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Review Article

3.5 billion years of reshaped Moho, southern Africa

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ABSTRACT

According to some previous studies, Archean continental crust is, on global average, apparently thinner than Proterozoic crust. Subsequently, the validity of this statement has been questioned. To provide an additional perspective on this issue, we present analyses of Moho signatures derived from recent seismic data along swaths 2000 km in length across southern Africa and its flanking ocean. The imaged crust has a near continuous age range between ca. 0.1 and 3.7 billion years, and the seismic data allow direct comparison of Moho depths between adjacent Archean, Proterozoic and Phanerozoic crust. We find no simple secular change in depth to Moho over this time period. In contrast, there is significant variation in depth to Moho beneath both Archean and Proterozoic crust; Archean crust of southern Africa displays as much crustal diversity in thickness as the adjacent Proterozoic crust. The Moho beneath all crustal provinces that we have analysed has been severely altered by tectono-metamorphic and igneous processes, in many cases more than once, and cannot provide unequivocal data for geodynamic models dealing with secular changes in continental crust formation. These results and conclusions are similar to those documented along ca. 2000 km swaths across the Canadian Shield recorded by Lithoprobe. Tying the age and character of the Precambrian crust of southern Africa to their depth diversities is clearly related to manifold processes of tectono-thermal 'surgery' subsequent to their origin, the details of which are still to be resolved, as they are in most Precambrian terranes. Reconstructing pristine Moho of the early Earth therefore remains a formidable challenge. In South Africa, better knowledge of 'fossilised' Archean crustal sections 'turned-on-edge', such as at the Vredefort impact crater (for the continental crust), and from the Barberton greenstone belt (for oceanic crust) is needed to characterize potential pristine Archean Moho transitions.

Contents

1. Introduction

Knowledge about the thickness and nature of Archean crust is of fundamental importance to constructing geodynamic models of the early

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Earth; and to test for secular changes in tectonic processes over Earth history. Throughout the 1970s and 1980s, thermal arguments that continental crust should thicken with age, as observed for oceanic crust, or that there was no discernable difference in its thickness with age, held sway (e.g. Jarchow and Thompson, 1989; Meissner, 1986; Mueller and Ansorge, 1989; Rudnick, 1995; Taylor and McLennan, 1985) until studies in the early 1990s showed that crust of preserved Archean segment (cratons; >2.5 Ga) is, on global average, apparently thinner than crust of Proterozoic domains (Durrheim and Mooney, 1991, 1992, 1994; Mooney et al., 1998). In these studies, depth to Archean Moho was estimated to range between 27 and 40 km thick, whilst this depth beneath Proterozoic crust varied between 40 and 55 km (Durrheim and Mooney, 1991, 1992). In addition, Archean crust was reported to be of low velocity, and with a Moho of low impedance contrast relative to Proterozoic crust (e.g. Gibbs, 1986).

Subsequently these studies were criticised because the seismic data on which they were based were of low quality and selectively used, rendering the results statistically unjustified (Wever, 1992). Indeed, all the studies acknowledged that there was insufficient data to conclusively resolve the issue. None-the-less, influential modern studies on the evolution of continental crust continue to indiscriminately use the global analyses of Durrheim and Mooney (1991, 1992), stating that Archean continental crust is thinner (and less underplated with high velocity basaltic crust) than Proterozoic and more recent crust (e.g. Hawkesworth et al., 2010; Hynes, 2001; Keller and Schoene, 2012; Moyen and van Hunen, 2012; Petitjean et al., 2006), despite evidence to the contrary based on high quality data of the world's most comprehensive crust–mantle transition survey across the well-mapped, iconic Phanerozoic–Precambrian terranes of Canada (e.g. Lithoprobe, see Clowes, 2010; Cook et al., 2010; Eaton, 2005; Hammer and Clowes, 1997; Percival et al., 2012, and references therein).

Seismic profiles documenting reflective lower crust with a well-defined Moho have been recorded in numerous Proterozoic and Archean regions (Clowes, 2010; Hammer and Clowes, 1997; Mooney et al., 1998). Whilst older global datasets suggested that Moho impedance contrasts and continuity tend to decrease with age due to gradual disruption of reflective laminations (e.g., Gibbs, 1986), such a simple correlation with age is not supported by the comprehensive Canadian datasets (Cook et al., 2010; Hammer and Clowes, 1997). Instead, well-defined reflection Moho appears to correlate better with major deformational episodes tectonic and/or igneous underplating (Cook and Erdmer, 2005; Cook et al., 2010; Eaton, 2005). Whilst this reinforces the concept that ductile shear and the rheological properties of the Moho transition zone, lower crust and upper mantle play important roles in influencing the thickness and reflective character, Moho reflectivity is not a robust indicator with which to distinguish between specific tectonic processes, such as extension, compression, transpression, or delamination (Eaton, 2005; Hammer and Clowes, 1997). Other geological observations must be integrated to resolve the origin of specific Moho transitions (Cook, 2002; Cook and Erdmer, 2005; Cook et al., 2010).

To further test for variation in crustal thickness over deep-time, we present analyses of Moho signatures derived from more recent seismic data across southern Africa. Our analysis allows direct comparison of Moho depths between adjacent Archean, Proterozoic and Phanerozoic crust, recommended by Wever (1992) as a prerequisite to test the global scale generalisations of a crustal age versus thickness relationship.

To clarify our nomenclature in this paper, we specifically refer to 'shields' as stabilised, post-Archean continental domains in which Archean cratons are often embedded. In the literature the terms Kalahari Craton, Kalahari Shield and Kaapvaal Craton are often used indiscriminately, and interchangeable; and Kalahari Craton is sometimes erroneously used for the combined Kaapvaal and Zimbabwe Cratons. We use of 'Azanian Craton' rather than 'Kalahari Craton': the Archean Kaapvaal and Zimbabwe Cratons amalgamated along the Archean Limpopo Belt to form a single unit referred to as the Azanian Craton (de Wit, 2007; McCourt et al., 2004). Subsequent accretion in the Proterozoic by a number of surrounding orogenic belts, such as the Natal–Namaqua Mobile Belt, inculcated the Azanian Craton into the Kalahari Shield that stabilised around 1.0 Ga; and later into a still larger Southern African Shield following younger peripheral orogenesis of the Pan African (ca. 600–700 Ma) and further stabilisation by 500 Ma (see inset in Fig. 1a).

The seismically imaged crust includes that of the Kaapvaal Craton (Fig. 1), the world's best preserved and studied Archean crust (and lithosphere), with an age range between 2.5 and 3.7 Ga (Adam and Lebedev, 2012; Carlson et al., 1996, 2000; Taylor et al., 2012, and references therein). The Kaapvaal Craton is flanked to the south by Mesoproterozoic crust of the Natal–Namaqua Mobile Belt (NMMB), a medium grade granulite terrain ranging in age from ca. 1.0 to 2.0 Ga (Cornell et al., 2006; Eglington, 2006). The latter is overlain along the coastal mountains of southernmost Africa by Palaeozoic sedimentary rocks (Cape–Karoo sequences) that were subjected to deformation in the late Palaeozoic (0.25 Ga) to form the 'Cape Fold Belt' (Lindeque et al., 2007, 2011; Tankard et al., 2009). During the early Mesozoic this fold-and-thrust belt was in part eroded and covered by 2–7 km of Mesozoic sedimentary sequences, and subsequently exhumed again in the late Cretaceous during the Kalahari Epeirogeny (de Wit, 2007) to form the Cape Mountains (Scharf et al., 2013; Tinker et al., 2008). During the upper Mesozoic the entire southern African continental margin was affected also by widespread crustal extension related to the opening of the southern oceans, forming a wide, relative shallow continental shelf, known as the Agulhas Bank (ca. 100,000 km^2, about the size of Iceland), dissected by numerous Jurassic–Cretaceous rift-basins (e.g. Tinker et al., 2008). The southern edge of this continental shelf (known as the Diaz Marginal Ridge) abruptly terminates against the Agulhas Falkland Fracture Zone (AFFZ, Fig. 1). The Agulhas Bank is underlain by Mesoproterozoic crust with a shallow Moho (30–33 km, Durrheim, 1987; Hales and Nation, 1972; Parsiegla et al., 2009; Stankiewicz et al., 2008). South of the AFFZ is early Cretaceous oceanic crust that, farther south still, is overlain by a late Cretaceous oceanic plateau (the Agulhas Plateau), which has been interpreted in the past as a displaced section of continental crust (Scrutton, 1973; Tucholke et al., 1981), but is now recognised to comprise entirely of oceanic crust (Gohl et al., 2011; Parsiegla et al., 2008).

Before proceeding with our analyses, it is important to clearly state what is meant by the Moho. Geophysicists conventionally define the Moho as the first order P-wave velocity discontinuity representing the crust–mantle boundary. These velocities are typically estimated using refraction seismic experiments. However, the definition of the base of the crust based on velocities alone is often imprecise and open to debate. With refraction surveys virtually always verifying the jump in velocities, seismic reflection surveys often indicate a complex set of reflectors with no certain discontinuity level (see Doyle, 1995, for a review). Furthermore, the geophysically defined boundary is not necessarily coincident with the petrological Moho, defined by compositional change of the material (e.g. Mengel and Kern, 1992), but for which there is still no scientific consensus about its origin or petrologic significance (e.g. Eaton, 2005). Given that most of the results presented here are from seismic reflection and receiver function analyses, we define the Moho for the purposes of this article as a geophysical boundary, where abrupt changes in the elastic properties of the material lead to reflection or phase conversion of seismic waves.

It is imperative to point out also that using seismic waves to estimate the depth of the Moho, or any other discontinuity, knowledge of seismic velocities throughout, and beneath, the crust must be assumed. The depths calculated from the receiver function analysis require accurate information on both P-wave and the S-wave velocities, and these are typically taken as constant for the entire crust. This is clearly a simplification, and with these analyses alone it could be impossible to state whether observed Moho depth variations are real, or represent a change in either (or both velocities). The uncertainties associated with these would be at least 10%, which translates to 4–5 km depth. Crustal structures obtained from seismic reflection velocities are usually better constrained, as the velocity variations with depth can be accurately estimated from, e.g. cores. However in the study area this could only be determined for the upper 4–5 km from drill-core obtained on the Kaapvaal Craton. Below that velocities are estimated based on expected crustal rock types. Given these constraints, our analysis from receiver function and deep reflection studies, along sections where they can be directly compared, indicates excellent agreement within the uncertainties outlined above (e.g. Fig. 1c). The most accurate estimates of depth to Moho are given when reflection seismic lines coincide with refraction profiles. This way accurate velocities obtained from tomography can be projected into the reflection profiles, and depths to reflectors can be accurately calculated. This was done for all the Inkaba lines, onshore and offshore.

