Mantle seismic structure beneath the Kaapvaal and Zimbabwe Cratons

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ABSTRACT

We present results of seismic tomography for a broad region of southern Africa using data from the seismic component of the Kaapvaal Project, a multinational, multidisciplinary experiment conducted in the late 1990s. Seismic images provide clear evidence of mantle structures that mimic the surface geology across the region and provide important constraints on subcrustal structure associated with Archean cratons. Specifically, a thick (~250 to 300km) mantle keel exists beneath the Kaapvaal craton; a slightly thinner (~225 to 250km) keel exists beneath the Zimbabwe craton and parts of the Archean Limpopo mobile belt. Mantle velocities lower than surrounding regions are evident across a broad swath beneath the surface expression of the Bushveld Complex, a ~2.05 Ga layered mafic intrusion. These reduced velocities may be due to mantle refertilization during intrusion of Bushveld magmas, or they may be caused by a thermal perturbation of more recent origin, perhaps related to the ~183 Ma Karoo magmatic event.

Introduction

Understanding the formation and evolution of Archean cratons remains one of the holy grails of earth science. The existence of these ancient continental cores indicates that cratonic formation was widespread before ~2.5 Ga, and a wide range of petrologic, geochemical, and seismological evidence suggests that these nuclei apparently formed under conditions unique to the Archean (*e.g.*, Jordan, 1988; Rudnick *et al.*, 1998; Carlson *et al.*, 2000; James and Fouch, 2002; Niu and James, 2002). It is not surprising that conditions in the Archean were somewhat different than processes operating on Earth today, but it is striking that these dissimilarities may be of a fundamental nature. Understanding the relationship of continental structures to the evolution of both the crust and mantle of continental interiors is integral to improving our understanding of cratons.

Nearly three decades ago, Jordan (1975) advocated the term "tectosphere" to describe the deep layer of the Archean mantle that remains attached to the continental craton through geologic time. By that model, extensive melt depletion of the cratonic mantle produced compositionally-induced low density refractory residues, resulting in cold cratonic roots that are isostatically balanced against warmer, yet compositionally denser, mantle beneath ocean basins. The term tectosphere is meant to distinguish this stable mantle root beneath cratons from the rest of the convecting mantle and to differentiate the Archean cratonic keel from post-Archean lithospheric mantle.

The existence of a chemically and physically distinct mantle tectosphere suggests processes or conditions unique to the Archean. Cratons are characterized by both low heat flow (*Jones*, 1988) and a very low geothermal gradient (McDonough and Rudnick, 1998; Jaupart and Mareschal, 1999; Nyblade, 1999; Rudnick and Nyblade, 1999) relative to Proterozoic mantle. Cratonic mantle xenoliths show that the tectospheric root, or "keel", is composed of highly depleted peridotite with low normative density and high seismic velocity (Boyd and McCallister, 1976; Jordan, 1979; James *et al.*, 2004). Refractory keels beneath cratons extend to mantle depths of at least 200km or deeper (*e.g.*, Jordan, 1975; Lerner-Lam and Jordan, 1987; Jaupart and Mareschal, 1999; Rudnick and Nyblade, 1999; Shapiro *et al.*, 1999; Ritsema and van Heijst, 2000; James *et al.*, 2001; James *et al.*, 2003).

Variations in crustal and crust/mantle boundary structure also support the concept of major differences in Archean and post-Archean craton formation processes. For instance, Archean crust is typically thinner than surrounding Proterozoic terranes (*e.g.*, Durrheim and Mooney, 1991; Durrheim and
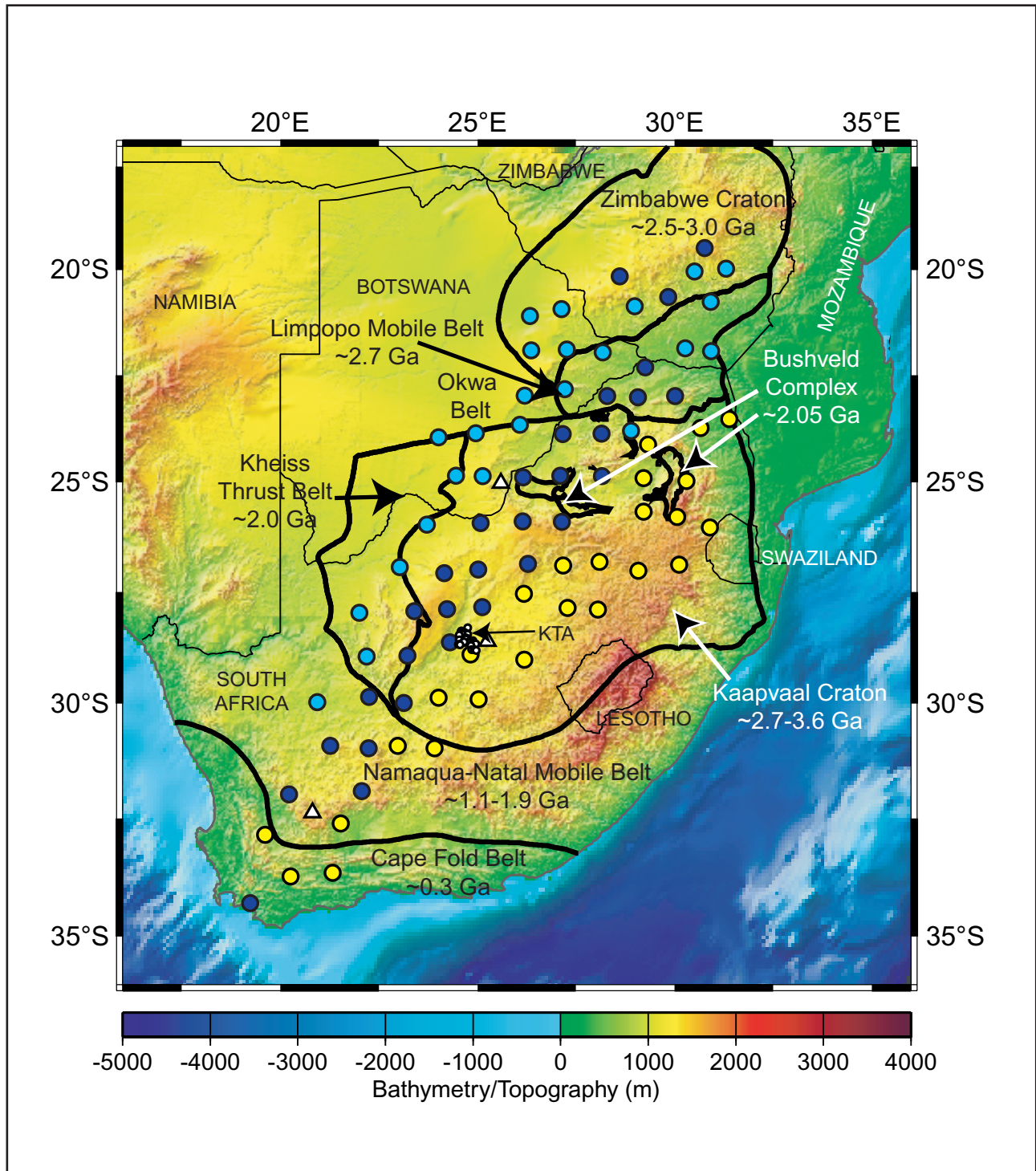
Figure 1. Map showing topography, principal geologic provinces, and station locations of the Southern Africa Seismic Array (SASE). Circles denote the fifty-five broadband seismic stations of the larger Kaapvaal Seismic Array (KSA), which were installed in April 1997 in South Africa, Botswana, and Zimbabwe. A total of 79 sites were occupied over the two-year deployment. Stations in light blue were re-deployed in April 1998 to sites indicated in yellow. Dark blue stations remained in the same location throughout the experiment period. Small open circles denote the dense Kimberley Telemetered Array (KTA), which was installed for a period of five months during the KSA deployment. Data from three GSN broadband stations located in the region (open triangles) are also incorporated in the tomographic analysis. The array extends from the Cape Fold Belt in the south, through the Proterozoic Namaqua-Natal mobile belt, across the Kaapvaal Craton and Bushveld Complex, through the Archean Limpopo Mobile Belt and into the Zimbabwe Craton. In the west, the array covers part of the Kheis and Okwa Proterozoic fold and thrust belts of Botswana and western South Africa.

