MAGNETOTELLURIC STUDIES ACROSS THE DAMARA OROGEN AND SOUTHERN CONGO CRATON

TSHEPO DAVID KHOZA

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Doctor of Philosophy

University of the Witwatersrand
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and
Dublin Institute for Advanced Studies
School of Cosmic Physics
Geophysics Section

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DECLARATION

I hereby declare that this thesis is entirely my own work, except where acknowledgement is made in the text. It is being submitted for the degree of Doctor of Philosophy at the University of the Witwatersrand, Johannesburg, South Africa. It has not been submitted before for any degree or examination at any other university.

I agree that the Library may lend or copy this thesis upon request.

Johannesburg, February 2016

Tshepo David Khoza
Archean cratons, and the Proterozoic orogenic belts on their flanks, form an integral part of the Southern Africa tectonic landscape. Of these, virtually nothing is known of the position and thickness of the southern boundary of the composite Congo craton and the Neoproterozoic Pan African orogenic belt due to thick sedimentary cover. In this work I present the first lithospheric-scale geophysical study of that cryptic boundary and define its geometry at depth. The results are derived from two-dimensional (2D) and three-dimensional (3D) inversion of magnetotelluric data acquired along four semi-parallel profiles crossing the Kalahari craton across the Damara-Ghanzi-Chobe belts (DGC) and extending into the Congo craton. Two-dimensional and three-dimensional electrical resistivity models show significant lateral variation in the crust and upper mantle across strike from the younger DGC orogen to the older adjacent cratons. The Damara belt lithosphere is found to be more conductive and significantly thinner than that of the adjacent Congo craton. The Congo craton is characterized by very thick (to depths of 250 km) and resistive (i.e. cold) lithosphere. Resistive upper crustal features are interpreted as caused by igneous intrusions emplaced during Pan-African magmatism. Graphite-bearing calcite marbles and sulfides are widespread in the Damara belt and account for the high crustal conductivity in the Central Zone. The resistivity models provide new constraints on the southern extent of the greater Congo craton, and suggest that the current boundary drawn on geological maps needs revision and that the craton should be extended further south.

The storage possibilities for the Karoo Basins were found to be poor because of the very low porosity and permeability of the sandstones, the presence of extensive dolerite sills and dykes. The obvious limitation of the above study is the large spacings between the MT stations (> 10km). This is particularly more limiting in resolving the horizontal layers in the Karoo basin. However the 1D models provide layered Earth models that are consistent with the known geology. The resistivity values from the 1D models allowed porosity of the Ecca and Beaufort group lithologies to be calculated. It is inferred that the porosity values are in the range 5-15 % in the region below the profile. This value is considered too low for CO$_2$ storage as the average porosity of rock used for CO$_2$ is generally more than 10 to 12 percent of the total rock unit volume.
To my wonderful family
My humanity is bound up in yours, for we can only be human together.
— Desmond Tutu

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My time at the Dublin Institute for Advanced Studies (DIAS) has allowed me to meet and interact with a lot of people and scientists from different backgrounds, many of whom I consider friends.

I would like to thank my supervisors Professor Alan Jones (DIAS) and Dr. Sue Webb (University of the Witwatersrand), both of whom provided excellent guidance, support and inspiration during my studies. Their continued encouragement and motivation has allowed me to complete this research and without them I would not have made it. The visit by Prof. Jones to BHP Billiton encouraged me to take up a PhD studentship under his supervision and this proved to be turning point in my research career, for this I would like to extend my sincere thanks to him.

I would especially like to thank Dr. Mark Muller who provided incredible support and guidance throughout my time at DIAS. I would like to thank BHP-Billiton, particularly Mr. Wayne Pettit, for funding my MSc research and providing the opportunity to continue my studies. Without this support, my work would not have been possible. The South African Centre for Carbon Capture and Storage (SACCCS) is thanked for providing additional funds for enabling me to research a hugely interesting and vital topic of CO\textsubscript{2} capture and storage.

The SAMTEX project has been a pleasure and privilege to work on not least because of the support and enthusiasm shown by the many co-workers and helpers during the field work. I sincerely like to thank all of them for having such a great time in southern Africa, particularly I like to thank Colin Hogg, Marion Miensopust and Pieter Share.

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I would like to thank all my friends for their support and their persuasion to relax every now and then. My family continue to be a major source of encouragement and I thank them sincerely for their continual faith, love and support.

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Some of the work conducted as part of the SAMTEX project has been published in major journals and presented in some international conferences as per below:

PEER-REVIEWED PAPERS


vii
ORAL PRESENTATIONS


• A. G. Jones, R. L. Evans, M. Muller, **Khoza, T.D.,** *Lithospheric geometries revealed through electromagnetic imaging: SAMTEX (Southern Africa MagnetoTelluric Experiment) observations and results Invited,* American Geophysical Union Fall Meeting 2011, 5th-9th December 2011, San Francisco, California, USA.


POSTER PRESENTATIONS

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2.13 Static shift correction ........................................ 57
  2.13.1 Curve fitting methods ................................... 57
  2.13.2 Statistical methods ...................................... 59
  2.13.3 Other dimensionality and directionality tools .......... 60
2.14 Magnetotelluric data modelling and inversion ............ 60
  2.14.1 Forward modelling ...................................... 60
  2.14.2 1D and 2D Inversion modelling ....................... 61
  2.14.3 3D Inversion ........................................... 62
3 SOUTHERN AFRICAN MAGNETOTELLURIC EXPERIMENT 64
3.1 SAMTEX ....................................................... 64
3.2 Summary of SAMTEX results .................................. 66
  3.2.1 Southern African diamond prospectivity ................. 66
  3.2.2 Kaapvaal craton .......................................... 68
  3.2.3 Rehoboth Terrane ........................................ 71
  3.2.4 Zimbabwe craton, Magondi Belt ......................... 73
  3.2.5 Limpopo belt ............................................ 74
3.3 Synthesis of the SAMTEX results ............................. 74
4 GEOLOGY, TECTONICS AND GEOPHYSICS OF SOUTHERN AFRICA 76
4.1 Southern African geology ...................................... 76
  4.1.1 Kalahari craton .......................................... 76
  4.1.2 Damara orogen .......................................... 80
  4.1.3 Congo craton ............................................ 87
4.2 Tectonic models proposed for Damara belt evolution ........ 88
  4.2.1 Ensialic or modified aulocogen model ................. 88
  4.2.2 Subduction model ...................................... 93
4.3 Geological and geophysical constraints on proposed models 95
  4.3.1 Paleomagnetism ......................................... 95
  4.3.2 Thermal profiles ...................................... 96
4.4 Previous geophysical studies ................................. 96
  4.4.1 Resistivity studies in Damara belt ..................... 96
  4.4.2 Seismic, magnetic and thermal results .................. 99
ii INTERPRETATION OF THE 2D AND 3D MODELS 105
5 MT DATA IMAGING AND ANALYSIS 106
5.1 Data imaging .................................................. 106
  5.1.1 Pseudosections ......................................... 108
  5.1.2 Niblett-Bostick depth maps ............................. 111
  5.1.3 Phase tensor maps ..................................... 114
5.2 Strike angle and decomposition analysis ...................... 118
<table>
<thead>
<tr>
<th>Chapter</th>
<th>Section</th>
<th>Title</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>6.1</td>
<td>2D Isotropic and Anisotropic inversion</td>
</tr>
<tr>
<td></td>
<td>6.1.1</td>
<td>2D Isotropic inversion</td>
</tr>
<tr>
<td></td>
<td>6.1.2</td>
<td>The ETO-KIM profile</td>
</tr>
<tr>
<td></td>
<td>6.1.3</td>
<td>The DMB profile</td>
</tr>
<tr>
<td></td>
<td>6.1.4</td>
<td>The NEN profile</td>
</tr>
<tr>
<td></td>
<td>6.1.5</td>
<td>The OKA-WIN profile</td>
</tr>
<tr>
<td>6</td>
<td>6.2</td>
<td>2D Anisotropic inversion</td>
</tr>
<tr>
<td>6</td>
<td>6.3</td>
<td>Robust features of the 2D models</td>
</tr>
<tr>
<td>6</td>
<td>6.4</td>
<td>Interpretation of 2D resistivity models</td>
</tr>
<tr>
<td>7</td>
<td>7.1</td>
<td>3D MT data</td>
</tr>
<tr>
<td></td>
<td>7.2</td>
<td>Inversion parameters</td>
</tr>
<tr>
<td></td>
<td>7.3</td>
<td>3D Inversion results</td>
</tr>
<tr>
<td></td>
<td>7.3.1</td>
<td>Damara belt conductor inter-connection</td>
</tr>
<tr>
<td></td>
<td>7.4</td>
<td>3D Inversion of 2D profiles</td>
</tr>
<tr>
<td></td>
<td>7.4.1</td>
<td>3D Inversion of ETO-KIM profile</td>
</tr>
<tr>
<td></td>
<td>7.4.2</td>
<td>3D inversion of DMB profile</td>
</tr>
<tr>
<td></td>
<td>7.4.3</td>
<td>3D Inversion of NEN profile</td>
</tr>
<tr>
<td></td>
<td>7.4.4</td>
<td>3D inversion of OKA-WIN profile</td>
</tr>
<tr>
<td>7</td>
<td>7.5</td>
<td>Interpretation of 3D inversion results</td>
</tr>
<tr>
<td></td>
<td>7.5.1</td>
<td>Mantle lithosphere</td>
</tr>
<tr>
<td></td>
<td>7.6</td>
<td>Correlation of Resistivity and Seismic Velocity</td>
</tr>
<tr>
<td>8</td>
<td>8.1</td>
<td>The origin of the Damara belt crustal conductor</td>
</tr>
<tr>
<td></td>
<td>8.2</td>
<td>Magnetotelluric evidence of thick Congo craton lithosphere</td>
</tr>
<tr>
<td></td>
<td>8.3</td>
<td>Damara orogen perspective on the evolution of Gondwana</td>
</tr>
<tr>
<td></td>
<td>8.4</td>
<td>Global Context: Comparison to Similar Tectonic Environments</td>
</tr>
<tr>
<td>8</td>
<td>8.5</td>
<td>Is there incipient rifting in the Kalahari?</td>
</tr>
<tr>
<td>9</td>
<td>9.1</td>
<td>What is carbon capture and storage?</td>
</tr>
<tr>
<td></td>
<td>9.2</td>
<td>MT application in carbon capture and storage</td>
</tr>
<tr>
<td>10</td>
<td>10.1</td>
<td>Carbon capture and Storage in South Africa</td>
</tr>
<tr>
<td></td>
<td>10.1.1</td>
<td>The Karoo Basin</td>
</tr>
<tr>
<td></td>
<td>10.1.2</td>
<td>Identification and characterisation of storage sites in the Karoo basin using MT data</td>
</tr>
<tr>
<td>Figure</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>--------</td>
<td>-----------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>1.1</td>
<td>Map of bedrock geology of Africa showing major tectonic blocks, modified from <em>Begg et al. (2009)</em></td>
<td>4</td>
</tr>
<tr>
<td>2.1</td>
<td>Induction vectors (real component, Parkinson convention) plotted on a profile across geoelectric strike for a 2-D Earth model. White triangles denote MT stations and black arrows the induction vectors. The circular rings show magnetic field lines at a point in time when current is into the page.</td>
<td>13</td>
</tr>
<tr>
<td>2.2</td>
<td>Resistivity ranges for common rock types, modified from <em>Martí (2006), Palacky (1988) and Miensopust (2011)</em></td>
<td>15</td>
</tr>
<tr>
<td>2.3</td>
<td>Resistivity model of the continental crust with a resistive upper crust underlain by a two-layer lower crust. (modified after <em>Jones (1987) and Schwarz (1990)</em>).</td>
<td>18</td>
</tr>
<tr>
<td>2.4</td>
<td>Sea floor resistivity as a function of depth (after <em>Constable (1990)</em>)</td>
<td>19</td>
</tr>
<tr>
<td>2.5</td>
<td>Electrical conductivity versus temperature for the most important mantle phases (dry and hydrous olivine, molten silicates, and molten carbonates) (after <em>Gaillard et al. (2008)</em> and references therein).</td>
<td>20</td>
</tr>
<tr>
<td>2.6</td>
<td>Mineralogical model of pyrolite mantle. The thicknesses of the D” layer varies according to the locality. The Mg-perovskite shows the (Fe, Al)-bearing composition (<em>Irifune and Ringwood, 1987</em>). Candidates for the lighter elements are: O, S, Si, C and H (after <em>Ono (2008)</em>)</td>
<td>21</td>
</tr>
</tbody>
</table>
Figure 2.7 Electrical resistivity profiles and the estimated water content in the mantle transition zone. The orange and bluish areas represent geophysically observed conductivity profiles in the Pacific from Kuvshinov et al. (2005) and the continental mantle from Olsen (1998), Tarits et al. (2004) and Neal et al. (2000), respectively. The thick solid line represents the electrical conductivity of olivine, wadsleyite and ringwoodite without water. Dashed lines indicate the electrical conductivity of hydrous olivine, wadsleyite and ringwoodite as a function of water content (red: 1.0 wt%; green: 0.5 wt%; blue: 0.1 wt%). Light green solid line denotes the previous experimental result of Xu et al. (2000). The electrical conductivity of hydrous olivine was estimated from the average of three crystallographic axes. In the olivine stability field, thick and thin lines indicate the electrical conductivity estimated from different conditions of oxygen fugacity, with MoMoO$_2$ and NiNiO buffers, respectively (Yoshino et al., 2008).

Figure 2.8 Comparison of global conductivity models (labelled according to their bibliographic references) as derived from geo-electric experiments (after Campbell et al. (1998)). The result for the Australian hemisphere is labeled CBCW (solid squares); CS 1988 (open squares) is the Australian hemisphere result of Campbell and Schiffmacher (1988). The various models are labelled according to their bibliographic references.

Figure 2.9 Summary of inversion results for, respectively, the four-layer model with fixed interface depths (left) 50 km, 440 km and 670 km, the three-layer model with variable mantle interface depth (middle) and the three-layer model with all fixed interfaces (right). Solid lines show the best fitting models, grey shadings show models with the misfit $\chi^2$ within 0.2 per cent from the minimum for each particular parametrization (from Velinsky et al. (2006)).

Figure 2.10 3D body of conductivity $\sigma_2$ embedded in a homogeneous half-space of conductivity $\sigma_1$, illustrating the scale dependence of dimensionality of the Earth. Galvanic distortion will persist at long periods depending on the skin-depth and dimensions of the 3D body. Modified from Simpson and Bahr (2005).

Figure 2.11 N layered half space model
A 2D model showing areas of differing resistivities separated by a boundary striking in x direction. Due to Ohm’s law and current being preserved the $E_y$ field is discontinuous across the boundary. The EM field is decoupled into 2 components with $E$ field parallel to strike called the TE mode and $B$ field parallel to strike called the TM mode.

Example representations of anisotropy in earth materials. Anisotropy often is represented in terms of minute alternating planes of different conductivity with strike (sometimes also called azimuth) and dip which may be non-parallel to geometric strike. Figure from Heise and Pous (2001) and Weidelt (1999).

A galvanic distortion model (after Utada and Munekane (2000)) where the Earth is envisaged to have regional 3D structures at depth and the surface contains a layer comprising 3D inhomogeneities.

Galvanic phenomena. The secondary fields $E_s$ adds to the primary fields $E_p$ for a conductive inclusion and subtracts from the $E_p$ for a resistive inclusion (after Jiracek (1990)).

An example of current channelling phenomena over a highly conductive 1 ohm-m thin dyke embedded in a homogeneous 1000 ohm-m resistive half-space.

Example of current channelling from an MT site in the SAMTEX database.


Two sites from the SAMTEX database displaying effects of static shift.

Continental crustal model illustrating the effect of applying distortion.

Structurally induced static shift model. Forward responses on all models the first and last models shows apparent resistivities of both modes converging to 100Ωm, while for the centre model one of the modes is shifted upwards as a result of the structure.

Topographically induced static shift model. Modelled apparent resistivity and phase curves across a topographic step, illustrating the effects of topography on the responses.
Figure 2.23 BBMT and LMT instruments used for most of the phases of the SAMTEX project. The receivers boxes, MTU 5 s and MTU 5As, are able to record data in the MT data range (0.001 s - 50 000 s) and the AMT-MT range (0.0001 s - 50 000 s) respectively. The MTU 5 s/MTU 5s record up to two electric and three magnetic channels simultaneously. The BBMT instruments are manufactured by Phoenix Geophysics (Canada), while the LMT are manufactured by LVIV center (Ukraine).

Figure 2.24 A typical field set-up for recording MT data. The recording unit, placed at the centre of the survey site, is powered by 12V batteries and placed in a protective case metal box. Four electrodes orientated N-S and E-W are buried 45-50 m away from the ground electrode to record the variations in electric field. The magnetic field components are recorded by two N-S and E-W oriented horizontal coils and if possible one vertical broad-band coil. The GPS antenna provides the MTU 5A with the site geographic coordinates and more importantly with a continuous time signal. The photos are from the phase four SAMTEX field work.

Figure 2.25 Example of time series recorded over a 24hr period at a site KXT003, showing the electric (E_x, E_y) and magnetic (B_x, B_y) field responses.

Figure 2.26 Example of splitting and windowing a time series segment.

Figure 2.27 An example of windowed segment after applying Discrete Fourier Transform and the resulting power spectra.

Figure 2.28 Graphical representation of the phase tensor. The lengths of the ellipse axes, which represent the principle axes of the tensor, are proportional to the principle values of the tensor: \( \Phi_{\text{max}} \) and \( \Phi_{\text{min}} \). If the phase tensor is non-symmetric, a third coordinate invariant is needed to characterize the tensor: the skew angle \( \beta \). The direction of the major axis of the ellipse is given by \( \alpha - \beta \) and defines the relationship between the tensor and the observational reference frame \((x_1, x_2)\). Redrawn from (Caldwell et al., 2004).

Figure 2.29 An example of curve fitting technique using DC electric method, after Stephen et al. (2003). There is an observable DC shift in the YX mode as it is not coincident with the DC results at higher frequencies.
Figure 2.30 Theoretical long period correction of distorted MT curve $\rho_a$, shifted to coincide with the global sounding curve (h-line), modified from Mikhlin (1984).

Figure 3.1 SAMTEX stations.

Figure 3.2 An image of the resistivity at 200 km. Also shown on the figure are kimberlite locations; red means known to be diamondiferous, green means known to be non-diamondiferous, and white means not defined or unknown, after Jones et al. (2009a).

Figure 3.3 Electrical more conductive directions (in red), scaled by maximum phase difference, for the lithospheric upper mantle, and the shear-wave splitting results (both high and low quality results plotted in green, but sites with no detectable splitting omitted). ZC: Zimbabwe craton, KC: Kaapvaal craton. LB: Limpopo belt. NN: Namaqua-Natal mobile belt. CFB: Cape Fold belt. KB: Kheis and Proterozoic fold and thrust belt. Blue line: N-S trending Colesberg Magnetic Lineament (CBL). Purple lines: E-W trending Thabazimbi Murchison Lineament (TML), as plotted by Silver et al. (2001).

Figure 3.4 A proposed model for the origin and structure of anisotropy in southern Africa. Taken from Hamilton (2008).

Figure 3.5 (top and middle) Anisotropic models for the Kaapvaal data. Conductivity in the direction perpendicular to the profile ($R_{xx}$) and parallel to the profile ($R_{yy}$) are shown. (bottom) The levels of anisotropy calculated as the difference in $\log_{10}(\text{resistivity})$ between the $R_{yy}$ and $R_{xx}$ models, after Evans et al. (2011).
Electrical resistivity models for profile KIM-NAM (after Muller et al. (2009) derived from 2-D smooth inversion of decomposed MT station responses for (a.) 25° E of N strike azimuth and (b.) 45° E of N azimuth. (c.) Estimates of the depth of penetration achieved at each individual station location for both the TE (red) and TM (blue) modes, and overlaid on the 25° inversion model. The models are entirely unconstrained (and should be ignored) in areas beneath the maximum depth of penetration. (d.) 2-D inversion RMS misfit error at each station. The surface extent of the geological terranes is shown in (a.) and abbreviations used as follows: Ghanzi-Chobe/Damara (DMB), Rehoboth (RBT), Western Kimberley Block (KBW), Eastern Kimberley Block (KBE) and Witwatersrand Block (WB). Black dashed lines in (a.) and (b.) indicate interpreted depth to base of lithosphere, where well constrained, and red diamonds indicate the depth to the base of the chemically depleted lithosphere as defined by Cr/Ca-in-pyrope barometry from kimberlitic concentrates.

The 2D smooth inversion model (vertical exaggeration equals 1.0) in relation to the known surface extent of geological terranes. The arrows above the image of the resistivity structure show the crustal extents of the Limpopo Belt, Zimbabwe craton, Magondi Mobile Belt, and Ghanzi-Chobe Belt (GCB) with respect to MT sites of the ZIM line, adapted from the regional scale geological terrane boundaries based on potential field data (Webb, 2009). The extent of the Okavango Dike Swarm (ODS), known from magnetic data, is indicated, as well as an estimated extent of the brine aquifer related to the Makgadikgadi salt pan complex. The dominant resistivity features related to the main geological terranes are labeled, and the question mark indicates the area of missing data coverage. Two dominant middle to lower crustal conductors are also apparent (compare with inversion results from the northern crustal part of the profile).
Figure 4.7 Summary time-space diagram for the Damara Orogen based on $^{40}\text{Ar}/^{39}\text{Ar}$ data from Gray et al. (2006). The plot encapsulates the complexity of magmatic, metamorphic, deformational and cooling patterns across the orogen as part of Pan-African orogenesis. WKZ: Western Kaoko Zone; CKZ: Central Kaoko Zone; EKZ: Eastern Kaoko Zone; TPMZ: Three Palms Mylonite Zone; VMZ: Village Mylonite Zone; PMZ: Purros Mylonite Zone; AMZ: Ahub Mylonite Zone; ST: Sesfontein Thrust; GMZ: Guantegab Mylonite Zone; OmL: Omaruru Lineament (Shear Zone); OkL: Okahandja Lineament (Shear Zone).

Figure 4.8 Major and minor fault structures in the Damara belt, mapped geophysically and through field mapping, overlain on the regional aeromagnetic map of Namibia. AL: Autseib Lineament, OML: Omaruru Lineament, OKL: Okahanja Lineament, MA=Matchless Amphibolite, KI=Kamanjab Inlier (map modified from Corner (2008), Singletary et al. (2003) and Begg et al. (2009)).

Figure 4.9 Schematic cross-sections showing evolution of a typical aulacogen (adapted from Hoffmann, Dewey and Burke (1974). Stages (1) and (2) incorporate features observed in the Damara orogen. After stage (2) the evolution of the Damara orogen deviates from this model.

Figure 4.10 (A) Schematic diagram showing the suggested crust and mantle structure of the graben stage of the Damara geosyncline (Martin and Porada, 1977a). (B) The orogenic stage of the Damara belt, showing (B) gravitational instability of the upper mantle which causes detachment and sinking of dense lithospheric layer with concomitant ascent of the diapir. (C) Further sinking of dense slabs causes folding of the weakened crust; the heat of the growing diapir causes high-grade metamorphism and anatexis, as well as gravitational instability and granitic plutonism within the crust.

Figure 4.11 Tectonic evolution of the Damara belt, Congo and Kalahari craton, highlighting a full Wilson Cycle processes (modified from Gray et al. (2007)).
A. Schematic APWP for Laurentia and Baltica (grey swath), and Congo and Sao Francisco paleopoles (dots), plotted in a Laurentian reference frame. Congo’s APWP has been rotated with respect to Laurentia according to the Euler Pole: \(150^0, 7^0, 185^0\) counter clockwise. B. Corresponding reconstruction of Rodinia according to a 1010 Ma Laurentian pole, with the Congo craton (white) rotated as above. Also shown is Dalziel \(1991\) Congo reconstruction (dark grey), based on geologic observations. C. Schematic combined Laurentia-Baltica APWP (grey swath), with the seven Kalahari paleopoles (dots), for the 1100-950 Ma time interval, rotated into coincidence and plotted in a Laurentian reference frame. D. Corresponding reconstruction of the Kalahari craton with respect to the rest of Rodinia with paleolatitudes according to a 1010 Ma Laurentian pole. E. Proposed Rodinia reconstruction with all cratons rotated with respect to a 1010 Ma Laurentian paleopole according to the Euler poles. Grenvillian orogenic belts highlighted in black stipple after. AM=Amazonia craton; A=Australia craton; BA=Baltica-Fennoscandia.; C=Congo craton; CMG=Coates Land-Maudheim-Grunehogna Province; E=Ellsworth-Whitmore Mountain Block; EA=East Antarctica; G=Greenland; I=India; K=Kalahari craton; M= Madagascar; RP-Rio de la Plata craton; S=Siberia craton; SF=Sao Francisco craton; WA=West Africa craton. (after \(Weil et al., 1998\))

Pressure-Temperature profile for the Damara mantle derived from xenoliths samples (modified from \(Whitehead et al., 2002\)).

The map showing array in relation to tectonic features. Fourier transform amplitudes and phases at a period of 28 minutes. The ellipse represents horizontal field polarization (\(de Beer et al., 1975\)).

Two-dimensional resistivity models of the crust derived from inversion of (a) TM, (b) TE, and (c) joint TM+TE mode magnetotelluric data. The main features of the resistivity models are the generally high conductivity of the mid-crust in the central part of the profile and the narrow, subvertical conductivity anomalies are attributed to basement shear zones. AF: Autseib Fault, WF/OL: Waterberg Fault/Omaruru Lineament. Study from (\(Ritter et al., 2003a\)).
Figure 4.16 Estimates of lithospheric thickness based on the tomographic models, calculated using the empirical parametrization of Priestley and McKenzie (2006). The Damara belt ranges between 160 to 180 km and it is over 200 km for the Congo and Kalahari cratons.

Figure 4.17 Stacked seismic receiver functions (SRF) at two stations located in the Damara belt. The black line shows the mean stack while the gray shaded areas indicate the 2σ bootstrap error bounds. Major converted phases, LAB and Moho, are labeled. The black arrows indicate LAB depths reported by Kumar et al. (2007).

Figure 4.18 Regional magnetic map of Southern Africa, showing the SAMTEX stations (black stations are the focus of this thesis, white stations are the other SAMTEX stations). This data is courtesy of the Council for Geoscience South Africa.

Figure 4.19 (a) Map showing the location of heat flow sites (dark circles are from study of Ballard et al. (1987)). (b) Heat flow versus distance from the cratonic margin from the study of Ballard et al. (1987) and the references therein (solid and open circles).

Figure 4.20 Heat flow map of northern Namibia, derived from data of Ballard et al. (1987) and the references therein.

Figure 5.1 SAMTEX stations overlain on the regional magnetic map of Southern Africa. The red dots indicate the locations of the profiles (ETO-KIM, DMB, NEN, OKA-WIN, RAK-CPV and WIN) that are modelled in this thesis. The blue dots show the location of the KIM-NAM profile of Muller et al. (2009)’s study. The magnetic data are courtesy of the Council for Geoscience South Africa.

Figure 5.2 Map showing sites names overlain on the geology map of Namibia (modified from Corner (2008)). KL=Kamanjab Inlier, AL=Autseib Lineament, OML=Omaruru Lineament, OKL=Okahanja Lineament.

Figure 5.3 Pseudosection plot of the ETO-KIM profile, showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted triangles show site locations.

Figure 5.4 Pseudosection plot of the DMB profile, showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

Figure 5.5 Pseudosection plot of the NEN profile, showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.
Figure 5.6  Pseudosection plot of the OKA profile, showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

Figure 5.7  Pseudosection plot of the WIN profile (oriented SW-NE), showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

Figure 5.8  Pseudosection plot of the WIN profile (oriented E-W), showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

Figure 5.9  Pseudosection plot of the RAK profile (oriented E-W), showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

Figure 5.10  Niblett-Bostick depth estimates for all stations along the ETO-KIM profile for both the TE (red bars) and TM (yellow bars) modes. While this is based on 1D approximations it gives an indication as to the depth sensitivity of the MT response for each site.

Figure 5.11  Niblett Bostick depth maps for various depths.

Figure 5.12  The total integrated electrical conductivity, or conductance (S), from 40 km to 200 km. This depth range is approximately the mantle lithosphere from the average base of the crust to the average base of the lithosphere. The colours represent log_{10}(S).

Figure 5.13  Phase Tensor invariant ($\phi_{min}$) azimuth after (Caldwell et al., 2004) averaged over the 1-10s decade. One ellipsoidal principal axis indicates the direction of maximum induction. It can be seen that along a magnetic anomaly that corresponds to the Autsieb Lineament there is a rotation in the direction of the ellipses.

Figure 5.14  Phase Tensor invariant ($\phi_{min}$) azimuth after (Caldwell et al., 2004) averaged over the 10-100 s decade. One ellipsoidal principal axis indicates the direction of maximum induction. It can be seen that along a magnetic anomaly that corresponds to the Autsieb Lineament there is a rotation in the direction of the ellipses.

Figure 5.15  Phase Tensor invariant ($\phi_{min}$) azimuth after (Caldwell et al., 2004) averaged over the 100-1000 s decade. One ellipsoidal principal axis indicates the direction of maximum induction.
Figure 5.16 Results of MT STRIKE analysis along the ETO-KIM, DMB, NEN, OKA-WIN and RAK profiles, using Groom and Bailey (1989) decomposition. The analysis was done for the period bands indicated, with the shortest periods (0.01-10 s) showing crustal depths and the longest periods (100-1000 s) showing mostly upper mantle lithospheric depths. The black lines indicate the geo-electric strike azimuth and the coloured squares show RMS misfit error, normalised by observational errors in the observed MT response, at each site.

Figure 5.17 Sensitivity to strike direction for the four N-S profiles crossing the Damara belt, shown for crustal Niblett-Bostick depths (10-50 km) shows that only a few sites are strongly sensitive to strike direction, where low RMS values(<2) are defined over a limited range of azimuths. Other sites are relatively insensitive to strike direction and are characterized by low RMS values.

Figure 5.18 Sensitivity to strike direction for the four N-S profiles crossing the Damara belt, shown for upper mantle Niblett-Bostick depths (50-150 km) shows that only a few sites are strongly sensitive to strike direction, where low RMS values(<2) are defined over a limited range of azimuths. Other sites are relatively insensitive to strike direction and are characterized by low RMS values.

Figure 5.19 (a) Decomposition parameters obtained from GB analysis of MT sites along the ETO-KIM profile. (b) Final regional azimuths, relative to North, for depth ranges and sites. Final regional geo-electric strike direction is 50°.

Figure 5.20 (a) Decomposition parameters obtained from GB analysis of MT sites along the DMB profile. (b) Final regional azimuths, relative to North, for depth ranges and sites. Final regional geo-electric strike direction is 50°.

Figure 6.1 Two-dimensional isotropic model of the ETO-KIM profile. AL: Autseib Lineament. The Congo craton is clearly mapped, comprising resistive mantle lithosphere. The Moho boundary is inferred from seismic receiver functions.
Isotropic 2D inversion model for the DMB profile. The RMS variation of each station along the profile and the 100 km conductance profile are shown. Final RMS was 2.58. Highlighted on the inversion model are the Damara belt and Congo craton eLAB. The black horizontal line shows the crustal boundary and thin white line depicts the Niblett-Bostick depth of penetration for each station. A is the crustal granites and B corresponds to the location of the thrust faults, clearly dipping to the south based on MT data. 

TM modelled data vs original data, shown for the apparent resistivity and phase responses. The difference in the apparent resistivity and phase responses shows model sensitivity, where most of the structure in the final model is influenced primarily by the phase data.

TE modelled data vs original data, shown for the apparent resistivity and phase responses. The difference in the apparent resistivity and phase responses shows model sensitivity, where most of the structure in the final model is influenced primarily by the phase data.

2D Isotropic model of the NEN profile, with the cratonic segment highlighted. A: crustal granites and the Central Damara belt conductors a shown. Both TE and TM mode apparent resistivity and phase were modelled but with a higher error floor (50%) set for the TM apparent resistivity first, due to the high resistivity dominance in the model. The final model resulted in RMS 2.27, with 10% error floors for TE apparent resistivity and 5% error floors for TE and TM phases. The red dots shows the RMS recovered from model derived from using the Winglink, while the black data shows the original data.

Two-dimensional isotropic inversion models of the OKA-WIN profile, derived from joint inversion of the TE and TM mode data. ODS: Okavango Dyke Swam. The Kalahari craton in this part of the region is made up of the Rehoboth Terrane. The seismic line represent the S-wave receiver functions of a station located within the Damara belt.

Isotropic (a) and anisotropic (b and c) lithospheric models of the ETO-KIM profile. Conductivity in the direction perpendicular to the profile (Rxx) and parallel to the profile (Ryy) are shown. AL: Autseib Lineament.
A model showing the combined KIM-NAM profile from Muller et al. (2009) and the 2D isotropic ETO-KIM profile (see Figure 6.1), depicting the current tectonic structure from the Kaapvaal craton in the SW to the Congo craton in the NW. The KIM-NAM model was derived from 2D smooth inversion of joint apparent resistivity and phases for both TE and TM responses decomposed for 45° E of N azimuth. The algorithm used for the KIM-NAM profile was of Rodi and Mackie (2001) implemented in WinGLink®. The dotted black lines shows the approximate depth to base of crust.

Two-dimensional anisotropic model of the DMB profile, showing resistivity variations in the XX (perpendicular to the profile) and YY (parallel to the profile) directions. There is little anisotropy evident in the model, apart from crustal geometric variation of feature A, which appears thicker in the XX direction and feature B which is more conductive in the YY direction.

Anisotropic model of the NEN profile showing resistivity structure in the direction parallel (YY) and perpendicular (XX) to profile. No significant anisotropy is observed in the data. RMS is 2.44.

A generic lithospheric model (A) and its 2D inversion result (B) done in order to test the ability of MT data to resolve the LAB. The inversion appears to recover the full structure, for station spacing of 20 km (similar to the ETO-KIM profile) Inversion parameters: 100 Ohm.m starting model, both TE and TM mode data was used with the smoothness parameter $\tau=3$. A 2.5 % Gaussian noise was added to the data.

Independent 2D inversion of the TM and TE mode for a generic lithospheric model shown in Figure 6.11 to give and indication of which structures are dominant on which mode. While both modes are able to resolve the broad structure, the TM mode underestimates the eLAB by about 30 km.
Figure 6.13 (a) Crustal-scale resistivity model of the ETO-KIM profile, showing the main features resolved by the inversion process. The mid-lower crustal conductors, upper crustal Pan-African granites and the Archean basement (Kamanjab Inlier) are the major features of this model. For comparison the geological section (b), modified from Gray et al. (2007), is included. AL: Autseib Lineament, OML: Omaruru Lineament, OKL: Okahanja Lineament, MA=Matchless Amphibolite, SFZ: Southern Foreland Zone.

Figure 7.1 Location of the MT sites (black filled circles) used for 3D inversions, overlain on the basement map of southern Africa, modified from Precambrian basement maps of Singletary et al. (2003) for Botswana and Corner and Swart (1997) for Namibia.

Figure 7.2 Mesh/grid used in producing 3D inversion models having dimension 64,78,52 in x, y, z respectively. A 100 Ohm.m starting model was used for inversion runs. Top figure shows the NS 2D section across the middle of the grid, while the bottom grid is the depth slice.

Figure 7.3 Variation in RMS with respect to iteration. The inversion took 69 hours to complete on the STOKES cluster of the Irish Centre of High Performance Computing.

Figure 7.4 Depth of penetration interpolated across the 3D area (A) and shown for each station (B). The Central Damara conductor (DGCC), which results in limited penetration is particularly distinct on the image.

Figure 7.5 3D perspective view of the inversion results, fitting the full impedance tensor $Z_{xx}, Z_{yy}, Z_{xy}, Z_{yx}$. The resulting RMS was 3.66. CC=Congo craton, DGCC=Damara-Ghanzi-Chobe belts conductor.

Figure 7.6 Resistivity maps derived from the 3D inversion at three depths. The black-filled circles show station locations. Also shown is the 20 km resistivity slice overlain on the geology map in Figure 7.1. The correlation of the DGC conductor with the Central zone is clear from this map.
Figure 7.7 2D profiles of ETO-KIM, DMB, NEN, OKA-WIN transects extracted from 3D inversion model. Profile locations derived are shows beside each image. The S-receiver function response is projected on the DMB profile as it is in closest proximity. The interpreted MOHO and LAB conversions, the latter indicated by a black arrow from Kumar et al. (2007), are shown. The LAB conversion (180 km) derived by Hansen et al. (2009) was not particularly well resolved as the signal at these depths is close to bootstrap error limits ($2\sigma$).

Figure 7.8 Variation in RMS with respect to iteration. The inversion took 69 hours to complete on the STOKES cluster of the Irish Centre of High Performance Computing.

Figure 7.9 Starting model used to test the continuity of the Damara crustal conductor (10 Ohm.m) starting at 10 km depth.

Figure 7.10 RMS fits for each station (top) and plotted on a map (bottom), giving an indication of the resulting site-by-fit when using as input to the inversion a model shown in Figure 7.9.

Figure 7.11 Three-dimensional inversion of the ETO-KIM profile data, showing the variation in resistivity from Proterozoic Damara belt to the Archean Congo craton. RMS fit was 2.04 after inverting all four components of the impedance tensor. The dashed lines shows the top of the crustal conductor.

Figure 7.12 Three-dimensional inversion model derived from inverting only the DMB profile data. The resulting RMS was 1.68, after inverting the full components of the impedance tensor ($Z_{xx}, Z_{yy}, Z_{xy}, Z_{yx}$). A: upper crustal granites, B: Central Damara conductor.

Figure 7.13 Three-dimensional inversion of the NEN profile. Similar features to the other profiles are observed. The resulting RMS is 1.71.

Figure 7.14 Three dimensional inversion result of the OKA-WIN profile data. The resistive, thick Congo craton lithosphere is mapped to the north, the Okavango Dyke Swarm (ODS) is clearly imaged as resistive crustal feature. The central Damara conductor appears fragmented in this profile but this is the result of the profile orientation.

Figure 7.15 Africa Array seismic stations.

Figure 7.16 Fundamental mode Rayleigh wave dispersion curves extracted for stations in the Congo craton and the Damara belt.
Figure 7.17 Rayleigh-wave group-velocity maps, estimated for different periods. The dispersion curves were extracted for each MT station as such the spacing between stations places additional resolution limits to those imposed by event-station technique.

Figure 7.18 A comparison between seismic velocity perturbation, seismic lithospheric thickness estimates and electrical resistivity derived from 3D inversion.

Figure 7.19 Velocity model of Fishwick at 100 km

Figure 7.20 Stacked S-wave Receiver Functions (SRFs) at two Africa Array stations located within the Damara belt (TSUM and LSZ). The black line represents the mean stack while the grey shaded areas indicate the 2σ bootstrap error bounds. The LAB and Moho depths are indicated but the LAB estimates are characterised by high errors (Hansen et al., 2009).

Figure 8.1 (a) Crustal-scale resistivity model of the ETO-KIM profile, showing the main features resolved by the inversion process. The mid-lower crustal conductors, upper crustal Pan-African granites and the Archean basement (Kamanjab Inlier) are the major features of this model. For comparison the geological section (b), modified from Gray et al. (2007), is included. AL: Autseib Lineament, OML: Omaruru Lineament, OKL: Okahanja Lineament, MA=Matchless Amphibolite, SFZ: Southern Foreland Zone.

Figure 8.2 The current tectonic structure from the Kaapvaal craton in the SW to the Congo craton in the NW. The dotted line shows the inferred lithosphere-asthenosphere boundary.

Figure 8.3 Proposed Damara Orogen, Congo and Kalahari cratons evolutionary model.

Figure 8.4 Comparison of MT models derived from Archean-Proterozoic terranes in Australia (Selway et al. (2006), Selway et al. (2009)) and Canada (Spratt et al. (2009)).

Figure 8.5 Earthquake epicentres plotted on a Goggle Earth image, showing the distribution (clusters shown in white polygon) of earthquakes over a 20 year period. The location of the OKA-WIN profile is indicated. Their appears to be a linear trend of earthquakes associated with those on the East African Rift valley.

Figure 8.6 Development of a classic rift, characterised thinned lithosphere accompanied by a localised deep seated high conductivity mantle anomalies.
Figure 9.1  CCS storage cycle, including and illustration of possible CO\textsubscript{2} storage options in underground geological formations (Benson and Cook, 2005)  

Figure 9.2  Theoretical study of Archie’s Law showing variation in rock resistivity $\rho_r$ and porosity for various values of $m$. The fluid resistivity, $\rho_w$, was assumed to be 1 Ohm.m.  

Figure 9.3  Porosity-resistivity-salinity nomogram that can be used to estimate porosity from bulk resistivity measurement.  

Figure 9.4  Summary of the main electrical conductivity mixing models that are currently available (modified from Glover (2007)).  

Figure 10.1  Distributions of litho-stratigraphic units in the Main Karoo Basin (reproduced from Woodford and Chevallier (2002))  

Figure 10.2  Cross Section of the Main Karoo Basin (reproduced from Woodford and Chevallier (2002))  

Figure 10.3  Stratigraphic chart with the major litho-stratigraphic subdivisions of the Karoo Supergroup in the main Karoo Basin of South Africa (after Catuneanu et al. (2005)).  

Figure 10.4  A: Porosity and Permeability of sandstones and siltstones obtained from deep boreholes located in the Main Karoo Basin (after Roswell and de Swart (1976), table from Woodford and Chevallier (2002)). B: Drill information from shallow core boreholes in the Karoo Basin (Woodford and Chevallier, 2002)  

Figure 10.5  Map showing porosity and permeability trends, estimated from boreholes located in the Karoo basin (modified from (Viljoen et al., 2010) and (Cloete, 2010)). Boreholes MA1-69 and WE1-69 are located within the proposed storage site with porosities on the former estimated to be in the range 4-10 %.  

Figure 10.6  Possible deep saline formation storage opportunities onshore and offshore in Mesozoic basins along the coast of South Africa and for deep coalfields of the Karoo basin (Compiled by the Council for Geoscience in Cloete (2010)).  

Figure 10.7  MT profile plotted on litho-stratigraphic map of the Karoo basin  

Figure 10.8  MT stations used for CCS investigation, overlain on regional stratigraphic map of the Karoo Basin  

Figure 10.9  Karoo geological section from south-west to northeast section through the Cape and Karoo Basins, with the location of the MT profile highlighted. Note that section does no include the Drakensburg and Group lithologies. The Dwyka Group is very thin in the section (Viljoen et al., 2010).
Figure 11.1  MT responses for some sites along the profiles. The TE and TM mode apparent resistivity are plotted against increasing period (proxy for depth) ................................................................. 192

Figure 11.2  Sensitivity to strike direction shown for different Niblett-Bostick depths shows that only a few sites (e.g., KIM021 and KXT010) are moderately sensitive to strike direction, where low RMS values(<2) are defined over a limited range of azimuths. Other sites are relatively insensitive to strike direction and are characterized by low RMS values. ................................................................. 193

Figure 11.3  Results of MT strike analysis along profile using GB decomposition. Data period are labelled, with increasing periods a proxy for increasing depths. Double-headed vectors show 2-D geological strike direction, scaled in length by phase difference between the TM and TE mode. ................................. 194

Figure 11.4  Example of MT responses from a few stations along the profile after $45^0$ strike decomposition. (see Figure 10.6 for their location) 195

Figure 11.5  Pseudosection plot of the undecomposed responses along the profile. Please note that sites KIM0113-KIM017 were added here to derive this particular map but will not be used for further formal 2D inversion modelling as they fall outside the target area. 196

Figure 11.6  Pseudosection plot of the decomposed responses (azimuth 45) along the profile. ................................................................. 197

Figure 12.1  Deep electrical sounding curves in the Karoo basin (near Queenstown), which show a general increase in the resistivity structure due to dolerites. Also shown is the geoelectric section derived from sounding data near Beaufort West showing high sediment resistivities in the Karoo Trough diminishing to the north (van Zijl, 2006) ................................................................. 199

Figure 12.2  1D model of sites KIM021 and KIM014 ............................... 200

Figure 12.3  1D model of sites KXT004 and KIM215 ............................... 201

Figure 12.4  1D models of MT sites KXT009 and KXT010 located at the southern part of the profile ............................... 202
Figure 12.5 1D models with various stratigraphic layers within the Karoo basin indicated. For the Occam and sharp boundary 1D models, the depths are real modelled depths, whereas for the Bostick model the Niblett-Bostick depth approximation is used. Three 1D models are shown in each plot: the Occam model (thick black lines), the Bostick model (thin continuous lines) and the sharp boundary layer models (shaded zones). The stratigraphic positions and depth of each layer was inferred from the SA1/66 borehole on Figure 10.3.