1.1. Kaapvaal Craton

Understanding the nature of the continental Moho underlying the Archean Kaapvaal Craton was one of the primary targets of the Kaapvaal Craton Project and the South Africa Seismic Experiment (Carlson et al., 1996, 2000; de Wit et al., 2004, and references therein). Key questions included how the Moho differs within the craton, and whether it remains a pristine Archean discontinuity or has been reshaped in more recent times.

The Kaapvaal Craton can be divided into two different crustal domains of different ages, separated by an N–S divide known as the Colesberg lineament (Figs. 1, 2; Brandl and de Wit, 1997; Schmitz et al., 2004). Deep seismic reflection data show this lineament to be a complex, deep crustal tectonic shear system with a west dipping listric geometry rooted in the middle crust (e.g. Fig. 2; de Wit and Tinker, 2004). East of the Colesberg lineament, the Eastern Block of the Kaapvaal Craton sector (also referred to as the Witwatesrand Block) is older, with granitoid basement ages ranging between ca. 3.0 and 3.66 Ga (Armstrong et al., 2006; de Wit et al., 2011; Eglington and Armstrong, 2004; Xie et al., 2012; Zeh et al., 2011), variably deformed at ca. 3.4 and 3.2 Ga, but stabilised by 3.1–3.0 Ga (e.g. Schoene et al., 2008; Taylor et al., 2012). There are no deep seismic reflection sections across this part of the craton to verify deeper crustal structures.

The Western Block (also known as the Kimberley Block) ranges in age between 2.7 and 3.0 Ga, and has been affected by major compressive tectonism towards the end of the Archean, but was stable by 2.5 Ga (Schmitz et al., 2004). Deep seismic reflection data (vibroseis) has revealed a complex tectonically stacked crust along listric thrusts (which locally affect the Moho), overprinted at ca. 2.7 Ga by major listric normal faults related to rifting and extensive contemporaneous bimodal volcanism of the Ventersdorp Large Igneous Province — VLIP (Armstrong et al., 1991; Hatton, 1995; Johnson et al., 1996; Tinker et al., 2002) to produce a relatively flat and sharp Moho between 38 and 40 km (Fig. 2; de Wit and Tinker, 2004), consistent with local high resolution receiver function analyses (James et al., 2003). Petrology and thermochronology on felsic xenoliths from the lowermost crust of this region indicate ultra-high temperature metamorphism, and likely significant ductile flow and partial melting here at this Ventersdorp time (Schmitz and Bowring, 2003). Thus, crust down to the Moho here has a complex tectonic history and the present character of the lower crust and the depth to Moho do not represent a pristine Archean Moho transition.

The Witwatersrand Block of the Kaapvaal Craton was eroded and peneplained before the Neoarchean. In many places up to 10 km of granitoid crust were removed before ca. 2.9 Ga, after which terrestrial clastic deposits (i.e. the Witwatersrand and Pongola sequences) and then shallow marine rocks (i.e. the Transvaal Group) covered the peneplain (Fig. 3). Thereafter this regional peneplain was reworked repeatedly close to sea level, satisfying freeboard principles (e.g. Hynes, 2001) for nearly 2.5 Gyrs, until late Cretaceous epeirogenic activity rapidly uplifted the craton by up to 2 km (de Wit, 2007). The cause and rate of this uplift are still a matter of intense debate (compare for example Bell et al., 2003; Brown et al., 2002; Burke and Gunnell, 2008; Flowers and Schoene, 2010; Guillocheau et al., 2012; Nyblade and Sleep, 2003; Rouby et al., 2009; Tinker et al., 2008; Torsvik et al., 2010 — and an overview of this debate in de Wit, 2007), but the details are beyond the scope of this paper.

Receiver function analyses of teleseismic events recorded by the Kaapvaal Array (Nguuri et al., 2001; Stankiewicz et al., 2002) found the Moho depth beneath most of the craton to be 35–45 km, though some values over 50 km have been detected in the northern section of the craton (James et al., 2001, 2003; Kwadiba et al., 2003; Wright et al., 2003). Sharp velocity contrasts were observed across the discontinuity, leading to conclusions that the seismic Moho is clearly defined (Fig. 4). Whilst the results of Nguuri et al. (2001) assume a constant Poisson's ratio, more recent studies show this is an oversimplification (Kgaswane et al., 2009; Nair et al., 2006; Youssof et al., submitted for publication) and their Moho depth estimates are slightly different. The study of Nguuri et al. (2001) is based on the assumption that Vp/Vs = 1.73 across the entire region as shown in Fig. 1. Nair et al. (2006) tested what effect Vp/Vs would have on Moho depth calculation. Their biggest deviation, from 1.73 to 1.816, shows the depths to be less than 5% off (<2 km of error for Moho of 40 km). Most of Nair et al.'s (2006) value uncertainties put them between 1.71 and 1.75; their results compared to Nguuri et al. (2001) show some very small deviations, and the larger scale variations that are the scope of this study would not be affected.

In addition, Nui and James (2002) divide the Archean crust into upper and lower units and interpreted the lower crust composition as felsic to intermediate with an average density of 2.86 g/cm^3. Youssof et al. (submitted for publication) use this density to derive an anisotropic model which suggests that the deep crust is layered, and includes a lower unit of ca. 20 km thick with at least 10% anisotropy. This layer is underlain by mantle transitional zone with ca. 5% anisotropy. Both units have similar trends of fast polarisation (ca. 40°N). Previous workers had identified this anisotropy to be restricted to the subcontinental mantle below Moho (e.g. Adam and Lebedev, 2012; Silver et al., 2001, 2004; Vinnik et al., 1995).

Northern sections of the Eastern Block reach Moho depths up to 50+ km, which is a value more typical for Proterozoic crust according to Durrheim and Mooney (1991, 1992). The main area of this thickness is found directly underlying the Proterozoic Bushveld igneous complex, a major, ca. 8 km thick mafic–ultramafic intrusive complex in the uppermost crust related to a Large Igneous Province (LIP) around 2.05 Ga (Cawthorn and Webb, 2001). The thick crust here extends as far as the Okwa Terrane in Botswana. The velocity contrast across the Moho in these regions is not distinct, suggesting a diffuse boundary of mafic layering. Nguuri et al. (2001) comment that such modified Moho is observed also beyond the surface geology and aeromagnetic signatures of the Bushveld Complex, suggesting that lower crustal and Moho modification might have been more widespread than can be inferred from surface maps. This is consistent with the suggestions that significant amounts of mafic material were added to the lower crust during the intrusions of the Paleoproterozoic Bushveld LIP at ca. 2.05 Ga (Cawthorn and Webb, 2001; Mapeo et al., 2004; Wright et al., 2003), and as expected from numerical models for the emplacement of such a large structure (e.g. Sobolev et al., 2011).

By contrast, there is no evidence from seismic or crustal xenolith analyses for evidence of significant amounts of mafic lower crustal material beneath the south and central parts of the Kaapvaal Craton, and the significant seismic anisotropy in the lower crust can be best explained by metamorphism and ductile flow (Youssof et al., submitted for publication) as suggested previously from the xenolith studies (Schmitz and Bowring, 2000, 2003). Such thermo-tectonic processes have clearly operated here repeatedly: 1 — During high grade lower crustal metamorphism and partial melting at 2.7 Ga, overlapping with the Ventersdorp LIP, as carefully documented from lower crustal xenoliths retrieved by kimberlites intruding the south and central parts of the craton (Schmitz and Bowring, 2003); 2 — During high grade lower crustal metamorphism and partial melting flanking the southern margin of the craton at 1.1 Ga, associated with the Natal–Namaqua Orogeny

(Schmitz and Bowring, 2000, 2004); and 3 — During substantial geothermal perturbations of 150–200 °C in the Mesozoic (ca. 0.1–0.2 Ga) beneath both the craton and the NMMB, associated with the Karoo LIP and widespread kimberlite activity during the extended break-up of Gondwana (Bell et al., 2003; de Wit, 2007; Jelsma et al., 2004). The effect of these thermotectonic events on reshaping the Moho is explored further below.

Youssof et al. (submitted for publication) propose a crustal model where isotropic felsic upper crust is separated from the lower crust by high grade foliated gneisses. Ductile stretching and flow of the lower crust during the multiphase metamorphic and melting processes would likely have caused local detachment between the crust and mantle, accentuating the sharpness and relatively flat nature of the older Archean Moho as is observed beneath this region (e.g. de Wit and Tinker, 2004; James et al., 2003).

1.2. Zimbabwe Craton

The Zimbabwe Craton is similar to the Kaapvaal Craton in many aspects, comprising a granite–greenstone terrain that has been significantly tectonised (Blenkinsop et al., 1997; Jelsma and Dirks, 2002). At surface, the crystalline mid-crust of this craton has a distinctly higher proportion of greenstones to granitoids (ca. 1:5) than that of the Kaapvaal Craton (1:10). The Zimbabwe Craton, in general, is Neoarchean in age

