Anomalous lithosphere beneath the Proterozoic of western and central Australia: A record of continental collision and intraplate deformation?

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**ABSTRACT**

A new surface wave seismic tomography model of Australia is presented which provides a means of investigating the lithospheric structure beneath the Proterozoic regions in the west, north and centre of the continent with improved resolution and reliability. The dominant feature of the model is a region of low seismic wavespeeds in the uppermost mantle, at approximately 75 km depth, beneath central Australia. The zone of slow wavespeeds is underlain by a region of fast wavespeeds, more typical of continental lithosphere. This layered velocity structure, and strong positive wavespeed gradient, makes the shallow anomaly hard to explain in terms of high mantle temperatures and typical steady-state continental geotherms. A possible thermal explanation requires the impact of the redistribution of high heat producing elements within the crust. Alternatively, a mineral or minerals with low seismic velocities, such as amphibole, in the shallowest part of the lithosphere, with a more conventional lithology in the deep continental root below, could explain the seismic wavespeeds. The anomaly is located directly beneath the zone where the Australian cratons amalgamated in the Proterozoic and where, subsequently, there have been periods of intraplate tectonic activity; suggesting a correlation between the prolonged history of deformation and a highly unusual lithospheric structure. In western Australia, the Capricorn Orogen and Pilbara Craton have a similar lithospheric thickness, whereas a thicker lithosphere is observed beneath the Yilgarn to the south. In northern Australia, large regions appear to be underlain by fast wavespeeds, similar to those observed beneath the Yilgarn Craton. Variations in the shear wavespeeds beneath the Arunta also indicate that there is not always an obvious correlation between the overlying surface geology and seismic structures observed in the upper mantle.

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1. Introduction

Throughout the Proterozoic, the evolution of the Australian continent was dominated by episodes of accretion and continent–continent collision (e.g. Myers et al., 1996; Betts et al., 2002). The prolonged active tectonic history is clear from the mapped geology of the continent. The relationship between the surface structure and the underlying lithospheric structure is less clear. Variations in lithospheric structure may increase the likelihood of certain geological processes at the surface, or the tectonic history of surface deformation may also be represented in the lithosphere. Using an improved dataset of seismic waveforms, it is now possible to obtain better resolved surface wave tomographic images. Thus, the variations in the shear wavespeed of the upper mantle of central Australia may be examined to place constraints on the Proterozoic, and subsequent, evolution of the continent.

1.1. Tectonic setting

The shield region of central and western Australia consists of three main blocks: the West, North and South Australian Cratons (after Myers et al., 1996; Betts et al., 2002), with a Proterozoic history dominated by accretion and continent–continent collision (Fig. 1). Within the West Australian Craton, the Capricorn Orogen records a series of Palaeoproterozoic events associated with the amalgamation of the Pilbara and Yilgarn Cratons. Deformation occurred in four main pulses from 2200 Ma to ca. 1650 Ma. Repeated reactivation continued in an intraplate setting including further basin formation and orogenic activity (Cawood and Tyler, 2004). The North Australian Craton is also an amalgamation of
blocks: the Kimberley Craton in the northwest; the Tennant Creek in the centre; the Mt Isa block to the east; and the early Proterozoic Arunta inlier on the southern margin. Many of the blocks were likely to have been accreted during the 1870–1840 Ma Barramundi Orogeny (Etheridge et al., 1987). The accretion of the West and North Australian Cratons corresponds to major phases of continental collision and supercontinent assembly world-wide (Cawood and Tyler, 2004; Betts et al., 2002). The South Australian Craton includes the Archean to Mesoproterozoic Gawler Craton in the centre and west, and the Mesoproterozoic Curnamona Province in the east (Betts et al., 2002).

The joining of the major cratons to form the Australian shield occurred during the Proterozoic. A number of differing interpretations exist concerning the exact nature and timing of associated events. Myers et al. (1996) proposed that the three cratonic blocks were separated through the Palaeoproterozoic, and tectonic activity from 1300 to 1100 Ma led to the assembly of the shield region. More recent work on the Rudall complex to the east of the Pilbara Craton (Bagas, 2004) suggests that the collision of the West and North Australian Cratons occurred earlier, at ca. 1800 Ma. Betts and Giles (2006) also suggest that the North and West Australian Craton collided between 1790 and 1770 Ma, with the amalgamation of the nucleus of the Gawler Craton and the North Australian Craton occurring between 1750 and 1690 Ma. Giles et al. (2004) propose that the Curnamona province (South Australia Craton) and eastern Mt Isa Block (North Australia Craton) were adjacent until approximately 1.5 Ga, before rifting and re-organisation placed the South Australian Craton in its present configuration. While there is more than one reconstruction that fits the geological data, it is apparent that the southern margin of the North Australian Craton was the location for a number of accretionary events over a prolonged timespan.

