Surface wave tomography: Imaging of the lithosphere–asthenosphere boundary beneath Central and Southern Africa?

Article in Lithos · November 2010
DOI: 10.1016/j.lithos.2010.05.011

1 author:

Stewart Fishwick
University of Leicester

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Surface wave tomography: Imaging of the lithosphere–asthenosphere boundary beneath central and southern Africa?

S. Fishwick

Department of Geology, University of Leicester, University Road, Leicester, LE1 7RH, United Kingdom

ARTICLE INFO

Article history:
Received 5 July 2009
Accepted 20 May 2010
Available online 8 June 2010

Keywords:
Southern Africa
Lithosphere
Asthenosphere
Surface wave tomography
Seismic discontinuities
Kimberlites

ABSTRACT

The lithosphere–asthenosphere boundary (LAB) remains a controversial subject in Earth sciences, and beneath cratonic regions appears to be a particularly difficult boundary to consistently image. Seismic methods give different indicators on the velocity structure of the upper mantle: tomographic models provide estimates of the velocity variations at a variety of lateral scales, but have limited vertical resolution; receiver function techniques provide good indication of the depth to seismic discontinuities, but less information on the absolute velocities. This study assesses whether the different methods give consistent estimates for the depth of the LAB in southern Africa.

Using a surface wave dataset with nearly 12,000 paths in the African region, new tomographic models of central and southern Africa are calculated. To show the non-unique nature of tomography, results are presented for two different parameterisations. The models indicate varying velocity structure beneath the cratonic regions of central and southern Africa, which yield estimates of the LAB depth from around 150 km depth in Tanzania, to approximately 200 km depth beneath the Kalahari Craton, down to depths of 225–250 km beneath parts of the Congo.

At a broad-scale these depth estimates are compatible with geothermometry from kimberlite xenoliths. In regions such as Tanzania, the kimberlite magmatism is observed to occur along strong horizontal gradients in upper mantle seismic velocity structure — potentially edge features in lithospheric structure. In contrast, a detailed comparison beneath the Kalahari Craton indicates that in this region, and given the present resolution of the tomography, there is no clear link between the kimberlites and velocity gradients. However, in general the kimberlites do not sample the regions of fastest seismic velocities.

The relationship between LAB depth estimates from the tomographic modelling and those estimates from receiver functions is not clear. Results from receiver function techniques beneath southern Africa tend to place discontinuities at either shallower (100–150 km), or deeper (300–350 km), depths than the thermally defined LAB estimates. As such, particular care should be taken in automatically associating a discontinuity from fast to slow seismic velocities as the same location as a thermally defined lithosphere–asthenosphere boundary.

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1. Introduction

Interpreting the nature and depth of the lithosphere–asthenosphere boundary (LAB) remains a somewhat controversial subject within Earth sciences. Early definitions of the lithosphere (e.g. Barrell, 1914) were based on the mechanical strength of the outer layer of the Earth. However, in modern usage the term lithosphere has evolved such that it can be used to describe the tectonic plate, or in relation to a thermal boundary layer, chemical boundary layer or effective elastic thickness (for further discussion see e.g. Anderson, 1995; Artemieva, 2009; Eaton et al., 2009). Within this paper the underlying idea is to look at the LAB within the plate tectonic framework (e.g. Eaton et al., 2009) rather than in the original terms of the strong layer.

Seismically the observations of a low velocity zone (LVZ) in the shallow upper mantle (e.g. Gutenberg, 1948) have often been equated with the asthenosphere, and for many studies the seismically simple parametrisation of a high velocity lid over a low velocity zone have been used to model the velocity structure of the upper mantle (e.g. Brune and Dorman, 1963); the transition from high velocity lid to low velocities marking the LAB. While the low velocity zone has been explained by partial melt (e.g. Anderson and Sammis, 1970), recently authors have shown that in many regions the reduced velocities can be explained simply by the variation in temperature with depth (Faul and Jackson, 2005; Stixrude and Lithgow-Bertelloni, 2005). Within the lithospheric mantle velocities will decrease due to the high temperature gradient, whilst once within the adiabatic regime the effects of pressure will begin to dominate and velocities increase.

With the onset of seismic tomography, where the velocity models are obtained through 2D or 3D inversions and do not necessarily have
a distinct step in seismic velocity, a variety of techniques have been used to estimate the depth to the lithosphere–asthenosphere boundary. Amongst the different methods used are: the depth to the maximum velocity gradient (e.g. van der Lee, 2002), the depth to a certain velocity anomaly above a global reference model (e.g. Simons and van der Hilst, 2002) and changes in the orientation of seismic anisotropy (e.g. Plomerová et al., 2002).

Recently it has become more common to attempt to directly convert the seismic velocities into physical parameters such as temperature (e.g. Goes et al., 2000; Priestley and McKenzie, 2006). Estimates of lithospheric thickness (in terms of the thermal boundary layer) can then be made directly, without the need of using the proxies such as velocity perturbation or gradient. During the same time period receiver function methods to accurately estimate the depth to seismic discontinuities have become techniques frequently used, and thus give alternative LAB depths in-line with the earlier high velocity lid–low velocity zone approach.

This paper will look at the comparison of LAB estimates for central, and particularly, southern Africa. A new tomographic model is presented using surface wave data for over 11,900 source–receiver paths in the African continent and adjacent plate margins. The tomographic model is then converted to lithospheric thickness using the empirical parameterisation of Priestley and McKenzie (2006). The thickness model is discussed with regard to previous surface wave studies and the location of kimberlites. The results are then compared to recent studies of seismic discontinuities in the upper mantle beneath the Kaapvaal Craton. Do tomography and receiver functions provide us with one consistent, and physically meaningful, model of the lithosphere–asthenosphere boundary?

