below 150 km depth velocities beneath the northern and southern parts of the craton are similar, with velocities reaching a minimum of 4.45-4.5 km/s at a depth of 175 km.

The Namaqua-Natal Belt has a high velocity (4.6 km/s) lid that is slightly slower than that beneath the southern Kaapvaal Craton. The lid is underlain by a low velocity zone with a minimum velocity of 4.35 km/s. We do not find a difference in velocity structure between the Namaqua and Natal Provinces. The Kheis Province has lid
velocities similar to the southern Kaapvaal Craton, but a LVZ beneath the lid that is similar to the LVZ found under the Namaqua-Natal Belt.

3.6.2 Comparison to Previous Studies

We now compare our models for shear wave velocity structure to those reported in previous studies. Because there is evidence of radial anisotropy in southern Africa (Saltzer, 2002), we compare our results only to other published SV models. We also limit our comparisons to studies that present models only for the terrains included in our study area. As reviewed in section 3.3.2, many studies present models representative of average structure across many terrains with various ages, and we do not include these studies in our comparisons. In addition, we focus our comparisons primarily on the velocity structure of the Kaapvaal Craton because that is where our model has the best resolution.

Other published SV velocity models for the Kaapvaal Craton are shown in Figure 3-15. Relative to our shear wave model, the model by Hansen et al. (2008) is faster at depths greater than 240 km and slower at shallower depths. The Hansen et al. model fits the phase velocity measurements published by Li and Burke (2006), illustrating the trade-off between velocities at different depths when fitting dispersion curves. Nevertheless, when taking into account uncertainties in the models, the models are generally similar. The model from Larson et al. (2006) has a faster lid than our model, and shows little reduction in velocity beneath the lid. Li and Burke (2006) show a model with a lid that is faster than our lid, but which is comparable to our model at depths greater than 225 km. A study of shear wave velocity in southern Africa by Chevrot and Zhao (2007) does not
report average shear wave velocities, but by comparing our shear wave velocity maps to their velocity perturbation maps, we observe similar patterns in velocity variation at depths greater than 200 km. At depths of 100-150 km, Chevrot and Zhao (2007) show lower velocities in the southeastern portion of the craton compared to other parts of the craton, which we do not observe.

Figure 3-15: Velocity models for the Kaapvaal Craton from several studies.

The shear wave models of Li and Burke (2006) and Larson et al. (2006) were calculated from the inversion of Rayleigh wave phase velocities, and it is instructive to compare the dispersion curves to see why our models of mantle structure differ from theirs. Figure 3-16 compares our observed and predicted phase velocities for the southern Kaapvaal Craton to those reported by Li and Burke (2006) and Larson et al.
(2006). At all periods, phase velocities from Larson et al. (2006) are equal to or faster than our measured velocities. Phase velocities from Larson et al. (2006) were measured using the two-station method along raypaths that were primarily oriented in a northeast-southwest direction. This direction is consistent with the fast direction of anisotropy found in our study (Figure 3-10) and previous studies (Silver et al., 2001; 2004; Fouch et al., 2004b), which might explain why their model is faster than ours.

Overall, our phase velocities are similar to those measured by Li and Burke (2005), with Li and Burke’s (2006) being slightly faster at periods less than 100s and slightly slower at periods greater than 100s (Figure 3-16). Li and Burke (2006) also used a slightly thicker (3 km) crust in their model than the one we used. Modeling tests show that the differences in the shear wave models shown in Figure 3-15 result from the variations in the phase velocity curves (Figure 3-16) and from the different crustal models used, rather from a difference in modeling technique (Appendix C).

![Figure 3-16: Observed phase velocity curves and predicted curves for the southern Kaapvaal Craton.](image-url)
Velocities in our shear wave model for the craton are also consistent with shear wave velocities estimated from xenoliths (James et al., 2004). James et al. (2004) find a maximum velocity of 4.71 km/s in the upper mantle lid at a depth of 50 km, and a minimum velocity of 4.49 km/s at a depth of 200 km. Within our model uncertainties, the xenolith-derived velocities are similar to the maximum (4.67 km/s) and minimum (4.49 km/s) velocities in our model within comparable depth intervals.

Figure 3-13 shows a comparison between our average models and those from Li and Burke (2006) for regions where our study areas overlap. For the northern Kaapvaal Craton, Li and Burke (2006) find lid velocities that are faster than ours, but similar minimum velocities beneath the lid. For the Western Namaqua-Natal Belt, we find mantle lid velocities that are comparable to those from Li and Burke (2006), within the variability of our model. At depths between 125 km and 275 km, we find a greater reduction in velocity relative to Li and Burke’s model, but the variability in our model indicates that the differences between our studies may not be resolvable. Upper mantle lid velocities for the Kheis Province are similar between our study and Li and Burke (2006), but at depths greater than 150 km, our velocities are up to 0.15 km/s slower than those reported by Li and Burke (2006).

In the northern section of the Kaapvaal Craton, we observe that velocities in the upper 150 km are 0.05 to 0.1 km/s slower than in the southern section of the craton. Our model uncertainties may be insufficient to confidently interpret these velocity variations. However, in their P and S wave tomography images of the Kaapvaal Craton, Fouch et al. (2004) also find low velocities beneath the Bushveld Complex, as well as to the west of the Bushveld Complex, compared to velocities at similar depths beneath the southern
Kaapvaal Craton, which is consistent with our results. Fouch et al. (2004) suggest that the lower velocities in the northern Kaapvaal Craton may be due to refertilization and metasomatism during the emplacement of the Bushveld Complex ~2.1 Ga.

3.6.3 Comparison with Other Cratons

To address the question of whether or not the upper mantle beneath southern Africa is thermally perturbed, we now compare our model for the Kaapvaal Craton to SV velocity models for other Archean cratons. We also discuss results from previous studies of xenoliths in the Kaapvaal Craton and other Archean cratons that have relevance to upper mantle thermal structure. We focus on the southern Kaapvaal Craton because our resolution is best in this part of our model and also to avoid complications from possible alteration of northern Kaapvaal Craton lithosphere by the Bushveld Complex.

Figure 3-17 shows our average model for the southern Kaapvaal Craton with SV models for other Archean cratons. The models for the Slave (Chen et al., 2007) and Tanzania (Weeraratne et al. (2003) Cratons were obtained from phase velocities measured using the two plane wave approximation method employed in this study (Forsyth and Li, 2006). Simons et al. (1999) measured phase velocities for the Western Australian Craton using a Partitioned Waveform Inversion method (Nolet, 1990). Darbyshire et al. (2007) used a monte carlo method (Shapiro et al., 1997) to model phase velocity measured using a two-station method, and presented shear wave velocity profiles for many two-station paths across the Superior Province. We show the average of their profiles from the central part of the Superior Province (Figure 3-17).
Considerable variation exists between the cratons. The Slave Craton, Superior Province, and Western Australia Craton have average velocity reductions of 2.5, 4.1, and 1.9 percent, respectively, from the maximum lid velocity to the minimum velocity in the LVZ, while the average velocity reduction for the Tanzania Craton is 12.1 percent. Li and Burke (2006) find an average velocity reduction of 5.3 percent for the Kaapvaal Craton, while our results indicate a slightly lower velocity reduction of 3.9 percent. These results indicate that changes in velocity in the upper mantle beneath the Kaapvaal Craton fall within the range of velocity variations observed beneath most other cratons and therefore are not necessarily anomalous. Given this result, we find little evidence for a thermal anomaly within the upper mantle beneath the Kaapvaal Craton. The velocities beneath the Tanzania Craton, however, are much lower than beneath the other cratons and are likely indicative of thermally perturbed upper mantle.

The similarity between the upper mantle structure of the Kaapvaal Craton and other Archean cratons is also reflected in xenolith data. Rudnick and Nyblade (1999) used a large set of pressure-temperature (P-T) estimates for mantle xenoliths to characterize the average geotherm for the Kalahari Craton (including both the Kaapvaal and Zimbabwe Cratons) and to estimate crustal and mantle heat production. By comparing the Kalahari pressure-temperature profile to those from other Archean cratons, they were able to draw comparisons between the thermal structures of the cratons. They found that P-T estimates for the Superior Craton are similar to those for the Kalahari Craton, indicating a similar geotherm for the two cratons. The P-T estimates for the Siberian Craton are, on average, also similar to the Kalahari Craton, but exhibit a large amount of scatter. For the Slave Craton, however, P-T estimates indicate a cooler
geotherm than for other Archean cratons (Rudnick and Nyblade, 1999). These findings are consistent with the velocity model in Figure 3-17 showing that the upper mantle beneath the Slave Craton is somewhat faster than for the Kaapvaal Craton and the Superior Province.

Figure 3-17: Velocity models for several cratons. Models are shown for the Slave Craton (Chen et al., 2007), the central Superior Craton (Darbyshire et al., 2007), the Western Australia Craton (Simons, 1999), the Tanzania Craton (Weeraratne et al., 2003), and for the southern Kaapvaal Craton (this study).

The thickness of the lithospheric mantle lid beneath Archean cratons is difficult to determine because of the uncertainty of how to best define the base of the lithosphere seismically. Some studies estimate lithospheric thickness based on absolute velocity comparisons, while other studies use the depth of the velocity minimum in the LVZ or
else the depth of the maximum negative velocity gradient. If we choose the depth at which the maximum negative velocity gradient occurs as the base of the lithosphere, we obtain a lithospheric thickness of ~170km for the Kaapvaal Craton. Regardless of how the base of the lithosphere is chosen, Figure 3-17 illustrates that the lid thickness beneath the Kaapvaal Craton is comparable to lid thicknesses beneath other Archean cratons, and that for all cratons shown, the base of the lithosphere probably lies between 150 and 200 km depth.

3.7 Summary and Conclusions

In this study, we use seismograms from SASE, AfricaArray, and GSN stations to measure Rayleigh wave phase velocities using the two plane wave approximation method of Forsyth and Li (2005). The method that we use also solves for azimuthal anisotropy and accounts for finite frequency effects of Rayleigh wave propagation. We find phase velocities that are generally similar to previous studies, with phase velocities that are faster in the southern Kaapvaal Craton and slower for the Bushveld Complex and the surrounding mobile belts. We also do not observe evidence of systematic velocity variations between the Namaqua and Natal Sectors of the Namaqua-Natal Belt.