Following the amalgamation of the shield, this central region continued to be subject to intraplate orogenic activity. Wingate et al. (2004) relate mafic igneous rocks, found across western and central Australia, with a large Mesoproterozoic (ca. 1070 Ma) igneous province formed above a plume head located beneath central Australia. During the Neoproterozoic a centralian superbasin developed where the three cratonic regions had collided (Walter et al., 1995). Zhao et al. (1994) suggest this superbasin formed as a result of extension and thinning of the lithosphere due the upwelling of a later mantle plume, centred on the eastern edge of the Gawler. In the Phanerozoic, deformation continued to take place in central Australia, despite its great distance from any of the plate boundaries active at that time. The Petermann Orogeny (600–520 Ma) occurred along the southern margin of the Amadeus Basin, while the Alice Springs Orogeny (400–300 Ma) occurred along its northern margin. Sandiford and Hand (1998) relate the location of deformation to regions containing high heat producing basement rocks, which were blanket ed under thick layers of sediments. Roberts and Houseman (2001) suggest that clockwise rotation of a north Australian block relative to a southern block, could cause the pattern of deformation and crustal thickening that are observed. The narrow zone of deformation implies that the blocks to the north and south are stronger, perhaps due to colder temperatures at the Moho.
From the available measurements, central Australia has an anomalously high heat flow (Cull, 1982), with an average value of greater than 80 mW m\(^{-2}\). This value is far higher than the global average of heat flow data in Proterozoic terranes, and McLaren et al. (2003) show that it is likely to be caused by high heat production in crustal rocks, rather than by a mantle source of heat flow or by transient heat flow related to neotectonic activity. McLaren et al. (2005) suggest that the high heat production is important in defining the Proterozoic evolution of Australia, due to the additional tectonic control that the high heat producing elements (HPEs) can produce through their weakening of the lithosphere. Pyysyweck and Beaumont (2004) have performed numerical modelling, which suggests that a crust containing high HPEs, particularly if distributed in the deep crust, is susceptible to tectonic activity.

1.2. Previous seismic work

Previous seismic investigations into the lithospheric structure of Proterozoic Australia have been carried out at both local and continental scales. The majority of local studies have focused on crustal structure, using reflection data (e.g. Goleby et al., 1989), teleseismic travel times (e.g. Lambeck and Burgess, 1992; McQueen and Lambeck, 1996) or a combination of the two (Korsch et al., 1998). Within central Australia there has been a major displacement in crustal and mantle structure, to the south of the Arunta Block thrust faulting appears to displace the Moho along the Redbank Thrust Zone (Goleby et al., 1989). The travel time anomalies also indicate dipping interfaces, extending to at least 50 km depth and in some cases down to at least 80 km (McQueen and Lambeck, 1996).

The deployment of mobile networks of broad-band seismometers began in 1993 with the initial SKIRR deployment (van der Hilst et al., 1994). Five more deployments of seismic stations gave continental coverage across Australia with approximately 400 km spacing between the stations. Since SKIRR there have been a number of other temporary networks focused on particular regions of the continent. The improved frequency range and three-component nature of broad-band seismometers allows far greater exploitation of the seismogram, opening up a greater range of studies than previously possible (Kennett, 2003).

Clitheroe et al. (2000) investigated the crustal structure across the continent using data from both permanent seismic stations and the temporary broad-band seismic stations of the SKIRR and KIMBA projects. Using receiver functions, they determined the depth to the Moho at 65 stations across Australia and, incorporating this information with reflection and refraction studies, presented a map of crustal thickness for Australia. Collins et al. (2003) added further reflection and refraction data to improve the model, however the data points remain both disparate and unevenly spaced, making the interpolation and interpretation of crustal depth in central Australia difficult.

Surface wave tomography has been used to map the variations in shear wavespeed across the whole continent (e.g. Zielhuis and van der Hilst, 1996; van der Hilst et al., 1998; Simons et al., 1999, 2002; Debayle and Kennett, 2000; Yoshizawa and Kennett, 2004; Fishwick et al., 2005). Global seismic studies have suggested that there is a relationship between increasing lithospheric thickness and an increasing age of the overlying crust (Polet and Anderson, 1995). All the studies of continental Australia show a large difference in S-wave velocity between Phanerozoic eastern Australia and the older cratonic regions of central and western Australia. Simons and van der Hilst (2002), however, indicated that on scale lengths of less than 1000 km the relationship between the age of the overlying geology in Australia and the lithospheric thickness is more complicated.

