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Tectonic model of the Limpopo belt: Constraints from magnetotelluric data

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ABSTRACT

Despite many years of work, a convincing evolutionary model for the Limpopo belt and its geometrical relation to the surrounding cratons is still elusive. This is partly due to the complex nature of the crust and upper mantle structure, the significance of anatectic events and multiple high-grade metamorphic overprints. We use deep probing magnetotelluric data acquired along three profiles crossing the Kaapvaal craton and the Limpopo belt to investigate the crust and upper mantle lithospheric structure between these two tectonic blocks. The 20–30 km wide composite Sunnyside-Palala-Tshipise-Shear Zone is imaged in depth for the first time as a sub-vertical conductive structure that marks a fundamental tectonic divide interpreted here to represent a collisional suture between the Kaapvaal and Zimbabwe cratons. The upper crust in the Kaapvaal craton and the South Marginal Zone comprises resistive granitoids and granite-greenstone lithologies. Integrating the magnetotelluric, seismic and metamorphic data, we propose a new tectonic model that involves the collision of the Kaapvaal and Zimbabwe cratons *ca.* 2.6 Ga, resulting in high-grade granulite Limpopo lithologies. This evolutionary path does not require a separate terrane status for each of the Limpopo zones, as has been previously suggested.

1. Introduction

Precambrian regions hold the key to understanding the tectonic processes that prevailed during the early and middle Archaean. Many questions still remain to be answered regarding the amount of heat available at the time, the onset and dominance of early plate tectonic processes and crustal generation these processes and/or by plumes. The greatest impediment is the relative paucity of preserved Archaean rocks, compared to inferred crustal generation during the Archaean, and the identification of Precambrian structures is often masked by secondary tectonic events.

The highly complex Limpopo belt in Southern Africa provides a natural laboratory to investigate these questions and elucidate the possible geological processes taking place, where structural, metamorphic and geochemical data are available. The Limpopo belt is an Archaean-aged high-grade metamorphic complex located between the Kaapvaal and Zimbabwe cratons (Fig. 1). It has an ENE–WSW trend and comprises three zones, the Northern and Southern Marginal Zones and the Central Zone, separated from each other by major thrust faults or strike-slip shear zones. Metasediments, granitoids and gneisses comprise a significant component of the rock outcrop. Due to its Archaean affinity and relatively good surface exposure, the Limpopo belt has been a focus of a number of geological, structural, metamorphic and geophysical studies (Van Reenen et al., 1987; Roering et al., 1992; De Beer and Stettler, 1992; Rollinson, 1993; Durrheim et al., 1992). More recently, a Geological Society of America Memoir (207) on the origin and evolution of high-grade Precambrian gneiss terranes focused specifically on the Limpopo belt (van Reenen et al., 2011).

Several models have been suggested regarding the formation and deformation of the Limpopo belt and these are presented and discussed in Section 3. Seismic tomography models suggest that the lithospheric mantle beneath the Limpopo belt is generally similar to that of the Kaapvaal and Zimbabwe cratons, i.e., fast velocities, implying thick, cold lithosphere (Li, 2011; James et al., 2001). In contrast the crustal structure of the Limpopo belt is highly complex, with evidence of polytectonic events that include metamorphism, magmatism, crustal uplift and structural deformation (Kramers et al., 2011). The nature of the horizontal movements during continental accretion, including the orientation and rate of plate movements during the Archaean, is not fully described. In addition, the deep geometry of the margin between the Kaapvaal craton and

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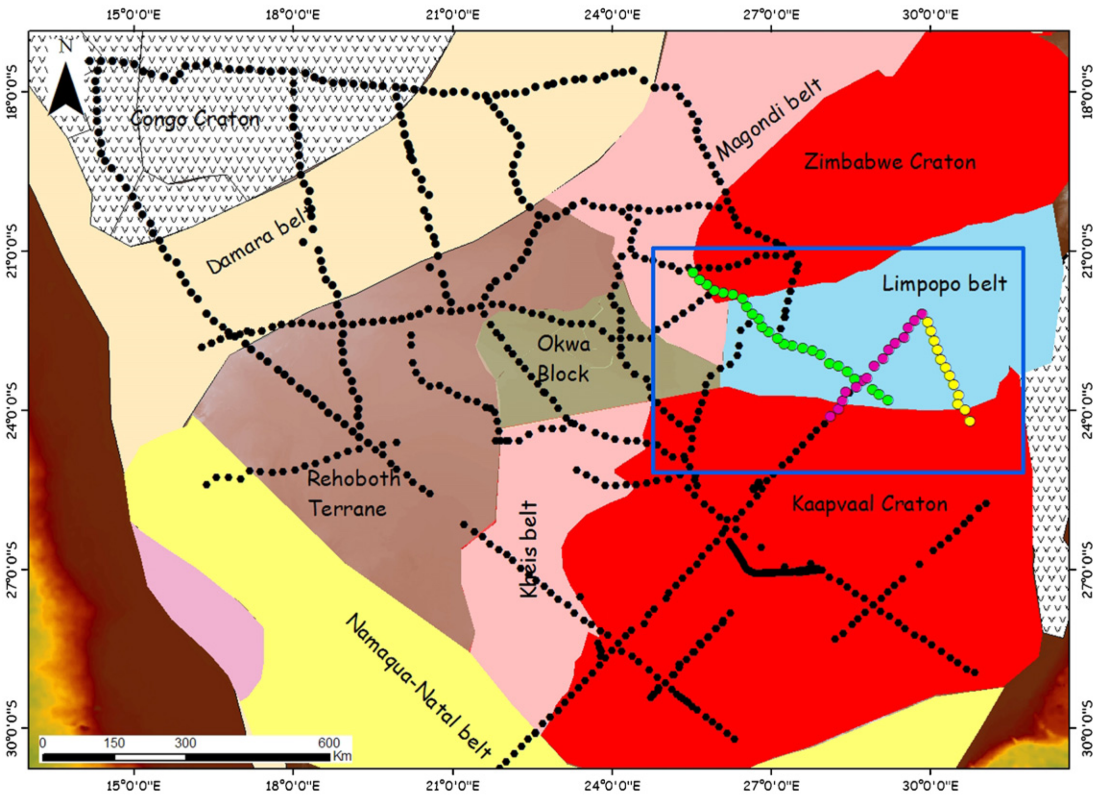


Fig. 1. SAMTEX stations (black circles) overlain on the regional tectonic map of Southern Africa, modified from Begg et al. (2009) and Webb (2009). The yellow, purple and green filled circles show locations of the MT stations for the LOW, KAP and LIM-SSO profiles respectively that are the focus of this work. The bathymetry data are from ETOPO1 courtesy of Anamte and Eakins (2009). The blue rectangle shows the area of interested, shown in Fig. 2.

the Limpopo belt, including the shear zones separating each zone, is not fully known.

In this article we attempt to address several of these issues, particularly the nature of the geometry of the shear zones separating the cratonic units, and investigate the exotic terrane status of the Limpopo belt, particularly for the Central Zone. To this end we use magnetotelluric (MT) data to image the crust and the mantle lithosphere beneath the Limpopo belt and Kaapvaal craton. The deep-probing MT technique has been used successfully in Precambrian regions all over the world to elucidate their tectonic history (Heinson, 1999; Davis et al., 2003; Selway et al., 2006, 2009; Spratt et al., 2009). In Southern Africa, the MT method has been used to image Archaean and Proterozoic boundaries and to understand the tectonic history of Archaean lithosphere (Jones et al., 2005, 2009; Hamilton et al., 2006; Muller et al., 2009; Evans et al., 2011; Miensopust et al., 2011; Khoza et al., 2011). The primary physical parameter being investigated is electrical resistivity of sub-surface materials. Given its sensitivity to resistivity contrasts, the crustal models derived from MT data are very informative in mapping basement features, the location of deep seated fault blocks (Selway et al., 2006; Spratt et al., 2009) and crustal melts (Wei et al., 2001; Unsworth et al., 2004, 2005; Le Pape et al., 2012). In the Earth's crust the primary conducting mineral phases are saline waters, graphite, sulphides, iron oxides and, in active regions like Tibet, partial melt. Mantle electrical resistivity is primarily sensitive to temperature variation and water content (Jones et al., 2012; Evans, 2012) and, to a lesser extent chemical composition and pressure. As part of the highly successful *S*outhern *A*frican *M*agneto *T*elluric *EX*periment (SAMTEX) we have collected MT data along several profiles crossing the Limpopo belt and its bounding terranes: the Kaapvaal craton to the south, Zimbabwe craton to the north and the Magondi belt to the west (Fig. 1). The LOW profile (yellow filled circles, Fig. 1) crosses the northern Kaapvaal craton, over the Hout River Shear Zone, which is thought to represent the southern limit of the Limpopo belt, into the Central Zone.

The KAP profile (pink filled circles, Fig. 1), part of which was the focus of the two-dimensional Kaapvaal craton study of Evans et al. (2011), is re-modelled here using newly developed three-dimensional (3D) techniques and also to do a holistic focused study of the Limpopo–Kaapvaal boundary. The LIM-SSO profile (green-filled circles, Fig. 1) crosses the Kaapvaal craton, Bushveld complex, Limpopo belt, Magondi belt and/or southern Zimbabwe craton and terminates close to Orapa kimberlite field. The Martin's Drift kimberlite cluster (Fig. 2) is about 50 km from MT site SSO103 (Fig. 2). The Orapa and Martin's Drift kimberlites erupted about 93 Ma and 1350 Ma respectively (Haggerty et al., 1983; Jelsma et al., 2004). These profiles, crossing the Limpopo belt and its surrounding terranes, were picked to provide a spatially adequate database to perform 3D magnetotelluric inversion and to understand the nature of the crustal and upper mantle geometry of the geology along the Limpopo belt.

2. Geological background

We define in Table 1 some acronyms that will be referred to consistently in the text.