We have inverted the phase velocity measurements for a quasi-three dimensional shear wave velocity model using the inversion technique of Julia et al. (2000), and compared the shear wave model to models from other studies. The models are generally similar and the small differences in the models that we find result from differences in phase velocities for periods of 40 to 100 seconds and a priori constraints used for crustal structure.
We find a high velocity upper mantle lid beneath the Kaapvaal Craton and a reduction in shear wave velocities of 3.9 percent within a LVZ below the lid. When compared to the velocity structure of other cratons, the upper mantle velocity structure beneath the Kaapvaal Craton does not appear to be anomalous. This conclusion is supported by seismic velocity estimates derived from xenolith compositions, as well as P-T estimates from xenoliths.

Consequently, our results do not support the presence of a thermal anomaly in the upper mantle beneath southern Africa, and therefore the high topography of southern Africa is likely supported by some other mechanism. Two alternative hypotheses have been proposed by other studies, which would be consistent with our findings. Nyblade and Sleep (2003) proposed that the uplift observed in southern Africa might be due to heating from the tail of a Mesozoic plume that was stationary beneath southern Africa for ~26 Ma or longer. They argue that the plume tail could have heated the lithosphere enough to create ~1km of plateau uplift, but that the perturbation to the thermal structure of the lithosphere might not be sufficient to be detected seismically.

Alternatively, some studies have suggested that uplift in southern Africa may be due to buoyancy from the anomalous structure in the lower mantle beneath southern Africa that forms the African Superplume. For example, Lithgow-Bertelloni and Silver (1999) suggested that the low velocity anomaly in the lower mantle beneath southern Africa may represent a large-scale low density mantle upwelling, which can provide dynamic uplift sufficient to support the high topography observed in southern Africa. This suggestion is supported by numerical modeling of mantle flow over the past 75 Ma by Conrad and Gurnis (2003), as well as by Simmons et al. (2007), who jointly inverted
seismic, gravity, and topography data to obtain a three-dimensional model of the superplume.

3.8 References


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Chapter 4

Upper Mantle Shear Wave Velocity Structure Across East Africa from the Inversion of Rayleigh Wave Phase Velocities

4.1 Abstract

In this study, we evaluate candidate models for the Cenozoic uplift and rifting of the East African Plateau. We measure Rayleigh wave phase velocities using a two plane wave approximation, accounting for finite frequency effects and anisotropy, with seismic data from the Tanzania Broadband Seismic Experiment and the AfricaArray East African Seismic Experiment. We then invert these phase velocities for a quasi-three dimensional shear wave velocity model of the upper mantle. We find that the Tanzania Craton is characterized by a high velocity lid extending to a depth of 150-200 km. These high velocities also extend into the Uganda Basement Complex and the Tanzania Divergence Zone to the north and east of the craton, respectively. This may indicate that these regions together form a rigid tectonic block, surrounded by rheologically weaker mobile belts where rifts have formed. We find along strike variation in uppermost mantle velocities for both the Western and Eastern Rift Branches, with the lowest velocities found beneath the Cenozoic volcanic fields. Our model also indicates that the entire East African Plateau is underlain by a broad, thick low velocity zone, extending into the transition zone, which is indicative of a large thermal anomaly underlying the entire East African Plateau. Based upon the lateral extent, thickness, and velocity reduction in this region, together with the likely extension of the anomaly into the transition zone and connection with thermal anomalies to the north, we infer that uplift of the East African
Plateau is due to a large thermal anomaly in the upper mantle, which is likely part of the African Superplume.

### 4.2 Introduction

The East African Plateau comprises a significant portion of one of the largest topographic anomalies on Earth, the African Superswell, which also includes the southeastern Atlantic Ocean basin, the Southern African Plateau, the Ethiopian Plateau, and the Arabian Shield (Nyblade and Robinson, 1994). In East Africa, the anomalous topography of the Superswell is commonly attributed to a thermal anomaly in the upper mantle associated with the development of the Cenozoic East African Rift System (EARS), inferred from the presence of an upper mantle low velocity zone, as well as from gravity and petrological observations (e.g., Simiyu and Keller, 1997; Ritsema et al., 1998; Chesley et al., 1999; Weeraratne et al., 2003; Furman, 2007; Huerta et al., 2009). The origin and size of the thermal anomaly, however, are poorly understood, as is the geodynamic connection between it and the formation of the East African Rift System. Some studies propose that the thermal anomaly and rifting are due to a non-plume source (e.g., King and Ritsema, 2000), while others suggest that one or more mantle plumes are the cause of the thermal anomaly and hence the Cenozoic uplift and rifting (e.g., Burke, 1996; Ebinger, 1997; Nyblade et al., 2000; Park et al., 2006).

Previous seismic studies have primarily focused on the Eastern Branch of the EARS due to the number of seismic stations that have been deployed in Kenya and central and eastern Tanzania over the past few decades. Some seismic studies have found evidence of a westward dipping low velocity zone (LVZ) in the upper mantle beneath
East Africa (e.g., Ritsema et al., 1998; Owens et al., 2000, Nyblade et al., 2000; Park et al., 2006), suggesting that the LVZ may extend down through the upper mantle and into the lower mantle. However, the limited resolution imposed by data availability made it difficult for these studies to determine if the westward dipping low velocity structure observed beneath the Eastern Rift Branch continues at depth beneath the Tanzania Craton connecting to a similar LVZ under the Western Rift Branch, and if it also extends into the lower mantle.

The shape, size, and depth extent of the LVZ provide key constraints on the origin of the thermal anomaly in the mantle beneath East Africa. Therefore, in this study, we use a new dataset obtained from seismic stations deployed in Uganda and western Tanzania, combined with data from previous seismic projects to investigate seismic velocity structure in the upper mantle beneath the entire East African Plateau, including both the Eastern and Western Branches of the EARS. We use a two plane wave method to measure Rayleigh wave phase velocities across most of East Africa, and invert these phase velocities for a quasi-three dimensional shear wave model of the upper mantle. Our shear wave velocity model provides a much more complete image of the upper mantle LVZ than the images shown in previous studies, enabling us to re-evaluate candidate models for the origin of uplift, rifting, and volcanism in East Africa.

4.3 Background

4.3.1 The Geology of East Africa

Our study area focuses on the central portion of the EARS, where the rift bifurcates into an eastern and a western branch around the Archean Tanzania Craton.
The Tanzania Craton is a topographically high area with approximate dimensions of 1000 km by 500 km (Goodwin, 1996). Most of the craton consists of granite, greenstone belts, gneisses, and amphibolites that were formed by at least 2.5 Ga (Cahen et al., 1984).

The Tanzania Craton is surrounded by Proterozoic mobile belts, and is thought to represent a rigid tectonic block around which the mobile belts formed and subsequent rifting occurred. To the southwest is the oldest of the mobile belts, the Ubendian Belt, which was formed between 2.0-2.1 Ga. The northeast-trending Usagaran Belt (~1.9 Ga) abuts the northwest-trending Ubendian Belt to the south-southeast of the craton. The northeast trending Kiberian Belt lies north of the Ubendian Belt, west of the Tanzania Craton, and was formed c. 1.3 Ga (Cahen et al., 1984). The Mozambique Belt to the east and southeast of the craton is formed of multiple assemblages accreted to the craton margin between 570 Ma to 1.2 Ga (Cahen, 1984; Key et al., 1989). In the north, the Tanzania Craton is bordered by the Ruwenzori Fold Belt (2.1 Ga) that extends across southern Uganda. To the north of the Ruwenzori Fold Belt is the Ugandan Basement Complex, which consists of metamorphic rocks, primarily migmatitic gneisses and granulites (Leggo, 1974). The ages of these rocks are uncertain, but are estimated to be between 1.8 and 2.5 Ga, and may be contiguous with the Archean Bomu-Kibalian Craton of central Africa (Cahen et al., 1984; Goodwin, 1996). It is speculated, however, that some deformation of the Uganda Basement Complex may have occurred during the Mozambique orogeny, c. 650 Ma (Leggo, 1974).

Cenozoic rifting in the East African Plateau is a part of the East African Rift System, which extends from the Afar Triple Junction in the north to the Malawi Rift in
the south and forms the boundary between the Somalian and Nubian Plates. The Eastern Rift Branch includes the Kenya (Gregory) Rift in the northern part of our study area. The Kenya Rift consists of several roughly north-south trending rift segments, where rifting initiated 23 Ma (Smith and Mosley, 1993). To the south, the Eastern Rift Branch widens into a 300 km wide area of diffuse rifting, called the North Tanzanian Divergence Zone, which surficially extends into the eastern margin of the Tanzania Craton (Chorowicz, 2005) (Figure 4-1).

Rifting in the Western Rift Branch did not begin until about 12-7 Ma (Ebinger, 1989). However, unlike the Eastern Rift Branch and northern rifts of the EARS, there is little evidence for age progression along strike of the rift (Nyblade and Brazier, 2002). The Western Rift Branch, including its presumed southern extension, the Malawi Rift, includes at least 32 segments. Ebinger (1989) used balanced cross-sections to estimate that the crust has been thinned by less than 10 km, but because of averaging across the rift and rift flanks, this estimate is difficult to confirm using seismic methods. The northern part of the Western Rift Branch consists of, from north to south, the Lake Albert and Lake Edward Rifts (together referred to as the Ugandan Rifts) and the Lake Kivu Rift, followed by the Lake Tanganyika Rift in the central portion of the rift branch. Farther south is the northwest trending Rukwa Rift and the roughly north-south trending Malawi Rift (Figure 4-1). The rifts in the southern portion of the Western Rift Branch are thought to occur, in some place, along reactivated Karoo faults (Daly et al., 1989).
Figure 4-1: Map of East Africa showing the locations of geologic provinces. Approximate locations of volcanic provinces within our study area are shown by gray swaths. Inset map shows the location of the larger map with reference to Africa.
Volcanism and uplift are thought to have been contemporaneous with the initiation of rifting c. 23 Ma for the Eastern Rift Branch (Saggerson and Baker, 1965; Baker et al., 1971; Shackleton, 1978; King, 1978; Coffin and Rabinowitz, 1983) and c. 7-12 Ma in the Western Rift Branch (Ebinger, 1989; Nyblade and Brazier, 2002). Although seismicity rates are much higher in the Western Rift Branch, volcanism is greater in the Eastern Rift Branch, where volcanic material is approximately 10 times more voluminous than in the Western Rift Branch (Ebinger et al., 1989). Figure 4-1 illustrates the location of volcanics within our study region. Volcanism in the Eastern Rift Branch covers a broad area including most of the Kenya Rift and the northern part of the Tanzanian Divergence Zone. In the Western Rift, however, volcanism is confined to a few isolated regions. In the northern portion of the Western Rift Branch, the two major volcanic regions are the Virunga and Kivu volcanic fields (Ebinger, 1989). To the south, volcanism occurs primarily in the Rungwe volcanic field between the Lake Rukwa and Lake Malawi Rifts.