Mooney, 1994; Clitheroe *et al.*, 2000; Nguuri *et al.*, 2001; Assumpçao *et al.*, 2002; Niu and James, 2002; Stankiewicz *et al.*, 2002), although there is not consensus on this issue (*e.g.* Rudnick and Fountain, 1995). Furthermore, the crust-mantle discontinuity (the "Moho") beneath Archean regions is commonly sharper than that beneath post-Archean regions (*e.g.*, Nguuri *et al.*, 2001; Assumpçao *et al.*, 2002; Niu and James, 2002). The lower crust beneath cratons may also be less mafic than beneath Proterozoic terranes (Griffin and O' Reilly, 1987; Durrheim and Mooney, 1994; Niu and James, 2002).

The southern African subcontinent provides a natural laboratory for detailed examinations of tectospheric keel structure. The region has long been the focus of intense geologic study by virtue of its economic importance. The Kaapvaal Project (Carlson *et al.*, 1996) was undertaken with the key objective of extending our understanding of the formation and evolution of cratonic crust and mantle. Data from the Project have provided important constraints on the geologic, petrologic, geochemical, and geophysical structure of the Kaapvaal craton and surrounding terranes. The centerpiece of the Project, the Southern Africa Seismic Experiment (SASE), has provided a unique opportunity to image crust and mantle structure over a wide range of spatial scales beneath southern Africa.

Results from seismic data collected during the SASE have yielded important new constraints on crust and mantle structure beneath southern Africa. Studies have included work on crustal thickness, Moho sharpness, and lower crustal composition (*e.g.*, Nguuri *et al.*, 2001; Niu and James, 2002; Stankiewicz *et al.*, 2002), upper mantle discontinuity structure (*e.g.*, Gao *et al.*, 2002; Stankiewicz *et al.*, 2002), 1D surface wave velocity and anisotropy structure (*e.g.*, Freybourger *et al.*, 2001; Saltzer, 2002), and mantle seismic anisotropy (*e.g.*, Silver *et al.*, 2001; Fouch *et al.*, 2004b; Silver *et al.*, 2004). More recent work has been undertaken to integrate these results and provide better constraints on the nature of craton formation and evolution (*e.g.*, James and Fouch, 2002; Shirey *et al.*, 2002; James *et al.*, 2003; James *et al.*, 2004). In addition, SASE data have begun to provide constraints on lower mantle seismic velocity and anisotropy (*e.g.*, Ni *et al.*, 2002; Ni and Helmberger, 2003a).

This paper provides recent results of mantle seismic velocity variations using data from the Kaapvaal Seismic Array and is an update of preliminary work published by James *et al.* (2001). We present images from three dimensional (3D) tomographic inversions for P-wave and S-wave velocity variations within the upper mantle beneath the Kaapvaal and Zimbabwe cratons and their adjacent mobile belts. We interpret the seismic results in terms of the geology, geochemistry and petrology of the crust and mantle beneath southern Africa and consider their implications for craton formation and evolution.

Data and Methods
The Southern Africa Seismic Experiment
The SASE was comprised of two passive broadband seismometer arrays deployed across southern Africa (Figure 1), and formed the seismic component of the Kaapvaal Project.