2. Methodology

2.1. Data

The data for the surface wave tomography study come from a mixture of permanent seismic stations located on the African continent, surrounding ocean islands, Arabia, southern Europe, and across to eastern Turkey. The onset of the Africa Array project (Nybøl et al., 2008) is increasing the station distribution within the continent, and data from 2006 for three of the Africa Array stations has been included. However, the addition of data from a number of temporary experiments is crucial in improving the path coverage across the continent. Data have, therefore, been used from seismic deployments in Tanzania (Nybøl et al., 1996), Kaapvaal Craton — southern Africa (Carlson et al., 1996), Ethiopia–Afar (Stuart et al., 2002), Ethiopia–Kenya (Nybøl and Langston, 2002), Saudi–Arabia (Vernon et al., 1996), eastern Turkey (Sandvol et al., 2003), Azores and MIDSEA (van der Lee et al., 2001), Cape Verde (Lodge and Helffrich, 2006), Seychelles (Collier et al., 2004), Cameroon (Tibi et al., 2005), Dhofar (Tibi et al., 2007) and Namibia (GEOFON–Potsdam) (Fig. 1).

The seismic sources used are mainly restricted to the region surrounding Africa: the mid-ocean ridge system to the east, south and west; the collisional system within the Mediterranean; plate boundaries in and around Arabia to the north-east; some of the larger events within the African continent also provide data towards the model (Fig. 1). The main advantage of using this dataset is that all source–receiver pairs are within the area of interest such that the results of the tomographic inversion are not affected by potential of smearing of velocity anomalies towards either events or stations outside of the region.

2.2. Surface wave tomography

The procedure used to generate the tomographic images is a series of stages. Initially a waveform selection process (Debayle, 1999) is undertaken to ensure that only high quality waveforms are submitted to the inversion. The seismogram is assessed using estimates of the signal-to-noise ratio at the periods of interest (50–120 s), and events must be at a sufficient distance (>1000 km) from the receiver to avoid complications from the source region.

A slightly modified version of the automated procedure of Debayle (1999), based on the waveform inversion scheme of Cara and Lévêque (1987) is used to calculate a 1D average model of the shear wavespeed for the path between the source and receiver. In this study, the model is constrained by information from the surface waveform at periods between 50 and 120 s for the fundamental and first four higher modes. Higher modes are useful in surface wave inversions as they have a different depth sensitivity to the fundamental mode. While the amplitude of all surface waves decay exponentially with depth, the higher modes sample increasingly deeper parts of the Earth (see e.g. Debayle et al., 2001). The inclusion of higher mode information therefore has the potential to significantly improve the vertical resolution of a surface wave study, however, the higher mode data are very dependent on the events that are used. For the African region, where many of the events are from small, shallow events associated with the mid-ocean ridge system there is unfortunately less higher mode information available than in a region surrounded by subduction zones.

As there is a non-linear dependence between the model parameters and the data the waveform inversion will only work with a limited perturbation from the starting model. While the scheme of Cara and Lévêque (1987) uses secondary observables of the waveform to increase the domain of quasi-linearity, the choice of starting model will still affect the final 1D model. For the mantle structure a 1D global Earth model, such as PREM (Dziewonski and Anderson, 1981), is commonly used. However, for the region used in this study, where some paths will be purely oceanic and some paths predominately continental (or even cratonic), this will not be ideal. One limitation of using a 1D reference model is that the velocity variations in the final model may be underestimated. Damping towards an average velocity that is neither as slow as the oceanic velocities or as fast as cratonic regions, will naturally limit the range of observed velocities. In order
to improve this stage of the inversion procedure, rather than using a 1D reference model, the large scale structures are incorporated by using a subset of the data and initially calculating a preliminary tomographic model to indicate the long wavelength (>1000 km) velocity variations. Path specific starting models can be created by computing the path-average velocity through this background model. An alternative method would be to incorporate the variations from a smoothed version of a global tomographic model.

Multiple waveform inversions are then used in order to improve the reliability of the final set of path-average models. For each seismogram the waveform inversion is run on four occasions with a different starting model, and at least three out of the four final models must show similar characteristics for the path to be used. Within the uppermost mantle the starting model varies between ±3% from the path-average structure defined through the preliminary tomographic model. The crustal model also incorporates lateral heterogeneity and is defined as the path-average structure through the model 3SMAC (Nataf and Ricard, 1996), and remains fixed within the inversion. The resulting 1D models are defined at 25 km depth intervals from below 50 km depth. Fig. 2 illustrates the results of this method for an event occurring at the Mid-Atlantic Ridge recorded at the station CM23, part of the array in Cameroon (Tibi et al., 2005). In this example the starting models all have a weak low velocity zone which becomes more pronounced in the final models. Above 150 km there is very good consistency in the final models despite the large variations within the starting models. Below this depth the final models begin to diverge suggesting that the resolution is decreasing. However, in the full depth range shown the spread of velocities in the final models remains smaller than the starting variation, indicating that there is still some resolution down to depths of at least 350 km. Both the consistency between the resulting models (Fig. 2) and the a posteriori error estimates from the waveform inversion give information on the reliability of the wavespeed profile with depth.