4.3.2 Previous Studies of East Africa

Using a combination of Rayleigh wave phase velocities and receiver functions, Last et al. (1997) report crustal structure for the southern Tanzania Craton, the Ubendian Belt, and the Mozambique Belt using data from the Tanzania Broadband Seismic Experiment. They report that crustal thickness varies between 37 and 42 km for the Tanzania Craton. For the Mozambique and Ubendian Belts, crustal thickness ranges are 36-39 km and 35-44 km, respectively. They conclude that crustal thickness does not vary
greatly across the East African Plateau, and that the crust of the Mozambique Belt has not been significantly thinned due to the diffuse rifting of the Tanzanian Divergence Zone.

Dugda et al. (2005) used receiver functions to measure Moho depth and Poisson’s ratio for stations across the Mozambique Belt surrounding the Kenya Rift as well as for the Ethiopian Plateau. For the region surrounding the Kenya Rift, they report Moho depths ranging from 33 to 42 km. Their findings are consistent with previous studies, which show little variation in crustal thickness away from the rift basins. Their findings for the plateau surrounding the Kenya Rift are also comparable to Moho depths they reported for the Ethiopian Plateau, outside of the Ethiopian rift basins.

Julia et al. (2005) used joint inversion of receiver functions from the Tanzania Broadband Experiment with phase velocities from Weeraratne et al. (2003) and group velocities from Pasyanos et al. (2001) to model crustal and upper mantle structure for the stations of the Tanzania Seismic Broadband Experiment. They find crustal thicknesses consistent with previous studies (e.g., Last et al., 1997; Langston et al., 2002), varying only between 38 and 42 km throughout their study area, with greater crustal thicknesses beneath the craton and western mobile belts. They find a greater reduction in velocity beneath the mobile belts and less velocity reduction beneath the Tanzania Craton when compared to the velocity model from Weeraratne et al. (2003).

A recent study by Kim et al. (2009) imaged crustal structure within the Rukwa Rift of the Western Rift Branch by both modeling waveforms for an earthquake that occurred and was recorded within the rift as well as H-K stacking and waveform modeling of receiver functions, offering new insights into crustal structure within the rift basins of the Western Rift Branch. They estimated crustal thickness within the rift to be
between 33 km and 38 km. They thus concluded that if the crust of the southern portion of the Western Rift Branch has been thinned, it has been only by a few kilometers.

Nolet and Meuller (1982) used a combination of body wave delay times and surface wave velocities to estimate both crust and upper mantle structure in both the Eastern and Western Rift Branches. Beneath the Eastern Rift Branch, they find evidence of a 40 km thick crust underlain by low mantle velocities extending to at least 220 km, which is the resolution limit of their model. They report a 35 km thick crust and faster mantle velocities beneath the Western Rift Branch. They also observe a strong reflector at a depth of 140 km beneath the Western Rift Branch. Combining their velocity models with gravity data, they propose that this may represent an old diapir of chemically distinct material, and that more recently, such a diapir may have risen beneath the Eastern Rift Branch.

Beneath the Kenya Rift, Park et al. (2006) use data from the Kenya Rift International Seismic Project (KRISP) to build a P wave tomography model beneath the Kenya Rift. They find a region of low velocity directly beneath the rift that extends to the west of the rift at depths greater than 150 km to a depth of at least 300 km. Park et al. conclude that their results are inconsistent with a single mantle plume beneath the Tanzania Craton, and suggest that it is instead consistent with a connection to the larger African Superplume to the southwest.

Ritsema et al. (1998) used teleseismic data from the Tanzania Broadband Seismic Experiment to build P and S wave tomographic models of the upper mantle. Their results show a thick keel beneath the craton, with fast velocities extending to a depth of at least 200 km, supporting the idea that the Tanzania Craton is a rigid tectonic block around
which rifting has occurred, without significant internal deformation. They find evidence of reduced velocities beneath the Kenya Rift and the Malawi Rift to depths of at least 400 km, but their study had insufficient resolution to determine whether such a low velocity anomaly exists beneath the Western Rift Branch.

Owens et al. (2000) used the three dimensional velocity model from Ritsema et al. (1998) to stack receiver functions to model transition zone structure beneath the Tanzania Craton and the Eastern Rift Branch. The “410” and “660” discontinuities are typically attributed to the $\alpha$-olivine to $\beta$-spinel and $\gamma$-spinel to perovskite+magnessiowustite transitions respectively, which have opposite Clapyron slopes. Therefore, for a hot thermal anomaly, one would expect to see a depression of the “410” and an elevation of the “660”, while an elevation of the “410” or depression of the “660” would indicate a cool anomaly. Owens et al. (2000) report evidence of a broad depression of the 410 discontinuity and thinning (~30-40 km) of the transition zone beneath the eastern side of the Tanzania Craton and the Eastern Rift Branch. However, they find that there are many local variations and complex arrivals around the 660 discontinuity, making it difficult to discern topography along that discontinuity. These results indicate a thermal anomaly in the upper mantle that extends at least into the shallow part of the transition zone, but which may not extend deeper than 660 km.

A more recent study by Huerta et al. (2009) reported a depressed “410” and “660” discontinuity. Huerta et al. (2009), however, note that in areas with a higher geotherm, additional phase transitions can occur, which may mask the $\gamma$-spinel to perovskite+magnessiowustite transition. They attribute the depressed “660” to a
transition from majorite to perovskite, which is consistent with the thermal anomaly of +350 degrees predicted by their depressed “410”.

Nyblade et al. (2000) review the seismic tomography model of Ritsema et al. (1998) and the receiver function study of Owens et al. (2000) and discuss the implications for mantle dynamics beneath the central portion of the EARS. They propose that the seismic structure of the upper mantle reported in these two studies could be best fit by a mantle plume beneath the eastern part of the Tanzania Craton that has ponded above the 410 discontinuity. They speculate that this ponding of plume material may extend beneath the craton to the west and under the Western Rift Branch.

Using a combination of Pn velocity measurements, results from previous seismic and xenolith studies, and measurements of gravity and heat flow from previous studies, Nyblade and Brazier (2002) propose a model for the temporal and kinematic evolution of the Eastern and Western Rift Branches. They propose that as the Eastern Rift Branch propagated southwards, it encountered the eastern margin of the rigid lithosphere of the Tanzania Craton at c. 10-12 Ma. Stresses from the extensional system were then transmitted across the craton, without causing significant deformation within the craton. Rifting was then initiated in the weaker lithosphere of the mobile belts to the west of the craton, forming what is now the Western Branch of the EARS.

Weeraratne et al. (2003) use data from the Tanzania Broadband Seismic Experiment to measure phase velocities using the two plane wave approximation method employed in this study (Forsyth and Li, 2005). Their study area is similar to, but slightly smaller than ours, but because of the distribution of stations used, their resolution is best in the southern portion of the Tanzania Craton and the Eastern Rift Branch. They find
thick lithosphere beneath the Tanzania Craton extending to a depth of ~170 km, underlain by a low velocity zone, where velocities are reduced by up to 12.1 percent relative to the maximum lid velocity. They find a greater reduction in velocity beneath the Eastern and Western Rift Branches, but these areas of reduced velocities beneath the rifts are more shallow and confined to depths of less than 200 km.

A study of shear wave splitting measurements by Walker et al. (2004) found that anisotropic structure of the East African Plateau is complex, but that the dominant fast direction beneath the rift branches is roughly aligned with the strike of the rift. They find, however, considerable variations in observed fast directions outside of the rifts, and the overall complicated pattern of anisotropy they attribute to fossilized anisotropy, asthenospheric flow around the Tanzania Craton or through channels of thinned lithosphere, and partially molten lenses and dikes in the lithosphere.

Other geophysical and geochemical techniques have also been employed to study the upper mantle structure and evolution of the East African Plateau. For example, Simiyu and Keller (1997) used a large number of gravity measurement across the East African Plateau to determine anomalies in the gravity field. They find a long wavelength negative Bouguer gravity anomaly across the Plateau, with smaller, more narrow negative anomalies beneath the two rift branches. They attribute these negative gravity anomalies to a broad region of low densities in the upper mantle that shallows beneath the rift branches, an interpretation that is consistent with a large thermal disturbance in the upper mantle.

Nyblade (1997) used temperature-depth measurements across the East African Plateau to characterize heat flow beneath the Tanzania Craton and the surrounding
mobile belts. He finds that, away from the Cenozoic rifts, the average heat flow for the Tanzania Craton is 33 mW/m², which is slightly lower than the global average for Archean cratons. In the Mozambique, Kibaran, and Ubendian Belts, he reports heat flow estimates of 64 mW/m², 67 mW/m², and 109 mW/m², respectively. The reported heat flows indicate that the mobile belts are higher than global averages for Proterozoic terrains, but that when accounting for uncertainties, this difference may not be resolvable. He concludes that, if the mantle beneath the East African Plateau has been thermally perturbed, there has been insufficient time for the perturbation to extend through the lithosphere of the Tanzania Craton.