The primary component of the SASE was the Kaapvaal Seismic Array (KSA), which consisted of 55 broadband seismic stations deployed across southern Africa from April 1997 to July 1999. Approximately half the stations were re-deployed to new sites in April and May 1998 for a total of 79 stations. Average station spacing for the KSA was ~100km and had a footprint of nearly 1 million km^2 (Figure 1). Permanent broadband stations BOSA, LBTB, and SUR supplemented KSA data, providing a total of 82 stations used in the analyses presented here. Nearly all stations used in the KSA were comprised of Streckheisen STS2 broadband seismometers and RefTek high resolution 24-bit digitizer/dataloggers with GPS timing, and recorded continuously at 20 samples/s. At the time of installation, the KSA was one of the largest temporary broadband seismic networks ever deployed. The array spans geological provinces across South Africa, Zimbabwe and Botswana, and traverses geologic terranes ranging in age from Early Archean to Phanerozoic.

Relative Arrival Time Analysis
We analyzed broadband seismic waveforms to determine relative arrival times of phases P, PKPdf, S, and SKS via a multi-channel cross-correlation technique (VanDecar and Crosson, 1990). In this method, all possible waveform pairs for a given phase recorded at an array from a single event are cross-correlated to determine the best-fitting relative arrival time for each waveform. The resulting data are therefore not absolute travel times, but rather relative arrival times across the array. The advantage of this analysis method is that it enables one to quantify errors in the arrival time dataset, which is not possible with most traditional arrival time phase picking methods.

Azimuthal coverage is good for both P- and S-wave datasets (Figure 2), although the northeast and southwest quadrants (middle East/Himalayan and Andean regions, respectively) provide the majority of events for both models. The timing accuracy for most of the dataset is about 0.001 s, and all data with potential timing errors greater than 0.01s were eliminated from the analysis. Record section waveforms examples for both P and S phases are shown in Figure 3, which demonstrate the very high signal-to-noise and overall excellent quality of data used in this study. Typical relative arrival time standard deviations for our dataset are ~0.03 s for P-waves and ~0.06 s for S-waves. These low standard deviations in relative arrival time result in very small velocity perturbations, and are significantly below the resolution of the final inversions. The P-wave inversion results presented in this study are based on 8693 rays from 234 events;
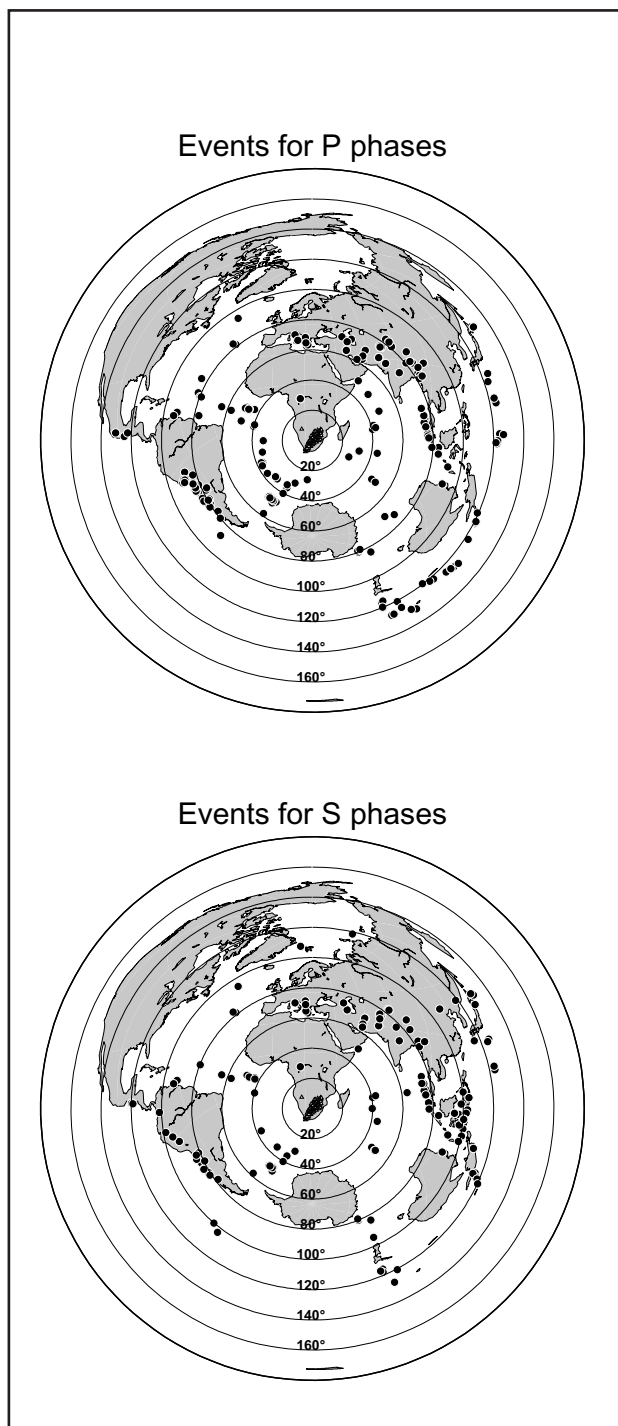
Figure 2. Location map of events (closed circles) used for tomographic inversions. Map is centered on the approximate center of the KSA. Locations are from the National Earthquake Information Center (NEIC) bulletin. A total of 234 events are used in the P-wave inversion; 152 events are used in the S-wave inversion.

the S-wave results are based on 5117 rays from 152 events.

Inversion for Seismic Velocity Structure

We used the inversion method of VanDecar (1991) to determine seismic velocity structure across the KSA.

A full description of the method, including recent modifications from the original technique developed by VanDecar (1991), can be found in Fouch *et al.* (2004a).

A significant benefit of this method is that velocity variations in these models represent the minimum variation in seismic velocities required to fit the dataset. While the results presented in this paper may therefore somewhat underestimate variations in mantle velocity structure, spatial variations are indeed real and should be viewed as the minimum seismic velocity variations across the region. We note that this inversion method cannot provide estimates of average seismic velocity at a given depth, but does provide very good estimates of lateral seismic velocity gradients across a region.