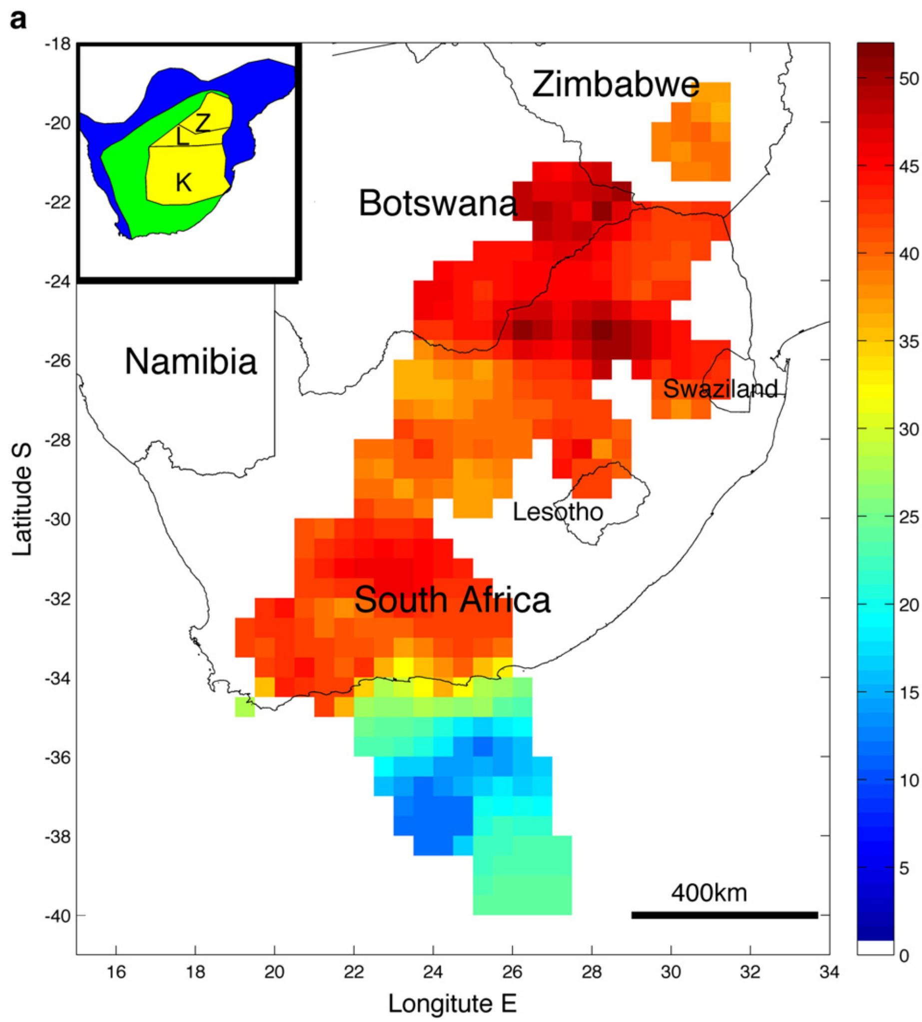

Fig. 1. a. Crustal thickness beneath sections of southern Africa inferred from geophysical experiments. Values are taken from receiver function studies of the Kaapvaal Craton seismic array (Nguuri et al., 2001; Stankiewicz et al., 2002), and seismic reflection and refraction experiments carried out in the framework of the Inkaba yeAfrica project (Parsiegla et al., 2009; Stankiewicz et al., 2008). The map is presented on a 0.5 degree grid, with depths estimated using a weighted average of results from the stations nearest to each grid point. Inset: crustal domains of southern Africa: yellow = Archean Azanian Craton [Kaapvaal + Zimbabwe cratons + Limpopo Belt]; enlarged in Paleo- and Mesoproterozoic (green) to form the Kalahari Shield; and expanding further still to form the Southern African Shield in the Neoproterozoic (blue). See text for further explanation. b. Same as (a), with tectonic features and locations of stations and sections indicated. Black and white triangles indicate the location of stations for which the crustal thickness is known; black = Kaapvaal Craton Project; white triangles and lines = Inkaba yeAfrica Project refraction and reflection lines, respectively (for clarity only selected stations are shown; for details see Stankiewicz et al., 2008). Green, broken and solid lines = vibroseis lines (6 and 16 seconds, respectively). Long solid black lines ABC and DEF are sections shown in Fig. 4. Line G–H is a receiver function section to complement the composite 16 seconds vibroseis line (solid red line), which has been projected into G–H for direct comparison as c; see text. Where the Inkaba and Kaapvaal profiles overlap (including the vibroseis lines) they agree with each other within uncertainties (e.g. c). Short grey lines (ca. NE–SW trending) = averaged orientations of lower crustal and upper mantle anisotropies (from Silver et al., 2001; Vinnik et al., 1995; Youssof et al., 2013; see text for details). Major geological/tectonic boundaries outlined in blue: ZC: Zimbabwe Craton; LB: Limpopo Belt; SB: Shashe Block; WB, EB: Western and Eastern Blocks of Kaapvaal Craton; KB: Kheis Belt; BC: Bushveld Complex (in yellow); VD: Vredefort Dome; AGC: Ancient Gneiss Complex; NNMB: Namaqua Natal Mobile Belt; CFB: Cape Fold Belt. *Onshore lineaments.* 1: Colesberg, 2: Inoyka, 3: Thabizimbi–Murchison, 4, 5: southern and northern boundary of the Limpopo Belt. *Offshore features.* OB: Outeniqua Basin; DMR: Dias Marginal Ridge; AB: Agulhas Bank; AFFZ: Agulhas–Falkland Fracture Zone; AP: Agulhas Plateau. c: Moho depth along the transect from the Western Block (WB) of the Kaapvaal Craton to Eastern Block (EB) in the vicinity of the Vredefort dome; average age-span between crustal formation and stabilisation for the two blocks is also shown. Comparison of analysis from receiver function and deep reflection studies, along sections where they can be closely compared. Blue: This study. Red: extrapolated from vibroseis transection in de Wit and Tinker (2004). Locations of the transects are shown in b (black line, G–H = from receiver function data; intersecting red line = composite of adjacent deep vibroseis lines). Note that due to the absence of stations directly along the vibroseis section, the transections do not coincide precisely. Given the uncertainties in both methods of at least 10% (e.g. 4–5 km depth), the depth to Moho indicates an excellent agreement. See text for further explanation.

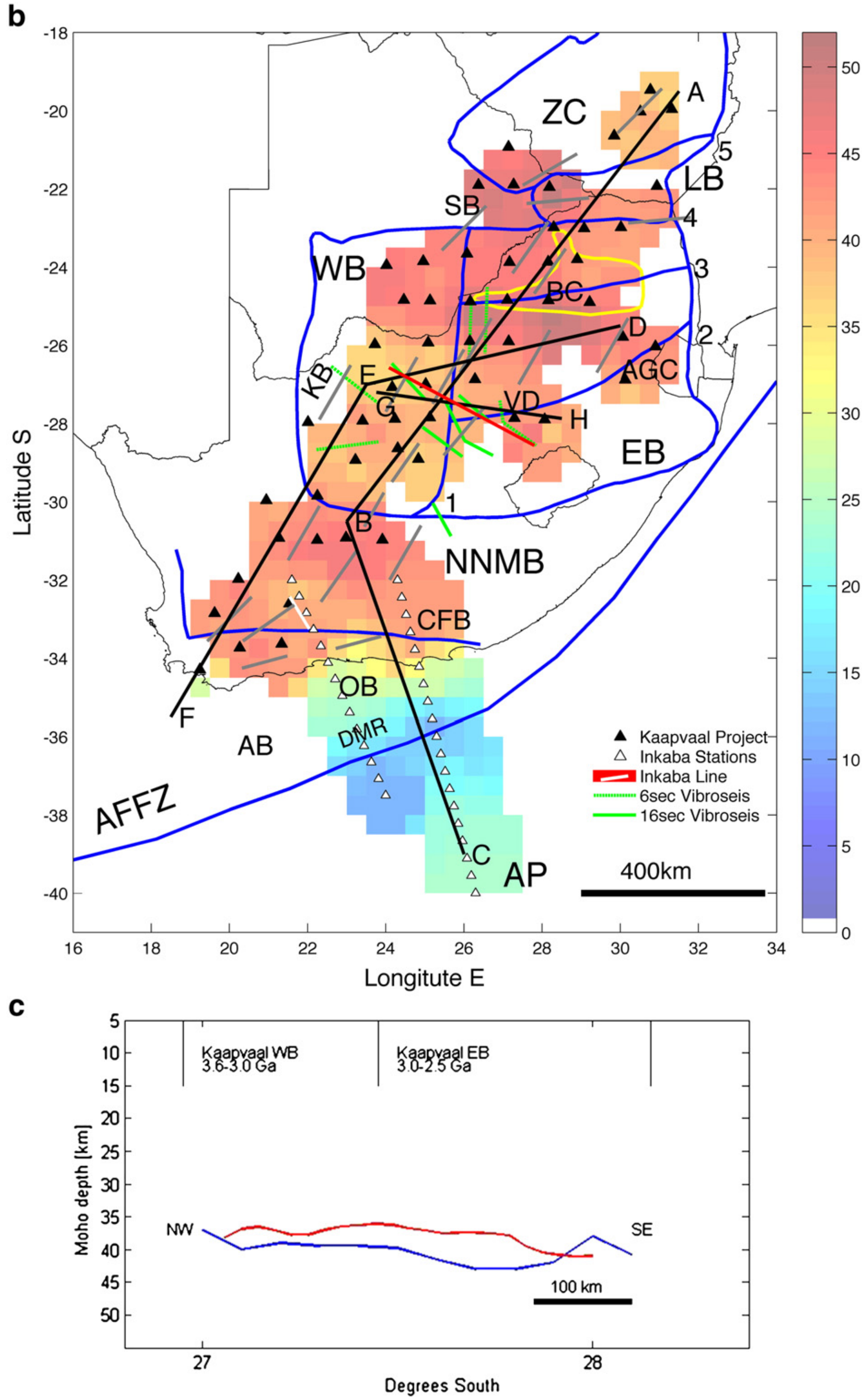

Fig. 1 (*continued*).

(2.6–3.0 Ga), younger than the general stabilisation of the Eastern Block of the Kaapvaal Craton (3.0–3.1 Ga), and in this respect is more akin to the Western Block. Work in northeastern Botswana (Bagai, 2008; Bagai et al., 2002; McCourt et al., 2004) has shown that this is also true for the Archean rocks there, and that there may be continuation between the Zimbabwe Craton and the Western Block of the Kaapvaal Craton (Fig. 1). However, in the south-central part of the Zimbabwe Craton, older granitic gneisses and greenstone belts of the Tokwe terrain (3.0–3.6 Ga) flank the Limpopo Belt (Jelsma and Dirks, 2002; Zeh et al., 2009). These older rocks resemble parts of the older Eastern Block of the Kaapvaal Craton, such as the Barberton region in South Africa, the ancient Gneiss Complex in Swaziland, and farther south in Natal (Schoene et al., 2008, 2009; Taylor et al., 2012; Xie et al., 2012; Zeh et al., 2009).

In the central part of the Zimbabwe Craton Moho depths between 35 and 40 km have been measured (Barton and Klemd, 2010; Gore et al., 2009). This is a stark contrast to the south-western part of the craton, and in the Limpopo Belt, where depths near 50 km have been recorded.

1.3. Limpopo Orogenic Belt

The rocks of the Limpopo Belt and its western extension, known as the Shoshe Belt in Botswana, in general are high grade granitoids and supracrustal rock suites representing exhumed Archean–Paleoproterozoic mid- to lower-crust. Kimberlite xenolith mineralogy suggests that the lower crust of the belt is underlain by thick layers of mafic granulite and eclogite (Pretorius and Barton, 2003).

The Limpopo belt is characterised by a complex orogenic history between 2.0 and 3.66 Ga, but there appears to be no simple, single thermo-chronology for the tectono-metamorphic history of the Limpopo Belt as a whole (Barton et al., 2006). Its rocks were affected by Mesoarchean melting at 3.14 Ga, and again by regional deformation and partial melting in the Neoarchean (2.6–2.7 Ga), which is particularly well preserved along the northern and southern margins of the belt. Subsequently, a strong tectono-thermal overprint affected the Limpopo Belt at the end Paleoproterozoic (2.02–2.06 Ga), which is most prominently preserved within the central parts of the belt (Zeh et al., 2009). Consequently, the upper crustal geology and geochemistry of the central Limpopo Belt's high-grade gneisses are distinct from the granite–greenstone crust of the two flanking cratons, justifying its identification as a separate crustal block (Barton et al., 2006; de Wit et al., 1992; McCourt and Vearncombe, 1992; Zeh et al., 2009, 2011). The uppermost 20–30 km of this Archean crust had been exhumed by the end of the Paleoproterozic (Barton et al., 2006; Blenkinsop et al., 1997; Zeh et al., 2009), but the details of the ca. 700 Myrs exhumation history between the two main metamorphic events (e.g. between 2.7 and 2.0 Ga) remain obscured.

