Crustal structure of Precambrian terranes in the southern African subcontinent with implications for secular variation in crustal genesis

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Summary

New estimates of crustal thickness, Poisson’s ratio and crustal shear wave velocity have been obtained for 39 stations in Angola, Botswana, the Democratic Republic of Congo, Malawi, Mozambique, Namibia, Rwanda, Tanzania and Zambia by modelling P-wave receiver functions using the H-κ stacking method and jointly inverting the receiver functions with Rayleigh-wave phase and group velocities. These estimates, combined with similar results from previous studies, have been examined for secular trends in Precambrian crustal structure within the southern African subcontinent. In both Archean and Proterozoic terranes we find similar Moho depths [38–39 ± 3 km SD (standard deviation)], crustal Poisson’s ratio (0.26 ± 0.01 SD), mean crustal shear wave velocity (3.7 ± 0.1 km s⁻¹ SD), and amounts of heterogeneity in the thickness of the mafic lower crust, as defined by shear wave velocities ≥4.0 km s⁻¹. In addition, the amount of variability in these crustal parameters is similar within each individual age grouping as between age groupings. Thus, the results provide little evidence for secular variation in Precambrian crustal structure, including between Meso- and Neoarchean crust. This finding suggests that (1) continental crust has been generated by similar processes since the Mesoarchean or (2) plate tectonic processes have reworked and modified the crust through time, erasing variations in structure resulting from crustal genesis.

Key words: Cratons; Crustal structure; Africa.

1 Introduction

Whether present-day processes of crustal formation were also the most dominant processes in the first 1–2 billion years of Earth’s history remains an unresolved question. Knowledge about the nature of Precambrian crust obtained from seismic data provides key constraints on geodynamic models of crustal genesis and evolution, and is therefore important for addressing questions about secular variation in crustal formation. Some studies going back to the early 1990s using compilations of seismic refraction data (e.g. Durheim & Mooney 1991, 1994), have suggested that there may be significant differences in crustal structure between Archean and Proterozoic terranes, indicating secular changes in crustal genesis, while others (e.g. Rudnick & Fountain 1995; Rudnick & Gao 2003) have argued for little change in crustal structure throughout the Precambrian. This debate has continued with more recent studies. For example, Thompson et al. (2010) using seismic images of crustal structure in the Canadian Shield, and Abbott et al. (2013) using previously reported seismic models of crustal structure in southern Africa, Australia and Canada, have argued for secular evolution in Precambrian crustal genesis. On the other hand, Tugume et al. (2013), using seismic images of crustal structure in eastern Africa and Stankiewicz & De Wit (2013), using swaths of seismic data across southern Africa, find no discernible differences in Archean and Proterozoic crustal structure.

In this study, we report new estimates of crustal structure (crustal thickness, shear wave velocity and Poisson’s ratio) across southern and eastern Africa, and use them in combination with other published results for the southern Africa subcontinent to investigate whether the conclusions reached by Tugume et al. (2013) for African crustal structure hold for a much larger dataset. Not only does our study expand on the number of Precambrian terranes examined in Africa, but it also allows for the comparison of continental crust pre- and post-3.0 Ga (i.e. Mesoarchean versus Neoarchean crust).

The new estimates of crustal structure come from Angola, Botswana, the Democratic Republic of Congo, Malawi, Mozambique, Namibia, Rwanda, Tanzania and Zambia. Estimates of Moho depth, crustal shear wave velocities and Poisson’s ratio for 39 new broad-band stations in those countries have been determined by modelling P-wave receiver functions using the H-κ stacking method (Zhu & Kanamori 2000) and a joint inversion of P-wave receiver functions and Rayleigh-wave dispersion measurements (Julià et al. 2000, 2003). Results from the previous studies were obtained by modelling P-wave receiver functions using similar methods.
including the $H$--$k$ stacking method (Midzi & Ottemoller 2001; Nguuri et al. 2001; Dugda et al. 2005; Nair et al. 2006; Gallacher & Bastow 2012; Tugume et al. 2012; Youssouf et al. 2013), the slant-stacking method (Last et al. 1997) and joint inversion methods involving $P$-wave receiver functions and Rayleigh-wave dispersion curves (Julià et al. 2005; Kgaswane et al. 2009; Tokam et al. 2010; Tugume et al. 2013). The data ensemble used in this study includes estimates for many Archean and Proterozoic terranes, allowing for a more comprehensive examination of Precambrian crustal structure in the southern African subcontinent than was done by Tugume et al. (2012, 2013).

As a point of clarification, throughout this paper Moho depth and crustal thickness are used interchangeably in reference to the total crustal thickness.

2 BACKGROUND GEOLOGY

The southern African subcontinent is an amalgamation of many Precambrian terranes (Fig. 1), including several Archean cratons and Proterozoic orogenic belts that have not been affected by any significant Phanerozoic tectonic activity. In this section, the geology of each terrane for which there are seismic estimates of crustal structure is briefly reviewed. A short summary of the crustal structure for each terrane from previous studies is provided in the Supporting Information.

2.1 Eastern Africa

The Archean Tanzania Craton forms the nucleus of the Precambrian tectonic framework in eastern Africa (Cahen et al. 1984). The craton is made up of two blocks separated by the Dodoma Schist Belt. The northern Nyanzian Block is comprised of greenstone belts and granites ranging in age from 2.80 to 2.66 Ga, overlain by a molasse that is intruded by 2.60 Ga granitoids. The southern Dodoman Block consists of 2.93–2.85 Ga granodiorites, granitic gneisses, migmatites and other high-grade metamorphic rocks (Cahen et al. 1984; Schultet 1997; Manya & Maboko 2003; Begg et al. 2009 and references therein).