We inverted P- and S-wave relative delay times independently for structure beneath the array. In these inversions, we evaluated a range of smoothing and damping parameters, and found that for a broad range of both parameters, the strength and distribution of velocity perturbation variations did not change significantly. For both the P- and the S-wave inversions we parameterized identically the model as splines under tension constrained by a series of regularly spaced knots (Figure 4). We designed the grid based on the resolution of the data set; within the interior portion of the model, the knots are spaced 50km apart in depth and 0.5 degrees in both latitude and longitude. Knots in the outer regions of the model are spaced 100km apart in depth and 1.0 degree in both latitude and longitude for a total of 85,211 grid points in the model. We applied arrival time delay corrections at each station for both elevation differences using the GTOPO30 digital elevation model and lateral variations in crustal thickness based on receiver function results from Nguuri *et al.* (2001). In the linear inversion presented here, we used as a starting model a modified version of the IASP91 one-dimensional (1-D) standard earth model (Kennett, 1991). These modifications to IASP91 more appropriately represent cratonic upper mantle velocities determined by seismic velocity analysis of mantle xenoliths (James *et al.*, 2004). However, a starting model based on unmodified IASP91 produced nearly identical results, characteristic for this form of seismic travel time tomography. The inversion solves simultaneously for the slowness perturbation field, earthquake relocations, and station corrections. Earthquake relocation parameters absorb contributions to the