The Limpopo belt is an ENE–ESW trending high grade Archaean metamorphic complex situated between the lower-metamorphic grade granite-greenstone Zimbabwe and Kaapvaal cratons. The three geologically-defined zones that make up the complex, the Northern Marginal, Central and Southern Marginal zones, are separated from each other by variously dipping shear zones (Fig. 2).

The *Northern Marginal Zone* (NMZ), which comprises granite-greenstone material (magmatic enderbites), is separated from the Zimbabwe craton to the north by the southward-dipping North

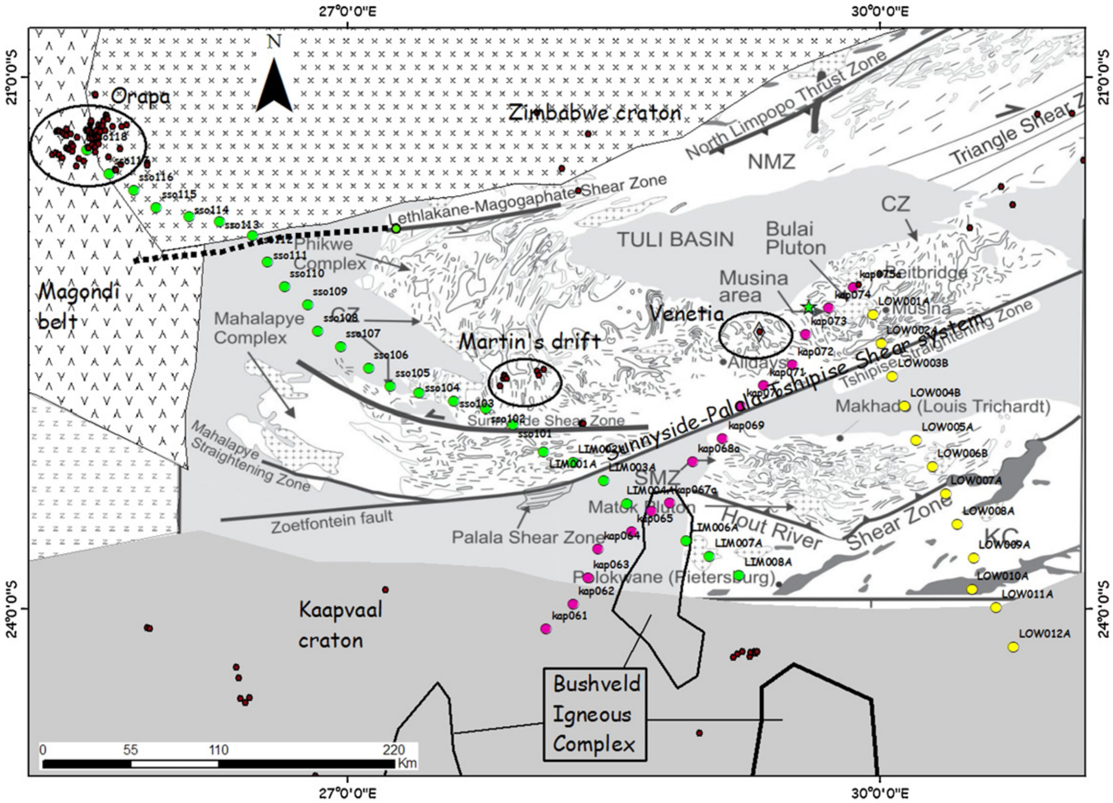


Fig. 2. Geology of Limpopo belt (modified from Kramers et al., 2011) showing major structural and geologic features. The MT station locations of the LOW (yellow circles), KAP (purple-filled circles) and LIM-SSO (green-filled circles) are shown. The red dots indicate the location of known kimberlites, including the Orapa, Martin's Drift and Venetia kimberlite fields. CZ = Central Zone; KC = Kaapvaal craton; NMZ = North Marginal Zone.

Limpopo Thrust Zone (NLTZ). The southern limit of the NMZ (the northern limit of the CZ) is marked by the south-dipping Triangle Shear Zone (TSZ) (Roering et al., 1992; Kamber et al., 1995a; Kramers et al., 2011).

The *Central Zone* (CZ) is a 3.3–2.5 Ga high grade zone that comprises metasediments, S-type granitoid gneisses and supracrustal rocks. Two periods of high metamorphic grade metamorphism are recorded in the CZ between 2.7–2.6 Ga and the other at 2.0 Ga (Smit et al., 2011). The 530 Ma diamondiferous Venetia kimberlite cluster is located within the CZ of the Limpopo belt (see Fig. 2). The Palala-Tshipise Straightening Zone (PTSZ) separates the CZ from the SMZ in NE South Africa (McCourt and Vearncombe, 1992) and has an ENE to NE trend. However, in SE Botswana the southern margin of the CZ is marked by a composite 40 km wide NW–SE striking linear structure, made up of gneisses of the Sunnyside Shear Zone and mylonites of the Palala-Tshipise Shear Zone. This complex composite structure is thus referred to as the Sunnyside-Palala-Tshipise Shear System (SPTSS) and is inferred from mineral lineation studies to have sub-vertical dip (Horrocks, 1983; McCourt and Vearncombe, 1992).

The *Southern Marginal Zone* (SMZ) is a 60 km zone that consists chiefly of enderbitic and charnokitic gneisses and, unlike the CZ and NMZ, the SMZ experience a single metamorphic event at 2.72–2.65 Ga only. The north-dipping Hout-River Shear Zone (HRSZ) is thought to mark the boundary between the SMZ and the Kaapvaal craton to the south. The 1.9 Ga Soutpansberg basin, which forms a 40 km wide, 300 km long, 7 km thick volcano-sedimentary trough, partially cross-cuts the PTSZ and developed as a graben-like basin (Tankard et al., 1982; Kamber et al., 1995a,b; Schaller et al., 1999). Granitoid gneisses and NE-trending greenstone belts (i.e., Murchison, Pietersburg and Giyani belts) make up the geological composition of the north-eastern Kaapvaal craton (Rollinson, 1993). The SMZ and the Kaapvaal craton show uniform and low $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ isotope ratios in metapelites and leucocratic granitoids (Kreissig et al., 2000) suggesting that the SMZ is a high-grade equivalent of the Kaapvaal craton. This result is in contrast to a separate terrane model for the SMZ proposed by Rollinson (1993), who argued that the different crustal histories and the separation of the SMZ, CZ and NMZ by major thrust faults pointed to separate Limpopo belt terranes. Similarly, the Zimbabwe craton, the NMZ and the CZ show elevated $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ isotope ratios, implying that these three terranes cannot be regarded as separate units from each other and were derived from potentially the same source, either in the mantle or crust, having high U/Pb ratios (Barton et al., 2006; Andreoli et al., 2011; Kramers et al., 2001, 2011).

The geometry of the shear zones bounding and separating the marginal zones and central zones from the cratonic blocks have been central to some of the proposed collisional models of the

Table 1
Description of acronyms referred to in the article.

Acronym	Description
KC	Kaapvaal craton
SMZ	South Marginal Zone
CZ	Central Zone
NMZ	North Marginal Zone
ZC	Zimbabwe craton
TSZ	Triangle Shear Zone
PTSZ	Palala-Tshipise Shear Zone
SPTSS	Sunnyside-Palala-Tshipise Shear System
NLTZ	North Limpopo Thrust Zone
HRSZ	Hout-River Shear Zone
SSZ	Sunnyside Shear Zone

evolution of the Central belt (Durrheim et al., 1992; De Beer and Stettler, 1992; Van Reenen et al., 1987; Roering et al., 1992; Treloar et al., 1992). The Kaapvaal craton forms the core of the composite Kalahari craton and formed in Archaean times *ca.* 3.7–2.6 Ga (de Wit et al., 1992). Post-stabilization processes include the development of sedimentary basins (e.g., Witwatersrand basin) at *ca.* 3 Ga and were followed by extensive volcanism (Ventersdorp magmatism) (Eglington, 2004). The very widespread platform sedimentation of the Transvaal Supergroup that overlies the Ventersdorp Supergroup represents exceptional early stability of the Kaapvall craton. Bushveld complex (Fig. 2), which intruded into the Kaapvaal craton and1 upper-most Transvaal sequence, is the largest known layered intrusion on Earth and its emplacement significantly altered the thermal and chemical structure of the Kaapvaal craton at 2.06 Ga (James et al., 2001; Evans et al., 2011). The Zimbabwe craton, which is also part of the greater Kalahari craton, is itself thought to be composed of a number of distinct tectonostratigraphic terranes, which consist of 3.5–2.95 Ga gneissic rocks overlain by 2.92 Ga assemblage of mafic and felsic volcanic rocks at its core (Blenkinsop and Vinyu, 1997; Kusky, 1998). Several 3.5–2.6 Ga greenstones belts complete the lithological profile of the Zimbabwe craton (Blenkinsop and Vinyu, 1997).

3. Tectonic models of Limpopo belt

The complex nature of the geological and structural relationships in the Limpopo belt has led to several plate tectonic and non-plate tectonic models being proposed for its evolution. These were detailed and reviewed by Kramers et al. (2011) and are summarized here.

3.1. The Neoarchaean Himalayan model

Treloar et al. (1992) were the first to propose a model of the Limpopo belt involving continental growth by accretion followed by shortening, which is similar to the Mesozoic evolution of Tibetan plateau during India–Asia collision. Central to this model was the observation of a regional structural pattern that suggest NW–SE compression, resulting in crustal thickening that also involved folding and NW-directed thrusting and lateral extrusion of crustal blocks along SW- to WSW-trending shear zones. Treloar et al. (1992) thus argued that terrane status (i.e., that each unit formed separately as unique/discrete crustal or lithospheric blocks prior to amalgamation and accretion as the Limpopo belt) for the SMZ, CZ and NMZ was not required, and that the Limpopo belt was a result of a crustal deformation event that included much of the Kaapvaal and Zimbabwe cratons. Roering et al. (1992) argued that the granulite terrane was a result of crustal thickening in response to the northward thrusting of the Kaapvaal craton over the Zimbabwe craton along the Triangle Shear Zone, *ca.* 2.7–2.6 Ga. This was followed by a metamorphic event and subsequently isothermal decompression during which rocks moved upward and spread outward onto the adjacent cratons, during a period of widespread anatexis, creating what has since being called a pop-up structure (Roering et al., 1992).