Geochemical analyses of basalts across the East African Plateau indicate that magmas in that region are composites of lithospheric mixing and plume-like sources (Furman, 2007). Isotope studies of basalts in the Kenya Rift and the Western Rift Branch indicate their source magmas have greater crustal and lithospheric assimilation than do the magmas from the Ethiopian Rift to the north, where lithospheric thinning has been more dramatic (Furman, 2007). Basalts from the Kenya Rift and the Kivu Rift have Sr-Nd-Pb isotope values similar to high-µ values characteristic of ocean islands, which are also found in the basalts of the Ethiopian Rift (Chaffey et al., 1989; Furman and Graham, 1999). These results are consistent with magma sourced from a deep mantle plume with some mixing of lithospheric materials (Furman, 2007).

Xenolith studies have also contributed to our knowledge about the dynamics of the upper mantle beneath the EARS. Chesley et al. (1999) conducted a study of xenoliths from the eastern Tanzania Craton and, using a Re-Os isotope analysis, found that xenoliths from depths less than 140 km indicate a refractory mantle at those depths that
formed ~2.8 Ga. They also found that the chemical composition of deeper xenoliths indicated a primitive mantle beneath the craton, with similarities to what is found for plume-related oceanic islands.

The origin of the uplift of the East African Plateau and associated rift formation of the EARS is much debated. Some studies suggest that uplift was initiated due to a plume, possibly associated with the African Superplume. Global tomographic studies have reported a broad low velocity anomaly in the upper mantle beneath the EARS that tilts to the southwest with depth (Ritsema et al., 1999; Grand, 2002). It has been suggested that this anomaly may have a connection to the African Superplume beneath southern Africa, but resolution in the mid-mantle is insufficient to confirm or refute this assertion. Other studies suggest that the uplift in this area may be due to a smaller hot-spot type plume. This plume might be located beneath the Kenya Rift, (e.g., Green et al., 1991; Burke, 1996; Ebinger and Sleep, 1998) or may be located under the eastern margin of the Tanzania Craton (e.g., Nyblade et al., 2000). Yet another proposed source of uplift is edge flow convection, as proposed by King and Anderson (1995). In this model, non-uniform heat flux from the thick lithosphere of the Congo Craton and the adjacent thinner lithosphere of the East African Plateau creates a small scale thermal upwelling (King and Ritsema, 2000). These models are discussed in more detail in section 4.5 and are evaluated using the results of this study.
4.4 Phase Velocity Inversion

4.4.1 Data Processing

We use data from several sources across east Africa, including stations from the Global Seismic Network (GSN), the Tanzania Broadband Seismic Experiment (June 1994-May 1995), the AfricaArray permanent seismic network, and Phase 1 (August 2007-December 2008) and Phase 2 (December 2008-June 2010) of the AfricaArray East African Seismic Experiment (Figure 4-2). This combined dataset includes data from 61 broadband stations.
Figure 4-2: Locations of stations used in this study. Squares indicate GSN stations, and inverted gray triangles indicate permanent AfricaArray stations. White circles are stations from the Tanzania Broadband Seismic Experiment. White and black triangles are stations from Phase 1 and Phase 2 of the AfricaArray East African Seismic Experiment, respectively. Rift faults and the craton boundary are the same as shown in Figure 4-1. The boundaries of the study area are shown as white lines.
We use data for 93 events recorded by Tanzania Broadband Seismic Experiment stations that were windowed and filtered by Weeraratne et al. (2003), the results of which were published in that paper. Following a similar methodology to Weeraratne et al. (2003), we add data for an additional 89 events that were recorded by the AfricaArray East African Seismic Experiment between August 2007 and August 2009. Events that were clearly recorded by at least five stations and had a high signal-to-noise ratio were selected for inversion. We also checked for timing problems and interference from other phases on the unfiltered waveforms before the data was included in the inversion. Because it is important to model both constructive and destructive interference of the two plane waves within our study area, low signal-to-noise ratio at a single station for a generally well-recorded event was not sufficient cause to eliminate the record. All events used had a Mw of 5.4 or greater (Figure 4-3). This expanded dataset gives us good azimuthal coverage within the Tanzania Craton and the surrounding mobile belts and rifts, as is illustrated by the ray coverage shown for 50 second Rayleigh waves in Figure 4-4.
Figure 4-3: Event locations. White circles are events from Weeraratne et al. (2003) that were recorded on the Tanzania Broadband Seismic Experiment. Gray circles are events that were recorded on Phase 1 of the East African Seismic Experiment. Black circles are events that were recorded on Phase 2 of the East African Seismic Experiment. Large gray circles indicate distance intervals of 30 degrees from our study area.
Each seismogram was filtered and windowed separately for 14 periods ranging from 20 to 182 seconds. At each frequency, a zero-phase-shift, 10 mHz wide bandpass filter centered at that frequency was used to filter the seismograms. For each event a window length was chosen for each period so that the record at each station contained the

Figure 4-4: Raypath coverage for 50 second Rayleigh waves. Station symbols are described for Figure 4-2. Rift faults and craton boundary are the same as shown in Figure 4-1.
primary Rayleigh wave arrival. This process of windowing each seismogram around the arrival of the Rayleigh wave limits noise and interference from body wave arrivals and the surface wave coda, while maintaining records of equal length. For each filtered and windowed seismogram, discrete Fourier analysis was used to measure the phase and amplitude for that record.

4.4.2 Model Parameterization and Inversion Methodology

Traditional surface wave tomography methodology assumes that a surface wave impinging on a study area or station can be represented by a single plane wave, propagating along a great-circle path. While this approximation is accurate for a sufficiently small area on a spherically homogenous Earth, it does not account for distortions of the wave field created by heterogeneities of a realistic Earth model. In this study, we apply a method developed by Forsyth and Li (2005) that accounts for deviations in the incoming wavefield by approximating it with two hypothetical plane waves based upon phase and amplitude measurements at the stations prior to inverting for velocity parameters. Initially, velocity is held constant, while the phase, amplitude and propagation azimuth of the two waves are solved for using an iterative simulated downhill simplex annealing method (Press et al., 1992). A secondary, linearized inversion (Tarantola and Valette, 1982), minimizing errors in a least-squares sense, is then employed to simultaneously solve for corrections to the velocity model and wave parameters.

This method utilizes a Cartesian coordinate system on a flat Earth and assumes that the complex incoming wave field can be approximated by two plane waves. This
assumption is most realistic when station aperture is small, thereby limiting the
sensitivity of the propagation of the waves to the curvature of the Earth (Donald Forsyth,
personal communication). Therefore, for earthquakes that were recorded on stations
separated by large distances, we have opted to process records from stations in smaller
groups, rather than as a single event. Because of the compact and non-linear station
configuration of Phases 1 and 2 of the AfricaArray East African Seismic Experiment,
earthquakes recorded by each phase have been treated as one event. Earthquakes
recorded by the Tanzania Broadband Seismic Experiment, however, have been processed
as three events (i.e., solving for three sets of wave parameters), each recorded by one of
three overlapping sections of the array as indicated in Figure 4-5.
A series of inversions were conducted for each period, starting with an inversion for an average one-dimensional phase velocity dispersion curve representing the entire study area. For this initial inversion, we use the average phase velocities for the Tanzania...
Craton from Weeraratne et al. (2003) as our starting model. Subsequent inversions were iterative, each using the phase velocities from the previous, simpler inversion as a new starting model. The next inversion included a solution for one-dimensional phase velocity curves for each geologic region indicated in Figure 4-6. The velocity curves for each region were then used as starting models to invert for velocities at each node, creating two-dimensional velocity maps. This iterative approach of refining the starting model for each inversion limits errors that might be introduced into the two dimensional phase velocity maps from a starting model that poorly represents average structure in the region.

We use a Gaussian sensitivity function to account for the sensitivity to heterogeneities off of the paths of the two plane waves within our study area for the one-dimensional inversions. It has been shown, however, that because surface waves include a finite number of frequencies, a Gaussian approximation is valid only when the characteristic wavelength of a period being used is smaller than the size the structure being measured (e.g., Zhou et al., 2004; Yang and Forsyth, 2006a). In a study by Zhou et al. (2004), finite frequency sensitivity kernels for both the phase and amplitudes of surface waves were calculated for full travel paths from the source to receiver. Zhou et al. (2004) showed that the polarity and amplitude of the sensitivity of surface waves to off path structure varies away from the assumed great-circle travel path and also varies in amplitude with distance along the travel path. The sensitivity kernels of Zhou et al. (2004) were applied to regional surface wave tomography, using both one and two plane waves, by Yang and Forsyth (2006a). Yang and Forsyth (2006a) also demonstrated that
there is improved resolution of small features when the finite frequency effects of surface wave propagation are accounted for in this way. We therefore follow the method developed and applied by Yang and Forsyth (2006a; 2006b) for accounting for these finite frequency effects of surface wave propagation in our two dimensional inversions. To further improve our resolution, we also solve for an estimate of average azimuthal anisotropy at each period.

Our study area is 1883 km by 1664 km (interior node size 1440 km x 1220 km) and includes the entire Tanzania Craton, the surrounding mobile belts, and both the Eastern and Western Branches of the EARS (Figures 4-1, 4-2). The area was divided into a set of interior and exterior nodes. Interior nodes were spaced 0.5 degrees apart and exterior nodes were spaced at 1 degree, giving a total of 837 nodes (Figure 3-6). An a priori error estimate was assigned to each node to determine damping values. This error estimate was higher for exterior nodes (1.5 km/s) than interior nodes (0.15 km/s) to allow the exterior nodes, which are outside of our primary areas of interest, to preferentially absorb any abnormalities in the incoming wave field that could not be accounted for by using the two plane wave approximation.
4.4.3 Phase Velocity Results

In Figure 4-7, we show an average phase velocity curve representing all of our study area, and the average phase velocity curve for southern Africa (Chapter 3) for
reference. At all periods, the phase velocity curve for East Africa is slower than that for southern Africa. The difference between the two curves is most pronounced at periods between 100 and 143 seconds. Figure 4-8 shows average phase velocity curves for each geologic region as indicated in Figure 4-6 and also shows phase velocities for the Kaapvaal Craton (Chapter 3). At all periods, especially at periods greater than 40 seconds, all regions within the East African Plateau, including the Tanzania Craton, are significantly slower than the Kaapvaal Craton. For some regions and periods, this difference is more than 0.2 km/s.