delay times from Earth structure outside of the array, while station corrections (~±0.4 s for P-waves; ~±0.75 s for S-waves) absorb contributions from unknown shallow structure that cannot be resolved by this technique. The inclusion of earthquake relocations and station corrections therefore restricts the resulting models of velocity perturbations to contain the least amount of structure required to satisfy the relative delay time observations within the estimated standard error discussed in the previous section.
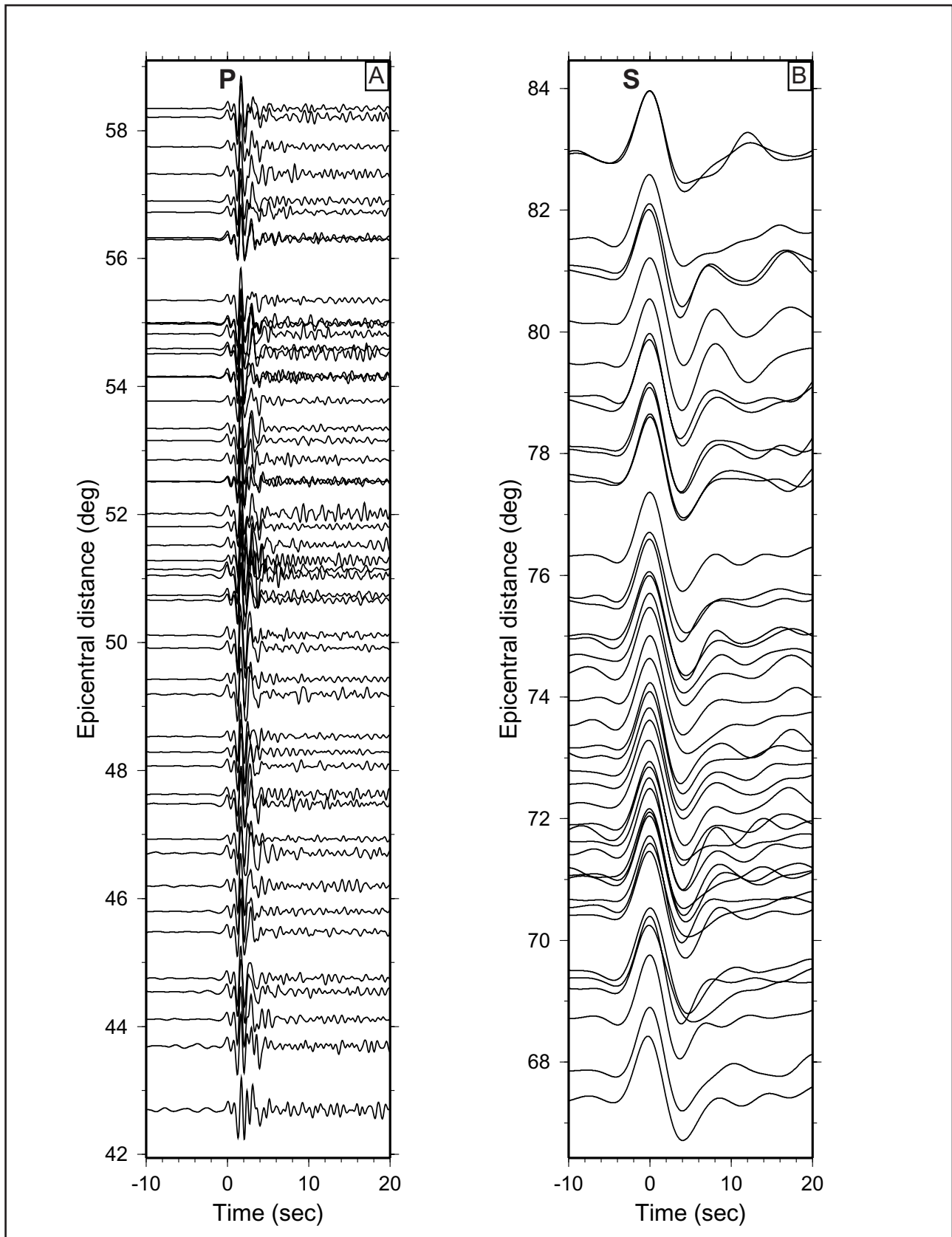
Figure 3. Record sections of typical velocity waveforms used for relative delay time analysis. Time series are aligned at 0 s on peak of phase used in analysis; amplitudes are normalized for all waveforms in record section. Typical delay time standard deviations for our dataset are ~0.03 s for P-waves and ~0.06 s for S-waves. (**a**) Vertical component waveforms from a South Sandwich Islands event (October 5, 1997; depth 274km; m_b 6.0) used for P-wave analysis. Waveforms have been bandpass filtered between 0.4 Hz and 2.0 Hz. (**b**) Transverse component waveforms from a Hindu Kush event in South Sandwich Islands (May 13, 1997; depth 196km; m_b 6.1) used for S-wave analysis. Waveforms have been bandpass filtered between 0.02 Hz and 0.2 Hz.
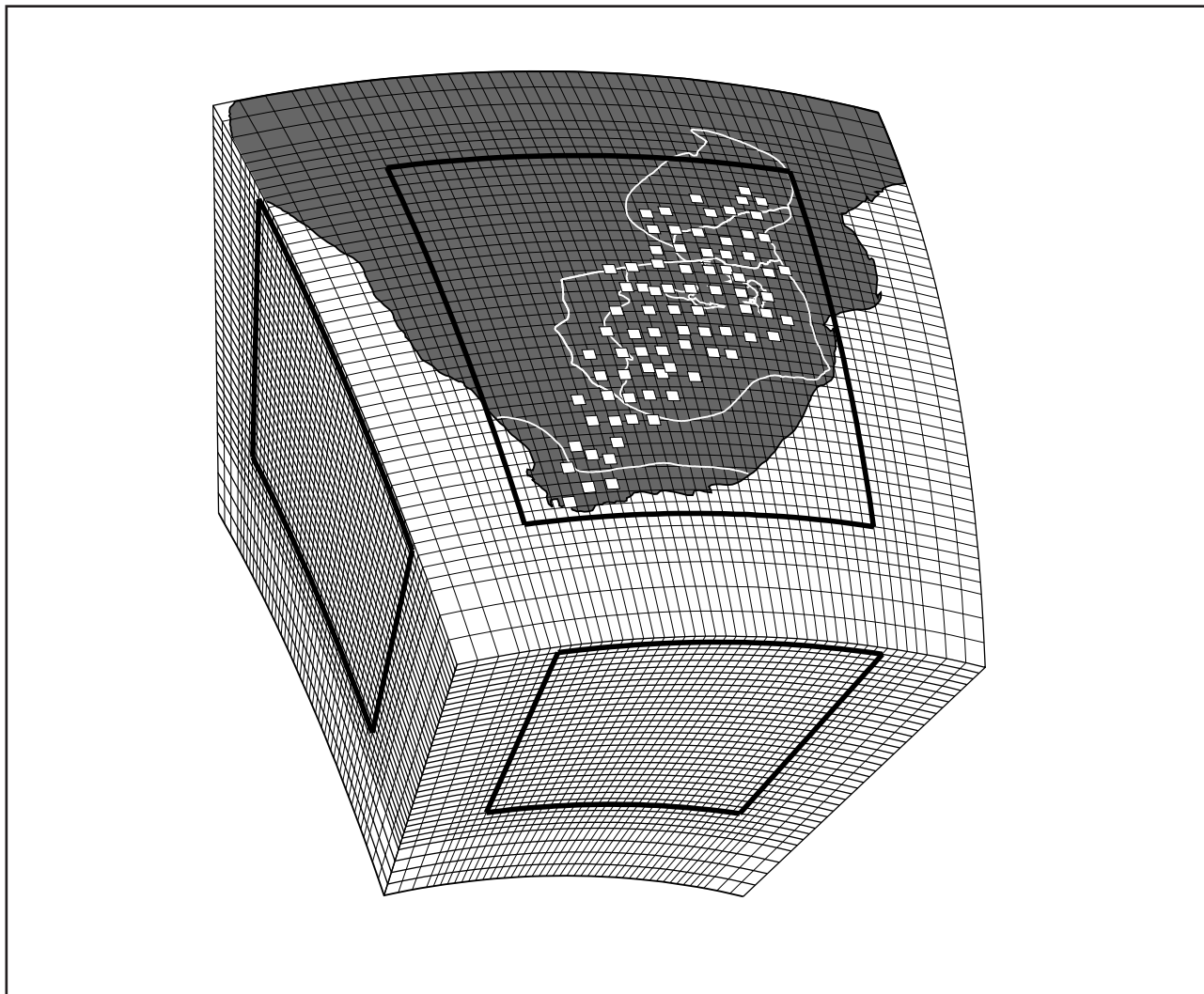
Figure 4. Three-dimensional perspective view of grid knots for splines under tension that constrain the velocity perturbation model determined in this study. Knot positions occur at thin black line crossings. Thick black lines indicate regions to which velocity maps and cross sections are confined. Open squares denote seismic stations; white lines denote geologic terrane boundaries labeled in Figure 1.