The complex history of the Limpopo Belt has been interpreted in the past to reflect an exhumed Neoarchean orogen, comprising a central exotic crustal block remobilised during collision between a passive margin of the Kaapvaal Craton and an active Andean-like margin flanking the Zimbabwe Craton creating a thick crustal root (e.g. Roering et al., 1992). This simplicity is no longer tenable (Barton et al., 2006; Zeh et al., 2009). Whilst the Archean part of the orogen likely comprises deformed Archean arc terranes flanking the Limpopo Belt, it is speculated that the thick lower crust is Paleoproterozoic and may be related to the emplacement history of the adjacent Bushveld Complex during transpression/transtension between the Kaapvaal and Zimbabwe Craton (now

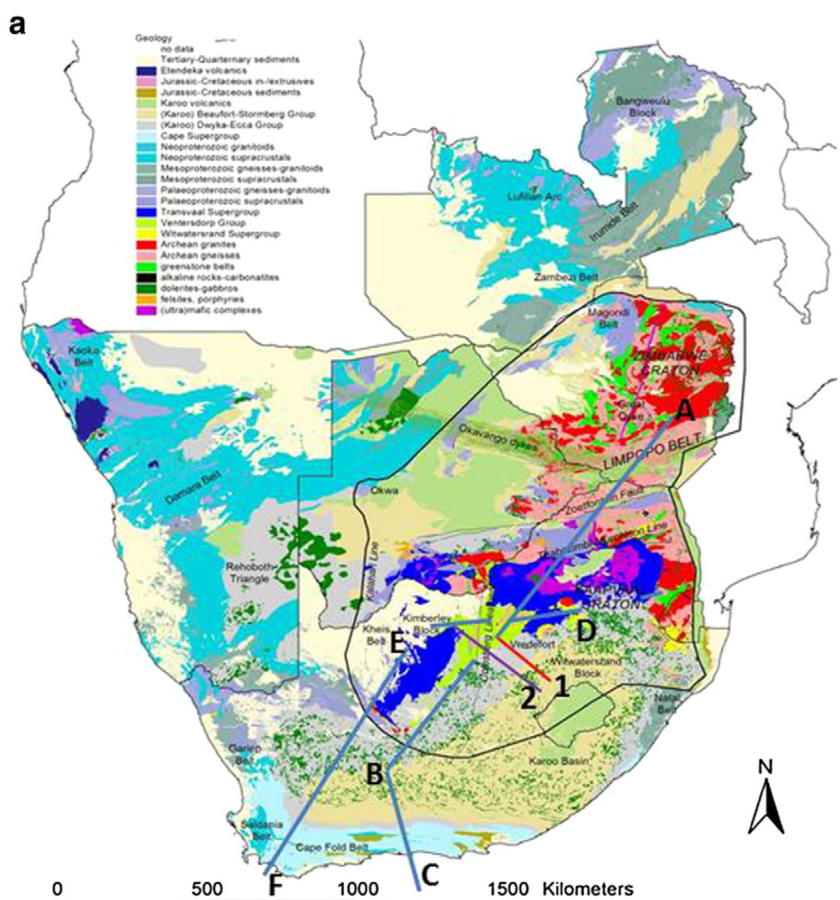

Fig. 2. a–c. a. Geology of southern Africa with locations and tectonic features referred to in the text. Note the outline of the Archean Kaapvaal and Zimbabwe Cratons is shown as one unit (thin black line), referred to as the Azanian Craton. Section lines ABC and DEF (blue), also shown in Fig. 1, along which depths to Moho are provided in Fig. 4. Lines 1 (red) and 2 (purple) represent the vibroseismic sections shown as b and c, respectively (map is modified from the frontispiece of the Inkaba yeAfrica special volume of the South African Journal of Geology volume 114, 2011). b. Vibroseismic reflection profiles across part of eastern margin of the Eastern Block (from de Wit and Tinker, 2004). (i) Interpreted and (ii) uninterpreted. Strong reflectors marked in black; blue dashed line best estimate of Moho. These lines clearly depict the listric extensional faults related to the Ventersdorp. Note the relative flat Moho at 38–40 km, which compares directly to the Moho depth determined from the receiver function along this line (see Fig. 1 c). The Vredefort dome, projected onto line DE-86, occurs about 100 km to the north. For further details see text and de Wit and Tinker (2004). c. Interpreted composite profile from 16 second vibroseis lines projected into a common NW–SE section across the central Kaapvaal Craton, and the Colesburg anomaly (from de Wit and Tinker, 2004). The crustal section comprises 5–10 km thick panels separated by tectonic discontinuities, similar to those observed in many Lithoprobe sections. The thickness of the Ventersdorp and Witwatersrand Supergroups is constrained by drill holes, as shown. Ages of the panels are interpreted from surface geology and drill cores (for detailed locations of sections and drill holes, see de Wit and Tinker, 2004, and text for details). Note that the depth to Moho is consistent with local high resolution receiver function analyses (Nguuri et al., 2001).

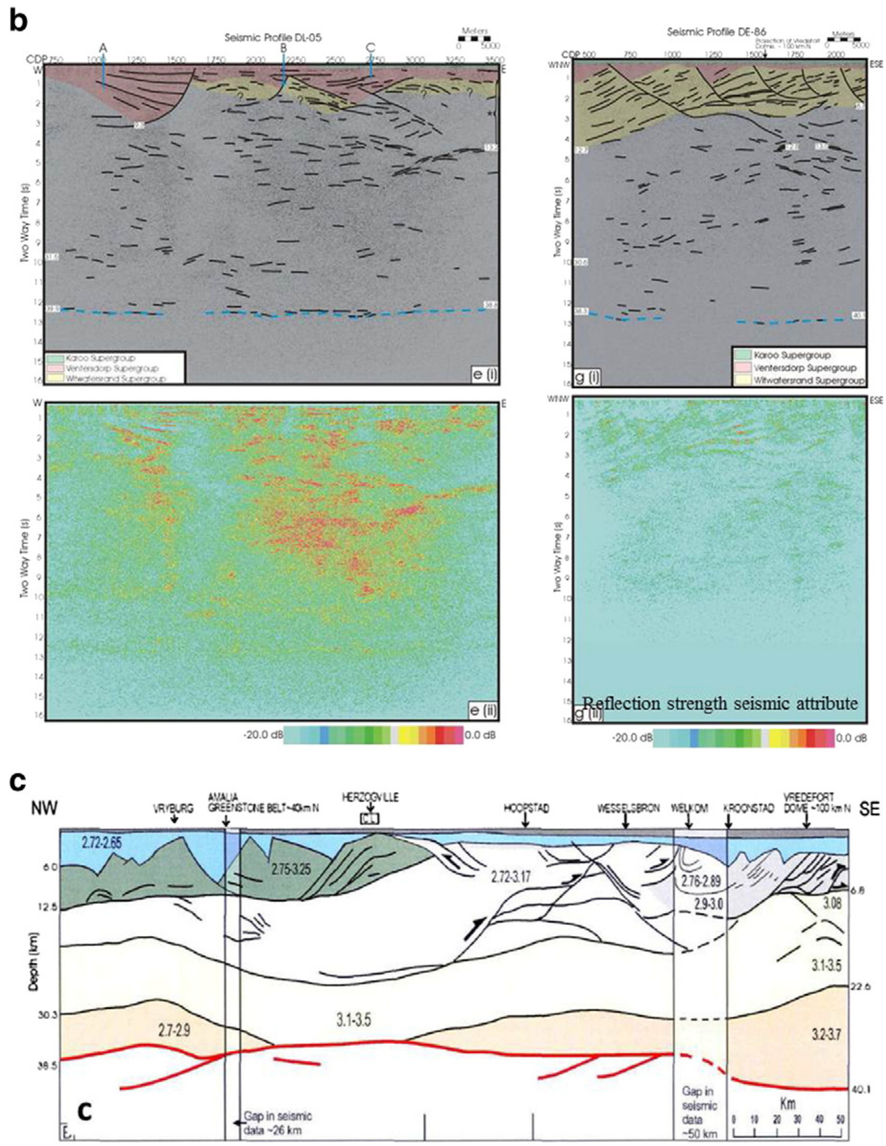

Fig. 2 (*continued*).

collectively known as the Azanian Craton). These interpretations remain part of an unresolved debate about the origin and evolution of the Limpopo Belt (Bagai, 2008; Barton et al., 2006; Zeh et al., 2009).

The Limpopo Belt is characterised by the most complex Moho structure found beneath the Archean crust of southern Africa. Early seismic studies using mine tremors as sources (Stuart and Zengeni, 1987) observed the Moho as a strong P-wave velocity contrast at depths between 30 and 35 km, but no seismic reflections were observed at this discontinuity during a subsequent reflection seismic experiment (Durrheim et al., 1992). Indeed, studies of receiver functions from the Kaapvaal Craton seismic array (Nguuri et al., 2001; Stankiewicz et al., 2002) found that in some cases the identification of the Moho is ambiguous, with weak Ps signal on receiver function analysis implying a structurally complex Moho at depths of up to, and even over 50 km (e.g. 56 km and possibly deeper, Barton et al., 2006). However, a clear crustal discontinuity at 30–35 km depth is also observed (Gore et al., 2009; Nguuri et al., 2001) — these studies, however, do not observe a significant velocity contrast across this discontinuity, and concluded it cannot be the Moho. Geothermobarometry studies on deep crustal xenoliths from the diamondiferous Venetia kimberlite that intrudes the central Limpopo Belt suggest the presence of a thick layer (>10 km) of medium pressure mafic granulite and eclogite at the base of the Moho (Barton and Klemd, 2010; Pretorius and Barton, 2003). We postulate therefore that the present-day Moho beneath the central and southern sections of the Limpopo Belt is deeper than 45–50 km, but is not clearly visible due to many kilometres of mafic underplating, likely related to the Bushveld and possibly also the earlier Ventersdorp LIP events.

Beneath the northern section of the belt the crustal thickness decreases rapidly to less than 40 km, similar to the values found beneath the Zimbabwe Craton. This is consistent with the geological surface observations that indicate a thin crustal slice of the Limpopo crust has been thrust northward across the Zimbabwe Craton (e.g. Barton et al., 2006). Beneath the southern margin of the belt, comprising high-grade Kaapvaal cratonic material thrust across and over the adjacent lower grade granite–greenstone Pietersburg Block of the Kaapvaal Craton, Moho depths similarly decrease across the terrane boundaries to 40–42 km.