Figure 12.6 A: Layered Karoo model based on known geological section and borehole (Figure 10.3). The resistivity were estimated from 1D inversion results (Figures 12.2, 12.3, 12.4, and 12.5). B: A forward MT response of one of the stations in A. C: Response from the deep electrical sounding work of van Zijl (2006).

Figure 12.7 1D forward model derived by assuming a layered Karoo stratigraphy.

Figure 12.8 Resistivity models derived using the 1D Occam inversion algorithm. Vertical bars beneath the stations indicate the sharp boundary layered 1D models. The background shows the 1D Occam inversion result, where the 1D models at each site. A good fit between borehole information and resistivity can be observed in the profiles.

Figure 12.9 2D inversion resistivity models derived from using smoothing parameter of 3 (top panel) and 0.1 (bottom panel).

Figure 12.10 Tradeoff curve determining an empirical solution for the best fitting smoothing parameter τ. The curve plots smoothness parameter on a logarithmic scale against the data misfit (RMS error). The best trade-off between both parameters can be found at the point with the highest curvature, in this case where τ = 3.

Figure 12.11 Independent 2D TE and TM mode inversion models showing moderate variation in resistivity structure, implying little anisotropic structure.

Figure 12.12 A 2D inversion model highlighting various stratigraphic horizons in the southern part of the MT profile. The NW portion is underlain by resistive basement rocks, possibly of the Kaapvaal craton. The 1D OCCAM models show the layered nature of the subsurface in the southern part of the profile.
Figure 12.13 2D models derived from joint TE and TM apparent resistivity and phase responses. The smoothing parameter tau was set at 0.5, with error floors for TE and TM apparent resistivity set at 15% and phase at 3%.  

Figure 12.14 2D Model responses compared to the measured data from joint inversion of the TE and TM mode apparent resistivity and phase data. Data up to 100 s was inverted. The RMS plots for each stations are shown.  

Figure 13.1 The borehole logs of the Beaufort sandstone (A and B) and Ecca sandstone (C) showing variation of porosity with depth. Data is from (Woodford and Chevallier, 2002)  

Figure 13.2 The variation in total dissolved solids (TDS), a measure of salinity, and source depth (in metres) of the groundwater. Note the increase in TDS with depth (data from Woodford and Chevallier (2002))  

Figure 13.3 TDS versus depth from borehole data (Woodford and Chevallier, 2002). These data were used to produce water resistivity estimates in the Figure 13.4  

Figure 13.4 Water resistivity estimates ($\rho_w$) vs depth, as calculated from the TDS values from Woodford and Chevallier (2002)  

Figure 13.5 Apparent porosity estimated from MT resistivity 1D and 2D models. Computations were done using Archie’s Law and the empirical relations of TDS and formation resistivity, calculated using Meju (2000)’s and Block (2001)’s equations.  

Figure 13.6 Apparent porosity estimated from MT resistivity 1D and 2D models compared to MA1-69 borehole porosity estimates. Computations were done using Archie’s Law and the empirical relations of TDS and formation resistivity.  

Figure 13.7 Apparent porosity estimated from MT resistivity 1D and 2D models compared to MA1-69 borehole porosity estimates. Lithological boundaries from Boreholes MA1-69, WE1-66 and WE1-69.

LIST OF TABLES
ACRONYMS

dgc Damara-Ghanzi-Chobe belts

elab electric lithosphere-asthenosphere boundary

sacccs South African Centre for Carbon Capture and Storage

samtex Southern African Magnetotelluric Experiment
## LIST OF SYMBOLS

<table>
<thead>
<tr>
<th>Symbol</th>
<th>SI Unit</th>
<th>Description</th>
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</thead>
<tbody>
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Continued on next page
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<td>$\Omega = \frac{\text{V}}{\text{A}}$</td>
<td>complex impedance tensor</td>
</tr>
<tr>
<td>$Z_{xx}, Z_{xy}, Z_{yx}, Z_{yy}$</td>
<td>$\Omega$</td>
<td>elements of the complex impedance tensor (in Cartesian coordinates)</td>
</tr>
</tbody>
</table>
INTRODUCTION

Magnetotellurics (MT) is a passive geophysical imaging technique where naturally varying electromagnetic (EM) fields are measured on the surface of the Earth. The modelling of these measured electric and magnetic fields allows us to infer the lateral and depth variations in the electrical structure of the subsurface. Unlike seismology which measures seismic impedances (P, S wave velocities and density) which are a function of bulk properties of rock matrix and varies by a few percent, electrical conductivity (and its reciprocal, resistivity), which is a function of minor constituents to rock matrix (but can be a bulk parameter) varies by about eight orders of magnitude. This fact alone makes the MT method far superior in imaging structural, lithological thermal and, if used with geochemical information, compositional variations in the Earth.

The MT method has been applied successfully on and off-shore, at different depth scales to investigate the lithospheric and thermal structure of the crust and upper to lower mantle. Apart from traditional seismic methods, MT is the only geophysical technique that is able to give us information about the physical and to some extent chemical structure of the deep lithosphere, using observed data.

In this work, the first ever lithospheric images of the lithosphere beneath the southern Congo craton and the Damara belt are presented. This work was done as part of the hugely successful multinational MT project called Southern African Magnetotelluric EXperiment (SAMTEX). Four (4) north-south (N-S) oriented MT profiles, labelled ETO, DMB, NEN, OKA, and one east-west (RAK) profile were modelled and the crustal and mantle lithospheric structure inferred. Additionally, and completely new area of research in the South African context, two (2) MT profiles across the central Karoo basin of South Africa were interpreted in an attempt to gain insight into the carbon capture and storage potential of South Africa.

This thesis is divided into five (5) parts. The first part (THE MAGNETOTELLURIC
METHODOLOGY) is divided into 3 chapters (chapter 2, chapter 3, chapter 4). The first chapter describes the historical development of the MT method (section 2.1), general theory (section 2.2), including a discussion on the MT impedance tensor (section 2.3) and the electromagnetic source fields relevant to the MT method (section 2.4).

The electrical conductivity of Earth’s materials and how they are imaged using MT is discussed in section 2.5. The dimensionality of the MT data is covered extensively in section 2.6 while distortion effects are discussed in section 2.7. The problem of static shift, which is a bane of the MT method, is covered extensively with practical and theoretical examples in section 2.8. This is followed by an overview of how MT data are acquired (section 2.9). The theoretical principles of the data processing (section 2.10), data analysis (section 2.11) and distortion correction (section 2.12) are presented. Finally in this chapter, the multi-dimensional modelling principles of the MT data are described (section 2.14).

The second chapter of the first part presents the summary of the SAMTEX project activities thus far (section 3.1) and the last chapter describes the geological background of the Southern African continent (section 4), including that of the Damara orogen, Congo and Kalahari cratons. Several tectonic models to be investigated are described in section 4.2, followed by a description of some of the geological and geophysical constraints on proposed models (section 4.3). Lastly section 4.4 gives an account of the previous geoscientific studies carried relevant for this research.

The second part of the thesis (INTERPRETATION OF THE 2D AND 3D MODELS) comprises four chapters (chapter 5, chapter 6, chapter 7, chapter 8). Chapter 5 deals with how MT data can be presented using pseudosections and electrical resistivity maps. Chapter 6 focus on the 2D isotropic and anisotropic inversion of the MT data on the profiles selected, while a full 3D modelling and interpretation are shown in chapter 7 and chapter 8 respectively.

The third part of the thesis (CARBON CAPTURE AND STORAGE) deals with an additional research component that involves the investigation of the carbon capture and storage potential of the Karoo basin of South Africa, using MT data. This part has five chapters to it. The first chapter 9 introduces the reader to carbon capture and storage (CCS) principles (section 9.1) including how the MT method can be applied to CCS (section 9.2). The second chapter 10 summarizes CCS activities in South Africa and explore possible storage sites in the Karoo basin of South Africa (section 10.1). The third chapter 11 deals with data processing; multi-dimensional modelling of the MT data is covered in the fourth chapter 12. The last chapter 13 concerns the
interpretation of the Karoo datasets and recommendations on the potential storage possibilities in the Karoo basin.

The fourth part of the thesis (CONCLUSIONS AND OUTLOOK) covers conclusions derived from the modelling and interpretation of 2D (section 14) and 3D (section 14) data. Furthermore, the interpretation in particular focuses on testing some of the traditional evolutionary models of the southern Congo and suggesting new tectonic models. In chapter 13 the results of CCS are summarized. The APPENDIX chapter A, contains papers published in international journals from the work done as pay of this PhD study.

1.0.1 Aims and motivation of my study

This work was originally planned as a two year Master of Science (MSc) by research but was subsequently converted to a PhD. The initial objectives of the MSc were to use the MT data to map the lithospheric structure of the Damara orogen and the southern extent of the Congo craton.

While a lot is known about the Kaapvaal craton lithosphere and indeed the greater Kalahari craton, very little is known about the deep geological structure that is north of the Kalahari craton. The great abundance of xenolith and seismic information has perhaps restricted Earth scientists to the Kaapvaal craton as those data provide an opportunity for multi-disciplinary research.

The greater Congo craton (see Figure 1.1) is to a large extent still enigmatic. Regional seismic tomography suggests that at its core, it is perhaps the thickest craton on Earth at 400 km. The craton itself is made up of an accumulation of Archean shields, the geometric structure of margins between these is largely unknown, which means that their evolution and influence on formation is still enigmatic.

In this work, the aim is to understand the southern-most boundary of the Congo craton and its relationship with the Kalahari craton and the Damara orogen in between them.

The conversion from an MSc to a PhD was motivated firstly by the huge amount of MT data available in the region and the opportunity it presented to study and understand continental collision processes prevailing in the Archean. The extension not only permitted more time for data analysis but also allowed for other datasets (i.e. seismic, gravity, magnetic and heat flow) to be incorporated, resulting in an com-
Comprehensive interpretation of the geology and lithosphere. The overarching objectives of this PhD project were thus to:

1. map the cratonic boundary between the Inland Branch of the Damara orogen that is emplaced between the Congo craton to the north and the Kalahari craton to the south,

2. understand the nature of the electric lithosphere-asthenosphere boundary (eLAB) beneath the Congo craton and the Damara-Ghanzi-Chobe belts (DGC),

Figure 1.1: Map of bedrock geology of Africa showing major tectonic blocks, modified from Begg et al. (2009)
3. test the validity of the subduction-related collisional model and the modified aulocogen model that have been proposed for the development of the Damara belt, and

4. develop a new tectonic evolutionary model of the Congo and Kalahari cratons and synthesize these with Gondwana formation and deformation.

By understanding the geometry of the margins, possible hypothesis could be inferred regarding formation and deformation processes during Gondwana accretion. The African tectonic landscape is a product of present and past rifting processes.

The famous East African Rift System (EARS) is one example of current extension tearing apart the African continent. It has been suggested, based on persistent seismic activity in NW Botswana, that the southern arm of the EARS extends as far south as the Okavango Delta. The availability of deep-probing MT data in this region provides a unique opportunity to test the rifting hypothesis. As such one of the major objectives in this work is to investigate the incipient rift theory and test its validity with available MT data. To this end, the questions being asked are (1) Is there evidence of classic rifting in a form of deep-seated localized high conducting mantle anomaly?, (2) Is incipient rifting process in this region initiated from the crust, along pre-existing shear zones?.

Another objective of my PhD research was to apply the MT method to test the suitability of the Karoo sediments to capture carbon dioxide (CO$_2$). This is the first time MT data has been applied in South Africa in this context. The work was funded by the South African Centre for Carbon Capture and Storage (SACCCS) and the aims were to quantitatively characterise a storage site identified from desktop study by SACCCS.

My involvement in the SAMTEX project started in February of 2008, during the fourth phase of data acquisition. I processed/re-processed and modelled all the data that is presented in this work. A significant amount of time was spent on obtaining good responses and inversion models as the testing of tectonic models depended heavily on the models generated. As a result, testing the robustness of the models formed a significant part of the research.
Part I

THE MAGNETOTELLURIC METHOD
THE MT METHOD

2.1 HISTORICAL DEVELOPMENT

The magnetotelluric (MT) method is still a relatively new technique when compared to other geophysical techniques such as seismology, magnetic and gravity methods. Since its theoretical development in the 1950’s, MT has become one of the fastest maturing scientific methods. The mathematical development of the MT method was proposed first by Russian scientist Tikhonov (1950). He showed the proportionality relationship, for low frequencies, between the electric field and the derivative of the horizontal magnetic field. Cagniard (1953), French scientist, later extended the theory and showed the mathematical amplitude and phase relations validity between the electric and magnetic field over a 1D layered Earth assuming a plane wave inducing field. These two authors are, rightly, credited with the development of the theory behind the MT technique (at least for the 1D case). The primary physical property being investigated when employing the MT method is electrical resistivity (or its converse conductivity).

The plane wave assumption of Cagniard (1953) was questioned by Wait (1954), the latter arguing that the proportionality relationship between the harmonic electric and tangential magnetic field is only valid for slowly varying waves (i.e., the fields should not vary appreciably in distance of the order of the skin depth. Price (1962) investigated and modified the theory of the MT method when distribution and dimensions of the ionospheric inducing field is considered, particularly when the latter is more extensive than the skin depth.

The development of MT data acquisition systems and sophisticated data processing and analysis methods has resulted in MT being applied more routinely in lithospheric-scale studies, minerals and environmental studies. The non-destructive nature of the MT method has resulted in data being collected in the most remote of regions. Perhaps the most attractive property of the MT method is its applicability at all
depth scales (i.e., from hundreds of metres to hundreds of kilometres). The depth of investigation is determined by the recording length.

2.2 MT Theory

For the purposes of the MT technique, the ionospheric source fields give rise to plane polarized electric ($\vec{E}$) and magnetic ($\vec{B}$) waves incident on the Earth’s surface described as follows:

$$\vec{E} = \vec{E}_0 e^{i\omega t - \vec{k}z}$$

$$\vec{B} = \vec{B}_0 e^{i\omega t - \vec{k}z}$$

where $\vec{E}_0$ and $\vec{B}_0$ are the origin amplitudes of the electric and magnetic fields respectively, $\omega$ is the angular frequency (period $T = \frac{2\pi}{\omega}$), $t$ is time, and the wavelength $\lambda = \frac{2\pi}{|\vec{k}|}$ and $\vec{k}$ is the wavenumber.

In general the propagation of electromagnetic waves in the Earth can be fully described by Maxwell’s equations. For linear isotropic materials, the inductive, conductive and capacitive properties that describes electromagnetic behaviour can be described by the following constitutive relationships:

$$\vec{J} = \sigma \vec{E}$$

$$\vec{D} = \epsilon \vec{E}$$

$$\vec{B} = \mu \vec{H}$$

where $\vec{E}$ ($V.m^{-1}$), $\vec{B}$ (Teslas $T$), $\vec{D}$ (Cm$^{-2}$), $\vec{H}$, $\vec{J}$ represent the electric field strength, magnetic induction, electric displacement, magnetic field strength, electric current density respectively. $\epsilon_0$ ($F.m^{-1}$) and $\mu$ (H.m$^{-1}$) are the dielectric permittivity and magnetic permeability of the substance and electric conductivity respectively.

The first constitutive relation (equation 2.3) is essentially the definition of Ohm’s Law, as it relates the material’s conductivity ($\sigma$) to the electric current density and the electric field. It follows from definition that the conductivity of a medium must be a tensor as both the electric field $\vec{E}$ and the current density $\vec{J}$ are vectors. If two of the orthogonal coordinate directions of the tensor are selected to lie in the direction of the
maximum and minimum conductivities, all non-diagonal elements of the tensor are zero. Considering isotropic (uniform in all directions) minerals or rocks, the three principal values of the conductivity are the same and conductivity can be treated as scalar.

The second constitutive relation (equation 2.4) relates the electric displacement \( \vec{D} \) to the electric field \( \vec{E} \) via another material specific property \( \epsilon \), electrical permittivity, the latter given by

\[
\epsilon = \epsilon_r \cdot \epsilon_0
\]

which relates the electrical permittivity of a vacuum (\( \epsilon_0 \)) to the relative permittivity (\( \epsilon_r \)) varying between 1 (vacuum) to 80.36 (water) at 20°C (Telford et al., 1976).

Electric permittivity is a measure of how efficient a dielectric material is to transmitting an electric field. At the very low frequencies used in MT dielectric permittivity can generally be ignored.

For a polarizable, magnetizable minerals Maxwell’s equations describe the behaviour of electromagnetic fields and how these fields interact with matter. These are described in the next section.

2.2.1 Maxwell’s equations

In 1873, James Clerk Maxwell published a two volume textbook (Maxwell, 1873) in which he described a set of four elegant equations that described electromagnetic theory and phenomena.

\[
\nabla \cdot \vec{E} = \frac{\eta}{\epsilon_0} \text{ [Gauss’s Law for an electric field]} \tag{2.7}
\]

\[
\nabla \times \vec{E} = -\frac{\partial \vec{B}}{\partial t} \text{ [Faraday’s Law of induction]} \tag{2.8}
\]

\[
\nabla \cdot \vec{B} = 0 \text{ [Gauss’s Law for magneticfield]} \tag{2.9}
\]

\[
\nabla \times \vec{H} = \vec{J} + \frac{\partial \vec{D}}{\partial t} \text{ [Ampere Maxwell Law]} \tag{2.10}
\]
where \( \eta \) is density of free electric charge and \( t \) is time. Using the vector identity
\[
\nabla \times (\nabla \times \vec{A}) = \nabla(\nabla \cdot \vec{A}) - \nabla^2 \vec{A}
\] (2.11)
and taking the divergence of Faraday’s Law (equation 2.8) results in
\[
\nabla(\nabla \cdot \vec{E}) - \nabla^2 \vec{E} = -\frac{\partial \vec{B}}{\partial t}
\] (2.12)
An assumption of the MT method is that there are no free charges in a layered Earth (i.e., \( \eta = 0 \)) and using Ampere’s Law (equation 2.10) yields,
\[
\nabla^2 \vec{E} = \mu \sigma \frac{\partial \vec{E}}{\partial t} + \mu \varepsilon \frac{\partial^2 \vec{E}}{\partial t^2}
\] (2.13)
An analogous relation can be determined for \( \vec{B} \) by taking the curl of the Ampere’s Law
\[
\nabla^2 \vec{B} = \mu \sigma \frac{\partial \vec{B}}{\partial t} + \mu \varepsilon \frac{\partial^2 \vec{B}}{\partial t^2}
\] (2.14)
Using equations 2.1 and 2.2, the above relations can be expressed as a function of angular frequency
\[
\nabla^2 \vec{E} + \mu \omega^2 \varepsilon (1 + \frac{\sigma}{\omega \varepsilon}) \vec{E} = 0
\] (2.15)
\[
\nabla^2 \vec{B} + \mu \omega^2 \varepsilon (1 + \frac{\sigma}{\omega \varepsilon}) \vec{B} = 0
\] (2.16)
where a propagation constant \( k \) (as a form from Helmholtz equation)
\[
k = \omega \sqrt{\mu \varepsilon (1 + \frac{\sigma}{\omega \varepsilon})}.
\] (2.17)
The first term in equation 2.17 describes displacement currents. Taking the free space values of permittivity and permeability results in the second term of equation 2.17 dominating, which describes conduction current (for typical MT spectral: \( 10^{-5} \text{s} < T < 10^5 \text{s} \)). Therefore equations 2.15 and 2.16 reduce to
\[
\nabla^2 \vec{E} + \omega \mu \sigma \vec{E} \nabla^2 \vec{B} + \omega \mu \sigma \vec{B}
\] (2.18)
Therefore the electric and magnetic fields pre-dominantly propagate via diffusion within the Earth (a process called quasi-static approximation).

In contrast, in the insulating air the fields travel via wave propagation. Note that since the magnetic permeability is assumed to be constant, \( \vec{B} \) and \( \vec{H} \) are interchangeable in
the previous equation.

Kalscheuer et al. (2008) showed that for frequencies between 10 - 300 kHz and considering a structure with resistivities greater than 1000 Ωm, the quasi-static approximation causes increasingly inaccurate forward responses and leads to inverse models with artificial structures. As such for very high frequencies such as the ones used in RadioMagnetotelluric (RMT), care must be taken in assuming the quasi-static approximation.

2.2.2 Apparent resistivity and phase

Equations 2.15 and 2.16 are second order differential equations, the solutions of which exists for a vertically incident plane wave. In a geometric system where z points vertically downwards, and the \( \vec{E} \) field is polarized along the x-axis, the \( \vec{H} \) field must be parallel to the y-direction, with the Poynting vector \( \vec{P} = \vec{E} \times \vec{H} \) describing the direction of propagation (Griffiths and College, 2010). The solution of this second order differential equations takes the form

\[
\vec{E} = E_1 e^{-\omega t} e^{-kz} + E_2 e^{-\omega t} e^{+kz} \tag{2.19}
\]

\[
\vec{B} = B_1 e^{-\omega t} e^{-kz} + B_2 e^{-\omega t} e^{+kz} \tag{2.20}
\]

where \( k^2 = \omega \mu \sigma \) is the wave number. Since the Earth does not generate energy but dissipates it as depth increases the incident field must decay to zero at infinite depths. This implies that the second part of the relation increases with depth and therefore its amplitudes \( E_2 \) and \( B_2 \) must be zero.

2.2.3 Fundamental assumptions of the MT method

The MT method has a number of assumptions that are made when considering induction in the Earth. These assumptions are made to simplify the theoretical and practical applications of the MT method (Vozoff, 1991; Simpson and Bahr, 2005).

1. Maxwell’s equations are valid and always obeyed.

2. The Earth does not generate electromagnetic energy, it only absorbs or dissipates it.

3. All fields can be regarded as conservative and analytic (i.e., the first derivative exists at all points) away from their sources.
4. The electromagnetic source fields utilised by the MT method may be treated as being uniform, plane-polarised electromagnetic waves, with near-vertical incidence to the Earth’s surface. This is commonly known as the plane-wave assumption and may be violated when the source field is too near, or in polar and equatorial regions (Wait, 1954; Price, 1962).

5. No accumulation of free charges is sustained in a layered Earth. However, in a 2D or 3D Earth, charges are accumulated and dissipated (on a cycle given by the frequency of interest) along conductivity discontinuities, producing the non-inductive static shift effect.

### 2.3 The MT Impedance Tensor

The relation between the horizontal electric ($E_x, E_y$) and magnetic ($B_x, B_y$) fields at a specific period is described by a second rank, frequency-dependant matrix $\vec{Z}$ as shown in equation 2.21.

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} B_x/\mu_0 \\ B_y/\mu_0 \end{pmatrix} \quad \text{or} \quad \vec{E} = \vec{Z} \frac{\vec{B}}{\mu_0} \tag{2.21}$$

An identical description of the transfer function was introduced by Weaver et al. (2000) wherein the MT tensor ($\vec{M}$) was derived using the $B$ instead of the $H$ field. Given that both the $\vec{Z}$ and $\vec{M}$ are complex, each matrix element contains the real and imaginary part

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} M_{xx} & M_{xy} \\ M_{yx} & M_{yy} \end{pmatrix} \begin{pmatrix} B_x \\ B_y \end{pmatrix} \quad \text{or} \quad \vec{E} = \vec{M} \vec{B} \tag{2.22}$$

#### 2.3.1 The geomagnetic transfer function

The relationship between the horizontal magnetic ($H_x, H_y$) and vertical magnetic ($H_z$) fields define a dimensionless quantity known as the geomagnetic transfer function or Tipper vector ($\vec{T}$):

$$H_z = (T_x, T_y) \begin{pmatrix} B_x/\mu_0 \\ B_y/\mu_0 \end{pmatrix} = (T_x, T_y) \begin{pmatrix} H_x \\ H_y \end{pmatrix} \tag{2.23}$$

Parkinson (1959, 1962) and Wiese (1962) devised a method of representing these tipper vectors as induction vectors. In practice the vectors are presented as two arrows
representing the real and imaginary components of the tipper vector on the xy plane as follows (Figure 2.1)

\[ \vec{T}_R = (\mathcal{R}(T_x), \mathcal{R}(T_y)) \],

\[ \vec{T}_I = (\mathcal{I}(T_x), \mathcal{I}(T_y)) \].

In the Earth, vertical magnetic fields are generated by lateral conductivity contrasts (Jones and Price, 1970; Jones, 1986; Simpson and Bahr, 2005) which means that induction vectors can be used to infer the presence and/or absence of conductive anomalies. When the Parkinson convention is used, the induction arrows point towards regions of anomalous current concentrations (as illustrated in Figure 2.1). The Wiese convention in contrasts uses arrows pointing away from regions of high internal conductivity. The tipper magnitude is given by

\[ |T| = \sqrt{|T_x|^2 + |T_y|^2} \]

and is proportional to the intensity of anomalous current concentrations that result in the vertical fields (Jones and Price, 1970). Jones (1986) showed for the Parkinson’s convention that in some cases (i.e., at sufficiently high frequencies) this assertion may be false in that the arrows in fact point away from anomalous zones whenever the surface observation site is at a location where return current flow associated with the anomalous horizontal electric field is sensed.
2.4 ELECTROMAGNETIC SOURCE FIELDS

The primary source fields used in MT exploration are in the exosphere. The magnetic field recorded at various points on the earth’s surface shows daily amplitude variations representing the effects of different current sources in the upper atmosphere (Campbell et al., 1998). Understanding the ionospheric electro-dynamics is critical understanding the natural source fields used in magnetotelluric and Geomagnetic Depth Soundings (GDS) experiments.

Many authors have studied the effects of solar activity on the Sq field (Rastogi and Iyer, 1976; Briggs, 1984; Olsen, 1998), but Takeda (2002) examined the correlation of the Sq amplitude with the electric conductivity of the ionosphere and with the sunspot number and found a positive correlations between the two. Furthermore, there is an apparent linear increase in Sq amplitude with the enhancement of the Hall and Pedersen currents, suggesting a local conductivity dependence.

Seasonal variations are also apparent in Takeda (2002)’s study where the Sq amplitude is higher in summer than in winter for the same conductivity value. This result was also observed by Olsen (1993) and Fukushima (1979) attributed to the effect of field-aligned currents which increase in amplitude in summer months and decrease in winter, since the Sq currents are in the same direction as the ionospheric currents in summer and reverse in winter. Dynamo current processes in the E region of the ionosphere (90-130 km) is the major source of Sq variations, away from polar regions (Campbell, 1989).

2.5 ELECTRICAL CONDUCTIVITY OF EARTH’S MATERIALS

In order to understand and interpret the electrical conductivity structure derived from MT method, it is important to know the electrical properties of rocks comprising the Earth’s interior. In this section the current knowledge on the electrical conductivity of rocks and material making up the Earth from the crust to the deep core is presented. Factors affecting the electrical conductivity of materials are investigated and their effects on the MT data interpretation are briefly discussed.

The Earth is made up of different types of rocks and fluids with varying bulk composition, thus one would intuitively expect its conductivity structure to vary vastly (see Figure 2.2). While attempts have been made to produce global conductivity models of the Earth, the results show huge variability, primarily because of the different geo-dynamic processes prevailing in the Earth and the sparse data coverage on
land and sea. The thermal structure of the Earth, elucidated from surface resistivity

![Resistivity ranges for common rock types](image)

Figure 2.2: Resistivity ranges for common rock types, modified from Marti (2006), Palacky (1988) and Miensopust (2011)

measurements is vital in providing an independent data constraint in understanding the evolution of the lithosphere. Xenolith geochemistry and seismology also provide information on the bulk composition of the Earth’s sub-surface. Petrological studies by Duba (1976), Shankland (1975), Constable et al. (1992), Jones et al. (2008) and Yoshino et al. (2008) have provided some constraints on the composition and temperature structure of the deep Earth. However in his classic paper, Duba (1976) questioned the relevance of the laboratory conductivity measurements and whether they represent in entirety the physical conditions in the Earth’s sub-surface. The delicate nature of these high-temperature-high pressure (HTHP) experiments naturally requires simulation of prevailing mantle thermodynamic and chemical conditions, measuring time and boundary conditions being some of the most vital components that require specific control.

Electrical resistivity of Earth materials varies by more than eight orders of magnitude. Figure 2.2 shows the dynamic range of resistivities of materials making up
the Earth composition; this makes the MT method more sensitive to anomaly detection than other geophysical (i.e., gravity and magnetic) techniques (Jones, 1999). The thermal structure of the Earth is generally inferred from heat flow studies but the magnetotelluric method, for reasons that will be discussed below, is also an effective tool for investigating the thermal structure of the lithosphere (Muller et al., 2009). The observed resistivity structure can be a consequence of many factors (discussed below) and unless other information (i.e., geology, geochemistry, etc.) is known about the area as to the possible causes of the conductivity, the interpretation of the MT data is subjective. It is thus imperative to understand some of the possible factors that may be responsible for the observed resistivity structure.

### 2.5 Electrical Conductivity mechanisms in Earth’s materials

In most Earth materials electric charge propagation occurs via three distinct mechanisms: **electronic**, **electrolytic** and **semi-conduction** (Jones, 1999; Nover, 2005; Simpson and Bahr, 2005).

**Electronic conduction** is more pronounced in good electrically conducting materials (i.e., metals, sulphides and graphite) as these materials have free electrons on their atomic structure. Nover (2005) notes that this conduction mechanism, although depending moderately on the thermodynamic parameters (i.e., pressure and temperature), can increase the conductivity of any medium by several orders of magnitude.

Within the Earth, **electrolytic conduction** dominates in fluid saturated rocks and partial melts due to presence of a conducting mineral phase, $H_2$ diffusion and the availability and mobility of free ions in solution.

**Semi-conduction** occurs in most common rock forming minerals, having few free electrons in the atomic structure, their temperature dependent conductivities described by an Arrhenius equation (Simpson and Bahr, 2005; Nover, 2005):

$$\sigma = \sigma_e e^{\frac{-E_e}{k_B T}} + \sigma_i e^{\frac{-E_i}{k_B T}},$$

(2.27)

where $E_e$ and $E_i$ are the activation energies for the extrinsic conduction (first term in Arrhenius equation) and intrinsic conduction at higher temperatures (second term) respectively, $k_B$ is the Boltzmann constant, $\sigma_e$ and $\sigma_i$ are the high-temperature asymptotes for each conduction type and $T$ is the absolute temperature.

Other physical factors contributing to the electrical conductivity of Earth’s materials include pressure, temperature, porosity, permeability, grain-size, presence of
conducting mineral phase, hydrogen diffusivity, anisotropy, presence of hydrous melt (H₂O, dissolved H+ ions), oxygen fugacity and iron content (Duba, 1976; Olhoeft, 1976; Heinson, 1999; Nover, 2005; Simpson and Bahr, 2005). The effect of these, particularly fluids (H₂O, CO₂), on conductivity at various levels within the Earth is still a topic of much research (Yoshino et al., 2008; Heinson, 1999). Conductivity of a rock material is enhanced if the conductive component (saline fluids, partial melt, graphite) forms an inter-connected network within the resistive host (Simpson and Bahr, 2005). The formation of this network depends on the presence of micro-structures and pores and the deformation history of a rock volume. Fractures and open pores facilitate the connectivity of the conducting fluid phase thus increasing the overall bulk conductivity of the rock. Archie (1942) presented an empirical formula (Archie’s law) relating the conductivity of the media to the conductivity of the electrolyte and porosity:

\[ \sigma = \sigma_p \eta^n, \]  

(2.28)

where \( \sigma \) and \( \sigma_p \) are the overall conductivities of the material and the conducting phase (electrolyte) respectively, \( \eta \) is the porosity and the exponent \( n \) is of order 1-2 (Jones (1999) and Shankland (1975)) varying according to the shape and interconnectivity of pores; 1 for high pore interconnection and 2 for low pore interconnection (Simpson and Bahr, 2005).

Salinity, ion charge and temperature are all factors affecting the conductivity of an electrolyte, and in materials with low porosity Archie’s empirical relation is modified such that the conductivity of the rock matrix is taken into account (Archie, 1942; Nover, 2005), such that

\[ \sigma = \sigma_r + (\sigma_p - \sigma_r)\eta^n. \]  

(2.29)

2.5.2 Continental and Oceanic Crust

The continental crust is made up of an intricate assemblage of sedimentary, igneous and metamorphic rocks each with a distinct evolutionary and deformation path, but being silica rich in bulk composition. The dynamic deformation processes induce fractures and pores in these rocks allowing fluids to fill, percolate and form an interconnected network, thus facilitating electrolytic conduction primarily in some geological terrains in the upper crust. The contribution to the bulk conductivity of the upper crustal material depends on the chemistry/salinity of the electrolytes.

The effect of water on conductivity is well documented (Schwarz, 1990; Jödicke, 1992; Yardley and Valley, 1997) even at mid-crustal depths. In the Soviet Kola deep drill-hole unusually large quantities of brines (up to 400 S/m) were observed at 11 km,
suggesting that fluid effects on conductivity of rocks and minerals can extend to
great depths (Jödicke, 1992). Similarly, the 9 km KTB borehole in Germany revealed
Archean brines up to 400 S/m (Emmermann and Lauterjung, 1997; Smithson et al., 2000).

Many MT studies around the world have indicated the presence of anomalous
high conductivity layers within various levels in the continental crust (Jodicke et al.,
1983; Beamish and Smythe, 1986; Bois et al., 1986; Green et al., 1986). The causes of these
anomalies is still an ongoing debate but it is generally accepted that in addition to
graphite and metallic oxides/sulphides, the presence of fluids in the lower crust can
explain most of the observed responses. In an attempt to account for the observed
anomalies in the crust, Jones (1987) presented a three-layer electrical conductivity
structure for the continental crust (Figure 2.3), where the first layer is a $10^4$ \(\Omega\).m upper
crust, reflecting the highly resistive fluid free silicates and carbonates rocks (Bedrosian,
2007). The lower crust he divides into two distinct parts, the difference being the
presence of interconnecting free water in each, the top part containing well connected
fluids ($10^2$ - $10^3$ \(\Omega\).m) and the bottom lower crust having pockets of fluids (resistivities in
the $10^2$ - $10^3$ \(\Omega\).m range). Saltwater with conductivity of up to 3 S/m saturates oceanic

![Figure 2.3: Resistivity model of the continental crust with a resistive upper crust underlain by
a two-layer lower crust. (modified after Jones (1987) and Schwarz (1990)).

sediments, and to some extent crustal rocks due to the former’s high porosity and
permeability, promoting electrolytic conduction of current as per Archie’s law. The
magnitude effect of this current conduction mechanism depends on the age of the
oceanic lithosphere (Palshin, 1996).
Figure 2.4 represents the summary of sea-floor resistivity as a function of depth (Constable, 1990), from interpretations of MT and borehole data. It is readily apparent from this conductivity structure that there is a general increase in resistivity from the sea-floor sediments to the depths of around 1-2 km perhaps due to closure of fluid bearing pores/cracks. There is then a sharp decrease in resistivity interpreted to represent partial melting at the base of the lithosphere, correlating with the seismic upper mantle low velocity zone (Heinson, 1999).

![Figure 2.4: Sea floor resistivity as a function of depth (after Constable (1990))](image)

2.5.3 Continental and Oceanic Mantle Conductivity

The resistivity of the continental lithospheric mantle is difficult to discern with absolute certainty due to it lying between relatively conductive lower crust and
2.5 Electrical Conductivity of Earth’s Materials

As discussed in detail by Jones (1999), in the same light, deep mantle resistivity mapping is also complicated by the screening effect of the electrical asthenosphere in addition to the rapid decrease of the power spectrum at larger periods.

Ultramafic minerals olivine and pyroxene are the dominant minerals in the upper mantle but most laboratory work has been done on single crystal olivine due to its higher volume. Duba (1976) published some of the initial results on the laboratory conductivity of olivine. Values of over 1,000 Ωm for the electrical resistivity of olivine in the sub-continental lithospheric mantle (SCLM) have been quoted by some authors (Constable et al., 1992).

Peridotite (ultramafic rock containing olivine, pyroxene and small proportion of spinel or garnet) is the most dominant rock that make up mantle composition (certainly true for 99% of samples brought up by kimberlites in Southern Africa). There have been various recent lab studies on dry and hydrous olivine, molten silicates and molten carbonates to explain the resistivity structure of the mantle. A brief summary of these is shown in Figure 2.5. Hydrous olivine or silicate melts have traditionally been invoked to explain the elevated conductivity in the mantle. However MT results, particularly in the Pacific (Baba et al., 2006) seems to suggest much higher conductivities and anisotropic conductivities in the mantle that cannot be explained by water.
dissolved in olivine. As such other conducting mineral phases must be invoked. Gaillard et al. (2008) presented experimental data showing that molten carbonates are by far the most conductive phase in the Earth’s upper mantle.

There exist three distinct discontinuities in the mantle at 410 km, 520 km and 660 km as indicated from the seismic velocity structure (Dziewonski and Anderson, 1981). These are attributed to phase transformations of olivine to wadsleyite, ringwoodite and perovskite+magnesiowustite. Figure 2.6 shows a summary of the phase changes occurring in the pyrolite mantle (Ono, 2008). Although in some places deep MT

![Figure 2.6: Mineralogical model of pyrolite mantle. The thicknesses of the D' layer varies according to the locality. The Mg-perovskite shows the (Fe, Al)-bearing composition (Irifune and Ringwood, 1987). Candidates for the lighter elements are: O, S, Si, C and H (after Ono (2008))](image)

studies are consistent with some of the laboratory measurements, the MT results consistently provide higher estimates of conductivity (Hirth et al., 2000), some of the primary reasons being the inadequate knowledge of the exact conduction mechanisms dominant in the mantle, the effect of mantle anisotropy, the screening effect of the electrical asthenosphere and current tectonic processes. Furthermore MT studies indicate higher mantle conductivities than laboratory derived olivine conductivities (Constable et al., 1992; Jones et al., 2001), suggesting that other factors may be influencing mantle conductivities, one of which is the hydrogen diffusion (Simpson and Bahr, 2005; Hamilton et al., 2006). A complete review is given by Simpson and Tommasi (2005);
however a linear relation describing the effect of hydrogen diffusivity on mantle mineral conductivity (olivine) is given by the Nernst-Einstein equation:

\[ \sigma = \frac{fDqc^2}{k_B T} \]  

(2.30)

where \( f \) is the unity factor, \( D \) is the diffusivity, \( c \) is the concentration of diffusing ions, \( q \) is the electrical charge, \( k_B \) is the Boltzmann constant and \( T \) is the absolute temperature.

Among other factors, hydrogen diffusivity is thought to be one of the main causes of electrical anisotropy. Based on experimental data on the solubility and mobility of hydrogen in olivine, Karato (1990) proposed that hydrogen (proton) may enhance electrical conductivity in deep mantle. This proposal was a clear departure from the traditionally held view of partial melt being the cause of enhanced conductivities in the mantle (Shankland et al., 1981) as relatively large melt fractions would be difficult to maintain in the dynamic Earth.

Studies by Yoshino et al. (2006) and Wang et al. (2006) showed the enhanced conductivity in olivine by hydrogen. What has largely been uncertain is the amount hydrogen required to explain the observed conductivities, mainly due to experimental limitations. Wang et al. (2006) estimated 0.01wt% of water in the asthenosphere from the electrical conductivity of olivine. Yoshino et al. (2008) investigated the effect of water on mantle transition zones (Figure 2.7) and observed that current conductivity profiles derived from MT experiments can be corroborated by a dry-mantle and that small amounts of hydrogen (\( H^+ \)) and iron (\( Fe^{2+} \) and \( Fe^{3+} \)) exert variable influence on the conduction efficiency of wadsleyite and ringwoodite. In the SCLM the dominant conduction mechanisms are likely to be ionic and electronic due to the movement of mobile charged ions (i.e., partial melt) and electrons (i.e., graphite or boundary film carbon) respectively.

Many geo-modellers have tried to present the global conductivity models, one of these studies being that of Campbell et al. (1998) as illustrated in Figure 2.8, comparing models from other regions. It is readily apparent that there is a general linear trend of increasing conductivities with increasing depth in the mantle and the various studies seem to be outlining some of the discontinuities mapped using seismic methods. Oceanic upper mantle is dry (<0.1 water) and highly resistive, estimated to be of the order \( 10^5 \) \( \Omega \)-m compared to the less resistive continental lithospheric mantle (Constable, 1990), this value derived from EM studies and very much consistent with laboratory determined values of olivine. Except at depths in the interval 300-400 km,
Figure 2.7: Electrical resistivity profiles and the estimated water content in the mantle transition zone. The orange and bluish areas represent geophysically observed conductivity profiles in the Pacific from Kuvshinov et al. (2005) and the continental mantle from Olsen (1998), Tarits et al. (2004) and Neal et al. (2000), respectively. The thick solid line represents the electrical conductivity of olivine, wadsleyite and ringwoodite without water. Dashed lines indicate the electrical conductivity of hydrous olivine, wadsleyite and ringwoodite as a function of water content (red: 1.0 wt%; green: 0.5 wt%; blue: 0.1 wt%). Light green solid line denotes the previous experimental result of Xu et al. (2000). The electrical conductivity of hydrous olivine was estimated from the average of three crystallographic axes. In the olivine stability field, thick and thin lines indicate the electrical conductivity estimated from different conditions of oxygen fugacity, with MoMoO$_2$ and NiNiO buffers, respectively (Yoshino et al., 2008).

MT experiments suggest some lateral variation of mantle electrical conductivity from continents to oceans, the former being significantly more conductive than the latter. Hirth et al. (2000) attributed it to varying water content in their respective materials in addition to other factors. One process that is thought to be the cause of increased volatile content in the mantle is rehydration associated with subducting lithosphere, where for instance Li et al. (2008) found up to 45 ppm H$_2$O in olivine, 402 ppm H$_2$O
2.5 Electrical Conductivity of Earth’s Materials

Figure 2.8: Comparison of global conductivity models (labelled according to their bibliographic references) as derived from geo-electric experiments (after Campbell et al. (1998)). The result for the Australian hemisphere is labeled CBCW (solid squares); CS 1988 (open squares) is the Australian hemisphere result of Campbell and Schiffmacher (1988). The various models are labelled according to their bibliographic references.

in orthopyroxene and 957 ppm H₂O in clinopyroxene. In comparison Peslier (2007) analyzed results of continental mantle samples in Southern Africa and found a range of water contents between 20-160 ppm H₂O.

Velinsky et al. (2006) used three years of CHAMP satellite vector magnetic data to produce a 1-D layered electrical conductivity inversion models for different parameterizations. The results, summarized in Figure 2.9, show the best models derived and the corresponding models with misfit of 0.2 from the minimum. A uniform 50 km thick crustal layer with conductance of approximately 0.1 S/m is resolved in all the models. The upper mantle conductivity is poorly resolved, a consequence of the resistive lithospheric mantle being embedded between conducting lower crust and electrical asthenosphere (Jones, 1999). The conductance value of 0.01 S/m attributed to the upper mantle by Velinsky et al. (2006) is a crude approximation. A lower mantle conductance of 6 S/m is estimated with fixed crust (50 km), upper mantle transition
(440 km) and lower mantle (670 km) seismically-based interfaces, consistent with the laboratory measurements of silicate perovskite (Ohta et al., 2008). Perovskite, a magnesium-silicate mineral, is believed to be the dominant phase in the lower mantle.

One of the most enigmatic areas of the Earth’s interior is the D" layer that forms the bottom of the lower mantle. This chemical boundary layer is believed to be the cause of the enhanced electrical conductivities at the base of the mantle and the source for mantle plumes. Seismic investigators have long known of the existence of this core-mantle D" boundary layer and attributed it to a post-perovskite phase change, perhaps a more metallic iron-rich alloy but the current mineral physics, mixing laws and mantle theoretical models cannot adequately account for its composition.

2.6 DIMENSIONALITY

The ability of the MT method to image effectively at all depth scales is complicated by the changes in the dimensionality of the subsurface. A much more complex Earth results in a much more complex impedance tensor. Consider a simple Earth model shown in Figure 2.10 showing a 3D body embedded in a homogeneous layered Earth. 1D responses will be obtained provided the period of interest are much smaller than the dimensions of the body. However with increasing periods (corresponding
skin depth) greater than the dimensions of the body, the effect of the 3D body on the EM responses decreases but the frequency independent galvanic distortion still persists. A multidimensional response results when the skin depth is comparable to the dimensions of the body.