Using new data (see below) in addition to the dataset of Fishwick et al. (2005), this paper focuses on the variations in shear wavespeed beneath the Proterozoic regions of central and western Australia.

2. Data and methods

2.1. Improvements to dataset

In 2003–2005, 20 broad-band seismometers were deployed for the TASMAC experiment in central and eastern Australia. In the work described in this paper, data from the TASMAC experiment, and new data from the joint pmd-crc and Research School of Earth Sciences experiment in western Australia (see Reading et al., 2007) have been added to the dataset of Fishwick et al. (2005). Some further data have been included from permanent seismic stations, particularly those on ocean islands to the west of Australia, in order to improve the azimuthal coverage. These new data give an additional 520 paths, an increase of almost 25% in comparison to the previous work (Fig. 2c). The increased data coverage improves resolution.
throughout Australia and also allows a comparison of the isotropic shear wavespeed model with an inversion incorporating azimuthal anisotropy.

### 2.2. Tomographic techniques

The method of surface wave tomography described by Fishwick et al. (2005) is employed in this work. In this technique, the surface wave tomography is a two-step process. Firstly, the path-average 1D shear wavespeed depth profiles are calculated, and secondly, many such 1D models are combined within a tomographic inversion to derive a 3D model of the velocity structure.

The 1D model for the path between the source and the receiver is calculated using the automated procedure of Debayle (1999), based on the waveform inversion scheme of Cara and Lévêque (1987). The 1D path average model is defined at 25 km intervals from 50 km depth down to the transition zone. To improve the reliability of the model, multiple starting models are used within the waveform inversion procedure (Fishwick et al., 2005). The starting models have different velocities between the Moho and 200 km depth, with variations between $-4.5\%$ and $+4.5\%$ from the reference model PREM (Dziewonski and Anderson, 1981). From 200 to 400 km depth the models slowly converge back to the starting models converge back towards the values of PREM. The consistency between the group of 1D path average models, derived from these differing starting models, provides an additional means to assess the reliability of the wavespeed profile with depth. The final path specific model used in the tomographic inversion is a weighted average of the similar 1D path average models accepted by the inversion scheme, and the spread of the models will be used in defining the weighting placed on each path within the inversion scheme.

The set of 1D models is then input into a tomographic inversion to derive a model of the 3D velocity structure. As the 1D models are defined at 25 km intervals, at each depth a 2D horizontal slice is independently calculated. Initially we invert for only the isotropic velocity structure and the spherical B-spline parameterisation scheme of Yoshizawa and Kennett (2004) is used to provide a smooth representation of the wavespeed variations. The length scale of horizontal smoothness is constrained by the distance separating the knot points within the spline function. The 2D horizontal slices can be constructed by using a linear inverse approach relating the velocity perturbation at a particular point to the path information at the same depth. The tomographic inversion is performed in two separate stages. Initially, an inversion with widely spaced knot points is performed to obtain the large-scale structure. In order to obtain the more detailed structure within the region this new tomographic model is then used as the reference model, for an inversion with closely spaced knot points (Fishwick et al., 2005). The inversion is regularised using a damping factor $\lambda$. If the damping is large we minimise the under-determined part of the solution, but will not minimise the data fit. If no damping is applied we minimise the data-fit, but no a priori information will be used to single out under-determined parameter (Menke, 1989). Various choices of $\lambda$ are applied and we choose an appropriate value based on the trade-off between fitting the data and remaining close to the large-scale structure of the initial inversion.

Azimuthal anisotropy can also be included in the tomographic procedure for the Rayleigh waves (e.g. Simons et al., 2002; Debayle and Kennett, 2003). In a weakly anisotropic medium the variation of wavespeed for the Rayleigh wave with respect to the azimuth $\theta$ will be predominantly dependent on the cosine $2\theta$, $\sin 2\theta$ terms (see Smith and Dahlen, 1973). Using a tomographic inversion code that incorporates this anisotropic component (K. Yoshizawa, 2004, pers. comm.) we are thus able to compare the results of an isotropic and anisotropic inversion.