The Tanzania Craton is surrounded by many Proterozoic mobile belts. The Palaeoproterozoic Usagaran Belt along the southeastern margin of the Tanzania Craton contains mostly supracrustal rocks metamorphosed to granulite facies at 1.92–1.89 Ga and 1.87–1.83 Ga granitoids that were partially derived from reworking and recycling of the Tanzania Craton (Cahen et al. 1984; Schultet 1997; De Waele et al. 2008; Begg et al. 2009 and references therein). To the southwest of the Tanzania Craton is the Palaeoproterozoic Ubendian Belt. It consists of granulite and amphibolite facies gneisses and metasedimentary rocks that formed during two orogenic events (~2.03–1.97 and ~1.93–1.86 Ga, respectively, Cahen et al. 1984; Lenoir et al. 1994; Schultet 1997; Begg et al. 2009 and references therein). Along the western side of the Tanzania Craton is the Mesoproterozoic Kibaran Belt. The north–south oriented terrane is comprised of amphibolite grade rocks formed during the Kibaran Orogeny (1.4 Ga), with intrusions of granites, ultramafic bodies and granitoids at 1.39, 1.3 and 1.0 Ga, respectively (Klerkx 1987; Begg et al. 2009 and references therein). The Mesoproterozoic Rwenzori Belt is a deformed volcano-sedimentary sequence (1.86–1.78 Ga) located to the northeast of the Kibaran Belt (Begg et al. 2009 and references therein). East of the Tanzania Craton is the Neoproterozoic Mozambique Belt. The Mozambique Belt is the longest terrane in Africa, extending along the east African coast from Mozambique to southern Egypt (Schultet 1997). The northern part of the belt consists of reworked Archean continental crust, while the southern part (most of Malawi and northern Mozambique) consists of Palaeoproterozoic to Neoproterozoic basement gneisses (Begg et al. 2009 and references therein). The Mozambique Belt is believed to be a continental collision zone formed by multiple collision events between 1200 and 450 Ma (Cahen et al. 1984; Shackleton 1986).

2.2 Central Africa

Central African Precambrian geology is dominated by the Congo Craton. The Congo Craton is composed of Mesoproterozoic to Palaeoproterozoic mobile belts amalgamated during the assembly of Gondwana (Goodwin 1996; De Waele et al. 2008), with regions of extensive Phanerozoic cover (De Waele et al. 2005). Archean basement rock is exposed in four shields around the edges of the craton (Begg et al. 2009 and references therein) but only three are relevant to this study. The Kasai Block on the southeastern edge of the Congo Craton is a heterogeneous 3.01 Ga granulite complex, metamorphosed between 2.9 and 2.6 Ga, and overlain by supracrustal rocks, which were later cut by 2.1 Ga granite plutons (Begg et al. 2009 and references therein). The Angolan Block on the southwest side consists of gneisses, metasediments, a 2.82 Ga gabbro-charnockite complex metamorphosed at 2.8–2.7 Ga and 2.6 Ga granitic intrusions (Begg et al. 2009 and references therein). The Ntem Complex, the northern part of the Gabon-Cameroon Shield in the northwest corner of the craton, consists of 3.22–2.99 Ga gneiss, gneisses and remnants of greenstone belts cut by 2.94–2.54 Ga granites (Begg et al. 2009 and references therein). The Ntem Complex was later reworked in the Palaeoproterozoic, during the Eburnean Orogeny (Tchameni et al. 2001).

Bordering the Congo Craton are several Proterozoic terranes. On the western side of the Congo Craton is the Palaeoproterozoic West Central African Craton. The terrane evolved between 2.5 and 2.0 Ga during the Eburnean Orogeny and contains reworked Archean rocks, metamorphosed and intruded by 2.1–1.92 Ga granitoids and overlain by rift-related sequences that may be connected with the break-up of Rodinia (Begg et al. 2009 and references therein). North of the Congo Craton lies the Neoproterozoic Oubanguides Belt which resulted from the collision between the Congo Craton, the Sao Francisco Craton and the West African Craton during the formation of Gondwana (Castaing et al. 1994; Totene et al. 2004). The mobile belt consists of 1.7–0.90 Ga sedimentary rocks over a 2.0 Ga basement (Begg et al. 2009 and references therein).

The Neoarchean Bangweleu Block lies to the southeast of the Congo Craton and is composed of crystalline basement of schist belts intruded by 2.73 Ga granitic and metavolcanic rocks (De Waele et al. 2006; Begg et al. 2009 and references therein). The terrane is considered to be a Neoarchean microcontinent that was strongly affected by tectonic events in the surrounding mobile belts (Begg et al. 2009 and references therein). To the southeast of the Bangweleu Block is the Mesoproterozoic Irumide Belt. The terrane consists of reworked Archean and Palaeoproterozoic basement rocks, overlain by 1.88 Ga rhyolites and sediments (Begg et al. 2009 and references therein). During the Irumide Orogeny at 1.02 Ga, the terrane was strongly deformed by isoclinal folding and thrusting, which resulted in crustal thickening without the addition of juvenile crust (De Waele et al. 2005; De Waele et al. 2006). Further south of the Irumide Belt is the Southern Irumide Belt. Composed of Mesoproterozoic gneisses, this mobile belt is interpreted as a stack of
arc-related terranes imbricated during the Irumide Orogeny (Begg et al. 2009 and references therein). The Zambezi Belt, south of the Irumide Belt, is a Neoproterozoic terrane that contains reworked Archean and Proterozoic gneisses, overlain by strongly deformed metasediments and granites formed during the amalgamation of Gondwana (Begg et al. 2009 and references therein). The mobile belt was strongly deformed between 870 and 820 Ma during the Pan-African Orogeny (Begg et al. 2009 and references therein).
Located on the southwest side of the Bangweleu Block is the Neoproterozoic Lufilian Arc. The terrane is composed of reworked Archean to Proterozoic basement overlain by Neoproterozoic volcanics, all thrust over the edge of the Congo Craton at 560–550 Ma during the assembly of Gondwana (Begg et al. 2009 and references therein).