1.4. Proterozoic Moho beneath Namaqua Natal Mobile Belt (NNMB)

This orogenic belt accreted to the southern edge of the Kaapvaal Craton at ca. 1.0–1.3 Ga (Cornell et al., 2006). The Mesoproterozoic crustal components of the NNMB date between 1.0 and 2.0 Ga and comprise a number of allochthonous terranes of high grade metamorphic

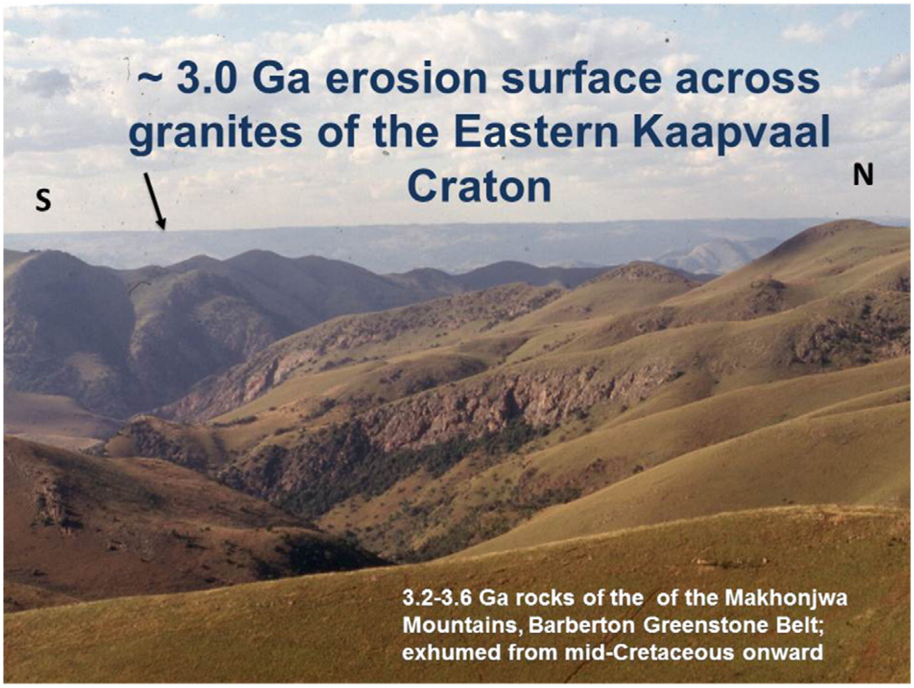

Fig. 3. Archean peneplain (ca. 3.0 Ga) cut across 3.08–3.66 Ga crystalline basement of the Eastern Block of the Kaapvaal Craton, including the Barberton Greenstone Belt rocks (foreground). By 3.0 Ga, about 7–10 km of early Archean rocks had been eroded from above the peneplain, and covered by shallow water terrestrial and marine sequences dated between 2.9 and 2.7 Ga, including the Transvaal Group, indicating that the Archean continent had stabilised by early Neoarchean times (de Wit, 2007). Numerous locally preserved subsequent sedimentary and volcanic sequences with marine incursions indicate that the peneplain surface was episodically rejuvenated close to sea level until the late Mesozoic, after which marine sediments are absent; the peneplain is now at about 1800 masl (modified from de Wit, 2007). Note how this relatively flat present day surface of the craton 'mimics' the relatively flat Moho ca. 40 km beneath it, and that both the top and the bottom of the Archean crust of Southern Africa represent complex palimpsests, as discussed in this paper.

granitoids tectonically interleaved with clastic and chemical sediments (Cornell et al., 2006; Dewey et al., 2006; Eglington, 2006; Eglington and Armstrong, 2004). Granulite metamorphism is of high temperature and relative low pressure type, and the exposed rocks of this belt reflect mid- to lower crustal thermotectonic processes of terrane accretion along a transpressive continental margin of the Kaapvaal Craton, followed by extension during tectonic collapse and exhumation (e.g. Dewey et al., 2006). Along the eastern sector of the margin, in Natal, seismic reflection data are consistent with models that suggest the NNMB rocks are thrust northwards across the southern part of the craton (de Wit and Tinker, 2004).

The Moho of the NNMB is generally sharp, with crustal thickness exceeding 40 km, possibly reaching up to 50 km in places (Harvey et al., 2001; Nguuri et al., 2001). But these are the results from the relatively few stations located on the belt during the Kaapvaal Craton experiment, and a much clearer image can be presented from recent controlled source seismic experiments carried out in the framework of the Inkaba yeAfrica project (de Wit and Horsfield, 2006).

A near vertical reflection seismic experiment (Lindeque et al., 2007, 2011) provided a clear image of the Moho structure below the NMMB along a ca. 100 km north–south profile from Slingersfontein to Prince Albert near the tectonic front of the Cape Fold Belt (Fig. 1). A sharp, continuous Moho, undulating slightly between 42 and 45 km depth is observed for the profile. Whilst most of the crystalline crust comprises stacked listric fragments, the lowermost 1–2 km of the NNMB crust along this line is continuous and characterised by increased reflectivity, sub-parallel to Moho, suggesting compositional differentiation and underplating. This may represent a layer of mafic gneiss, consistent with suggestions that the NNMB has been underplated by basalts (Dewey et al., 2006; Lindeque et al., 2011). From ca. 30°S, just north of the Cape Mountains, the Moho dips southwards, reaching depths close to 45 km beneath the frontal mountain ranges. Unfortunately, for logistical reasons the profile could not be extended across the Cape Mountains.

1.5. Proterozoic Moho beneath the Cape Mountains and Agulhas Bank (Outeniqua Basin)

Also in the framework of Inkaba yeAfrica (de Wit and Horsfield, 2006), two onshore–offshore seismic refraction profiles were carried out (Parsiegla et al., 2009; Stankiewicz et al., 2008). These near parallel profiles (Fig. 1), approximately 200 km apart, started in the Karoo Basin, traversed across the Cape Mountains, and continued offshore across the Agulhas Bank where it contains the Outeniqua Basin and the Agulhas Falkland Fracture Zone (AFFZ). The eastern profile extended farther south beyond the AFFZ onto the Agulhas Plateau (Fig. 1).

The results of these seismic experiments and the reflection seismic experiment discussed above also show a slight increase in Moho depth (by 3–5 km) beneath the exhumed Cape Fold Belt front. Farther south the Moho rises very abruptly, from depth of ca. 40 to ca. 30 km over a lateral distance of ca. 50 km (Stankiewicz et al., 2008). This shallowing occurs about 50 km inland of the present day coastline, beneath a coastal plain that was shaped as a Cretaceous–Eocene marine-cut platform, and from which a thickness of up to 7 km overburden was removed during Cretaceous (120–80 Ma) exhumation of the Cape Fold Belt, and a further 1 km or so during the Cenozoic (Scharf et al., 2013; Tinker et al., 2008).

The lower crust in this region exhibits very high P-wave velocities (>7 km/s, Stankiewicz et al., 2008). This is consistent with high lower crustal and upper mantle velocities calculated as early as 1972 (Hales and Nation, 1972). Crustal rocks with such velocities are likely to represent mafic igneous rocks, or their metamorphosed equivalents (amphibolites, mafic granulites). As most seismic studies of the NNMB in other locations (e.g. Green and Durrheim, 1990; Hirsch et al., 2008) did not identify such high velocities in the lower crust, this underplating is unlikely to be a Mesoproterozoic property of the NNMB basement, and is more likely to represent mafic material added to the base of the crust during younger magmatism. The most likely events to produce

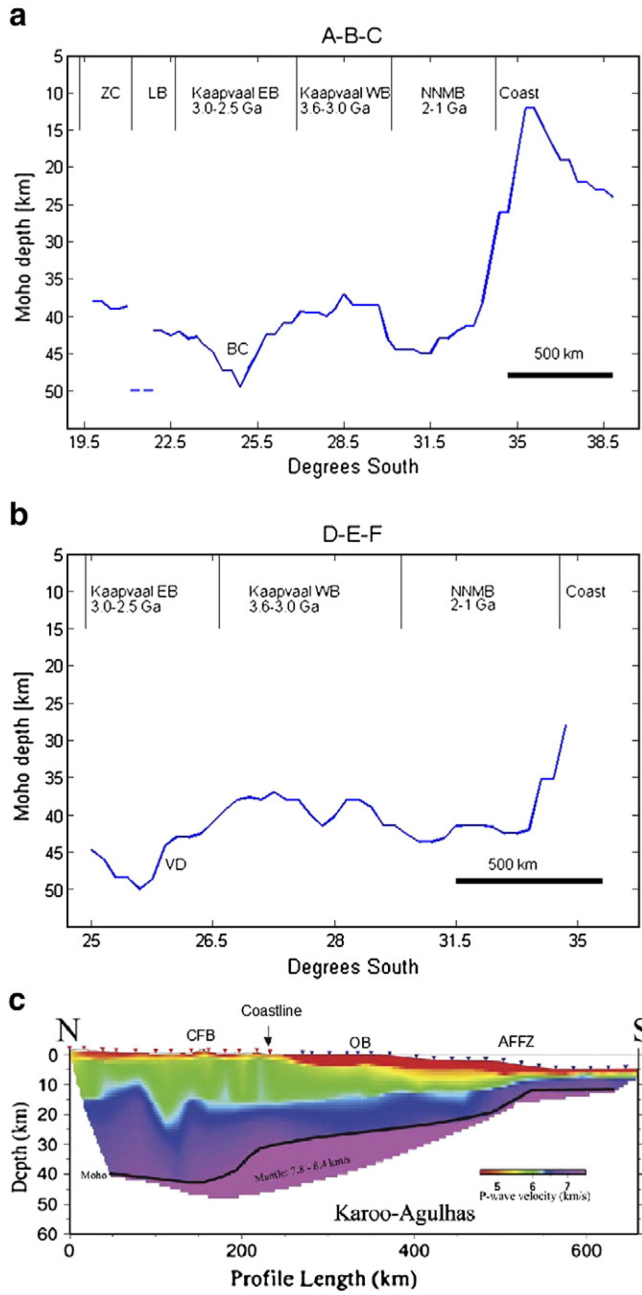

Fig. 4. a and b. Sections ABC and DEF summarising variations in depth to Moho across terranes of southern Africa, based on seismic data from Harvey et al., 2001; Parsiegla et al., 2009; Stankiewicz et al., 2002, 2008; see also Fig. 1. Note the range in age from 0.1 to 3.7 Ga. For location see Fig. 1. VD and BC mark the proximity to Vredefort Dome and Bushveld Complex. The dashed line under the Limpopo Belt shows a depth of Moho 50 km, which has been recorded (e.g. Nguuri et al., 2001), though due to sparse station coverage is not visible on our cross-section. c. Onshore–offshore P-wave velocity section across southern margin of Africa (1.0 to 0.1 Ga, respectively). Note the thickened crust beneath the Cape Fold Belt, the abrupt thinning inland from the present coast, and the gradual stretching towards the AFFZ.
Modified from Stankiewicz et al. (2008).

such volumes of underplating are those associated with Large Igneous Provinces (LIPs) in mid-Jurassic times (Karoo: ca. 182 Ma); the early Cretaceous (Etendeka: ca. 135 Ma) and/or the late-Cretaceous (Agulhas: ca. 100 Ma; see below).