### 2.3. Reliability

In order to give an impression of the reliability of the tomographic images, we perform a checkerboard test at 100 km depth (Fig. 2a and b). This style of synthetic test indicates whether there is the possibility to resolve a particular structure given the particular path coverage of our dataset. The results of the checkerboard test are predominantly dependent on the density and azimuthal distribution of data. We therefore present the path coverage alongside these results (Fig. 2c).

In this test, 1D models are created by averaging the velocity structure of the checkerboard input pattern for each path used in the inversion. As a means of simulating the limitations of the waveform inversion step, a component of noise has been added relative to the weighting for the path used in the actual inversion (Fishwick et al., 2005). The path-average 1D models are then used to reconstruct the imposed wavespeed structure.

The results indicate that good recovery of the input model is observed where there is a good path coverage, both in terms of number of paths, and azimuthal distribution of paths. In regions dominated by paths of one direction we can see smearing of the initial wavespeed structures. Throughout continental Australia the recovery of the input structure is generally good. The location and amplitude of the input anomalies are quite well recovered. Some northeast–southwest smearing of the input anomaly is observed in central Australia. The results suggest that variations in wavespeed occurring over short distances can be recovered within the continent.

A number of other factors can influence the recovery of structure in synthetic tests. For slices deeper than 100 km, while the path coverage remains the same, the noise in the path-average model is likely to increase as a smaller number of paths have good higher mode information and therefore the resolution will be slightly lower. Neglecting finite frequency effects has the consequence that the realistic resolution is slightly less than that recovered in these tests (Yoshizawa and Kennett, 2004).

The vertical resolution of the surface wave tomography is harder to estimate, due to the difficulty of tracking resolution in depth through the two-steps of the surface wave inversion (see Fishwick et al., 2005, for further details). It is worth noting that in the shallowest part of the models, there can be some smearing from crustal velocities, as the crustal model is fixed in the inversion. We use the global crustal model 3SMAC (Nataf and Ricard, 1996). The crustal model is incorporated into the path-average structure for the waveform inversion. Tests by Debayle and Kennett (2000) showed that there must be significant differences between the real crustal structure and 3SMAC, over a large portion of the path, for this to affect the output of the waveform inversion, and subsequently the tomographic models.

### 3. Results

#### 3.1. Depth slices

The models of the shear wavespeed variations are illustrated in Fig. 3. The models are plotted as perturbations from the 1D global reference model ak135 (Kennett, 1995). ak135 is chosen as it has a continental bias, in comparison to PREM (Dziewonski and Anderson, 1981), making it a good choice as the reference model for displaying the wavespeeds in the continental interior of Australia. The colour scale remains the same in all the images displayed, colour saturation occurs for wavespeed perturbations greater than $\pm 7.5\%$.

For the uppermost part of the lithosphere we present two adjacent models: at 75 and 100 km depth (Fig. 3a and b). At these
Fig. 3. Images of the isotropic shear wavespeed models from the tomographic inversion at: (a) 75 km, (b) 100 km, (c) 150 km, (d) 200 km, (e) 250 km depth. The models are plotted as perturbations from the global reference model ak135 (Kennett, 1995). The colour scale is the same for all images. On each image, the dashed lines indicate the outline of the three main cratonic blocks (Myers et al., 1996; Betts et al., 2002). (f) A map of the main geological units is also shown for reference (colour version available online).

At 75 km depth the West Australian Craton appears to be well defined by fast wavespeeds with the transition to slow velocities occurring near the Rudall Complex in the north and the Albany-Fraser in the southeast. A striking difference between these two images, is the large change in wavespeed in central Australia. Slow wavespeeds are modelled at 75 km depth beneath the area separating the three cratonic regions. By 100 km, however, fast wavespeeds, more typical of continental lithosphere, cover the majority of Australia.

The deeper models are plotted at 50 km intervals from 150 to 250 km depth. The reference velocities (ak135) are 4.51, 4.52 and 4.59 km/s, respectively. A zone of very fast wavespeed, with perturbations of around +7%, is observed in central Australia at 150 km depth. By 200 km depth this region is similar to the other areas showing fast wavespeeds, and by 250 km only weak velocity perturbations are imaged. To the north, regions with fast wavespeeds continuing to 250 km are observed beneath the Kimberley, the eastern Arunta and near the Tennant Creek Inlier. Further west there is contrasting structure beneath the Yilgarn and Pilbara Cratons, with the fast wavespeeds observed to a greater depth beneath the Yilgarn. The contrasting structure of the cratonic regions of west Australia is discussed by Fishwick et al. (2005).