2.3 Southern Africa

Two cratons are found in southern Africa. The Zimbabwe Craton consists of several terranes that formed between 3.6 and 2.4 Ga (Dirks & Jelsma 2002). The central Tokwe Gneiss terrane (East Tokwe Block) formed at 3.6–3.3 Ga and contains mafic fragments. Clastic sediments accreted against the western part of the Tokwe Gneiss terrane (West Tokwe Block) between 3.2 and 2.8 Ga and greenstone belts formed between 2.7 and 2.6 Ga, marking the stabilization of the craton. The last major tectono-thermal event affecting the craton was the Great Dyke around 2.58 Ga (Jelsma & Dirks 2002). The Kaapvaal Craton is a conglomerate of several terranes and predominantly consists of granitoids with gneisses and narrow greenstone belts that formed between 3.7 and 2.7 Ga. (De Wit et al. 1992; Eglington & Armstrong 2004; Begg et al. 2009 and references therein). The Swaziland terrane (>3.2 Ga) and the Witwatersrand terrane sutured together around 3.2 Ga, followed by the joining of the Pietersburg and Kimberley terranes between 3.0 and 2.8 Ga. The craton was later affected by several tectono-thermal events, including the Dominion (3.1 Ga), Witwatersrand (3.0–2.8 Ga), Ventsdorp (2.7 Ga), Transvaal (2.6–2.2 Ga) and Waterberg (2.0–1.8 Ga), and the emplacement of the Bushveld Complex (2.05 Ga, Eglington & Armstrong 2004; Johnson et al. 2006).

The Archean Limpopo Belt is an east–west trending zone of high-grade metamorphic rocks that separates the Zimbabwe Craton from the Kaapvaal Craton and probably formed during a 2.7–2.6 Ga collision between the two cratons (McCourt & Armstrong 1998; Kramers et al. 2006; Begg et al. 2009 and references therein). The terrane’s 3.3–3.1 Ga gneisses were affected by granulite facies metamorphism and granitoid magmatism at 2.7–2.57 Ga (Begg et al. 2009 and references therein). Peak metamorphism was at ca. 2.7 Ga with some metamorphism between 2.06 and 2.0 Ga linked with the Bushveld Complex (Begg et al. 2009 and references therein).

Several Proterozoic mobile belts encircle the Zimbabwe and Kaapvaal cratons. The Palaeoproterozoic Kheis-Okwa Belt runs along the western edge of the Kaapvaal Craton (Cornell et al. 2006; Begg et al. 2009). The Kheis terrane is a thin-skinned, eastward thrusting belt composed of 1.98 Ga basaltic and clastic sediments (Cornell et al. 2006; Begg et al. 2009). They were folded and metamorphosed at ca. 1.9 Ga and intruded by granite and mafic rocks between 1.27 and 1.12. The Okwa terrane has ca. 2.1 metamorphic basement rocks that may be underlain by Archean rocks (Cornell et al. 2006; Begg et al. 2009). Striking northeast–southwest to the east of the Kaapvaal Craton is the Palaeoproterozoic Rehoboth Province. The province consists of 1.79–1.73 Ga gneisses and migmatites (Begg et al. 2009 and references therein). To the south of the Kaapvaal Craton is the Mesoproterozoic Namaqua-Natal Belt. The fold belt is comprised of igneous and supracrustal rocks that accreted against the craton during the Namaqua Orogeny (1.2–1.0 Ga, Cornell et al. 2006). The Damara Belt runs northeast-southwest between the Congo Craton and the Rehoboth Province. This Neoproterozoic terrane is highly complex. The northern part consists of Neoarchean basement inliers overlain by 750 Ma rift sequences, while the southern part represents a passive margin folded and metamorphosed from 690 to 485 Ma during the assembly of Gondwana (Begg et al. 2009 and references therein).

3 DATA AND METHODS

3.1 Data

Data recorded on a number of networks were used for the receiver function analysis (Fig. 2). Data were used from 15 seismic stations in the AfricaArray East Africa Seismic Experiment phase 3 network in Zambia (ZP network code), which operated between August 2010 and July 2011, four stations in the AfricaArray Mozambique Network (XV network code), which operated between August 2011 and August 2013, one station each in Angola, Namibia and Botswana supported by an industry consortium between 2007 and 2010 (Congo Craton network), and three stations in Angola that belong to the National Meteorological Institute of Angola (INAMEt network). In addition, data from 11 permanent AfricaArray stations (AF network code) and two IRIS/GSN stations were used. Each station included a broad-band seismometer, a 24-bit data logger and a Global Positioning System clock, and the data were recorded continuously at either 20 or 40 samples per second.

For the joint inversion of the receiver functions and surface wave dispersion measurements, fundamental-mode Rayleigh-wave phase velocities at periods ranging from 20s to 182s were taken from O’Donnell et al. (2013) and fundamental-mode Rayleigh-wave group velocities at periods ranging from 10s to 105s were taken from Raveloson et al. (2015).

3.2 P-wave receiver functions

Receiver functions are a time series that show the response of Earth’s structure beneath the recording station (Langston 1979) and are commonly used for investigating crustal structure. The main phases in the waveform are the direct P wave, the P-to-S conversion at the Moho (Ps) and its reverberations between the Moho and the free surface (PpPs and PsPs + PpSs, Fig. 3). The amplitudes and arrival times of the phases place valuable constraints on crustal structure beneath the receiver (Langston 1979; Ligoria & Ammon 1999) and can be used to estimate the thickness of the crust and the bulk crustal P-wave velocity/shear wave velocity (VP/VS) ratio (e.g. Zandt & Ammon 1995; Zhu & Kanamori 2000).