Figure 2.10: 3D body of conductivity $\sigma_2$ embedded in a homogeneous halfspace of conductivity $\sigma_1$, illustrating the scale dependence of dimensionality of the Earth. Galvanic distortion will persist at long periods depending on the skin-depth and dimensions of the 3D body. Modified from Simpson and Bahr (2005).

2.6.1 1D Earth

The simplest most conceivable 1D Earth model is one where the subsurface consist of layers with different physical properties, resistivity for MT, but could equally well be different densities or seismic velocities. This idealised model is shown schematically in Figure 2.11. In this case the impedance tensor $\tilde{Z}$ can be expressed as

$$\tilde{Z}_{1D}(\omega) = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} = \begin{pmatrix} 0 & Z_{xy} \\ -Z_{xy} & 0 \end{pmatrix}. \quad (2.31)$$

The diagonal components ($Z_{xx}$, $Z_{yy}$) reduce to zero and the off-diagonal terms ($Z_{xy}$, $Z_{yx}$) have the same magnitude (but different sign to preserve right hand rule) since there are no lateral variations in conductivity. Wait (1954) derived an analytical solution to calculate surface impedance using an iterative process starting from the lowermost homogeneous layer:

$$Z_n = \frac{\omega \mu}{k_n} \left[ k_n Z_{n+1} + \omega \mu \tanh(k_n l_n) \right]$$

$$\omega \mu + k_n Z_{n+1} \tanh(k_n l_n) \right] \quad (2.32)$$
where

\[ k_n = (1 - i) \sqrt{\frac{\omega \mu \sigma_n}{2}} \]  

(2.33)

and the thickness of the \( n \)th layer is \( l_n = z_n - z_{n-1} \).

The apparent resistivity \( \rho_a \) of the layered media can thus be expressed in terms of a complex transfer function \( Z \):

\[ \rho_a = \frac{|Z(\omega)|^2}{\mu_0 \omega}. \]  

(2.34)

The phase \( \phi \) is accordingly given as the inverse tangent of the ratio of the real and imaginary part of \( Z \):

\[ \phi = \arctan \left( \frac{\Im Z}{\Re Z} \right). \]  

(2.35)

The Kramers-Kronig formulation describes the coupling of the apparent resistivity and phase due to causality by a dispersion relationship (Weidelt, 1972):

\[ \phi(\omega) = \frac{\pi}{4} - \frac{\omega}{\pi} \int_0^{\infty} \frac{1}{\omega^2 - \omega'^2} \log \left( \frac{\rho_a(\omega')}{\rho_o} \right) d\omega' \]  

(2.36)
The phase of a homogenous half-space can thus be calculated from the \( \rho_0 \) scaled apparent resistivity. For a two layer model equation 2.32 reduces to

\[
Z_s = \frac{\omega \mu}{k_1} \left[ \frac{k_1 + k_2 \tanh(k_1 z_1)}{k_2 + k_1 \tanh(k_1 z_1)} \right]
\]  

(2.37)

- **High Frequency Limit**: In the case where \( \tanh(k_1 z_1) \to 1 \), equation 2.37 reduces to \( Z_s = Z_1 \) and \( \rho_{a,S} = \rho_1 \). Since the skin depth is small at high frequencies, MT is only sensitive to the upper layer only.

- **Low Frequency Limit**: In the case where \( \tanh(k_1 z_1) \to 0 \), equation 2.37 reduces to \( Z_s = Z_2 \) and \( \rho_{a,S} = \rho_2 \). The skin depth is large for long periods therefore the MT response is dominated by resistivity of the lower half-space.

These frequency limits can be explored further if the bottom layer is resistive and conductive. Setting \( \lambda_1 >> z_1 \) implies that \( k_1 z_1 << 1 \) and thus, \( \tanh(k_1 z_1) \to k_1 z_1 \), resulting in

\[
Z_s = \frac{\omega \mu (1 + k_2 z_1)}{k_2 + k_1^2 z_1}
\]  

(2.38)

For a **resistive basement** (\( \sigma_2 << \sigma_1 \)) \( k_2 << k_1 \), then equation 2.38 reduces to \( Z_s = \omega \mu / k_1^2 z_1 \) and \( \rho_{a,S} = [\omega \mu (\sigma_1 z_1)]^{-1} \). Apparent resistivity depends on the **conductance** (i.e., conductivity-thickness product), illustrating an inherent non-uniqueness of the MT method which is the inability to independently resolve the conductivity and thickness of a layer. If either parameter is required other independent information is required.

For a **conductive basement** (\( \sigma_2 >> \sigma_1 \)) \( k_2 >> k_1 \), then equation 2.38 reduces to \( Z_s = \omega \mu z_1 \) and \( \rho_{a,S} = \omega \sigma_2 \mu z_1^2 \). This result illustrates that the apparent resistivity depends on the thickness of the layer and MT is sensitive to the top of the conductive layer.

### 2.6.1.1 Skin Depth in a layered Earth model

In a layered Earth model it is possible to estimate the depth of penetration or **skin depth** of MT responses. The skin depth equation 2.39 represents one method of estimating the depth of penetration:

\[
\delta(T) = \sqrt{\frac{2}{\mu_0 \sigma(T) \omega(T)}} \approx 0.5 \sqrt{\rho_a(T) T},
\]  

(2.39)

The skin depth gives the depth at which the EM wave of period \( T \) decays to \( \frac{1}{e} \) (or 37 %) of its original value in a homogenous halfspace medium of resistivity equal to the
apparent resistivity $\rho_a$ at that particular period.

*Niblett and Sayn-Wittgenstein (1960) and Bostick (1977)* presented depth approximation methods that are to be applied to the apparent resistivity curves only (i.e., $\rho_a$) not the phase. Unlike in equation 2.39, these two techniques imply an attenuation factor of approximately $\frac{1}{2}$ instead of $\frac{1}{e}$. *Jones (1983a)* elegantly showed that the Bostick and Niblett’s approximation are very equivalent in that they give exactly the same resistivity-depth profiles. The penetration depth for Niblett-Bostick estimation is given by

$$\delta_{NB} = \sqrt{\frac{\rho_a(T)}{2\pi\mu}}.$$  \hfill (2.40)

and the resistivity at this depth is given by

$$\rho_{NB}(\delta_{NB}) = \rho_a(T) \frac{1 + m(T)}{1 - m(T)},$$ \hfill (2.41)

where

$$m(T) = \frac{\partial \log(\rho_a(T))}{\partial \log(T)} = \frac{T}{\rho_a(T)} \frac{\partial \rho_a(T)}{\partial T}$$ \hfill (2.42)

is the gradient of the apparent resistivity curve on the log-log scale.

### 2.6.2 2D Earth

*Jones (1983b)* explored the conditions under which a body may be treated as a 2D feature. In general, if resistivity within the Earth varies with depth and laterally then it is considered 2D. This situation is illustrated in Figure 2.12 where two regions of different conductivities are separated by a long a boundary, that could represent a conductive fault or dike in a conductive environment. As a direct consequence of Ohm’s law (equation 2.3), charge build up is expected across a boundary. As was discussed in section 2.2 one of the boundary conditions that are adhered to in MT theory is that current density $\vec{J}$ is continuous across a boundary. An implication of the continuity of $\vec{J}$ is that electric field $\vec{E}$ across the boundary must be discontinuous given that $\sigma_1$ and $\sigma_2$ have different values. Thus the amplitude of the $\vec{E}$ field will be different on either side, achieved by charge build up on the boundary surface (*Jones and Price, 1970*).

Figure 2.12 B shows the behaviour of the electric current at a boundary. Given the high intensity of the incident current density in shallow parts the amount of
Figure 2.12: A 2D model showing areas of differing resistivities separated by a boundary striking in x direction. Due to Ohm’s law and current being preserved the $E_y$ field is discontinuous across the boundary. The EM field is decoupled into 2 components with $E$ field parallel to strike called the TE mode and $B$ field parallel to strike called the TM mode.

Charge built up is more intense than in deeper parts (i.e., current density decreases with skin depth). It is worth asking as to how far from the boundary will charge build up have an effect. Dawsonn et al. (1982) investigated this by considering an ocean-coast setting and found that within the bounds $-y_1 < y < y_2$ (see figure 2.12) the current will be perturbed by boundary charges and outside this boundary, current flow will have adjusted to the one expected for homogeneous half-spaces $\sigma_1$ and $\sigma_2$.

For the ideal 2D Earth shown in Figure 2.12, electric and magnetic fields are mutually orthogonal. The x direction defines what is known as geo-electric strike direction, along which electric field ($E_x$) induces magnetic fields in the vertical plane perpendicular to strike direction ($H_y$). The magnetic field parallel to the strike ($H_x$) on the other hand, induces electric fields only in the vertical plane perpendicular to strike. In this case Maxwell’s equations can be decomposed, in component form, into two independent modes describing currents flowing parallel and perpendicular to the strike direction (x), called the Transverse Electric (TE) or $E$ polarization and Transverse Magnetic (TM) or $B$ polarization modes respectively.
The TE mode is composed of \( E_x, B_y, B_z \):

\[
\begin{align*}
\frac{\partial E_x}{\partial y} &= \frac{\partial B_y}{\partial t} = i\omega B_z, \\
\frac{\partial E_y}{\partial z} &= \frac{\partial B_z}{\partial t} = -i\omega B_y, \\
\frac{\partial B_z}{\partial y} - \frac{\partial E_y}{\partial z} &= i\omega B_x,
\end{align*}
\tag{2.43}
\]

The TM mode on the other hand is composed of \( B_x, E_y, E_z \):

\[
\begin{align*}
\frac{\partial B_x}{\partial y} &= \mu_0 \sigma E_z, \\
-\frac{\partial B_x}{\partial z} &= \mu_0 \sigma E_y, \\
\frac{\partial E_z}{\partial y} - \frac{\partial E_y}{\partial z} &= -i\omega B_x,
\end{align*}
\tag{2.44}
\]

The impedance tensor presented in equation 2.21 can be factorized for this 2D case, in strike coordinates, into:

\[
\tilde{Z}_{2D} = \begin{pmatrix} 0 & Z_{xy}(\omega) \\ Z_{yx}(\omega) & 0 \end{pmatrix} = \begin{pmatrix} 0 & Z_{TE}(\omega) \\ Z_{TM}(\omega) & 0 \end{pmatrix}
\tag{2.45}
\]

The off-diagonal elements \( Z_{TE} \) and \( Z_{TM} \) represent both the TE and TM modes and usually have different magnitudes and opposite sign, the latter resulting in \( xy \) and \( yx \) phases being in 1st and 3rd quadrants respectively (if positive time dependency \( e^{+i\omega t} \) is used). The diagonal elements are zero since the electric components are related to the magnetic components only.

As it shall be seen in section 2.11, recorded MT data is normally not in strike coordinates, as such the diagonal components of the impedance tensor are not zero. If a 2D interpretation is desired, then the off-diagonal elements of the observed impedance tensor \( \tilde{Z}_{obs} \) must be minimised by decomposing the tensor by an angle \( \phi \) around the vertical axis using a rotation matrix \( \tilde{R}_\phi \). The resulting impedance tensor \( \tilde{Z}_{2D} \) in strike coordinates is derived using the relation:

\[
\tilde{Z}_{2D} = \tilde{R}_\phi \tilde{Z}_{obs} \tilde{R}_\phi^T
\tag{2.46}
\]

where

\[
\tilde{R}_\phi = \begin{pmatrix} \cos \phi & \sin \phi \\ -\sin \phi & \cos \phi \end{pmatrix} \quad \text{and} \quad \tilde{R}_\phi^T = \begin{pmatrix} \cos \phi & -\sin \phi \\ \sin \phi & \cos \phi \end{pmatrix}
\tag{2.47}
\]

are the rotation matrix and its transpose respectively.
2.6.3 3D Earth

As we shall see later in section 2.7 it might not be possible to find an ideal angle in which to minimise the diagonal elements. In this case the conductivity distribution varies in 3D with depth and two lateral directions (x and y). The MT tensor (\( \vec{Z} \)) cannot be decomposed into TE and TM modes and requires a full determination of all four elements:

\[
\begin{pmatrix}
E_x(\omega) \\
E_y(\omega)
\end{pmatrix}
= 
\begin{pmatrix}
Z_{xx}(\omega) & Z_{xy}(\omega) \\
Z_{yx}(\omega) & Z_{yy}(\omega)
\end{pmatrix}
\begin{pmatrix}
B_x(\omega) \\
B_y(\omega)
\end{pmatrix}
\]  

(2.48)

\( \vec{Z} \) is complex and can be factorised into real (\( \vec{X} \)) and complex (\( \vec{Y} \)) components

\[
\begin{pmatrix}
Z_{xx} & Z_{xy} \\
Z_{yx} & Z_{yy}
\end{pmatrix}
= 
\begin{pmatrix}
X_{xx} & X_{xy} \\
X_{yx} & X_{yy}
\end{pmatrix} + i
\begin{pmatrix}
Y_{xx} & Y_{xy} \\
Y_{yx} & Y_{yy}
\end{pmatrix}
\]  

(2.49)

This means that \( \vec{Z} \) will have both magnitude

\[
\rho_{\omega ij}(\omega) = \frac{1}{\mu_0 \omega} |Z_{ij}(\omega)|^2,
\]  

(2.50)

and phase

\[
\phi_{ij} = \arctan \left( \frac{\mathbb{I} Z_{ij}}{\mathbb{R} Z_{ij}} \right)
\]  

(2.51)

It must be noted that for 1D Earth a similar transfer function is derived (see equation 2.34), however it equals to the absolute value of the off-diagonal elements scaled by \( \mu_0 \) and \( \omega \).

The above discussions on dimensionality can be summarised on Table 2.1, which shows the dependance of each of the tensor elements with increasing complexity of the subsurface.

<table>
<thead>
<tr>
<th>1D</th>
<th>2D</th>
<th>3D</th>
</tr>
</thead>
<tbody>
<tr>
<td>( Z_{xx}=Z_{yy}=0 ) and ( Z_{xy}=-Z_{yx} )</td>
<td>( Z_{xx}=-Z_{yy} ) and ( Z_{xy} \neq -Z_{yx} )</td>
<td>( Z_{xx} \neq -Z_{yy} \neq Z_{xy} \neq Z_{yx} )</td>
</tr>
</tbody>
</table>

Table 2.1: Impedance tensor elements vs dimensionality of the Earth

In section 7 we will apply 3D inversion techniques to regional dataset in order to model what is clearly 3D data.
2.6.4 Electrical anisotropy

In general, the resistivity of Earth’s materials varies spatially in different directions, a feature termed electrical anisotropy. Mathematically, anisotropic conductivity is represented by a second rank 3D tensor where Ohm’s Law (equation 2.3) adopts the form (Vozoff, 1980):

\[
\begin{pmatrix}
    j_x \\
    j_y \\
    j_z
\end{pmatrix} =
\begin{pmatrix}
    \sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\
    \sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\
    \sigma_{zx} & \sigma_{zy} & \sigma_{zz}
\end{pmatrix}
\begin{pmatrix}
    E_x \\
    E_y \\
    E_z
\end{pmatrix}
\] (2.52)

The conductivity is symmetric (i.e., \( \sigma_{xy} = \sigma_{yx}, \sigma_{xz} = \sigma_{zx}, \) and \( \sigma_{zy} = \sigma_{yz} \)) for purely ohmic conduction and positive-definite (Weidelt, 1999; Pek, 1997). These properties allow the tensor to be diagonalized and expressed by the principal conductivities (\( \sigma_1, \sigma_2, \sigma_3 \)) and three (3) rotations angles (Pek, 1997, 2002; Pek and Santos, 2006; Heise et al., 2006) relating the tensor’s principal axes \( (x',y',z') \) to the reference frame \( (x,y,z) \). Following the notation of Pek (1997) and Pek (2002), who defined the three angles \( (\alpha_S, \alpha_D, \alpha_L) \) in Cartesian coordinate system (referred to as Euler angles), the conductivity was expressed as:

\[
\sigma(x,y,z) = \begin{pmatrix}
    \sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\
    \sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\
    \sigma_{zx} & \sigma_{zy} & \sigma_{zz}
\end{pmatrix} = \vec{R}_x(-\alpha_S)\vec{R}_x(-\alpha_D)\vec{R}_z(-\alpha_L)\begin{pmatrix}
    \sigma_1 & 0 & 0 \\
    0 & \sigma_2 & 0 \\
    0 & 0 & \sigma_3
\end{pmatrix} \vec{V} \] (2.53)

where \( \vec{V} \) is \( \vec{R}_z(\alpha_L)\vec{R}_x(\alpha_D)\vec{R}_z(\alpha_S) \) and \( \vec{R}_x \) and \( \vec{R}_z \) are rotation matrices given by

\[
\vec{R}_x(\alpha) = \begin{pmatrix}
    1 & 0 & 0 \\
    0 & \cos \alpha & \sin \alpha \\
    0 & -\sin \alpha & \cos \alpha
\end{pmatrix} \] (2.54)

and

\[
\vec{R}_z(\alpha) = \begin{pmatrix}
    \cos \alpha & \sin \alpha & 0 \\
    -\sin \alpha & \cos \alpha & 0 \\
    0 & 0 & 1
\end{pmatrix} \] (2.55)

where the subscript describes the respective rotation axis.

At the resolving scales of the MT method, the cause of bulk anisotropy may the
result of mixing in a preferred orientation of two or more media of differing resistivities (Wannamaker, 2005). Illustrated in Figure 2.13 is a simplified view of a material with two different conductivities illuminated with an oblique electric field showing that current vectors and electric field are not necessarily parallel. For this laminar

![Diagram of two-dimensional strike and anisotropy](image)

Figure 2.13: Example representations of anisotropy in earth materials. Anisotropy often is represented in terms of minute alternating planes of different conductivity with strike (sometimes also called azimuth) and dip which may be non-parallel to geometric strike. Figure from Heise and Pous (2001) and Weidelt (1999).

geometry current density over both objects is slanted toward the strike direction relative to the average electric field (Wannamaker, 2005). It has been shown that the resistivity normal ($R_n$) and parallel ($R_p$) to the laminae shown in Figure 2.13 can be calculated as follows (Heise and Pous, 2001; Wang and Fang, 2001):

$$R_n = R_2(1 - V_1) + R_1V_1$$  \hspace{1cm} (2.56)

and

$$\frac{1}{R_p} = \frac{1 - V_1}{R_2} + \frac{V_1}{R_1}$$  \hspace{1cm} (2.57)

where $V_1$ is the volume fraction of material 1.

Pek (2002) formalised the background theory for anisotropy in a 1D layered model and
showed that the electric fields can be expressed in terms of aggregate conductivities $\Lambda_{ij}$:

\[
\begin{align*}
\frac{\partial^2 E_x}{\partial z^2} + i\omega\mu_0 (A_{xx} E_x + A_{xy} E_y) &= 0 \\
\frac{\partial^2 E_y}{\partial z^2} + i\omega\mu_0 (A_{yx} E_x + A_{yy} E_y) &= 0
\end{align*}
\] (2.58, 2.59)

where

\[
\begin{align*}
A_{xx} &= \sigma_{xx} - \frac{\sigma_{xz}\sigma_{zx}}{\sigma_{xx}}, A_{xy} = \sigma_{xy} - \frac{\sigma_{xz}\sigma_{zy}}{\sigma_{zz}} \\
A_{yx} &= \sigma_{yx} - \frac{\sigma_{yx}\sigma_{zx}}{\sigma_{zz}}, A_{yy} = \sigma_{yy} - \frac{\sigma_{yz}\sigma_{zy}}{\sigma_{zz}}
\end{align*}
\] (2.60, 2.61)

and $A_{xy} = A_{yx}$ from symmetry (Wannamaker, 2005). Consequently, a full conductivity tensor of a layer (or half-space) is impossible to resolve from plane-polarised EM field responses. Azimuthally anisotropic structure incorporating the elements of the rank-3 layer tensors can be used to describe an anisotropic 1D layered Earth (Pek, 1997; Wannamaker, 2005).

The extension of anisotropic studies into 2D was advanced by Pek et al. (2003) who also included topography (Pek and Santos, 2006). Baba et al. (2006) applied an extension of Rodi and Mackie (2001) isotropic code, solving for resistivity parallel and perpendicular to depth assuming the direction of anisotropy is known and coincides with the strike direction. Li et al. (2003) presented a method for inverting 2D anisotropic structure using arbitrary anisotropic directions. More recently Mandolesi and Jones (2012) presented a new technique to jointly invert MT and seismic structure in a 2D environment.

The cause of observed anisotropy, particularly for the mantle, is still debated, with hydrogen diffusion along olivine a-axes (Mackwell and Kohlstedt, 1990; Bahr and Duba, 2000; Gatzemeier and Tommasi, 2006), and interconnectivity of a conductive mineral phase (e.g., graphite) and partial melt along grain boundaries (Jones et al., 1992; Mareschal et al., 1995) being the primary candidates for mantle materials. The presence of anisotropy in the lithosphere is increasingly being taken into account when doing MT interpretation (Baba et al., 2006; Martí et al., 2010; Evans et al., 2011) and can manifest itself in macro or micro-scale.

Crustal anisotropy can result from different orientations of fluid-filled structures (i.e., strain induced) or layering of materials with varying physical properties (Wannamaker, 2005; Miensopust and Jones, 2011). In the lithospheric mantle strain induced
crystal preferred orientation (CPO) of anisotropic minerals, hydrogen diffusivity in mantle minerals, or presence of partial melt are the main causes for observed anisotropy (Wannamaker, 2005; Yoshino et al., 2006). It follows that understanding electrical anisotropy with depth has implications for the interpretational aspects of continental evolution. Sénéchal et al. (1996), Eaton et al. (2004) and Hamilton et al. (2006) investigated the presence of electrical anisotropy for the Kaapvaal craton, Grenville Front and Great Slave region respectively in an attempt to look for a symbiotic relationship between MT and seismic anisotropy, their results varied from weak to strong correlation between observed seismic and electrical anisotropy. What is clear however is that tectonic processes, including those forming Neoproterozoic collisional terranes, result in observed anisotropy variations. Thus it is important to understand anisotropy at all scales in order to infer present and past tectonic processes.

2.7 MAGNETOTELLURIC DISTORTION EFFECTS

The fundamental objective of any geophysical technique is to obtain an accurate information about the structure of the Earth. This aim is often not easily achievable because of instrument, data processing and/or modelling limitations. In the case of MT studies, our inability to correctly model the Earth at all scales naturally prevents us from obtaining the maximum information about the sub-surface structure of the Earth.

In using deep-probing (i.e., long period) MT technique, the primary objective is to obtain the correct three-dimensional (regional) electrical resistivity structure with data that is free of the effects of near-surface features/structures. These shallow (local) inhomogeneities contribute to the measured electric and magnetic field and their effect should be removed from the data (Berdichevsky and Dmitriev (1976), Jiracek (1990), Park (1985), Wannamaker et al. (1984)). It is thus necessary to understand the nature of the distorting features and if possible their effects quantified adequately. The discussion that follows is an attempt to describe the nature of distortion, the physical principles governing the problem, how one recognises distorted MT data and some of the established methods used to remove distortion effects from measured data.

2.7.1 What is MT Distortion?

In general one can define distortion of MT data as a “phenomenon produced by the presence of shallow and local heterogeneities which are either too small compared to the target of interest or are too spatially under-sampled to allow for interpretation”. In general the
measured MT data provide responses from the combination of regional (3D) Earth structures and shallow features (Figure 2.14). The effect of the near-surface features on the regional response depends on the frequency of interest (Berdichevsky and Dmitriev, 1976; Jiracek, 1990). According to Chave and Smith (1994) and following the notation of Utada and Munekane (2000), the electric field at an arbitrary position \( \vec{r} \) in a 3D medium can be formulated as follows:

\[
\vec{E}(\vec{r}) = \vec{E}_0(\vec{r}) - \omega \mu_0 \sum_j \int_V dV' g(\vec{r}, \vec{r}') \delta \sigma_j(\vec{r}') \vec{E}(\vec{r}') + \nabla \left( \frac{1}{\sigma_0} \nabla \cdot \sum_j \int_V dV' [g(\vec{r}, \vec{r}') \delta \sigma_j(\vec{r}') \vec{E}(\vec{r}')] \right)
\]  

(2.62)  

(2.63)
where \( V_j \) is the scattering body and \( \delta \sigma_j(\vec{r}') \) is the anomalous conductivity within a uniform 1-D background conductivity for which an analytic solution to the Green’s function exist, the latter given by

\[
g(\vec{r}, \vec{r}') = \frac{e^{(\gamma_0 |\vec{r}-\vec{r}'|)}}{4\pi|\vec{r}-\vec{r}'|},
\]

where

\[
\gamma_0 = \sqrt{\omega \mu_0 \sigma_0}
\]

and

\[
\gamma_0 = \frac{1}{|\Re[\gamma_0]|} = \sqrt{\frac{2}{\mu_0 \sigma_0 \omega}}
\]

is the inductive scale length which, according to Jones (1988) can be taken to be given by either the real part of Schmucker’s C function (the depth of the maximum eddy current flow, Weidelt (1972)), or by the Niblett-Bostick depth (Niblett and Sayn-Wittgenstein (1960); Bostick (1977)).

Similarly the magnetic field expression at an arbitrary position \( \vec{r} \) is given by

\[
\vec{B}(\vec{r}) = \vec{B}_0(\vec{r}) + \mu_0 \nabla \times \sum_j \int_V dV' g(\vec{r}, \vec{r}') \delta \sigma_j(\vec{r}') \vec{E}(\vec{r}')
\]

In order to understand the problem of MT distortion it is necessary to look at various distortion types and compare and contrast their nature and effect on the measured data (Chave and Smith (1994), Groom and Bailey (1989)).

2.7.2 Galvanic Effects

Galvanic effects result from charge build-up around surfaces of structures. Primary electric fields produce electric charges on surfaces of conductivity variations and these in turn produce secondary electric fields which add to the primary fields. Figure 2.15 (Jiracek (1990)) illustrates the galvanic phenomena if the distorting body is more conductive than the host material (a) and a resistive body in a conducting host (b). The total electric field diminishes over a conductive body and enhanced over a resistive body. The secondary electric field is in phase with the primary electric field (Smith and Sandwell (1997)).

Figure 2.15 also illustrates quite well the problem of current channelling discussed by Jones (1983b) that may persists at long periods. In order to visualize and quantify
Figure 2.15: Galvanic phenomena. The secondary fields $E_s$ adds to the primary fields $E_p$ for a conductive inclusion and subtracts from the $E_p$ for a resistive inclusion (after Jiracek (1990)).

This current channelling phenomena, forward response models were produced over a thinly conductive dyke embedded in a homogeneous resistive half-space (Figure 2.16). Galvanic distortion can be described mathematically as was shown by Chave and Smith (1994) and Smith (1997). The observed electric field $\vec{E}_{obs}$ is proportional to the regional field $\vec{E}_{2D}$ scaled by the electric galvanic distortion matrix $\vec{C}$ given by the relation

$$\vec{E}_{obs} = \vec{C}\vec{E}_{2D},$$

(2.68)

The magnetic field of the galvanic distorted currents $\vec{B}_{dist}$ is also proportional to the regional electric field $\vec{E}_{2D}$, scaled by the magnetic distortion matrix $\vec{D}$, and is given by

$$\vec{B}_{dist} = \vec{D}\vec{E}_{2D}$$

(2.69)

The two distortion matrices are given by

$$\vec{C} = \begin{pmatrix} C_{11} & C_{12} \\ C_{21} & C_{22} \end{pmatrix} \quad \text{and} \quad \vec{D} = \begin{pmatrix} D_{11} & D_{12} \\ D_{21} & D_{22} \end{pmatrix}$$

(2.70)
Figure 2.16: An example of current channelling phenomena over a highly conductive 1 ohm-m thin dyke embedded in a homogeneous 1000 ohm-m resistive half-space

The observed magnetic field is the sum of the regional magnetic field $\vec{B}_{2D}$ and the galvanic distorted magnetic field $\vec{B}_{\text{dist}}$

$$\vec{B}_{\text{obs}} = \vec{B}_{2D} + \vec{D}\vec{E}_{2D}. \quad (2.71)$$

It was shown in equation 2.21 that

$$\vec{E}_{\text{obs}} = \vec{Z}_{\text{obs}}\vec{B}_{\text{obs}} \quad (2.72)$$

The regional impedance can also be written as

$$\vec{E}_{2D} = \vec{Z}_{2D}\vec{B}_{2D} \quad (2.73)$$

Therefore equation 2.72 can be rearranged by substituting the equations 2.68, 2.71 and 2.73 to yield

$$\vec{Z}_{\text{obs}} = \vec{C}\vec{Z}_{2D}(I + \vec{D}\vec{Z}_{2D})^{-1}\vec{R}^T, \quad (2.74)$$

where $I$ is the identity matrix. Given that the 2D regional reference frame (or strike coordinate) does not usually correspond to the observation coordinate system, a rotation angle $\alpha$ must be applied to yield

$$\vec{Z}_{\text{obs}}(\alpha) = \vec{R}(\alpha)\vec{C}\vec{Z}_{2D}(I + \vec{D}\vec{Z}_{2D})^{-1} \quad (2.75)$$
The magnetic galvanic distortion is usually neglected for the MT case (see for example Garcia and Jones (2002)), therefore equation 2.74 reduces to

\[ \vec{Z}_{\text{obs}}(\alpha) = \vec{R}(\alpha) \vec{C} \vec{Z}_{2D} \vec{R}^T(\alpha). \]  (2.76)

An example of real data affected by current channelling is shown in Figure 2.17. In section 2.11, some techniques that are used to remove galvanic distortion are presented.

### 2.7.3 Inductive Effects

The inductive effects are distinct from the galvanic effects in two ways, firstly only the magnetic fields are the primary distorted fields and unlike in the galvanic case there is a phase difference (0-90°) between the primary and secondary magnetic field components (Jiracek (1990), Utada and Munekane (2000), Berdichevsky and Dmitriev (1976)). In an idealised 3D case, the contribution of each shallow body to the inductive component is a function of the inductive scale length (\(\Lambda_j\)) and the size of the body, \(L_i\), (Utada and Munekane (2000)). The ratio between these two quantities is the induction number, \(M\) and maybe expressed as

\[ M = \frac{L_i}{\Lambda_j} = \frac{L_i}{\sqrt{\omega \mu_0 \sigma_j \delta}} \]  (2.77)

Time varying magnetic fields (Figure 2.18) produces secondary electric fields flowing in a closed loop, as per Faraday’s law, which in turn induces secondary magnetic fields. For small bodies or low frequencies the magnitude of the induction effect decreases (Utada and Munekane (2000), Jiracek (1990)), hence the contribution of the induction effects is negligible for long period studies. Magnetic field distortion needs to be considered, particularly at short periods (Chave and Smith (1994)).

Figure 2.17: Example of current channelling from an MT site in the SAMTEX database
Static shift remains one of the most problematic of MT problems yet to be solved. The complex nature of the problem results in static shift being the least understood of all distortion effects. It is generally defined as the upward/downward amplitude scaling in the apparent resistivity curve due to galvanic effects \((\text{Berdichevsky and Dmitriev (1976)})\). It can be the result of multi-dimensional conductivity contrasts having dimension less than the true penetration depth of the frequency of interest \((\text{Simpson and Bahr (2005)})\).

The phase curves are generally not affected by static shift, due to the primary and secondary fields being in phase, which provides a proxy for first pass recognition of the static shift effects from response curves. Figure 2.19 is an example of two sites from the SAMTEX database displaying effects of static shift. The phase response curves (lower panel) are generally similar for the two modes, but the apparent resistivity curves, although having the same general behaviour, are shifted relative to each other, causing enhanced or diminished impedance magnitudes.
Figure 2.19: Two sites from the SAMTEX database displaying effects of static shift

Figure 2.20 shows an example of distortion applied to a continental-type model. The distortion tensor applied, explained in more detailed below, is also shown. The characteristic shift in apparent resistivity curves is clearly evident while the phase response are consistent with each other. Galvanic effect results in the amplitude change in the apparent resistivity that is period independent, hence the shift occurs over a the whole period range (Smith and Sandwell (1997)). This shift is more pronounced if
Figure 2.21: Structurally induced static shift model. Forward responses on all models the first and last models shows apparent resistivities of both modes converging to 100Ωm, while for the centre model one of the modes is shifted upwards as a result of the structure.

the distorting bodies are more conductive than their resistive hosts.

Another instructive example of the causes of static shift is shown in Figure 2.21. This model consist of three layers with resistivities as shown, with the central model’s top layer having a structural discontinuity separating layers of different resistivity. At long periods the forward responses on the first and last models shows all modes converging to an apparent resistivity value of 100 Ωm, while for the central model one of the modes is shifted upwards for sites closer to the structure. The phase responses remain unchanged for all the models.

Another example of static shift is shown in Figure 2.22, this time the effect of topography is investigated. For sites outside the topographic step, the response curves for the two polarization modes are exactly similar in character and coincide. On the other hand, sites on the topographic step exhibit a clear shift in one of the modes.

From the discussions and forward models produced it is clear that static shift can be due to a number of factors and if it is to be corrected one needs to understand its possible causes. The preceding discussion is a review of the problem of MT distortion, a form of geologic noise, usually encountered in measured data. It is very clear that although a lot of progress has been made, particularly in the theoretical
understanding of distortion effects, satisfactory account of the problem of static shift remains one of the outstanding issues in MT. Choosing an appropriate method to correct for static shift requires knowledge of the target depth of interest. In section 2.11 several practical and theoretical attempts that have been used to correct for static shift in the data will be presented.

2.9 DATA ACQUISITION

The acquisition of MT data begins with selecting a geological target. In Southern Africa, the primary objective of the SAMTEX project was to map the deep lithospheric structure of Precambrian regions. The instrumentation used to record MT data has improved significantly over the last decades. During the four phases of the SAMTEX project, BBMT data was recorded using instruments manufactured by Phoenix Geophysics (Figure 2.23). MT data needed to be recorded for extended periods of time in order to get long period responses of the deep Earth. Given the large depths of interest the spacing between the sites was approximately 20 km, along main roads.

On the surface of the Earth, the horizontal time variations of the naturally occurring electric \((E_x, E_y)\) and magnetic \((B_x, B_y)\) fields are recorded. In cases where the ground is softer and easier to dig the vertical magnetic field \((B_z)\) can also be recorded. Two types of data can be recorded at an MT site: Broadband MT (BBMT) and/or
Figure 2.23: BBMT and LMT instruments used for most of the phases of the SAMTEX project. The receivers boxes, MTU 5 s and MTU 5As, are able to record data in the MT data range (0.001 s - 50 000 s) and the AMT-MT range (0.0001 s - 50 000 s) respectively. The MTU 5 s/MTU 5 record up to two electric and three magnetic channels simultaneously. The BBMT instruments are manufactured by Phoenix Geophysics (Canada), while the LMT are manufactured by LVIV center (Ukraine).

Long Period MT (LMT). The difference between the two recording is the type of magnetometer used and the amount of recording time. BBMT data is recorded over relatively short time period compared to LMT.

In addition to recording location coordinates and elevation, GPS antenna provides continuous time signal of the recorded electric and magnetic fields. Furthermore, GPS synchronization means that multiple stand-alone units can be deployed simultaneously in almost any terrain, with no need for interconnecting cables. A small number of magnetic channels can be combined with many electric channels to form an economical system of virtually any size. The deployment of synchronized remote reference stations permits sophisticated noise-reduction methods during processing. The MTC-50 magnetic sensors coils weigh approximately 8 kg, and measure only 144 cm in length. They provide MT data at frequencies up to 1 000 Hz and down to 0.00002 Hz (www.phoenix-geophysics.com). Non-polarising lead-chloride electrodes are used for measuring the electric fields.
In the first phase of the SAMTEX project the LIMS instruments, supplied by the Geological Survey of Canada were used to record the LMT data. These were subsequently replaced by the LEMI-417 systems (shown in Figure 2.23) manufactured by LVIV center in the Ukraine (www.isr.lviv.ua/lemi417.htm). The LEMI-417 system is a 7 component MT system (3 magnetic and four electric channels) using a flux-gate magnetometer. The system also has a built-in GPS receiver to provide satellite synchronization of the internal clock. The manufacturer notes that the main advantages of the LEMI-417 system are the very low temporal drift and high accuracy of measurements, making it especially efficient for deep sounding application. Furthermore, the system is highly sensitive and has unique low power consumption.

During the SAMTEX project BBMT data were typically recorded over 2-3 days and LMT data acquired up to 3 weeks. The length of the recording times necessitated choosing a site location far away from electrical noise. In southern African typical noise sources are DC train lines, electric fences, power lines, mining and animal activity. A typical field set-up is illustrated in Figure 2.24. The recording unit is placed at the centre of the survey site, connected to a 12V battery. The electrodes are buried in a 30 cm deep hole in bucket filled with salted mud in order to lower the contact resistance with the ground and keep the electrodes moist during long recording times. During SAMTEX, 90-100 m (i.e., 45-50 m distance from the centre for each electrode) dipole lengths for the N-S and E-W electrodes were used. The magnetic coils, oriented N-S and E-W, were also levelled and buried to depth of about 30 cm in order to reduce the ground motion effects and to protect them from animal/human interference. The recording for long-period data is used using similar layout, but the electrodes used, when deploying LVIV instruments, use a Cu – CuSO₄ solution instead of Pb – Cl. The fundamental difference however between BBMT and LMT recording is the magnetic sensors used.

2.10 DATA PROCESSING

The main objective of data processing is to derive, from measured electric and magnetic fields, estimates of EM impedance functions in the frequency domain, from which apparent resistivity information is derived. As with any time series processing, statistical techniques are employed to minimise the noise and derive resistivity information from the time series information (an example shown in Figure 2.25). Historically least squares (LSQ) statistical techniques were used to derive transfer functions estimates, assuming that noise follows a Gaussian distribution (Swift (1967), Rankin (1969), Sims et al. (1971)). For noise that is not Gaussian distributed, a rigorous
Figure 2.24: A typical field set-up for recording MT data. The recording unit, placed at the centre of the survey site, is powered by 12V batteries and placed in a protective case metal box. Four electrodes orientated N-S and E-W are buried 45-50 m away from the ground electrode to record the variations in electric field. The magnetic field components are recorded by two N-S and E-W oriented horizontal coils and if possible one vertical broad-band coil. The GPS antenna provides the MTU with the site geographic coordinates and more importantly with a continuous time signal. The photos are from the phase four SAMTEX field work and manual screening of the data had to be performed to remove obvious outliers. This procedure is inherently error-prone, hence the development of statistically robust methods such as derived by Jones and Jodicke (1984) (tscascade), Egbert and Booker (1986) (EMTF: Electromagnetic Transfer Function), Chave et al. (1987) (BIRRP: Bounded Influence Remote Reference Processing) and Larsen (1989), are used.

Jones et al. (1989) provides an excellent summary of some of these techniques. The general processing sequence generally involves three main processes: (1) data pre-conditioning, (2) time-to-frequency conversion and (3) MT transfer function estimation. The steps followed in this process are:

1. Pre-processing of the time series involve inspecting the data for inconsistent/null/bad data points and removing any trends in the series

2. Fractionalize the recording in segments of equal length
3. Multiply each segment with a suitable window function, e.g., the Hamming window to avoid spectral leakage

4. Fourier transform each segment

5. Calculate the auto- and cross spectra for each segment

6. Calculate the resulting impedance from the auto- and cross spectra

7. Calculate the mean and the error from the various estimates of the transfer function

The following section describes these processes in a more detail.

2.10.1 Data segmentation and frequency domain conversion

Splitting the time series into various M segments (containing N samples) allows for manual or automatic removal of outliers and trends, and robust statistical computations to be performed later in the processing sequence. The length of the segments depend on the longest period of interest. A tapering window is applied to each segment to reduce spectral leakages (Jones, 1977; Gubbins, 2004), which is common in spectral analysis when Fourier Transforms are applied to a time series with finite...
data length. A statistically superior result will be obtained when more segments are used. Hamming window is an example of windowing function that is applied to this effect (example illustrated in Figure 2.26).

Once the above mentioned pre-conditioning is performed, _Discrete (or Fast) Fourier transforms_ (Brigham, 1974), _Cascade decimations_ (Wight and Bostick) or _wavelet transforms_ (Garcia and Jones, 2008; Zhang, 1997; Trad and Travassos, 2000) are applied to the windowed time series segments to convert the recorded $\vec{E}$ and $\vec{B}$ components into the frequency domain and derive the spectrum of each segment and for each channel. From these evaluation, frequencies in the range 6-10 per period decade are optimally chosen and the final power spectra smoothed using a Parzen window (Marti, 2006). A spectral matrix (shown below) for each evaluation frequency is then derived, which comprises the auto and cross-spectra for each segment, made up of the field components and their complex conjugates (e.g., $B_x(\omega) \cdot B^*_x(\omega); B_y(\omega) \cdot B^*_x(\omega); B_z(\omega) \cdot B^*_x(\omega); E_x(\omega) \cdot B^*_x(\omega); E_y(\omega) \cdot B^*_x(\omega)$ (Marti (2006)).

![Figure 2.26: Example of splitting and windowing a time series segment](image-url)
For the same evaluation frequency, several spectral matrices will be stacked and potentially manually edited or weighted using statistical techniques. An example of a windowed time-series segment after applying Discrete Fourier Transforms is shown in Figure 2.27, including the spectrum.

Figure 2.27: An example of windowed segment after applying Discrete Fourier Transform and the resulting power spectra

### 2.10.2 Estimating the transfer function

The next processing step is estimating the transfer functions; the impedance tensor (Equation 2.78 and 2.79) and tipper vector (Equation 2.80).

\[
E_x(\omega) = Z_{xx}(\omega) \cdot H_x(\omega) + Z_{xy}(\omega) \cdot H_y(\omega) + \delta Z(\omega) \tag{2.78}
\]

\[
E_y(\omega) = Z_{yx}(\omega) \cdot H_x(\omega) + Z_{yy}(\omega) \cdot H_y(\omega) + \delta Z(\omega) \tag{2.79}
\]

\[
H_z(\omega) = T_x(\omega) \cdot H_x(\omega) + T_y(\omega) \cdot H_y(\omega) + \delta T(\omega) \tag{2.80}
\]

\(\delta Z(\omega)\) and \(\delta T(\omega)\) represent uncorrelated noise (in this case electrical noise), that is required because measurement errors make the transfer function equations inexact. The recorded electric and magnetic fields are used to estimate the transfer functions as per the equations above.
2.11 DATA ANALYSIS AND REMOVAL OF DISTORTION

In section 2.7 we introduced the concept of distortion and several ways in which MT data can be distorted and how one can recognise when distorted responses. In this section we look at methods that can be utilized to correct for primarily galvanic distortion effects. For the type of land survey concerned magnetic distortion effects are not critical but galvanic distortion must be removed.

Distortion analysis and removal is one of the most critical steps in the robust interpretation of MT data. It must be undertaken to justify the modelling approach (i.e., whether to apply 2D or 3D inversion methods) and to determine the dominant strike direction if data is to be modelled in 2D (Ledo and Jones, 2002; Jones et al., 2005a; Jones, 2012). Failing to do this step will simply yield unreliable resistivity models. Jones (2012) recognised that the inability of the MT method to model the variation in electrical resistivity from microscopic scales to larger scales (larger than our experimental design) is manifested as distortion in the MT responses. There are some cases where 2D (even 1D) treatment of data is sufficient, however this conclusion must be arrived at by applying several tests of dimensionality.

2.12 CORRECTING GALVANIC DISTORTION IN MT DATA

Various tools have been proposed (discussed below) to deal with the distorting effects on MT data. The reader is referred to an excellent synthesis of these methods by Jones (2012). These range from practical/experimental methods to theoretical/statistical techniques. The merit of each technique, discussed in more detail below, is a function of a combination of factors ranging from data quality, data density, and assumptions made about the dimensionality of the Earth’s sub-surface.

A presumption of the dimensionality of the Earth subsurface is usually employed to correct for distortion, i.e., a 3D/2D model where 2D regional response is distorted by local near-surface 3D features (Bahr, 1988; Groom and Bailey, 1989; Lilley, 1998; McNeice and Jones, 2001a) or a model where 3D regional structure is distorted by local 3D features (Ledo et al., 1998; Ledo, 2005; Utada and Munekane, 2000). The methods for correcting distortion can be classified into 2 categories: (1) Phase-based and (2) magnitude-based techniques.