One measure of the suitability of the tomographic images, is to compare the residuals between the data and an initial homogeneous velocity model, with the residuals of the data and the final model output from the tomographic inversion. For the models presented above the variance reduction is—75 km: 82.5%; 100 km: 86.4%; 150 km: 84.6%; 200 km: 69.2%; 250 km: 45.9%. The large variance reduction at depths between 75 and 150 km shows the models produce a very good fit to the data. It is also an indicator of the large variations in shear wavespeeds, in a region containing both oceans and old cratonic regions a homogenous velocity model cannot fit the data at these depths. The smaller variance reduction in the deeper part of the model is due to both the smaller perturbations in the data; a homogenous velocity model has a better fit to the data than at the shallowest depths, and also the decreas-
inversions, however, there is little difference. The variance reduction at 75 km depth is 84.2% and at 150 km depth is 84.7%. These are close to the values of the isotropic inversion, and could not be used to justify the incorporation of anisotropy. At 250 km depth the variance reduction of the anisotropic model is 54.8% almost 10% higher than the isotropic inversion; suggesting that the N–S direction of anisotropy is a robust feature.

An isotropic inversion with a finer parameterisation has also been tested. Using knot points separated by 1.5° rather than 2° leads to a small improvement in the data fit. However the greater number of unknowns (1.75×) is not justified by the small change in variance reduction. Due to the uncertainties as to whether the anisotropic model is justifiable, we continue to use the isotropic inversion with knot points at 2° intervals in the following section.

3.2. Cross sections

In order to aid the understanding of the structure of the features observed in the 2D models, we construct north–south cross sections through western and central Australia. The cross sections are computed through the isotropic model, so features would be slightly smoother if anisotropy had been included. Fig. 5 illustrates cross sections at 5° intervals from 120°E to 140°E. Covering a north–south profile of 25° the cross sections have a vertical exaggeration of approximately 1.8. Fig. 5a shows the cross sections as the perturbations from the reference model (ak135). As we are now comparing velocities at different depths, plotting the models as perturbations from a reference model places more importance on the velocity structure of the reference model. For example, in any location in our model, constant velocities with increasing depth could change from positive to negative perturbations, if there is a strong positive gradient in the reference model. For this reason, we also present the sections as the absolute velocities (Fig. 5b).

At 120°E the cross sections are predominantly through the Archaean regions of the West Australian Craton, and are worth discussing for comparison with cross sections further east. The differing depth extent of the fast wavespeeds beneath the Yilgarn and Pilbara is noticeable and the Capricorn Orogen appears to have a similar profile to the Pilbara, although slightly slower velocities at shallow depths. It is interesting to note that beneath the southern edge of the Yilgarn, to the south of 32°S, a distinct low velocity zone is observed at a depth of 150–250 km.

The cross sections at 125°E and 130°E pass through the regions of slowest wavespeeds in the shallow lithosphere, velocities can be lower than 4.3 km s⁻¹. Beneath the shallow layer of slow wavespeeds the fast wavespeeds, predominantly between 100 and 200 km depth, are observed. In the section at 130°E a low velocity layer at, or just below, 200 km is observed through the majority of the north and central part of the continental interior.

At 135° the cross section shows a different style of structure. Slightly slow wavespeeds are seen at the shallowest depths, with faster wavespeeds between 20°S and 30°S, seen at 100–200 km depth. From the perturbation cross section, it appears there are lower velocities, and perhaps a dipping structure beneath 200 km depth at approximately 25°. However, the absolute velocity section indicates that the reference model exaggerates the slow perturbations, while there are slightly lower wavespeeds the dipping structure is not as apparent. Fast velocities are observed in the deepest part of the model near 31°S and 22°S.

Another distinct change in style of cross section is observed at 140°E. A distinct low velocity zone at a depth of approximately 175–250 km depth is seen between 22°S and 29°S, with wavespeeds of about 4.4 km s⁻¹. To the north and south of this low velocity, slightly fast wavespeeds are continuous at all depths in the
cross section, although the perturbations from the reference model decrease with depth.