Resolution Tests

We performed a number of simple resolution tests with hypothetical structures that approximate the imaged structures to assess the seismic velocity models presented in this study. We emphasize that the tomographic images presented in this paper were determined using linear inversion and are therefore preliminary. Results of initial and rather crude resolution tests for the linear inversion are included in the electronic supplement to James et al. (2001) and can also be found at http://www.ciw.edu/kaapvaal/pubs/tomography/supplemental.pdf. Non-linear inversion results and a more detailed analysis of resolution tests are presented in a seperate paper (Fouch et al., 2004a). The preliminary resolution tests in James et al. (2001) show that the lateral extent of the cratons is very well recovered, and while downward smearing of structure occurs, the effect does not preclude good estimates of keel depth. Best resolution, based on the density of crossing ray paths, is achieved in the depth range ~150 to 500km. Shallow structures above ~80km are sampled by relatively few crossing paths from teleseismic events and are thus poorly resolved. Away from the edges of the model, we estimate resolution of the vertical extent of cratonic roots to within ~50km. The lateral resolution of velocity variations is substantially better; for most regions of the model, we estimate this value to be ~25km. Additionally, the P-wave images contain somewhat better resolution than the S-wave images due to a larger data set and more crossing rays. Both models, however, are spatially well-resolved with the exceptions noted above. The strength of velocity perturbations recovered in the resolution tests are reduced somewhat from the input model; this effect increases with depth in the keel. The strongest recovered velocities are close to the input model variations, suggesting that the largest observed velocity perturbations likely are only slightly underestimated from variations found in the real earth using this analysis method.
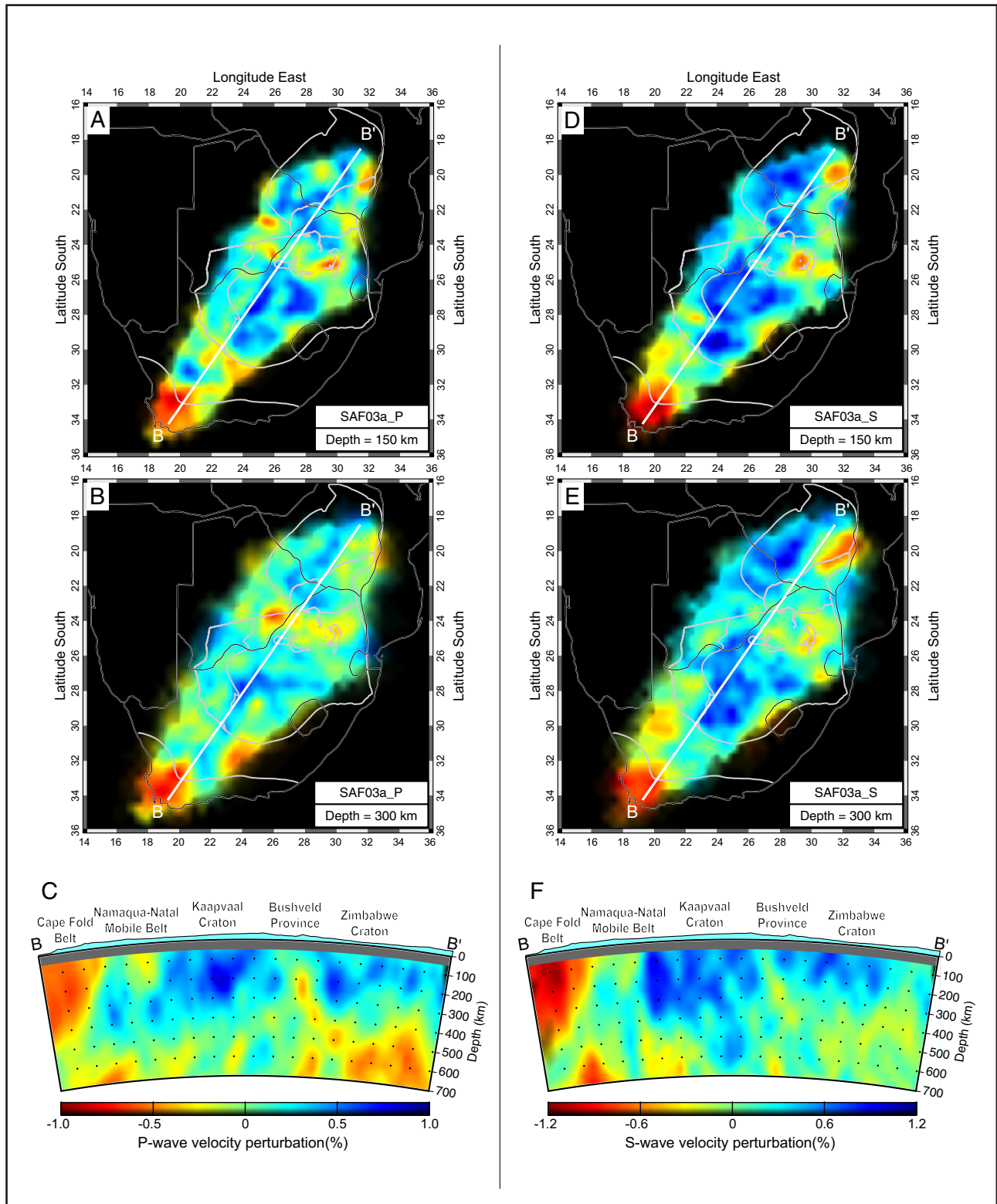
Figure 5. Horizontal and vertical cross sections for P- and S-wave velocity perturbation models determined by inversion of delay times after correction for elevation and crustal thickness. Color scales show percent velocity perturbation (note slightly different range for P- and S-models). Surface topography (light blue) is plotted at 20 times actual scale. Uppermost 50km in vertical sections (shaded area) denotes regions where station delay time residuals are incorporated in model calculations. Colors fade to black for ray hit counts less than 10 within the model to represent the most poorly-resolved regions of the model. Panels a-c are images of P-wave velocity perturbations; panels d-f are images of S-wave velocity perturbations. Panels a and b are P-wave depth slices at 150 and 300km, respectively; panels d and e are S-wave depth slices at 150 and 300km, respectively; panels c and f are vertical cross section views of P- and S-wave velocity perturbations, respectively, across B-B' profile shown in map view panels.

Mantle Seismic Structure Beneath Southern Africa

Results for the linear, smoothed, and damped relative travel time inversions for both P- and S-waves exhibit a range of trends that clearly relate to geologic structure at the surface (Figure 5). For the most part, the P- and S wave models are quite similar, and suggest that the velocity inversions are robust. These variations provide key insights into the relationship between crustal geology and deeper mantle structure beneath southern Africa.