The method presented by Groom and Bailey (1989) is widely used for reasons that will be given below. As such it is explained first.
2.12 Correcting galvanic distortion in MT data

2.12.1 Groom-Bailey decomposition

By far the most widely used galvanic correction approach is the one implemented by Groom and Bailey (1989), first presented by Bailey and Groom (1987); Groom (1988). The method, henceforth called GB technique, is popular for two main reasons: firstly, it is based on a physical model of distortion where a 2D regional structure is distorted by local 3D anomalies. This model was presented by Larsen (1977). Secondly, the effects of local 3D responses are separated from regional 1D/2D responses of interest. This process is done, as was shown in equation 2.76, by factorising the observed impedance $\vec{Z}_{\text{obs}}$ into a rotation matrix $\vec{R}$, distortion matrix $\vec{C}$ and a regional impedance tensor $\vec{Z}_{2D}$. Groom and Bailey (1989)'s parametrization separates the distortion matrix into determinable and indeterminable parts by taking the product of the gain factor $g$ and three matric tensors twist ($\vec{T}$), shear $\vec{S}$ and local anisotropy $\vec{A}$:

\[ \vec{C} = g \vec{T}\vec{S}\vec{A} \]  \hspace{1cm} (2.81)

where

\[ \vec{T} = \frac{1}{\sqrt{1 + t^2}} \begin{pmatrix} 1 & -t \\ t & 1 \end{pmatrix}, \text{ where } t = \tan \phi_t, \]  \hspace{1cm} (2.82)

\[ \vec{S} = \frac{1}{\sqrt{1 + s^2}} \begin{pmatrix} 1 & s \\ s & 1 \end{pmatrix}, \text{ where } s = \tan \phi_s, \]  \hspace{1cm} (2.83)

\[ \vec{A} = \frac{1}{\sqrt{1 + a^2}} \begin{pmatrix} 1 + a & 0 \\ 0 & 1 - a \end{pmatrix} \]  \hspace{1cm} (2.84)

In the above equations $t$, $s$ and $a$ are the twist shear and anisotropy parameters, respectively. The shear and twist contribute to the telluric orthogonality and therefore phase mixing (Groom, 1988; Thiel and Heinson, in press). Thus the shear and twist give an indication of the strength of distortion, with high values indicating strong distortion. The gain and anisotropy factors cannot be independently determined since the system of equations to be solved is under-determined. This results in apparent resistivity curves that are scaled in magnitude, without changing the phase curves. This well-known phenomena is called Static Shift and to this day continues to be the bane of the MT method (Jones, 1988, 2012). Unless additional constraints are available (described in the next section) the absolute values of resistivity will not be uniquely determined.
The GB technique was extended to include multiple sites and periods by McNeice and Jones (2001a) in statistical manner, but keeping the twist and shear site-dependent, which improved the estimates of azimuth over a range of periods and therefore more superior to single-site estimates of Groom and Bailey (1989). McNeice and Jones (2001a) made free the STRIKE code that decomposes data using the GB approach, allowing also depth based decomposition (instead of period), given the uncertainties in the depth of penetration from site-to-site. This approach is very useful given that the same period might sense completely different depths as such cases it is instructive to define a depth range instead of a frequency range for the analysis. In Chapter 5 we will apply the GB method on the MT data.

Another method has been proposed to interrogate the MT tensor for distortion. Bahr (1988) proposed a measure of distortion of impedance tensor by considering a rotationally-invariant parameter called the phase sensitive skew; $\alpha$. Although a lot of effort has been put into the 3D/2D distortion problem some practitioners have suggested ways to correct for distortion in the presence of 3D regional structures (Ledo et al., 1998; Garcia and Jones, 2002; Utada and Munekane, 2000).

2.12.2 The MT Phase tensor approach

This technique was first proposed by Caldwell et al. (2004) (see also Bibby et al. (2005) and a Comment and Reply by Caldwell and Brown (2007) and Moorkamp (2007)) and has the advantage of not making any assumptions about the dimensionality of the underlying resistivity structure (2D/3D) and even in the presence of galvanic distortion the phase tensor preserves the phase information.

The MT phase tensor approach is based on the premise that near surface heterogeneities affect only the amplitude of the electric field and not the phase relationship between electric and magnetic field, assuming galvanic distortion. Caldwell et al. (2004) describes the phase tensor as follows: the relationship between the observed MT impedance $\vec{Z}$ and the regional MT impedance $\vec{Z}_R$, of any dimensionality, in the presence of distortion $\vec{C}$, is given by the relation $\vec{Z} = \vec{C}\vec{Z}_R$, where $\vec{Z}_R = \vec{Z}_R + i\vec{Y}_R$. The distorted real and imaginary parts can be written respectively as $\vec{X} = \vec{C}\vec{X}_R$ and $\vec{Y} = \vec{C}\vec{Y}_R$.

The real second rank phase tensor is defined as the ratio of the real $\vec{X}$ and imaginary parts $\vec{Y}$ of the complex impedance $\vec{Z} = \vec{X} + i\vec{Y}$

$$\vec{\Phi} = \vec{X}^{-1}\vec{Y}$$

(2.85)
From the relation above:
\[ \Phi = \bar{X}^T \bar{Y} \]
\[ = (C \bar{X}_R)^{-1}(C \bar{Y}_R) \]
\[ = \bar{X}_R^{-1}C^{-1}C \bar{Y}_R \]
\[ = \bar{X}_R^{-1}\bar{Y}_R \]
\[ = \Phi_R. \]

The phase tensor can be written in Cartesian coordinates in terms of real and imaginary parts as a matrix

\[
\begin{pmatrix}
\Phi_{11} & \Phi_{12} \\
\Phi_{21} & \Phi_{22}
\end{pmatrix} = \frac{1}{\det(\bar{X})} \begin{bmatrix}
X_{yy}Y_{xx} - X_{xy}Y_{yx} & X_{yy}Y_{xy} - X_{xy}Y_{yy} \\
X_{xx}Y_{yx} - X_{yx}Y_{xx} & X_{xx}Y_{yy} - X_{yx}Y_{xy}
\end{bmatrix}
\]

where \( \det(\bar{X}) = X_{xx}Y_{yy} - X_{xy}X_{yx} \) is the determinant of \( \bar{X} \). The 2-D tensor can be characterized (in the general case) by a direction and three independent scalar quantities that are independent of the coordinate system used to express the tensor. The three coordinate invariants used by (Caldwell et al., 2004) are the maximum (\( \Phi_{\text{max}} \)) and the minimum (\( \Phi_{\text{min}} \)) tensor values and the skew angle \( \beta \) given by

\[
\beta = \frac{1}{2} \arctan\left(\frac{\Phi_{12} - \Phi_{21}}{\Phi_{11} + \Phi_{22}}\right)
\]

Since \( \Phi \) is a rank 2, non-symmetric tensor is it can be represented graphically by an ellipse (Bibby, 1986) as illustrated in Figure 2.28. The major and minor axes of the ellipse (\( \Phi_{\text{max}} \) and \( \Phi_{\text{min}} \)) represent the principal values of the tensor with the orientation of the major axis specified by the angle (\( \alpha - \beta \)). The phase tensor can be written in terms of its invariants as:

\[
\Phi = \tilde{R}^T(\alpha - \beta) \begin{pmatrix}
\Phi_{\text{max}} & 0 \\
0 & \Phi_{\text{min}}
\end{pmatrix} \tilde{R}(\alpha + \beta)
\]

where \( \tilde{R}^T \) is the transpose of the rotation matrix

\[
\tilde{R}(\alpha + \beta) = \begin{pmatrix}
cos(\alpha + \beta) & \sin(\alpha + \beta) \\
-\sin(\alpha + \beta) & \cos(\alpha + \beta)
\end{pmatrix}
\]

and

\[
\alpha = \frac{1}{2} \arctan\left(\frac{\Phi_{12} + \Phi_{21}}{\Phi_{11} - \Phi_{22}}\right),
\]
2.12 Correcting galvanic distortion in MT data

Figure 2.28: Graphical representation of the phase tensor. The lengths of the ellipse axes, which represent the principal axes of the tensor, are proportional to the principle values of the tensor: $\Phi_{\text{max}}$ and $\Phi_{\text{min}}$. If the phase tensor is non-symmetric, a third coordinate invariant is needed to characterize the tensor: the skew angle $\beta$. The direction of the major axis of the ellipse is given by $\alpha - \beta$ and defines the relationship between the tensor and the observational reference frame ($x_1, x_2$). Redrawn from (Caldwell et al., 2004).

and

$$\beta = \frac{1}{2} \arctan \left( \frac{\phi_{12} - \phi_{21}}{\phi_{11} + \phi_{22}} \right). \tag{2.91}$$

The principal values $\Phi_{\text{max}}$ and $\Phi_{\text{min}}$ can be described by

$$\Phi_{\text{max}} = (\phi_1^2 + \phi_3^2)^{1/2} + (\phi_1^2 + \phi_3^2 - \phi_2^2)^{1/2}, \tag{2.92}$$

$$\Phi_{\text{min}} = (\phi_1^2 + \phi_3^2)^{1/2} - (\phi_1^2 + \phi_3^2 - \phi_2^2)^{1/2} \tag{2.93}$$

where $\phi_1$, $\phi_2$ and $\phi_3$ are related to the trace ($\text{tr}(\phi)$), the determinant ($\text{det}(\phi)$) and the skew ($\text{sk}(\phi)$) of the tensor respectively,

$$\phi_1 = \frac{\text{tr}(\phi)}{2} = \frac{\phi_{11} + \phi_{22}}{2},$$

$$\phi_2 = (\text{det}(\phi))^{1/2} = (\phi_{11}\phi_{22} - \phi_{12}\phi_{21})^{1/2} \text{ and}$$

$$\phi_3 = \frac{\text{sk}(\phi)}{2} = \frac{\phi_{12} - \phi_{21}}{2}.$$
\[ \lambda = \frac{[(\phi_{11} - \phi_{22})^2 + (\phi_{12} + \phi_{21})^2]^{1/2}}{[(\phi_{11} - \phi_{22})^2 + (\phi_{12} + \phi_{21})^2]^{1/2}} \] (2.94)

The phase tensor is strictly not a decomposition technique as the impedance tensor free of distortions is not recovered, but the tensor expresses the changes in phase relationships with polarisation even in the 3D/3D general case (Table 2.2). This makes the method a very powerful imaging tool as was shown by Heise et al. (2006) in their trial-and-error forward modelling of a data set from the Taupo Volcanic Zone, New Zealand. In 2D and 3D/2D the strike of the structures is given by \( \alpha \).

### 2.13 Static Shift Correction

Several authors have implemented various theoretical and practical ways to correct for static shift in data. These can broadly be put into 2 categories: Curve fitting and Statistical methods (see for example Ogawa (2002)’s categories).

#### 2.13.1 Curve fitting methods

In this technique the distorted curve is shifted upward or downward to coincide with the undistorted curve, the latter determined in some practical or statistical manner. One of the ways to determine which direction to shift the curve is to conduct ground Time Domain Electromagnetic (TDEM) surveys (Stephen et al., 2003; Spitzer, 2001) and shift the responses to coincide with the TEM responses, since EM techniques like TDEM do not suffer from static shift (Simpson and Bahr, 2005).

Figure 2.29 shows an example of how DC resistivity was used to correct for static shift. At high frequencies the DRS response, which is from a DC resistivity survey, coincides with only one mode (TE/XY). The TM/YX mode is the one to be corrected as it has an observable DC shift. These techniques are only effective for correcting static shifts from near surface features (Simpson and Bahr, 2005; Simpson, 1998).
A rather different approach to correcting for static shift using short periods, such as

![Graph](image.png)

Figure 2.29: An example of curve fitting technique using DC electric method, after *Stephen et al. (2003)*. There is an observable DC shift in the YX mode as it is not coincident with the DC results at higher frequencies

in the example above, was proposed by Russian scientists (*Mikhlin, 1984*), described later by *Berdichevsky (1989)* and *Jiracek (1990)*, where long period curve correction was used. The idea was to use global average induction responses derived from a worldwide network of geomagnetic observatories to correct long period responses of clearly static shifted data. To this end, *Mikhlin (1984)* created a layered Earth model, the top layer of which had a resistive inclusion as shown in Figure 2.30. The synthetic response of this model is labelled $\rho_a$ and is scaled in amplitude relative to the response of the same model but without the inclusion, labelled $\rho_N$. The corrected curve $\rho_a'$ is obtained by shifting the $\rho_a$ curve to coincide with the long period branch of the magneto-variational responses (h-line). The disadvantage to this approach is the lack of resolution of the long period responses due to the sparse data coverage of the geomagnetic observatories.

*Jones (1988)* showed how curve shifting can be applied in a sedimentary environment
by treating a sedimentary layer with known modal characteristic resistivity in a parametric form, thus allowing shifting of TE and TM response modes to coincide with this resistivity value. This technique requires many observations to be made but can be applied successfully in existing data sets and does not require additional measurements.

2.13.2 Statistical methods

Various statistical techniques, mostly involving geometrical averaging of apparent resistivities, have been applied to the static shift problem (Berdichevsky and Dmitriev (1976)). Averaging techniques require many sites and detailed knowledge of the geological structure, such as in the Canadian Lithoprobe experiment (Jones (1988)) and the Electromagnetic – Array – Profiling (EMAP) (Torres-Verdin and Bostick (1992)). In this field set-up, there is continuous lateral sampling of the electric field and low pass filters are used to suppress static shifts effects (Simpson and Bahr (2005)), based on the assumption that secondary field effects sum to zero. Applying this technique to complex geological environments may not give desired results as this technique gives relative estimates of static shift.
2.13.3 Other dimensionality and directionality tools

The widely used telluric tensor decomposition techniques have presumptions about the Earth structure (i.e., 3D/2D or 3D/3D), and offer the most quantitative way to separate the effects of local heterogeneities from the three-dimensional regional effects.

Static shift remains a huge challenge in MT data analysis. How effective these techniques are depend on the presumptions about the dimensionality of the regional structure being sensed. It follows from the discussions above that in order to best handle distortions encountered in MT data one must consider the data collection and modelling procedures.

In an ideal case, one should look at imaging the 3D regional resistivity structure and correcting for local 3D structures that are not desired. This is a 3D/3D problem that some authors have tried to look at with variable success. In most of the techniques that solve the 3D/3D problem, many assumptions still have to be made and deviations from those renders the process inadequate. Lots of progress has been made over the years in trying to address the distortion problem and the distortion correction techniques that are available when complemented with good data collection techniques and processing methods, give an acceptable result. Experimenting with more than one technique of static shift correction is advisable so as to compare relative efficiency in dealing with different model types.

2.14 MAGNETOTELLURIC DATA MODELLING AND INVERSION

The next step after processing MT data is to take the distortion-free data and convert them to an image of resistivity variation with depth, from which geological inferences can be drawn. This process is accomplished by conducting forward and inversion modelling, either in 1D, 2D or 3D. The resulting model, found either by forward or inverse modelling processes, must have responses at the surface of the Earth that are consistent with observed responses.

The next section describes two modelling procedures that shows how Earth models showing resistivity variations with depth are developed.

2.14.1 Forward modelling

Forward modelling is an iterative trial-and-error process. The objective is for the MT practitioner to simulate the electromagnetic induction process by finding a geological
model that matches the observed MT responses. In other words, the user generates synthetic data, for example, using equation 2.32, if 1D a Earth model is desired, then compares the calculated responses with the observed ones. The input model is then modified where data are poorly fitted and the responses recomputed until a satisfactory fit is achieved. The input conductivity model will generally be discretised into cells or blocks of specific resistivity and then solving Maxwell’s equations using some form of approximation (e.g., finite element, finite difference or integral equation) to solve for EM fields everywhere in the mesh. The resulting synthetic responses, from which apparent resistivity and phases are calculated, can be compared with measured data and the RMS fit between them will indicate how close they are.

Simpson and Bahr (2005) showed that there are advantages to modelling data this way, for example deriving optimal data acquisition parameters when designing a field survey (i.e., the site spacing, periods required). Additionally, forward modelling responses might be compared with actual field data, if available, to test the confidence intervals of the data.

Forward modelling of MT data in 1D, 2D and 3D is now routine. Wannamaker et al. (1987), Mackie and Madden (1993d) and Pek (1997) presented 2D forward modelling codes based on finite-element, finite difference and anisotropic formulations respectively. However, finding a forward model that matches the observed responses reasonably well is a time consuming process. What is ideal is an automated process that uses the observed data as input and generates a geologically plausible (2D or 3D) model. This process is known as Inversion.

2.14.2 1D and 2D Inversion modelling

Inversion modelling involves taking the observed/recorded data and, through some automated process that minimises the misfit between the observed and forward responses, obtaining a reasonable model. The inversion problem is non-unique (i.e., multiple models can be obtained from the same MT data). A general inverse problem can be expressed as

\[ \vec{d} = G(\vec{m}) \]  

where data (impedance in the case of MT) is represented by \( \vec{d} = d^1, d^2, \ldots, d^N \), \( G \) is the forward operator (such as equation 2.32 for 1D modelling) and \( \vec{m} = m^1, m^2, \ldots, m^M \) are the model parameters (such as resistivity in MT case).

In MT, 1D and 2D inversion problem is no longer a theoretical problem (i.e., the
theory is well developed and put in practice). One (1D) and two (2D) inversion algorithms include those of Fischer and Quang (1981) (INVERS), Constable (1987); DeGrootHedlin (1990) (OCCAM 1D and 2D), Smith and Booker (1991) (Rapid Relaxation Inversion-RRI), Siripunvaraporn and Egbert (2000) (REBOCC) and Rodi and Mackie (2001) (RLM2DI/WinGLink).

Most 2D inversion schemes solve for minimum structure while others, principally that of Pek (1997), allow solving for anisotropic variations. The 2D inversion isotropic and anisotropic models presented in this thesis were derived using the code of Rodi and Mackie (2001), the forward part of which is described in Mackie et al. (1988) and is based on finite difference approximation of Maxwell’s equations. In addition the forward model is discretized using a rectangular mesh. Rodi and Mackie (2001)’s inversion is based on a non-linear conjugate gradient (NLCG) scheme to minimise a regularized objective function (Tikhonov regularization, (Tikhonov and Arsenin, 1977)).

2.14.3 3D Inversion

Three-dimensional (3D) inversion, although rapidly developing, is to large extent still in its infancy. In 3D Earth and in Cartesian coordinates, resistivity varies in all three spatial directions, \(x\), \(y\), \(z\).

There are two freely available 3D inversion codes: that of Siripunvaraporn et al. (2005a) (WSINV3DMT) and that of Egbert and Kelbert (2012) (Modular Electromagnetic-ModEM). The former code is by far the most widely used but it requires huge amounts of computer memory, as it is based on data-space regularization, so that an application to a normal PC is almost impossible. The code by Egbert and Kelbert (2012) is a huge improvement on the former due to its treatment of the Jacobian matrices and is well suited for huge datasets. For this reason, the 3D inversion modelling that is carried out in this thesis uses only the code of Egbert and Kelbert (2012) given the large datasets that are modelled.

The ModEM inversion scheme is, as the name implies, based on a modular design where the Jacobian and sensitivity matrices are factored into several components (i.e., data functionals, forward and adjoint solvers, model parameter mappings). The computation of the sensitivity matrix this way allows for faster computation times. Not only is ModEM suitable for large inversions (such as in our study) but also consistently resulted in models with low RMS and with geologically reasonable geometry of features. As such, we will only show the ModEM results here but the models we obtained with WSINV3DMT contained the same features. For an example of full im-
plementation of the ModEM algorithm, the reader is referred to Kelbert et al. (2012). As with the 2D models, various parameters were tested to obtain geologically-plausible models. The ModEM algorithm has provision to invert up to six combinations of impedance tensor elements (\(Z_{xx}, Z_{xy}, Z_{yy}, Z_{yx}\)) and vertical transfer functions (\(H_z\) tippers). Chapter 7 deals exclusively with modelling data in 3D using the ModEM code.

A question can be asked as to when is it appropriate to apply 3D inversion techniques? In geologically complex regions, 3D inversion model will render a reasonable model than 2D models. This is because no assumptions about dimensionality is made and all measured data, apparent resistivity, phases and tippers, can be modelled. Just like with 2D inversion models carried out in section 6, the input parameters including initial model, the number of periods, the quality of the data, the "smoothing parameters" and the regularization and the complexity of the study region are all important in obtaining a reasonable geological model.
The Southern African Magnetotelluric Experiment (SAMTEX) is a multi-national, multi-purpose scientific project that was conceived in the mid 1990s with the initial intention of determining the resistivity structure of the Kaapvaal craton. The main objective was to collect MT data along a NE-SW profile, called KAP line, starting from Cape province in the SW ending in northern Limpopo province of South Africa. The data collected from the KAP line was to coincide and compared with seismic data acquired as part of a much older seismic project, the Southern African Seismic Experiment (SASE).

In 2003 Dr. Rob Evans of the Woods Hole Oceanographic Institute (WHOI) submitted a proposal to the Continental Dynamics Programme of the National Science Foundation (NSF) to acquire MT data across the Kaapvaal craton. With funding granted, Phase 1 of data acquisition commenced in autumn of 2003 along the NE-SW KAP main line and along two orthogonal profiles, one of which crossing the gold-rich Witwatersrand basin.

The Dublin Institute for Advanced Studies, under the leadership of Prof. Alan Jones, joined the SAMTEX consortium and the project grew to three more data acquisition phases. Funding was secured by Prof. Jones from major international mineral exploration companies (i.e., De Beers, Rio Tinto and BHP Billiton SA Ltd), government research institutions (South African Department of Science and Technology, Council for Geoscience South Africa, Council for Scientific and Industrial Research, Botswana Geological Survey, Namibian Geological survey). Prof. Jones was also successful in acquiring additional funding from research grants from the National Science Foundation (EAR-0309584 and EAR-0455242) through the Continental Dynamics Program) and Science Foundation of Ireland (grant 05/RFP/ GEO001). A late partner in the SAMTEX consortium was ABB Sweden who joined the project after receiving...
a contract from the Namibian power utility, NAMPOWER, to map areas that might suitable for the placement of high-voltage power lines.

These funds led to SAMTEX project growing to a multinational project and by the end of Phase 4, in May 2008, MT data acquired at over 750 MT stations in an area in excess of a million square kilometers across three Southern Africa countries (Botswana, Namibia and South Africa) and at the time represented the largest known MT scientific experiment in the world (see map coverage on Figure 3.1). While the intention was to collect data in Zimbabwe, the logistical and security situation at the time led to the suspension of the field acquisition in that country.

Data quality was generally very high, especially in Namibia and Botswana, but was poor at some locations in South Africa, particularly close to the town of Kimberley and in the Witwatersrand Basin, due to the high amplitude electrical-noise generated by the DC power-supply to both the mines and railway lines.

My involvement in the SAMTEX project began in January 2008 during the fourth phase of the project. Dr. Mark Muller led the data acquisition at this stage and with the help of Claire Horan, Colin Hogg, Marion Miensopust, Pieter Share and Jan Schmoldt. In total, 120 sites were collected, with exact station number for each profile summarized in Table 3.1. Given the various organizations involved, the SAMTEX project had specific objectives.

---

**Academic objectives**

- Understanding the tectonic processes dominant during the early Earth by determining the structures and geometries of sub-cratonic lithospheric mantle beneath Southern Africa

- Compare and contrasts the results from Southern Africa with cratonic results from elsewhere and develop and test hypothesis of lithospheric formation and deformation during the Archean and post Archean.

**Industry objectives**

- Determine geometries of boundaries of cratons in Southern Africa

- Assess whether the MT method (an electromagnetic geophysical imaging technique) can be used for area selection for diamond exploration
3.2 SUMMARY OF SAMTEX RESULTS

With the above mentioned objectives it is important to summarise some key results obtained by analysing and modelling data acquired during the SAMTEX project. Most of the work has been presented in major international conferences by various people involved in the project (see section Publications for some examples). The DIAS website (www.dias.ie) lists these presentations in chronological order. The key findings are firstly summarized by tectonic region and then synthesized.

3.2.1 Southern African diamond prospectivity

Jones et al. (2009b) presented a electrical resistivity map (at 200 km depth) of Southern Africa and compared it to the location of kimberlites (Figure 3.2). This comparison
<table>
<thead>
<tr>
<th>Country</th>
<th>Profile</th>
<th>BBMT sites</th>
<th>LMT sites</th>
<th>AMT sites</th>
<th>Total number of sites</th>
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<td>4</td>
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<td>1</td>
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</tr>
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<td></td>
<td><strong>121</strong></td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>

Table 3.1: Summary of MT sites collected during Phase 4

suggested that diamondiferous kimberlites occurred on the margins of thick Archean cratons, which is a departure from the well known Clifford’s rule that says that kimberlites are emplaced on the centre of cratons. Furthermore, the younger Proterozoic lithosphere is relatively more conductive than the Archean cratons.

Images of electrical resistivity and electrical resistivity anisotropy at 100 km and 200 km, constructed through approximate one-dimensional methods, map resistive regions spatially correlated with the Kaapvaal, Zimbabwe and Angola Cratons, and more conductive regions spatially associated with the neighbouring mobile belts and the Rehoboth Terrane. Known diamondiferous kimberlites occur primarily on the boundaries between the resistive or isotropic regions and conductive or anisotropic regions.

Comparisons between the resistivity image maps and seismic velocities from models constructed through surface-wave and body-wave tomography show spatial correlations between high velocity regions that are resistive, and low velocity regions that are conductive. In particular, the electrical resistivity of the sub-continental lithospheric mantle (SCLM) of the Kaapvaal Craton is determined by its bulk parameters, so is controlled by a bulk matrix property, namely temperature, and to a lesser degree by iron content and composition, and is not controlled by contributions from intercon-
Figure 3.2: An image of the resistivity at 200 km. Also shown on the figure are kimberlite locations; red means known to be diamondiferous, green means known to be non-diamondiferous, and white means not defined or unknown, after Jones et al. (2009a)

connected conducting minor phases, such as graphite, sulphides, iron oxides, hydrous minerals, etc. This makes quantitative correlations between velocity and resistivity valid, and a robust regression between the two gives an approximate relationship of $V_s [\text{m/s}] = 0.045 \times \log(\text{resistivity [ohm m]}) + 4.5$.

3.2.2 Kaapvaal craton

Seismic and electrical anisotropy

Hamilton et al. (2006), also in Hamilton (2008), analysed the relationship between electrical and seismic anisotropy using the Phase 1-3 SAMTEX stations and the SASE stations. The results are summarised in Figure 3.3 where the most conductive directions (which are used as proxy for electrical anisotropy) are plotted and
compared against seismic shear-wave splitting results. Hamilton et al. (2006) found that there is no ubiquitous correlation between the seismic and MT anisotropy directions for crustal and mantle lithospheric depths. However there appears to be, at least qualitatively, a significantly better correlation between the MT asthenospheric results and shear wave splitting results in 3 regions analysed. This subtle similarity led to Hamilton et al. (2006) to propose a model shown in Figure 3.4 to explain the

Figure 3.3: Electrical more conductive directions (in red), scaled by maximum phase difference, for the lithospheric upper mantle, and the shear-wave splitting results (both high and low quality results plotted in green, but sites with no detectable splitting omitted). ZC: Zimbabwe craton, KC: Kaapvaal craton. LB: Limpopo belt. NN: Namaqua-Natal mobile belt. CFB: Cape Fold belt. KB: Kheis and Proterozoic fold and thrust belt. Blue line: N-S trending Colesberg Magnetic Lineament (CBL). Purple lines: E-W trending Thabazimbi Murchison Lineament (TML), as plotted by Silver et al. (2001)
origin of asthenospheric anisotropy. The proposed model has an anisotropic, but heterogeneous, lithosphere on the cratonic region, with better ordered structures and anisotropy in the lithosphere of mobile belts. The lithosphere is underlain by anisotropic asthenosphere, as well as a strongly anisotropic region below the thicker cratonic keels in the region. Asthenospheric mantle MT results appear to be better correlated with plate motion directions, as well as shear wave splitting fast axis directions promoting a wet asthenosphere, with strongly developed lattice preferred orientation.

Figure 3.4: A proposed model for the origin and structure of anisotropy in southern Africa. Taken from Hamilton (2008)

Electrical anisotropy

*Evans et al. (2011)* modelled the main KAP line and presented an anisotropic model of the Kaapvaal craton. Figure 3.5 shows the anisotropic models in 2 different directions, parallel and perpendicular to the profile. Within the central portions of the Kaapvaal craton are regions of resistive lithosphere about 230 km thick, in agreement with estimates from xenolith thermobarometry and seismic surface wave tomography, but thinner than inferred from seismic body wave tomography. The MT data are unable to discriminate between a completely dry or slightly “damp” (a few hundred parts per million of water) structure within the transitional region at the base of the lithosphere. However, the structure of the uppermost 150 km of lithosphere is consistent with enhanced, but still low, conductivities reported for hydrous olivine and orthopyroxene at levels of water reported for Kaapvaal xenoliths.
The electrical lithosphere around the Kimberley and Premier diamond mines is thinner than the maximum craton thickness found between Kimberley and Johannesburg/Pretoria. The mantle beneath the Bushveld Complex is highly conducting at depths around 60 km. Possible explanations for these high conductivities include graphite or sulphide and/or iron metals associated with the Bushveld magmatic event.

### 3.2.3 Rehoboth Terrane

*Muller et al. (2009)* presented a 2D electrical model of the Kaapvaal craton and the Rehoboth Terrane (KIM-NAM profile on Figure 3.6) and compared the lithospheric thickness. The 2D model reveal a heterogenous resistivity structure with thicker, resistive lithosphere beneath the Archean Kaapvaal compared to the relatively thinner less resistive Rehoboth Terrane. *Muller et al. (2009)* observed a clear relationship between electrical resistivity structure of the lithosphere and a tectonic stabilization age of the Rehoboth Terrane. Furthermore, a significant lithospheric thickness difference is observed between the Eastern and Western Kaapvaal. The present day thinner litho-
Figure 3.6: Electrical resistivity models for profile KIM-NAM (after Muller et al. (2009)) derived from 2-D smooth inversion of decomposed MT station responses for (a.) 25° E of N strike azimuth and (b.) 45° E of N azimuth. (c.) Estimates of the depth of penetration achieved at each individual station location for both the TE (red) and TM (blue) modes, and overlaid on the 25° inversion model. The models are entirely unconstrained (and should be ignored) in areas beneath the maximum depth of penetration. (d.) 2-D inversion RMS misfit error at each station. The surface extent of the geological terranes is shown in (a.) and abbreviations used as follows: Ghanzi-Chobe/Damara (DMB), Rehoboth (RBT), Western Kimberley Block (KBW), Eastern Kimberley Block (KBE) and Witwatersrand Block (WB). Black dashed lines in (a.) and (b.) indicate interpreted depth to base of lithosphere, where well constrained, and red diamonds indicate the depth to the base of the chemically depleted lithosphere as defined by Cr/Ca-in-pyrope barometry from kimberlitic concentrates.

Spheronic thickness in the Rehoboth Terrane accounts for the absence of diamondiferous kimberlites in the Rehoboth province.
3.2.4 Zimbabwe craton, Magondi Belt

*Miensopust et al.* (2011) presented a 2D model across the southern western Zimbabwe craton to map the extent of the craton margin. Figure 3.7 shows the 2D smooth inversion, the major feature being the thick, resistive Zimbabwe craton and less thicker Magondi belt. The Zimbabwe craton is characterized by thick (220 km) resistive lithosphere, consistent with geochemical and geothermal estimates from kimberlite samples of the nearby Orapa and Letlhakane pipes (175 km west of the profile). The lithospheric mantle of the Ghanzi-Chobe belt is resistive, but its lithosphere is only about 180 km thick. At crustal depths a northward dipping boundary between the Ghanzi-Chobe and the Magondi belts is identified, and two middle to lower crustal conductors are discovered in the Magondi belt. The crustal terrane boundary between

![Image of 2D smooth inversion model](image-url)
the Magondi and Ghanzi-Chobe belts is found to be located further to the north, and the southwestern boundary of the Zimbabwe craton might be further to the west, than previously inferred from the regional potential field data.

### 3.2.5 Limpopo belt

This work is published on Precambrian Research Journal and the paper is embedded in the Appendix section A.

## 3.3 SYNTHESIS OF THE SAMTEX RESULTS

There is no doubt that the SAMTEX project has achieved its objectives set out in the proposals. The mapping of the Southern African lithosphere has been unprecedented. As with any scientific experiment, the more data are collected, analysed and modelled, the more questions are raised. This has certainly been the case with the SAMTEX project. The results summarised above have shown that a lot of previously unknown information has been gained as a direct result of SAMTEX.

However, there are questions that still remain to be answered regarding the lithospheric evolution. The correlations between seismic and electrical anisotropy is not fully understood, i.e., 1) is the observed anisotropy a global phenomena?, 2) is it restricted to certain parts of the lithosphere and most importantly what are the causes of anisotropy. Hamilton et al. (2006) provided one possible cause for asthenospheric anisotropy observed in the Kaapvaal craton. The asthenosphere beneath the Kaapvaal craton was also found by Evans et al. (2011) to be more conducting in one direction than a perpendicular direction, implying anisotropy.

Before the work of Muller et al. (2009) the Rehoboth Terrane was largely unknown. The MT models for the Rehoboth showed conclusively that the Rehoboth is a much younger component of the composite Kalahari craton and, like the Kaapvaal craton, had its lithosphere modified by post-stabilization metasomatic events. This result, and the thinner lithosphere, is used to account for the absence of diamondiferous kimberlites in the region.

The SAMTEX industry partners, who look for minerals deposits such as kimberlites, understand better now the places to target for further exploration. The work of Jones et al. (2009b), which has global implications, showed that the areas to target are those on margins of cratons, or specifically on regions where there are resistivity gradients. In the same light (and it be shown later) Chapter 9 shows how MT data can be used to
investigate the potential of a sedimentary basin to store CO$_2$. MT is used as a targeting tool in this respect. The Karoo basin is believe to hold a significant amount of shale gas. However what is not quantified is specific sites where gas might be stored. The global environmental implications of CO$_2$ that is released into the atmosphere are known. Capturing and storing these harmful gases in subterranean reservoirs offers one technique in mitigating these effects. To this end MT is used to map places that are most suitable to storing carbon.

The Congo craton study is the main focus of this thesis. A lot of SAMTEX still remains to be modelled and analysed, particularly in key regions like the enigmatic Okwa terrane in central Botswana. Furthermore there remain some gaps that remains to be filled. The extension of acquisition into Zimbabwe will help us understand better the evolution of the Limpopo belt development.
4

GEOLOGY, TECTONICS AND GEOPHYSICS OF SOUTHERN AFRICA

4.1 SOUTHERN AFRICAN GEOLOGY

The African tectonic landscape is made up of some of the oldest continental materials on Earth (Figure 4.1). Southern Africa particularly boasts some of the world’s most spectacular geological features: the oldest rocks (e.g., Barberton greenstone belts), oldest and biggest known meteorite (Vredefort impact structure), largest igneous intrusion (Bushveld Complex), which holds the world’s largest platinum resources, some of the largest sedimentary basins (e.g., Witwatersrand and Transvaal basins) also with abundance deposits. In addition, some of the oldest plant, animal and hominid fossils have been unearthed in southern Africa; making this region very diverse, and for Earth science studies, a unique place that is without equal on the globe (McCarthy and Rubidge, 2006).

Perhaps the biggest advantage of southern African geology is that it transcends the entire geological time scale (see Figure 4.2). The Archean Kaapvaal craton located in South Africa, forms the central core of the composite Kalahari craton. Jacobs et al. (2008) defined the Kalahari craton to be composed of the Kaapvaal, Limpopo, Zimbabwe cratons and Rehoboth Terrane. We follow this definition in this thesis.

4.1.1 Kalahari craton

The Kalahari craton is a composite craton comprising several Archean and Neoproterozoic blocks all with distinct but related evolutionary paths. The Kalahari craton acted as a stable cratonic block with respect to the later tectonics of the Namaqua-Natal mobile belt to the south and the Pan African orogeny to the north. The Kalahari craton is separated from the greater Congo craton in the north by the northeast-southwest (NE-SW) Damara-Ghanzi-Chobe belts (herein called DGC); Damara in Namibia and Ghanzi-Chobe in Botswana.
The lithospheric architecture of Africa

4.1 Southern African Geology

Figure 4.2: Geologic time scale with linear time axis. Courtesy of www.geology.com.
The Kaapvaal Craton (see Figure 4.3) consists predominantly of granitoids with gneisses and narrow greenstone belts (Begg et al., 2009; Tankard et al., 1982; de Wit et al., 1992; Key and Ayres, 2000; James and Fouch, 2002; Poujol et al., 2003). Several terranes can be recognized on the basis of structural trends, the ages and distribution of greenstone belts, and granitoid ages, but much of the craton is covered by Upper Archean basins, and terrane boundaries therefore are based on geophysical data.

The Southeastern Terrane contains the oldest rocks, including 3.7-3.2 Ga gneisses, and the Mesoarchean Barberton greenstone belt experienced major 3.1-3.0 Ga magmatism. It is overlain by the 3.1-2.87 Ga Pongola Basin, the 3.07-2.7 Ga Witwatersrand Basin, and the 2.7-2.6 Ga Ventersdorp Basin.

The Central Terrane is a narrow, NE-trending block consisting of 3.34-3.0 Ga gneisses, metamorphosed at ca. 2.6 Ga. The Pietersburg Terrane consists of 2.9-2.8 Ga TTG gneisses and greenstone belts, intruded by 2.7-2.6 Ga granitoids. The western parts of
these three terranes are overlapped by 2.85-2.66 Ga supracrustal rocks, and then by 2.6-2.5 Ga Transvaal Basin rocks; the amalgamation of the eastern terranes occurred at 2.85 Ga.

The Western Terrane is elongated N-S. It contains greenstone belts dated to 3.0-2.78 Ga on a basement of 3.25-2.88 Ga gneisses and granitoids. The 2.75 Ga granitoids occur on either side of the Colesburg Lineament, which marks the suture between the Western Terrane and those to the east. The northern part of the Craton was intruded by the 2.06-2.05 Ga Bushveld Igneous Complex (Cawthorn et al., 2007).

The Limpopo belt, like the Damara belt, developed during collision of two landmasses, the Zimbabwe and Kaapvaal cratons, during the period 2.7-2.6 Ga. It contains 3.3-3.1 Ga gneisses affected by granulite-facies metamorphism and granitoid magmatism at 2.7-2.57 Ga (van Reenen et al., 2011; Kramers et al., 2011). The northern and southern marginal zones of the block are extensively deformed, and contain (possibly allochthonous) rocks correlative with the adjacent cratons. The peak of metamorphism was at ca. 2.56 Ga in the northern marginal zone, and at ca. 2.7 Ga in the southern zone. Metamorphism also may have affected the belt between 2.06 and 2.0 Ga, linked with events in the adjacent Kheis-Okwa-Magondi Belt, and the intrusion of the Bushveld complex. The reader is refered to the latest MT models attempting to explain the evolution of the Limpopo belt in Appendix A.

The Zimbabwe Craton contains a 3.5-2.8 Ga TTG complex, unconformably overlain by the 2.8-2.7 Ga Bulawayan Supergroup (komatiites, flood basalts, and sediments) and overprinted by ca. 2.6 Ga granitoids. The Great Dyke and its satellites cut the craton at 2.57 Ga; during Proterozoic time the block remained uplifted, and no sedimentary basins formed (Begg et al., 2009).

4.1.2 Damara orogen

The Damara belt is a 1000 km long and 500 km wide Pan African orogenic belt (Figure 4.4). The term “Pan African” refers to a 870 to 550 Ma tectono-thermal event during which mobile belts developed around Archean cratons, as part of an orogenic cycle that resulted from the amalgamation of continental domains. The Pan-African event therefore is used to describe tectonic, magmatic, and metamorphic activity of Neoproterozoic to earliest Palaeozoic age, especially for crust that was once part of Gondwana (Kroner and Mainz, 2004). The Pan-African orogenic cycle is a time equivalent with the Cadomian Orogeny in western and central Europe and the Baikalian in Asia.
The Damara orogen is a three branch system reflecting part of West Gondwana suture; a collisional triple junction between the Congo, Kalahari cratons in southern Africa and Rio de Plata craton in South America (Gray et al., 2008; Frimmel and Frank, 1998) (see Figures 4.5). The exact positioning of fold belts and cratonic segments is

Figure 4.4: Cross section view across the Damara belt modified from Gray et al. (2008)

Figure 4.5: Map of the supercontinent of Gondwana showing the location of the Damara Orogen (from Foster et al. (2009)).
Figure 4.6: Position of Neoproterozoic fold belts and magmatic arcs between the various cratonic blocks of southwestern Gondwana. The timing of basin opening and closure in each belt is indicated. Note that the cratonic affinity of the Paranapanema Block is unresolved and shown here only tentatively. CDPT Cuchilla Dionsio-Pelotas Terrane, CFT Cabo Frío Terrane, CT Coastal Terrane, CUT Curitiba Terrane, LAT Luis Alves Terrane, MT Marmora Terrane, SM/OT Serra do Mar-Oriental Terrane, PSZ Purros Shear Zone, SBSZ Sierra Ballena Shear Zone, CSZ Colenso Shear Zone; modified from Frimmel et al. (2010)

The DGC is subdivided into three zones based on distinct metamorphic, structural and geological patterns (see Figure 4.4): the Northern, Central and Southern...
Zones. These sequences are underlain by a mosaic of reworked basement terranes. We discuss each of these zones separately below.

4.1.2.1 Archean to Mesoproterozoic basement \( \geq 1.5 \text{ Ga} \)

The granites and dioritic ortho-gneisses of the Andib Terrane, which are located in the Kaoko belt, are the oldest (2.5-2.6 Ga) basement lithologies in the Damara orogenic system (Seth et al., 1998; Goscombe et al., 2003). The orthogneisses of the Epupa Complex are dated at 2.1 Ga and appear to have a younger Mesoproterozoic component (Miller, 1983a). The Kamanjab Inlier comprises 1.5 Ga granitoids and metarmorphic rocks and is thought to represent the leading edge of the Congo craton (Burger et al., 1976; Tegtmeyer and Kröner, 1985; Goscombe et al., 2003; Gray et al., 2008).

The 1.3 Ga Kunene Complex (KC) of NW Namibia and SW Angola is composed mainly of anorthosites (Ashwal and Twist, 1994; Morais et al., 1998; Mayer et al., 2004), but some authors suggest it may be a layered mafic intrusion (Silva, 1990). The Kunene Complex, like the Kamanjab Inlier, is believed to represent part of the Congo craton basement (Gray et al., 2008). Among the anorthosite massifs, the Kunene Complex is one of the largest in the world (15,000 km²), possibly second in size only to the Lac-Saint-Jean Complex of Canada (Mayer et al., 2004). Isotope studies (\(^{143}\)Nd/\(^{144}\)Nd) conducted on the KC in the Complex, as well as the high anorthite content have been invoked to suggest that the parent magma of KC was derived from a mantle-source, and thus that the emplacement of KC anorthosites reflect an episode of continental accretion rather than a within-continent differentiation (Mayer et al., 2004). Furthermore, the initial \(^{143}\)Nd/\(^{144}\)Nd of the Lufinda dyke (0.510766 ± 24) implies instead a minor addition of crustal material into the parental magma. Collectively, the emplacement of the Kunene Complex requires an extensional setting and a significant thermal anomaly at the margin of the Congo Craton during the early Kibaran cycle (Mayer et al., 2004).

4.1.2.2 Northern Zone

The western Northern Zone (Ugab domain) is influenced by transition between basinal turbidite facies and platform carbonates (Miller, 1983a; Passchier and Paciullo, 2002). The Northern Zone underwent metamorphism from ca. 540 to 530 Ma (Goscombe, 2004) and it represents the junction between Kaoko and Inland Branches, and is intruded by granites (Coward, 1983; Maloof, 2000; Passchier and Paciullo, 2002; Goscombe, 2004). The Northern Zone is characterized by the occurrence of Pre-Damara basement rocks (Welwitschia and Kamanjab inliers) that belongs to the Proterozoic Huab Gneiss domain. These inliers are flanked by Damaran rocks, comprising the Nosib, Swakop...
and Mulden group series consisting of pebbly feldspathic quartzites, carbonates and fine clastic sedimentary rocks. Alkaline and peralkaline rhyolite ash-flow tuffs dominate but other felsic extrusive rocks and mafic to intermediate intrusive rocks also occur. The great thickness of these lavas and tuffs (up to 6 km; (Miller, 1983a)) and their localized occurrence at the northern margin of the orogen seem to indicate that emplacement of these igneous rocks was associated with crustal extension. Jung and Mezger (2000) suggested that this igneous activity coincides with continental breakup at the southern margin of the Congo craton.