4. Discussion

The wavespeed perturbations that are observed in the upper mantle can be caused by variation in a number of physical properties and/or variation in chemical composition. Seismic wavespeeds are most sensitive to temperature (e.g. Duffy and Anderson, 1989), and the effect of anelasticity is particularly important (Karato, 1993). Anelasticity greatly increases the influence of temperature on shear wavespeed. The effect of composition is an area of considerable debate. The results of Griffin et al. (1999) suggests that compositional differences between Archaean and Phanerozoic subcontinental lithospheric mantle may lead to velocity differences of 5%. However, Goes et al. (2000) used these different compositions within their modelling of mantle temperatures and found that the compositional differences alone cause effects of less than 1%. Recent studies by Lee (2003) and Schutt and Lesher (2006) also give wavespeed perturbations of between 0.5 and 2% for variations in peridotite composition. Other factors such as water and partial melt (Karato and Jung, 1998; Hammond and Humphreys, 2000) and grain size (Paul and Jackson, 2005) can have a significant effect on the seismic wavespeed.

4.1. Shear wavespeed anomalies beneath Proterozoic Australia

4.1.1. Wavespeed variations with depth in the central region of Australia

In the shallow lithosphere, one of the most obvious features of the model is the large region of anomalously slow wavespeeds throughout much of central Australia. These slow velocities are underlain with faster wavespeeds from 100 km depth, creating a very strong positive velocity gradient in the uppermost mantle. From the tomographic images alone it is possible to ascertain the extent, but not the likely causes of the slow wavespeeds. A number of possible explanations are discussed below.

As temperature is often the dominant cause of wavespeed anomalies in the upper mantle it is reasonable to consider that this zone of slow wavespeeds may be a region of the mantle with higher temperatures. Within old continental lithosphere, a steady-state temperature profile can be estimated (e.g. Jaupart and Mareschal, 1999; McKenzie et al., 2005). The temperature profile is constrained by a number of factors: crustal heat production, mantle heat production, and transfer of heat at base of the lithosphere. The heat producing elements (HPE) of K, Th, Ur, are more abundant in the crust than the mantle. Rudnick et al. (1998) gave a preferred value for heat production in bulk-crustal compositions in the order of $0.6-1 \mu \text{W m}^{-3}$, while cratonic mantle heat production is estimated to be approximately $0.03 \mu \text{W m}^{-3}$. Using these thermal constraints Shapiro and Ritzwoller (2004) reject models with a positive velocity gradient in the shallow lithosphere, and similarly Faul and Jackson (2005) were not able to match increasing velocities in the uppermost mantle with their modelling of seismic wavespeeds. Therefore, due to the very strong positive shear wavespeed gradient, a steady-state continental geotherm is unable to explain the low velocities.

A second possibility relating to temperature, is high heat production in the crust, causing elevated mantle temperatures. Central Australia is a region with high heat production, the granites have an average of $4.6 \mu \text{W m}^{-3}$ (McLaren et al., 2003). However, it is generally considered that the majority of high heat producing ele-
ments will be found in the upper crust, as tectonic activity will progressively redistribute elements through the crust (Sandiford and McLaren, 2002). If HPEs were distributed lower in the crust during the Proterozoic it is unlikely they remain there in the present day. Any possible contribution from HPEs will be very dependent on their thickness and position in the crust and at what age they were redistributed into their present location. Modelling of temperature profiles suggests that it is possible to create positive temperature gradients in the upper mantle due to crustal heat production (D. McKenzie, 2006, pers. comm.). Whether this is possible over the large area containing low velocities, and with the tectonic history of the region, would require quantitative modelling, and thus a better estimate on the uncertainties of the absolute velocities.

The faster wavespeeds beneath this region extend to around 200 km depth. These velocities suggest that the temperature in this part of the mantle is low and the thick lithosphere in this region will limit the input of mantle derived heat into the crust. The low temperatures required to obtain the relatively thick lithosphere and fast wavespeeds are one of the reasons that indicate that high surface heat flow is predominantly controlled by crustal heat production (McLaren et al., 2003) rather than by heat transfer in the convecting mantle.

Composition can also be considered as an alternative physical property that could give the wavespeed structure beneath the region. Although it is not possible to get large variations in shear wavespeed due to simple peridotitic compositional differences. It is possible that a more exotic lithology could be a cause of the low wavespeeds. For example, serpentinitization of the mantle wedge at subduction zones is one area where much slower seismic velocities can be observed, with 20% serpentinitization reducing shear wavespeed velocities from about 4.6 to 4 km/s (Hyndman and Peacock, 2003; Christensen, 1966). However, most serpentine minerals are only stable at very low temperatures, and over a long tectonic history would be likely to have undergone dehydration reactions.