3.2. Terrane accretion models

Rollinson (1993) and Barton et al. (2006) proposed models describing the Limpopo belt formed by accretion of separate terranes of unrelated origin that constitute the NMZ, CZ and SMZ. In Rollinson (1993)'s model the distinct crustal evolutions of the zones, supported by prominent shear zones separating them, implied these blocks accreted together prior to the collision of between the Kaapvaal and Zimbabwe cratons in Neoarchaean and warrant their consideration as discrete terranes.

Barton et al. (2006) proposed a similar accretion model, but, unlike Rollinson (1993), the process involved a complex assembly of a large number of terranes between *ca.* 2.7 and *ca.* 2.04 Ga, where the SMZ, CZ, NMZ, Zimbabwe craton, Phikwe and Beit Bridge complexes, accreted, in subduction settings, to form migrating arcs that led to development of juvenile crust. This Turkic-type accretion was proposed by Sengor and Natal'in (1996) as the principal craton building process through Earth's history. In this model the Beit bridge and Phikwe complexes, in addition to the terranes defined by Rollinson (1993), show distinct P–T–t (pressure–temperature–time) paths and metallogenic signatures, that suggest they are separate terranes. Furthermore, the lack of S-type granitoid magmatism, ophiolites and syntectonic sedimentary basins led Barton et al. (2006) to argue against the continent–continent collision model (Kramers et al., 2011).

3.3. Transpression model for the central zone

In contrast to McCourt and Vearncombe (1992), who argued for a dip- or oblique-slip movement along shear zones (see below), Kamber et al. (1995a,b) interpreted the movement along the TSZ and PTSZ as being dextral-transcurrent that recorded the Paleoproterozoic transpressive collisional event which resulted in crustal thickening and uplift of the CZ. This model was later expanded by Holzer (1998) and Schaller et al. (1999).

3.4. Other models

Other models that have been invoked include that of McCourt and Vearncombe (1992) who, based on Pb isotopic and structural grounds, interpreted the CZ to be an exotic terrane that was emplaced from NE to SW as a thrust sheet facilitated by the Triangle-Tuli-Sabi and Sunnyside-Palala structures which acted as complimentary lateral ramps. Based on Pb isotope data on igneous rocks and the dip slip shear movement along these structures, which resulted in the CZ being the structurally the highest in relation to the marginal zones, McCourt and Vearncombe (1992) argued that the CZ may have represented a part of the overriding plate during the Neoarchaean Limpopo orogeny (Kramers et al., 2011).

4. Previous geophysical studies of the Limpopo belt

Several studies have been performed, each attempting to image the geometry and structure of the Limpopo belt. These are summarized in Fig. 3.

The geoelectric, seismic reflection and gravity results of De Beer and Stettler (1992) and Durrheim et al. (1992) formed the basis from which many of the collisional models were formulated (Fig. 3D). The northward and southward dip of HRSZ and the NLTZ respectively lend support to the 'pop-up' model derived by Roering et al. (1992), as outlined above. Furthermore, the gravity results delineated high density rocks in the upper crust of the NMZ and SMZ. The PTSZ did not correspond with any seismic reflection signature which led De Beer and Stettler (1992) to propose that it is a near vertical fault that penetrates the entire crust. The work of James et al. (2001) used P-wave and S-wave delay times from a broadband seismic array to map high velocity mantle roots extending to depths of 250–300 km beneath the Kaapvaal and Zimbabwe cratons and the Limpopo belt (Fig. 3B). In a related crustal study, Nguuri et al. (2009) analyzed receiver functions to map the crustal structure of the Limpopo belt (Fig. 3A). The NMZ was found to have a 37 km thick crust, similar to the Zimbabwe craton, while the SMZ had a 40 km thick crust similar to Kaapvaal craton and the Moho in both zones generated strong P-to-S conversions. However, the Moho structure from Nguuri et al. (2009)'s study appeared more

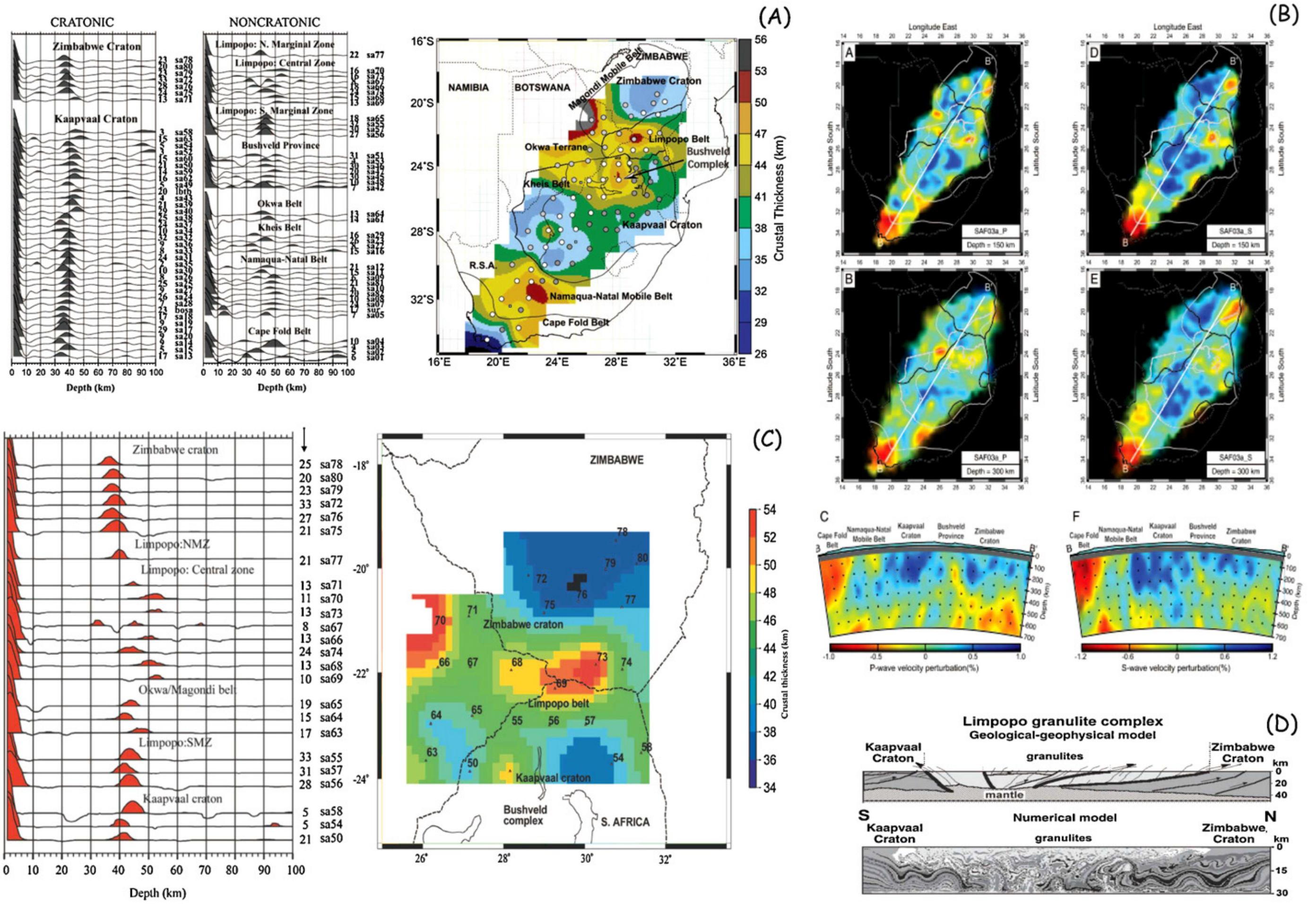


Fig. 3. Summary of geophysical studies undertaken over the Limpopo belt. The study of Nguuri et al. (2009) focussed on defining the crustal structure across the Kaapvaalcraton, Limpopo and Zimbabwe craton using seismic receiver functions (A), while the tomography work of James et al. (2001) focused on the deep mantle lithospheric structure (B). De Beer and Stettler (1992) and Durrheim et al. (1992) used reflection seismic data to study the crustal shear zones bounding the marginal zones of Limpopo belt (D) and more recently Gore et al. (2010) used seismic receiver functions to study the crustal structure of the Limpopo Central Zone.

complex beneath the CZ, corresponding with weaker P-to-S conversions.

In a second attempt, using additional seismic stations, Gore et al. (2010) analyzed receiver function data and derived a Moho map (Fig. 3C) beneath the entire Limpopo belt which appears to show thick (more than 50 km) crust for the Kaapvaal and Zimbabwe cratons, which are consistent with estimates from Nguuri et al. (2009)'s results. As noted by Gore et al. (2010), the Limpopo belt has a low elevation relative to the adjacent cratons, which is puzzling on isostacy grounds given the deeper Moho. The authors explain this discrepancy by interpreting the CZ as a remnant of a deep-rooted crustal block that did not fully rebound during denudation. The notion of a dense lower-crustal/upper-mantle root beneath the CZ is supported by the positive Bouger anomaly and could possibly be the result of magmatic underplating (Gore et al., 2010; Gwavava et al., 1992; Ranganai et al., 2002; Kramers et al., 2011).