Before discussing the tomographic models in detail it is necessary to have some illustration on the reliability of the images. The nature of the parameterisation clearly impacts the potential resolution of the tomography. Very sharp lateral transitions in wavespeed will be smoothed such that the strongest velocity gradient occurs over the same lateral distance as the knot point spacing, and small horizontal gradients will be observed over a wider region (Fishwick et al., 2008). For isotropic tomography the inversion is performed in two steps (see Fishwick et al., 2005). Initially a large scale model is calculated using knot points at 8° intervals. This new model is then used as the reference for a final inversion with smaller spacing between knot points. The advantage of this two-step approach is that in the final stage we damp towards the large scale features within the data rather than a reference model that may not be representative of the actual seismic velocities.

3. Tomographic models

3.1. Reliability

The path-specific 1D model used in the tomographic inversion is a weighted average of the similar final models accepted by the waveform inversion scheme. The final set of path-average 1D models is used within a tomographic inversion to derive a model of the velocity structure. In this work over 11,900 1D models have been defined, giving excellent path inversion to derive a model of the velocity structure. In regions with dense networks of temporary seismometers, there is the potential for high resolution studies.

The series of 2D slices are calculated at 25 km intervals in depth using the inversion scheme of Yoshizawa and Kennett (2004). The reconstruction of each depth slice within the model is achieved by a linear inverse problem relating the velocity perturbation at a particular point to the path information at the same depth. The velocity distribution is expanded as a set of spherical B-splines, thus incorporating Gaussian type smoothing around the knot points. The inversion is solved using a damped least squares inversion scheme. The damping controls the allowed variation from an a priori reference model, strongly damped models will be forced to stay close to the reference model, while little or no damping allows the data misfit to be minimised. The appropriate damping is chosen independently for each tomographic inversion through the comparison of the data misfit and model variance.

The data are weighted based on our estimate of the reliability of the 1D model, using information from both the a posteriori error estimate of the waveform inversion and the consistency of the models produced from the multiple starting models (Fig. 2). At this point it would also be possible to introduce a weighting to normalise the contribution due to differing numbers of paths in different regions. In this study this additional normalisation is not introduced, data are simply weighted based on the path-specific information. Testing suggests that for isotropic inversions this choice is reasonable, although a balanced path coverage may be more important when incorporating azimuthal anisotropy.

For isotropic tomography the inversion is performed in two steps (see Fishwick et al., 2005). Initially a large scale model is calculated using knot points at 8° intervals. This new model is then used as the reference for a final inversion with smaller spacing between knot points. The advantage of this two-step approach is that in the final stage we damp towards the large scale features within the data rather than a reference model that may not be representative of the actual seismic velocities.

Fig. 2. Results from the waveform inversion procedure for an event occurring on the Mid-Atlantic Ridge recorded at the station CM23, part of the temporary array in Cameroon (Tibi et al., 2005). The light grey curves show the four different upper mantle starting models used for the inversion procedure, the black curves illustrate the final path-average models generated by within the inversion. The crustal model remains fixed to 3SMAC (Nataf and Ricard, 1996) for all inversions.
Waveform inversion stage is simulated by adding noise relative to the estimates of uncertainty resulting from the actual inversion of the real data. The general checkerboard pattern is recovered throughout Africa and much of the surrounding ocean (Fig. 3). The amplitude of the anomalies is generally underestimated due to the noise added to the data and the regularisation of the inversion scheme. At shallower depths the recovery of structure and true amplitudes would be improved, due to the smaller uncertainties on the path-average models for the real data. In contrast, at greater depths the uncertainties in the final model will increase due to the larger uncertainties observed in the real data at these greater depths (Fig. 2).

This style of test only assesses the inversion for a particular depth slice — the checkerboard structure is only in 2D. The true vertical resolution is hard to estimate in the two-stage approach used in this surface wave tomography. As the depth slices are independently computed it is not possible to calculate the exact interdependence between the different layers (see also Simons et al., 2002). However, more complete synthetic tests for similar studies (Priestley et al., 2008) have indicated that with the inclusion of the higher mode information the resolution will be in the order of 25–50 km for depths shallower than around 250 km. Ideally, a full test including the computation of synthetic seismograms using the spectral-element method would provide an improved understanding of the resolution and limitations of the method. A global dataset has been created for benchmarking purposes (Qin et al., 2006), and was used in the recent upper mantle study by Lebedev and van der Hilst (2008). This benchmarking test will not illustrate the resolution for a regional study with a very high density of path coverage. Adding noise to simulate the uncertainties of the waveform inversion is, therefore, a reasonable alternative.

Uncertainties in the absolute velocities are even harder to quantify. The difficulty in calculating the interdependence between the different depth slices means that any estimates made by varying the damping, or working with subsets of the data, are only an estimate for the particular vertical parameterisation that is employed. For the present dataset and style of parameterisation, reasonable choices in damping only vary the absolute velocities by about ±1%, or about ±0.05 km s\(^{-1}\). This will be an underestimate of the true uncertainty in absolute velocity.

3.2. Results

Fig. 4 illustrates the results from the tomographic inversion scheme at 100, 175 and 250 km depth. The models are plotted as perturbations from the global reference model AK135 (Kennett, 1995), for all the images the colour scale is the same with saturation at perturbations greater than ±7.5%. Contour intervals are plotted at 2% intervals. Two different final parameterisations have been used: the whole African model is shown with knot points at spacing of 3° (Fig. 4a) and a second higher resolution model is shown for central and southern Africa with knot points at 1.5° intervals. The higher resolution model has the potential to illustrate more detailed variations in wavespeed, but implicit in the goal of higher resolution is a greater uncertainty in the reliability of the model. Showing two different interpretations (choices in parameterisation) of the data illustrates the non-uniqueness of tomographic methods.