Velocities are generally fastest for the Tanzania Craton, and velocities in the region directly north and west of the craton are comparable to those for the Tanzania Craton. Velocities for both the Eastern and Western Rift Branches are slower than for the craton. At periods less than 67 seconds, the Western Rift Branch is slower than the Eastern Rift Branch. At longer periods, though, the Eastern Rift Branch is slower. At the longest periods we model, 167 and 182 seconds, the velocities for the Eastern and Western Rift Branches are comparable.
Figure 4-7: The average one dimensional phase velocity dispersion curve for eastern Africa is shown by a solid line. Error bars indicate one standard deviation. For comparison, the average one dimensional phase velocity dispersion curve for southern Africa is shown by a dashed line.
Figure 4-9 shows 2-dimensional phase velocity maps for selected periods. At all periods, the Tanzania Craton is faster than the surrounding regions. This is most pronounced at periods greater than 25 seconds, and velocities are generally greatest in a linear region trending north-south through the center of the craton. At all periods, the region directly north and west of the Tanzania Craton has velocities comparable to those observed within the craton. At periods less than 50 seconds, there are regions of low velocities focused beneath the Western Rift Branch, the Malawi Rift, and the Kenya Rift. At periods longer than 50 seconds, low velocities are primarily found beneath the Kenya Rift and the Malawi Rift, although diffuse areas of low velocities can also be found beneath the Western Rift Branch and beneath other regions of the East African Plateau.
Error bars shown in Figures 4-7 and 4-8 indicate formal uncertainties of one standard deviation as calculated in the phase velocity inversion. These mathematical calculations of uncertainty are probably unrealistically small. Model uncertainties for the two dimensional inversions give more realistic measures of uncertainty and are calculated from the diagonals of the covariance matrix (Appendix E). Phase velocity maps in Figures 4-9 and 4-10 are cropped to show only areas where the model uncertainty is less than 0.06 km/s, which is intermediate compared to values chosen by previous studies (e.g., 0.04 km/s for Weeraratne et al., 2003; ~0.08 km/s for Li and Burke, 2006).

Figure 4-9: Phase velocity maps for selected periods. Rift faults and craton boundaries are the same as shown in Figure 4-1.
To further examine the resolution of our phase velocity models, checkerboard resolution tests were conducted and selected periods are shown in Figure 4-10. Resolution tests for additional periods are shown in Appendix E. At periods up to 125 seconds, input checkers were two degree squares alternating between positive and negative 5 percent velocity anomalies, and for periods of 143 seconds and greater, input checkers were three degree squares. At periods of 125 seconds and less, the locations of checkers are well resolved, but less than half of the amplitude of the input velocity anomalies is recovered. For periods of 143 seconds and greater, two degree checkers were not well resolved, but three degree checkers can be resolved (Appendix E). Resolution is best in the central region of the Tanzania Craton, where the raypath coverage is greatest. The areas with the poorest resolution are in the southwestern and the southeastern parts of the model domain.
As described in section 4.4.2, our inversion for two dimensional phase velocity maps includes a solution for the average azimuthal anisotropy at each period, which is illustrated in Figure 4-11. For periods of 25 seconds and less, the fast direction is roughly east-west, and probably represents anisotropy due to crustal structure. At periods greater than 25 seconds, the azimuth of the fast propagation direction varies between periods, but is generally oriented north-south, which is consistent with fast directions reported by previous studies (e.g., Weeraratne et al. 2003; Walker et al. 2004). At periods of 143

Figure 4-10: Checkerboard resolution tests for selected periods. Input checkers were two degree squares for the maps at 25, 67, and 100 seconds and were three degree checkers for the map at 143 seconds. Maps are clipped to show only areas with a model uncertainty of 0.06 km/s or less. Rift faults and craton boundaries are the same as shown in Figure 4-1.
seconds and below, we observe azimuthal anisotropy of 3 percent or less. At 167 and 182 seconds, however, we observe a significant increase in the percentage of anisotropy, to approximately 7 and 11.5 percent respectively. While we do not eliminate the possibility that this may represent a real increase in anisotropy at greater depths, the decreased number of raypaths and lateral resolution at these depths indicate that this apparent increase in anisotropy may not be well resolved.

Figure 4-11: Fast direction for azimuthal anisotropy and percent change in velocity at each period.
4.5 Shear Velocity Inversion

4.5.1 Inversion for Shear Wave Velocity

We follow the method of Park et al. (2008) to obtain a quasi-three dimensional shear wave velocity model for the East African upper mantle by inverting our phase velocity measurements. To create a quasi-three dimensional model, we first constrain crustal structure based on a priori information, and then invert for a one dimensional shear wave profile at each of our interior nodes (Figure 4-6), following the method of Julia et al. (2000). We then use a Gaussian weighted smoothing parameter with a characteristic length of 80 km applied across the nodes to create the quasi-three dimensional shear wave model from our one dimensional models.

To compute shear wave velocity profiles at each node, using the same node locations and spacing as the interior nodes of the phase velocity inversions, we follow the inversion methodology described by Julia et al. (2000). This inversion technique uses an iterative damped generalized linear least-squares inversion that was designed to jointly invert both dispersion and receiver function data. To apply this joint inversion method in our study, where we include only phase velocities, we designate the weight assigned to receiver function data to be equal to zero. The technique of Julia et al. (2000) inverts observed data, which in our case are phase velocities, with model smoothness and weighting parameters to solve for changes to a starting velocity model. In the case of Rayleigh wave phase velocities, the inversion is accomplished by minimizing a function describing their behavior from Russell (1987).
To build a starting model representative of each geologic region indicated in Figure 4-5, we use crustal thickness and velocity estimates (Table 4-1) that are consistent with those obtained in previous studies (e.g., Julia et al., 2008; Tugume et al., 2009) over an IASP91 mantle as an initial model to invert the average phase velocity curves for each geologic region (Figure 4-7). We include a one km thick layer above and below the Moho with no vertical smoothing to constrain the Moho depths and still allow large variation in velocity across the boundary. For all other layers above the 410-discontinuity, a linear smoothing parameter was used to prevent large and unrealistic contrasts in velocity between adjacent layers, which Rayleigh waves are not sensitive to. At depths greater than 410 km, the models were set equal to IASP91.

The results from these preliminary inversions were used as starting models for each node in our study area. As described above for the inversion for the starting models, in the one dimensional inversions at each node, velocities below 410 km were fixed to IASP91. Smoothing was applied for all layers above 410 km, except for the 1 km thick layers above and below the Moho, effectively fixing the Moho depth, while allowing velocities above and below it to vary.
4.5.2 Shear Wave Velocity Results

Depth slices from our shear wave structure model are shown in Figure 4-12, and cross-sections through the model are shown in Figure 4-13. Additional depth slices and cross-sections are shown in Appendix E. The Tanzania Craton is the dominant fast feature with velocities up to 4.65 km/s. High velocities extend to the north beneath the Ruwenzori Belt, and to the west and southeast beneath the Ubendian, Kibaran, and
Mozambique Belts. The fastest velocities in the craton and the deepest extent of fast velocities lies along a roughly north-south axis that runs through the western part of the craton.

We also observe a region of reduced velocity beneath the entire study area at depths greater than ~225 km. Velocities less than 4.4 km/s are found in all areas of our model at depths greater than 225 km, and in most regions, velocities remain below 4.4 km/s to the 410-discontinuity, where our models are tied to IASP91. The lowest velocities are less than 4.15 km/s, and are generally found beneath the rifts at depths between 250 and 350 km.
Velocities are also slow throughout the upper 200 km of the mantle beneath the rifts. Velocities beneath the Kenya Rift are between 4.2 and 4.35 km/s in the upper 200 km, but are faster to the south in the Tanzanian Divergence Zone, where some parts of the rift are underlain by faster velocities, up to 4.5 km/s, similar to those beneath the Tanzania Craton at similar depths, although the area has been affected by rift faulting.

Figure 4-12: Shear wave velocity maps at selected periods from the quasi-three dimensional shear wave model. Locations for the cross-sections (Figure 4-13) are shown on the 250 km depth slice (d). Rift faults and craton boundary are the same as shown in Figure 4-1.
over a 300-400 km wide region. At depths between 200 and 350 km, however, velocities are similar (<4.3 km/s) to those found along the strike of the Eastern Rift Branch to the north and south of the Tanzanian Divergence Zone.

In the Western Rift Branch, velocities in the upper 200 km vary significantly along the strike of the rift. In the northern Western Rift Branch, beneath the Kivu and Uganda Rifts, velocities are slightly faster than beneath the Kenya Rift at comparable depths, ranging between 4.3 and 4.45 km/s. Velocities in the upper 200 km are slowest beneath the Kivu Rift, where there is columnar structure in which velocities are 0.1 km/s slower than the surrounding mantle. To the south, beneath the Lake Tanganyika Rift, velocities are faster, ranging between 4.35 and 4.55 km/s. For the Malawi Rift, velocities are also slow (4.25-4.45 km/s) and decrease to less than 4.15 km/s at greater depths.
Figure 4-13: Cross-sections through our quasi-three dimensional shear wave velocity model. The paths of the cross-sections are shown in Figure 4-12(d). UR – Ugandan Riffs; VR – Kivu Rift; TR – Tanganyika Rift; UBC – Ugandan Basement Complex; TC – Tanzania Craton; MR – Malawi Rift; KR – Kenya Rift; KB – Kibaran Belt; MB – Mozambique Belt; UB – Ubendian Belt; TDZ – Northern Tanzania Divergence Zone; RR – Rukwa Rift

Figure 4-14 shows average one dimensional shear wave profiles for the regions indicated in Figure 4-5, with error bars showing one standard deviation for each depth interval. For the Tanzania Craton and regions directly to its north and west, the velocities
in the upper 150 to 200 km are faster than AK135 (Kennet et al., 1995), but are slower than AK135 at greater depths, reaching an average minimum of 4.2 km/s at a depth of 290 km. There is an approximate 10 percent decrease in velocity from the fastest lid velocity to the slowest upper mantle velocity in the average model of the Tanzania Craton. The Western Rift Branch is slower than AK135 at all depths, except between 150 and 200 km, where velocities are comparable to AK135. The Eastern Rift Branch is slower at all depths than AK135. All regions are slower than AK135 at depths greater than 200 km.
Figure 4-14: Average shear wave velocity models for geologic regions (Figures 4-1 and 4-6) within our study area. Gray swaths indicate one standard deviation for each layer. For comparison, AK135 is shown by a dashed line.
4.6 Discussion

Our phase velocities inversions indicate that the Tanzania Craton is, at all periods, slower than the Archean Kaapvaal Craton of southern Africa. And, the Eastern and Western Rift Branches are both slower than the Tanzania Craton at all periods. At periods less than 60 seconds, the Western Rift Branch exhibits the slowest velocities, however, at periods longer than 60 seconds, velocities are comparable to or faster than the velocities in the Eastern Rift Branch. We find evidence of azimuthal anisotropy that is oriented in a generally north-south direction. This is broadly consistent with the direction of anisotropy reported by previous studies (e.g., Weeraratne et al., 2003; Walker et al., 2004) as well as with the general trend of the EARS. We find an increase in the amount of anisotropy at the longest periods, which may indicate an increase in anisotropy with depth, but because of limited raypath coverage at these periods, this apparent increase may not be well resolved.