5. The magnetotelluric (MT) method and data

The magnetotelluric method is an electromagnetic (EM) sounding technique that has evolved rapidly since its first theoretical description in the 1950s. By measuring the time variations on the surface, of the horizontal electric (E_x, E_y) and horizontal and vertical magnetic (H_x, H_y and H_z) fields induced in the subsurface, we can derive the lateral and vertical subsurface variations of electrical resistivity. The ratios of the EM fields are related, in the frequency (ω) domain, by an impedance $Z_{xy}(\omega)=E_x(\omega)=H_y(\omega)$, from which the *apparent resistivity* (i.e., the resistivity of a homogeneous half space)

$$\rho_{a,xy}(\omega)=\frac{1}{\omega\mu}\left|\frac{E_x(\omega)}{H_y(\omega)}\right| \tag{1}$$

and *impedance phase*

$$\phi_{xy}(\omega)=\frac{\arctan(E_x(\omega))}{H_y(\omega)} \tag{2}$$

can be estimated (Chave and Jones, 2012).

In Southern Africa, we have acquired broadband (periods from 0.001 s to 8000 s) and long period (15 s to over 10,000 s) magnetotelluric data as part of the SAMTEX project (Fig. 1). More than 750 stations of data were collected in Namibia, Botswana and South Africa over four field seasons along various profiles from 2003 to 2008, with station spacings of approximately 20 km and 60 km for broadband and long period data, respectively. The orientations of the profiles were chosen to transect over specific geological features of interest. Magnetotelluric broadband data were collected using Phoenix Geophysics (Toronto) MTU5 instruments, and long period data were acquired with LVIV (Ukraine) LEMI systems. In this work we model MT data collected along three profiles: LOW, LIM-SSO and KAP (Fig. 2).

The NE–SW KAP line was the focus of the Kaapvaal lithospheric study of Evans et al. (2011). We focus here specifically on the

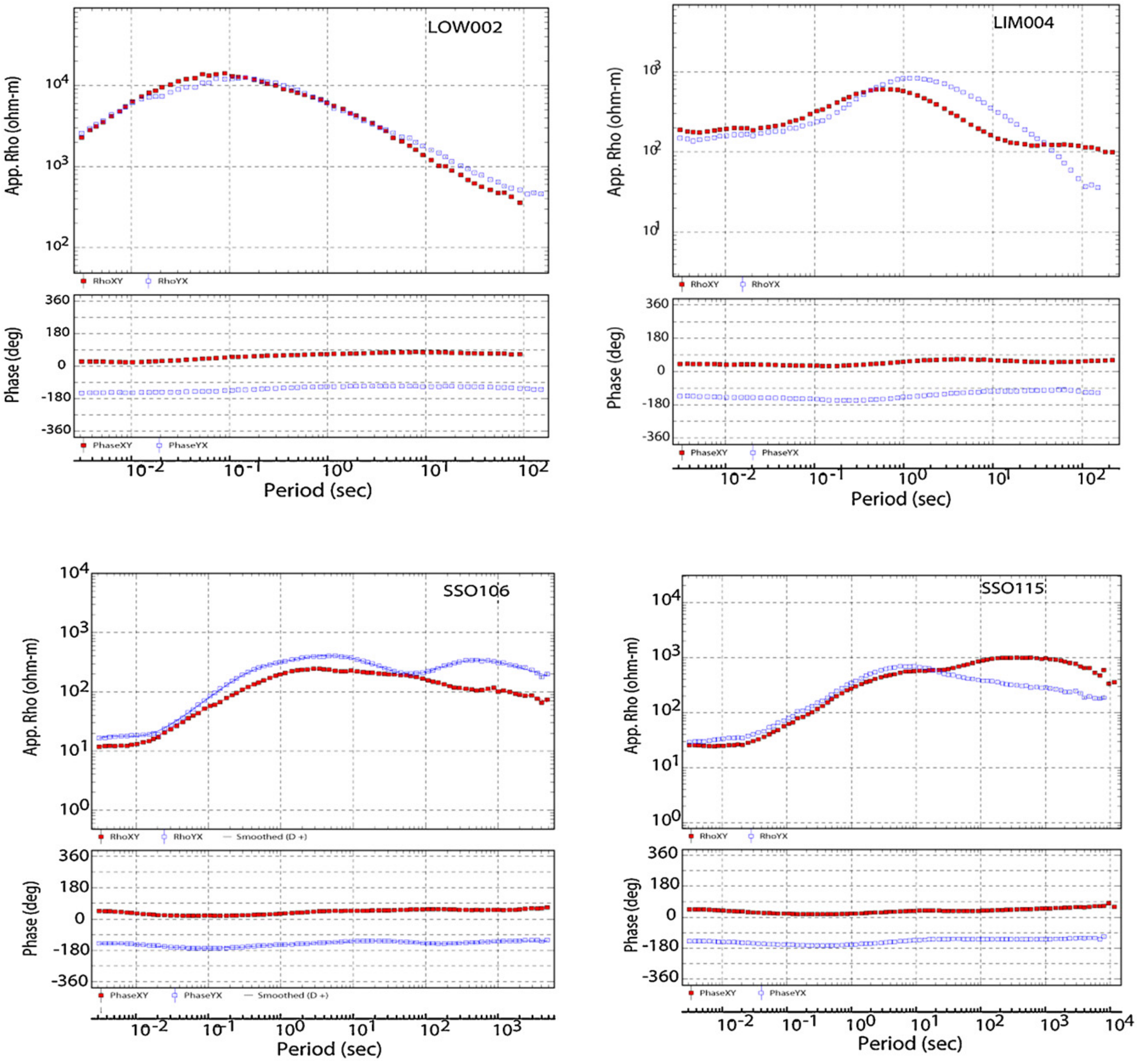


Fig. 4. MT responses curves for 4 sites, showing apparent resistivity and phase plotted against increasing period (proxy for depth). The red squares show the transverse-electric (TE) mode data, and the blue squares show the transverse-magnetic (TM) mode data.

northern half of the profile traversing over the Limpopo belt. There were several motivations for remodelling this part of the profile. Firstly, the work of Evans et al. (2011) focussed principally on defining the Kaapvaal craton lithosphere not the Limpopo belt. Secondly Evans et al. (2011) developed lithospheric models using 2D techniques. While 2D modelling is able produce first-order regional features, it cannot account for 3D data complexities, particularly off-profile features. To address this deficiency we apply a newly-developed 3D inversion algorithm (Egbert and Kelbert, 2012) to derive information on the 3D structure of the Limpopo belt. To this end only 14 stations from the KAP profile were modelled, crossing the northern limb of the Bushveld Igneous Complex (BIC), the SMZ and the CZ (purple sites in Fig. 2).

The almost NNW-SSE LOW profile comprises 12 stations crossing the northern Kaapvaal craton in the south, extending into the SMZ and part of the CZ (yellow sites in Fig. 2). Vertical magnetic field (Hz) data were recorded at 3 stations only on the LOW profile. Due to logistical and security concerns of going into Zimbabwe at the time of the survey, the LOW profile was terminated close to the South Africa–Zimbabwe border.

The NW–SE LIM-SSO profile comprises 25 stations (13 of which recorded Hz data) crossing (from SE to NW) the Kaapvaal craton, the northern limb of the BIC, the western Limpopo belt and the south-western margin of the Zimbabwe craton (green sites in Fig. 2).

5.1. MT data and processing

The recorded electric and magnetic time series data were processed using standard robust processing methods of Jones and Jodicke (1984), Egbert (1997) and Chave and Thompson (2004) (methods 6, 7 and 8 in Jones et al., 1989).

Given that multiple sites were recorded simultaneously, we employed remote referencing methods (Gamble et al., 1979) to reduce bias effects and improve the quality of the estimated MT responses. The resulting responses are shown in Fig. 4 for four representative stations on and off the Limpopo belt, where variation