P-wave receiver functions were computed for each station using teleseismic events with magnitudes equal to 5.5 and greater and at epicentral distances between 30° and 90° (Fig. 4, see Supporting Information for list of events). For calculating the P-wave receiver functions, the originally recorded seismograms were de-trended, tapered, high pass filtered above 0.05 Hz to remove low frequency noise from the instrument response, and low pass filtered below 8 Hz. Afterwards, the waveforms were decimated to 10 samples per second to avoid aliasing, and cut to 10s before and 110s after the first P-wave arrival. Next, the horizontal components were rotated to the great circle path to obtain radial and transverse components. The vertical component was then deconvolved from the radial and tangential components using the iterative time-domain deconvolution method (Ligoria & Ammon 1999). A maximum of 500 iterations were used in the deconvolution method (Ligoria & Ammon 1999).

For each teleseismic event, radial and tangential receiver functions were computed for a Gaussian width factors of 1.0 (f ≤ 0.5 Hz)
Figure 2. Topographic and geologic map of the southern African subcontinent showing the locations of the temporary and permanent broad-band seismic stations used in this study. Precambrian terrane boundaries are the same as in Fig. 1. Descriptions of the networks not provided in this paper can be found in Tugume et al. (2013), Tokam et al. (2010), Kgswame et al. (2009), Nair et al. (2006), Dugda et al. (2005) and Last et al. (1997). Seismic stations are labelled for which new estimates of crustal structure are provided in this study.
and 2.5 ($f \leq 1.25$), typical for receiver function analysis. Lower frequencies result in longer wavelength receiver functions, which are better for observing long period phases from the lower crust and mantle, while higher frequency receiver functions can reveal detailed phases from shallow crustal structure (Owens & Zandt 1985; Ligorria & Ammon 1999; Julià 2007).

Next, the receiver functions were assessed for quality using a least squares minimization of the difference between the original radial component and the predicted radial component. The predicted radial component was generated by the convolution of the original vertical component and the radial receiver function. A similarity of 85 per cent or greater was used to select receiver functions for further analysis. In addition, events that resulted in transverse receiver functions with large amplitudes were not considered for further processing, even if they passed the 85 per cent criterion. Fig. 5 illustrates the quality of receiver functions at one station.

### 3.3 $H$–$\kappa$ stacking method

Crustal thickness estimated solely from the delay time between the Moho Ps phase and direct P-wave trades off strongly with the crustal $V_p/V_s$ ratio. Therefore, the $H$–$\kappa$ stacking method of Zhu & Kanamori (2000) was applied to the receiver functions to obtain estimates for crustal thickness ($H$) and bulk $V_p/V_s$ ratio ($\kappa$). The $H$–$\kappa$ stacking method reduces the ambiguity in $H$ and $\kappa$ by incorporating the later multiple converted phases from the Moho (PpPs and PsPs + PpSs). In this method, the receiver functions are transformed to the $H$–$\kappa$ parameter space by an objective function:

$$s(H, \kappa) = \sum_{j=1}^{N} w_j r_j(t_1) + w_j r_j(t_2) - w_j r_j(t_3),$$

where $t_j$ are the traveltimes of the three main $P$-to-$S$ converted phases from the Moho (Ps, PpPs and PsPs + PpSs), $w_j$ are weights assigned to each phase (sum of $w_j = 1$), $r_j$ is the receiver function amplitude for the $j$th receiver function and $N$ is the number of receiver functions. $S(H, \kappa)$ reaches its maximum when optimal values for $H$ and $\kappa$ are determined, satisfying a simple layer over a half-space crustal model. The values of $H$ and $\kappa$ are taken as estimates for the Moho depth and $V_p/V_s$ ratio near the receiver (Zhu & Kanamori 2000).

In order to apply the $H$–$\kappa$ stacking method to the receiver functions, weights for the converted phases (eq. 1) and an average crustal
A $V_p$ of 6.5 km s$^{-1}$ was chosen because it is consistent with average crustal P-wave velocities determined from previous studies of Precambrian crust in the area (Fuchs et al. 1997; Julià et al. 2005; Tugume et al. 2012 and references therein). A weighting system of $w_1 = 0.4$, $w_2 = 0.3$ and $w_3 = 0.3$ was employed to place equal weight on all the Moho phases. For a few stations (CHAM, GETA, KTWE, LSZ, MTVE, MWEN, MZM and PORTQ) the third phase (PsPs + PpSs) was not clearly observed and so the weights were readjusted to $w_1 = 0.5$, $w_2 = 0.5$ and $w_3 = 0.0$.

Results from the $H$-$\kappa$ stacking method for crustal thickness and Poisson’s ratio obtained from the $V_p$/$V_s$ ratio are tabulated in Table 1, and an example for one station is illustrated in Fig. 6 (see Supporting Information for results from all stations). Receiver functions with a Gaussian filter of 1.0 were used for all stations except MTVE, MWEN, MKUS and MUFZ, where a Gaussian of 2.5 was used. For these stations the Moho converted phases were more clearly observed at higher frequencies than at lower frequencies. The $H$-$\kappa$ stacking method was performed on both 1.0 and 2.5 Gaussian filters for all stations, and similar estimates were obtained.