An alternative compositional lithology that could lead to slow seismic velocities is a large amphibole content. Theoretical calculations on the mineralogy around subduction zones shows that adding 25% paragasitic amphibole could lead to velocity decreases in the order of 5% (Hacker et al., 2003; Hacker and Abers, 2004). Paragasitic amphiboles are found in upper mantle-derived xenoliths, and due to worldwide occurrence, are recognised as a ubiquitous hydrous phase in the uppermost mantle (e.g. Dawson and Smith, 1982). In comparison to serpentinite, which is constrained to a narrow zone in the mantle wedge and very low temperatures, paragasite can be found in a larger region, being stable to 30 kbar at a temperature of 925 °C (Niida and Green, 1999).

If a lithology containing large volumes of paragasite was the cause of the low velocities, it is not simple to explain its present location. The upper mantle xenoliths containing abundant amphibole are from lherzolites, rather than from ancient sub-continental lithospheric material. Beneath the region of slow wavespeeds, faster velocities more typical of continental lithosphere are found, suggesting that if the slow wavespeeds are caused by composition a layered structure has formed during some time interval in the tectonic evolution of central Australia.

A third possible cause of low velocities in the upper mantle is a region of partial melt. Recent experimental work shows that partial melt will have a strong impact on both wavespeed and attenuation (e.g. Faul and Jackson, 2005). However, this would also be related to very high temperatures, and for the reasons mentioned above, this is unlikely. Furthermore, the increased attenuation due to melt is not observed in the body wave results of Kaiho and Kennett (2000).

It is worth noting that due to the uncertainties in the a priori crustal model (see Section 2.3) and their effect on the inversion results, we are not attempting to quantitatively model the absolute velocities obtained for the tomography with exact temperatures and compositions; the discussion remains at a qualitative level. Although some of the anomaly could be caused by smearing of crustal structure, the region of low velocities predominantly trends...
EW in central Australia and then extend towards the NW into the Canning Basin. In contrast, the regions of thickest crust in the models of Clitheroe et al. (2000) and Collins et al. (2003) follows a predominantly NS trend from central Australia into the North Australian Craton, with very thick crust also observed beneath Mt Isa. The crust beneath the Canning Basin is much thinner.

An entirely independent dataset also confirms that slow wavespeeds are observed in the uppermost mantle of central Australia. Fig. 6 illustrates the arrival times of S waves from less than 20° (Kaiho and Kennett, 2000). In the central region, positive residuals, and thus slow propagation relative to the reference model ak135 (Kennett, 1995), are observed for paths from earthquakes near Tenant Creek to stations to the south and southwest. These paths are sensitive to the uppermost mantle structure, and indicate that the slow wavespeeds in the surface wave models are not simply an artifact of smearing crustal velocities.

A major conclusion from the velocity anomalies, is that they cannot be explained by a typical steady-state continental geotherm. This conclusion does not only relate to the slow wavespeeds, but also to the very strong positive velocity gradient observed in the uppermost mantle. Even if the wavespeeds at 75 km depth were increased by 2–3%, e.g., due to any systematic bias from the crust, the positive velocity gradient remains and the conclusion holds.

We are not aware of any other similar observations, from surface wave studies, of slow wavespeeds in the shallow lithosphere observed in tectonic settings worldwide which might correspond with that of central Australia. Results from travel time tomography within the Archaean and Proterozoic region of Wyoming and Colorado, indicate a low wavespeed anomaly overlying a north-dipping fast wavespeed anomaly (Yuan and Dueker, 2005). The authors interpret the dipping structure as a Proterozoic slab remnant, with the variations in wavespeed predominantly caused by chemical heterogeneities. The results from the study of Yuan and Dueker (2005) are along a 2D profile, and show relative velocity anomalies. It is therefore difficult to compare the results with the more extensive region of low velocities observed within Australia. Future work will be aimed at quantitatively modelling the wavespeeds within central Australia. Further development of the tomographic technique with the use of Monte-Carlo methods (e.g. Sambridge and Mosegaard, 2002) to give a better estimate of the uncertainties is required before this can be considered.

4.1.2. Fast velocities beneath the Capricorn Orogen

In contrast to central Australia, the mantle beneath the Capricorn Orogen shows fast shear wavespeeds at shallow depths. In this region, the wavespeeds of around 4.7 km/s (a +5% variation from the reference model) appear to be typical of fast cratonic lithosphere, and can be reasonably explained by low temperatures beneath the Moho. Using the results comparing velocities of peridotitic composition (e.g. Schutt and Lesher, 2006) it is possible that a small percentage (%1) of the anomaly could be related to the increase in velocity due to the chemical composition of more depleted lithospheric mantle.