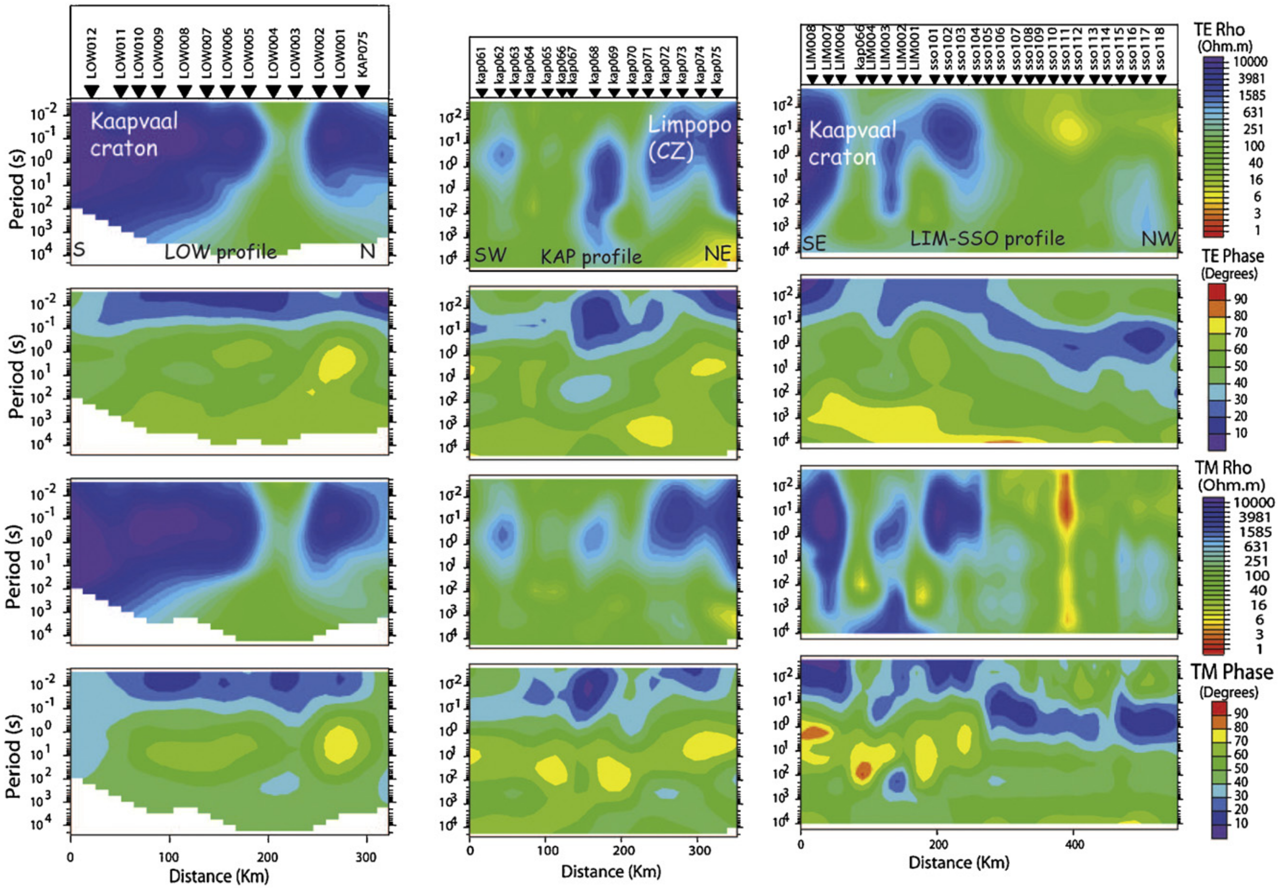


Fig. 5. Pseudosections of TE and TM mode data for the three profiles, showing apparent resistivity and phases as a function of period increasing downwards (proxy for depth). Some features, like the Kaapvaalcraton upper crust and the resistive lithologies of the Limpopo belt Central Zone are readily recognizable.

in apparent resistivity is plotted as a function of period, the latter being a proxy for depth (i.e., the longer the period the deeper the depth of penetration).

Data quality was generally very good for most stations on the KAP and LIM-SSO profiles. The LOW profile sites suffered from long period distortion and at most stations we were only able to model periods up to 300 s. However, the general resistive nature of the crust, inferred from resistivity studies of De Beer et al. (1991) in the SMZ and northern Kaapvaal craton (high and low grade granitoids), implied that depth of penetration up to 100 km was assured (also estimated with 1D Niblett-Bostick approximations).

Fig. 5 shows pseudo-section plots of apparent resistivity and phases for the LOW, KAP, LIM-SSO profiles giving an indication of the lateral variation of resistivity with period (i.e. depth). The apparent resistivity and phase pseudo-sections are shown for both the transverse magnetic (TM) and transverse electric (TE) modes. High resistivity values are represented by blue (cold) colours, whereas red (hot) colours indicate low resistivity values. Although these images are distorted due to the representation of distances along the abscissa versus log (period) instead of true depth, some major features can already be recognized. The LOW profile in particular reveals the resistive lithologies of the Kaapvaal craton, SMZ and Central Zones of the Limpopo belt. We will discuss key features that are resolved from 3D inversion modelling of these data in a Section 7.

6. 3D inversion modelling

The motivation for doing 3D instead of 2D inversion is that no assumption about the dimensionality of the data had to be made. Furthermore, modelling the four components of the impedance together with vertical transfer functions (Hz) enable us to define the nature of the structures that would otherwise not be resolved by applying 2D inversion modelling, thus we are able to gain added information about the resistivity distribution at depth.

In total, MT data from a total of 51 stations were modelled. The modular EM code of Egbert and Kelbert (2012) was used to generate 3D models. The apparent resistivity error floors were set to 10% for the diagonal and 15% for the off-diagonal elements of the impedance tensor (i.e., if the errors were less than 10% or 15% of the amplitude of the impedance, they were set to that level, if they were more, they were unchanged). The phase error floors were set to 5% and a constant absolute error floor in Hz was set at 0.01%. The size of the 3D grid was 79, 72 and 52 cells on the north, east and vertical downwards direction, respectively. The impedance elements (Z_{xx}, Z_{xy}, Z_{yy}, Z_{yx}) and H_z were modelled with the smoothing parameter tau (τ) set to 3 and using a 100 ohm-m half-space as an input model. The final model produced, converged to an average RMS of 3.61 (Fig. 9). In order to validate the features observed in the 3D model and test the resolution of nearby off-profile conductors, we conducted 3D inversion of each of the 2D profiles separately (3D/2D). Siripunvaraporn et al. (2005) demonstrated the advantages of modelling data this way using synthetic examples and, in a more recent study, Patro and Egbert (2011) applied similar technique to model data from the Deccan Volcanic Province. The principal advantage is that off-profile features are correctly located spatially, and are not artificially placed beneath the profile, as can be the case in 2D. In order to maintain consistency we use the same inversion parameters (i.e., four impedance elements Z_{xx}, Z_{xy}, Z_{yy}, Z_{yx} were modelled, using a tau value of 3 and 100 ohm-m half-spaces as input).

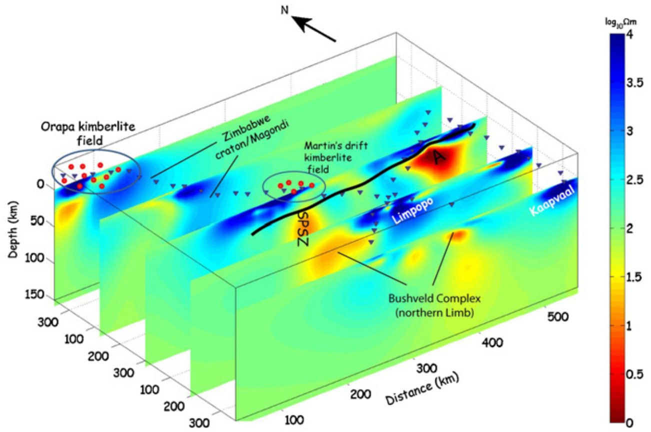


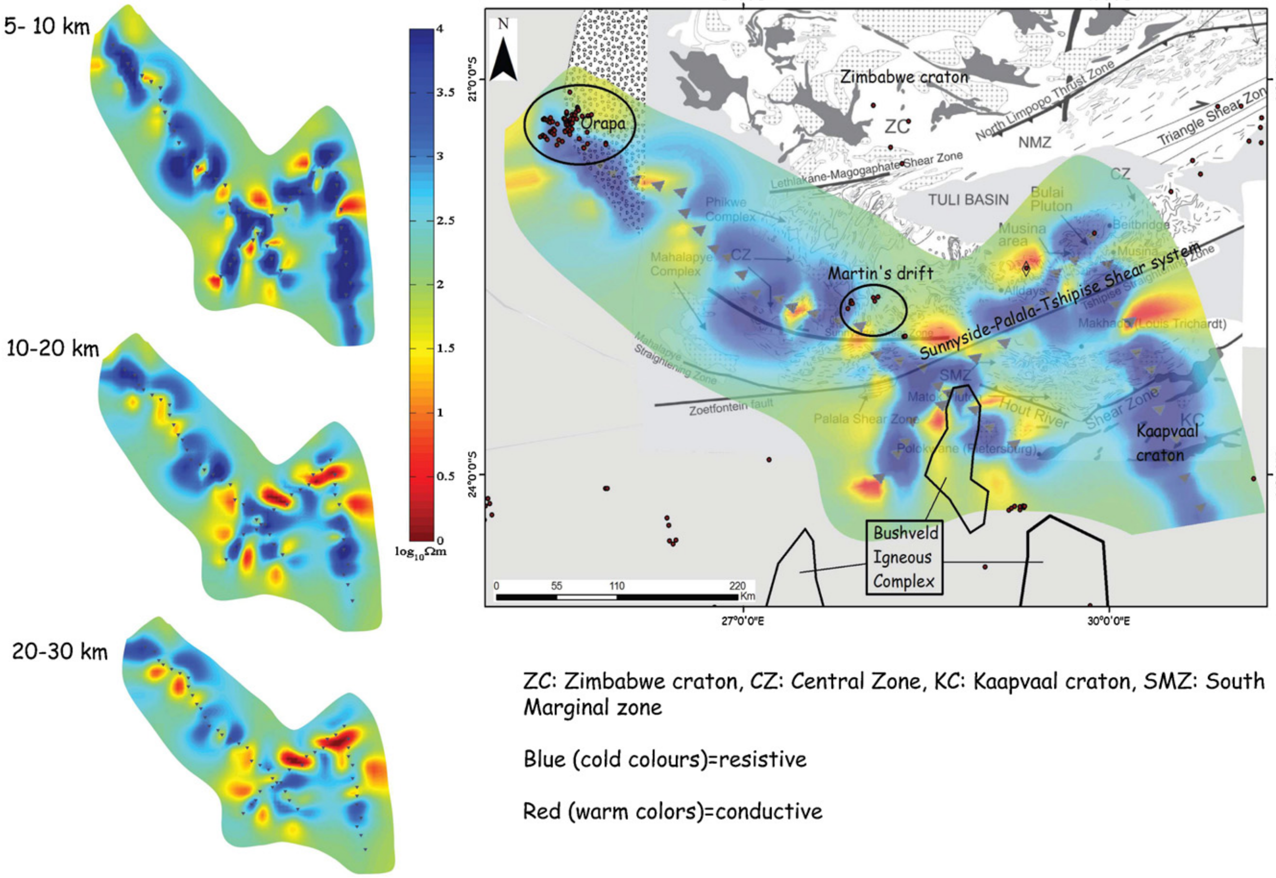
Fig. 6. 3D E-W perspective view showing the variation in resistivity laterally and depth across the Limpopo belt. The location of the Orapa and Martin's Drift kimberlite fields are projected. The dark solid line shows the approximate trace of the Palala-Tshipise-Sunnyside Shear system. SPSZ: Sunnyside-Palala Shear Zone. The conductive Feature A is explained more in the text.