At 100 km depth much of southern Africa shows fast velocity perturbations from the reference model, typically associated with cold lithosphere. Along the western margin of the continent the velocities vary from fast (offshore Congo), to slow (central Angola), to fast (southern Angola, northern Namibia) and back to slow (central-southern Namibia) (Fig. 4a). These variations correlate with long wavelength gravity anomalies, and in Angola have been linked to dynamic support for the recent uplift of the Bie Dome (Al-Hajri et al., 2009). In the east African rift, the improved resolution of the fine parameterisation can be observed. A narrow zone of low velocities associated with the eastern branch of the rift system in Kenya is clearly visible (Fig. 4b). The southern end of these low velocities is observed in the array study of Weeraratne et al. (2003), however the less detailed parameterisation (Fig. 4a), typical of continent-scale
studies, does not separate this feature from a general region of lower velocity in eastern Africa.

The models at 175 km depth show a continuation of the fast velocities beneath much of the Congo and Kalahari Craton, whereas beneath the Tanzanian Craton the velocities are now slightly slower than the reference model. Across southern Africa the results from this study at depths of 100–175 km are quite consistent with the recent continent-scale tomography presented by Begg et al. (2009). This consistency is particularly encouraging given the different datasets used — SV Rayleigh wave tomography in this study and SH body-wave tomography in Begg et al. (2009). Beneath the Kalahari Craton, the high resolution model (Fig. 4b) suggests that there are quite significant variations in wavespeed at 175 km depth. The fastest velocities within this region are observed in the northern Kaapvaal around the South Africa–Botswana border. On one hand, encouragingly, these results are reasonably consistent with the array study of Li and Burke (2006), although their even finer parameterisation allows more detailed structure to be recovered. On the other, there still remain significant differences between the velocity variations observed in different high resolution studies for both surface (Li and Burke, 2006; Chevrot and Zhao, 2007) and body waves (Touch et al., 2004), thus making it harder to have confidence in interpreting the detailed velocity structure.

In the deepest models shown (250 km depth) there are very few parts of southern Africa that have seismic velocities faster than the global reference model. Noticeably slow velocities are observed beneath parts of the East African rift system, and a large (800°400 km) region of slow velocities is also seen beneath much of Angola and Namibia. At this depth the model with knot points at 1.5° intervals shows little difference to the lower resolution model. This is not surprising; given the increased uncertainty in the path-average models with depth (e.g., Fig. 2), and the wider sensitivity kernels of the long period surface waves, recovering small-scale anomalies becomes more difficult at greater depth.

4. Discussion — lithospheric thickness estimates

4.1. Surface wave tomography

One strategy for converting the seismic velocities into an estimate of lithospheric thickness is to apply the method of (Priestley and McKenzie, 2006). They used thermal models of the Pacific and P–T estimates from Kimberlites alongside a global tomographic model of shear velocity in order to find an empirical relationship linking velocity, temperature and pressure. Using their parameterisation the
velocities from the tomographic model are converted into temperature estimates. In this study the lithospheric thickness is then defined by calculating at what depth the temperature profile reaches the isentrope for a potential temperature of 1315 °C. This is a slightly more simplistic approach in comparison to fitting theoretical temperature profiles, as used by Priestley and McKenzie (2006), but gives similar estimates in thickness. Fig. 5 shows the thermal lithospheric thickness estimates for southern Africa for the tomographic model with knot points spaced at 3° intervals. The location of kimberlites from the Consorem database (Faure, 2006) are also indicated. Known diamondiferous locations are plotted in red, and unknown and non-diamondiferous kimberlites are plotted in brown.

The results indicate significant differences in thickness beneath the cratonic regions of southern Africa. Beneath the Congo region lithospheric thickness varies from greater than 220 km in the west to around 180 km in the east (Fig. 5). The majority of previous tomographic studies have indicated fast velocities, typical of cold cratonic lithosphere, extending to depths of at least 200 km beneath the Congo Basin (e.g. Sebai et al., 2006; Priestley et al., 2008; Begg et al., 2009). Using group velocities, (Pasyanos and Nyblade, 2007; Pasyanos, 2010) also suggested that the lithosphere was thickest beneath the western part of the Congo, but gave significantly thinner values (140–160 km) beneath the central part of the basin. Given the good path coverage of the different models, the most probable reasons for the differences in tomographic models are the vertical parameterisations used, and the stronger sensitivity to shallow (crustal) structure for group velocities.

The thinnest lithosphere of the cratonic regions is observed beneath Tanzania (Fig. 5). Given that the same waveform inversion technique is used it is surprising that the thickness estimate in this study varies so much from the work of Priestley et al. (2008), who estimate very thick lithosphere in this region. The synthetic tests of Priestley et al. (2008) indicate that their model is well resolved in this region, as do the tests performed here. Given that the two approaches use very similar inversion codes, the differences must be in the treatment of the data. In this study the code is only semi-automated with a manual comparison of multiple inversions for each path and thus one possibility is that the fully-automated inversion is accepting a few data with mis-constrained 1D models, which are then mapped into the high velocity zone beneath this region. However, Priestley et al. (2008) use cluster analysis on their dataset, which should remove any outliers, and so presently no explanation for the differences is completely satisfactory.