From the cross sections (Fig. 5a and b) at 120° E we can see that the velocities beneath the Capricorn Orogen are slightly lower than beneath either the Yilgarn or Pilbara Cratons, and it appears that the depth extent of the fast wavespeeds is similar to the Pilbara and significantly less than beneath the Yilgarn. In fact the depth slices (Fig. 3) suggest that the western edge of the Capricorn Orogen has an even thinner region of fast wavespeeds.

At depths of 150–200 km the eastern edge of the Orogen, near the Rudall Complex shows similar fast wavespeeds as observed in the region of central Australia. Therefore, one possible explanation for the change in structure from west to east beneath the Capricorn is that the region of lithosphere beneath the eastern edge of the Orogen are also affected by the collision of the West and North Australian Cratons as well as that of the Pilbara and Yilgarn. At depths of 150–200 km it appears that it is the later collision of the larger scale cratonic regions that dominates the seismic structure.

4.1.3. Contrasts in velocities within the North Australian Craton

The North Australian Craton is believed to be an amalgamation of various blocks of Proterozoic or Archaean age (Betts et al., 2002), and the results of the surface wave tomography suggest that there are distinct variations in the lithospheric structure beneath the region.

The depth slices (Fig. 3d and e) suggest that the lithosphere beneath the Kimberley Block continues to a greater depth than the regions immediately adjacent. These results remain in agreement with the previous work of Fishwick et al. (2005). We interpret the deep anomaly to be related to older Archaean lithosphere, with the younger Proterozoic orogenic belts, which surround the Kimberley Block, having a thinner lithosphere. The presence of thick lithosphere in this region is also in agreement with isotopic data from diamondiferous kimberlites (Graham et al., 1999), and at a regional scale the edge of the thicker lithosphere may also be a control on the occurrence of diamond deposits (Jaques and Milligan, 2004).

Beneath the eastern Arunta, and extending north close to the edge of the Tennant Creek Block there is a region where fast wavespeeds extend to depths of 200–250 km. Diamond deposits are also observed close to the northern limit of this zone of fast wavespeeds (Jaques and Milligan, 2004). The occurrence of similar wavespeeds in this region in comparison to the wavespeeds observed beneath the Yilgarn and Kimberley suggests there may be an Archaean lithospheric root. It is also possible that the thick roots may not simply be a feature of Archaean mantle, but also relate to regions of early Proterozoic mantle as well. Further dating of mantle xenoliths is required before strong conclusions can be drawn on the age relationship.

In the western Arunta the fast wavespeeds are only observed to a depth of around 150 km. A zone of lower velocities appears to extend northwards towards the eastern edge of the Kimberley Block. The contrasting wavespeed structure beneath the eastern and western Arunta indicates that there is not always a simple correlation between the overlying surface geology and the structure of the lithospheric mantle beneath. The variations in structure may relate to both the formation of the lithosphere, and also the different processes that have acted on different regions of the lithosphere over geological time.

4.2. Summary

The new surface wave tomography model of Australia clearly shows a region of slow shear wavespeeds in the uppermost mantle beneath central Australia. This agrees with observations from an independent set of body wave data (Kaiho and Kennett, 2000). At 75 km depth, the location of this region of slow wavespeeds is well constrained between the three main cratonic blocks, and the boundary with the West Australian Craton is particularly evident. Beneath the region of slow wavespeeds, fast wavespeeds are observed between 100 and 200 km depth. The combination of slow and fast velocities creates a very strong positive velocity gradient in the uppermost mantle.

Most velocity variations in the upper mantle are commonly attributed to temperature. Within central Australia it is not possible to link typical steady-state continental geotherms to the strong velocity gradient and slow wavespeeds in the shallowest part of the model. There are two main alternatives that can explain the observed wavespeeds. It is possible that the impact of high heat production in the crust is being seen, this is very dependent on when
the HPEs have been redistributed within the crust, but can lead to low velocities in the shallowest mantle and a positive velocity gradient. An alternative explanation requires compositional variations. The low velocities indicate something other than a peridotitic composition, and one possibility is the presence of large quantities of amphibole within the lithosphere. Whatever the cause of the velocity anomaly, its location in central Australia, at the position where the three main cratonic blocks amalgamated (Myers et al., 1996; Betts et al., 2002), and where there has since been intraplate tectonic activity, suggests that the prolonged history of continental collision and subsequent deformation has created a highly unusual lithospheric structure. It is hoped that the new models presented will stimulate further debate as to the evolution of the lithosphere beneath the Proterozoic surface geology of north and central Australia.

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