7. Results

Our final 3D model (in perspective view) from inverting all data simultaneously is shown in Fig. 6, and horizontal depth slices through the volume are shown in Fig. 7. The blue triangles show the locations of the MT stations and the off-profile extent of the sections is limited to the corresponding depth-footprint. Highlighted on both figures are the SMZ, CZ, PTSZ, HRZ, Bushveld complex and Kaapvaal and Zimbabwe cratons. On comparing the 10 km depth slices to the locations of major shear zones, it is clear that the PTSZ corresponds to a major resistivity contrast down to lower crustal level. In order to obtain an indication of the geometry and extent of the structures at depth, 2D sections were extracted from the 3D model and are shown in Fig. 8. We will now highlight the major features resolved for each profile.

7.1. LOW profile

The crust in the southern part of the model is dominated by resistive features in the SMZ and the Kaapvaal craton. There is a significant resistivity break, up to 20 km in lateral distance, in the model at sites LOW003 and LOW004 that correlates spatially with the Soutpansberg basin (see Fig. 2). The 1.9 Ga elongated Soutpansberg Basin occurs within the SMZ on the southern side of the PTSZ and partly transcends the location of the PTSZ. The 3D/2D models suggest that the basin extends off-profile to the east and west confirming its E–W orientation. The location of the PTSZ is characterized by a conductive signature. Based on its lack of seismic response, Durrheim et al. (1992) and De Beer and Stettler (1992) interpreted the PTSZ to have a sub-vertical dip. The sites KAP074 and LOW001 are clustered around the 2.57 Ga Bulai pluton. Compositionally the pluton is charnokitic, made up of granites and granodiorites and as a result it appears as a resistive feature. Conductive Feature B beneath the SMZ occurs at 35 km depths and similar to the enigmatic feature imaged by De Beer et al. (1991) using DC resistivity and LOTEM methods. Moho



ZC: Zimbabwe craton, CZ: Central Zone, KC: Kaapvaal craton, SMZ: South Marginal zone

Blue (cold colours)=resistive

Red (warm colors)=conductive

Fig. 7. Crustal depth sections derived from 3D inversion model. Three crustal depth are shown. The 10 km section is overlain on the geological map of the Limpopo belt (after Kramers et al., 2011). The main features are highlighted, including the locations of the Orapa and Martin's drift kimberlite fields which are in close proximity to the LIM-SSO profile.

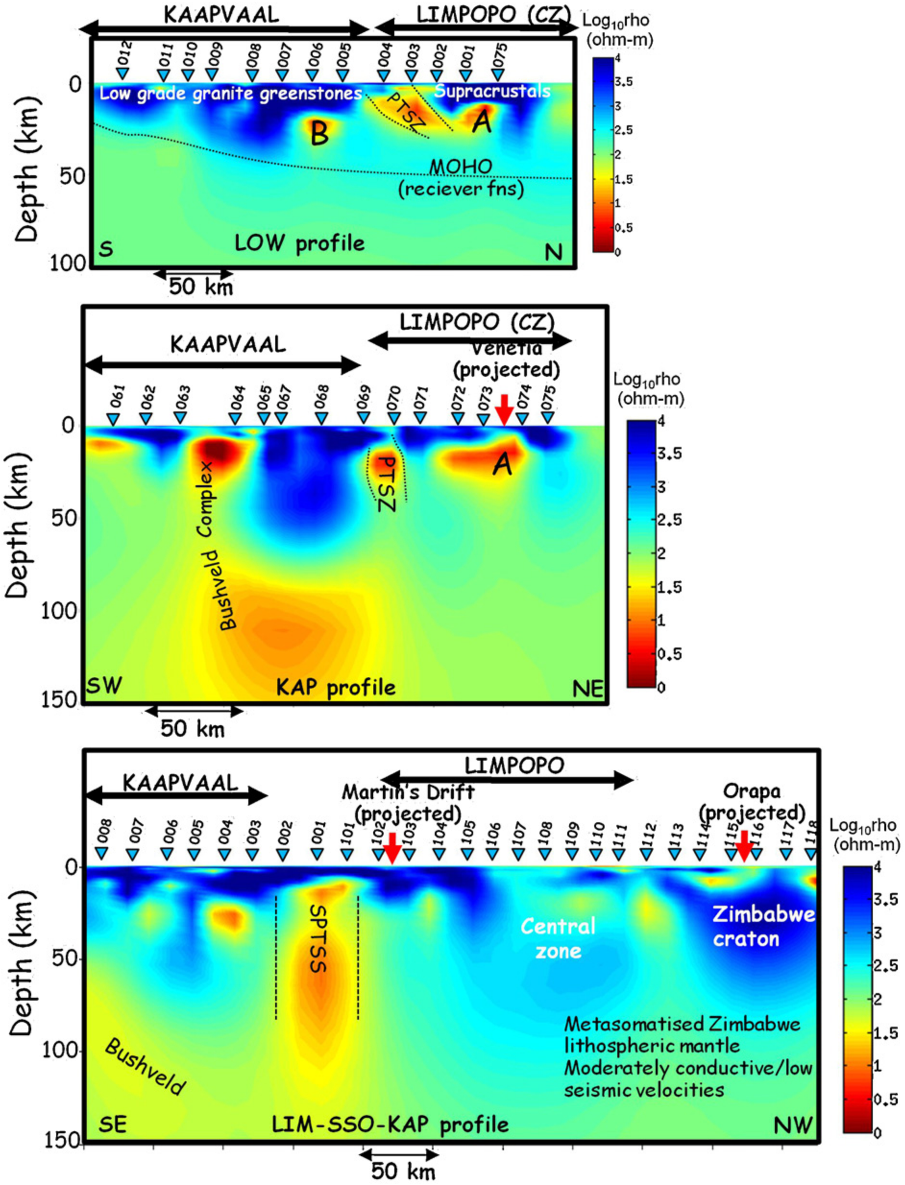


Fig. 8. LOW, KAP and LIM-SSO profiles derived from 3D inversion model. The Moho depth on the LOW profile was derived from the seismic receiver function study of Gore et al. (2010) (note the profile depth is 100 km). The location of Venetia kimberlite is projected on the KAP profile and is shown as red arrow. Similarly, the Orapa and Martin's Drift kimberlite clusters are shown on the LIM-SSO profile. Features A and B are referred to in the text. PTSZ: Palala-Tshipise Shear Zone, SPTSS: Sunnyside-Palala-Tshipise Shear System.

depth was estimated by Gore et al. (2010) from seismic receiver functions, and infers a relatively thicker crust beneath the CZ.

7.2. KAP profile

Similar features to the LOW profile are observed on the KAP profile. The resistive granite-greenstones lithologies of the Kaapvaal craton to the south are mapped in the upper crust. A significant conductivity break in crustal structure is observed with a conductive anomaly corresponding to the northern limb of the Bushveld complex. The PTSZ is situated along sites KAP068, KAP069 and KAP070 and is similarly evident as a conductive feature. The resistive upper-crustal Beitbridge Complex lithologies extend to a depth of 10 km and are underlain by conductive Feature A. Given the tectonic implications of the location and geometry of the PTSZ, several 3D forward models were generated in order to obtain a geometrical model that matches the observed resistivity responses. To this end, we tested models where a 15–20 km conductive zone (approximately 10 Ωm, akin to PTSZ) is embedded in a resistive media (10,000 Ωm, akin to the CZ and Kaapvaal). Various dip angles were tested and the model with the conductive zone having a sub-vertical dip returned almost similar responses to those of KAP068 and KAP069 (Fig. 8).

7.3. LIM-SSO profile

The SE part of this profile is characterized by resistive low grade lithologies related to the Kaapvaal craton and a conductive feature extending to depths of over 100 km attributed to the northern limb of the Bushveld Igneous Complex. The prior 2D isotropic and anisotropic models of Evans et al. (2011) mapped the Bushveld complex as a mantle conductive feature extending to depths in excess of 150 km. The composite SPTSS is positioned on the geological map between MT sites LIM002 and SSO101. This region corresponds to a vertically-dipping conductive feature on the MT

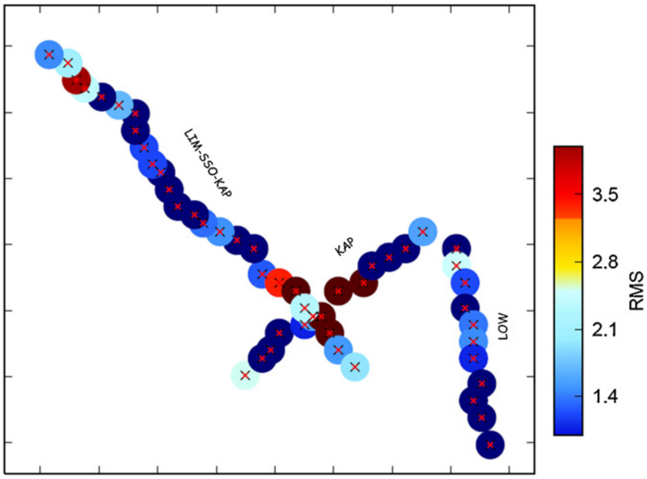


Fig. 9. RMS misfit between modelled data and resulting model plotted from each station.

model. The SPTSS is composed of gneisses and mylonites (McCourt and Vearncombe, 1992), therefore the observed elevated conductivities are interpreted to map the lateral and depth extent of the SPTSS. (McCourt and Vearncombe, 1992) mapped the SPTSS as a 40 km wide linear shear system that maps the boundary between the CZ and the SMZ. From the resistivity models we infer that the SPTSS has approximately 50 km lateral extent and that it represents a fundamental crustal suture zone. The locations of the Martin's drift and Orapa kimberlite clusters are projected on the model and the lithosphere beneath it is characteristically resistive and thick.

8. Interpretation

Several features are resolved from 3D inversion modelling: (1) the upper crustal resistive granite-greenstone of the Kaapvaal and SMZ, (2) the lower crustal conductors labelled A and B on the LOW profile and A on the KAP profile, (3) the conductive feature associated with the composite Sunnyside-Palala-Tshipise Shear System (SPTSS).