In contrast, the seismic velocities illustrated in this study (Fig. 4) are very consistent with the results obtained in the array tomography of Weeraratne et al. (2003). The similarities between two studies at two such different scales, alongside the receiver function study of the transition zone in this region (Huerta et al., 2009), suggest that the results are reliable. One possible explanation for the thin lithosphere could be the impact of thermal upwelling on the base of the lithosphere. Whether this is a thermal thinning of lithosphere, as implied in the direct translation to lithospheric thickness, or other major modification of the base of the sub-continental lithospheric mantle resulting in very low seismic velocities, is difficult to assess.

Beneath the Kalahari Craton in southern Africa the estimates of lithospheric thickness are in the vicinity of 180–200 km for much of the craton, with slightly greater thicknesses close to the South Africa–Botswana border. In this region the estimates are similar to those of Priestley et al. (2008). Using an absolute velocity of 4.55–4.6 km s⁻¹ (Li and Burke, 2006) also suggested a depth to the LAB of 180 km, similar to the estimates from the empirical parameterisation. In contrast, the body-wave tomography for the Kaapvaal region have shown fast velocities extending to depths in excess of 300 km (Fouch et al., 2004), and thus a suggestion of much thicker lithosphere. Priestley and Tillmann (2009) illustrate that these results might not be completely incompatible and can arise from the same structure, simply due to differences in method and regularisations between the body-wave and surface-wave tomography. Alternatively, the variations could be explained by the different behaviour of Vp and Vs with temperature and pressure. For a modelled Archaean Craton Afonso et al. (2008) observe a decrease in Vs within the lithospheric mantle, while Vp monotonically decreases.

On the Namibian margin the thickness estimates can be compared to the results of Fernández et al. (2010). From the modelling of petrology, mineral physics, gravity anomalies and heat flow the authors estimate the thermal lithospheric thickness to be increasing gradually from 100 to 125 km in the oceanic region, with a sharp increase to 175 km in the continental domain (Fernández et al., 2010). In the present study (Fig. 5) the lithospheric thickness on the Namibian margin increases gradually from 120 to 140 km in the oceanic domain, before rapidly increasing to thicknesses greater than 160 km beneath the continent. Given the different methods used, and the different lateral resolution of the studies, the similarity of the results is encouraging.

4.1.1. Limitations

While the technique used to calculate lithospheric thickness does provide estimates compatible with the thermal models beneath oceans and some kimberlite data, there are a number of assumptions that limit our interpretations. The original empirical parameterisation (Priestley and McKenzie, 2006) requires good knowledge of the temperature structure beneath the ocean, and that the velocity model used is an accurate representation of this structure. Including a fit to kimberlite data was required to constrain the parameterisation but also raises questions as to whether the present day seismic structure should generate temperature profiles compatible with profiles derived from kimberlites of different ages. Additionally, the empirical nature of the parameterisation implies that the most reliable results in this study will be obtained if the seismic velocities presented have been created using a similar inversion scheme to the velocities used by Priestley and McKenzie (2006). Importantly, the waveform inversion procedure of Debye (1999), or slight modification thereof, have been used in both this study and by Priestley and McKenzie (2006) and therefore the
vertical parameterisation will be the same. Different tomographic inversion schemes and regularisations, have been used and therefore create uncertainties in the temperatures. For example, a tomographic model with significantly less damping (regularisation) than that used by Priestley and McKenzie (2006) will have larger variations in temperature than those that may actually be present.

A comparison with estimates of the temperature–velocity relationship from mineral physics data can also be made. Using the method of Afonso et al. (2008), Griffin et al. (2008) estimate that at a depth of 100 km beneath Archaean cratons and for a temperature of 800 °C the shear velocities would be around 4.71 km s⁻¹. Although these are anharmonic velocities, at this temperature the effects of anelasticity should be minor. Using this velocity (4.71 km s⁻¹) and depth (100 km) the empirical parameterisation gives temperatures of approximately 355 °C. Considering Priestley and McKenzie (2006) Fig. 5a these differences can be expected, at low temperatures the velocities of the empirical parameterisation are much slower than those predicted by either the methods of Paul and Jackson (2005) or Goes et al. (2000). Despite these limitations the method can be used as a good proxy for lithospheric thickness. However, the absolute temperature profiles generated from the empirical parameterisation must be treated with caution.

One of the difficulties with using only a surface wave dataset is that it is very hard to separate out the effects of other physical parameters beyond temperature. This problem exists whether using an empirical parameterisation or the mineral physics derived relationships between velocity and temperature. Grain size, water, melt, and varying composition will all have an effect on the seismic velocities. While changes in composition appear to only have a small (1–2%) effect on seismic velocity (e.g. Schutt and Lesher, 2006; Griffin et al., 2008), this will affect the temperatures estimates. For example, using the empirical parameterisation, at 200 km depth velocities of 4.55 km s⁻¹ will be converted to 1400 °C and velocities of 4.62 km s⁻¹ will be converted to 1250 °C; variations in temperature that will affect the estimates of thickness.

Furthemore, both the vertical parameterisation and the vertical resolution will affect the reliability of the temperatures and thickness estimates. Many surface wave models appear to show increasing velocities in the uppermost mantle (Moho — ~100 km depth) (e.g. Fishwick and Reading, 2008; Lebedev et al., 2009; Pedersen et al., 2009). This seismic velocity distribution cannot easily be explained by one single composition and a cratonic geotherm, and will lead to physically unrealistic temperature profiles. One possibility is that these errors may be caused by the smearing of slow crustal velocities into the uppermost mantle, however the number of different seismic models that show this feature suggest that the shallowest mantle remains poorly understood. Ideally, velocity models compatible with simple physical models should be tested (e.g. Shapiro and Ritzwoller, 2004; Pedersen et al., 2009), deviations from these physically consistent models will then suggest an increased complexity of structure.