The most robust feature in the tomographic images is that regions of highest velocity mantle coincide closely with the boundaries of both the Archean Kaapvaal and Zimbabwe cratons. The region of maximum positive velocity perturbations (blue regions, Figure 5) outlines the present-day core of the Kaapvaal craton, from the southernmost Proterozoic Bushveld Complex south-southwest to the inferred contact with the Proterozoic Namaqua-Natal mobile belt in the southwestern region of the array. North of the Bushveld Complex, similar mantle structure exists across both the Archean Limpopo mobile belt and Zimbabwe craton. The strong positive velocity perturbations across the region suggest that the seismically-defined keel extends to depths of ~250 to 300km beneath the Kaapvaal craton, and perhaps slightly less (~225 to 250km) beneath the Limpopo mobile belt and Zimbabwe craton.

A significant break in the highest velocity regions is evident near the ~2.05 Ga Bushveld Complex (Walraven and Hattingh, 1993; Buick et al., 2001) (Figure 5). Lower than average mantle velocities near the Bushveld Complex appear to extend not only into the mantle beneath the layered intrusion itself, but also well to the west, particularly in the higher-resolution P-wave models. While the overall pattern of reduced velocities across the Bushveld Complex is well-resolved, the localized higher order variations of the P-wave perturbations (i.e., structures <25km in lateral dimension) are not well-resolved.

Another significant velocity anomaly exists in the uppermost mantle near the Proterozoic Namaqua-Natal mobile belt south and west of the Kaapvaal craton. This mobile belt is thought to be the remnants of a major north to south convergent margin that extended as far north as the Zimbabwe craton (de Wit et al., 1992), and is characterized by velocity perturbations slightly lower than those observed beneath cratonic regions. We observe patches of higher velocity material in the 200 to 400km depth range beneath this region, however, and these higher velocities appear to exhibit continuity with the high velocity material beneath the adjacent Kaapvaal craton. Finally, the reduced velocity perturbations beneath the Cape fold belt are likely due to poor resolution in this region and possibly lower mantle low velocity anomalies (i.e., Ritsema et al., 1998; Ni and Helmberger, 2003b). We therefore do not discuss further the Cape Fold belt anomalies in our analysis.

Discussion

The depth extent of cratonic roots has long been a contentious issue. Part of the debate has been with regard to Jordan's tectosphere hypothesis (Jordan, 1975; 1978), in which the concept of deep, cold, and chemically distinct keels beneath continents was contested when first proposed (e.g. Okal and Anderson, 1975; Anderson, 1979; Sclater et al., 1980). More recently, however, the tectosphere hypothesis has been strengthened substantially by petrologic and geochemical studies of mantle nodules, particularly data from southern Africa. For instance, Re-Os model age determinations show that mantle nodules erupted from even the greatest depths beneath the craton (~200km) are Archean in age (Pearson et al., 1995; Carlson et al., 2000). Moreover, recent analyses of xenolith pressure and temperature data suggest that the craton geotherm likely intersects the mantle adiabat between depths of 160km and 300km. Current "best" estimates place the mantle adiabat intersection in the range of 220 to 250km beneath the cratons if the Brey and Kohler (1990) geothermobarometry methodology is assumed (Rudnick et al., 1998; Rudnick and Nyblade, 1999; James et al., 2004). Other thermobarometric methods, however, may produce lower geotherms that imply greater depths to the mantle adiabat intersection (James et al., 2004).

The keel thickness estimates from petrologic and geochemical data are consistent with the results of this study, which show that positive velocity perturbations beneath the cratons extend to depths of at least 250km, and possibly as deep as 300km. With the exception of the reduced velocity region near the Bushveld Complex, the mantle beneath the Zimbabwe and Kaapvaal cratons, and the Limpopo mobile belt exhibits a strikingly coherent pattern of high velocity anomalies to depths of at least 250km. These results are in general agreement with regional seismic studies of several other cratonic regions (e.g., VanDecar et al., 1995; van der Lee and Nolet, 1997; Ritsema and van Heijst, 2000; Goes and van der Lee, 2002), which suggest keel thicknesses as great as 300km in some areas. The high velocity perturbations that extend below 300km in the southern Kaapvaal could be real, although resolution tests suggest that these deeper high velocity anomalies may be due to vertical smearing of velocities in the inversion process (Fouch et al., 2004a). Additionally, independent evidence from heat flow studies does not suggest such great keel thicknesses, although such estimates are highly model dependent (Jones, 1988; Rudnick and Nyblade, 1999).

The velocity perturbations observed in this study are consistent with those computed from recent surface wave studies. For example, a 1-D isotropic regional study by (Zhao et al., 1999) found a maximum of a ~3% increase in velocities across the southern Africa region. Similarly, a 3-D, isotropic, regional study of fundamental mode Rayleigh waves by Ritsema and van Heijst (2000) found similar degrees of velocity perturbations, and a maximum keel thickness of ~250km. A pair of 1-D

studies using the KSA dataset incorporated anisotropy into the analyses and found a maximum keel thickness of ~220km (Freybourger *et al.*, 2001; Saltzer, 2002). While these studies do not agree on the location and strength of the lithospheric anisotropic zone, they clearly show higher than average velocities extending to depths >200km.

We find no evidence for a low seismic velocity layer beneath subregions of the southern African keel system. If a ubiquitous, homogeneous low-velocity zone (LVZ) existed beneath the entire region, we would not be able to resolve it with our tomographic analysis method. Given the complexity in keel structure observed in this study, a simple homogeneous zone of reduced velocities across the region is not likely; we therefore should be able to image areas containing LVZ structure if present. The absence of an LVZ is consistent with several recent surface wave studies of the region that possess good vertical resolution (Ritsema and van Heijst, 2000; Freybourger *et al.*, 2001; Saltzer, 2002; Gore, 2004; Nguuri, 2004), but it is not consistent with previous 1-D isotropic studies (Cichowicz and Green, 1992; Qiu *et al.*, 1996; Priestley, 1999). The results of Freybourger *et al.* (2001) and Saltzer (2002), however, used KSA data to demonstrate that a low-velocity zone beneath the high velocity keel may be artificially derived if anisotropy is not included in the 1-D calculation. While some thermal blanketing may influence velocities immediately beneath the craton, the velocity perturbations we find in the 200 to 300km depth range are too high to reflect "cool" asthenospheric mantle. Even to depths of ~400km, the velocity perturbations we observe are too high for suboceanic mantle velocities.