4.1.2.3 Central Zone

The granite-dominated Central Zone, metamorphosed under high-T/low-P conditions between 538 and 505 Ma, is floored by an attenuated Congo-like craton, with dome and basin fold interference (e.g., Jacob and Bunting (1983); Coward (1983); Miller (1983b); Kroner (1984); Kisters and Neumaier (2004)), and possible extensional structures (e.g., Oliver (1994)). The Central Zone has two phases of metamorphism at 538 to 516 Ma and 511 to 505 Ma with granitic magmatism between ca. 560 Ma and ca. 470 Ma (Jacob and Armstrong, 2000; Jung Hee Suh and Mezger, 2003).

The thicker basinal sequence has undergone intermediate-T/intermediate-P (Barrovian series) metamorphism and is now part of the 100 km wide Southern or Khomas Zone (Kasch, 1983a; Kukla, 1993). This zone consists of homoclinaly north dipping Kuiseb Schist with transposed foliation and schistosity (Miller, 1983b; Kukla and Charlesworth, 1988; Kukla and Stanistreet, 1991), which represents a major shear zone interface transitional into the basement-cored fold and thrust nappes of the Southern Margin Zone (Miller, 1983b; Coward, 1983). Based on U-Pb ages for metamorphic monazite, Kukla (1993) suggests an age of 525-515 Ma for peak metamorphism in the Kuiseb Schist; a lower limit is given by the intrusion of the 505 Ma post-tectonic Donkerhuk granite (Kukla, 1993).

The metamorphic pattern across the Damara Orogen was summarised by Gray et al. (2006) and is shown in Figure 4.7. This figure shows the space-time diagram of complex magmatism and deformational patterns in the Damara orogen. In the Central Zone, the metamorphic grade increases from east to west reaching high-grade conditions with local partial melting in the coastal area (Goscombe et al., 2003; Goscombe, 2004). These high-grade conditions culminated in low pressure high-temperature granulite-facies conditions with the temperature estimated to have been between 680 deg/C and 750 deg/C at 5 to 6 kbar (Jung et al., 1998; Jung Hee Suh and Mezger, 2003). The timing of orogenic events and the age of the peak of metamorphism within the Central Zone of the Damara orogen is constrained by combined U-Pb monazite
Figure 4.7: Summary time-space diagram for the Damara Orogen based on $^{40}$Ar/$^{39}$Ar data from Gray et al. (2006). The plot encapsulates the complexity of magmatic, metamorphic, deformational and cooling patterns across the orogen as part of Pan-African orogenesis. WKZ: Western Kaoko Zone; CKZ: Central Kaoko Zone; EKZ: Eastern Kaoko Zone; TPMZ: Three Palms Mylonite Zone; VMZ: Village Mylonite Zone; PMZ: Purros Mylonite Zone; AMZ: Ahub Mylonite Zone; ST: Sesfontein Thrust; GMZ: Guantegab Mylonite Zone; OmL: Omaruru Lineament (Shear Zone); OkL: Okahandja Lineament (Shear Zone).
ages and Sm-Nd garnet-whole rock ages indicating that the time span of high-grade metamorphism ranges from ca. 540 Ma to 470 Ma with an age of ca. 510 Ma for the (main) peak of regional metamorphism (Jung and Mezger, 2000; Jung Hee Suh and Mezger, 2003).

4.1.2.4 Southern Zone

The southern zone of the Damara belt is about 100 km wide and is defined by a wide zone of intense, north-dipping, shear-dominated transposed fabrics and basement-cored fold-nappes bordering the Kalahari Craton (Southern Margin Zone). The Southern Zone underwent peak temperatures of c. 600 deg/C and pressures of 10 kbar (Kasch, 1983a) (see also Figure 4.7). Shear bands, developed in Kuiseb Formation schist and units of the Southern Margin Zone indicate north-over-south movement in a north-south transport direction.

Gray et al. (2008) interpreted the major south-directed bulk shear strain deformation was responsible for crustal-scale under-thrusting of the Kalahari Craton northwards, as well as continued thrusting and crustal thickening along the margins of the orogen. The associated crustal thickening and burial along this margin led to the Barrovian metamorphism and significant magmatic underplating related to extension in the lower part of the overriding plate, led to marked magmatism and younger, high-T/low-P metamorphism in the Central Zone (Gray et al., 2006, 2007, 2008).

4.1.2.5 Major fault structures in the Damara

Ground geological and magnetic mapping has revealed the presence of complex crustal faults in the Damara belt (see Figure 4.8). The major prominent structures that has tectonic significance are the Autseib Lineament (AL), Omaruru Lineament (OML) and Okahanja Lineament (OML). The AL delineates the boundary between the Central Zone and the Northern Zone and has one of the most prominent long wavelength anomalies, striking NE-SW (Miller, 1983c; Corner, 2008). Geophysical mapping by Eberle et al. (1995) suggested that the AL has a northward dip. Cretaceous intrusives rocks appear to be offset along the AL, which means it was reactivated post Damara sedimentation (Clemson et al., 1999).

The OML is a NE-SW trending fault zone between the lower (magnetically quiet) and upper (magnetically anomalous) Damara belt sequences. The OML has a corresponding gravity high signature (Corner, 2008). The offset of Triassic sedimentary rocks and the truncation of units in the Cretaceous Erongo complex (Holzförster et al., 1999) and the differences in paleo-temperatures (Raab et al., 2002) are thought to be evidence of
Figure 4.8: Major and minor fault structures in the Damara belt, mapped geophysically and through field mapping, overlain on the regional aeromagnetic map of Namibia. AL: Autseib Lineament, OML: Omaruru Lineament, OKL: Okahanja Lineament, MA=Matchless Amphibolite, KI=Kamanjab Inlier (map modified from Corner (2008), Singletary et al. (2003) and Begg et al. (2009))

Mesozoic and Cretaceous reactivation along the OML, with estimated 2 km vertical displacement.

The OKL separates the Southern Zone from the Central Zone and also has a magnetic signature. Down-faulting and uplift movement in the SE and NW respectively along the OML has exposed pre-Damara sequences (Miller, 1983c; Corner, 2008).

4.1.3 Congo craton

The Congo craton, which occupies a large part of central and southern Africa, comprises several Archean shields, the southern part comprising the Angolan shield, which consists of granite-gneisses and metasediments, overlain by Paleoproterozoic cover (Begg et al., 2009). Thick Kalahari sedimentary cover leaves the southern Congo
Archean basement rocks are exposed in four shields located around the edges of the Congo Craton; marked differences in geological history among these shields suggest they represent separate terranes, but the terrane boundaries are not defined.

The Gabon-Cameroon Shield in the NW corner of the craton contains 3.2-2.99 Ga migmatites, gneisses, and remnants of greenstone belts, cut by 2.97-2.54 Ga granites (Cahen et al., 1984). The 2.15-2.05 Ga supracrustal cover rocks (Fransevillian Formation) are cut by 2.04-1.92 Ga granites (Tchameni and Pouclet, 2000).

The Bomu-Kibalian Shield (NE edge) contains 3.42 Ga tonalitic gneisses and 3.0-2.6 Ga greenstone belts, as well as metamorphosed 2.51-2.46 Ga granites and the Kasai Shield (SE edge) is a heterogeneous 3.01 Ga granulite complex, metamorphosed at 2.9-2.6 Ga. It is overlain by metamorphosed supracrustal rocks, which were cut by granite plutons at ca. 2.1 Ga (Goodwin, 1996).

The Angolan Shield (SW Angola) consists of gneisses, metasediments, and a 2.82 Ga gabbro-charnockite complex and the basement rocks were metamorphosed at 2.8-2.7 Ga and intruded by granites at 2.6 Ga (Carvalho et al., 2000). A Paleoproterozoic (2.15-2.1 Ga) cover sequence is cut by 2.1-2.0 Ga Eburnean granitoids.

4.2 Tectonic Models Proposed for Damara Belt Evolution

There are two main tectonic models that have been proposed for the development of the Damara belt:

1. The **Ensialic or modified aulocogen model**, and

2. The **Subduction model**

There is still a lack of consensus on the preferred model. The next section describes the technical merits of these two models.

4.2.1 Ensialic or modified aulocogen model

Aulocogens, first noticed by Shatski (1946), are defined as long troughs, containing accumulations of sediments, extending into continental cratons from fold belts (Burke, 1977). In particular they have the following properties (Burke, 1977):

1. extension of orogen far into a craton with gradual change in size,
2. a thick gently folded sedimentary sequence,

3. location of the junction between the aulocogen and the orogen at a deflection in the fold belt,

4. the occurrence of horst-like features within an aulocogen that influenced sedimentation,

5. an early history of the aulocogen as a fault-bounded graben and a later history as a broader superimposed basin itself sometimes involved in later faulting

Aulacogens extend from their junction with the associated geosyncline several hundred kilometres into the platforms, and are on the whole wider than normal continental rifts (e.g., Rhine graben, East African rifts) and are filled with sediments and volcanics of geosynclinal thicknesses (8-12 km) (Barnes and Sawyer, 1980). An aulocogen model is characterized by a sequence of three main events (Hoffmann, Dewey and Burke, 1974; Martin and Porada, 1977a), shown in Figure 4.9:

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Figure 4.9: Schematic cross-sections showing evolution of a typical aulacogen (adapted from Hoffmann, Dewey and Burke (1974). Stages (1) and (2) incorporate features observed in the Damara orogen. After stage (2) the evolution of the Damara orogen deviates from this model.
(1) **Graben stage**: A first stage of faulting with associated volcanism produces a graben filled with several kilometres of coarse clastic sediments topped by mudstones and carbonate rocks.

(2) **Downwarping stage**: Volcanism and renewed subsidence now also engulf the margins of the former graben. This wider basin is filled with quartzite, mudstone and turbidite that pass in the shallower parts of the basin into stromatolitic carbonate-rock-rich shelf deposits. Calc-turbidites are deposited in the deeper part. A gigantic olistostrome is regarded as the beginning of a molasse phase consisting of red mudstone, siltstone and sandstone with halite and gypsum casts. The olistostrome was intruded by large laccoliths of quartz diorite to granodiorite.

(3) **Postgeosynclinal stage**: Renewed faulting, tilting, slight compression, erosion and deposition of conglomerate. The latter occurred simultaneously with transcurrent faulting. However, given the amount of folding and the intense metamorphism observed in the Damara belt, it seems unlikely that a simple aulacogen model, starting with rifting, igneous activity and subsidence can account for the whole Damara orogeny. Furthermore, there is no evidence of subsidence in the Damara belt. Therefore Martin and Porada (1977a), Martin and Porada (1977b), Kröner (1977) and Trompette (1997) proposed a modified aulacogen model for the Damara belt (see Figure 4.10).

According to this model, the geosynclinal development began on a sialic crust with the formation of three graben systems (Figure 4.10A) which evolved toward widely spaced aulacogens. The aulacogens are thought to have started their development as keystone grabens above rising asthenospheric diapirs. The rifts were filled with coarsely clastic sediments, and the rifting was in some areas accompanied by alkalic, rhyolitic to basaltic igneous activity. During this stage a considerable thinning of the crust was caused by fault-block rotation, erosion of the elevated rift shoulders and ductile flow in the lower part of the crust.

During the next stage cooling of the diapir and/or movement of the whole plate away from the diapir led to a subsidence of geosynclinal dimensions and the filling of this basin with the sediments and subordinate volcanics of the Hakos and Khomas Subgroups.

The third stage which initiated the orogeny is attributed to the transformation of basaltic rocks, intruded during the diapiric phase into the lower crust and the transition zone between crust and mantle, into eclogite. The resulting gravitational instability (Figure 4.10B) between the lithospheric upper mantle layer and the as-
thenosphere is thought to have caused this heavy slab to become detached from the lighter sialic crust and to sink into the asthenosphere with a concomitant ascent of hot asthenosphere material to the base of the crust. The rise in temperature caused high-grade metamorphism and anatectic mobilisation (Figure 4.10C) in the geosynclinal sediment pile and the sialic pre-Damara basement and thereby initiated the paroxysmal phase of the orogeny.

The development of the Damara fold belt occupied a time-span of about 300 m.y. In developing the ensialic tectonic model at the time, Martin and Porada (1977a) (also in Martin and Porada (1977b)) emphasised the following observations:
4.2 TECTONIC MODELS PROPOSED FOR DAMARA BELT EVOLUTION

4.2.1.1 Evidence for ensialic/modified aulocogen model

- the absence of ophiolitic rocks,
- the lack of MORB signature,
- the absence of a suture zone separating the two cratons
- the presence of older crystalline ensialic basement
- the unchanged position of the cratonic segments in the Pan-African times, implying that two plates were not rotated relative to one another at any stage.

The Matchless Amphibolite is believed to represent part of the older oceanic crust and has some MORB signature (Martin and Porada, 1977a). Barnes and Sawyer (1980) argued that the ensialic or modified aulocogen model cannot be applied for the Damara belt for the following reasons. Firstly, the linear shear zones exhibited a complex asymmetrical structural pattern with up to 80% shortening across the belt. Secondly, the markedly asymmetric metamorphic pattern broadly follows the structural pattern forming two distinct, parallel metamorphic belts of relatively high (northern belt) and low (southern belt) geothermal gradients, respectively. Abundant granitic intrusions occur in the high-grade metamorphic belt. Thirdly, the evolution of the Damara igneous rocks; the early (Nosib) igneous rocks are alkali; mid-Damara (Matchless Member) amphibolites resemble oceanic-floor basalts.

As a result, Barnes and Sawyer (1980) argued that the depleted upper-mantle material representing oceanic lithosphere was tectonically emplaced into the Damara metasediments during early tectonism. An extensive calc-alkali suite (the Salem Suite) intruded the high-grade metamorphic belt during a long period spanning most of the Damara tectonism. Therefore a model invoking the formation of alkali rocks, followed by the development of oceanic crust, initiation of north-westward subduction and ocean closure terminating in continental collision was considered to explain the major features.

Kukla and Stanistreet (1991) refuted the ensialic models based on sub-marine fan type sedimentary patterns in the Khomas trough and the structural features that exhibit compressional regime. These were all thought to indicate that the Damara meta-sedimentary sequence was deposited within imbricate thrust sheets which were floored by thick pelitic units. Kukla and Stanistreet (1991) also argued that the Matchless Amphibolite and the associated pelagic sedimentary rocks represented a upper parts of an oceanic crustal sequence, thereby interpreting the sediments to be
4.2 TECTONIC MODELS PROPOSED FOR DAMARA BELT EVOLUTION

4.2.2 Subduction model

This is the most popular model favoured for the development of the DGC. The Damara belt is thought to have developed from the collision of the Kalahari craton to the south with the greater Congo craton to the north (Coward, 1981; Miller, 1983c; Goscombe et al., 2003; Gray et al., 2007, 2008). In this model the DGC appears deposited in a Late Proterozoic accretionary prism that developed during convergence of Kalahari and Congo cratons.

Figure 4.11: Tectonic evolution of the Damara belt, Congo and Kalahari craton, highlighting a full Wilson Cycle processes (modified from Gray et al. (2007))

to have undergone a full Wilson Cycle (see Figure 4.11). Rifting between the Congo
and Kalahari cratons has been inferred to develop along three northeast-trending intra-continental rifts between ca. 840 Ma and ca. 730 Ma (Miller, 1983b), although dating of Northern Zone rift-related rhyolites has provided ages 746 Ma and 747 Ma (Hoffman et al., 1996). Synrifting silicic volcanism in the Central Zone took place at ca. 750 Ma (de Kock et al., 2000). Eventual development of an oceanic spreading center and growth of oceanic crust at ca. 700 Ma in the southern part led to development of the Khomas ocean basin (Miller, 1983b).

There are conflicting tectonic hypotheses for closure of the Khomas basin (Porada, 1979, 1983; Martin and Porada, 1977a). The Khomas basin has been depicted either as an intracontinental basin between the Congo and Kalahari cratons with turbidites sitting on attenuated continental lithosphere (Porada, 1979, 1983; Hawkesworth, 1983), or as an ocean basin floored in part by oceanic lithosphere that closed by subduction beneath the leading edge of the Congo craton to form an Andean-type margin with development of voluminous magmatism (Barnes and Sawyer, 1980; Miller, 1983b; Kasch, 1983b). Basement-cored domal culminations in the Central Zone (Jacob and Bunting, 1983; Puhan, 1983; Kroner, 1984; Oliver, 1994; Kisters and Neumaier, 2004) and basement-type isotopic signatures in some granites require the presence of attenuated Congo continental crust in this part of the Khomas ocean basin, probably not unlike the present-day Japan Sea, which has a complex distribution of attenuated continental crust, rifted crust, and oceanic crust (Tamaki, 1995).

Closure of the former Khomas ocean basin involved high-angle convergence with overthrusting at the margins to give a doubly vergent orogen. Apart from the laterally continuous, shear zone-hosted Matchless amphibolite there is no classic suture. This unit consists of intensely deformed basalt, pillow basalt, chert, and gabbro, with tholeiitic geochemistry (Schmidt and Wedepohl, 1983), as well as dispersed serpentinite bodies (boudins). The distinct metamorphic zonation of the Inland Branch, with Barrovian metamorphism on the orogen flanks, reflects structural thickening at the craton margins, while Andean- or Cordilleran-style high-T/low-P metamorphic conditions in the Central Zone and accretionary prism-like features of the Southern Zone Kuiseb Schist (Kukla and Stanistreet, 1991) reflect subduction beneath this zone (cf. Maloof (2000)).

Arguments against subduction-related closure of the Khomas ocean basin have also been based on granite geochemistry (Hawkesworth, 1983), as well as the lack of blueschists and eclogites or typical subduction zone low-T/high-P metamorphism (Kroner, 1982). Furthermore the basalt chemistry has been used to argue for an ensialic origin with basalts formed on thin crust (Schmidt and Wedepohl, 1983), but see Kasch
4.3 GEOLOGICAL AND GEOPHYSICAL CONSTRAINTS ON PROPOSED MODELS

(1983a) for a discussion. The dominance of S-type granites and the absence of clearly mantle-like isotopic signatures even in mafic I-type granitoids have also been invoked as argument against subduction (McDermott and Hawkesworth, 1990). In addition the positions of the respective cratons through time, the sizes of the ocean basins between them and the positions and directions of subduction zones that closed the ocean basins are still not well known (Gray et al., 2006).

In summary, evidence for subduction-related development of the Damara belt is summarized below.

4.2.2.1 Evidence of subduction:

• MORB-type geochemistry of the Matchless Amphibolite belt (Gray et al., 2008), possibly representing relict oceanic lithosphere,

• The distinct metamorphic patterns, Barrovian on orogen flanks and Andean- or Cordilleran-style high-T/low-P in the Central Zone, reflecting subduction beneath this zone.

• The kinematics of the Southern Zone schists, combined with geochemistry of the more primitive Central Zone diorites and syenites supports northward subduction of the Khomas Ocean lithosphere beneath the attenuated leading edge of the Congo Craton,

4.3 GEOLOGICAL AND GEOPHYSICAL CONSTRAINTS ON PROPOSED MODELS

4.3.1 Paleomagnetism

The have been very few studies of paleomagnetic studies in the Damara belt. This paucity of data has in many ways led to the contrasting models proposed above. McWilliams and Kröner (1981) presented paleomagnetic results from the Nosib, Otavi, and Mulden groups of the Damara Supergroup, which showed that no great relative movements have occurred between the Congo and Kalahari cratons during the interval of Pan-African tectonism in the Damara belt, effectively ruling out the continental collision that was preceded by large cratonic separation and closure of Khomas ocean. This led to McWilliams and Kröner (1981) to suggest an evolutionary model for the Damara belt which involves rifting, heating, and stretching of the lithosphere underneath the Damara belt, followed by delamination of the subcrustal lithosphere.

Weil et al. (1998) incorporated McWilliams and Kröner (1981)'s data and presented
a synthesis of paleomagnetic data from all over the world, in an attempt to reconstruct the positions of continental blocks during Rodinia times. These results are summarised in Figure 4.12. The results of Weil et al. (1998) suggest that the Congo and Sao Francisco cratons were connected throughout the 1100-800 Ma intervals (Figure 4.12A and B). Furthermore, the paleopoles, a paleogeographical position for the 1100-950 Ma interval with the Kalahari rotated 35° counterclockwise away from the Congo craton (Figure 4.12C and D). This reconstruction does not disagree with either the modified aulocogen and subduction models above but does support the existence of a rift structure that created an embayment of oceanic floor in the Khomas basin of the Damara belt (Porada, 1989).

4.3.2 Thermal profiles

The Neoproterozoic thermal and compositional structure beneath the Damara belt is difficult to discern due to the multiple tectono-magmatic events that have occurred since stabilization. Whitehead et al. (2002) constructed a thermal profile from mantle peridotite xenoliths discovered in a Cretaceous nepheline plug in the Damara Belt (Figure 4.13). In this study, Whitehead et al. (2002) observed that the geothermal gradient beneath the Damara belt was substantially hotter (90 mW/m²) than that beneath Rehoboth (45 mW/m²) and was attributed to the post collisional tectonothermal events associated with regional Cretaceous magmatism. Furthermore, the inferred mantle stratigraphy was interpreted to represent the juxtapositioning of relatively fertile Phanerozoic-type mantle beneath refractory peridotite. This mantle stratigraphy was proposed to reflect the rifting phase in the Damara belt in the Cretaceous time, perhaps from plume related melts.

4.4 Previous geophysical studies

4.4.1 Resistivity studies in Damara belt

Previous electrical and electromagnetic studies of the DGC include the late 1970s and early 1980s magneto-variational and Schlumberger soundings by de Beer et al. (1975), de Beer et al. (1976), de Beer et al. (1982) and Van Zijl and de Beer (1983). These studies (Figure 4.14) were the first to discover the along-strike crustal conductor within the DGC that extends from Namibia into Botswana. Serpentinized lower crust was, at the time, thought to be the cause of the elevated conductivity.

In 1998 and 1999 the GeoForschungsZentrum (GFZ) conducted a crustal-scale study (Figure 4.15) along a 200 km-long profile with a site spacing of 4 to 12 km and a
focused 3D array of 60 sites with a site spacing of 500 m to 2 km across the Waterberg Fault-Omaruru Lineament (Ritter et al., 2003b). In addition to mapping the middle to
lower crustal DGC conductor, Ritter et al. (2003b) imaged a very resistive upper crust (granites) and two sub-vertical conductors, the Autseib and Omaruru Lineaments. The authors invoked graphite concentration to explain the conductivity along upper crustal lineaments and middle to lower crustal anomaly, the latter being along shear deep-seated shear planes. In a related study, Weckmann et al. (2003) used MT phases to reveal a conductive ring structure in the shallow crust, with crustal-scale electrical anisotropy and an elongated conductor running sub-parallel to the Waterberg Fault-Omaruru Lineament.

Muller et al. (2009) presented a lithospheric-scale resistivity study of the Rehoboth terrane, partially extending into the DGC belt, which showed significant lithospheric thickness, electrical and thermal variations from older to younger terranes. Muller et al. (2009) found the lithospheric thickness of the DGC to be approximately 160 km.
Further east on the eastern border of Botswana, Miensopust et al. (2011) mapped the Ghanzi-Chobe belt having resistive 180 km thick lithospheric mantle.

### 4.4.2 Seismic, magnetic and thermal results

#### 4.4.2.1 Seismic tomography

Using a surface wave dataset with nearly 12,000 paths in the African region, Fishwick (2010) derived a tomographic models of central and southern Africa, indicating varying velocity structure beneath the cratonic regions. Estimates of the LAB depths from around 150-180 km were found for the Damara belt (Figure 4.16). The Kalahari craton had an LAB depth of approximately 200 km and the Congo craton returned LAB depths in the range 225-250 km. At a broad-scale these depth estimates are compatible with geothermometry from kimberlite xenoliths.
Figure 4.15: Two-dimensional resistivity models of the crust derived from inversion of (a) TM, (b) TE, and (c) joint TM+TE mode magnetotelluric data. The main features of the resistivity models are the generally high conductivity of the mid-crust in the central part of the profile and the narrow, subvertical conductivity anomalies attributed to basement shear zones. AF: Autseib Fault, WF/OL: Waterberg Fault/Omaruru Lineament. Study from (Ritter et al., 2003a)

4.4.2.2 S-reciever functions

Hansen et al. (2009) presented a study of S-wave receiver functions (SRF), estimating the Moho depth to be 35 km, for the TSUM station (see Figure 4.17). Stations LSZ and TSUM are located in the Damara belt. The LAB conversions for both LSZ and TSUM correspond to 142 and 180 km depths respectively, however, in both cases, this signal is on the edge of the boot-strap error limits and is not well resolved.

4.4.2.3 Magnetic mapping

Magnetic maps, derived from measuring the variations in the Earth’s magnetic field, provide information about the concentration of magnetic mineral (e.g., magnetite, sulphides, oxides) and the crustal geological structure. A regional magnetic map of Southern Africa is shown in Figure 4.18. This map has been used as a first pass
Figure 4.16: Estimates of lithospheric thickness based on the tomographic models, calculated using the empirical parametrization of Priestley and McKenzie (2006). The Damara belt ranges between 160 to 180 km and it is over 200 km for the Congo and Kalahari cratons.

delineation of regional tectonic features. Similar magnetic patterns such as observed in the Rehoboth Terrane, have provided guidance to where the SAMTEX stations were placed. However, given the ambiguity of the magnetic patterns and the fact that they inherently map shallow structures, this map should be used as a guide.

4.4.2.4 Thermal structure

Heat flow measurements in Namibia and Botswana are surprisingly sparse (see Figure 4.19). The only data available is that of Ballard et al. (1987) (and references therein) which show data from 25 sites in Botswana and Namibia. A gridded map of these heat flow was produced and is shown in Figure 4.20. These data shows the Kaapvaal craton having a mean value of 47 mW.m² while the Damara belt has an average heat flow of approximately 66 mW.m². Ballard et al. (1987) recognized the insufficient data points but argued that a correlation could be made between the high heat flow and younger mobile belts. This observation has since been confirmed from MT data studies of Muller et al. (2009) who observed relatively low heat flow in Archean cratons (e.g., Kaapvaal) compared to Proterozoic terranes (e.g., Rehoboth or...
Figure 4.17: Stacked seismic receiver functions (SRF) at two stations located in the Damara belt. The black line shows the mean stack while the gray shaded areas indicate the 2σ bootstrap error bounds. Major converted phases, LAB and Moho, are labeled. The black arrows indicate LAB depths reported by Kumar et al. (2007). A similar conclusion was made by Whitehead et al. (2002) (see Figure 4.13).
Figure 4.18: Regional magnetic map of Southern Africa, showing the SAMTEX stations (black stations are the focus of this thesis, white stations are the other SAMTEX stations). This data is courtesy of the Council for Geoscience South Africa.
Figure 4.19: (a) Map showing the location of heat flow sites (dark circles are from study of Ballard et al. (1987)). (b) Heat flow versus distance from the cratonic margin from the study of Ballard et al. (1987) and the references therein (solid and open circles).

Figure 4.20: Heat flow map of northern Namibia, derived from data of Ballard et al. (1987) and the references therein.
Part II

INTERPRETATION OF THE 2D AND 3D MODELS
5.1 DATA IMAGING

The SAMTEX database has a wealth of information (Figure 5.1). In choosing the profiles to be modelled and interpreted, the motivation was to work on regions that are less well-known and could add to the knowledge of the tectonics in the region. The regions that are north of the Kaapvaal Craton, i.e., Damara belt and greater

![Diagram of Southern Africa showing SAMTEX stations and profiles.](image)

Figure 5.1: SAMTEX stations overlain on the regional magnetic map of Southern Africa. The red dots indicate the locations of the profiles (ETO-KIM, DMB, NEN, OKA-WIN, RAK-CPV and WIN) that are modelled in this thesis. The blue dots show the location of the KIM-NAM profile of Muller et al. (2009)’s study. The magnetic data are courtesy of the Council for Geopscience South Africa.

Congo craton, are to a large extent enigmatic and these are the focus of this thesis.
The geology of these regions was described in chapter 4.

The MT data that will be analysed are from the N-S oriented profiles labelled ETO-KIM, DMB, NEN, OKA-WIN and the E-W RAK-CPV (see Figure 5.1). These data provide decent spatial coverage that will enable lateral geometric definition of variations in the geology. Furthermore, the distribution of the MT sites in a rectangular fashion (Figure 5.2) allowed 3D inversion to be conducted (see chapter 7). The average distance between the profiles is 380 km and the station spacing along the profiles is approximately 20 km.

Figure 5.2: Map showing sites names overlain on the geology map of Namibia (modified from Corner (2008)). KI=Kamanjab Inlier, AL=Autseib Lineament, OML=Omaruru Lineament, OKL=Okahanja Lineament.
5.1.1 Pseudosections

Preliminary analysis of MT data, before any formal inversions are done, can give useful information about the subsurface variations in resistivity. This first-pass presentation of MT data is usually in a form of pseudosections. These are gridded maps of apparent resistivity and phase data against increasing period (proxy for increasing depth) that present a qualitative first pass view of the variation of resistivity with period (i.e., depth). The pseudosections were created using the WingLink package using 10 periods per decade, smoothing factor of 5, spline weight of 7, interpolation radius 7 and horizontal (x) distance of 10 km. In the pseudosection plots the warm (e.g., red) and cold (e.g., blue) colours show conductive and resistive structures respectively.

Strike directions were determined using the STRIKE code and TE and TM mode defined by decomposing the data to a 50° direction. The data were projected on a profile perpendicular to the strike direction.

5.1.1.1 ETO-KIM profile

The ETO-KIM profile partly crosses the NW Kalahari craton and transcends the Damara belt and the southern Congo craton in northern Namibia (see Figure 5.2). The TE and TM mode apparent resistivity pseudosection (Figure 5.3) indicate a highly resistive feature that corresponds to the spatial position of the KI. The resistive features of the in the Damara belt, in the crust are clearly seen. The significance of these will be explained in later sections. This profile is an extension of the KIM-NAM profile that was modelled in Muller et al. (2009)’s study of the Kaapvaal craton and the Rehoboth Terrane. The ETO-KIM profiles crosses major Damara belt lineaments: Autseib Lineament (AL) at station ETOo09, The Omaruru Lineament (OML) at site ETOo04, the Okahanja Lineament (OKL) at site KIM447 and the Matchless Amphibolite belt (MA) at site KIM444 (see Figure 5.2). Additionally the profile crosses the Kamanjab Inlier (KI) from site ETO012 to ETO019.

5.1.1.2 DMB profile

The DMB profile also crosses the same structures as the ETO-KIM profile and extends further south into the Kalahari craton. Given its continuity into the Kalahari craton, past the WIN profile, the profile is the longest of all the profiles modelled. The pseudosection plot of the DMB profile is shown in Figure 5.4 and reveals the deep- and shallow-seated resistive features mapped also in the ETO-KIM profile.
Figure 5.3: Pseudosection plot of the ETO-KIM profile, showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted triangles show site locations.

5.1.1.3 *NEN profile*

The pseudosection plot of the profile is shown in Figure 5.5 and shows similar features such as the previous two profiles, with an additional resistive feature further south that is placed in the Kalahari craton.
Figure 5.4: Pseudosection plot of the DMB profile, showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

5.1.1.4 OKA-WIN

The OKA-WIN profile is more difficult to model given the geometry in which the data was collected. To avoid additional projection-related distortions that might be introduced, two pseudosections plots were produced. The first pseudosection plot (shown in Figure 5.6) uses only the OKA stations (see Figure 5.2) and the second pseudosection was created using the SW-NE trending WIN stations (Figure 5.7). The north and south resistivity features mapped on the NEN profile are evident on this profile.

5.1.1.5 WIN profile

The E-W WIN profile has different orientation from the other profiles but the pseudosection plot (Figure 5.8) is included here for correlation purposes. The resistive feature mapped to the south of the NEN and OKA-WIN profiles is also seen on this section. However, given its orientation, this profile could not be modelled using 2D inversion. But the data was used for 3D inversion as no assumptions needed to be made regarding the strike direction.
Figure 5.5: Pseudosection plot of the NEN profile, showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

5.1.1.6 **RAK profile**

Similar to the WIN profile, the RAK profile trends in the E-W direction and the pseudosection is presented here for feature-correlation purposes (see Figure 5.9). The TE and TM phase data suggest the presence of a continuous (E-W) resistive feature at depth. The continuity of this feature, which could represent the southern extent of the Congo craton, will be addressed using formal 2D and 3D inversion modelling. Like the WIN profile, data from the RAK profile was used for 3D inversion. It is important to note that the pseudosection plots provide a quick qualitative view at the data and while some features can already be distinguished, their geometry, vertical and lateral extent can be quantified from 2D or 3D inversion models.

5.1.2 **Niblett-Bostick depth maps**

In order to have confidence in the results of our inversions, it is important to analyse the depth to which MT responses are able to penetrate. One way of estimating these
Figure 5.6: Pseudosection plot of the OKA profile, showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

depths was shown in section 2.6.1.1 (equation 2.40).

Figure 5.10 show the estimated depth of penetration for the TE and TM mode responses for the ETO-KIM profile. It important to note that these penetration depths are based on 1D estimates, and that for complex 3D structure may not represent the true depths. Given the variation in resistivity structure along each of the profiles (as observed in the pseudosection plots), it is expected that the penetration depth estimates will vary along the profile. Furthermore, given the inductive nature of EM fields, regions of lower resistivity will be characterized by lower depth of penetration.

The Nibblet-Bostick (NB) depth transformation from apparent resistivity and phase against period to layer resistivity against depth was shown to be very effective in mapping regional resistivity contrasts of the Southern African lithosphere by Jones et al. (2009b). Similar maps are constructed here for 50, 100, 150 and 200 km depths for the data from stations shown in Figure 5.2 only, the results are shown in Figure 5.11. In Figure 5.11 depth slices of maximum resistivity, obtained by rotating the
Figure 5.7: Pseudosection plot of the WIN profile (oriented SW-NE), showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

apparent resistivity and phase data through 180°, then deriving the NB transformed resistivity-depth data, and determining the largest value of resistivity at the particular depth of interest (Jones et al., 2009b). As was noted by Jones et al. (2009b) these maps are conservative since maximum resistivity is robust in that it is only affected by significant conductivity bodies and is less affected by distortion effects.

Figure 5.11 also shows strong resistivity contrasts associated with different age terranes. The more resistive Congo craton, particularly at depths more 50 km is readily seen. The Pan-African Damara belt appears less resistive. It must be appreciated that the resistivity maps on Figure 5.11, while providing spatial differentiation between young and old lithosphere, are qualitative representation of the variation of electrical resistivity with depth. The quantitative information, like the depth to the eLAB, will be inferred from the formal 2D and 3D inversion models in chapters 6 and 7.
5.1.2.1 Lithospheric conductance

Figure 5.12 shows a map of integrated electrical conductivity, i.e., conductance, from minimum depth of 40 km to maximum depth of 200 km. Note that the map is in log (Siemens) and shows again the more conductive regions within the Damara belt.

5.1.3 Phase tensor maps

In section 2.12.2 we introduced the Phase Tensor approach as a very useful dimensionality indicator. The main advantage of this technique is that it makes no assumptions about the dimensionality of the subsurface, as such the derived the phase tensor is the phase tensor of the true regional response (Jones, 2012). When the phase tensor is plotted graphically, as shown in Figures 5.13, 5.14 and 5.15, it is usually represented by an ellipse, where the principal directions of the ellipse correspond to the direction of greatest inductive response (i.e., a response owing to an underlying 1D or 2D regional structure.). This direction is the closest equivalent to a 2-D geoelectric strike direction (Caldwell et al., 2004; Bibby et al., 2005).
Figure 5.9: Pseudosection plot of the RAK profile (oriented E-W), showing the TE and TM mode apparent resistivity and phase variations versus period. The black inverted black triangles show site locations.

The phase tensor maps in Figures 5.13, 5.14 and 5.15 are plotted for period decades 1-10 s, 10-100 s and 100 - 1000 s, respectively, where the color of the ellipse give the azimuth of a major ellipsoidal axis ($\Phi_{max}$). The shapes of the ellipses give an idea of the dimensionality of the subsurface. A 1D structure can be assumed when the phase tensor is circular. The Rehoboth Terrane appears to be dimensionally less complex than the Damara belt and Congo craton. It interesting to note the change in ellipsoid direction in the region of the Autseib Lineament (Figure 5.13), which also correspond to the major magnetic anomaly, indicating that this is major crustal feature. In general the majority of the ellipses appear to have a NE-SW direction, corresponding to the major structural features.

There appears to be very little period variation in phase tensor azimuth, mainly in the Rehoboth Terrane (i.e., there is no significant depth variation in the azimuths). In the DGC, period variation is noticeable which, given the structural and geological
Figure 5.10: Niblett-Bostick depth estimates for all stations along the ETO-KIM profile for both the TE (red bars) and TM (yellow bars) modes. While this is based on 1D approximations it gives in indication as to the depth sensitivity of the MT response for each site.

complexity, is to be expected. North of the DGC period variation is minimal on most sites.
Figure 5.11: Niblett Bostick depth maps for various depths.
Figure 5.12: The total integrated electrical conductivity, or conductance (S), from 40 km to 200 km. This depth range is approximately the mantle lithosphere from the average base of the crust to the average base of the lithosphere. The colours represent log_{10}(S).

5.2 STRIKE ANGLE AND DECOMPOSITION ANALYSIS

The discussion in sections 2.6 and 2.11 were concerned with the analysis and decomposition methods that are employed for MT data analysis. In this section the results of applying these techniques to the data are shown. The geo-electric strike analysis and decomposition was done using the extended version of *Groom and Bailey (1989)* method, which performs not only the single-site STRIKE decomposition, but also the multi-site and multi-frequency and depth analysis, as was shown by *McNeice and Jones (2001b)*. Figure 5.16 shows the results of applying the single site GB analysis from sites on all the profiles. Given that the large area concerned and that dimensionality of the sub-surface can vary regionally along the profiles, GB analysis was carried out for the ETO-KIM, DMB, NEN, OKA-WIN and RAK profiles.
Figure 5.13: Phase Tensor invariant ($\phi_{\text{min}}$) azimuth after (Caldwell et al., 2004) averaged over the 1-10s decade. One ellipsoidal principal axes indicates the direction of maximum induction. It can be seen that along a magnetic anomaly that corresponds to the Autsieb Lineament there is a rotation in the direction of the ellipses.

The analysis was done for the period bands indicated, with the shortest periods (0.01-10 s) showing crustal depths and the longest periods (100-1000 s) showing mostly upper mantle lithospheric depths. The black lines indicate the geo-electric strike azimuth and the coloured squares show RMS misfit error, normalised by observational errors in the observed MT response, at each site.

As can be observed in Figure 5.16 most sites show low RMS misfits between the GB model and the data, which suggests that sub-surface is broadly 1D or 2D. In some sites on the northern end of the ETO-KIM profile, relatively high RMS misfits are observed. The responses from these few sites show current-channelling effects which were caused by high conducting power stations infrastructure in the vicinity of the Ruacana falls in NW Namibia. If severe current channelling effects were observed the sites were not used or the long period responses were discarded. The strike direction for most sites varies slightly along each profile from site-to-site, but broadly
5.2 STRIKE ANGLE AND DECOMPOSITION ANALYSIS

Figure 5.14: Phase Tensor invariant ($\phi_{\text{min}}$) azimuth after (Caldwell et al., 2004) averaged over the 10-100 s decade. One ellipsoidal principal axes indicates the direction of maximum induction. It can be seen that along a magnetic anomaly that corresponds to the Autsieb Lineament there is a rotation in the direction of the ellipses.

approximately NE-SW for most of the sites.

Figure 5.17 shows the sensitivity to strike direction for the four N-S profiles crossing the Damara belt, shown for upper mantle Niblett-Bostick depths (50-150 km). Only a few sites are strongly sensitive to strike direction, where low RMS values (<2) are defined over a limited range of azimuths. Other sites are relatively insensitive to strike direction and are characterized by low RMS values.

A similar map is shown for depth range 50-150 km (Figure 5.18). Majority of the sites display low RMS misfit at this depth range for all the profiles, which means that for range of strike azimuth, these sites are insensitive to the strike direction.

In order to provide an idea of the derived decomposition parameters from GB analysis, strike analysis maps are shown for the ETO-KIM (Figure 5.19) and DMB (Figure 5.20) profiles. What is shown are the GB decomposition parameters for sites
along profile (a) and the variation in azimuth for each site. The color representation distinguishes several depth ranges as shown. It is readily clear that the different depth ranges result in relatively similar strike azimuths. Furthermore, there is little azimuthal variation with depth for most sites (b).

To this effect, the final regional geo-electric strike direction is on average $50^\circ$, and this was used for further 2D modelling.
Figure 5.16: Results of MT STRIKE analysis along the ETO-KIM, DMB, NEN, OKA-WIN and RAK profiles, using Groom and Bailey (1989) decomposition. The analysis was done for the period bands indicated, with the shortest periods (0.01-10 s) showing crustal depths and the longest periods (100-1000 s) showing mostly upper mantle lithospheric depths. The black lines indicate the geo-electric strike azimuth and the coloured squares show RMS misfit error, normalised by observational errors in the observed MT response, at each site.
Figure 5.17: Sensitivity to strike direction for the four N-S profiles crossing the Damara belt, shown for crustal Niblett-Bostick depths (10-50 km) shows that only a few sites are strongly sensitive to strike direction, where low RMS values (<2) are defined over a limited range of azimuths. Other sites are relatively insensitive to strike direction and are characterized by low RMS values.
Figure 5.18: Sensitivity to strike direction for the four N-S profiles crossing the Damara belt, shown for upper mantle Niblett-Bostick depths (50-150 km) shows that only a few sites are strongly sensitive to strike direction, where low RMS values (<2) are defined over a limited range of azimuths. Other sites are relatively insensitive to strike direction and are characterized by low RMS values.
Figure 5.19: (a) Decomposition parameters obtained from GB analysis of MT sites along the ETO-KIM profile. (b) Final regional azimuths, relative to North, for depth ranges and sites. Final regional geoelectric strike direction is $50^\circ$. 
Figure 5.20: (a) Decomposition parameters obtained from GB analysis of MT sites along the DMB profile. (b) Final regional azimuths, relative to North, for depth ranges and sites. Final regional geo-electric strike direction is $50^\circ$. 
6

2D INVERSION OF THE ETO-KIM, DMB, NEN AND OKA-WIN PROFILES

6.1 2D ISOTROPIC AND ANISOTROPIC INVERSION

The following section deals with the isotropic and anisotropic inversion of the MT data, as the presence of anisotropy is being increasingly being taken into account when conducting MT data interpretation. It is important to understand the nature of anisotropy at all scales in order to infer present and past tectonic processes.

6.1.1 2D Isotropic inversion

Two-dimensional (2D) isotropic electrical resistivity models were derived for the ETO-KIM, DMB, NEN, and OKA-WIN profiles using the finite difference code of Rodi and Mackie (2001). This code uses a non-linear conjugate gradient (NLCG) scheme to search for the smoothest best-fit model (Mackie and Madden, 1993a,b; Rodi and Mackie, 2001).

Several inversion parameters were tested in order to obtain a robust, low RMS and geologically-acceptable model. The regularization parameter tau (τ) is the “smoothing” operator that controls the Tikhonov regularisation trade-off between model smoothness and fitting the data. A tau value of 5 was used for the two models, after a rigorous search for the tau parameter that gives the most favourable compromise between RMS fit and model smoothness using the L-curve approach (Hansen and O’Leary, 1993).

A 100 Ωm homogeneous half-space was used as the starting model for all inversion runs. The systematic inversion approach that was followed included fitting initially the TE and TM phase responses preferentially by allocating high error floors (50%) for the apparent resistivities. The next step was to include the apparent resistivities after some iterations by reducing the error floor, and lastly the vertical magnetic field data...
were included. The reason behind inverting the phase responses first is that they are generally free of static shift effects (Jones, 1988). The alpha and beta parameters in the inversion code, which control the horizontal smoothness and vertical regularization weighting, were set to 1 and 0 respectively after testing. The final error floors were 5% for phase, 10% for apparent resistivity and 0.02 for vertical fields.

The smoothest models available were sought since the inductive nature of the MT technique allows us to assume a smoothed structure at depth. The next subsections show the anisotropic models for each of the profiles.