Within the upper mantle the vertical resolution in the order of 25–50 km means that it is not possible to resolve any rapid changes (discontinuities) in seismic velocity using surface waves. Using the thermal definition of the lithosphere, variations in seismic velocity should be gradual, and one would not expect to see a discontinuity at the depth of the thermal LAB. However, discontinuities are observed in the upper mantle and have been used to estimate LAB thicknesses, so it is important to consider what physical properties these can be related to.

### 4.2. Seismic discontinuities

Receiver functions offer one of the best approaches to observe rapid variations in seismic velocity in the upper mantle. The P-receiver function, where incoming P waves are converted into S-waves, has frequently been used to estimate depths to the major seismic discontinuities such as the Moho, and transition zone. In contrast, when investigating discontinuities at depths consistent with an expected LAB, the results from P-receiver functions can be harder to interpret due to the possibility of arrivals being masked by reverberations from the Moho. More recently, the S-receiver function technique (using S-waves converted into P waves) has been used to investigate discontinuities in the upper mantle. The advantage of this technique is that the converted phases arrive before the direct phase and thus avoid the interference of multiples from shallow discontinuities such as the Moho. One disadvantage is that the resolution is lower for S-receiver functions due to the longer periods of the incoming S-waves. For a review of receiver function imaging of the lithosphere–asthenosphere boundary see Ryche et al. (2010–this volume).

In the present study we focus on the Kalahari Craton of southern Africa. Since the deployment of the SASE array (Carlson et al., 1996), it has become one of the most studied cratonic regions in terms of seismic data, and offers the ideal opportunity to look at the variety of discontinuities that are observed by receiver functions, how these compare with surface wave estimates, and the relationship to the LAB.

Using the P-receiver function Gao et al. (2002) suggested that there were no discernible discontinuities between the Moho and 410 discontinuity, and that the transition zone thickness does not vary beneath the region. In contrast, using P-receiver functions and data from permanent seismic stations Ryche and Shearer (2009) showed evidence for a strong discontinuity in the upper mantle, but at much shallower depths than expected for the LAB beneath a craton. Fig. 6b shows the results for the LBTB station in Botswana (Ryche and Shearer, 2009). A discontinuity, resulting from a decrease in seismic velocity, is observed at around 100 km depth. The authors note that their observations require a change in velocity that could not be caused by thermal variations alone. For similar discontinuities beneath the oceans and phanerozoic terranes the preferred explanation is a change in hydration or partial melt, however this seems unlikely for cratonic areas as it implies a decoupling within the zone assumed to be the lithospheric mantle (Ryche and Shearer, 2009).

Fig. 6c shows the results from the S-receiver study of Wittlinger and Farra (2007). While the P-receiver function study of the SASE array (Gao et al., 2002) did not observe any clear discontinuities in the upper mantle, the S-receiver functions suggest that there is distinct layering of structure beneath the craton. Wittlinger and Farra (2007) associate the observed receiver functions with a radially anisotropic upper mantle, increasing velocities between 200 and 300 km and then a negative velocity gradient from around 300 km depth to the transition zone. They thus associate the base of the tectospheric root of the Kaapvaal to be at about 300 km depth. A similar LAB depth from S-receiver functions in southern Africa was also estimated from permanent stations (Kumar et al., 2007), although the results at the southern Africa stations (BOSA, LSZ, and LBTB) suggest that there may also be discontinuities observed at shallower depths.

Savage and Silver (2008) utilise both P and S-receiver functions for this region. They suggest that in the P-receiver functions discontinuities observed in the depth range of 150–200 km are artifacts due to multiple reflections. In the S-receiver functions they observe a distinct discontinuity at around 150 km depth beneath the northern Kaapvaal. Using both long and short period waves their preferred velocity structure does not require radial anisotropy, but instead a velocity decrease of 4.5% split between two discontinuities around 10 km apart near 150 km depth (Savage and Silver, 2008). Rather than associating this discontinuity with the LAB they suggest that it is most likely related to a rapid change in composition, and is the result of refertilisation of the lithosphere by Fe- and volatile rich melts. Using S-receiver functions Hansen et al. (2009) observe a similar discontinuity at 150 km depth beneath the southern Kaapvaal—a region where Savage and Silver (2008) found only weak evidence of a discontinuity. Moorkamp et al. (2010) jointly invert receiver function, surface wave dispersion and magnetotelluric data beneath one station of the SASE array. They also
observe a low velocity zone at depths of around 150 km which appears to be linked to a zone of higher conductivity.

One of the difficulties in assessing models from receiver functions is that using Occam’s principle of obtaining the simplest model that adequately fits the data has led seismologists to frequently obtain models with the fewest number of layers; e.g., with a constant high velocity lid and then a low velocity zone beneath. While these parameterisations are the simplest models that describe the seismic velocity characteristics, they are not the simplest models in terms of the physical properties of the upper mantle. For a constant composition velocities are expected to decrease within the lithospheric mantle (e.g. Faul and Jackson, 2005; Stixrude and Lithgow-Bertelloni, 2005).