On the LOW and KAP profiles (in South Africa) this shear zone is labelled the Palala-Tshipise Shear Zone (PTSZ) and it extends westward (into NE Botswana) where is links with the Sunnyside Shear Zone (SSZ) and becomes a composite SPTSS. In the region of sites KAP071 to KAP075 (Fig. 2), the upper crust is composed of Beit Bridge Complex lithologies, which comprises quartzo-feldspathic rocks with a granitic bulk compositions (Klemd et al., 2003). These are evident in the LOW and KAP profiles (Fig. 8) as resistive features. Below these rocks, at 10 km depth, is a conductive (approximately 10 Ωm) Feature A. Seismic results (De Beer and Stettler, 1992; Durrheim et al., 1992) indicate a reflector at 10 km depth below the Beit Bridge complex, corresponding to the top of the conductive Feature A. Conductive middle to lower crust is observed globally (Jones et al., 1992; Hyndman et al., 1993) and in the Canadian Cordillera, for example, coincident reflective and conductive middle to lower crust is observed (Marquis et al., 1995), the causes of which are thought to be, principally, trapped saline pore-fluids for Phanerozoic regions or graphite for Precambrian regions. Pretorius (2003) combined xenolith-derived P–T mineral equilibria data with seismic information to derive the crustal and upper mantle lithology structure beneath the 520 Ma Venetia kimberlite cluster. This structure consists of supracrustal rocks down to 10 km depth, overlying mafic amphibolite rocks and restites (garnet-quartz rocks, granulite and eclogite).

According to Pretorius (2003), the amphibolites/restites were derived from partial melting of a remnant subducted Fe–O rich Archaean oceanic crust and gravitationally settled at deeper levels due to their higher densities (3.2–4.5 g/cm^3). However it is unlikely that Feature A corresponds to the amphibolite in that silicate rocks are usually resistive (Chave and Jones, 2012; Evans, 2012); therefore another conducting mineral phase must be present to account for the observed high conductivities. Conductive Feature A extends from 10 km depth to about 50 km, although, given the shielding effect of conductive features, the bottom depth is possibly overestimated. The top of the conductor overlain by a resistor is usually wellresolved with the MT method. There is a clear spatial correlation between the location of Feature A with high density rocks (Ranganai et al., 2002) and the interpreted thickened crust (Gore et al., 2010). In their review of age determinations in the Limpopo Complex, Kramers and Mouri (2011) have commented on the sharp age peak of the 2.0 Ga event in the Central Zone, and they and Kramers et al. (2011) have suggested that there may have been underplating by Bushveld complex related mafic magmas. They argued that this could also explain the gravity anomaly and poorly defined Moho in the region. Given the known conductive signature of the Bushveld complex, this could provide an alternative explanation of Feature A. It is thus worthwhile to investigate the possible causes and tectonic significance of this Feature A.

In the Earth's crust, fluids, interconnected sulphides/oxides, graphites, or high conducting metamorphic rocks are potential candidates for material that can give rise to observed elevated conductivities. The presence of fluids, particularly in Precambrian terranes, is diffcult to discern due to the complex evolution of metamorphic rocks. There is evidence of prograde and retrograde metamorphism, where rocks have undergone more than one deformation event in the Limpopo belt. For this reason, Barnes and Sawyer (1980) argued that it is unlikely that fluids will have remained stable in the crust since Archaean times, given their short residence times. Also argued by Yardley and Valley (1997), with interesting discussion by Wannamaker (2000) and response by Yardley and Valley (2000). Goldfarb et al. (1991) gives evidence for at least 70 Ma residence times of water.

Fluid inclusion studies in the Limpopo belt have focused in the Central Zone (Hisada and Miyano, 1996; Hisada et al., 2005; Tsunogae and van Reenen, 2007; Huizenga et al., 2011) and the South Marginal Zone (van Reenen and Hollister, 1988; van Reenen et al., 1994; van den Berg and Huizenga, 2001; Touret and Huizenga, 2001). In the SMZ, fluid inclusion studies (and high-temperature reaction texture) in granulites (Touret and Huizenga, 2001; van den Berg and Huizenga, 2001) indicate presence of brines and CO_2 rich-fluids during peak granulite facies metamorphism. van den Berg and Huizenga (2001) suggest that the brines represents remnants of preserved connate water. Studies by Huizenga et al. (2011) in the Central Zone confirmed the presence of CO2-rich-fluids in high-temperature Mg-rich garnets co-existing with brines, similar to the SMZ.

While presence of fluids in the CZ and SMZ is known, questions can be asked as to (1) which tectonic process introduced CO_2 in the crust, (2) how widespread are they, and, more importantly for electrical conductivity, (3) what is the nature of the inter-connectivity of the fluids? The fluid inclusion studies undertaken on material from the Limpopo belt are unable to estimate absolute fluid content and, as such, questions (2) and (3) are beyond the scope of this study, but we address here the first question and suggest a possible mantle source for CO_2, in a subduction setting.

Sm–Nd and Lu–Hf results suggest that the southern Zimbabwe craton was an active magmatic arc from the south to the southwest, characterized by subduction of oceanic lithosphere *ca.* 2.7–2.6 Ga (Bagai et al., 2002; Kampunzu et al., 2003; Zhai et al., 2006; Zeh et al., 2009; Kramers et al., 2011; Kramers and Zeh, 2011). Furthermore, the CZ, NMZ and Zimbabwe craton have similar elevated U/Pb ratios and U, Th concentrations (Kramers et al., 2001; Barton et al., 2006; Andreoli et al., 2011), implying that all three represent parts of the same plate where the CZ is the shelf, the NMZ the active margin and the Zimbabwe craton being the overriding plate (Kramers et al., 2011). These results contradict the Turkic-type model that requires separate terrane status for the CZ, NMZ and Zimbabwe craton. The presence of subducting carbonate-rich oceanic lithosphere would release CO_2 fluids into the lower crust and react with the mantle peridotite of shelf region (CZ), evidence of the latter process in the CZ stems from zircon Lu–Hf data on the charnockitic Bulai pluton sample which shows remelting of older crust (Zeh et al., 2007). However, the SM-Nd studies of Harris et al. (1987) suggested that the pluton might have been derived from anetexis of juvenile crust. In addition, petrographic study of 65 samples exposed in a 10 km^2 area, approximately 13 km west of site KAP074 (Fig. 2), indicated the presence of graphite and magnetite (Klemd et al., 2003). In the same study, CO_2-rich inclusions were found in garnet rims of metapelites.

It is therefore reasonable that graphite was precipitated from the CO_2-rich fluids. We therefore propose that the observed conductivity of Feature A is due to the presence of graphite and minor accessory minerals like magnetite, in the crust. The graphite reduction mechanism was invoked to explain the high conductivity signatures of the Fraser fault and also proposed for Yellowknife Fault. The conductive feature B is located in the lower crust. De Beer and Stettler (1992) mapped it, but provided no explanation at the time as to its possible cause and significance. In our models and those of De Beer et al. (1991) conductive Feature B coincides with the position of the HRSZ at depth; it is not known if Features A and B is are tectonically related. However given that two independent studies using different techniques have now confirmed its presence it is reasonable to infer that conductor B is a pervasive feature in the crust.

Conductivity in the Earth's lower crust has been observed from MT studies and debated for many decades (Edwards et al., 1981; Gregori and Lanzerotti, 1982; Jones et al., 1992; Touret and Marquis, 1994; Glover, 1996; Yang, 2011), but there, is yet, no consensus. van Reenen and Hollister (1988) suggested a subdivision of the SMZ into a northern granulite zone and a zone of retrograde hydration in the south, the former being the result of prograde melting reactions without involving fluids (Stevens, 1997) whereas the latter is associated with CO_2 and brine rich fluids sourced from devolitisation reactions (van den Berg and Huizenga, 2001). Feature B is located within this hydrated subzone but it is unlikely that fluids are responsible for the observed conductivities given their short residence times in the crust, therefore graphite (and minor conductive phases like FeO and sulphides) are suggested to be likely candidates for increased conductivities.

Central to the pop-up model proposed for the Limpopo development was the geophysical mapping studies by De Beer and Stettler (1992) and Durrheim et al. (1992), who imaged the northward and southward dip geometry of the HRZ and TSZ respectively. The PTSZ exhibited no seismic response in the study of Durrheim et al. (1992), leading to the suggestion that it was a vertically-dipping feature. While our results do not dispute the previous results of De Beer and Stettler (1992) and Durrheim et al. (1992), we map, for the first time, the PTSZ as a conductive feature and suggest that it in fact represent a fundamental suture between the Kaapvaal and Zimbabwe cratons. The PTSZ appears to be sub-vertical on the KAP and LIM-SSO profiles and dips slightly to the north on the LOW profile. On a mantle lithospheric scale, the PTSZ correlates with the discontinuity observed between the 300 km thick Kaapvaal/SMZ block and the 250 km thick Zimbabwe/NMZ/CZ block (Fouch, 2004). The PTSZ comprises conductive mylonite and ultramylonite rocks, similar to those found in some parts of the Bushveld complex (McCourt and Vearncombe, 1992).

The conductive region in the Bushveld crust is related to metallic sulphides and oxides widespread in the complex (the Bushveld complex is the largest resource for platinum group metals). The location of the Northern Limb of the BIC is on a junction between the Hout River Shear Zone and the Palala-Tshipise Straightening zone. These shear zones are both characterized by high conductivity signature. Kamber et al. (1995b) noted that high grade metamorphism at 2.0 Ga, which was a result of collision, was coeval with dextral transcurrent shear movement of the PTSZ, suggesting a transpressive collisional event. Given that the Soutpansberg basin is located in the SMZ just south of the PTSZ, it is possible that the deposition of volcano-clastic sediments in the trough developed on the extensional side of the transpressive shear zone system (post CZ uplift), an observation suggested by Kramers et al. (2011). This model is in contrast with the aulocogen model proposed by Jansen (1975), and is in agreement with a half-graben setting suggested by Bumby et al. (2002) and Tankard et al. (1982).