6.1.2 The ETO-KIM profile

Figure 6.1 shows the isotropic smooth electrical resistivity model, which reveals a heterogeneous electrical resistivity for the mantle lithosphere. The Damara belt crustal conductor remains evident in the lithospheric model, as well as the resistive crustal granitoids. The most striking feature on the MT models is the resistive (i.e., colder), thicker lithosphere interpreted to be the electrical manifestation of the southern Congo craton.

![Figure 6.1: Two-dimensional isotropic model of the ETO-KIM profile. AL: Autseib Lineament. The Congo craton is clearly mapped, comprising resistive mantle lithosphere. The Moho boundary is inferred from seismic receiver functions.](image)

6.1.3 The DMB profile

Figure 6.2 shows the 2D inversion model of the DMB profile, including the RMS variation for each station and the conductance profile at 100 km. Figures 6.3 and 6.4 show the comparison between responses from resulting 2D inversion and the original
Figure 6.2: Isotropic 2D inversion model for the DMB profile. The RMS variation of each station along the profile and the 100 km conductance profile are shown. Final RMS was 2.58. Highlighted on the inversion model are the Damara belt and Congo craton eLAB. The black horizontal line shows the crustal boundary and thin white line depicts the Niblett-Bostick depth of penetration for each station. A is the crustal granites and B corresponds to the location of the thrust faults, clearly dipping to the south based on MT data.

The electrical resistivity models, particularly on the OKA-WIN profile (Figure 6.6) suggests that the LAB beneath the DGC is at a depth of ~150 km, consistent with a 50 mW.m$^{-2}$ steady-state conductive lithospheric geotherm. (Pollack and Chapman,
Figure 6.3: TM modelled data vs original data, shown for the apparent resistivity and phase responses. The difference in the apparent resistivity and phase responses shows model sensitivity, where most of the structure in the final model is influenced primarily by the phase data.

Also evident on the models is the resistive (i.e., colder), thicker lithosphere north of the DGC, interpreted to correlate with the southern edge of the Congo craton. The eLAB beneath the Congo craton appears to be much deeper at about 240 km. The Okavango Dyke Swam (ODS), with an approximate width of 100 km, appears as a very resistive (10,000 Ωm) crustal feature apparently extending to depths of 60 km. The dykes are mainly doleritic in composition and are hosted by granite gneisses, which would account for the high resistivities. The large strike (1500 km) extent of the dyke swam suggests a pervasive emplacement along zones of crustal weakness during continental breakup.
Figure 6.4: TE modelled data vs original data, shown for the apparent resistivity and phase responses. The difference in the apparent resistivity and phase responses shows model sensitivity, where most of the structure in the final model is influenced primarily by the phase data.

6.2 2D ANISOTROPIC INVERSION

The anisotropic code used is an extension of the isotropic code discussed above (same regularization), with an additional parameter ($\tau_{iso}$) that controls the degree of anisotropy permitted by the inversion (Evans et al., 2005; Baba et al., 2006; Evans et al., 2011; Le Pape et al., 2012), that is how close the resistivities $\rho_{xx}$, $\rho_{yy}$ and $\rho_{zz}$ will be to each other. A low $\tau_{iso}$ value (0) will enforce a totally anisotropic result; that means independent TE and TM mode inversions. The limitation of the code is that the anisotropic direction must coincide with the geo-electric strike direction. The resulting 2D inversion models show resistivity in two different directions; one parallel ($R_{xx}$) and the other perpendicular ($R_{yy}$) to the geo-electric strike. The $R_{zz}$ model is not shown here as it is particularly difficult to resolve due to the fact that most of the current flow in the MT problem is horizontal.

Figure 6.7 shows the isotropic (a) and anisotropic (b-c) lithospheric scale 2D models derived for the ETO-KIM profile. The anisotropic models suggest elevated mantle conductivities in directions parallel to the profile ($R_{yy}$) particularly for the lower-most mantle beneath the Congo craton. As mentioned above, various candidates for mantle anisotropy include (but are not limited to) strain induced crystal preferred orientation (CPO) of anisotropic minerals, hydrogen diffusivity in mantle minerals and/or presence of partial melt (Wannamaker, 2005; Yoshino et al., 2006). Due to the lack of
mantle xenoliths samples and age dates in the northern Namibia region (though there are kimberlites), it is unknown (at this time) if the mantle is dominated by hartzburgite, lherzolite or dunite. Xenolith-defined temperature profiles derived in the Kundelungu plateau in the Democratic Republic of Congo (NE of Namibia), suggest that the Congo craton lithosphere experienced melt induced metasomatic episodes (Batumike et al., 2009) ca 32 Ma ago with LAB depths estimated at 175 km. This result is consistent with a recent thermobarometric study of Ashchepkov et al. (2011), who analyzed mantle xenocrysts from the Catoca kimberlite cluster in northern Angola, and inferred a geotherm of 37-40 mWm$^{-2}$ and a fertile clinopyroxene-rich layered mantle lithosphere with metasomatic signatures.

If these geothermal conditions are representative for the larger Congo craton lithosphere to the south in Namibia, then it is reasonable to expect a slightly hydrated lower mantle lithosphere, which could explain the enhanced conductivities. This
Figure 6.6: Two-dimensional isotropic inversion models of the OKA-WIN profile, derived from joint inversion of the TE and TM mode data. ODS: Okavango Dyke Swam. The Kalahari craton in this part of the region is made up of the Rehoboth Terrane. The seismic line represent the S-wave receiver functions of a station located within the Damara belt.

Figure 6.7: Isotropic (a) and anisotropic (b and c) lithospheric models of the ETO-KIM profile. Conductivity in the direction perpendicular to the profile (Rxx) and parallel to the profile (Ryy) are shown. AL: Autseib Lineament.
would though be in sharp contrast to the Kaapvaal Craton, that has a hydrated upper lithosphere and a dry lower lithosphere (Peslier et al., 2010). Furthermore, given its lower geothermal gradient profile, it can inferred that the Congo craton lithosphere must be thicker at its core than, say the Kaapvaal craton lithosphere, as the latter shows a slightly elevated geotherm of 41-44 mWm$^{-2}$ (see results of Muller et al. (2009)). Figure 6.8 shows the ETO-KIM and Muller et al.’s [2009] KIM-NAM profiles,

![Figure 6.8: A model showing the combined KIM-NAM profile from Muller et al. (2009) and the 2D isotropic ETO-KIM profile (see Figure 6.1), depicting the current tectonic structure from the Kaapvaal craton in the SW to the Congo craton in the NW. The KIM-NAM model was derived from 2D smooth inversion of joint apparent resistivity and phases for both TE and TM responses decomposed for 45° E of N azimuth. The algorithm used for the KIM-NAM profile was of Rodi and Mackie (2001) implemented in WinGLinK®. The dotted black lines shows the approximate depth to base of crust.](image)

depicting the current tectonic geometry derived from modelling MT data. Major Archean cratonic segments (Kaapvaal and Congo) are delineated from Proterozoic (Rehoboth Terraine) and Neoproterozoic (Damara belt) regions by their thickness and resistivities, confirming again the general correlation of older lithosphere being thick and resistive, whereas younger lithosphere is thin and shows elevated conductivity. Anisotropic 2D inversion models were also created for the DMB (Figure 6.9) and NEN (Figure 6.10) profiles.
6.3 Robust Features of the 2D Models

The inherent non-uniqueness of the inversion process makes it difficult to test the model parameters and the resolution of the resulting model. There are, however, approaches that can be followed to place bounds on feature dimensions and properties. A few effective approaches include using a preferred model as a starting model or perturbing a feature of interest by either removing it or changing its dimensions and then assessing the resultant change in misfit *Evans et al.* (2011); *Spratt et al.* (2009). The following tests were done for the MT data for the ETO-KIM profile using *Rodi and Mackie* (2001)'s WINGLINK package. Similar tests were done for all the other profiles and the results were found to be consistent.
1. **LOCAL MINIMA TEST**

This test tries to ensure that the inversion result is not simply a local minimum in misfit space. Once an inversion result is found and features of interest identified, the user has the option to perturb the feature’s resistivity and restarting the inversion process. If the feature returns, then the assumption is that it is required by the model. This test was performed for the model shown in Figure 6.1, by making the resistivity of both the Damara belt conductor (10 Ohm.m) and the Congo craton (10,000 Ohm.m) 100 Ohm.m. Not only did the two features return, but the RMS misfit did not change significantly. This means that two features are required by the model and can thus be considered to be robust.

2. **HIDDEN TRADE-OFF TEST**

This is a more "constrained" version of the local minima test above in the once a specific feature in the model is changed, the inversion can be restarted to solve for the smoothest closed to that perturbed model. As was noted by Evans et al. (2011) this test ensures that there are not any hidden trade-offs between...
different regions of the model. If there are, then the resultant model will have a similar misfit to the preferred model but with different features.

3. **RESISTIVITY "LOCK" TEST**

A feature of interest can have its resistivity unchanged or "locked" during the inversion process. A weighting factor can be applied such that degree to which the perturbation is locked can change if required by data. This test is somewhat "risky" in that a smooth model is sought but the "locked" feature is forced to not be smooth. If there are constraints on the geometry of features being locked, from seismic information for instance, then this test can be effective.

4. **SYNTHETIC TESTS**

The ability of the MT data to resolve geologic features depends, among other things, on the station coverage, the periods being modelled and of course the nature and geometry of rocks being models. The easiest test of the ability of a certain configuration of MT stations to resolve a feature of interest is to construct an "ideal" model and produce MT forward responses. These responses can then be inverted in 2D or 3D and the a comparison made with the "ideal" model. In this case, for instance, a generic model comprising thick (260 km) resistive (10,000 Ohm.m) lithosphere depicting the Congo craton, which has on its margins thinner (150 km) less resistive (2000 Ohm.m) lithosphere depicting the Damara belt, was constructed as shown in Figure 6.11. Generating forward responses (then adding 2.5 % Gaussian noise) for MT station configuration such as the ETO-KIM profile, and inverting these to see produced the result in Figure 6.11 B. In general the lithospheric structure was broadly recovered with eLAB thickness of 150 km and 250 km for younger ("Damara belt") and older ("Congo craton") respectively. The depth to eLAB in the mantle at adiabatic conditions is defined as the depth where the electrical resistivity has a value of 100 Ohm.m. This test implies that the station spacing of 20 km (akin to the ETO-KIM profile) and data up to 10,000 s, is adequate to resolve the Achaean and Proterozoic lithosphere.

A further test of reproducibility of MT data for lithospheric thickness is to model the two modes independently. This test gives an indication of which structures are dominant in which mode. The test was done for the generic lithospheric model in Figure 6.11 A and the results are shown in Figure 6.12.
The middle to lower crustal conductive zone and ( structure dominated by three main features that correlate relatively well with known

Figure 6.11: A generic lithospheric model (A) and its 2D inversion result (B) done in order to test the ability of MT data to resolve the LAB. The inversion appears to recover the full structure, for station spacing of 20 km (similar to the ETO-KIM profile)

Inversion parameters: 100 Ohm.m starting model, both TE and TM mode data was used with the smoothness parameter \( \tau = 3 \). A 2.5 % Gaussian noise was added to the data.

The two inversion models are able to recover the general structure of the generic model with one notable difference; the TM mode underestimate the eLAB by about 30 km, particularly for the Proterozoic "Damara" lithosphere. This value will be even less when considering the conductive nature of the Damara crust. However it can be inferred, based on these tests, that the generic lithospheric structure can be confidently resolved.

6.4 INTERPRETATION OF 2D RESISTIVITY MODELS

The 2D inversion model (Figure 8.1) of the ETO-KIM profile reveals a complex crustal structure dominated by three main features that correlate relatively well with known structural and geological features: (1) the upper crustal resistive structures, (2) the middle to lower crustal conductive zone and (3) the high resistivity Kamanjab Inlier.
Also evident in the MT model is the presence of the lower crustal conductive feature anomalies in the DGC upper-crust. Outcrops, particularly in the Central Zone. Geological exposures indicate the presence of magnetites, sulphides and graphite which could explain the presence of conductive anomalies in the DGC upper-crust.

Figure 6.12: Independent 2D inversion of the TM and TE mode for a generic lithospheric model shown in Figure 6.11 to give an indication of which structures are dominant on which mode. While both modes are able to resolve the broad structure, the TM mode underestimates the eLAB by about 30 km.

to the north. The upper crustal resistive features are mapping the syn-and post-tectonic granitic intrusions that form a significant component of the Damara belt outcrops, particularly in the Central Zone. Geological exposures indicate the presence of magnetites, sulphides and graphite which could explain the presence of conductive anomalies in the DGC upper-crust.

Also evident in the MT model is the presence of the lower crustal conductive feature
Figure 6.13: (a) Crustal-scale resistivity model of the ETO-KIM profile, showing the main features resolved by the inversion process. The mid-lower crustal conductors, upper crustal Pan-African granites and the Archean basement (Kamanjab Inlier) are the major features of this model. For comparison the geological section (b), modified from Gray et al. (2007), is included. AL: Autseib Lineament, OML: Omaruru Lineament, OKL: Okahanja Lineament, MA=Matchless Amphibolite, SFZ: Southern Foreland Zone.

described previously (Van Zijl and de Beer, 1983; Ritter et al., 2003b). The Kamanjab Inlier (KI), which forms the basement upon which Damara sediments were deposited ca 770 Ma, is imaged as a very resistive feature. Compositionally, the Kamanjab inlier comprises granitoids and metamorphic sequences (Goscombe et al., 2003), which accounts for its high resistivity (over 10 000 Ωm). The surface boundary of the inlier is traced on the geological map between sites ETO011 and ETO012; however the MT model reveals that this boundary could be extended even further south between ETO009 and ETO010. Furthermore, the boundary of the Kamanjab Inlier at depth appears to have a gentle dip to the south, terminating beneath the Damara north-central zone.

The Autseib Lineament (AL) is one of the most prominent magnetic anomalies in the Southern Africa (Corner, 2008). It separates the Damara Central Zone from the Northern Zone and extends north-westward into Botswana (see Figure 4.8). The AL, a former extensional fault reactivated as recently as the Mesozoic (Clemenson et al., 1997;
Corner and Swart, 1997), is now a thrust fault and appears on the resistivity model as an upper crustal conductor, steeply dipping to the south. Evidence for modern reactivation along the pre-existing feature was presented by Raab et al. (2002) using apatite fission results.

The resistivity of the cratonic segments ranges from 5 000 $\Omega$m to over 30 000 $\Omega$m at 100 km depths, consistent with laboratory-derived dry olivine resistivities (Constable et al., 1992; Jones et al., 2009a; Fullea et al., 2011). The southernmost extent of the Congo craton lithosphere is prominent and at a depth of about 150 km directly beneath the Autseib lineament, where there is significant change in resistivity from 30 000 $\Omega$m to about a 1000 $\Omega$m.

The LAB beneath the Congo craton on the NEN profile is inferred at depths of approximately 250 km. It is interesting to note a thick resistive lithosphere in the south of the profile. While current tectonic outlines suggests that this is part of the composite Kalahari craton, the possibility exists that it could also be part of the enigmatic Okwa terrane. A complete modelling of the Kalahari data (one EW profile and one NS profile), on which the Okwa Terrane is located, is currently being done and the understanding of those results will provide constraints on its extent and the possible relation to the Rehoboth
The motivation for modelling the Earth in 3D (instead of 2D) is multi-fold. Firstly, the Earth is 3D and it must be modelled as such. While we collect MT data along 2D profiles, the data we acquire and subsequently process results in a full 2x2 impedance tensor, with four complex tensor elements for each period, relating the electric and magnetic field components. Modelling all four components \((Z_{xx}, Z_{xy}, Z_{yy}, Z_{yx})\) not only requires no assumptions about the strike directions but perhaps more importantly provides added information about the electrical resistivity structure. Secondly, MT data often shows that the validity of the 2D assumption mentioned above fails due to influence of 3D subsurface structures. In the ETO-KIM profile for instance, the dimensionality and phase tensor analysis shows that 3D induction effects are present in the data, as evidenced by high RMS, twist and shear values. Modelling these data in 2D will lead to incorrect results and the true geological structure will not be fully recovered. In order to ensure that the features of the 2D models are robust it is necessary that comparison be made with 3D inversion results. Thirdly the spatial data coverage of SAMTEX stations warrants data to be treated in 3D. Furthermore, given geological complexities may occur as a result of orogen convergence, it is reasonable to expect 3D manifestation at depth.

In order to test the multi-dimensional properties of the sub-surface and have full confidence in the resistivity features, the data along the ETO-KIM, NEN, DMB, OKA-WIN, and RAK-CPV profiles were inverted using the 3D algorithm. The spatial geometry, site spacing and frequency range are key parameters that must be considered in order for 3D modelling to take place. The distribution of stations that will be used to produce a 3D inversion model is shown in Figure 7.1. Eighty five (85) MT sites having the best quality responses were selected out of a total of 204 stations making up the 6 profiles above.
In this section two different inversion schemes were applied to model our data in 3D. The \textit{WSINV3DMT} code of Siripunvaraporn et al. (2005a) and the newly developed modular electromagnetic modelling and inversion \textit{ModEM} scheme of Egbert and Kelbert (2012) were used. The \textit{WSINV3DMT} is based on the 2D data space Occam’s inversion of Siripunvaraporn and Egbert (2000) which seeks the smoothest model to fit the data. The \textit{ModEM} inversion scheme not only is suited for large inversions (such as in our case study) but also consistently resulted in low RMS models and geologically reasonable geometry of features. As such only the \textit{ModEM} results will be shown here.

As with the 2D models, various parameters were tested to obtain geologically plausible models. The \textit{ModEM} algorithm has provision to invert up to five combinations of impedance tensor elements ($Z_{xx}$, $Z_{xy}$, $Z_{yy}$, $Z_{yx}$) and vertical transfer functions ($H_z$ tippers). Various combination of inversion parameters including varying the percentage...
of error floors as shown in Table 7.1, we tested and compared: A total of 85 stations

<table>
<thead>
<tr>
<th>Inversion type</th>
<th>Impedance elements</th>
<th>Impedance error floors (%)</th>
<th>Tipper error floors (%)</th>
<th>Iterations</th>
<th>RMS</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$Z_{xx}$, $Z_{xy}$, $Z_{yy}$, $Z_{yx}$</td>
<td>10 and 30 -</td>
<td>-</td>
<td>49</td>
<td>3.66</td>
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<tr>
<td>2</td>
<td>$Z_{xy}$, $Z_{yx}$</td>
<td>10 -</td>
<td>-</td>
<td>84</td>
<td>4.50</td>
</tr>
<tr>
<td>4</td>
<td>$Z_{xy}$, $Z_{yx}$</td>
<td>10 0.1</td>
<td>48</td>
<td>72</td>
<td>4.95</td>
</tr>
<tr>
<td>5</td>
<td>$Z_{xx}$, $Z_{xy}$, $Z_{yy}$, $Z_{yx}$</td>
<td>10 0.1</td>
<td>-</td>
<td>48</td>
<td>7.85</td>
</tr>
</tbody>
</table>

were carefully chosen and noisy responses edited and non-ideal stations removed. The MT responses for each site were then decimated to 4 periods per decade and 18 periods values within the range 0.01-15,000 s were modelled. Not all stations had vertical transfer functions and for those that did (when using option 4 and 5 in Table 7.1) a constant error floor in $H_z$ was set at 0.1 %. The mesh was discretized to be fairly regular on the centre, with cell widths of approximately 20 km (Figure 7.2). The mesh

Figure 7.2: Mesh/grid used in producing 3D inversion models having dimension 64,78,52 in $x, y, z$ respectively. A 100 Ohm.m starting model was used for inversion runs. Top figure shows the NS 2D section across the middle of the grid, while the bottom grid is the depth slice.
also comprised a total of 64 cells in the X direction, 78 in Y direction and 52 cells in the Z direction, with most stations preferentially adjusted to be located on the centre of cells.

It should be noted that the intensive computational requirements involved in inverting the full 3-D MT impedance tensor data and tippers necessitates that the numbers of model and data parameters be reduced in comparison to the 2D inversion. About 9 periods per decade were, for example, used to produce the 2D inversion models in contrast to 4 for 3D models. Similarly, the mesh size for 2D inversion for each profile had on average 180 columns and 70 rows. Several starting models which included different resistivity half-space models of 100 ohm-m and 200 ohm-m were tested and compared with the former resulting in lower RMS models. Table 7.1 shows several combinations of inversion parameters tried and the resulting RMS values. The inversion runs were set for a maximum of 120 iterations and required 36 to 72 hours of computation (Figure 7.3) on the STOKES (SGI Altix ICE 8200EX) 320 node cluster of the Irish Centre of High Performance Computing (ICHEC www.ichec.ie).

As expected the more data one adds to the inversion the higher the resulting

![Figure 7.3: Variation in RMS with respect to iteration. The inversion took 69 hours to complete on the STOKES cluster of the Irish Centre of High Performance Computing](image)

RMS. All models were generally similar, with minor geometrical variations, but the
major lithospheric structures were consistent. Due to the extensive size of the study area, it is important to estimate the depth of investigation at all the stations in order to gain confidence in the derived 2D and 3D models. Figure 7.4 (A) shows the maximum Niblett-Bostick penetration depths interpolated across the 3D data space, with the Central Damara conductor being most prominent as it results in limited penetration of EM fields. Generally, for most stations penetration depths were in excess of 200 km (Figure 7.4 (B)). It must be recognised that the penetration depths are only from a 1D approach, but give a qualitatively accurate view. In the next section, the inversion results in a form of 3D depth maps and sections will be presented.

Figure 7.4: Depth of penetration interpolated across the 3D area (A) and shown for each station (B). The Central Damara conductor (DGcc), which results in limited penetration is particularly distinct on the image.
7.3 3D inversion results

Given that several inversion parameters were tested in order to derive a geologically reasonable model, several output maps are presented for comparative purposes. Figure 7.5 shows a perspective view of the inversion model, resulting from modelling the full impedance tensor $Z_{xx}, Z_{yy}, Z_{xy}, Z_{yx}$. The resulting RMS was 3.66. CC=Congo craton, DGCc=Damara-Ghanzi-Chobe belts conductor.

Figure 7.5: 3D perspective view of the inversion results, fitting the full impedance tensor $Z_{xx}, Z_{yy}, Z_{xy}, Z_{yx}$. The resulting RMS was 3.66. CC=Congo craton, DGCc=Damara-Ghanzi-Chobe belts conductor.

The 3D inversion results suggest that the DGC conductor is extensive along geological strike of the Damara orogen, as is evident on the depth slices (Figures 7.5 and 7.6). Spatial comparison with geological maps suggests that this conductor resides mainly in the Central Zone of the DGC, the top of which is at approximately 20 km. The depth to the base of the conductive feature is generally poorly-resolved due to the inductive nature of the MT technique and the limited penetrative ability of the EM fields below conductors.

The nature of 3D inversion is such that more data (i.e., full impedance tensor) are modelled; as a result one expects the 2D and 3D results to differ. Additionally, variable grid size, lack of data, misfit distribution and trade-off parameter will introduce variations in the models. The geometric variations of cratonic lithosphere observed between the profiles could be the result of this very fact. That being the case, the general features of the model that are robust are the cratonic lithosphere to the
Figure 7.6: Resistivity maps derived from the 3D inversion at three depths. The black-filled circles show station locations. Also shown is the 20 km resistivity slice overlain on the geology map in Figure 7.1. The correlation of the DGC conductor with the Central zone is clear from this map.

north and the presence of the Central Damara conductor, which were also observed in the 2D models.

In order to obtain an indication of the variation in geometry of the tectonic features across the region, 2D profiles were extracted out of the 3D inversion model, and the results are shown in Figure 7.7 for the ETO-KIM, DMB, NEN and OKA-WIN profiles from west to east. The TSUM S-wave receiver function site is also shown on the DMB profile. The modelled responses fit the observed data well, as illustrated in
Figure 7.7: 2D profiles of ETO-KIM, DMB, NEN, OKA-WIN transects extracted from 3D inversion model. Profile locations derived are shown beside each image. The S-receiver function response is projected on the DMB profile as it is in closest proximity. The interpreted MOHO and LAB conversions, the latter indicated by a black arrow from Kumar et al. (2007), are shown. The LAB conversion (180 km) derived by Hansen et al. (2009) was not particularly well resolved as the signal at these depths is close to bootstrap error limits (2σ).

Figure 7.8, which shows the RMS difference for all impedance tensor elements of the modelled responses and the resulting 3D inversion model.
Figure 7.8: Variation in RMS with respect to iteration. The inversion took 69 hours to complete on the STOKES cluster of the Irish Centre of High Performance Computing

7.3.1 Damara belt conductor inter-connection

The results of the 3D inversion suggests that the Damara belt crust contains two significant features: (1) the upper crustal resistive structures and (2) the middle to lower crustal conductive layer. This structure is evident in all the profiles. It is thus a reasonable possibility that the crustal conductor is pervasive throughout the Inland branch of the Damara orogen. However, given the large profile separations (minimum 200 km to maximum 400 km), quantitative test of this assumption must be done in
order to test it validity.

To this effect a constrained starting model (instead of 100 Ohm half-space) was constructed comprising a 10 Ohm.m conductor in the centre of the Damara belt, from 10 km depth to 50 km (Figure 7.9). As it was mentioned earlier the "closer" the input starting model is to the "true" model the better the resulting fit to the data and thus

Figure 7.9: Starting model used to test the continuity of the Damara crustal conductor (10 Ohm.m) starting at 10 km depth.

Figure 7.10: RMS fits for each station (top) and plotted on a map (bottom), giving an indication of the resulting site-by-fit when using as input to the inversion a model shown in Figure 7.9.
optimal resulting resistivity structure. If the resulting RMS fit, particularly of MT sites over the feature, for using a constrained input model is better than that obtained from using a homogeneous 100 Ohm.m half-space, then is it reasonable to assume that the Damara conductor is extensive throughout the crust. More importantly perhaps is whether the conductive feature returns when the inversion is started with constrained input. When using the starting model shown in Figure 7.9 the resulting RMS fit of the data is shown in Figure 7.10. In general low RMS misfit is observed for most stations indicating that a conductor along the DGC is required by the model.

7.4 3D INVERSION OF 2D PROFILES

In order to validate the features observed in the 3D model and test the resolution of nearby off-profile conductors, 3D inversion models of the each of the 2D profiles were done separately (3D/2D). Siripunvaraporn et al. (2005b) 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).

7.4.1 3D Inversion of ETO-KIM profile

Figure 7.11 shows the resistivity model of the ETO-KIM profile. Three major features are resolved: (1) the upper crustal (<10 km) resistive lithologies of the Pan African granite intrusions, (2) the Damara belt middle-lower crustal conductor and (3) the thick resistive lithosphere to the NE associated with the Congo craton.

7.4.2 3D inversion of DMB profile

All the components of the impedance tensor were used in the inversion and the result is shown Figure 7.12. The error floor were set at 10 % for the off-diagonal elements ($Z_{xy}, Z_{yx}$) and 30% for the diagonal elements ($Z_{xx}, Z_{yy}$). Structures in the model are similar to the structures observed in the 3D inversion of data from all profiles (Figure 7.7). There are slight variations in geometry of some the features, but this is expected considering that, only profile data is modelled.
7.4 3D Inversion of 2D Profiles

Figure 7.11: Three-dimensional inversion of the ETO-KIM profile data, showing the variation in resistivity from Proterozoic Damara belt to the Archean Congo craton. RMS fit was 2.04 after inverting all four components of the impedance tensor. The dashed lines shows the top of the crustal conductor.

Figure 7.12: Three-dimensional inversion model derived from inverting only the DMB profile data. The resulting RMS was 1.68, after inverting the full components of the impedance tensor ($Z_{xx}, Z_{yy}, Z_{xy}, Z_{yx}$). A: upper crustal granites, B: Central Damara conductor.

7.4.3 3D Inversion of NEN profile

Figure 7.13 shows the resulting resistivity section along the NEN profile which maps similar features as observed in other profiles (Figures 7.11, 7.12 and 7.14). The striking difference between the models of the NEN profile and that of inverting full dataset of all profiles simultaneously is the estimated depth to the eLAB. The lithospheric thickness estimated from the inversion of NEN data only appears to be lower (<200 km) than that recovered from inverting all the data from other profiles together (>220 km). The inversion process is, of course, non-unique and the result is influenced by the number of stations and periods inverted. In inverting for single profile data rather than multiple profiles, the number of data being fitted is less, which imply necessarily that the resulting model features will be different.
Figure 7.13: Three-dimensional inversion of the NEN profile. Similar features to the other profiles are observed. The resulting RMS is 1.71.

7.4.4 3D inversion of OKA-WIN profile

Figure 7.14 shows a NS slice extracted from the 3D inversion block of the OKA-WIN profile. Given the curved nature of the profile, the NS section is displayed rather than the profile, in order to better observed resistivity variations. In addition to the crustal Damara conductor and the thick cratonic lithosphere, the Okavango Dyke Swam is mapped as a resistive crustal feature. The dykes are resistive because they are composed of ultra-mafic dolerites.

7.5 INTERPRETATION OF 3D INVERSION RESULTS

The resistive granitic plutons comprise a significant percentage of the geological outcrop of the Central Zone, and reflect massive amounts of volcanism. There are three plausible mechanisms that could result in huge voluminous melt generation specific to the Damara Belt Central Zone: (1) volcanism generated during the initial 750 Ma rifting phase, (2) decompression melting of subducted Khomas oceanic lithosphere (560 Ma), and/or (3) melt derived from crustal recycling of turbiditic material during final 540-520 Ma collision (Gray et al., 2007). Isotopic studies on granites (Jung Hee Suh and Mezger, 2003), however, indicate that the granites in the Central Zone lack positive mantle signatures (i.e., positive $\varepsilon_{Nd}$, low $\delta^{18}O$). S-type granites, which are particularly most significant in volume in the Central Zone, have similar chemical and isotopic signatures to the Neoproterozoic turbidites in the southern zone, indicating a fertile “turbiditic” crustal source at mid-crustal levels by dehydration melting (Gray et al., 2007).
Figure 7.14: Three dimensional inversion result of the OKA-WIN profile data. The resistive, thick Congo craton lithosphere is mapped to the north, the Okavango Dyke Swam (ODS) is clearly imaged as resistive crustal feature. The central Damara conductor appears fragmented in this profile but this is the result of the profile orientation.

Jung Hee Suh and Mezger (2003) suggests that the Damara Central Zone, which has high thermal gradients (30 – 50 °C/km) and high temperature/moderate pressure metamorphic conditions, underwent significant middle to lower crustal low-pressure melting that generated the voluminous post-tectonic granites imaged by the MT model. The elevated gradients correlate with the observed conductivity structure of the Damara Central Zone. More subdued geothermal gradients are observed in the Barrovian metamorphism-dominated marginal zones of the north and southern Damara belt.

The detection of the well-known crustal conductor in the Damara belt still requires explanation, as its understanding and characterization has tectonic implications. Issues that still need attention regarding this conductor include its possible causes and, most importantly, its apparent spatial along-strike extent. These will be addressed now.
Ritter et al. (2003b) attributed graphite along fossilized shear zones as the cause of the observed conductor, based on field evidence of graphite-bearing marble units. Geological exposures indicate the presence of accessory sulphides and oxides (Ben Goscombe, 2012 pers comm) which could explain the presence of conductive anomalies in the DGC upper-crust. Massive sulfide mineralization (Cu-Zn-Au) is also found, for example, in the Matchless Amphibolite belt in the Damara Southern Zone; the latter being thought to represent a possible relict oceanic lithosphere (Barnes and Sawyer, 1980). This observation suggests that sulphides could have been introduced in the Damara crust in the same manner (subduction) as for the longest-known crustal conductivity anomaly, namely the North American Central Plains (NACP) conductor (Jones et al., 1993, 2005b). Furthermore, if the Matchless Amphibolite represents fossil oceanic lithosphere, it is reasonable to expect the presence of carbon in graphite form, which would elevate the conductivity significantly as it would provide an efficient conduction mechanism.

While hydrothermal crustal fluids are interpreted to be present in some locations in the mid- to lower crust, they are highly unlikely to give rise to the high conductivities observed due to their relatively short residence times in the crust (Bailey, 1990) and lack of supporting field evidence of alteration. Thus, a combination of graphite and sulphides offers the most plausible explanation for the increased conductivity in the Damara central zone.

The apparent large strike extent of the Central Zone conductor could be a result of regional metamorphic processes or structural placement of interconnected graphite/-sulphides and speaks to coherent along-strike tectonic processes, as for the NACP above. The DGC Central zone is a high T/low P metamorphic zone, bounded to the north and south by zones of intermediate temperature and pressure metamorphism, which define regions of structural thickening (Maloof, 2000; Gray et al., 2007). It is possible that during periods of accretion and marked shortening thickening, graphite was preferentially emplacement along grain boundaries, and deep shear movement facilitated graphite inter-connectivity, an argument that was also invoked by Ritter et al. (2003b). This argument is supported by Kisters and Neumaier (2004), who observed that during peak high temperature/low pressure metamorphism (circa 538 Ma-505 Ma) the central Damara zone underwent pure shear deformation with lateral-parallel orogen stretch. Furthermore, microscopic investigations in the DGC central zone reveal that graphite shows a strong lattice-preferred orientation in the mylonitic core zones. This strain-induced graphite, which might have been concentrated during collision of Congo and Kalahari cratons, would explain the large lateral (along strike) extent of the observed conductivity anomalies.
In summary, combining the electrical and metamorphic observations, the abundance of resistive S-type granites in the high-T Central Zone resulted from remelting of turbidites, facilitated by magmatic under-plating at lower crust levels, during the final collisional phase of the Congo and Kalahari cratons. The conductivity in the mid- to lower crust appears to have multiple possible causal sources, with graphite along deep-seated shear zones being the most probable and sulfide mineralization being more dominant in the upper crust.

7.5.1 Mantle lithosphere

Mantle electrical resistivity is more sensitive to temperature variations than it is to composition (Jones et al., 2009a), thus ‘colder’ and thicker cratonic lithosphere is generally well-imaged with the MT method and can be distinguished from ‘warmer’, generally thinner, mobile belts. The top of conductive zones overlain by resistive features, such as the boundary between cratonic lithosphere and asthenosphere, can be efficiently imaged to within 10% with high-quality MT data (Jones, 1999). For smoother models, such as those derived in our case, the lithosphere-asthenosphere transition is not sharply defined, but rather appears as a broad gradational boundary as consequence of smooth regularization.

There are several definitions of what the LAB is (see for example Eaton et al. (2009)). Mantle adiabatic temperature and composition will determine the electrical conductivity of the asthenosphere (Evans et al., 2011) but, for a completely dry peridotitic rigid lithosphere, the onset of the transition to asthenospheric conditions tends to occur at conductivities in the order of $10^{-2}$ S/m and greater (global averages are for conductivities of order 0.04 - 0.2 S/m for the asthenosphere (Jones, 1999)). While this transition is not sharply imaged for the central Damara belt, due to the shielding effect of the conductive mid-lower crust, for sites outside this conductive zone, where depth of penetration is larger, the boundary appears to be at depths of around 160 km.

Similar depth estimates were found by Muller et al. (2009). Fishwick (2010) estimated, from surface wave tomographic models, lithospheric thickness in the range 160 km - 180 km from the DGC. The DGC’s signal for the lithosphere-asthenosphere boundary (LAB) from the SRFs at station TSUM is not well resolved, due to high $2\sigma$ error bounds, but is nevertheless estimated at 180 km ± 20 km (see Figure 6.6 for the OKA profile for example). Thus all three methods, MT, surface wave tomography and SRF, give an LAB of 160 - 180 km.
Mantle xenoliths, showing lherzolitic and harzburgitic composition, found in the Cretaceous nephelinite plug within the Central Zone of the DGC, 17 km east of Swakopmund, suggest an elevated geotherm (90 mW/m² from Whitehead et al. (2002)) which is significantly higher than the Rehoboth Terrane (45 mW/m² from Muller et al. (2009)). This composition is typical for Proterozoic mantle. The elevated geothermal gradient in the Damara belt is possibly related to the post-collisional tectono-thermal events, evidence of which is widespread and in the form of granites particularly in the Central Zone. Furthermore, Whitehead et al. (2002) infers a heterogeneous, oxidized and fertile-to-refractory mantle beneath Damara Central zone. The xenolith geotherm is, however, based on relatively few samples, and, if representative at all, reflects upper lithospheric mantle temperatures at the time of kimberlite eruption (75 Ma) suggesting lateral temperature and lithospheric thickness variations from the Rehoboth (180 km) to the DGC (160 km). As geochemical and age constraints for the DGC lithospheric mantle are absent, our data and results provide no constraints on whether the DGC lithospheric-mantle might consist of a younger component, created and stabilized coeval with Pan African orogenesis.

What is immediately clear is that the interpreted boundary of the Congo craton by Corner (2008) and Begg et al. (2009) does not correlate with the mapped extension of the cratonic lithosphere from the electrical resistivity results. These authors have the margin close to MT station NENo22 on the NEN profile (Figure 7.1). The resistivity model suggests that this margin is approximately 100 km further south, close to station NEN118, and has a steep southward dip up to 80 km depth. The Tsumkwe (in Namibia) and Nxa-Nxau kimberlite fields on the border between Namibia and Botswana are located on an area of resistivity gradients which spatially correlates with lithospheric thickness change from a thinner mobile belt to a thicker craton (see the NEN model for example). A similar trend was remarked upon by Jones et al. (2009a) where, in general, diamondiferous kimberlites appear to occur on areas of lateral resistivity and velocity gradients at the edges of cratons, indicating shallowing of lithospheric roots.

The Okavango Dyke Swam, which crosses the OKA-WIN profile at an area between OKA008 and OKA004, is imaged as a resistive feature on this profile. Compositionally, the ODS is made up of dolerite dykes of varying widths ranging from 0.2 to 67 m, which Miensopust et al. (2011) showed manifest themselves as an anisotropic feature in 2D models. The ODS is hosted in similarly resistive granite-gneiss host rocks, such as the Kwando complex, the Rooibok Complex (analogous to Matchless Amphibolite in Namibia) and various Paleoproterozoic igneous basements obscured by younger strata.
The relation between seismic velocity and electrical resistivity has long been recognised. In the case of Southern Africa, Jones et al. (2009b) showed qualitatively that velocity gradients coincide with resistivity gradients for mantle lithosphere down to >150 km. Furthermore Jones et al. (2009c) showed that there is an advantage to combining seismic and EM data, given that conductivity is exponentially dependent on temperature whereas the shear and bulk moduli have only a linear dependence in cratonic lithospheric rocks.

To this end, a comparison is made between electrical resistivity data and fundamental mode Rayleigh group velocities. These data were constructed from tele-seismic data recorded with Africa Array stations shown in Figure 7.15. The group velocity dispersion curves (Figure 7.16) were constructed using event-station method for all event-station pairs in and around Southern Africa. The dispersion curves measured were inverted for 2-D group velocities maps, called group velocity tomography. The group velocities were extracted for each of the stations along profiles shown in Figure 7.1.

Vdovin et al. (1999) recognised that Rayleigh waves between 30 and 75 s are strongly sensitive to lower-crustal velocities and particularly crustal thickness. Furthermore

Figure 7.15: Africa Array seismic stations

(though magnetically distinct). Thus, while its field outcrops is well constrained, it is not readily distinguishable from its resistive hosts on the MT model.
7.6 Correlation of Resistivity and Seismic Velocity

a map of Rayleigh wave velocities at a period of 50 s can be inversely related to the Moho depth, which means that low velocities, at 50 s, result largely from thickened crust. At a 100 s, Rayleigh wave velocities map the uppermost-mantle (80-150 km) and periods up to 150 s are sensitive to depth of 200 km. Figure 7.17 shows the estimated group velocity maps across north central Namibia at various periods, indicated heterogeneities related to crustal (20 to 50 s) and upper mantle structure (>100 s). The 20 s, 75 s and 100 s maps all show a high velocity maps, which suggest deep seated root, but in comparison to the electrical resistivity results, this anomalous zone appears to be further north. The differences are ascribed to the sparse nature of the seismic stations, from which dispersion curves were derived. Furthermore the limited resolution of the phase velocity results, particularly for the longer periods, results "smooth" results. However the existence of the high velocity root, as suggested by the resistivity model, is supported. The low velocity anomaly in the west of the map in all maps is also recognised in the velocity map of Fishwick (2010) as shown in Figure 7.19. The resistivity and seismic results differ in this region in that the is no correlation of high resistivity (i.e. thick lithosphere) with high velocities as one would expect for craton regions. The differences are thought to be the result of the lack of seismic stations in northern Namibia. It is interesting to note that the LAB in this part of the Damara belt was estimated by Hansen et al. (2009) to be 160 km (Figure 7.20), from S-wave receiver functions, while seismic estimates of LAB from body-wave tomography were as low as 120 km (Fishwick, 2010)
Figure 7.17: Rayleigh-wave group-velocity maps, estimated for different periods. The dispersion curves were extracted for each MT station as such the spacing between stations places additional resolution limits to those imposed by event-station technique.
Figure 7.18: A comparison between seismic velocity perturbation, seismic lithospheric thickness estimates and electrical resistivity derived from 3D inversion.
Figure 7.19: Velocity model of Fishwick at 100 km
Figure 7.20: SRFs at two Africa Array stations located within the Damara belt (TSUM and LSZ). The black line represents the mean stack while the grey shaded areas indicate the $2\sigma$ bootstrap error bounds. The LAB and Moho depths are indicated but the LAB estimates are characterised by high errors (Hansen et al., 2009).
MT EVIDENCE OF HIGH ANGLE CONTINENTAL CONVERGENCE

In the next section a new tectonic model is derived based on the features observed in the MT results presented previously. The model derived is put in the Gondwana context.

8.1 THE ORIGIN OF THE DAMARA BELT CRUSTAL CONDUCTOR

A consistent feature in the MT results is the presence of a conductive structure in the Damara belt. Conductive zones in the crust are commonly known in rift zones such as in Rhine Graben and Rio Gande rift (Jiracek et al., 1983), but are also found in continental regions. Various possibilities exists that can account for crustal conductivity (1) inter-connected fluids, (2) metallic sulphides, graphite, hydrated minerals and/or (3) partial melt. In the case of the Damara Orogen, fluid inclusion studies (Walter, 2004) suggests the presence of CO$_2$-rich fluids in vein-quartz samples in the Brandberg region, NW Namibia. It is not enough to just have the presence of fluids, they must be inter-connected. Given the lack of knowledge about the inter-connectivity, the effect of fluids in enhancing conductivity cannot be ascertained.

There is no evidence of partial melt. The peak metamorphic temperatures were in the range from 600$^\circ$C – 700$^\circ$C, circa 530-500 Ma (Goscombe et al., 2005). The surface heat flow in the Damara belt is in the range 50-60 mWm$^{-2}$ (Nyblade, A. and Pollack, H., 1993), some 40-50% lower than areas with known partial melt or rifting in the crust, e.g., Rio-Grande rift (107 mWm$^{-2}$) or Cordilleran Rift (300 mWm$^{-2}$) (Lysak, 1992).

Combining the electrical and metamorphic observations Figure 8.1, the abundance of resistive S-type granites in the high-temperature Central Zone resulted from remelting of turbidites, facilitated by magmatic under-plating at lower crust levels, during the final collisional phase of the Congo and Kalahari cratons. The conductivity in the middle to lower crust appears to have multiple possible causal sources, with graphite...
Another consistent feature in the resistivity models is the thick, resistive cratonic lithosphere mapped in north-central Namibia. The Congo craton is characterized by very thick (to depths of 250 km) and resistive (i.e., cold) lithosphere. It was shown in the preceding sections that the thick cratonic lithosphere is pervasive throughout northern Namibia and north-western Botswana.

In general, the MT models suggest that the Congo craton lithosphere extends significantly southward beneath the Northern Platform sequences of Damara belt. The Damara belt lithosphere to be more conductive and significantly thinner than that of the adjacent Congo craton. The southern and northern DGC passive mar-

Figure 8.1: (a) Crustal-scale resistivity model of the ETO-KIM profile, showing the main features resolved by the inversion process. The mid-lower crustal conductors, upper crustal Pan-African granites and the Archean basement (Kamanjab Inlier) are the major features of this model. For comparison the geological section (b), modified from Gray et al. (2007), is included. AL: Autseib Lineament, OML: Omaruru Lineament, OKL: Okahanja Lineament, MA=Matchless Amphibolite, SFZ: Southern Foreland Zone.

along deep-seated shear zones being the most probable and sulfide mineralization being more dominant in the upper crust.