In the same way that it is becoming more common to assess surface wave dispersion estimates in terms of physically based parameterisations (Shapiro and Ritzwoller, 2004; Pedersen et al., 2009), one possibility is to use \textit{a priori} models based on physical properties and then inverting receiver functions for the minimal divergence from these starting models, rather than a simple seismic reference model. Although this is unlikely to change the number, or magnitude, of the discontinuities, the different background model should give a better understanding of the depth and cause of discontinuities within the upper mantle.

4.3. Anisotropy

Changes in the pattern of seismic anisotropy can be used to attribute a depth to the lithosphere–asthenosphere boundary (e.g. Plomerová et al., 2002). Eaton et al. (2009) showed that in a dry rheology dislocation creep, and thus seismic anisotropy, will be present in the lower part of the lithosphere and in the asthenosphere. The depth to the LAB is approximated by choosing a particular strain rate, and is found to be close to the depth of the thermal boundary layer. In contrast, if the lithosphere is dry and the asthenosphere is wet, then the transition to a dislocation creep dominated rheology will occur as a sharp boundary and the depth is controlled completely by the presence of water (Eaton et al., 2009).

Sebai et al. (2006) used both Rayleigh and Love waves to produce anisotropic models of the velocity structure beneath Africa. Beneath the cratonic regions they noted a general reduction in radial anisotropy down to depths of about 300 km, although the very different path coverage of the Rayleigh and Love datasets makes detailed comparisons difficult. Beneath the Kalahari Craton they observed very weak anisotropy at 180 km depth. The systematic changes in azimuthal anisotropy at this depth beneath the cratonic regions was suggested to be the signature of the cratonic root (Sebai et al., 2006).

Although it is relatively easy to incorporate azimuthal anisotropy into the inversion scheme for surface wave studies (e.g. Sebai et al., 2006) it is much harder to distinguish real anisotropy from an apparent anisotropy that can be traded off against velocity heterogeneity. Initial inversions on the present dataset suggested a change in direction of anisotropy at a depth between 175 and 250 km depth, similar to the thermal LAB estimates. However, synthetic tests indicate that this cannot easily be differentiated from possible artifacts due to path coverage and the heterogeneity–anisotropy trade-off.

4.4. Kimberlite localities

Although kimberlites have long been associated with Archaean cratonic regions (Clifford, 1966) underlain by thick lithospheric
mantle, the causes of their detailed distribution remains an area of discussion. At crustal depths tectonic controls may be terrane boundaries, dyke swarms or fracture zones (e.g., Jelsma et al., 2009) and diamondiferous deposits frequently occur close to gradients or discontinuities in potential field data (Jaques and Milligan, 2004). Within the mantle, thick lithosphere is required to obtain the diamond stability field, distinct changes, or strong lateral gradients in the thickness of the lithospheric mantle, may be a deep–sealed feature that can control the location of kimberlites. O’Neill et al. (2005) use numerical modelling to illustrate that these rapid changes in lithospheric thickness result in large stress gradients, and can influence melt migration.

Comparisons of tomographic images and the location of diamondiferous deposits have been made at the continental scale for both Australia (Jaques and Milligan, 2004; O’Neill et al., 2005) and Africa (Priestley et al., 2008). For the Kaapvaal region more detailed comparisons have made with both seismic velocity images (e.g. Shirey et al., 2002; Griffin et al., 2003; Begg et al., 2009) and with electrical conductivity (Jones et al., 2009). One of the difficulties in assessing the relationship between the kimberlites and tomographic models is that observations are strongly dependent on the choice of the reference velocity and the colour scheme used to represent velocity perturbations. An alternative method is to simply plot the horizontal gradient of the velocity in order to make a direct evaluation of edge features (Fishwick, 2006).

Fig. 7 illustrates the variations in horizontal velocity gradient for the high resolution tomographic model at a depth of 175 km. In Tanzania, the kimberlite database (Faure, 2006) shows a linear trend of kimberlites running roughly NNN–SSE through the centre of the Tanzanian Craton. The Mwadui pipe, part of the northern cluster of kimberlites, is the largest diamondiferous kimberlite to be mined worldwide (Stiefenhofer and Farrow, 2004). Although the kimberlites appear to cut through the centre of the surface outline of the craton, Fig. 7 shows that the kimberlites follow close to a clear gradient in seismic wavespeed, related to a change in lithospheric thickness. To the west thicknesses exceed 160 km, with thinner depth estimates to the east (Fig. 5). Garnet geochemistry from Tanzanian xenoliths suggests that the depleted mantle is also only observed to depths of around 160 km (Griffin et al., 2003). However, given that the tomographic study shows present day velocities, and the Mwadui kimberlite has an estimated age of 52 Ma (Stiefenhofer and Farrow, 2004) it is difficult to tell if the deep-seated mantle structure is the cause of the location of kimberlites, or is the result of their formation.

A SW–NE corridor of kimberlites is observed through Angola and into the Democratic Republic of Congo (DRC). In this region there are some weak gradients in the seismic velocities with faster velocities generally to the North, but no coherent lineation of the gradient as observed in relation to the Tanzanian kimberlites. The lithospheric thickness estimates from the tomographic model are 180–200 km, increasing towards the DRC. Garnets from the Mbuji Mayi kimberlite field in the DRC indicate cool geotherms and a lithospheric thickness of 210 km (Batumike et al., 2009) — in good agreement with the seismically derived model.