In our shear wave structure model, the Tanzania Craton is a prominent feature, and fast velocities associated with it extend not only to the mapped boundary of the craton, but also farther to the north and the southeast. We find that the maximum lid velocities for the Tanzania Craton are comparable to those of the Kaapvaal Craton and other Archean cratons (Chapter 3). This indicates that any thermal anomaly in the upper mantle has not destroyed the lithospheric keel of the Tanzania Craton, as has occurred elsewhere (i.e., the north China Craton; Huang et al., 2009).
We find that the thickness of the lithosphere beneath the Tanzania Craton is between 150 and 200 km, depending on whether the bottom of the lithosphere is chosen as the middle or the base of the maximum negative velocity gradient (Figure 4-14). This is consistent with the lithospheric thickness reported for the Tanzania Craton by Weeraratne et al. (2003) and for other Archean cratons (see Figure 3-15, Chapter 3), but is much thinner than the lithospheric thickness of >250 km reported by Priestley et al. (2008). We find that, within the craton, thickness may vary by more than 50 km. Depth slices through our shear wave model in Figure 4-12 show that fast velocities defining the cratonic lithosphere extend deepest along a north-south zone that runs through the western part of the craton. The cross-section through the southern craton (4-13e) also illustrates the thicker region of fast velocities beneath the western craton, as well as the general thinning of the lithosphere towards the Indian Ocean to the east. The thicker lithosphere to the west may indicate that the lithosphere beneath the Tanzania Craton has been thinned in the east, possibly due to heating by a thermal upwelling in that region as has been suggested by previous studies (e.g., Ebinger, 1997; Nyblade et al., 2000; Park et al., 2006), and as discussed in greater detail below.

The northern boundary of the Tanzania Craton has been tentatively defined by the southern extent of the Ruwenzori Fold Belt (Cohen et al., 1984), which may be thrust over the craton. Farther north, the Ruwenzori Fold Belt may also be thrust over the Ugandan basement complex, the age of which is uncertain. The tectonic relationship between the basement of the Tanzania Craton and the Ugandan basement complex is not well documented, so we take the opportunity here to compare the seismic structure across the Ruwenzori Fold Belt. Our depth slices show that velocities throughout the upper
mantle are similar between the portion of the Tanzania Craton beneath Lake Victoria and the area directly to the north of it. The cross-section in Figure 4-13b includes the Tanzania Craton, the Ruwenzori Fold Belt, and the Ugandan basement complex. We find no seismic evidence at any mantle depth that the Ruwenzori Fold Belt represents a through-going lithospheric structure separating two cratonic blocks. This finding is consistent with the suggestion made by many previous investigators that the two rift branches have formed in the weaker mobile belts surrounding a central rigid block composed of the Tanzania Craton and the Ugandan basement complex.

We also observe an eastward extension of the high velocities indicative of cratonic lithosphere into the northern Tanzania divergence zone and the Mozambique Belt. This finding is consistent with previous studies that have noted high lithospheric mantle velocities in this region (Brazier et al., 2000) and crustal thicknesses that indicate little, if any crustal thinning due to the diffuse rifting in the region (Last et al., 1997; Dugda et al., 2005). This finding suggests that the Tanzania Craton lithosphere extends at depth to the east beneath the Mozambique Belt in northern and central Tanzania.

Our results show three zones in the upper 200 km where velocities are slowest, beneath the Kenya Rift, the Kivu and southern Ugandan Rifts, and the Malawi Rift. These are all regions with Cenozoic volcanism. The largest of these low velocity regions, volumetrically, is beneath the Kenya Rift, which is more volcanically active than the Western Rift Branch (Ebinger, 1989; Furman, 2007). The other two regions of low velocity in the upper 200 km are focused beneath smaller, Cenozoic volcanic centers, the Virunga volcanic field and the Rungwe volcanic field (Figure 4-1).
Our model is consistent with previous studies that have found an apparent westward dipping low velocity structure beneath the Eastern Rift Branch (Ritsema et al., 1998; Park et al., 2006), however, our results indicate that structure beneath the Eastern Rift Branch is part of a broad low velocity structure that extends beneath the Tanzania Craton and connects to the Western Rift Branch. This finding is also consistent with global tomography models, which indicate a large-scale low velocity zone beneath the entire East African Plateau (e.g., Grand, 2002; Ritsema et al., 1999; Simmons et al., 2007). Thus, the westward dip of the low velocity feature beneath the Eastern Rift Branch that has been reported by previous studies (e.g., Ritsema et al., 1998; Park et al., 2006) is an apparent dip caused by the lack of resolution beneath the western part of the East African Plateau in these earlier studies.

Our preferred shear velocity model indicates that the low velocity zone beneath the East African Plateau extends to at least 410 km. Because of the decreasing sensitivity of surface waves to structure at increasing depths, it is necessary to constrain velocities at some depth in our inversion to prevent unrealistic structures in the deepest parts of the model. The periods for which we have measured phase velocities have their peak sensitivity to mantle structure at depths shallower than the transition zone, and so we have opted to constrain velocities at depths greater than 410 km to IASP91. This approach leaves open the question of whether or not the LVZ extends into the transition zone or is confined to depths above it. To address this resolution issue, Figure 4-15 shows tests for each geologic region, in which velocities are constrained to IASP91 at depths of 300, 350, and 410. It is clear from these tests that our phase velocity observations cannot be fit without including a low velocity zone that extends at least to a
depth of 410 km. The models that are fixed to IASP91 at shallow depths routinely underpredict phase velocities at periods between 50 and 100 seconds, and overestimate the phase velocities at longer periods. The match between predicted and observed phase velocity decreases as the depth at which the model is fixed to IASP91 decreases (Figures 4-15 and 4-16).

Our phase velocity measurements, including periods up to 182 seconds, provide only limited resolution in the transition zone. However, to test the possibility that the low velocity zone we observe in the upper mantle may extend into the transition zone, we performed an additional one dimensional inversion for each region while fixing the velocities in the transition zone to be 5 percent slower than IASP91. The Clapeyron slope for the olivine to β-spinel transition predict that for a 5 percent slow transition zone, the transition would occur at a depth of 420 km (Bina and Helffrich, 1994), so we fix the velocity transition to that depth. The results of these tests indicate that the observed phase velocities can be better fit by a slow transition zone, however, this improvement is smaller than our uncertainties and therefore may not be well constrained. Results for a quasi-three dimensional model with a 5 percent slow transition zone are shown in Appendix E. Previous receiver function studies are consistent with this finding that the LVZ extends into the transition zone (e.g., Owens et al., 2000; Huerta et al., 2009).

Seismic velocity at a given depth is a factor of many material properties including chemical composition, water content, grain size, partial melts, and temperature. Therefore, it is possible that some of the velocity variations we observe in our study are due to factors other than thermal anomalies. However, variations in chemical composition is estimated to only account for velocity variations of less than one percent
(e.g., Sobolev et al., 1996; Goes et al., 2000; Griffin et al., 2003), and water content is unlikely to be a factor due to the long absence of subduction in this region. Thus, we attribute the low velocities in the upper mantle beneath the East African Plateau primarily to elevated temperatures.

Figure 4-15: One dimensional shear wave inversions for each region constrained to IASP91 at 300 (green), 350 (blue), and 410 km (red).
Figure 4-16: Observed phase velocities (black) and phase velocity dispersion curve predicted for the one dimensional shear wave inversions constrained to IASP91 at 300 (green), 350 (blue), and 410 km (red) (Figure 4-15). Purple curves indicate the predicted phase velocity curves for shear wave models with a 5% slow transition zone and a “410” discontinuity depressed to 420 km. Error bars indicate an uncertainty of 0.06 km/s.
To estimate the temperature anomaly represented by the low velocity zone we observe, we assume a conversion factor of 1 Kelvin for a velocity change of 0.0012 km/s. This conversion factor was determined assuming a grain size of 10 mm, at an average temperature of 1300 degrees C, and for a period of 100s (Faul and Jackson, 2005; Weins et al., 2008). At a depth of 300 km, where our average shear wave profile for East Africa has a minimum velocity, velocities are approximately 0.5 km/s less than IASP91. This represents a temperature anomaly of 415K, which is broadly consistent with the temperature anomaly predicted in the transition zone from receiver functions (Owens et al., 2000; Huerta et al., 2009).