Farther south the crust continues to thin, albeit much more gradually, for another 250 km underneath the Agulhas Bank, Outeniqua Basin and the Diaz Marginal Ridge (Fig. 1), until the Agulhas–Falkland Fracture Zone (AFFZ) is reached. The crust north of the AFFZ is continental of NNMB-type (Lindeque et al., 2011), but has been extensively stretched during the opening of the Southern Oceans. Parsiegla et al. (2009) performed an analysis of crustal stretching factors and concluded that two separate stretching episodes are necessary to explain the geometry of the Moho. These authors propose that rifting associated with the break-up of East- and West-Gondwana in middle to late Jurassic times, which gave rise to seafloor spreading in Riiser Larsen and Weddell Seas off Antarctica (e.g. Eagles and König, 2008; König and Jokat, 2006), resulted in crustal thinning over the whole Agulhas Bank and up to at least 50 km inland of the present day coast. The second episode is likely to have been the result of shear motion along the active Agulhas Falkland transform system in the early Cretaceous times. This is consistent with seismic reflection and well data (Broad et al., 2006; McMillan et al., 1997), which exhibit an upper Valanginian unconformity (135 Ma), as well as evidence for strike-slip faulting throughout the southern Outeniqua Basin.

1.6. Oceanic crust

Immediately south of the Diaz Marginal Ridge (DMR), the Moho rises from ~25 to ~15 km over a lateral distance of ~50 km. This rise, and the average P-wave velocity of ~6 km/s near it, is the typical characteristics of a Continent–Ocean Transition (COT) zone. Typical oceanic crust (120–160 Ma) is observed in the Agulhas Passage, until the Agulhas Plateau (80–100 Ma). This later structure rises about 2.5 km above the surrounding ocean floor, and the Moho underneath it is found at a depth between 20 and 22 km (e.g. Gohl et al., 2011; Parsiegla et al., 2008, 2009). The first attempt to measure the Moho depth here was remarkably accurate: Graham and Hales (1965) obtained a value of 21 km from offshore gravity measurements. This is almost an expected value for continental thickness, which previously led to suggestions that the Plateau is a displaced section of the African continent (e.g. Scrutton, 1973; Tucholke et al., 1981). However, large magnetic anomalies within the Plateau (Le Pichon and Heirtzler, 1968) and the lack of significant return of continental rocks from dredge hauls (Tucholke et al., 1981) can only be explained by assuming it is of volcanic origin.

Recent seismic experiments (Gohl and Uenzelmann-Neben, 2001; Gohl et al., 2011; Parsiegla et al., 2008; Uenzelmann-Neben et al., 1999) presented velocity models and identified seismic reflectors in the crust. High velocities associated with volcanic material and evidence for lava flows have conclusively demonstrated the Plateau to be of volcanic origin. Parsiegla et al. (2008) presented a model mapping in detail the extrusive cover and intruded layers of middle and lower crust. The lower two-thirds of the crustal column exhibit P-wave velocities of more than 7.0 km/s, increasing to 7.5 to7.6 km/s at the crustal base. These velocities suggest that the lower crust was accreted by large volumes of mantle-derived material to form an over-thickened oceanic layer (Gohl et al., 2011). Gohl et al. (2011) and Uenzelmann-Neben et al. (1999) estimate that the formation of the Plateau was related to the deep Bouvet plume and occurred between 100 and 80 Ma, making it by far the youngest, but by no means the thinnest crust along the transect presented here.

1.7. Summary of results and tectono-thermal implications

Fig. 1 summarises the recent results of the depth to Moho from the seismic data described above, and Fig. 4 presents a number of sections along ca. 600–2000 km swaths across the southern African crust. Depth to Moho in individual regions is as follows: the Eastern part of the Kaapvaal Craton: 44.2 +/− 3.2 km; the Western part of the Kaapvaal Craton: 40.1 +/− 2.7 km; the Zimbabwe Craton 38 +/− 1.6 km; the Limpopo Belt: 40–50; the NNMB: 43.1 +/− 1.9 km; the Agulhas Bank continental shelf: stretched from 30 to 25 km; the oceanic crust: 8 +/− 1.5 km; and the Agulhas Oceanic Plateau: 20–22 km.

The thinnest cratonic crust is Neoarchean (dated between ca. 2.5 and 3.0 Ga), and is mostly confined to the southern part of the

Western (Kimberley) Block, which is younger than the thicker Eastern (Witwatersrand) Block (ca. 3.0–3.6 Ga). Geology and vibroseismic data (Fig. 2) imply that this thinning is inherited from significant late Archean crustal stretching that took place during subaerial Ventersdorp rifting (ca. 2.7 Ga). The teleseismic data (Receiver Function analyses) indicate that thinnest Archean section of cratonic crust is felsic (consistent with the xenolith data from this region), and around 40 km for the southwestern Kimberley Block with a range across the entire block of ca. 37–43 km. Deep vibroseis probing (eight 16 second TWT lines; Fig. 1) along a composite NW–SE section across the central sector of the Western (Kimberley) Block penetrated well below a clear Moho, confirming its depths between 36 and 41 km (13–16 seconds TWT; de Wit and Tinker, 2004; Fig. 2). This work also revealed a composite crust comprising a tectonic stack of moderate to shallow, mostly west-dipping crustal sections down to Moho that lacks mafic underplating. This early stacking is likely related to the accretion between the Eastern and Western Bocks of the Kaapvaal Craton around 2.8–2.9 Ga, but is also overprinted by pervasive extension at ca. 2.7 Ga (de Wit and Tinker, 2004). The lowest values of Moho depths occur in the general region affected by extensive crustal-scale listric normal faulting and asymmetric graben formation associated with regional extension tectonics during outpourings of the 2.7–2.65 Ga mafic volcanic lavas of the Ventersdorp LIP (Hatton, 1995; Tinker et al., 2002). Detailed analyses of seismic reflection data reveal that Ventersdorp volcano-sedimentary sequences are thickest along the western margin of the Craton, including in asymmetric rifts with thick volcano-clastic sequences that can be traced more than 600 km across this sector of the Kaapvaal Craton. The shape of this Ventersdorp material is wedge-like, thickening towards the western Craton margin where the Ventersdorp basalt sequences reach up to 9 km thick. This margin has consequently been interpreted as an Archean equivalent of a modern 'passive' continental–ocean transition, with a calculated low effective elastic thickness of 7.5 to 10 km between ca. 2.7 and 2.0 Ga (Tinker et al., 2004).

During the Ventersdorp times, crustal extension and magmatism was more diffuse farther west across the Eastern (Witwatersrand) Block of the Craton, but it affected almost the entire Craton none-theless. For example, it coincides with an elevated palaeo-thermal gradient peak of at least 40–65 °C at ca. 2.7 Ga affecting the mid-crust as far east as the present eastern margin of the Kaapvaal Craton, and as far north as the central Limpopo Belt. This is corroborated by ultra-high temperature metamorphism at 2.72 Ga in the lower and middle crust determined by U/Pb thermochronology on minerals from lower crustal xenoliths retrieved in kimberlites from the south central region (Schmitz and Bowring, 2003); and by high-grade anatectic melting of the middle crust in the Ancient Gneiss Complex in Swaziland at 2.73 Ga (Taylor et al., 2010, 2012). Unfortunately there is no seismic data anywhere across the Ancient Gneiss Complex to test for depth to Moho and possible mafic underplating at 2.7 Ga.

Stratigraphic and basin analyses indicate that after 2.0 Ga, the passive western continental margin of the craton recovered to its present day effective elastic thickness value of 60 to 70 km (Doucouré and de Wit, 2003; Tinker et al., 2004), following tectonic loading across the western edge of the Kaapvaal Craton by east-directed thin-skinned thrusts between ca. 1.8 and 1.9 Ga, the age of regional Olifantshoek deformation in the tectonic Kheis Belt that flanks the entire western margin of the craton (Tinker et al., 2004).

Old diamonds (ranging from ca. 1.5 to 3.2 Ga; Pearson, 1999; Pearson et al., 1998; Shirey and Richardson, 2011; Shirey et al., 2002) retrieved from kimberlites in the general area of relative thin cratonic crust indicate that the low elastic strength of the Archean lithosphere in the Paleoproterozoic here was not due to the lack of thick, mechanically strong and relatively cool subcontinental mantle lithosphere. The low palaeo-elastic crustal thickness was therefore most likely related to the relative hot extended lower crust during Ventersdorp times at ca. 2.7 Ga, consistent with a location near the western passive continental margin of the craton throughout the late Archean to Paleoproterozoic. This is apparently still reflected in its relative shallow felsic crust here, despite subsequent partial recovery during thermal subsidence and Transvaal marine transgressions (de Wit and Tinker, 2004; Tinker et al., 2002).

The oldest, Paleo- to Meso-Archean section of the craton corresponds to the thickest cratonic crust, with an average thickness around 44 km (ranging from 41 to 48 km). However, the thickest component is near-coincident with the area covered by the 2.05 Ga Bushveld Igneous Complex (>60.000 km^2 in aerial extent). It is likely therefore that a contribution to this thickness reflects Paleoproterozoic addition of mafic material underplated below the lower granitic crust, and by up to 8 km thick intrusions into the upper crust (James et al., 2001; Nguuri et al., 2001; Wright et al., 2003). It is likely that this mafic underplating extended north also beneath the Limpopo Belt (Gore et al., 2009). A 5–10 km addition of mafic material to the lower crust across such a wide region is consistent with the seismic data and numerical simulation of plume-like activity during punctuated emplacement of a large volume of mantle melt to form a late Paleoproterozoic LIP (e.g. Sobolev et al., 2011).

Thus, whilst Archean rigidity of this crust of the Eastern Block was already well established by ca. 3.0 Ga (de Wit et al., 1992, 2011; Moser et al., 2001), the lowermost crust of this cratonic fragment experienced at least two significant episodes of remobilisations and mafic underplating (at 2.7 and 2.0 Ga) that significantly reshaped an early Archean Moho.

We can confidently conclude then that neither of the Kaapvaal crustal blocks nor the Limpopo block appears to have preserved pristine Archean thicknesses from the time they were first stabilised. In the case of the Western and south-central Eastern Blocks, thinning and lower crustal melting occurred about 2.7 Ga, and mid-crustal melting at this time extended in many places as far as the eastern Margin of the Craton. Additionally, the northeast of the Eastern Block, and likely the Limpopo Belt, were underplated by hot mafic mantle material around 2.0 Ga. Earlier, the crust of both blocks had already experienced extensive tectono-thermal remobilisation histories. For example, widespread lower crustal melting occurred during a prolonged period of extensional deformation throughout most of the Eastern Block at 3.08–3.14 Ga, creating a thick brittle 'carapace' of upper granite-felsic volcanic crust at that time (Armstrong et al., 2006; de Wit et al., 1992; Dirks et al., 2009; Hart et al., 2004; Moser et al., 2001; Schoene et al., 2008; Taylor et al., 2012; Zeh et al., 2009; e.g. see Fig. 3). By contrast the Western Block along the Colesberg Lineament and the flanks of the Limpopo Belt likely had formed thickened crust during the accretion tectonism in the early Neoarchean. Other more local Archean events are beyond the scope of this paper, but the reader is referred to Eglington and Armstrong (2004) for further documentation.