Formal uncertainties for $H$ and $\kappa$ were estimated using a bootstrapping method to repeat the summation procedure 200 times with random, resampled data from the original dataset (Efron & Tibshirani 1991). Additional uncertainties in $H$ and $\kappa$ arise from the chosen mean crustal $V_p$. Therefore, the $H$-$\kappa$ stacks were recomputed for $P$-wave velocities of 6.3 and 6.7 km s$^{-1}$ to place lower and upper error bounds on $H$ and $\kappa$ for a range of crustal $P$-wave velocities. The combination of the two methods yields an overall uncertainty in the Moho depth at each station that is $\sim 3$-4 km. The formal uncertainties obtained from using a mean crustal $V_p$ of 6.5 km s$^{-1}$ are given in Table 1.

For two stations (KISZ and SENA), the crustal multiples (PsPs and PpSs) were not clearly seen and the $H$-$\kappa$ stacking method could not constrain the crustal thickness and $V_p$/$V_s$ ratio. Therefore, the Moho depth was estimated using the Moho Ps arrival time, an assumed crustal $V_p$ of 6.5 km s$^{-1}$, an assumed Poisson’s ratio of 0.26 and eq. (2) from Zandt & Ammon (1995). The receiver functions for these stations can be found in the Supporting Information and the estimated depths are reported in Table 1.

### 3.4 Joint inversion of Rayleigh-wave dispersion curves and receiver functions

Crustal shear wave velocity models were obtained for each station by jointly inverting the receiver functions with Rayleigh-wave phase and/or group velocities. Receiver functions constrain shear wave velocity contrasts at interfaces between mediums, and single mode Rayleigh-wave dispersion curves place constraints on averages of the absolute shear wave velocity at depth. Therefore, the combination of the two datasets provides tighter constraints on the shear wave velocity structure at depth and bridges the resolution gaps inherent in each dataset individually. The joint inversion method used is a linearized, dumped least squares method developed by Julià et al. (2003) that incorporates a priori constraints.

To perform the inversion, receiver functions and dispersion measurements that have sampled the same area near a station are first selected. The receiver functions are binned and stacked by similar ray parameter in the groups 0.04–0.049, 0.05–0.059, 0.06–0.069 and 0.07–0.079, to account for moveout from differing incident angles. The groups used for a particular station depend on the spread of ray parameters associated with the receiver functions for that station, and so the ray parameter groups vary from station to station. For each station, receiver functions were binned and stacked for two sets of overlapping frequency bands corresponding to Gaussian bandwidths of 1.0 ($f \leq 0.5$ Hz) and 2.5 ($f \leq 1.25$ Hz). Inverting receiver functions at several frequency bandwidths help distinguish sharp discontinuities from gradational ones in the velocity models (Cassidy 1992; Julià et al. 2005; Julià 2007).

The starting model used in the joint inversion is an isotropic medium with a 38-km-thick crust and a linear shear wave velocity increase in the crust from 3.4 to 4.0 km s$^{-1}$. The crust lies over a flattened Preliminary Reference Earth Model (PREM) for the mantle (Dziewonski & Anderson 1981). The mantle shear wave velocity structure is modelled to 290 km and then constrained to be equal to 5 per cent lower than PREM to better fit the longer period dispersion velocities, for stations in Malawi, Rwanda, Tanzania and Zambia within or near to the East African rift system (Julià et al. 2005). For stations in Angola, Botswana, the Democratic Republic of Congo, Mozambique and Namibia away from the East African rift system, the shear wave velocity below 290 km depth is constrained to PREM (Julià et al. 2005). The model is parametrized with layers of constant velocity that increase with depth. Layer thicknesses are 1 and 2 km at the top of the model, 2.5 km between 3 and 65 km depth, 5 km between 65 and 265 km depth and 10 km below a depth of 265 km. Poisson’s ratio was fixed at 0.25 for crustal layers and at PREM values for mantle layers.

Uncertainties in the velocity models were estimated using the approach in Julià et al. (2005) by repeating inversions for a range of parameters, constraints and Poisson’s ratios. We found uncertainties in the crustal shear wave velocity to be about 0.1 km s$^{-1}$ and about 0.2 km s$^{-1}$ in the upper mantle. These uncertainties in the velocity translate to uncertainties of $\pm 2.5$ km in crustal thickness.

Results from the joint inversion technique are tabulated in Table 1 and the results for one station is illustrated in Fig. 7. Results for all stations can be found in the Supporting Information. Shear wave velocities for typical lower crustal lithologies derived from experimentally determined $V_p$/$V_s$ ratios have shown that the shear wave velocities in the lower crust do not likely exceed 4.3 km s$^{-1}$ and that velocities greater than 4.3 km s$^{-1}$ are typical for mantle lithologies (Christensen & Mooney 1995; Christensen 1996). Therefore, following the approach of Kgaswane et al. (2009) for southern Africa, Tokam et al. (2010) for western Africa and Tugume et al. (2013) for eastern Africa, in the joint inversion models the Moho is defined as the depth at which the shear wave velocity exceeds 4.3 km s$^{-1}$. For most stations there is a significant increase in velocity or a velocity discontinuity at the depth where the velocity exceeds 4.3 km s$^{-1}$, except for stations MKUS, MPIK, SERJ, MZM, KTWE and LSZ, where the shear wave velocity increase is gradational from the lower crust to the upper mantle.

We also obtain estimates of the thickness of the mafic layering in the lower crust from the joint inversion models. Previous studies that have examined continental crustal structure (e.g. Holbrook et al. 1992; Christensen & Mooney 1995; Rudnick & Fountain 1995; Rudnick & Gao 2003) have reported that common lower crustal mafic lithologies, such as amphibolites, garnet-bearing and garnet-free mafic granulate and mafic gneiss, have high shear wave velocities ($\sim 3.9$ km s$^{-1}$) while intermediate-to-felsic lithologies have lower shear wave velocities ($< 3.9$ km s$^{-1}$). Therefore we define the mafic lower crust as layers with shear wave velocities between 4.0 and 4.3 km s$^{-1}$.