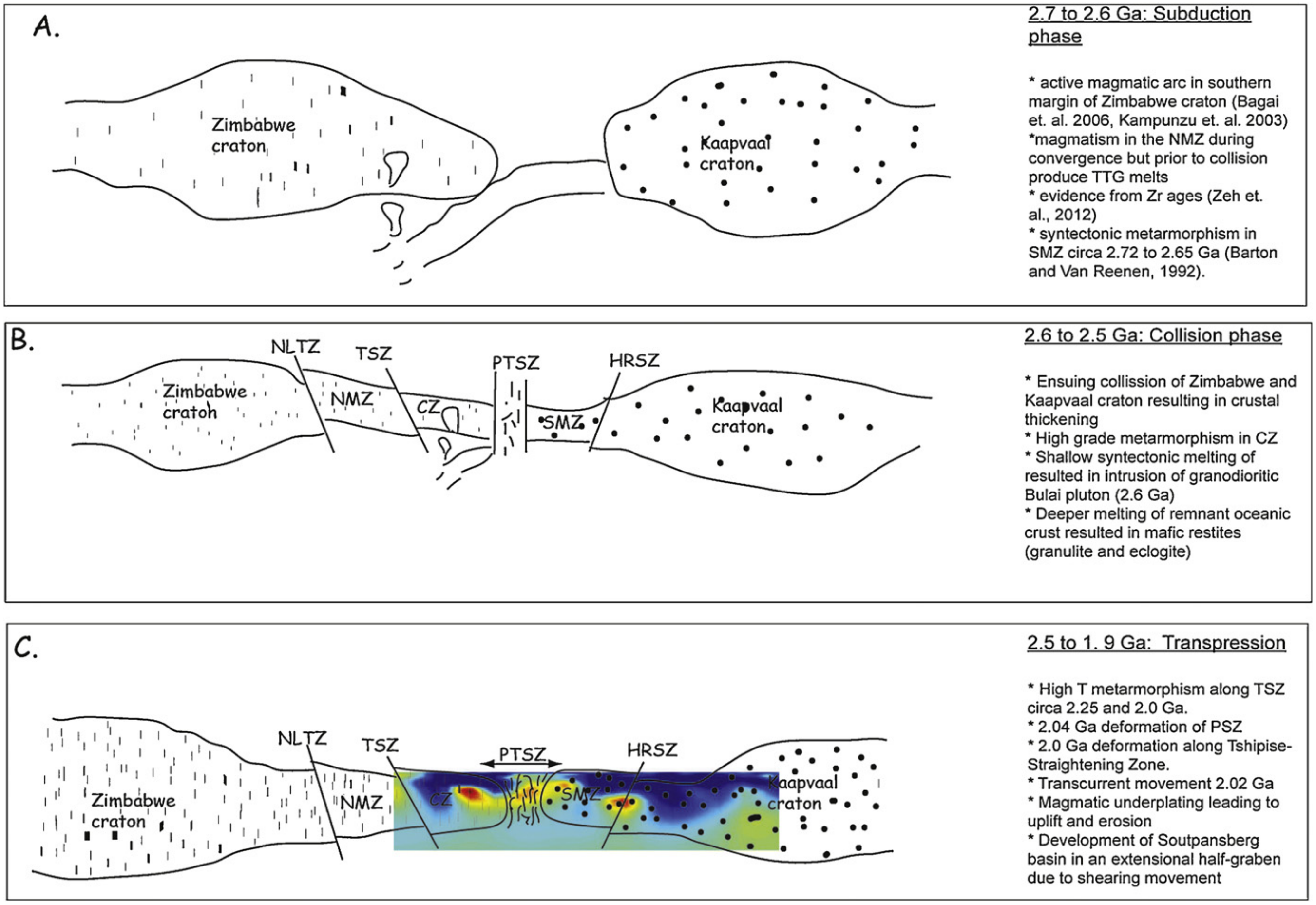
The geometry of inter-cratonic sutures have played a significant role in the evolution of the African tectonic landscape, by focussing ascending magmas and areas of localized rifting (Begg et al., 2009). The close spatial proximity of the PTSZ and the Venetia and Martin's drift kimberlite suggest that these shear zones could have possibly acted as conduits to ascending magma, leading to the emplacement of kimberlite volcanic material. The projection of the Orapa kimberlite on the LIM-SSO MT model suggests that it plots on the part of resistive thick lithosphere that is an extension of the Zimbabwe craton, as was suggested by Miensopust et al. (2011). The resistive lithosphere in this region on the LIM-SSO profile extends to 150 km depth. Seismic tomography maps, however, infer a seismically slow mantle beneath the Orapa kimberlite field (James et al., 2001) which was attributed to intrusion events related to mid-Proterozoic collision of the Okwa and Magondi belts (Shirey et al., 2002; Griffin et al., 2003) that significantly modified the-then Archaean lithosphere.

9. Towards a tectonic model

From the discussions above and combining the resistivity models and the recent metamorphic results, we propose a model for the evolution of the Limpopo belt (illustrated in Fig. 10) that involves three main tectonic processes, namely (1) *subduction* phase at 2.7–2.6 Ga, (2) *collision* phase at 2.6–2.5 Ga and (3) *transpression* phase at 2.2–1.9 Ga. In this model, the Neoarchaean collision of the Kaapvaal and Zimbabwe cratons, preceded by oceanic lithospheric subduction beneath the latter, is followed by Paleoproterozoic transcurrent/transpression along shear zones. The SPTSS represents a major suture zone between the Kaapvaal and Zimbabwe cratons.

The *Subduction* phase (Fig. 10A): petrological results suggest that the southern Zimbabwe craton, in effect the NMZ, was an active magmatic arc characterized by subduction of oceanic lithosphere *ca.* 2.7–2.6 Ga (Bagai et al., 2002; Kampunzu et al., 2003; Zeh et al., 2009). Magmatism was during convergence but prior to collision of the Kaapvaal craton with the Zimbabwe craton. At 2.7 Ga, however there is no evidence that the SMZ was an accretionary margin, but there is an indication that high grade metamorphism was prevalent (Kramers et al., 2011).

The *Collision* phase (Fig. 10B): the ensuing collision between the Kaapvaal and Zimbabwe cratons gave rise to the observed high-grade metamorphism in the CZ and resulted in thickened crust. Shallow syntectonic melting resulted in the intrusion of the granodiorites, such as the 2.6 Ga Bulai pluton, which shows signatures of old and juvenile crustal anetexis at this time. Deeper melting of remnant oceanic crust resulted in CO_2-rich fluids migrating into the crust and precipitating graphite. The collisional suture zone between the Kaapvaal and Zimbabwe cratons is located in the CZ as



2.7 to 2.6 Ga: Subduction phase

* active magmatic arc in southern margin of Zimbabwe craton (Bagai et. al. 2006, Kampunzu et. al. 2003)
*magmatism in the NMZ during convergence but prior to collision produce TTG melts
* evidence from Zr ages (Zeh et. al., 2012)
* syntectonic metarmorphism in SMZ circa 2.72 to 2.65 Ga (Barton and Van Reenen, 1992).

2.6 to 2.5 Ga: Collision phase

* Ensuing collission of Zimbabwe and Kaapvaal craton resulting in crustal thickening
* High grade metarmorphism in CZ
* Shallow syntectonic melting of resulted in intrusion of granodioritic Bulai pluton (2.6 Ga)
* Deeper melting of remnant oceanic crust resulted in mafic restites (granulite and eclogite)

2.5 to 1. 9 Ga: Transpression

* High T metarmorphism along TSZ circa 2.25 and 2.0 Ga.
* 2.04 Ga deformation of PSZ
* 2.0 Ga deformation along Tshipise-Straightening Zone.
* Transcurrent movement 2.02 Ga
* Magmatic underplating leading to uplift and erosion
* Development of Soutpansberg basin in an extensional half-graben due to shearing movement

Fig. 10. Schematic illustration of the tectonic evolution of the Limpopo belt from 2.7 Ga derived from the combination of the MT results and the metamorphic studies (not to scale). The resistivity image of the LOW profile is projected in the background.

the composite conductive mylonitic feature: the Sunnyside-Palala-Tshipise-Shear zone system.

The *Transpression* phase (Fig. 10C): the PTSZ was reactivated at 2.04 Ga with transcurrent movement at 2.02 Ga (Holzer, 1998) as a result of the eastward movement of the Zimbabwe craton (coevally with Magondi belt) relative to Kaapvaal craton. Crustal exhumation and uplift, a result of Bushveld age magmatic underplating *ca.* 2.03–1.95 Ga, of the CZ was followed by erosion and the subsequent deposition of the Soutpansberg basin at 1.9 Ga in an extensional graben-like setting (Kamber et al., 1995a,b; Schaller et al., 1999).

10. Conclusions

The Limpopo belt has been given many geological tags (i.e., mobile or orogenic belt, complex, terrane), in essence to separate it from the Kaapvaal and Zimbabwe cratons. From the discussions above, the Limpopo belt is perhaps best viewed as a plate tectonic manifestation of polytectonic structural and metamorphic activities that resulted from the horizontal collision between the Kaapvaal and Zimbabwe cratons. Geophysical studies presented in this work, supported strongly by metamorphic results, appeal to an evolutionary path involving the collision between the Kaapvaal and Zimbabwe cratons, with the Palala Shear Zone representing a fundamental suture. To this end the Zimbabwe craton represent an overriding plate margin with the NMZ being the active margin and the CZ the leading shelf. Thus, evolutionary models proposing the separation of the Limpopo belt into separate terranes are not required. Questions still remain however regarding the timing of the metamorphic events (particularly in the SMZ), orientation and rate of plate of movement in the Archaean. However the presented model, based on all available data including the new MT data, suggests that horizontal collisional movement is the most plausible of all the models that have been presented thus far.

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