The presence of a broad LVZ beneath the entire study region and the thermal anomaly that it represents has important implications for understanding the uplift of the East African Plateau and the origin of the EARS. Three primary models have been proposed for the uplift of the East African Plateau and rifting (Figure 4-17). The first model suggests that small-scale convection occurs, either due to stretching of the lithosphere (Buck, 1986; Mutter et al., 1988; Webb and Forsyth, 1998), or due to edge flow convection from the Congo Craton (King and Anderson, 1995; King and Ritsema, 2000) (Figure 4-17a). However, the depth and lateral extent of the low velocity zone imaged in our model is too large to be explained by edge flow convection, the effects of which are shallow and confined to depths of less than 350 km (King and Anderson, 1995).
A second possible model attributes the uplift and rifting of the EARS in East Africa to a mantle plume that rises beneath the Tanzania Craton, and has ponded in the upper mantle above the 410-discontinuity (e.g., Simiyu and Keller, 1997; Ebinger et al., 1997; Nyblade et al. 2000, Park et al., 2006) (Figure 4-17b). The broad LVZ with similar velocities throughout our study area, and the apparent thinning of the lithosphere beneath the eastern side of the Tanzania Craton is consistent with this hypothesis. The thickness (> 200 km) and lateral extent of the low velocity zone beneath the entire East African Plateau are large for a starting plume head. The volume of thermally perturbed mantle beneath the East African Plateau, assuming a minimal thickness of 200 km
beneath the entire East African Plateau, is 351 million cubic km. The estimate is larger than what has been modeled for other plumes (e.g., Sleep, 1997; Ebinger and Sleep, 1998), but is not necessarily larger than what is geodynamically possible for a plume originating at the core-mantle boundary (e.g., Campbell and Griffiths, 1990; Griffiths and Campbell, 1991; Campbell, 2005). However, we have given a minimum estimate of volume for the thermal anomaly beneath the East African Plateau, and have not accounted for the connection to the thermal anomaly beneath the Ethiopian Plateau to the north (Benoit et al., 2006a; 2006b), nor the likely extension of the thermal anomaly into the transition zone. When this broad anomalous region is considered together, along with the temperature contrast that we estimate for the anomaly, mantle plume models cannot easily account for this volume of thermally perturbed mantle.

A third geodynamic model that has been proposed suggests that there is a direct thermal connection between the African Superplume in the lower mantle beneath southern Africa and the thermal anomaly in the upper mantle beneath eastern Africa (Figure 4-17c). Global tomographic studies (Ritsema et al., 1999; Grand, 2002; Simmons et al., 2007) lend support to this model. The broad thermal anomaly we observe in the upper mantle beneath the East African Plateau is more consistent with this model than a starting plume head model.

Additional support for a broad thermal anomaly in the upper mantle beneath all of East Africa comes from gravity studies and geochemical studies. A gravity study by Simiyu and Keller (1997) finds evidence for a broad long wavelength negative gravity anomaly across the Tanzania Craton and the Eastern and Western Rift Branches. Shorter wavelength, higher amplitude negative anomalies over the two rift branches are
superimposed on the longer wavelength anomaly. An early study of gravity and elastic plate thickness across the EARS determined that the high topography of the East African Plateau is over-compensated by buoyant low-density mantle material, a finding that is also consistent with a broad thermal upwelling (Ebinger et al., 1989).

Geochemical evidence for a thermal upwelling is provided by both xenolith studies and geochemical analysis of basalts. Chesley et al. (1999), for example, found that the composition of xenoliths from beneath the eastern Tanzania Craton indicates a primitive mantle, consistent with a deep thermal upwelling. Isotope studies of basalts in the Kenya Rift and the Western Rift Branch indicate high-µ magma composition, which is typical of lower mantle plume-like sources, as reviewed in section 4.3.2 (Pik et al., 2006; Furman, 2007)

4.7 Summary and Conclusions

In this study, we invert data from the Tanzania Broadband Seismic Experiment together with new data from GSN and the AfricaArray East African Seismic Experiment stations for phase velocities using the two plane wave approximation method of Forsyth and Li (2005). We find that phase velocities beneath the Tanzania Craton and the areas directly north and west of the craton are faster, at all periods, than those beneath the Western and Eastern Rift Branches. At the shortest periods, the Western Rift Branch is slower than the Eastern Rift Branch, but at periods longer than 50s, this relationship is reversed. Anisotropy is found at all periods, but seems most pronounced at periods of 167 and 182 seconds. The fast direction for all periods is roughly north-south, generally aligned with the trend of the EARS.
Using crustal models consistent with recent studies (e.g., Julia et al., 2005; Tugume et al., 2009), we invert our phase velocities for a quasi-three dimensional model of shear wave velocity. Our shear wave velocity model indicates that the lithosphere of the Tanzania Craton is fast to depths between 150 and 200 km. The fast velocities characteristic of the Tanzania Craton extend to the north into the Uganda Basement Complex and to the east into the Tanzanian Divergence Zone, which may indicate that these regions together form a rigid, undeformed block, around which rifting occurs in the weaker lithosphere of the mobile belts. Both the Eastern and Western Rift Branches are generally slower than the Tanzania Craton. Both branches, however, show variable structure along strike in the upper 200 km of the mantle. Slowest velocities are found beneath regions with Cenozoic volcanism.

At depths greater than 225 km, we find a low velocity zone, with velocities reduced by more than 10 percent relative to lid velocities, across the entire East African Plateau, which extends to and likely into the transition zone. Based upon the velocity reduction compared to IASP91, we estimate that this low velocity zone represents a thermal anomaly of approximately 415 Kelvin, which is consistent with thermal anomalies estimated for the transition zone (e.g., Huerta et al., 2009). The size of the inferred thermal anomaly is, at minimum, larger than is typically expected of a staring plume head, and is likely part of the African Superplume.

In summary, based upon the broad, thick low velocity zone we observe, and the evidence from other seismic, geophysical, and geochemical studies, we conclude that the East African Plateau is underlain by a broad thermal anomaly throughout the upper mantle that extends into the transition zone (Figure 4-16). Because this thermal anomaly
is larger than is typically associated with starting plume heads, we conclude that it is part of the northeastern extension of the African Superplume imaged by global tomographic studies (e.g., Ritsema, 1999; Grand, 2002; Simmons et al., 2007).

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Chapter 5

Conclusion

In this thesis, I have used a variety of techniques to investigate crustal and upper mantle structure in three different tectonic regimes within the Afro-Arabian region to better understand seismicity, rifting, and uplift of areas within and surrounding the African Superswell. In Chapter 2, I discussed the depth extent of seismicity and implications for lithospheric rheology in the convergent margin of the Zagros Mountains. In Chapter 3, I examined upper mantle shear wave velocity structure beneath the southern African Kaapvaal Craton and surrounding mobile belts to evaluate seismic evidence for an upper mantle thermal anomaly supporting the high elevations of the Southern African Plateau. In Chapter 4, I modeled the seismic velocity structure of the upper mantle depths to evaluate three possible geodynamic models for the Cenozoic uplift and rifting in East Africa.

In Chapter 2, I presented source parameters of six moderate-sized earthquakes that nucleated within the Zagros Mountains, which were determined through a combination of depth phase modeling and moment tensor analysis. I found that, although all six events were reported as having lower crustal or upper mantle source depths in global earthquake catalogs, my results indicated that all six events nucleated within the upper 11 km of the crust. Previous studies of seismicity within this region have debated whether the lower crust and upper mantle beneath the Zagros Mountains might be aseismic. My results, however, lend support to the assertion that seismicity in the Zagros
Mountains is limited to the upper crust. The results reported in Chapter 2 also suggest that global catalogs may overestimate source depths in this region.

In Chapter 3, I investigated the upper mantle seismic structure beneath southern Africa. Southern Africa is characterized by anomalously high topography, which some previous studies have suggested may be supported by a thermal anomaly in the upper mantle that is manifest seismically as a low velocity zone (e.g., Priestley, 1999; Li and Burke, 2006; Priestley et al., 2006; Wang et al., 2008). The existence of such a low velocity zone is debated, however, and some other studies have found little seismic evidence for a low velocity zone in the upper mantle to support the presence of a thermal anomaly (e.g., Zhao et al., 1999; Freybourger et al., 2001; James et al. 2001; Saltzer, 2002; Larson et al., 2006; Chevrot and Zhao, 2007; Hansen et al., 2009). I used a combination of data from the Southern African Seismic Experiment, AfricaArray and Global Seismic Network stations, to measure phase velocities using a two plane wave approximation method (Forsyth and Li, 2005). These velocities were inverted for a quasi-three dimensional shear wave velocity model. By comparing the shear wave velocity structure beneath the Kaapvaal Craton to the upper mantle velocity structure beneath other Archean Cratons, I found that the reduction in velocity beneath the Kaapvaal Craton is not anomalous relative to other Archean cratons. I thus found little seismic evidence for an upper mantle seismic anomaly indicative of a thermal anomaly beneath the Kaapvaal Craton supporting the high topography observed in southern Africa.

In Chapter 4, I applied the same methodology from Chapter 3 to investigate upper mantle seismic structure in East Africa and to evaluate possible models for the origin of
the rifting and uplift of the East African Plateau. The East African Plateau, like the Southern African Plateau, is characterized by anomalously high topography. In East Africa, seismic, petrologic, and gravity studies provide evidence that this high topography is due to buoyancy from a thermal anomaly in the upper mantle (e.g., Ritsema et al., 1998; Simiyu and Keller, 1997; Chesley et al., 1999; Owens et al., 2000; Nyblade et al., 2000; Nyblade and Brazier, 2002; Weeraratne et al, 2003, Park et al., 2006). The size and source of the thermal anomaly, however, has been greatly debated. Using data from the Tanzania Broadband Seismic Experiment and from the AfricaArray East African Seismic Experiment, I measured phase velocities and inverted them for a quasi-three dimensional shear wave model. I found that the entire East African Plateau is underlain by a broad low velocity zone, which extends into the mantle transition zone, indicating that the low velocity anomaly beneath East Africa is broader than has been proposed in previous studies. Based upon my observations, I evaluated possible geodynamic models for the uplift and rifting of the plateau. The size of the anomaly is not consistent with the edge flow. Given that the anomaly appears to extend into the transition zone it is unlikely to represent a typical plume. The size of the anomaly is most consistent with an upper mantle extension of the African Superplume.