Within our reported uncertainties, we find a 1-to-1 correlation between our Moho estimates from the $H$-$\kappa$ stack method and our Moho estimates from the joint inversion method, for all but a few
The following table presents the crustal structure results for the stations used in the study. The data includes information on tectonic regions, stations, latitude, longitude, elevation, Poisson’s ratio, and various depths and velocities. For example, the depth from stacking P-wave receiver functions is labeled as Depth (1), and the depth from jointly inverting P-wave receiver functions and surface wave dispersion curves is labeled as Depth (2). Notations such as N, Lat (°), Long (°), W, and Avg Vs (km s\(^{-1}\)) are used to denote specific measurements. Notes at the bottom of the table provide additional details on the data sources and assumptions used in the calculations.
4 RESULTS

In this section, we review the crustal thickness, mean crustal shear wave velocity, thickness of the mafic lower crust and the mean crustal Poisson’s ratio for terranes of similar age. To do this, estimates obtained in this study are combined with previously published results obtained using similar techniques, as described and referenced in the Supporting Information and summarized and referenced in Table 2. Estimates for a terrane’s crustal structure are averages of estimates from all the stations within that terrane. Table 2 gives the estimates for each terrane ordered by Precambrian age and in Fig. 9 the estimates are shown graphically.

4.1 Mesoarchean terranes

The estimates for average crustal thickness are very similar between the Mesoarchean terranes and range from 36 to 39 km. The Swaziland and Pietersburg terranes in the Kaapvaal Craton both have an average crustal thickness of 39 km and a fairly thick mafic lower crust (Swaziland Terrane = 14 km and Pietersburg Terrane = 12 km). The Witswatersrand Terrane and the Kimberley Terrane, also in the Kaapvaal Craton, have average crustal thicknesses of 37 km, but the mafic lower crust in the Witswatersrand Terrane is 7 km and only 2 km in the Kimberley Terrane. The East Tokwe Block in the Zimbabwe Craton has an average crustal thickness of 36 km and a mafic lower crustal thickness of 12 km. Poisson’s ratio for the terranes range from 0.25 to 0.28. The estimates for the average crustal shear wave velocity range from 3.6 to 3.7 km s$^{-1}$.

4.2 Neoarchean terranes

The average crustal thickness in the Neoarchean terranes is quite variable, ranging from 36 to 45 km. In the Zimbabwe Craton, the West Tokwe Block has an average crustal thickness of 36 km and a mafic lower crustal thickness of 4 km. The Limpopo Belt is slightly thicker and has an average thickness of 41 km, and a mafic lower crustal thickness of 14 km. In the Tanzania Craton, the Dodoman and Nyanzian blocks have an average crustal thickness of 38 km, and mafic lower crustal thicknesses of 3 and 4 km, respectively. The Bangweleu Block has an average crustal thickness of 41 km and no mafic lower crust. The Angola Block has a 38 km thick crust with a mafic lower crustal thicknesses of 13 km, and the Kasai Block has a 37 km thick crust with a 3 km mafic lower crust. In contrast, the Ntem Complex in Cameroon is much thicker than any
Figure 7. Results from the joint inversion for station TEZI. (a) Receiver function stacks with low Gaussian frequency (1.0) stacks on top and high Gaussian frequency (2.5) stacks on the bottom. Each group is ascending in ray parameter bin. (b) Rayleigh-wave phase (top curve) and group velocities (lower curve). (c) The shear wave velocity model.

4.3 Palaeoproterozoic terranes

Average crustal thickness in the Palaeoproterozoic terranes range from 33 to 41 km. The average crustal thickness is 45 km and the mafic lower crustal thickness is 23 km. Most of the Neoarchean terranes have a Poisson’s ratio of 0.25. However, in the Kasai Block it is 0.23, in the West Tokwe Block and Ntem Complex it is 0.26, in the Nyasian Block it is 0.27, in the Angola Block it is 0.30. The estimates for average shear wave velocity are similar between the Neoarchean terranes, averaging between 3.6 and 3.7 km s⁻¹. The exception is again the Ntem Complex, with a fast average crustal shear wave velocity of 3.9 km s⁻¹.

4.4 Mesoproterozoic terranes

The average crustal thickness in the Mesoproterozoic terranes ranges from 33 to 41 km. The Namaqua-Natal Fold Belt has an average crustal thickness of 33 km and a mafic lower crustal thickness of 12 km, however the reported crustal thickness for this terrane has a large standard deviation of ±6 km. The Irumide Belt has an average crustal thickness of 41 km and a mafic lower crustal thickness of 4 km, while the average Southern Irumide crust is 39 km thick with a 3-km-thick mafic lower crust. In eastern Africa, the Kibaran Belt has a 40-km-thick crust and a 4-km-thick mafic lower crust. Comparatively, the average crustal thickness of the Rwenzori Belt is 39 km with a 2-km-thick mafic lower crust. Poisson’s ratio for all the Mesoproterozoic terranes is 0.25 except for in the Irumide and Southern Irumide Belt. The Irumide Belt has a Poisson’s ratio of 0.23, while in the Southern Irumide it is 0.28. The average crustal shear wave velocity is similar for all Mesoproterozoic terranes, averaging between 3.7 and 3.8 km s⁻¹.