It is safe then to conclude that on a regional scale, Neoarchean compressional tectonism at ca. 2.9 Ga, followed by extensional tectonism associated with the Ventersdorp LIP-event at 2.7 Ga, and Paleoproterozoic basaltic underplating at 2.0 Ga, significantly reshaped the original Mesoarchean and Neoarchean Moho of the Eastern Block, and the Western Block, respectively. These two different processes – tectonic and magmatic – changed the seismic and petrologic characters of the Archean lower crust beneath most if not all of the Kaapvaal Craton and, with contemporary melting both sharpened and diffused the original Moho transition, respectively. Beneath the Bushveld Complex especially, Archean crustal thickness and characteristics must have changed beyond recognition, since the mega-scale of this Proterozoic intrusion must have induced a complete restructuring of the Moho (c.f. Sobolev et al., 2011).

1.8. Remnants of Archean Moho structures exposed at surface

The overview above implies that reconstructing the thickness and internal structure of pristine Archean crust and Moho remains a formidable challenge. One way to realise this quest is to focus on geological, geophysical and geochemical studies of crustal sections with potential Mesoarchean Moho transitions exposed at surface. Both Archean oceanic and continental examples have been identified on the Kaapvaal Craton. The former in the craton's greenstone belts, especially in the Barberton greenstone belt that contains Mesoarchean ophiolite fragments (3.2–3.4 Ga) related to suprasubduction processes (Furnes et al., 2013a, in press). These oceanic crustal segments were tectonically stacked during 3.2–3.1 Ga accretion processes of the Eastern Block of the Kaapvaal (Armstrong et al., 2006; de Wit et al., 2011; Furnes et al., 2012, 2013a; Taylor et al., 2012). In the case of the Barberton greenstone belt, well exposed sections containing juxtaposed chemical enriched and depleted plutonic ultramafic components, which were deformed at high temperatures, occur in tectonic contact with 2–3 km thick basaltic crustal sections (de Wit, 2004; de Wit et al., 2011) that do not display oceanic plateau geochemistry (Furnes et al., 2012, 2013a, in press). It is therefore highly probable that 'normal' oceanic crust–mantle transitions are preserved in the plutonic sequences. The true original thickness of this Archean ocean crust preserved on the Kaapvaal Craton cannot yet be precisely determined, since the crustal and mantle sections are mostly tectonically separated, but is highly unlikely to have been greater than 10 km (de Wit et al., 1987, 2011; Furnes et al., 2012, 2013a). Work is on-going to restore the sections, but appropriate geophysical data is still wanted to further guide this work.

In the case of continental crust, a unique Archean 'crust-on-edge' profile is exposed at the Vredefort circular dome, interpreted as a mega-impact site with a central uplift that formed at 2.02 Ga near the centre of the Kaapvaal Craton across which the vertical architecture of Mesoarchean crust can be viewed (Gibson and Reimold, 2008; Hart et al., 1981; Slawson, 1976). The impact structure has exposed a Mesoarchean crustal section 20–36 km thick, possibly down to the Archean Moho (Hart et al., 1990, 2004; Moser et al., 2001), although there is considerable controversy about this interpretation (e.g. Lana et al., 2003, 2004). Below we briefly highlight the Vredefort example which clearly provides crucial information not directly available through present day deep seismic analyses, but may provide a solid guideline, together with the crustal xenolith data, with which to unravel the character of pristine Archean continental crust

1.9. Is Archean Moho exposed in the Vredefort dome?

In the Neoarchean crust of the Kaapvaal, the Vredefort dome provides a continuous, albeit very poorly exposed, >20 km-thick section through granitic rocks representing middle to lower crust. The section is unconformably overlain by upper crust comprising more than 15 km of overturned Neoarchean sedimentary and volcanic rocks. Roughly restored, this yields a total thickness of some 36 km of crust, in turn underlain by ultramafic rocks with residual mantle geochemistry.

The middle crustal layer of this section is an 8–12 km thick granite gneiss domain composed of 3.08–3.1 Ga massive to slightly foliated granites. The lower crust component is a ca. 10–12 km thick charnockites-enderbite (granulite) banded gneiss complex dated between 3.1 and 3.5 Ga. It also contains relic fragments of ca. 3.5 Ga supracrustal rocks derived from greenstone belt material, but there is no evidence of extensive mafic granulites here (Lana et al., 2004).

The contact between the middle and lower crust is transitional over a distance of several kilometres (Armstrong et al., 2006; Hart et al., 1990, 2004; Lana et al., 2003, 2004; Moser et al., 2001). The lower crustal rocks are in turn underlain by depleted ultramafic rocks, but the contact with the granite crust is not exposed (Hart et al., 1990).

Originally a radial section across the Vredefort dome was interpreted to reveal a continuous Archean continental 'crust-on-edge' section ca. 36 km thick (Hart et al., 1981; Slawson, 1976) that terminated in the centre of the dome with an underlying, but unexposed, depleted ultramafic section (Hart et al., 1990; penetrated through cored-drilling). For nearly three decades this section was interpreted as a relatively pristine transect through Archean continental crust into the upper mantle across a relatively sharp Moho transition zone. Subsequent field work has shown that the section is geologically much more complicated to restore yet with any degree of confidence (e.g. Armstrong et al., 2006; Gibson and Reimold, 2008; Hart et al., 2004; Lana et al., 2004).

Structural work suggests the section comprises a stacked sequence of different granitoid fragments (Hart et al., 2004) that make up a deformed arc-like complex (Armstrong et al., 2006; Gibson and Reimold, 2008). However there is unresolved debate about whether the lower crustal section is predominantly 3.5 Ga (Hart et al., 2004) or is younger than 3.2 Ga (Armstrong et al., 2006), and at present there is no consensus on how to geologically restore the crustal section prior to impact; the outcrop is simply too poor, and the geophysics is simply not of the fidelity yet to compensate for the lack of surface exposures. However, if the lower section of the continental crust at Vredefort is representative of the Mesoarchean, then its distinct felsic nature supports the model of an Archean crust–mantle transition in which there is little mafic (granulite) material in the lower crust (c.f. Durrheim and Mooney, 1994).

Since the Vredefort section may indeed be the world's only preserved section through Archean crust directly underlain by mantle material, it deserves new attention through combined deep geophysics and drilling.

Finally, if the interpretation of the ca. 36 km crust-on-edge section has merit, and because the Vredefort dome is today underlain by 36–40 km crust with a subhorizontal Moho (Durrheim et al., 1991; Wright et al., 2003), the present Moho must be entirely annealed and of post-2.02 Ga age; it likely formed by crustal flow and melting following impact, and is not therefore pristine Archean Moho. Because this younger Moho can be connected and correlated with the 'flat' regional seismic Moho of the craton, this provides the most compelling evidence that the present day Moho is mostly a relatively young subhorizontal tectono-thermal surface that has been repeatedly rejuvenated to truncate across the complex stacked crust and amalgamated blocks that make up the Archean craton. Thus the Moho beneath the Kaapvaal Craton geometrically simulates the craton's erosional peneplain at surface: both the top and the bottom of the crust of Southern Africa are complex palimpsests.

2. Discussion and concluding remarks

The most important first order observation of our analyses shows that there is no simple age–depth relationship, and no obvious thickening of crust from the Archean to the Present. If anything our swaths across the 2000 km of crust dated between 100 and 3600 million years shows the opposite — a general decrease in depth to Moho towards the present, except beneath the Agulhas Plateau. But the latter is likely related to relative recent crustal plume activity linked to the African lower mantle superswell (e.g. Burke et al., 2008; Gohl et al., 2011; Torsvik et al., 2010). Above all, however, the results cannot be interpreted to reflect a simple secular change of Moho over this time period. Results from the Kaapvaal Craton and surrounding terrains indicate that there is significant variation in depth to Moho beneath both Archean and Proterozoic crust. The data also shows that the Archean crust of southern Africa displays as much crustal diversity in thickness as its adjacent Proterozoic crust; and that these variations are similar to those recovered across present continents elsewhere at averaged regional scales (e.g. 5×5 or $2 \times 2°$ as in CRUST 5.1 and its successor 2.0, respectively; Mooney et al., 1998; http://igppweb.ucsd.edu/~gabi/crust2.html), except below large active mountain systems. Debate about depth to pristine Archean Moho based on seismic analyses of southern Africa crust therefore remains unresolved.

Wever (1992) argued that the data presented from all Archean terrains was selective, in that sometimes minimum values for crustal thicknesses of Archean crust were compared to maximum thicknesses of Proterozoic crust, thus compromising the statistical significance of the results. He also pointed out that some relatively thick Archean regions were ignored where later tectonic thickening could be inferred (e.g. the Kapuskasing structure of the Superior Craton, Canada; Percival and West, 1994). Since information for tectonic thickening (or thinning) or significant crustal erosion is not always objectively available from surface observations of Archean cratons, the original depth to Moho of these regions becomes prone to subjective analyses (Wever, 1992). Comparisons of Archean and Proterozoic crustal structure should be considered, therefore, in unison with knowledge of local geology, since differences that do exist may be more related to local or regional tectono-thermal overprints than due to secular changes, so that only adjacent terrains should be compared before original thicknesses may be restored. In that respect, the tectonically amalgamated Archean Kaapvaal Craton and the Proterozoic NNM Belt provide a near unique opportunity to explore such comparison; and at first glance the seismic Moho looks remarkable simple as a relatively young feature that cuts across the tectonic boundary between these two blocks, truncating older tectonic fabrics in the overlying crust.

But there is a further uniqueness in being able to compare the above mentioned findings with those from Canada, in order to test our findings against global scale generalisations of a crustal age versus thickness relationship. It is important to be able to do this against the Canadian Lithoprobe data-base because of its unrivalled quality on both continental-scale and regional details that have been carefully integrated with crustal geology and geochronology (Clowes, 2010; Percival et al., 2012).

First, there are many apparent similarities between our analyses of the Kalahari Shield and the Moho findings across the Archean and Proterozoic sections of the Canadian Shield. But, compared to the Kaapvaal Craton, the Neoarchean crust in Canada appears to have more pristine Neoarchean Moho preserved beneath the Superior Province (32–40 km), since in several places moderate-dipping Archean sutures zones can be traced across near flat sections of Moho into the underlying mantle (Calvert et al., 1995; Cook et al., 2010; Ludden and Hynes, 2000). Beneath the Western Block of the Kaapvaal Craton, which is a near age- and rock-equivalent of the Superior Province, similar features beneath the Moho have been proposed to exist (Fig. 2), but these are based on poor seismic images and cannot be traced with similar confidence across the Moho into overlying crust or be shown as unequivocally Archean in age (de Wit and Tinker, 2004).