A comparison of the upper mantle structure beneath southern and eastern Africa in Chapter 3 and 4 of this thesis indicates that uplift in these areas comes from different sources. Phase velocities beneath the Kaapvaal Craton are higher than beneath the Tanzania Craton at all periods. Although lid structure is similar for both cratons, the shear wave model for the Kaapvaal Craton indicates a narrow zone of low velocities reduced by only 3.9 percent, relative to the highest average lid velocity. For the Tanzania
Craton, however, the shear velocity model indicates a broad, thick low velocity zone with velocities reduced by more than 10 percent. Thus, temperatures beneath the Tanzania Craton must be elevated to a higher temperature over a broader region than beneath the Kaapvaal Craton. I concluded that while a broad thermal anomaly beneath the Tanzania Craton is a likely source of uplift for the East African Plateau, there is not significant seismic evidence for a similar thermal anomaly at upper mantle depths supporting the Southern African Plateau.

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Appendix A

Supplementary Figures for Chapter 2

This Appendix contains supplementary figures for Chapter 2. Each figure shows the fits for the full waveform inversion and depth phase analysis of one event. These figures show results from moment tensor inversion and depth phase modeling for five of the six events studied here. They are formatted to match Figure 2-1 from Chapter 2. A) Fit of full waveform synthetics to observed data from moment tensor inversion. Observations are shown as a solid line, while synthetics are shown as a dashed line. The bar beneath each set of waveforms indicates a scale of 100 seconds. B) The best fitting focal mechanism at each depth is shown plotted against RMS error. C) Observed depth phases are shown bracketed by synthetics for the maximum and minimum possible source depths. The bar beneath the seismograms indicates a time scale of 5 seconds. Records from NORSAR arrays are stacked to improve signal-to-noise ratios, but where GSN stations are used, records from only a single station are shown. D) Our best double couple solution and source depth are shown along with the double couple CMT solution and source depth.
Figure A-1: Event #2
Figure A-2: Event #3
Figure A-3: Event #4
Figure A-4: Event #5
Figure A-5: Event #6
Appendix B

List of Events Used for Surface Wave Tomography in Southern Africa

This appendix includes a table with details about the events used to conduct the surface wave tomography in southern Africa in Chapter 3. North and east latitudes and longitudes are positive. Locations are taken from the WHDF and PDE catalogs.
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Table B1. Event List for Surface Wave Tomography in Southern Africa (Ch. 3)
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Appendix C

Supplementary Figures for Chapter 3

This Appendix contains supplementary figures for Chapter 3. Figures C-1, C-2, C-3 and C-4 show additional figures from the two dimensional phase velocity inversion. Figure C-1 shows phase velocity maps for each period that we inverted. Figure C-2 shows the model uncertainty. Figures C-3 and C-4 show checkerboard tests with input checkers of 2 and 3 degree squares respectively on saturated scales of plus and minus five percent of the average velocity at each period. Figures C-1, C-3, and C-4 are cropped to show only those areas with a model uncertainty less than 0.06 km/s at 50 seconds. Figure C-5 shows tests of different smoothing parameters for the quasi-three dimensional shear wave model, and Figure C-6 shows slices through our shear wave model at 50 km intervals. Figure C-7 shows our reproduction of the shear wave model reported by Li and Burke (2006) using their reported phase velocities and crustal and uppermost mantle structure.
Figure C-1: Phase velocity maps for southern Africa for all periods (17-167s). (continued below)
Figure C-1: (continued) Phase velocity maps for southern Africa for all periods (17-167s). (continued below)
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Figure C-1: (continued) Phase velocity maps for southern Africa for all periods (17-167s).
Figure C-2: Model uncertainty for southern Africa for all periods (17-167s). (continued below)
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Figure C-2: (continued) Model uncertainty for southern Africa for all periods (17-167s). (continued below)
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Figure C-2: (continued) Model uncertainty for southern Africa for all periods (17-167s).
(continued below)
Figure C-3: Checkerboard resolution test for southern Africa for all periods (17-167s). Input checkers were 2 degree squares alternating plus and minus five percent of the average velocity for that period. (continued below)
Figure C-3: (continued) Checkerboard resolution test for southern Africa for all periods (17-167s). Input checkers were 2 degree squares alternating plus and minus five percent of the average velocity for that period.
Figure C-3: (continued) Checkerboard resolution test for southern Africa for all periods (17-167s). Input checkers were 2 degree squares alternating plus and minus five percent of the average velocity for that period.
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Figure C-4: Checkerboard resolution test for southern Africa for long periods (111-167s). Input checkers were 3 degree squares alternating plus and minus five percent of the average velocity for that period.
Figure C-5: Tests of smoothing parameters for the quasi-three dimensional shear wave velocity model. Slices are shown at a depth of 200 km for four smoothing lengths as indicated. Velocities shown in this figure are for a flat-earth velocity model.
Figure C-6: Depth slices thought our quasi-three dimensional shear wave velocity model at 50 km intervals.
Figure C-7: Reproduction of shear wave velocity profile using the phase velocities and crustal model reported by Li and Burke (2006). Part a) shows phase velocities reported by Li and Burke (2006) in black, and phase velocities predicted by our reproduced model in red. Part b) shows Li and Burke’s southern Kaapvaal Craton, converted to flat earth velocities (black), and our reproduced shear wave model produced by inverting their phase velocities and fixing the crust and uppermost mantle layer to the velocities and thicknesses they used in their study.
Appendix D

List of Events Used for Surface Wave Tomography in East Africa

This appendix includes a table with details about the events used to conduct the surface wave tomography in East Africa in Chapter 4. North and east latitudes and longitudes are positive. Locations are taken from the WHDF and PDE catalogs.
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Appendix E

Supplementary Figures for Chapter 4

This Appendix contains supplementary figures for Chapter 4. Figures E-1, E-2, E-3, and E-4 show additional figures from the two dimensional phase velocity inversion. Figure E-1 shows phase velocity maps for each period that we inverted. Figures E-2 shows model uncertainty. Figure E-3 shows two degree checkerboard tests for each period, and Figure E-4 shows three degree checkerboard tests for selected periods. Figure E-5 shows a resolution test of a negative five percent velocity anomaly beneath the Western and Eastern Rift Branches at a period of 50s. Figure E-6 and E-7 show depth slices at 50 km intervals and cross-sections through our shear wave model, respectively.
Figure E-1: Phase velocity maps for East Africa for all periods (20-182s). (continued below)
Figure E-1: (continued) Phase velocity maps for East Africa for all periods (20-182s). (continued below)
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Figure E-1: (continued) Phase velocity maps for East Africa for all periods (20-182s). (continued below)
Figure E-2: Model uncertainty for East Africa for all periods (20-182s). (continued below)
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Figure E-2: (continued) Model uncertainty for East Africa for all periods (20-182s).

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Figure E-2: (continued) Model uncertainty for East Africa for all periods (20-182s). (continued below)
Figure E-3: Checkerboard resolution test for East Africa for all periods (20-182s). Input checkers were 2 degree squares alternating plus and minus five percent of the average velocity for that period. (continued below)
Figure E-3: (continued) Checkerboard resolution test for East Africa for all periods (20-182s). Input checkers were 2 degree squares alternating plus and minus five percent of the average velocity for that period.
Figure E-3: (continued) Checkerboard resolution test for East Africa for all periods (20-182s). Input checkers were 2 degree squares alternating plus and minus five percent of the average velocity for that period.
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Figure E-3: (continued) Checkerboard resolution test for East Africa for all periods (20-182s). Input checkers were 2 degree squares alternating plus and minus five percent of the average velocity for that period.
Figure E-4: Checkerboard tests for periods of 143, 167, and 182 seconds. Input checkers were three degree squares alternating plus and minus five percent of the average velocity at that period.
Figure E-5: Resolution test at 50s for a 5 percent slow anomaly under the Eastern and Western Rift Branches. Rectangles indicate the approximate locations of input anomalies.
Figure E-6: Depth slices through our quasi-three dimensional shear wave velocity model at 50 km intervals.
Figure E-7: Additional cross-sections through our quasi-three dimensional shear wave velocity model. Locations of the cross-sections are shown on the 250 km depth slice (Figure E-6d).
Vita
Aubreya Nicole Adams

EDUCATION

Pennsylvania State University, anticipated Summer 2010, Ph.D. in Geosciences

* A seismic investigation of crust and upper mantle structure beneath the Zagros Mountains and the Southern and East African Plateaus

University of Florida, August 2004, B.S. in Geosciences, Summa cum Laude

RESEARCH AND INDUSTRY EXPERIENCE

Pennsylvania State University, August 2004 – present, Graduate Researcher
Chevron, Houston, Texas, May – August 2006, Seismology Intern
Shell, New Orleans, Louisiana, May – August 2005, Geophysics Intern

TEACHING AND MENTORING EXPERIENCE

Pennsylvania State University, August – December 2008, Personal Undergraduate Tutor
Pennsylvania State University and University of Witwatersrand, May – August 2008, Undergraduate Research Mentor; Classroom and Field Camp Instructor
Pennsylvania State University, January – May 2007, Teaching Assistant
Pennsylvania State University, August – December 2005, Lab Instructor
Pennsylvania State University, August – December 2004, Teaching Assistant

FIELD EXPERIENCE

Broadband Seismic Array Demobilization & Reinstallation, November - December 2008
Pennsylvania State University, Uganda and Tanzania
Broadband Seismic Array Servicing, November 2007; August – September 2007
Pennsylvania State University, Uganda
Field Camp, 2007, ExxonMobil, Monterrey, Mexico
Broadband Seismic Station Installations, January 2006, Pennsylvania State University, Zambia

HONORS AND AWARDS

Pennsylvania State University:
Department Colloquium 1st, 3rd Place Oral Presentation by Pre-Comp. Student, April 2006, 2007

University of Florida:
Danker Scholarship for Outstanding Graduating Undergraduate, July 2004
“Estwing” Field Camp Award, July 2004
President’s Honor Roll, December 2002 – May 2004
Florida Bright Futures Scholarship, August 2000 – August 2004

PUBLICATIONS


ABSTRACTS

AfricaArray Annual Meeting, May 2008