8.2 MAGNETOTELLURIC EVIDENCE OF THICK CONGO CRATON LITHOSPHERE

Another consistent feature in the resistivity models is the thick, resistive cratonic lithosphere mapped in north-central Namibia. The Congo craton is characterized by very thick (to depths of 250 km) and resistive (i.e., cold) lithosphere. It was shown in the preceding sections that the thick cratonic lithosphere is pervasive throughout northern Namibia and north-western Botswana.

In general, the MT models suggest that the Congo craton lithosphere extends significantly southward beneath the Northern Platform sequences of Damara belt. The Damara belt lithosphere to be more conductive and significantly thinner than that of the adjacent Congo craton. The southern and northern DGC passive mar-
gin sequences, which were deposited on a rift margin, appear to be overthrust onto both cratonic nuclei during continental amalgamation at 495 Ma, supporting studies by Goscombe (2004), to give a doubly vergent orogen. This result means that the current tectonic lines mapping the southern edge of the Congo craton need to be revised and drawn further south. The current tectonic geometry of the Damara

![Current tectonic structure](image)

Figure 8.2: The current tectonic structure from the Kaapvaal craton in the SW to the Congo craton in the NW. The dotted line shows the inferred lithosphere-asthenosphere boundary.

The base of the Damara Orogen electrical lithosphere is mapped at depths of around 160 km. The Kalahari craton segments (Rehoboth, Kaapvaal and Zimbabwe segments) have been mapped and their depth extends inferred from the MT data (>200 km). Mantle xenoliths, show lherzolitic and harzburgitic composition, with elevated geothermal gradient. This composition is typical for Proterozoic mantle. The elevated geothermal gradient in the Damara belt is possibly related to the post-collisional tectonothermal events, evidence of which is widespread and in the form of granites particularly in the Central Zone.

The xenolith geotherm is, however, based on relatively few samples and, if representative at all, reflects upper lithospheric mantle temperatures at the time of kimberlite eruption (75 Ma) suggesting lateral temperature and lithospheric thickness variations from the Rehoboth (180 km) to the DGC (160 km). As geochemical and age constraints for the DGC lithospheric mantle are absent, the MT data and results provide no constraints on whether the DGC lithospheric mantle might consist of a younger component, created and stabilized coeval with Pan-African orogenesis.
Combining the electrical and metamorphic observations, the abundance of resistive S-type granites in the high-temperature Central Zone of the DGC resulted from remelting of turbidites, facilitated by magmatic under-plating at lower crust levels, during the final collisional phase of the Congo and Kalahari cratons. The structural complexity of the DGC (i.e., numerous linear faults) including the large strike extent (50-60 degrees) supports the high-angle collision of the Kalahari and greater Congo Craton. There is a distinct lithospheric depth variation from the thick Archean cratons in the south to the thinner DGC and thicker Congo craton in the north. Figure 8.3 is an attempt at an evolutionary model of the Damara orogen in relation to the Congo and Kalahari cratons. The closure of the basin between the Kalahari and Congo craton (A), resulted from the northward migration of the Kalahari and the southern migration of the Congo craton. During the ensuing collision, magmatic underplating in the lower crust resulted in the emplacement of the intrusives in the Damara Central Zone (B). It is at this stage that deep seated crustal structures were developed within the Damara orogen. The present day tectonic configuration (C) is one where the
8.4 Global Context: Comparison to Similar Tectonic Environments

It is interesting to note the striking similarity between our MT models and other Archean-Proterozoic margins around the world in imaging steeply conductive features separating the two terrains. For example, the Central Australian Suture Zone, which forms the contact between the North Australian craton and the Proterozoic Warumpi province (Selway et al., 2006, 2009), is mapped as a dipping conductor. Also, in Australia, the Errabiddy Shear Zone forms a conductive suture between the Yilgarn Craton and the Glenburgh Terrane (Selway et al., 2006, 2009) (Figure 8.4). The margin between the Slave craton and the Wopmay orogen (Spratt et al., 2009; ?) is preserved as steeply dipping conductive structure. In the case of the Damara orogen, the Autseb Lineament separates Archean terrain from the younger Proterozoic segment.

All of these features were subjected to later reactivation post-continental collision. This observation suggests that steeply dipping crustal-scale conductive structures are...
perhaps characteristic of the Proterozoic-Archean margins and can be used to infer continental accretion. This means that perhaps broad generalization can be inferred about similar mode of evolution for Archean/Proterozoic margins.

**8.5 Is there incipient rifting in the Kalahari?**

There has been some suggestions that the East African Rift System extends into north-western Botswana, based on persistent seismicity (Reeves, 1972) (Figure 8.5), forward modeling of magnetic and gravity data (Kinabo, 2007) and anomalous heat flow (Chapman and Pollack, 1974). All geophysical experiments done on the Okavango

![Figure 8.5: Earthquake epicentres plotted on a Google Earth image, showing the distribution (clusters shown in white polygon) of earthquakes over a 20 year period. The location of the OKA-WIN profile is indicated. Their appears to be a linear trend of earthquakes associated with those on the East African Rift valley.](image)

Rift Zone (ORZ), before ours, focussed on shallow (less than 1 km) structure (Modisi et al., 2000; Kinabo, 2007), and revealed the ORZ as a half-graben feature.

If indeed continental rifting is taking place, then one might expect a classic rift
signature (Figure 8.6) of thinned lithosphere accompanied by a localised deep seated high conductivity mantle anomalies, as is observed for example in the Rio Grande Rift (Hermance and Pedersen, 1980; Jiracek et al., 1983; Hermance and Neumann, 1991). None of these features are observed on the electrical resistivity models, particularly along the OKA-WIN profile. The lack of elevated mantle conductivity suggests that incipient continental rifting, if taking place, is generated from pre-existing structural features in the crust, as proposed by Birt et al. (1997) for the Kenya Rift, and propagates downward without any mantle upwelling signature.

It is must be mentioned that more recently detailed seismic data was collected in Botswana by the Woodshole Oceanographic Institute (WHOI) in collaboration with the Dublin Institute for Advanced Studies and the Geological Survey of Botswana, to study incipient rifting. These results have not been published as yet. It is hoped that these result will add significantly in understanding the not only rifting in the Kalahari, but the global processes that can explain rifting all over the world.
Part III

CARBON CAPTURE AND STORAGE
Carbon capture and storage (CCS), as the name implies, broadly refers to the process of collecting carbon dioxide (CO₂) released by industrial power plants and injecting it in geologically suitable formations in the Earth’s upper-most crust for long-term storage. Sometimes called carbon sequestration, CCS is the most practical and efficient process for mitigating the emmissions of greenhouse gases. Geologically suitable formations in this context refers to high porosity rocks (e.g., sandstones) at minimum depth of 800 m, overlain by low porosity cap-rocks (e.g., shales). In addition to capturing and storage of carbon, mechanisms must be put in place for the monitoring after injection.

In 2007 the Inter-governmental Panel on Climate Change (IPCC) released a report (Pachauri and Reisinger, 1997) which was meant to synthesize the changes in global climate including the effects of that change on the natural ecosystems and society in general. The major findings of that work was that global greenhouse emmissions have increased since the dawn of industrialization, with 70% increase between 1970 and 2004. The report recommends CCS in geological formations as one of the efficient means of curbing the emmissions.

To this effect South African (SA) government made significant strides in the establishment, in 2009, of a Centre for Carbon Capture and Storage, a division of the South African National Energy Development Institute (SANEDI, formerly SANERI), which is tasked with undertaking CCS research and technical development. This was followed, in 2010, by a publication of an Atlas (Cloete, 2010), which identified and ranked potential CO₂ storage sites largely in Mesozoic basins along the coast (Outeniqua, Orange and Durban/Zululand basins) and the Karoo basin (see also the technical report of Viljoen et al. (2010)). The results contained in the atlas were derived largely on literature review of all available historic boreholes, petrophysical, geological and geophysical data. Detailed work on individual basins has not been done and the authors of the atlas recognised the need of a quantitative study to be...
carried out on the proposed CCS sites in an effort to image and characterise their potential storage capacity.

In this context, the purpose of this chapter is to report on the research conducted using MT data collected in the Karoo Basin over one of the identified CCS sites. A key physical parameter in assessing storage potential of rocks is porosity. MT data is used to investigate the porosity and permeability properties of Karoo strata and to determine whether they can be used to store anthropogenic gases at depth.

9.1 WHAT IS CARBON CAPTURE AND STORAGE?

CCS is analogous to how oil and gas are naturally trapped in underground formations. The same geological formations that have allowed oil/gas to be stored for thousands of years below the surface can be used to store CO₂. Indeed injecting CO₂ underground has been used to enhance oil/gas recovery (EOR) in some fields. An ideal CCS site must comprise a porous and permeable media (e.g., sandstone) where CO₂ can be injected and stored, overlain by an impermeable cap rock (e.g. shale) that will retain (by dissolution or adsorption) the injected material and prevent it from moving escaping into the atmosphere (Benson and Cook, 2005; Cloete, 2010; Viljoen et al., 2010).

Geological storage options for deep injection of CO₂ include depleted oil and gas formations, deep unmineable coal seams and deep saline formations (Figure 9.1). The storage of CO₂ is currently being done in a number of projects worldwide, e.g., the Sleipner oil field in North Sea (Saline formation storage), the Weyburn project in Canada (EOR), the CO₂CRC Otway Project in Australia, the Salah project in Algeria (EOR) and the Shute Creek Gas Project (EOR) in Unites Stated of America (Benson and Cook, 2005; USA Department of Energy, 2010).

As one might expect, storing CO₂ in depleted oil/gas fields is naturally the most attractive for several reasons. Firstly, the structural integrity of the formations and traps is demonstrated and the oil and gas that originally accumulated in traps did not escape (in some cases for many millions of years). Secondly, the geological structure and physical properties of most oil and gas fields have been extensively studied and characterized (Benson and Cook, 2005).

Storage of CO₂ in deep saline formations that are saturated with aqueous ground fluids. These formations are widespread and contain enormous quantities of water,
what is carbon capture and storage?

Figure 9.1: CCS storage cycle, including and illustration of possible CO$_2$ storage options in underground geological formations (Benson and Cook, 2005)

but are unsuitable for agriculture or human consumption (Viljoen et al., 2010).

CO$_2$ storage in **unmineable coal seams** involves injecting gaseous CO$_2$ through wells, which will flow through the fracture system of the coal, diffuse into the coal matrix and be adsorbed onto the coal micropore surfaces, freeing up gases with lower affinity to coal (i.e., methane) (Benson and Cook, 2005).
A natural question to ask is why is MT suitable for investigating the CCS potential of subsurface sedimentary strata. The resistivity of a rock formation is a function of four parameters: (1) the porosity of the rock that is occupied by a fluid, (2) the degree of interconnection of the fluid, (3) the resistivity of the host rock and (4) the salinity of the groundwater (Unsworth et al., 2007a). As was pointed out earlier, the porosity and permeability of rocks is a key physical property that must be quantified.

In 1942, Archie (1942) observed that the resistivity of saturated bulk rock $\rho_r$ is directly proportional to the resistivity of the aqueous fluid $\rho_w$ in its pores resistivity, given by the relation

$$\rho_r = F \rho_w. \quad (9.1)$$

where $F$ is called the Formation factor which describes the effect of the presence of the rock matrix (Glover et al., 2000; Glover, 2000). For a material with 100% porosity, $F = 1$. In rocks $F$ takes values usually between 20 and 500. Note that formation factor has is unit-less because it is the ratio of two resistivities. The formation factor includes both the effect of the variable porosity and the effect of the tortuous pathways that the current is forced to take through the conducting fluid due to the presence of the insulating rock grains. It can be seen, therefore, that the formation factor is related to the porosity of the rock and the connectivity of the pore spaces (permeability). The natural complexity of pore systems in rocks means that the formation factor cannot be expressed simply as a function of porosity and connectivity in a theoretically rigorous way (Glover, 2007).

Intrinsic (or specific) permeability $\kappa_i$ is represented by the relation (Bates and Jackson, 1980)

$$\kappa_i = C d^2 \quad (9.2)$$

where $C$ is the dimensionless constant related to configuration of the flow paths and $d$ is the average or effective pore diameter. The units of measurement for permeability are m$^2$ or the more commonly known non-SI units Darcy (D) or milli-Darcy (mD); 1mD = 10$^{15}$m$^2$. The permeability of sandstone is in the range 0.001 - 700 mD, while that of shale varies in the range 0.00001-0.1 mD. For purposes of CO$_2$ storage, the permeability of the reservoir rocks (sandstone) should be as high as possible while that of the caprock should be as low as possible to prevent CO$_2$ migration from the reservoir (Viljoen et al., 2010). Archie (1942) also observed that an empirical relation existed between the Formation Factor $F$ and the rock porosity ($\phi$) such that

$$F = \phi^{-m}, \quad (9.3)$$
Table 9.1: Cementation factor for different consolidated and unconsolidated (sandstones) media

<table>
<thead>
<tr>
<th>Cementation factor (m)</th>
<th>Material</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.0-1.4</td>
<td>Igneous and metamorphic rocks</td>
</tr>
<tr>
<td>1.4-2.0</td>
<td>Sandstones (higher values in consolidated sandstones)</td>
</tr>
<tr>
<td>2.0-2.6</td>
<td>Carbonates</td>
</tr>
</tbody>
</table>

where m is the cementation factor describing how consolidated the sediments are. A rock with 100% porosity for instance, the cementation factor is zero. Table 9.1 shows typical values of m for various lithologies. A constant "a" is sometimes placed before the porosity term in equation 9.4 such that \( F = a\phi^{-m} \) and so \( F = a \) when \( \phi = 1 \), however, as was noted by Glover (2000) there is no physical justification for this term. It arises from applying a best fit equation to \( F \) versus porosity data, and should be avoided.

Equations 9.1 and 9.3 are often combined resulting in the relation

\[
\rho_r = \rho_w \phi^{-m}.
\]

in what is known as Archie’s first law. Thus by knowing the resistivity of geological formation \( \rho_r \) and that of aqueous fluids \( \rho_w \) we can, in principle, determine rock properties (porosity and permeability) needed for reservoir characterization using Archie’s Law given in equation 9.4.

Figure 9.2 shows a theoretical computation of rock resistivity against porosity for varying m constants. It is clear that, in general, porosity decreases with increasing resistivity. The resistivity of the earth generally increases with depth in the crust. Porosity decreases as a result of pressure induced compaction of rocks. MT measurements allow us to estimate the resistivity of geological formation \( \rho_r \). The resistivity of groundwater (\( \rho_w \)) can be estimated from borehole data or using established empirical relations. The relation given by Block (2001), for instance, allows \( \rho_w \) to be estimated from the salinity or Total Dissolved Solution (in ppm) data such that

\[
\rho_w = 4.5(TDS)^{-0.85}
\]

Meju (2000) also found the following empirical relation to hold

\[
\rho_w = 6139 \times TDS(ppm)^{-0.9852}
\]

where TDS refers to the Total Dissolved Solids, which is a measure of salinity. It is thus, possible to estimate the porosity using equation 9.4.
Porosity is a fraction between 0 and 1 (or 0-100%). Typically it ranges from less than 0.01 for solid granite to more than 0.5 for coarse grained unconsolidated material (Table 9.2). According to Bell (1983) sedimentary rocks with round granules, in the absence of bounding matrix, can theoretically have porosities in the 26 to 48 %, but this reduces to less than 26 % in real rocks as they generally deformed and compacted. In section 13.1, TDS estimated from borehole located in the Karoo basin will be used to determine formation water resistivity, which in turn is used to estimate porosity, given the MT derived bulk resistivities. Furthermore, two dimensional inversion and modelling techniques will be used to provide the electrical resistivity structure of the Karoo sediments in the top 5-10 km. The data will then be inverted assuming a one-dimensional earth and a set of resistivities obtained. These resistivities are
<table>
<thead>
<tr>
<th>Medium</th>
<th>Porosity Range</th>
<th>m</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>clean sand</td>
<td>0.12-0.40</td>
<td>1.3</td>
<td>Archie (1942)</td>
</tr>
<tr>
<td>consolidated sandstones</td>
<td>0.12-0.35</td>
<td>1.8-2.0</td>
<td></td>
</tr>
<tr>
<td>glass spheres</td>
<td>0.37-0.40</td>
<td>1.38</td>
<td>Wyllie and Gregory (1955)</td>
</tr>
<tr>
<td>binary sphere mixtures</td>
<td>0.147-0.29</td>
<td>1.31</td>
<td></td>
</tr>
<tr>
<td>cylinders</td>
<td>0.33-0.43</td>
<td>1.47</td>
<td></td>
</tr>
<tr>
<td>disks</td>
<td>0.34-0.45</td>
<td>1.46</td>
<td></td>
</tr>
<tr>
<td>cubes</td>
<td>0.19-0.43</td>
<td>1.47</td>
<td></td>
</tr>
<tr>
<td>prisms</td>
<td>0.36-0.52</td>
<td>1.63</td>
<td></td>
</tr>
<tr>
<td>8 marine sands</td>
<td>0.35-0.50</td>
<td>1.39-1.58</td>
<td>Jackson et al. (1978)</td>
</tr>
<tr>
<td>glass beads (spheres)</td>
<td>0.33-0.37</td>
<td>1.2</td>
<td></td>
</tr>
<tr>
<td>quartz sand</td>
<td>0.32-0.44</td>
<td>1.43</td>
<td></td>
</tr>
<tr>
<td>rounded quartz sand</td>
<td>0.36-0.44</td>
<td>1.4</td>
<td></td>
</tr>
<tr>
<td>shaley sand</td>
<td>0.41-0.48</td>
<td>1.52</td>
<td></td>
</tr>
<tr>
<td>shell fragments</td>
<td>0.62-0.72</td>
<td>1.85</td>
<td></td>
</tr>
<tr>
<td>fused glass beads</td>
<td>0.02-0.38</td>
<td>1.5</td>
<td>Sen et al. (1981)</td>
</tr>
<tr>
<td>fused glass beads</td>
<td>0.10-0.40</td>
<td>1.7</td>
<td>Schwartz and Kimminau (1987)</td>
</tr>
<tr>
<td>sandstone</td>
<td>0.05-0.22</td>
<td>1.9-3.7</td>
<td>Doyen (1988)</td>
</tr>
<tr>
<td>polydisperse glass beads</td>
<td>0.13-0.40</td>
<td>1.28-1.40</td>
<td>de Kuijper et al. (1996)</td>
</tr>
<tr>
<td>fused glass beads</td>
<td>0.10-0.30</td>
<td>1.6-1.8</td>
<td>Pengra and Wong (1999)</td>
</tr>
<tr>
<td>sandstones</td>
<td>0.07-0.22</td>
<td>1.6-2.0</td>
<td></td>
</tr>
</tbody>
</table>

Table 9.2: Porosity range and Archie’s exponent m (cementation factor) for different consolidated and unconsolidated media (from Lesmes and Friedman (2005))

then converted to apparent porosities using Archie’s law shown in equation 9.4. The modelling of the data in 1D is valid since it is known that the Karoo basin stratigraphy comprises layered sedimentary rocks, a variety of sandstones inter-bedded with shales, overlain by Drakensberg basalts (see Figures 10.1, 10.2).

Another method in which electrical resistivity can be used to determine porosity of rocks is that of using nomograms. This principle is illustrated in Figure 9.3, which shows a porosity-resistivity-salinity nomogram that can be used to estimate porosity from bulk resistivity measurement. Thus if one has temperature/salinity
measurements (say from boreholes) and resistivity from MT results, plotting these on the nomogram and connecting them with a best fit line would yield the resistivity of the pore fluid, which would in turn be used to estimate porosity. In combination with existing geological knowledge of the sedimentary basins, this research will develop new Karoo-specific CO₂ storage geological model based on electrical conductivity data and characterise potential carbon storage sites for areas around Lesotho, that have already being identified as potential storage site.

MT cannot give the same vertical detail as a seismic reflection survey. The best resolution is achieved in MT when determining the depth to the top of a conducting layer. MT data can however, provide valuable information on bulk material properties. Furthermore, while the conductivity and thickness are not readily resolved with MT data, the conductance (i.e., the conductivity-thickness product) can be efficiently determined. There other mixing models that can be used to estimate porosity of substances from resistivity values (Figure 9.4).

Figure 9.3: Porosity-resistivity-salinity nomogram that can be used to estimate porosity from bulk resistivity measurement.
### 9.2 MT Application in Carbon Capture and Storage

#### Figure 9.4: Summary of the main electrical conductivity mixing models that are currently available (modified from Glover (2007))

<table>
<thead>
<tr>
<th>Conducting phases</th>
<th>Equation</th>
<th>Reference</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Parallel model</strong></td>
<td>( \sigma_{\text{eff}} = \sum_{i=1}^{n} \phi_i \sigma_i )</td>
<td>Guégan and Palczanskas (1994), Luo et al. (1994), Somerton (1992)</td>
<td>Arithmetic mean. Parallel layers of constant arbitrary thickness with conductivity ( \sigma_i ) arranged axially to current flow.</td>
</tr>
<tr>
<td><strong>Perpendicular model</strong></td>
<td>( \frac{1}{\sigma_{\text{eff}}} = \sum_{i=1}^{n} \frac{1}{\phi_i \sigma_i} )</td>
<td>Guégan and Palczanskas (1994), Luo et al. (1994), Somerton (1992)</td>
<td>Harmonic mean. Parallel layers of constant arbitrary thickness with conductivity ( \sigma_i ) arranged normally to current flow.</td>
</tr>
<tr>
<td><strong>Random model</strong></td>
<td>( \sigma_{\text{eff}} = \frac{1}{n} \sum_{i=1}^{n} \sigma_i^h )</td>
<td>Guégan and Palczanskas (1994), Luo et al. (1994), Somerton (1992)</td>
<td>Geometric mean. Arbitrary shaped and oriented volumes of conductivity ( \sigma_i ) distributed randomly.</td>
</tr>
<tr>
<td><strong>Hashin-Shtrikman upper bound</strong></td>
<td>( \sigma_{\text{eff}} = \frac{\phi_1 \sigma_1 + (1 - \phi_1) \sigma_2}{\phi_1 + (1 - \phi_1) \sigma_2} )</td>
<td>Hashin and Shtrikman (1962)</td>
<td>Commonly denoted HS+. Derived from effective medium considerations.</td>
</tr>
<tr>
<td><strong>Hashin-Shtrikman lower bound</strong></td>
<td>( \sigma_{\text{eff}} = \frac{\phi_1 \sigma_1 + (1 - \phi_1) \sigma_2}{\phi_1 \sigma_1 + (1 - \phi_1) \sigma_2} )</td>
<td>Hashin and Shtrikman (1962)</td>
<td>Commonly denoted HS-. Derived from effective medium considerations.</td>
</tr>
<tr>
<td><strong>Waff model</strong></td>
<td>( \sigma_{\text{eff}} = \frac{\sigma_1 + (1 - \phi_1) \sigma_2}{1 + (1 - \phi_1) \sigma_2} )</td>
<td>Waff (1974)</td>
<td>Based on concentric spheres of varying sizes with volume of core (fractional volume of phase 1) to volume of shell (fractional volume of phase 2) ratio constant. Functionally equivalent to HS+.</td>
</tr>
<tr>
<td><strong>Modified brick-layer model</strong></td>
<td>( \sigma_{\text{eff}} = \frac{\sigma_1 (\sigma_2 \phi_1^3 - 1) - \sigma_2 \phi_2^3}{\sigma_1 (\phi_1 - \phi_1^3) - \sigma_2 (\phi_2 - \phi_2^3)} )</td>
<td>Schilling et al. (1997)</td>
<td>Modified to allow validity to be extended to cover the range 0.00 ( \leq \phi_i \leq 1.00 ) (0%-100%). Almost coincident with HS+.</td>
</tr>
</tbody>
</table>

**Models with variable exponents**

| Lichteneker-Rother equation | \( \sigma_{\text{eff}} = (\sigma_1^{n_1} (1 - \phi_2) + \sigma_2^{n_2} \phi_2)^{\frac{1}{n}} \) | Lichteneker and Rother (1936), Kervin (1982) | Derived from the theory of functional equations under appropriate boundary conditions. Formally the same as Archie’s law if \( \phi_2 = 0 \). |
| Lichteneker-Rother equation (generalized) | \( \sigma_{\text{eff}} = (\sum_{i=1}^{n} \phi_i^{n} \sigma_i)^{\frac{1}{n}} \) | Lichteneker and Rother (1936), Kervin (1982) | Logical extension of the Lichteneker and Rother model made in this work. |
| Russian equation | \( \sigma_{\text{eff}} = \sigma_1 \phi_1^{\frac{1}{n}} \) | Bussign (1983) | Derived from effective medium theory. |
| Conventional Archie’s law | \( \sigma_{\text{eff}} = \sigma_1 \phi_1^{p} \) | Archie (1942) | Derived empirically, but provable analytically for special cases. |
| Modified Archie’s law | \( \sigma_{\text{eff}} = \sigma_1 (1 - \phi_2)^p + \sigma_2 \phi_2^p \) | Glover et al. (2006a) | Derived from the conventional Archie’s law by considering boundary conditions implied by geometrical constraints. |

where \( p = \frac{\log(1 - \phi_2)}{\log(1 - \phi_1)} \).
10.1 CARBON CAPTURE AND STORAGE IN SOUTH AFRICA

South Africa has huge amounts of coal reserves that contribute immensely (75 percent) to the total national and international energy consumption. The country also emits over 400 million tonnes of CO$_2$ annually year of which approximately 60 percent are capturable and therefore potentially available for carbon capture and storage (Cloete, 2010). The change in climate/weather patterns is thought to be caused by the accumulation of greenhouse gases, such as carbon dioxide in the Earth’s atmosphere.

In order to reverse this trend, a novel approach is to store anthropogenic CO$_2$ in subterranean geological formations, on or offshore. The final step in the process involves continuous monitoring to ensure carbon stays in the geological formations. The requirements for an ideal site to store CO$_2$ are (Bachu, 2000, 2003; Benson and Cook, 2005):

- **porous and permeable** rock reservoir rock to allow injection and storage, overlain by
- **impermeable seal rock** to allow retention of the stored gas,
- adequate capacity and injectivity,
- a sufficiently stable geological environment to avoid compromising the integrity of the storage site,
- basin characteristics (tectonic activity, sediment type, geothermal and hydrodynamic regimes)

In South Africa, there are very few onshore sites/localities that have been identified which have the potential to store CO$_2$; one of these is the Karoo Basin which is the largest sedimentary basin in South Africa as it occupies 60% of the landmass (Figure 10.1).
10.1.1 The Karoo Basin

The Late Carboniferous to Early Jurassic Karoo Basin (Figure 10.1) is a 12 km thick sedimentary basin underlain by a relatively stable basement, on which glacial deposits and sediments were accumulated. The Cape Fold Belt forms along the southern margin of the Karoo basin. Lithologically, the Karoo basin comprises up to 6 km of clastic sedimentary strata capped by at least 1.4 km of basaltic lava (Figure 10.2) (Smith, 1990) and (Svensen et al., 2006)). The sediments were deposited from the Late Carboniferous to the Mid-Jurassic (Figure 10.3), in an environment ranging from partly marine (the Dwyka and Ecca groups), to fluvial (the Beaufort Group and parts of the Stormberg Group) and aeolian (upper part of the Stormberg Group) (Svensen et al., 2006). Dolerite dykes and sills intruded the Karoo sediments during the Jurassic period. While the rest of the basin is relatively undeformed the south-western part is folded as a result of the Cape orogeny (Figure 10.2). The Karoo Basin has extensive coal deposits, which gives rise to the possibility that they can be used for CO₂ storage or for enhancing methane recovery (Viljoen et al., 2010). While several oil/gas exploration boreholes have been drilled in the basin, none has returned significant

![Figure 10.1: Distributions of litho-stratigraphic units in the Main Karoo Basin (reproduced from Woodford and Chevallier (2002))](image-url)
Figure 10.2: Cross Section of the Main Karoo Basin (reproduced from Woodford and Chevallier (2002)).

<table>
<thead>
<tr>
<th>Supergroup</th>
<th>Group/Formation</th>
<th>Age</th>
<th>SA1/66 Borehole Intercept</th>
<th>Basin evolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Karoo Supergroup</td>
<td>Beaufort Group</td>
<td>258-228 Ma</td>
<td>457 m</td>
<td>Fluvial accumulation</td>
</tr>
<tr>
<td></td>
<td>Ecca Group</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Upper Ecca</td>
<td>270-276 Ma</td>
<td>1280 m</td>
<td>Turbidite influx (from south)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lower Ecca</td>
<td>279 Ma</td>
<td>1280 m</td>
<td>Lacustrine deposition</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Dwyka Group</td>
<td>289 Ma-342 Ma</td>
<td>(borehole ends in Dwyka) 3100 m</td>
<td>Post-glacial marine transgression</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cape Supergroup</td>
<td></td>
<td>500 Ma</td>
<td></td>
<td>Local glacial diamictites top of Witteberg Group</td>
</tr>
</tbody>
</table>

Figure 10.3: Stratigraphic chart with the major litho-stratigraphic subdivisions of the Karoo Supergroup in the main Karoo Basin of South Africa (after Catuneanu et al. (2005)).

economic deposits (Van Vuuren et al., 1998). A desktop study was compiled by the Council for Geoscience (SA) to identify possible sites (see the South African Atlas on
CO₂ sequestration, Cloete (2010)). Porosity and permeability of geological formations in the lower Karoo were found to be rather low (<26%) by international standards for sequestration sites (See Figure 10.4 and 10.5). Woodford and Chevallier (2002) found

<table>
<thead>
<tr>
<th>Bore No.</th>
<th>Formation analyzed</th>
<th>Depth interval (m)</th>
<th>No. of samples</th>
<th>Percentage porosity (average)</th>
<th>Permeability range (md air)</th>
</tr>
</thead>
<tbody>
<tr>
<td>KL1</td>
<td>Beaufort Ecca</td>
<td>14 - 1341</td>
<td>41</td>
<td>1.9</td>
<td>0.6 - 1.9</td>
</tr>
<tr>
<td>SA1</td>
<td>Beaufort Ecca</td>
<td>34 - 1215</td>
<td>84</td>
<td>0.5</td>
<td>0.2 - 2.9</td>
</tr>
<tr>
<td>SK2</td>
<td>Ecca</td>
<td>741 - 1361</td>
<td>3</td>
<td>0.8</td>
<td>0.4 - 0.5</td>
</tr>
<tr>
<td>SPI</td>
<td>Beaufort Ecca</td>
<td>401 - 1938</td>
<td>12</td>
<td>0.5</td>
<td>0.0 - 0.1</td>
</tr>
<tr>
<td>MA1</td>
<td>Beaufort Ecca</td>
<td>427 - 1390</td>
<td>29</td>
<td>8.8</td>
<td>3.1 - 18.1</td>
</tr>
<tr>
<td>LA1</td>
<td>Ecca</td>
<td>1540 - 1600</td>
<td>67</td>
<td>5.6</td>
<td>12.1 - 11.2</td>
</tr>
<tr>
<td>MA1 + W1</td>
<td>Beaufort Ecca</td>
<td>620 - 635</td>
<td>7</td>
<td>1.8</td>
<td>1.8</td>
</tr>
<tr>
<td>SMA</td>
<td>Beaufort Ecca</td>
<td>80 - 850</td>
<td>77</td>
<td>6.2</td>
<td>1.3 - 12.5</td>
</tr>
<tr>
<td>ME1</td>
<td>Middle Ecca</td>
<td>780 - 900</td>
<td>18</td>
<td>11.0</td>
<td>7.7 - 12.8</td>
</tr>
</tbody>
</table>


Figure 10.4: A: Porosity and Permeability of sandstones and siltstones obtained from deep boreholes located in the Main Karoo Basin (after Roswell and De Sward, 1976), table from Woodford and Chevallier (2002). B: Drill information from shallow core boreholes in the Karoo Basin (Woodford and Chevallier, 2002)

that porosity occurs along bedding planes between sedimentary layers. However this observation was limited to the few hundred meters (200 m) below surface, and as
10.1 CARBON CAPTURE AND STORAGE IN SOUTH AFRICA

Figure 10.5: Map showing porosity and permeability trends, estimated from boreholes located in the Karoo basin (modified from (Viljoen et al., 2010) and (Cloete, 2010)). Boreholes MA1-69 and WE1-69 are located within the proposed storage site with porosities on the former estimated to be in the range 4-10%.

such is not a good indicator of porosity in all Karoo basin at depths below 200 m. But given that no quantitative study has been done anywhere in South Africa to assess the basin to store carbon, the potential for CO₂ sequestration is underestimated in many respects.

A recent study of the Karoo Basin was completed by Shell (Golder, 2011), in an effort to characterise the basin for gas potential, found that the main lithologies are characterised by very low primary porosity (<0.1% to <0.02%) and permeability. Deep core holes (up 5.5 km) drilled in the late 1960’s (Botha et al., 2002) confirmed that the permeabilities of the sandstone/siltstone are very low, and vary from 0.00 to 2 mD for the Beaufort Group and 0.00-0.1 mD for Ecca Group.

Perhaps the biggest geological limitation to carbon sequestration in the Karoo basin is the presence of numerous dolerite dykes/sills as these complicate the hydrological fluid processes/flow and can also divide the basin into relatively small compartments.
or confined aquifers. Furthermore, faulting and deformation is often caused by dolerite intrusions, which increases the potential for fluid escape/migration along fault induced pathways (Botha et al., 2002; Woodford and Chevallier, 2002). Saghafi et al. (2008) noted however, that dolerite induced metamorphic process (i.e., heating) enhances the adsorption and a secondary porosity properties of the coal seams, which result in the increase in their gas storage ability. Dolerite intrusions would, according to Saghafi et al. (2008), act as stratigraphic gas traps. As was noted by Van Vuuren et al. (1998) and Viljoen et al. (2010), the discovery of oil occurrences in fractures next to dolerite dykes supports the observation that dykes can act as stratigraphic/structural traps.

10.1.2 Identification and characterisation of storage sites in the Karoo basin using MT data

The Atlas on the carbon sequestration potential in South Africa (Cloete, 2010) identified several sites onshore and offshore, shown in Figure 10.6. These localities were

![Map of South Africa showing possible storage sites](image)

Figure 10.6: Possible deep saline formation storage opportunities onshore and offshore in Mesozoic basins along the coast of South Africa and for deep coalfields of the Karoo basin (Compiled by the Council for Geoscience in Cloete (2010)).

classified and ranked according to storage potential, based on review of available
geoscientific data. In the Karoo basin, the areas identified occur on the SW and NE margins of the Lesotho border, was given fairly low CO₂ capture potential for its low porosity and permeability, including the presence of dykes and sills. It is in the SW region that MT data exists and for this reason this area is the focus of this research.

There are over 750 MT sites covering most of the South Africa’s sedimentary basins (including Karoo, Wits, Kalahari basins). MT data has been used at various scales to map the electrical conductivity of the subsurface. Unlike the lithospheric studies (see section 3 for details), this investigation will focus on short period (i.e., shallow depth) data that will allow detailed mapping of the uppermost crustal units.

The MT data that will be used were collected in the summer of 2008 in the fourth phase of SAMTEX. These data form part of the KIM-NAM profile (Muller et al., 2009) and were added with the sole purpose of mapping the south-eastern boundary of the Kaapvaal craton (Figure 10.7). The MT stations that will be used are labelled

![Figure 10.7: MT profile plotted on litho-stratigraphic map of the Karoo basin](image)

KXT (for Kaapvaal Extension) and are shown in Figure 10.8. The MT stations are less than 100 km from the Lesotho border and the average spacing between the sites is
18 km. There are about 20 MT stations that will be used for this research but only 11 of those are located over the potential storage site outlined in the Atlas. As a result only 12 stations, KXT001 to KXT010 and KIM020 and KIM021 will be used for formal 1D and 2D inversion. The profile transcends most of the Karoo lithologies and the stratigraphy in this region is well constrained by deep borehole (MA1-69), the location of which is shown in Figure 10.8.

The geological section in this part of the Karoo is shown in Figure 10.9 and illustrates the layered nature of the lithologies. The MT profile is highlighted on the section and main sedimentary layers are those of the Beaufort, Ecca and Dwyka groups. The Drakensburg basalts do not form part a big component of the geology at the location of the MT profile. The "expected lithologies" to be mapped are thus sandstones and shales of the Beaufort and Ecca group and the diamictite of the Dwyka group.

Given the layered nature of the stratigraphy in this region, MT data will be modelled...
Figure 10.9: Karoo geological section from south-west to northeast section through the Cape and Karoo Basins, with the location of the MT profile highlighted. Note that section does no include the Drakensburg and Group lithologies. The Dwyka Group is very thin in the section (Viljoen et al., 2010).

using 1D techniques in order to address a few questions:

- What is the depth to basement beneath the Karoo basin?
- Can MT assist in the estimation of the porosity and permeability of the Karoo sediments?
- What is the thickness of the individual lithologies?
- Could the Karoo basin, in general, be suitable for CCS?
11.1 DATA PROCESSING

The same data processing methodologies described in section 2.10 were used to derive the apparent resistivity and phase data (as a function of period) from the recorded time series. The recorded electromagnetic time series data were processed using robust techniques of Jones and Jodicke (1984), (method 6 in Jones et al. (1989)), Egbert and Booker (1986) and the heuristic approach of Jones et al. (1989), as implemented in the Phoenix processing software SSMT 2000, to derive MT impedance estimates that were subsequently converted to apparent resistivities and phases.

Data quality was generally good for most sites; however where severe distortion (e.g., current channelling) effects were observed that could not be treated, those sites were not used for further modelling (e.g., site KXT06 was severely distorted). Figure 11.1 shows examples of MT responses from six sites along the profile. The Rho+ algorithm of Parker (1980) was used to check for internal consistency between the apparent resistivity and phase responses, including discarding spurious data points.

Unlike in the lithospheric studies where longer period MT data were modelled, only data in the period range $10^{-2} - 10^1$ s were used for uppermost crustal studies for CCS. The depth of penetration for each site was estimated using the Nibblett-Bostick approach already described in equation 2.39. The resistivity of Karoo sediments, in the depth range 0-3 km, from previous resistivity studies of van Zijl (2006) appears to vary from 500 - 1200 Ohm.m. This results in the average NB depth of penetration of over 10 km for the period range modelled. In the case of the Karoo it is appropriate to use 1D depth approximations, like Niblett-Bostick, give the layered nature of the stratigraphy.
11.2 Strike Analysis

The single site decomposition and analysis technique of *Groom and Bailey (1989)* and multi-site and multi-period extension by *McNeice and Jones (2001)*, collectively called the GB technique and described in section 5, was used for analysing Karoo data. Figure 11.2 shows the GB single site analysis results and the strike sensitivity with depth variations. There appears to be very little strike variation with depth down to 2 km. Site KIM021 shows "anomalous" sensitivity with RMS variations of more than 2. This variation is attributed to 3D structure as is observed by splitting in the TE and TM phase responses from early periods $10^{-2}$ to over $10^{2}$ s. On the surface site KIM021 is not located over or close to a fault structure that can be traced on the surface, as...
11.2 Strike Analysis

Figure 11.2: Sensitivity to strike direction shown for different Niblett-Bostick depths shows that only a few sites (e.g., KIM021 and KXT010) are moderately sensitive to strike direction, where low RMS values (<2) are defined over a limited range of azimuths. Other sites are relatively insensitive to strike direction and are characterized by low RMS values.

such the 3D nature of the responses are difficult to discern given the layered nature of the stratigraphy in this part of the Karro basin. However it is well known that the dolerite dykes and sills are numerous and their structure, geometry and dimension can vary greatly. While dykes could possibly result in the increased dimensionality of the response, it is highly unlikely given that the response on either side of MT site KIM021 exhibit a relatively 1D/2D responses (i.e., not split in the phase responses).

The variation in site-to-site azimuth is perhaps better visualised in Figure 11.3, where the general strike orientation is between 40° and 46°, with an average of 45°. This value is consistent with the strike azimuth derived by Muller et al. (2009) for the KIM-NAM profile. As a result MT responses were decomposed to 45° strike angle for 2D modelling and the resulting responses are shown in Figure 11.4. Figures 11.5 and 11.6 show, respectively, gridded pseudo-section plots of the undecomposed and
Figure 11.3: Results of MTstrike analysis along profile using GB decomposition. Data period are labelled, with increasing periods a proxy for increasing depths. Double-headed vectors show 2-D geoelectrical strike direction, scaled in length by phase difference between the TM and TE mode.

decomposed (to 45°) data. In general there is not significant difference between the two images. It should be noted however that these images provide a “first-pass” look at the data and are largely qualitative. Formal 2D inversion model approaches will be presented in the next chapter from which quantitative information, like conductance and thickness of geological strata, will be derived.

However, pseudosection plots, particularly the phase responses, suggests a layered nature of the subsurface to the south. Furthermore, the NW part of the profile appears to be underlain by a resistive basement, possibly of the Kaapvaal craton. Indeed the SE margin of the Kaapvaal craton is mapped between sites KXT003 and KXT004, precisely where the resistivity contrast is observed on the pseudo-sections. Thus the phase data is not affected by static shift and it provides more reliable qualitative
Figure 11.4: Example of MT responses from a few stations along the profile after 45° strike decomposition. (see Figure 10.6 for their location)

information. It is also readily clear from Figures 11.5 and 11.6 that the higher the resistivity the lower the phase magnitude. This is behavior is expected, particularly for a 1D Earth model.
Figure 11.5: Pseudosection plot of the undecomposed responses along the profile. Please note that sites KIM0113-KIM017 were added here to derive this particular map but will not be used for further formal 2D inversion modelling as they fall outside the target area.
Figure 11.6: Pseudosection plot of the decomposed responses (azimuth 45) along the profile.
12

12.1 1D FORWARD AND INVERSE MODELLING

As was already alluded to in the previous sections, MT responses are the manifestation of 3D subsurface geology, even when we collect 2D profile data. In some cases however the resistivity responses can be successfully modelled using layered Earth models, particularly when apriori information is available on the distribution and thicknesses of sedimentary layers. The Karoo basin has been extensively drilled in some places in search for natural gas and coal deposits. As a result a good deal is known about the layered structure and thickness of strata in the basin.

In this section we model the MT responses using 1D techniques for several purposes:

- to gain an understanding on the ability of the MT data to resolve layered Karoo strata,
- to build a geologically reasonable 1D section of the profile in order to image the continuity of different strata
- use the 1D model as input/prior model to constrain the 2D inversion model.

Several 1D forward and inversion procedures exist and were referred to in chapter 2. In this section, the OCCAM (Constable, 1987) and BOSTICK (Bostick, 1977) models will be derived. The OCCAM and BOSTICK models are smooth models, which provide resistivity distribution versus depth. By fitting data using smooth models the risk of over-interpreting the data or modelling arbitrary discontinuities is simple layered models is reduced. Smith (1988) proposed the use of the Spearman’s statistic to test for over-fitting or under-fitting of MT data.

One-dimensional modelling, like any other modelling, provides meaningful results if aprori information, in a form of borehole, geological and/or geophysical data, is
known. There have been numerous boreholes drilled the Karoo basin and some DC and Schlumberger arrays electrical resistivity data acquired. van Zijl (2006) provide a review of these resistivity studies and developed conductivity 1D section for the region closest to the MT profile in the southern part of the Karoo. The resistivity section, Figure 12.1, shows a general increase in resistivity from the surface to a depth of up to 3 km, ranging from 300 Ohm-m to 2000 Ohm-m. Also shown are the apparent resistivity responses of some stations on the region.

Figures 12.2, 12.3, 12.4 and 12.5 show the 1D OCCAM and BOSTICK models of several responses along the profile. In each case the invariant resistivity and phase responses has been calculated. The invariant curve, thus lies, graphically between the TE and TM mode curves. For each of the stations on the MT profile the OCCAM and BOSTICK models were computed. The Occam inversion was run with 15 iterations using the maximum number of layers (43) to derive the smoothest model. A sharp boundary model was also computed for 15 iterations to produce a layered model from the smooth OCCAM model. Layer thicknesses were manually edited to match those determined from boreholes logs shows in Figures 10.3 and 10.4. In order to produce a layered model from the smooth Occam model, a ‘sharp boundary’ layered model
inversion (Figure 12.5) was also run over 10 iterations with predefined maximum RMS fit of 5%. Layer thicknesses and resistivities were manually edited and fixed to increase the match with the Occam models. The number of available layers in the sharp boundary computations is limited to 8 layers, a key limitation of Mackie and Madden (1993c)’s Winglink package. The southern-most stations KXT0009 and KXT010 show similar layered models where Stormberg, Beaufort, Ecca and Dwyka group rocks are resolved. There is gradual decrease in resistivity with depth, as was also noted by van Zijl (2006). The resistivity information from this section and the thickness of the layers determined from boreholes is used here to constrain the 1D and 2D models.