4.5 Neoproterozoic terranes

The Neoproterozoic terranes have similar average crustal thicknesses, ranging from 38 to 40 km. The Damara Belt has an average crustal thickness of 38 km and a 2-km-thick mafic lower crust. In southeastern Africa, the Zambezi Belt has an average crustal thickness of 39 km and a mafic lower crust of 4 km. Comparatively, the crustal thickness in the Lufilian Arc is 40 km with a 5-km-thick mafic lower crust. The Mozambique Belt has an average crustal thickness of 38 km and a mafic lower crustal layer 2 km thick. In western Africa, the crustal thickness of the Oubanguides Belt is 39 km and the mafic lower crust is 7 km thick. Estimates for Poisson’s ratio ranges from 0.26 to 0.28. The average crustal shear wave velocities in the Neoproterozoic terranes range from 3.6 to 3.8 km s⁻¹.

5 DISCUSSION

Are there patterns in mean crustal thickness, crustal Poisson’s ratio, mean crustal velocity and lower crustal velocity that show age-dependent trends through the Archean and Proterozoic? Here we address that question, determining if the conclusion reached by Tugume et al. (2013) that there are no secular trends is consistent with our expanded dataset. We also compare our estimates to several previous studies that have examined the structure of Precambrian crust and discuss the implications of our estimates for understanding crustal genesis from the Mesoarchean and onwards.
The average crustal thickness for Archean terranes is 38 ± 3 km (SD), the average crustal thickness for Proterozoic terranes is 39 ± 3 km (SD) and the average crustal thickness for all Precambrian terranes is 39 ± 3 km (SD, Table 2). While some terranes are thicker or thinner than the average 39 km, most Precambrian terranes have thicknesses that fall within 1 SD of the mean thickness. The West Central African Belt (33 km) and the Okwa Terrane (43 km), for which there is only one data point, are exceptions, as is the Ntem Complex.

Granitic rocks with a felsic composition have a Poisson’s ratio of 0.24, while intermediate lithologies such as diorite have a Poisson’s ratio around 0.27. Rocks with mafic lithologies like gabbro have values around 0.30 (Tokay & Vavakin 1982; Christensen 1996). The average Poisson’s ratio for many of the Precambrian terranes is between 0.25 and 0.27 with an overall average of 0.26 ± 0.01 (SD), which is indicative of bulk felsic to intermediate lithologies for the continental crust. A few terranes are more felsic (i.e. Irumide Belt, 0.23), or more mafic (i.e. Pietersburg Terrane 0.28, Angola Block 0.30, Southern Irumide Belt 0.28 and Lufilian Arc 0.28) (Table 2).

The average crustal shear wave velocity for all Archean and Proterozoic terranes is between 3.6 and 3.8 km s\(^{-1}\) with an overall average of 3.7 km s\(^{-1}\) (Table 2). Given the uncertainties, all terranes fall within 1 SD of the mean velocity.

Most Precambrian terranes have a high shear wave velocity (between 4.0 and 4.3 km s\(^{-1}\)) layer in the lower crust that is between 2 and 23 km thick. The thickness of this mafic layer is quite variable. In Archean terranes, the average thickness is 8 ± 6 km (SD) and in Proterozoic terranes the average thickness is 6 ± 3 km (SD). All Precambrian terranes exhibit a mafic lower crust except for the Bangweleu Block, where a high velocity layer at the base of the crust was not found.

In summary, we find little evidence for secular trends in crustal thickness, Poisson’s ratio, mean crustal velocity and lower crustal thickness for the southern African continent. Importantly, we find as much variability in crustal structure within an age range (i.e. Mesoarchean, Neoarchean, Palaeoproterozoic, Mesoproterozoic and Neoproterozoic) as there exists across all the Precambrian terranes. In addition, we find little difference in crustal structure between Meso- and Neoarchean terranes. Thus our findings are consistent with the conclusion reached by Tugume et al. (2012, 2013), which was based on a more limited dataset for Africa.

5.1 Comparison to other studies

Durrheim & Mooney (1991, 1994), Thompson et al. (2010), and Abbott et al. (2013) concluded that average Archean crust is different from post Archean crust, arguing for a secular trend in crustal genesis. In contrast to these studies, our findings do not indicate a change in crustal structure as a function of age. Rudnick & Fountain (1995), Zandt & Ammon (1995), Tugume et al. (2012, 2013) and Stankiewicz & De Wit (2013), on the other hand, found no significant difference between the mean thickness, mean velocity, or composition of Archean and Proterozoic terranes. Our estimates are consistent with these studies. However, our estimates indicate a larger variability in the lower crustal structure within terranes of a similar age, as well as between terranes of different ages, than suggested by Rudnick & Gao (2003). This finding is illustrated in Fig. 10, which shows a plot of the thickness of the crust versus the thickness of the high shear wave velocity layer in the lower crust. The data are grouped by terrane age and show that for Precambrian terranes of all ages, there is significant variability in the thickness of the high velocity layer in the lower crust and that there is no correlation with crustal thickness.

5.2 Implications for crustal genesis

The estimates from this study show little evidence for secular trends in crustal structure for terranes in the southern African subcontinent.
spans some 3.6 Ga of Earth’s history. Evaluated in the context of secular variation and the models initially presented by Durrheim & Mooney (1991) and Rudnick & Fountain (1995), this finding suggests one of two things. Either there have been few changes over much of Earth’s history in the processes that formed the crust of the southern African subcontinent, or crustal structure has been modified to the point where secular trends are no longer observed. If similar modes of crustal genesis have been active throughout much of Earth’s history, then modern day plate tectonic processes that form continental crust, such as island-arc accretion and arc magmatism, would appear to be the dominant crustal forming processes from at least the Mesoarchean.