Given the large number of seismic stations within the SASE array and the good path coverage (Fig. 1), and the large number of kimberlites in the region a comparison of tomography and kimberlite distribution around the Kaapvaal region (e.g. Shirey et al., 2002; Griffin et al., 2003; Begg et al., 2009). There is no clear relationship between the horizontal velocity gradient at 175 km depth and the location of the kimberlites. However, a close examination of the absolute velocities in the tomographic model (Fig. 8) indicates that the diamondiferous kimberlites are generally not sampling the regions of fastest velocity. One of the difficulties with this detailed assessment is that even the high resolution surface wave model has lateral resolution of approximately 200 km, and the gradients in velocity will be dependent on the regularisation (damping) applied to the inversion. Furthermore different high resolution studies still produce different images of the mantle (Li and Burke, 2006; Chevrot and Zhao, 2007), highlighting uncertainties in the models. The narrow aperture of the SASE array may limit the resolution on the edges of the craton. The increasing instrumentation in southern Africa from the Africa Array project (Nyblade et al., 2008) should reduce the problems associated with the station distribution for future models.

Similarly, detailed comparisons of the lithospheric thickness estimates from kimberlites with the estimates from the tomographic

Fig. 7. Greyscale image of the horizontal gradient of seismic velocity using the high resolution tomographic model at 175 km depth (Fig. 4b). The absolute values of the gradient are strongly dependent on the damping used within the inversion and are not shown. The locations of kimberlites from the Consorem database (Faure, 2006) are plotted as diamonds (black — diamondiferous, and grey — non-diamondiferous/unknown).

Fig. 8. Absolute velocities in southern Africa using the high resolution tomographic model at 175 km depth. The locations of kimberlites from the Consorem database (Faure, 2006) are plotted as diamonds (red — diamondiferous, and brown — non-diamondiferous/unknown).
models remain difficult. Firstly, the mantle thermobarometry of xenoliths of southern Africa can be considered as recording a snapshot of a thermally evolving system, rather than being representative of the steady state of the lithospheric mantle (e.g. Bell et al., 2003). Kobussen et al. (2008) show an example of this evolving state for two suites of xenoliths on the southwest margin of the craton. It is therefore not simple to make a direct evaluation of the present day thermal state from all xenoliths. Secondly, as discussed previously, the conversion between velocity and lithospheric thickness is not without complications.

5. Summary

The results from the surface wave tomography suggest that there is significant difference in lithospheric thickness beneath the cratonic regions of central and southern Africa. Taking a basic approach, where the lithosphere is defined as the layer where heat transport is predominately by conduction and converting seismic velocities to temperatures using the method of (Prieske and McKenzie, 2006), cratonic thicknesses vary from around 150 to 160 km beneath Tanzania to around 200 km beneath much of the Kalahari Craton to >225 km beneath parts of the Congo. One possible explanation for these differences is to associate the variations in thickness to the recent thermal history of the regions: thinnest lithosphere beneath the cratonic region adjacent to the East African rift zone, and thickest lithosphere beneath the Democratic Republic of Congo where a mantle downwelling has been proposed (Crosw et al., in press: Forte et al., 2010).

The limitation of using surface wave tomography to estimate lithospheric thickness is that it is hard to resolve rapid variations in the vertical shear velocity structure and all the variations in velocity are accounted for in terms of temperature. Detailed interpretations of the structure in regions such as the Kalahari Craton remain difficult. The petrology suggests varied levels of depletion and metasomatism, and also a complex evolution over time of the lithospheric mantle. The tomography shows the present day snapshot of the velocity structure and separating the details between small changes in temperature and compositional differences is not possible. Despite this the broad variation in a thermally defined LAB depth between the different cratonic regions of central and southern Africa, show similarities to estimates from mantle xenoliths, and thus appear to be robust.

The results from receiver function studies suggest there may be a variety of discontinuities in the upper mantle. It is interesting to note that in South Africa most of the discontinuities are placed either above (100 km — P-receiver function (Rychert and Shearer, 2009), 150 km — S-receiver function (Savage and Silver, 2008; Hansen et al., 2009)) or below (300–350 km — S-receiver function (Wittlinger and Farra, 2007; Kumar et al., 2007)) the depths (200 km) where thermal estimates would place the LAB. These results suggest that we should be cautious in automatically associating a discontinuity from fast, to slow velocities as placing the LAB. These results suggest that we should be cautious in automatically associating a discontinuity from fast, to slow velocities as placing the LAB. These results suggest that we should be cautious in automatically associating a discontinuity from fast, to slow velocities as placing the LAB. These results suggest that we should be cautious in automatically associating a discontinuity from fast, to slow velocities as placing the LAB. These results suggest that we should be cautious in automatically associating a discontinuity from fast, to slow velocities as placing the LAB.

Acknowledgements

Significant thanks must go to the many, many people who have been involved in the establishment of permanent seismic stations, and the deployment of the numerous temporary networks of seismometers that have been used within this study. The facilities of the IRIS Data Management Centre were used to access much of the data, and also the GEOFON Data Center of GFZ Potsdam. Eric Debyale and Kazu Yoshizawa are thanked for the use of their inversion codes, and Ben Ellis for help drafting a figure. This work on African structure began while at Bullard Laboratories, University of Cambridge, and colleagues there are thanked for stimulating discussions. The opportunity to write this article for the special issue came from Sue O’Reilly. Timely discussion with many participants at the ESF Exploratory Workshop on Defining the Lithosphere–Asthenosphere Boundary, held at the Dublin Institute of Advance Studies, helped refine the content of the paper. Suggestions from Juan-Carlos Afonso and one anonymous reviewer have also improved the manuscript. The work has been funded by NERC New Investigator grant NE/G000859/1.

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