Second, the Moho transition across the Archean Superior Province to the Mesoproterozoic Grenville Belt represents the equivalent transition across the Archean–Mesoproterozoic Moho boundary along our Kalahari Shield transect, since the Grenville Belt is a near age- and rock-equivalent of the NNM Belt, and the tectonic evolution of both belts was spatially closely linked during the time of the Mesoproterozoic amalgamation of Rodinia supercontinent (Lindeque et al., 2011). In both cases, the Neoarchean crust thickens from 40 to 45 km where it under-thrusts their respective Mesoproterozoic provinces (compare de Wit and Tinker, 2004 and Ludden and Hynes, 2000). But there are also many local and regional variations in Moho relief, and in places the Moho beneath the Grenville Belt reaches 50 km (Ludden and Hynes, 2000). Nevertheless, the relatively flat Moho sections beneath the Grenville and the NMMB are remarkably similar in the way they truncate their overlying, duplexed crust.

Much of the Moho beneath both the Archean and Mesoproterozoic provinces of Canada has also been extensively deformed by shearing and plastic flow during several episodes of metamorphism and melting of the lower crust (Ludden and Hynes, 2000). In many places a sharp near-flat Moho is discordant to the listric crustal structures above it due to 'tectonic shingling' (c.f. Cook, 1986), as is the case with similar structures above the relatively thin and flat Moho of the south-central part of the Kaapvaal Craton (de Wit and Tinker, 2004; James et al., 2003). The relatively low Archean crustal thickness in different parts of the western Superior province has been attributed to tectonic thinning by ductile shear and flow, as is the case for the south western Kaapvaal Craton described in this paper. Thus, whilst the Lithoprobe sections should be considered as a global baseline for late Archean crustal characteristics, these Archean terranes have also been subjected to later transformation. Therefore, here too, the precise variations of original Archean crustal thicknesses remain open to interpretation and require careful integration (and restoration) with knowledge of thermo-tectonism documented at surface, before a true formation thicknesses can be confidently established (Cook et al., 2010).

In all Lithoprobe transects, investigators have invoked modern plate tectonic processes to explain the seismic structures, regional tectonic histories and variations in Moho depths and characters, implying that relationships between tectonics and Moho structure also can be extended into the Archean (e.g. Eaton, 2005; Cook et al., 2010). It is beyond the scope of this paper to explore further details of plate tectonic interpretations for the Azanian Craton and Kalahari Shield. Suffice it to say that, similar to the Canadian examples, the observations in southern Africa are also compatible with such lithosphere processes (de Wit et al., 1992; Furnes et al., 2012, 2013a; Taylor et al., 2012, and references therein).

In summary, the Archean Kaapvaal Craton is not likely to reveal pristine Archean crustal thickness variations, and its greatly remobilised form cannot therefore provide unequivocal data for Archean geodynamic models dealing with secular changes in continental crust formation.

The only pristine Mesoarchean continental section that may preserve original Moho and crustal characteristics is exposed near the surface in and around the Vredefort impact site (ca. 2.02 Ga). There is at present no evidence from the Vredefort 'crust-on-edge' section that its thickness was substantially less than modern continental crust. However, whilst the interpretations of the field observations may eventually converge, the global significance of using this single site to reconstruct a plate tectonic Archean Earth will always remain controversial.

This mirrors the controversies related to the depth to Moho of Archean oceanic crust, believed to have been 15 to 23 km thick based on thermal arguments (e.g. van Thienen et al., 2004, and references therein), but for which there is no direct observational evidence. By contrast, field evidence from ophiolite fragments preserved in the Barberton Greenstone of the Kaapvaal Craton Belt (e.g. Furnes et al., 2012, 2013a, in press) suggests that preserved Archean oceanic sections, when restored, may have been thinner by a factor of two to three compared to the theoretical values based on calculations with potential mantle temperatures much greater than today (e.g. de Wit, 2004; de Wit et al., 1987, 2011; Furnes et al., 2012, 2013a, in press). But again, the global significance of these single locations from which to reconstruct an Archean plate tectonic Earth remains unresolved.

Finally, the regional crustal model of the Kaapvaal Craton includes a fast, felsic isotropic upper crustal layer (ca. 2.8–3.0 Ga) above a chemically intermediate and highly anisotropic, layered lower crust of 3.0–3.5 Ga age (e.g. Youssof et al., submitted for publication), consistent with observations in the crustal sections at Vredefort and the Ancient Gneiss Complex (e.g. Gibson and Reimold, 2008; Moser et al., 2001; Taylor et al., 2012). Youssof et al. (submitted for publication) interpret the substantial anisotropy in the lower crust as a result of a gneissic fabrics formed during tectono-magmatic processes that were active after cratonic assemblage (e.g. 2.0–2.7 Ga). The crustal contribution to the total anisotropy of the craton is surprisingly high and the depleted mantle of the cratonic area has 30–40% weaker residual anisotropy than previously believed (James et al., 2003; Nair et al., 2006; Nguuri et al., 2001; Nui and James, 2002; Silver et al., 2001, 2004). But this anisotropy clearly cuts across all major sub-vertical Archean shear zones, except within and flanking the central Limpopo Belt (Fig. 1; Silver et al., 2001, 2004; Youssof et al., submitted for publication), some of which transect the entire crust here down to Moho and possibly beyond in places (de Wit and Tinker, 2004; Good and de Wit, 1997). The anisotropy also transects the Mesoproterozoic east–west tectonic fabrics of the NMMB crust, and cuts across the tectonic boundary of the Kaapvaal-NMMB at a high angle. The fact that the anisotropy in the subcontinental mantle and the lower crust is sub-parallel (Fig. 1) suggests they have a common post-Archean origin.

The new data analyses of Youssof et al. (submitted for publication) show, for the first time, strong azimuthal crustal anisotropy in the lower crust of the entire southern African shield, and must therefore be relatively young and acquired predominantly during ultra-high temperature deformation and annealing post-1.0 Ga, and likely post-0.25 Ga because the mantle and the crustal anisotropy cut across all Mesoproterozoic tectonic fabrics mapped at surface (shear zones, faults, foliations), including the 1.0 Ga tectonic boundary between the Kaapvaal Craton and the NNMB and the 0.25 Ga tectonic front of the Cape Fold Belt. Thus, this anisotropy is likely related to a period of time during which the Gondwana/Africa plates moved near 50° latitude from the end-Paleozoic (300 Ma) to end-Mesozoic (65 Ma), and a further ca. 10° during the Cenozoic in a direction subparallel to the azimuth of the anisotropy (e.g. Buiter et al., 2012; de Wit, 2007; Vinnik et al., 1995; for an excellent animation of this movement see: http://www.reeves.nl/upload/SouthAtlantic1.gif). Throughout this time-span there was substantial episodic heating in the sub-continental mantle in response to plume activity of the Karoo LIP (ca. 180 Ma) and the Bouvet LIP (ca. 100 Ma), as well as widespread associated kimberlites activity across a wide region of the Kalahari Shield (e.g. Bell et al., 2003; Jelsma et al., 2004).

Subcontinental mantle seismic anisotropy was originally attributed to plate motion processes during the Mesozoic–Cenozoic (Vinnik et al., 1995). Notwithstanding the subsequent recognition of remnant Archean anisotropic fabrics preserved in the mantle lithosphere (Adam and Lebedev, 2012; Silver et al., 2004), this late Phanerozoic plate motion now appears the most parsimonious solution to the age of the more extended crust–mantle anisotropy found by Youssof et al. (submitted for publication). This is also because it is now also more clearly established that the horizontal plate movement during the Cretaceous was associated with heating of both the oceanic lithosphere (by the Karoo and Bouvet hotspots) and the continental lithosphere during extensive contemporaneous kimberlite activity across the Kalahari Shield and the epeirogenic uplift of southern Africa (e.g. de Wit, 2007; Torsvik et al., 2010). This coupled crust–mantle anisotropy also confirms that the motion of the Gondwana/Africa plates did not delaminate the Archean crust substantially from its subcontinental mantle, as is evident from the presence of old diamonds in the young kimberlites (e.g. Pearson, 1999; Pearson et al., 1998; Shirey and Richardson, 2011; Shirey et al., 2002) that confirm an overlapping age between the old cratonic mantle keel and its overlying crustal lid; and now also from the relative young age of the regional Moho.

The fundamental question that lingers on in Craton studies like the Kaapvaal Craton then is how substantial and repeated heating and thinning of its crust associated with large mantle derived volumes of the world-class LIPs that have affected this craton episodically since its Archean stabilisation (e.g. the Ventersdorp, the Bushveld, and Karoo LIPs) can be reconciled with the retention of old ages of lithospheric mantle diamonds brought to surface by Paleoproterozoic to Recent kimberlites. Apparently widespread heating processes that regionally affected old crust with ages commensurate to those reflected in the diamonds have not pervasively affected its underlying lithospheric mantle. It is likely therefore that the Moho beneath the Kaapvaal Craton predominantly reflects a recurring *thermo-chemical* erosion zone along which tectonic décollement was restricted, so that the crust and the depleted mantle lithosphere retained cohesion for more than 3.2 Gyr along a complex long-lived palimpsest.

It would be of great interest to calculate the amount of strain needed to reset the anisotropy both in the crust and mantle, yet retain their overall cohesion, but such calculations would depend on a significant number of (presently) unknown parameters, particularly the scale and rates of transient geothermal gradients before and after LIPS events, kimberlite activity, and fluid mobility during related metamorphic/melting events. We are only now learning how to best interpret the geochemistry of mantle xenoliths and xenocrysts from different age kimberlites in terms of transient (palaeo-) geotherms and fluid mobility (e.g. Bell et al., 2003 and personal communications, 2013), so that reliable palaeo-strain gradients across the Moho remain to be determined.

In summary, confidently tying the age and character of the Archean and Proterozoic crust of southern Africa to their depth diversities is still to be resolved, and reconstructing pristine Moho of the early Earth remains a formidable challenge. More detailed comparison between surface exposures such as at Vredefort and deep geophysics is needed to make new inroads.

Finally, the entire Agulhas bank is underlain by a shallow Moho (10–30 km), reflecting Cretaceous crustal thinning of the Mesoproterozoic NNMB crust by more than 50% because there is also considerable new Cretaceous underplating seen in the refraction seismic results. In addition, in some oceanic sectors like the Agulhas Plateau, old oceanic crust has been significantly thickened to more than 20 km by plume activity. This crustal thickening process should have affected significant continental sectors of southern Africa, including the Archean Craton, several times during large recurring punctuated mantle perturbations (e.g. the Ventersdorp LIP at ca. 2.7 Ga; the Bushveld LIP at ca.2.0 Ga; the Karoo LIP at ca. 0.2 Ga and the Agulhas LIP at 0.1 Ga). Apart from beneath the Bushveld and the Agulhas Bank, the Moho transitions described along our swaths apparently do not reveal significant mafic underplating. The effects of these recurring mantle perturbations on the sub-continental mantle and the Moho still wait to be unravelled.

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