Alternatively, during the Mesoarchean, hotter mantle temperatures may have favoured mantle plumes as the primary generator of continental crust via intracontinental rifting, magmatism and underplating (Arndt & Davaille 2013). If plumes were the primary driver of continental crustal genesis prior to ~3.0 Ga, then secular trends in crustal structure may be evident only between Mesoarchean and younger crust. But, after 3.0 Ga, the rate of crustal growth may have decreased substantially coeval with the onset of crustal recycling and modification of the crust through subduction (Cawood et al. 2013; Hawkesworth et al. 2013), and so finding Mesoarchean crust that has not been reworked can be difficult. Indeed, the five terranes which formed in the Mesoarchean (Fig. 9) are overlain by the 3.1–2.87 Ga Pongola Basin, the 3.07–2.7 Ga Witwatersrand Basin and the 2.7–2.8 Ga Ventersdorf Basin, possibly indicating some degree of crustal modification in the Neoarchean (Begg et al. 2009 and references therein). Therefore, from our data ensemble it is difficult to make definitive statements about differences in Mesoarchean versus younger crustal structure or the lack thereof.

Consequently, while our findings do not show secular variation in crustal structure from the Mesoarchean and onwards, because of crustal recycling and modification by plate tectonic processes, which may have homogenized crustal structure across many terranes, drawing firm conclusions about secular variations in crustal genesis is not straightforward. The lack of secular changes in crustal structure may simply reflect crustal reworking and not the processes by which the crust was initially extracted from the mantle.

<table>
<thead>
<tr>
<th>Age</th>
<th>Terrain</th>
<th>Source (a): Crustal thickness from H-wave receiver functions and surface wave dispersion curves.</th>
<th>Source (b): Poisson’s ratio</th>
<th>Source (c): Mafic lower crust velocity from jointly inverting H-wave receiver functions.</th>
<th>Source (d): Unpublished data.</th>
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<tbody>
<tr>
<td>Mesoarchean</td>
<td>Swaziland Terrane (Kaapvaal Craton)</td>
<td>2 39 ± 7 6.7 0.26 6 14 3.7 7</td>
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<td>Neoarchean</td>
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<td>3 37 ± 0 36 ± 6.7 0.26 6 4 ± 1 3.7 6</td>
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<td>1 37 ± 7 38 ± 6.7 0.30 1 13 3.7 1</td>
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<td>Neoarchean</td>
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<td>5 45 ± 2 45 ± 6.7 0.26 4 23 ± 4 3.7 ± 0.1 5</td>
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<td>1 44 ± 7 38 ± 6.7 0.27 1 2 3.7 1</td>
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<td>Mesoproterozoic</td>
<td>Irumide Belt</td>
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Notes: N, number of stations. Avg depth (1): Results from H–κ stacking of P-wave receiver functions. Avg depth (2): Results from jointly inverting P-wave receiver functions and surface wave dispersion curves. Source (a): Crustal thickness from H–κ stacking and jointly inverting P-wave receiver functions and surface wave dispersion curves. Source (b): Poisson’s ratio from H–κ stacking of P-wave receiver functions. Source (c): Mafic lower crustal thickness and velocity from jointly inverting H-wave receiver functions and surface wave dispersion curves.
Figure 9. A plot of average crustal structure. The terranes are ordered from oldest (left-hand panel) to youngest (right-hand panel) and are shaded with the fill patterns by age (Mesoarchean, Neoarchean, Palaeoproterozoic, Mesoproterozoic and Neoproterozoic), which is the same as in Fig. 1 and shown in the key. The error bars on Moho depths are ±1 SD of the average for each terrane for which there are 3 or more stations. The grey shading represents the average thickness of layers in the lower crust with $V_s \geq 4.0$ km s$^{-1}$. The bold, dashed lines at 36 and 42 km show the average crustal thickness range (i.e. 39 km ± 3 km). The black bars indicate the number of stations in each terrane.

Figure 10. Plot showing average crustal thickness for each terrane versus the thickness of crustal layers with $V_s \geq 4.0$ km s$^{-1}$ (i.e. mafic lower crust).

### 6 SUMMARY

In this paper, we have obtained estimates of crustal structure for 39 new stations using the $H$-$\kappa$ stacking method and jointly inverting $P$-wave receiver functions with surface wave dispersion measurements. This includes estimates for several Precambrian terranes for which there were few previously published results of crustal structure (Bangweleu Block, West Central African Belt, Irumide Belt, Southern Irumide Belt, Zambezi Belt and Lufilian Arc).
After combining our new estimates with published results for the southern African subcontinent and examining crustal thickness, Poisson’s ratio and the shear wave velocity for some 30 terranes ranging in age from ~3.6 to 0.5 Ga, we find little evidence for secular trends in Precambrian crustal structure. For both Archean and Proterozoic crust, we find similar ranges in crustal thickness (38–39 ± 3 km SD), Poisson’s ratio (0.26 ± 0.01 SD), and shear wave velocity (3.7 ± 0.1 km s−1 SD), as well as similar amounts of heterogeneity in the lower crustal structure.

This finding can be interpreted in one of two ways, (1) either that continental crust has been formed by similar processes since the Mesoarchean, or (2) crustal reworking by plate tectonic processes has erased any secular trends in crustal structure that would indicate a change in crustal genesis at some point in Earth’s history.

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REFERENCES


SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this paper:

Supplemental File A. A summary of the seismic estimates of crustal structure from previous studies for the Precambrian terranes of the southern African subcontinent is provided in this supplemental file. Many of the estimates reviewed have been included in Table 2.

Supplemental File B. List of teleseismic events used.

Supplemental File C. Results from the $H$–$k$ stacking method for all stations.

Supplemental File D. Results from jointly inverting $P$-wave receiver functions and Rayleigh wave phase and group velocities.

Supplemental File E. Radial (left) and tangential (right) receiver functions versus ray parameter for stations KISZ and SENA where the Moho depth was estimated using the Moho Poisson’s ratio of 0.26, and eq. (2) from Zandt et al. (1995).

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