The most prominent upper mantle velocity anomaly within the Kaapvaal craton is associated with the Bushveld layered intrusion and its western extension into Botswana. The tomographic results are also consistent with evidence from geological and geophysical data that Bushveld Complex structures extend westward into Botswana (Ranganai *et al.*, 2002). While the zone of reduced mantle velocities in the region is clearly real, resolution tests reveal difficulties in quantifying the magnitude of the velocity contrast and suggest that this contrast may be somewhat smaller than that obtained from the data inversion (Fouch *et al.*, 2004a). Assuming the observed velocity reductions of ~0.5% in P and ~0.8% in S, elementary calculations from thermal/chemical/velocity relationships (Christensen, 1982) suggest that a temperature anomaly of ~100ºC would be required to produce the velocity reduction observed.

The reduced mantle velocities near the region of the Bushveld Complex suggest that considerable mantle modification has occurred in the region, probably at the time of intrusion of Bushveld magmas. One explanation of relatively low velocities beneath the Bushveld Complex is compositional "refertilization" (*i.e.*, iron enrichment) of the mantle during the Bushveld event.

Indeed, it is plausible that pervasive mantle metasomatism, as may have accompanied the Bushveld magmatism, could significantly reduce seismic velocities in the mantle. Jordan showed, for example, that refertilized cratonic samples have normative seismic velocities up to 1% lower and densities up to 2 to 3% higher than the depleted nodular peridotites (Jordan, 1979). Moreover, increased proportions of orthopyroxene can occur by reaction of metasomatic fluids with olivine, providing an attractive explanation for the substantially larger P-wave anomaly relative to the S-wave anomaly in the region (James *et al.*, 2004). A similar origin to the pattern of large positive P-wave anomalies relative to S has been observed for regions of the Wyoming craton in the western United States (Goes and van der Lee, 2002).

A temperature anomaly on the scale of ~100ºC, however, is impossible to verify with current data. While the mantle xenoliths from the Premier mine on the edge of the Bushveld Complex do exhibit very slightly higher geotherms than those from the undisturbed craton to the south (James *et al.*, 2004), the fact that the xenoliths were erupted ~1.2 Ga renders them meaningless for estimating present-day geotherms. Heat flow measurements (Jones, 1988) from the Bushveld region seem to indicate, however, that geotherms in the region of the Bushveld Complex are only marginally higher, if at all, relative to the surrounding region. What is clear is that the Bushveld Complex was a major mantle event as evidenced by recent data from Re-Os isotope analysis, which show that ~1.2 Ga mantle samples from the Bushveld province were heavily overprinted and reset with Proterozoic Bushveld Complex ages of ~2.05 Ga (Carlson *et al.*, 2000). The isotopic resetting of an entire volume of Archean Kaapvaal mantle almost certainly required material addition to the mantle (Carlson *et al.*, 2000; Shirey *et al.*, 2002).

Other interpretations of this enigmatic region are also plausible. For example, analyses of Premier mine xenolith samples notably indicate that neither peridotite major element composition nor orthopyroxene abundance differ significantly from xenoliths erupted through undisturbed Kaapvaal craton (F.R. Boyd, personal communication 2001). Conversely, eclogitic materials, if present (Shirey *et al.*, 2002), would tend to increase both average velocity and average density of the depleted peridotitic mantle and therefore cannot be an explanation for reduced seismic velocities. Finally, Duncan (A. Duncan, personal communiucation 2002) has suggested that the Bushveld anomaly observed in this study may be due to a thermal perturbation resulting from Karoo magmatism at ~183 Ma (Duncan *et al.*, 1997). Further work will be required to assess the validity of these various competing hypotheses.

Data provided by the SASE continue to be an important source for a broad range of studies currently in progress, which will provide further information regarding the seismic structure of the Kaapvaal region.

For instance, preliminary 2-station surface wave analyses have been performed by (Gore, 2004) and (Nguuri, 2004). (Larson *et al*., 2003) currently are extending this work, which will provide high-resolution estimates of lateral variations in crust and uppermost mantle velocity structure across the entire southern Africa region. In other related efforts, body wave relative delay times collected from this study are being combined with surface wave phase delay data to invert for mantle seismic velocities (van der Lee *et al*., 2001). These studies will provide better constraints on the depth extent of the cratonic keel across the region. In addition, Li (A. Li, pers. comm. 2003) is utilizing both fundamental mode Love and Rayleigh waves recorded by the SASE to constrain mantle anisotropy and velocity structure beneath the Kaapvaal region.

Conclusions

Tomographic images of seismic body wave perturbations beneath southern Africa exhibit a clear correspondence with geologic terrane boundaries. The mantle keel beneath the Kaapvaal craton appears to extend to a depth of ~250 to 300km. The maximum keel thickness beneath the Zimbabwe craton is somewhat less, reaching a local maximum of ~225 to 250km. The total contrast in velocity perturbation within both cratons is between 0.4% and 0.6%, consistent with (but not necessarily due to) compositional variations observed in mantle nodules. The Archean Limpopo mobile belt appears to be underlain by a mantle keel with a typical cratonic signature, despite a highly complex crustal structure. The mantle beneath the ~2.05 Ga Bushveld Complex exhibits anomalously reduced velocities, which may be due either to compositional variations resulting from mantle fertilization during Bushveld magmatism or to thermal variations from more recent magmatic events associated with 183 Ma Karoo volcanism.

Acknowledgments

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