Figure 12.6 shows a 2D resistivity model that has resistivity values based on the 1D inversion models shown in Figure 12.5 and the layer thickness constrained by the boreholes. A forward response was computed for hypothetical MT stations and an example of a typical response is also shown in Figure 12.6 B. This response is similar
to the deep sounding data modelled by *van Zijl* (2006) shown in Figure 12.1 C, illustrating the gradual increase in resistivity of the Karoo stratigraphy. MT sites KXT005 generally shows similar responses (Figure 11.1). In the same light a 1D forward response was computed varying the thickness (between 500-1500 m) of the Beaufort group rocks in particular as their thickness is thought to change rapidly in places (*van Zijl*, 2006; *Viljoen et al.*, 2010; *Cloete*, 2010). The response of maximum thickness variation illustrated in Figure 12.7 shows similar behaviour to the 2D responses. Figure 12.8 shows 1D resistivity models derived using the 1D OCCAM inversion algorithm. A good fit between borehole information and resistivity can be observed in the profiles.
12.2 2D INVERSE MODELLING

Two-dimensional (2D) inversion modelling was carried out using the NLCG scheme of Rodi and Mackie (2001), as implemented in the WINGLINK software package. This same package was used to create models in section 6. As already described in that section, there are several inversion user-controlled parameters that can be changed before and during the inversion process. Thus the programme is sensitive to the starting model, the parameters chosen, the data, the error floors, and the modes (TE or TM) used. An appropriate combination of these parameters and settings will give a reasonable geological model being sought. A systematic inversion process was followed whose purpose was to firstly "slowly" introduce structure in the model. Unless otherwise stated the resulting 2D inversion models were generated following the 5-step procedure outlined below:
Figure 12.5: 1D models with various stratigraphic layers within the Karoo basin indicated. For the Occam and sharp boundary 1D models, the depths are real modelled depths, whereas for the Bostick model the Niblett-Bostick depth approximation is used. Three 1D models are shown in each plot: the Occam model (thick black lines), the Bostick model (thin continuous lines) and the sharp boundary layer models (shaded zones). The stratigraphic positions and depth of each layer was inferred from the SA1/66 borehole on Figure 10.3.

- **Step 1**: TM phase only inversion - only TM data were used and the apparent resistivity error floor set to a high value (50%). The TM phase error floor was set high (20%), then successively reduced to 10% then 5%.

- **Step 2**: TE phase was added, with the TE apparent resistivity error floor set to a high value (50%), and the TE phase set high then successively reduced to a value about the same as that of TM phase.

- **Step 3**: TM apparent resistivity was added and then TM apparent resistivity error floors successively reduced (50% - 25% - 10%) to a value about twice that of the TM phase.
Figure 12.6: A: Layered Karoo model based on known geological section and borehole (Figure 10.3). The resistivity were estimated from 1D inversion results (Figures 12.2, 12.3, 12.4, and 12.5). B: A forward MT response of one of the stations in A. C: Response from the deep electrical sounding work of van Zijl (2006).

Figure 12.7: 1D forward model derived by assuming a layered Karoo stratigraphy.

- **Step 4**: TE apparent resistivity added - then corresponding erros floors successively reduces (50% - 25% - 10%) to a value about twice that of TM phase.
Figure 12.8: Resistivity models derived using the 1D Occam inversion algorithm. Vertical bars beneath the stations indicate the sharp boundary layered 1D models. The background shows the 1D Occam inversion result, where the 1D models at each site. A good fit between borehole information and resistivity can be observed in the profiles.

- **Step 5:** The statics were set on and the TE and TM apparent resistivity error floor set to same level as TM phase error floor.

There were no Tipper ($H_z$) data for the MT stations and as such tipper data could not be included in the inversion process.

One of the inversion parameters of the WINGLINK programs that significantly affect the inversion results is the $\tau$ parameter. This regularization parameter controls the trade-off between data misfit and model smoothness. In contrast to the Occam inversion, where the parameter varies with each iteration, the NLCG methods allows the user to determine a value for the smoothness parameter, and therefore to chose a desired misfit. The large value of $\tau$ results in a smoother model at the expense of worse data fit. It can be seen on Figure 12.9 that for $\tau = 3$ the model is significantly smoother than that where $\tau = 0.1$. In general, for lithospheric studies, one might desired a larger value of $\tau$ given the expected smooth nature of the subsurface at those depths. For crustal studies, where geological structure is more complex, a lower value is chosen in order to get a more "rounder" model.

There is however a formal approach, using the so-called *L-curve*, that is used to decide on the appropriate $\tau$ to use. This was discussed briefly in chapter 6 and involves conducting multiple inversions runs varying the value of $\tau$ and inspecting
the resulting RMS misfits. The $\tau$ value, representing point of maximum curvature is chosen as the most appropriate for the subsequent inversion runs. Furthermore, given that apparent resistivity responses are often susceptible to static shift effects, it is wise to conduct the $\tau$ test using only TE and TM mode phase data, i.e., by setting the apparent resistivity error floors to a high value (50\%) and the phase error floors to low value (3\%). Figure 12.10 shows the results of RMS vs the $\tau$ parameter for several inversion runs. The most appropriate $\tau$ value to use was chosen to be 3.

A simple test of the 2D nature of the sub-surface is to conduct an independent modelling of the TE and TM modes. In addition to the anisotropic nature, independent TE and TM modelling provides a first pass look at the association of the resulting resistivity features with individual modes. Figure 12.11 shows TE and TM mode resistivity models derived by setting the respective apparent resistivity error floors to 10\% and that of phase responses to 5\%. It is readily clear that there are no significant differences in both resistivity models, particularly the SE part of the profile. The resistive feature in the NW part of the profile is mapping basement rocks of the Kaapvaal craton on which the Karoo sediments were deposited.

The SE part of the profile in Figure 12.11 shows a layered resistivity structure
Figure 12.10: Tradeoff curve determining an empirical solution for the best fitting smoothing parameter $\tau$. The curve plots smoothness parameter on a logarithmic scale against the data misfit (RMS error). The best trade-off between both parameters can be found at the point with the highest curvature, in this case where $\tau = 3$.

in both the TE and TM mode inversion models. The top-most thin conductive (15 Ohm.m) layer underneath sites KXT005 to KXT002 shows the regolith cover sequence. A 500 m thick layer, with resistivity of 100 Ohm.m, possibly mapping the Stormberg Group rocks, is underlain by a 1 km thick 400 Ohm.m layer of the Beaufort group. The Ecca group sediments, which appear to be thickest below the KXT007, have a moderately low resistivity, with the base at 6.5 km. The Dwyka tillites, show increasing resistivity towards the basement. It is noted that while these resistivities are interpreted to correlate with individual Karoo group strata (see Figure 12.12), it is impossible to resolve alternating thin (possibly conducting) shale formations that are known to exist in the Karoo, given the inductive nature of the MT method and its limited ability to resolve very thin conducting layers. The shale formations, given their low porosity, are ideal cap rocks for preventing injected carbon from migrating upwards. Also shown in Figure 12.12 are the 1D OCCAM models of the southern sites, KXT009 and KXT010, that show the layered nature of the sequences. Several other parametrizations were modelled as shown in Figure 12.13. Comparing the model responses and the measured data gives an indication of how well the data were fitted during the inversion process and also how robust the features in the model are. This comparison is shown in Figure 12.14, for a some stations along the profile. Most sites return an RMS of <1.5 indicating a very good fit and confidence in the resulting model.
Figure 12.11: Independent 2D TE and TM mode inversion models showing moderate variation in resistivity structure, implying little anisotropic structure.
Figure 12.12: A 2D inversion model highlighting various stratigraphic horizons in the southern part of the MT profile. The NW portion is underlain by resistive basement rocks, possibly of the Kaapvaal craton. The 1D OCCAM models show the layered nature of the subsurface in the southern part of the profile.
Figure 12.13: 2D models derived from joint TE and TM apparent resistivity and phase responses. The smoothing parameter $\tau$ was set at 0.5, with error floors for TE and TM apparent resistivity set at 15% and phase at 3%.
Figure 12.14: 2D Model responses compared to the measured data form joint inversion of the TE and TM mode apparent resistivity and phase data. Data up to 100 s was inverted. The RMS plots for each stations are shown.
13.1 POROSITY ESTIMATES

Some historic boreholes in the Karoo basin have porosity estimates (see reports of Viljoen et al. (2010), Cloete (2010) and Woodford and Chevallier (2002)). The data from some of these borehole is shown in Figure 13.1 for Ecca group sediments and Table 10.4 for the Beaufort group sediments. The closest boreholes (MA1-69 and WE1-69) to

Figure 13.1: The borehole logs of the Beaufort sandstone (A and B) and Ecca sandstone (C) showing variation of porosity with depth. Data is from (Woodford and Chevallier, 2002)

the MT profile (approximately 50 km) are shown in Figure 10.5. In general porosity
estimates for the Beaufort group are in the range 4-10%. Most of the Karoo Basin

![Graph showing the variation in total dissolved solids (TDS) and source depth (in metres) of the groundwater. The graph indicates an apparent increase in TDS with depth.](image)

Figure 13.2: The variation in total dissolved solids (TDS), a measure of salinity, and source depth (in metres) of the groundwater. Note the increase in TDS with depth (data from Woodford and Chevallier (2002)).

has TDS in the range of 450-1000 mg/l (Woodford and Chevallier, 2002) (Figure 13.2). Given the knowledge of the TDS it is possible to determine the water resistivity using equation 9.5 or 9.6. This was done for Karoo TDS borehole data and the results are shown in Figure 13.4. As salinity rises the resistivity decreases (Figures 13.3 and 13.4). In general resistivity of fluids $\rho_w$ decreases with depth.

Determining porosity using equation 9.4 is the next step in the analysis. As shown before, a key parameter in Archie’s law is the cementation factor $m$. Typical values of $m$ include $m=1.8$-2.0 for consolidated sandstones and $m=1.3$ for unconsolidated sands. It can be shown that the case with $m=1$ corresponds to fluid distributed in cracks, while $m=2$ corresponds to fluid distributed in spherical, poorly connected,
pores \cite{Unsworth2007}. A value of $m=1.5$ represents an intermediate case and is used in this study as the preferred value. In order to test the dependance of porosity estimates on the cementation factor $m$, two $m$ values were used: $m =1.5$ and 2. In a similar light the formation resistivity water equations of \textcite{Meju2000} and \textcite{Block2001} were used to test the variation in porosity. Figures 13.5 and 13.6 shows porosity values estimated from applying Archie’s law, in combination with the formation water values on Figure 13.4. The rock resistivity values $\rho_r$ at each depth were read from the 1D (Figure 12.2) and 2D models (Figure 12.11) models and the formation water resistivity are as shown in Figure 13.4. Equation 9.4 was used for the computations of porosities shown in Figure 13.5. The porosities estimated using the empirical formula of \textcite{Block2001}, for $m = 1.5$ or $m = 2$ consistently return lower porosity estimates than that of \textcite{Meju2000}. Unconsol-
Figure 13.4: Water resistivity estimates ($\rho_w$) vs depth, as calculated from the TDS values from Woodford and Chevallier (2002)
idated sediments, i.e., lower \( m \) value, would of course have elevated porosities. This relation is clear for the MT estimated porosity in the Karoo. Furthermore, porosity as might be intuitive, should generally decrease with depth due to load pressure effects (i.e., \( \phi \propto \frac{1}{\text{depth}} \)). This is evident in Figure 13.5, where porosity varies from 5.6% at about 80 m depth to less than 1% at 800 m. The discrepancy between results from Meju (2000) and that of Block (2001) stems from the fact that the former’s empirical relation depend largely on the ionic content of the solution. Furthermore, Meju (2000)’s formuae is for leachate from a landfill site therefore the geochemistry of these fluids are not expected to have a strict relation to that of deep pore waters in the Karoo basin. Therefore the result of Block (2001) is more reasonable and appropriate to use for this study.

It should be noted that the porosities estimated from MT derived resistivity data represent apparent porosities, that is, they represents the interconnected pore volume or void space in a rock that contributes to fluid flow or permeability. Therefore isolated pores and volume occupied by water adsorbed on clay minerals or other grains is excluded in the estimates. It is worthwhile to compare the porosities estimated from borehole to those derived from MT data. In general, the MT derived porosities are less than those derived from boreholes, in some cases there is an order of magnitude difference (Figure 13.7). There could be number of explanations for the discrepancy. Firstly, the MT models represent to a large extent average or bulk resistivities of rock, not true resistivities. For instance, the MT models are unable to recover the complete structure of the Karoo in that some very thin layers (e.g., shales) were not resolved. Secondly, the inductive nature of the MT method is such that resolution is lost when depth increases, therefore depth estimates from MT data can be underestimated particularly where there are conductive units that might be fluid-filled. This observation can however be overcome with collecting very high-resolution data (closely-spaced stations and recording for longer to obtain better depth estimates.) Thirdly, borehole derived resistivities were estimated from rock specimen in the laboratory. There is a general trend in comparing laboratory data to recorded MT data, where the former consistently return relatively higher estimates of resistivity, while the latter is less. This observation holds true for water in the mantle (Yoshino et al., 2006; Gaillard et al., 2008), but is not necessarily true for the crust.

13.2 THE CO\textsubscript{2} STORAGE POTENTIAL OF THE KAROO BASIN

Given the low porosity estimates derived from resistivity data it is reasonable to assume that, what is being estimated is not primary porosity but secondary porosity, that is small cracks and fissures. Secondary porosity in the Karoo can be promoted by
13.2 The CO2 Storage Potential of the Karoo Basin

Figure 13.5: Apparent porosity estimated from MT resistivity 1D and 2D models. Computations were done using Archie’s Law and the empirical relations of TDS and formation resistivity, calculated using Meju (2000)’s and Block (2001)’s equations.

dolerite dykes and sills due to their heating effect and dissolution effects. Fractures could negatively affect the containment of CO2. Borehole studies in the Karoo have no estimates of secondary porosity.

Some exploration companies are conducting studies to look for gas deposits found in fractured shales (not sandstones) in the Karoo. Roswell and de Swart (1976) reports that shale gas occurs in isolated pockets and has low volume. The presence of dolerites dykes and sills and their influence on secondary porosity requires further investigation.
13.2 THE CO₂ STORAGE POTENTIAL OF THE KAROO BASIN

Figure 13.6: Apparent porosity estimated from MT resistivity 1D and 2D models compared to MA1-69 borehole porosity estimates. Computations were done using Achie’s Law and the empirical relations of TDS and formation resistivity.
Figure 13.7: Apparent porosity estimated from MT resistivity 1D and 2D models compared to MA1-69 borehole porosity estimates. Lithological boundaries from Boreholes MA1-69, WE1-66 and WE1-69.
Part IV

CONCLUSIONS, BIBLIOGRAPHY AND APPENDICES
CONCLUSIONS AND OUTLOOK

14.1 2D AND 3D MODELLING AND INTERPRETATION OF DAMARA AND CONGO CRATON

14.1.1 Crustal models

The upper crust of the DGC is characterized by both resistive and conductive structures, the former interpreted to be granitic intrusions in the Central Zone related to the Pan-African magmatic events. A mid-lower crustal conductor is confirmed the origin of which is observed in the DGC which we propose to be related preferential alignment of graphitic/serpentinitic/sulphide materials during the collision of Kalahari and Congo cratons, as part of Gondwana almagamation. The southward steeply dipping structure (corresponding to the Autseib Lineament) we propose is a feature that is distinct to Archean/Proterozoic margins and that it should be used to target and distinguish terranes of different ages.

14.1.2 Congo craton southern extent

In general, the MT models suggest that the Congo craton lithosphere extends significantly southward beneath the Northern Platform sequences of Damara belt. The southern and northern DGC passive margin sequences, which were deposited on a rift margin, appear to be overthrusted onto both cratonic nuclei during continental amalgamation at 495 Ma, supporting studies by Goscombe (2004), to give a doubly vergent orogen. This result means that the current tectonic lines mapping the southern edge of the Congo craton need to be revised and drawn further south.

It is interesting to note the striking similarity between our MT models and other Archean-Proterozoic margins around the world in imaging steeply conductive features separating the two terrains. For example, the Central Australian Suture Zone, which forms the contact between the North Australian craton and the Proterozoic
Warumpi province (Selway et al., 2006, 2009), is mapped as a dipping conductor. Also in Australia, the Errabiddy Shear Zone forms a conductive suture between the Yilgarn Craton and the Glenburgh Terrane (Selway et al., 2009). The margin between the Slave craton and the Wopmay orogen (Spratt et al., 2009) is preserved as steeply-dipping conductive structure. All of these features were subjected to later re-activation post continental collision. This observation suggests that steeply-dipping crustal scale conductive structures can be used to infer continental accretion and perhaps broad generalization can be inferred about similar mode of evolution for Archean/Proterozoic margins.

14.1.3 \textit{Rifting in the Okavango Delta?}

There has been some suggestions that the East African Rift System extends into northwestern Botswana, based on persistent seismicity (Reeves, 1972), forward modeling of magnetic and gravity data (Kinabo, 2007) and anomalous heat flow (Chapman and Pollack, 1974). All geophysical experiments done on the Okavango Rift Zone (ORZ), before ours, focussed on shallow (less than 1 km) structure (Modisi et al., 2000; Kinabo, 2007), and revealed the ORZ as a half-graben feature. If indeed continental rifting is taking place, then one might expect a classic rift signature of thinned lithosphere accompanied by a localised deep seated high conductivity mantle anomalies, as is observed for example in the Rio Grande Rift (Hermance and Pedersen, 1980; Jiracek et al., 1983; Hermance and Neumann, 1991). None of these features are observed on the electrical resistivity models, particularly along the OKA-WIN profile. The lack of elevated mantle conductivity suggests that incipient continental rifting, if taking place, is generated from pre-existing structural features in the crust, as proposed by Birt et al. (1997) for the Kenya Rift, and propagates downward without any mantle upwelling signature.

14.2 \textbf{Carbon Capture and Storage Summary}

The saline formation and coalbed storage possibilities for the Karoo Basins were found to be poor because of the very low porosity and permeability of the sandstones, and the presence of extensive dolerite sills and dykes.

The obvious limitation of the above study is the large spacings between the MT stations (> 10km). This is particularly more limiting in resolving the horizontal extent of individual units as there might be local structural variations that cannot be resolved using the current dataset in the Karoo basin. However the 1D models provide layered Earth models that are consistent with the known geology. The resistivity values from
the 1D models allowed porosity of the Ecca and Beaufort group lithologies to be determined. It is inferred that the porosities values are in the range 5-15% in the region below the profile. This value is considered too low for CO₂ storage as the average porosity of rock used for CO₂ is generally more than 10 to 12 percent of the total rock unit volume (Viljoen et al., 2010).

14.2.1 Carbon Capture and Storage Recommendations

• MT data can be used effectively for qualitative and quantitative modelling of storage site, including porosity estimates. For all onshore storage sites, it is recommended that high resolution MT data be collected on identified sites.

• Forward MT modelling has shown that a site spacing of 2 km could resolve the thickness and conductance (conductivity-thickness product) of layers in the Karoo basin.

• The Karoo has CCS storage potential but issues regarding low porosity and permeability in some parts need to be addressed. Furthermore, if technological developments in low porosity injectivity can be overcome, then the Karoo basin offers one of the largest potential storage of CO₂ in the world given its sheer size.

• Secondary porosity caused by dolerite dykes/sills needs to be investigated further. This was not done in this study given the limited time and data available.


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Part V

APPENDIX
This appendix presents all the papers published in international literature, with myself as being the first author. There are two other papers published with my colleagues where my contributions were data collection in Share, et al., 2014 and data modelling in Hobbs, et al.,(2013). These papers are consequently not included in this Appendix.

The papers below are all in published format, therefore the layout, format and figure numbering does not follow the rest of the thesis.
I wrote the material in the ALL the publications below. As a first author, I collected, processed, modelled and interpreted the data used in the publications. Additionally, I made all the figures shown. My co-authors contributed to the refinement of the text.
Lithospheric structure of an Archean craton and adjacent mobile belt
revealed from 2-D and 3-D inversion of magnetotelluric data:
Example from southern Congo craton in northern Namibia

T. D. Khoza,1,2 A. G. Jones,1 M. R. Muller,1 R. L. Evans,3 M. P. Miensopust,1,4
and S. J. Webb2

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[1] Archean cratons, and the stitching Proterozoic orogenic belts on their flanks, form an
integral part of the Southern Africa tectonic landscape. Of these, virtually nothing is
known of the position and thickness of the southern boundary of the composite Congo
craton and the Neoproterozoic Pan-African orogenic belt due to thick sedimentary cover.
We present the first lithospheric-scale geophysical study of that cryptic boundary and
define its geometry at depth. Our results are derived from two-dimensional (2-D) and
three-dimensional (3-D) inversion of magnetotelluric data acquired along four
semiparallel profiles crossing the Kalahari craton across the Damara-Ghanzi-Chobe belts
(DGC) and extending into the Congo craton. Two-dimensional and three-dimensional
electrical resistivity models show significant lateral variation in the crust and upper mantle
across strike from the younger DGC orogen to the older adjacent cratons. We find Damara
belt lithosphere to be more conductive and significantly thinner than that of the adjacent
Congo craton. The Congo craton is characterized by very thick (to depths of 250 km) and
resistive (i.e., cold) lithosphere. Resistive upper crustal features are interpreted as caused
by igneous intrusions emplaced during Pan-African magmatism. Graphite-bearing calcite
marbles and sulfides are widespread in the Damara belt and account for the high crustal
conductivity in the Central Zone. The resistivity models provide new constraints on the
southern extent of the greater Congo craton and suggest that the current boundary drawn
on geological maps needs revision and that the craton should be extended further south.

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of an Archean craton and adjacent mobile belt revealed from 2-D and 3-D inversion of magnetotelluric data: Example from

1. Introduction

[2] The African tectonic landscape is comprised of a
number of relatively stable Archean cratons that are flanked
and stitched together by younger Proterozoic fold belts.
The cratonic margins, and intracratonic domain boundaries,
have played a major role in the Phanerozoic tectonics of
southern Africa by focusing ascending magmas and localiz-
ing cycles of extension and rifting [Begg et al., 2009].

The characteristic feature of cratons is their thick litho-
sphere, but their formation and evolution remains elusive
and is the central focus of geoscientific studies. The identi-
fication of major tectonic events is complicated by lack of
knowledge of lithospheric structures and their geometries
beneath the surface expression of these orogens and cratons.
Deep-probing magnetotelluric (MT) information is provid-
ing critical data in aiding the understanding of the formation
and deformation of Pre-Cambrian orogens.

[3] In southern Africa, the greater composite Kalahari
craton is separated from the composite Congo craton by
the NE-SW trending Damara-Ghanzi-Chobe belts (Damara
belt is in Namibia, and its extension in Botswana is
called Ghanzi-Chobe), henceforth collectively called DGC
(figure 1), a component of Gondwanan Pan-African oro-
ogenic belts. Due to the huge amount of geochemical [Griffin
et al., 2003], seismic [James et al., 2001], and electrical
[Jones et al., 2009; Muller et al., 2009; Evans et al., 2011;
Miensopust et al., 2011] data, a lot is known about the
Kalahari craton, which includes in its core the Kaapvaal
craton. What is not fully understood is the lithospheric struc-
ture and geometry of the Congo craton, north of the Kalahari
craton, particularly its southern boundary with the Damara belt in north central Namibia and northwestern Botswana.

[4] The composite Congo craton comprises several Archean shields, the southern part comprising the Angolan shield, that consists of granite-gneisses and metasediments overlain by Paleoproterozoic cover [Begg et al., 2009]. The highly complex Damara-Ghanzi-Chobe (DGC) belt, forming part of the Pan-African orogenic system, records the Neoproterozoic collision between the composite Congo and the Kalahari cratons during the amalgamation of Gondwana [Daly, 1986; Prave, 1996; Hanson, 2003; Ritter et al., 2003; Gray et al., 2006]. Thick Kalahari sedimentary cover makes it difficult to investigate the geometry of the southern Congo craton beneath the DGC belt with geological mapping. In this work, we use the deep-probing magnetotelluric technique to circumvent this problem.

[5] The Southern African Magnetotelluric Experiment (SAMTEX) is a multinational scientific project that is applying deep-probing MT to understand the present and past tectonic process in southern Africa through mapping of physical parameters and their geometries. Data at over 750 MT sites were collected during four phases of acquisition. In the third and fourth phases, MT data were acquired along four N-S and two E-W profiles crossing the Damara belt and the southern Congo craton (Figure 1), in addition to the other profiles that will be presented elsewhere.

[6] This paper has two parts; the first part concerns the 2-D isotropic and anisotropic inversion modeling and 3-D inversion, the second part is the interpretation of the models. The 2-D crustal and lithospheric models presented are of the westernmost N-S ETO-KIM profile, which is an extension of the KIM-NAM profile published by Muller et al. [2009] (see Figure 2). We chose the ETO-KIM profile for 2-D inversion as it will provide strict comparison with models from the KIM-NAM profile and provide insight into the variation in lithospheric structure/thickness from a Proterozoic terrane to an Archean craton. The 3-D inversion results are derived from applying a newly developed 3-D modular electromagnetic modeling (ModEM) code of Egbert and Kelbert [2012] on MT data from 88 stations distributed across the DGC and the southern Congo craton (see Figure 3 for site locations). The overarching objective of our study is to map the deep electrical structure of the Congo craton and its flanking Damara orogen and to understand the nature of the cryptic boundary between these two geological domains.

[7] The magnetotelluric (MT) method has been used extensively in similar Archean and Proterozoic terrains [Jones et al., 2002; Wu et al., 2005; Jones et al., 2005a; Selway et al., 2006; Spratt et al., 2009; Muller et al., 2009; Miensopust et al., 2011] to image the geometry of cratonic blocks and infer the tectonic histories of Pre-Cambrian collisional margins. Muller et al. [2009], Evans et al. [2011], and Fullea et al. [2011] showed how electrical resistivity data can be used to infer the depth to the thermal thickness of the lithosphere by comparing the results with
Figure 2. Detailed geology of northwestern Namibia modified from Corner [2008]. Deep-seated NE-SW trending faults (Okahanja and Autseb lineaments) separating various Damara belt units are mapped. The thick black line shows the profile over which the geological section in Figure 4 was generated. Also shown is the Kamanjab inlier (KI) which forms a part of the Archean basement of the Congo craton. The stratigraphic subdivisions are based on metamorphic, structural, and surface geological mapping. CC = Congo craton outline inferred by Corner [2008], AL = Autseb Lineament, OML = Omaruru Lineament, OKL = Okahanja Lineament.

2. Geological Framework

[8] The Congo craton, which occupies a large part of central and southern Africa, comprises several Archean shields, the southwestern part of it comprising the Angolan shield that consists of granite-gneisses and metasediments overlain by Paleoproterozoic cover [Begg et al., 2009]. Thick Kalahari sedimentary cover leaves the southern Congo craton, and particularly its boundary with the DGC, virtually unknown and poorly understood.

[9] The Damara orogen is a three branch system reflecting part of West Gondwana suture and a collisional triple junction between the Congo and Kalahari cratons in southern Africa and Rio de Plata craton in South America [Gray et al., 2008; Frimmel, 1998]. Two coastal branches, the Kaoko belt and Gariep belt, lie parallel to the present Namibia coast while the Damara-Ghanzi-Chobe belt (DGC) forms the NE-SW trending “Inland branch” or “intracratonic branch” [Martin and Porada, 1977]. The DGC, which went through a complete Wilson cycle, records the high-angle convergence of the composite Kalahari craton to the south with the composite Congo craton to the north during Pan-African amalgamation of Gondwana [Gray et al., 2006]. The Brazilian orogens, Riberia, and Dom Feliciano complete the orogenic systems that define the amalgamation sutures of West Gondwana between the South American and African cratonic nuclei. The aggregation of West Gondwana involved the closing of the Adamastor Ocean between the Rio de la Plata and Kalahari cratons (forming the Damara Belt) and the Khomas Ocean between the Congo and Kalahari cratons (forming the Damara Belt) [Goscombe et al., 2003; Gray et al., 2006]. The lack of definitive evidence of subduction-related material (i.e., ophiolite and blueschist facies metamorphism) has often been cited as evidence against Khomas ocean closure as the Kalahari craton subducted northward under the leading edge of the Congo craton. Martin and Porada [1977], Porada [1989], and Trompette [1997] proposed that the Damara orogenic system developed in a continental rift setting around 1000 Ma, and the DGC represents a failed rift or aulacogen that was followed by basin closure at circa 520 Ma. However, mid-ocean ridge basalt-type geochemistry in the
intensely deformed pillow basalts, chert, and gabbros of the Matchless Amphibolite Belt in the DGC Southern Zone and the reported eclogites in the Zambezi Belt, which forms a continuation of the Damara belt in Zambia [Gray et al., 2008; Barnes and Sawyer, 1980], are increasingly providing support for the subduction hypothesis.

[10] The major geological units of the Damara belt (Figures 2 and 3) include passive margin carbonate sequences and basal rift-related and deep turbiditic sediments deposited in the period spanning the Neoproterozoic time from 770 to 600 Ma [Miller, 1983; Hoffmann, 1994; Prave, 1996; Frimmel, 1998; Goscombe, 2004]. The collision of the Kalahari and Congo cratons terminated the deposition of the marine sedimentary sequence in Neoproterozoic time, with peak metamorphism as a consequence of collision occurring in the Middle Cambrian to Lower Ordovician periods between 534 Ma and 508 Ma [Miller, 1983; Raab, 2002]. The sequence was intruded by granitoid material during Pan-African Damara orogeny in late Neoproterozoic-Cambrian times. Contrasting metamorphic (i.e., high temperatures/low temperatures in the center and margins of the orogen, respectively) and structural deformation patterns associated with the collision of the Kalahari and Congo cratons are observed across strike of the DGC crust. Pre-Damara basement is exposed as large Archaean-Proterozoic age inliers (Kamanjab and Kunene complex) in northern Namibia. The sedimentary sequences were deposited on top of these basement granitoids/gneisses, preserved partly as the leading edge of the Congo craton [Gray et al., 2008]. The envisaged crustal geometry is...
shown in Figure 4, and the tectonostratigraphic subdivisions with the DGC are based on the metamorphic, structural, and geological mapping. Moho depth is interpolated from off-shore seismics, and S wave receiver functions at a few locations (see location of SRF site by Hansen et al. [2009] in Figure 3).

Like the Congo craton, the Kalahari craton is a composite craton composed of several Archean and Proterozoic blocks. Herein, we follow the Jacobs et al. [2008] definition of the Kalahari craton, which includes the Archean Kaapvaal and Zimbabwe cratons, separated by the high-grade metamorphic Limpopo belt, in addition to Paleoproterozoic components (Rehoboth, Okwa, and Magondi terranes). The Kalahari craton acted as a stable cratonic block with respect to the later tectonics of the Namaqua-Natal mobile belt and the Pan-African orogeny that defined the final stages of the development of the DGC. Protracted postcollision tectonothermal events within the DGC included inter alia, episodic granite emplacements, metamorphism, and structural deformation [Gray et al., 2006]. The DGC belt is

Figure 4. Cross-sectional profile showing the simplified crustal architecture of the Inland Branch of the Damara Orogen (modified from Miller [1988] and Gray et al. [2006]. See Figure 2 for location of the profile (thick black line)). The Moho depth was estimated from off-shore seismic information [Fernández et al., 2010], and recent S wave receiver function results [Hansen et al., 2009] from the TSUM station shown in Figure 1.

Figure 5. Examples of magnetotelluric responses from six sites: three sites on the Congo craton and three sites from the Damara orogen (color-coded). Apparent resistivity and phase responses are plotted a function of increasing period (period being proxy for increasing depth). Both the transverse electric (XY) mode (currents flowing parallel to strike) and transverse magnetic (YX) mode (currents crossing strike) are shown. The data at most stations were generally good up to 8000 s.
overlain by thick Cenozoic Kalahari sediments, particularly in northwestern Botswana, and airborne magnetic and gravity patterns have been used as tools to map its spatial extent.

3. Previous Geophysical Studies

Previous electrical and electromagnetic studies of the DGC include the late 1970s and early 1980s magnetovariational and Schlumberger soundings by de Beer et al. [1975,
Figure 7. Niblet-Bostick depth estimates for MT stations along the ETO-KIM profile for both the TE (red bars) and TM (yellow bars) modes. While this is based on 1-D approximations, it gives in indication as to the depth sensitivity of the MT response for each site.

1976, 1982] and van Zijl and de Beer [1983]. These studies were the first to discover the along-strike crustal conductor within the DGC that extends from Namibia into Botswana. Serpentinitized lower crust was, at the time, thought to be the cause of the elevated conductivity.

[13] In 1998 and 1999, the GeoForschungsZentrum conducted a crustal-scale study along a 200 km-long profile with a site spacing of 4 to 12 km and a focused 3-D array of 60 sites with a site spacing of 500 m to 2 km across the Waterberg Fault-Omaruru Lineament [Ritter et al., 2003]. In addition to mapping the middle to lower crustal DGC conductor, Ritter et al. [2003] imaged a very resistive upper crust (granites) and two subvertical conductors, the Autseib and Omaruru Lineaments. The authors invoke graphite concentration to explain the conductivity along upper crustal lineaments and middle to lower crustal anomaly, the latter being along shear deep-seated shear planes. In a related study, Weckmann [2003] used MT phases to reveal a conductive ring structure in the shallow crust, with crust-scale electrical anisotropy and an elongated conductor running subparallel to the Waterberg Fault-Omaruru Lineament.

[14] Muller et al. [2009] presented a lithospheric-scale resistivity study of the Rehoboth terrane, partially extending into the DGC belt, which showed significant lithospheric thickness, electrical, and thermal variations from older to younger terranes. Muller et al. [2009] found the lithospheric thickness of the DGC to be approximately 160 km. Further east on the eastern border of Botswana, Miensopust et al. [2011] mapped the Ghanzi-Chobe belt having resistive 180 km thick lithospheric mantle.

4. Magnetotelluric Method

[15] Magnetotellurics (MT) is a passive-source electromagnetic sounding technique that uses the time-varying ionospheric magnetic fields as its source for periods greater than 0.1 s (i.e., deep crustal and mantle probing periods) [Chave and Jones, 2012]. Mathematically, MT responses are represented by a complex impedance tensor \( Z \) in the frequency domain (also known as the MT transfer function or MT response function), which defines the linear relation between the electric (\( E \)) and magnetic (\( H \)) field vectors. On surface, simultaneous measurements of the three magnetic field components (\( H_x, H_y, \) and \( H_z \)) and the two horizontal electric fields (\( E_x, \) and \( E_y \)) are recorded, processed, and modeled to obtain the subsurface resistivity structure of the Earth. The ratios of the electric and magnetic fields as a function of period (which, for an electromagnetic (EM) method, is a proxy for depth) gives the lateral and depth distribution of electrical resistivity [e.g., Jones, 1999; Chave and Jones, 2012]. For the ideal 2-D Earth case, two independent modes of electric field propagation can be defined: (1) the transverse electric (TE) mode describes currents flowing parallel to the strike direction, and (2) the transverse magnetic (TM) mode which describes currents flowing perpendicular to strike direction [Chave and Jones, 2012].

[16] The electrical resistivity of Earth materials varies by over several orders of magnitude, which makes the MT method a superior technique for investigating geological and structural changes in the crust and mantle of physical properties. In the crust, conducting mineral phases, like graphite and sulfides, produce conductivity anomalies. In active tectonic regions (e.g., Himalayas), partial melt is thought to be responsible for observed elevated conductivities [Unsworth et al., 2004; Le Pape et al., 2012]. In the mantle, electrical resistivity of mantle minerals (olivine, orthopyroxene, and clinopyroxene) is primarily controlled by temperature [Constable et al., 1992; Xu, 2000] and, to a lesser extent, compositional variations [Jones et al., 2009]. In addition, dissolved hydrogen has been found to increase the conductivity of olivine in the mantle [Karato, 1990] although the effect on conductivity is hotly debated [Karato, 1990; Yoshino et al., 2006; Jones et al., 2009; Karato, 2010; Jones et al., 2012]. Thus, resistivity models of the Earth provide directly the present geometrical arrangement of tectonic margins and the definition of the base of the lithosphere. Used in conjunction with geochemical and seismological information, it is possible to confidently derive the thermal
Figure 8. (a) Crustal-scale resistivity model of the ETO-KIM profile, showing the main features resolved by the inversion process. The mid-lower crustal conductors, upper crustal Pan-African granites, and the Archean basement (Kamanjab Inlier) are the major features of this model. (b) For comparison, the geological section, modified from Gray et al. [2007], is included. AL: Autoeib Lineament, OML: Omaruru Lineament, OKL: Okahanja Lineament, MA: Matchless Amphibolite, SFZ: Southern Foreland Zone.

4.1. Data Collection and Processing

The SAMTEX database comprises 750 MT stations distributed across three southern African countries: Botswana, Namibia, and South Africa (see Figure 1). During the third phase of SAMTEX, high-quality data were acquired along the N-S oriented subparallel profiles, herein referred to as DMB (Damara), NEN (Nyae-Nyae), and OKA-WIN (Okavango-to-Windhoek) (see Figures 2 and 3 for their location). The ETO-KIM profile (red circles in Figure 1) is the extension of the prior KIM-NAM profile (blue circles) published by Muller et al. [2009]; the objective of the latter was to determine the lithospheric structure and evolution of the Rehoboth Terrane and its relationship to the Kaapvaal craton. The NW-SE ETO-KIM profile and the northernmost E-W RAK (Ruacana-to-Katima) profile were both collected in the fourth phase. All profiles were designed to traverse over known and enigmatic tectonic terranes, with approximately 20 km separating the stations. The layout of the sites in Namibia and Botswana have a reasonable spatial distance that makes the region suitable for 3-D inversion. At each MT site, we recorded time variations of horizontal electric and magnetic fields for approximately 3 days, resulting in high-quality broadband data at periods between 0.002 s and 10,000 s. Vertical magnetic fields (Hz) were recorded at 17 sites along the ETO-KIM profile, 6 sites along the DMB profile, 5 sites along the NEN profile, and 7 sites along the OKA-WIN profile (difficult terrane precluded Hz measurements at every site). Phoenix Geophysics' instruments, MTU-5 and MTU-5A recording units and MTC-50 induction magnetometers, were used for broadband data acquisition. We also recorded long period data (10–30,000 s) at seven sites along the ETO-KIM profile, three sites on the DMB profile, four sites on the NEN profile, and four sites on the OKA-WIN profile over 3 week periods using LEMI-417M (Long Period Electromagnetic Instrument) systems manufactured by LVIV Centre of Institute of Space Research, Ukraine (http://www.isr.lviv.ua/lemi417.htm). The broadband and long period data were merged to produce a single MT response for a particular site. Where appropriate, the longer the period of recording, the deeper we image into the Earth. However, in more conductive regions, the signal penetration is attenuated due to the shielding effect of conducting materials (see section 8).

Data quality was generally good for most sites; however, where severe distortion (e.g., current channeling) effects were observed that could not be treated, those sites were not used for further modeling. Figure 5 shows an example of MT responses from four sites along the ETO-KIM profile (two sites within the Damara belt and two sites on the Congo craton). For each site, apparent resistivities and the phase lags between electric and magnetic fields evolution of Pre-Cambrian regions [Davis, 2003; Muller et al., 2009; Fullea et al., 2011].
5. 2-D Inversion of MT Data

[22] In general, the resistivity of Earth’s materials varies spatially in different directions, a feature termed electrical anisotropy. The cause of observed anisotropy, particularly for the mantle, is still debated, with hydrogen diffusion along olivine a-axes [Mackwell and Kohlstedt, 1990; Bahr and Duba, 2000; Gatzemeier and Tommasi, 2006; Jones et al., 2012] and interconnectivity of a conductive mineral phase (e.g., graphite) along grain boundaries [Jones et al., 1992; Mareschal et al., 1995], being the primary candidates for mantle materials. The presence of anisotropy in the lithosphere is increasingly being taken into account when undertaking MT interpretation [Baba et al., 2006; Marti et al., 2010; Evans et al., 2011; Miensopust et al., 2011; Le Pape et al., 2012] and can manifest itself in macroscale or microscale. Crustal anisotropy can result from different orientations of fluid-filled structures (i.e., strain-induced) or layering of materials with varying physical properties [Wannamaker, 2005; Miensopust and Jones, 2011]. In the lithospheric mantle, strain-induced crystal preferred orientation (CPO) of anisotropic minerals, hydrogen diffusivity in mantle minerals, or the presence of partial melt are the known main causes for observed anisotropy [Wannamaker, 2005; Yoshino et al., 2006]. Thus, understanding electrical anisotropy with depth has implications for the interpretational aspects of continental evolution. Ji et al. [1996], Hamilton et al. [2006], and Eaton et al. [2004] investigated the presence of electrical anisotropy for the Greenville Front, Kaapvaal craton, and Great Slave lake regions, respectively, in an attempt to look for a symbiotic relationship between MT and seismic anisotropy; their results varied from weak to strong correlation between observed seismic and electrical anisotropy. What is clear, however, is that tectonic processes, including those forming Neoproterozoic collisional terranes, result in observed anisotropy variations. Thus, it is important to understand anisotropy at all scales in order to infer present and past tectonic processes.

5.1. Isotropic Inversion

[23] We derived a 2-D isotropic electrical resistivity model for the ETO-KIM profile data using the finite difference code of Rodi and Mackie [2001]. This code uses a nonlinear conjugate gradient scheme to search for the smoothest best fit model [Mackie and Madden, 1993a, 1993b; Rodi and Mackie, 2001]. Several inversion parameters were tested in order to obtain a robust, low RMS and geologically acceptable model. The regularization parameter tau (τ) is the “smoothing” operator that controls the Tikhonov regularization trade-off between model smoothness and fitting the data. A tau value of 5 was used for the two models, after a rigorous search for the tau parameter that gives the most favorable compromise between RMS fit and model smoothness using the L-curve approach [Hansen and O’Leary, 1993]. A 100 Ωm homogeneous half-space was used as the starting model for all inversion runs. The systematic inversion approach that was followed included fitting initially the TE and TM phase responses preferentially by allocating high
error floors (50%) for the apparent resistivities. The next step was to include the apparent resistivities after some iterations by reducing the error floor, and lastly, the vertical magnetic field data were included. The reason behind inverting the phase responses first is that they are generally free of static shift effects [Jones, 1988]. The alpha and beta parameters in the inversion code, which control the horizontal smoothness and vertical regularization weighting, were set to 1 and 0, respectively, after testing. The final error floors were 5% for phase, 10% for apparent resistivity, and 0.02 for vertical fields. The reader is referred to Spratt et al. [2009] for an excellent application of the code and use of parameters. We chose to solve for the smoothest model available, since the inductive nature of the MT technique allows us to assume a smoothed structure at depth.

5.2. Anisotropic Inversion

[24] The anisotropic code we use is an extension of the isotropic code discussed above (same regularization), with an additional parameter ($\tau_{an}$) that controls the degree of anisotropy permitted by the inversion [Evans et al., 2005; Baba et al., 2006; Evans et al., 2011; Le Pape et al., 2012], which is how close the resistivities $\rho_{xx}$, $\rho_{yy}$, and $\rho_{zz}$ will be to each other. A low $\tau_{an}$ value (0) will enforce a totally anisotropic result, which means independent TE and TM mode inversions. The limitation of the code is that the anisotropic direction must coincide with the geoelectric strike direction. The resulting 2-D inversion models are shown for two different directions; one parallel ($R_{xx}$) and the other perpendicular ($R_{yy}$) to the geoelectric strike. The $R_{zz}$ model is not shown here as it is particularly difficult to resolve due to the fact that most of the current flow in the MT problem is horizontal.

6. 3-D Inversion

[25] The inversion of magnetotelluric data in three dimensions has developed rapidly over the last couple of decades, particularly in the last few years [e.g., Smith and Booker, 1991; Mackie and Madden, 1993a; Zhidanov et al., 2006; Sasaki and Meju, 2006; Farquharson and Craven, 2009; Siripunvaraporn and Egbert, 2009; Egbert and Kelbert, 2012]. The motivation for modeling the Earth in 3-D (instead of 2-D) is multifold. First, the Earth is 3-D, and it must be modeled as such for precise geological interpretation. While we collect MT data along 2-D profiles, the data we acquire and subsequently process is sensitive to 3-D conductivity variation and results in a full impedance tensor, with four complex tensor elements for each period, relating the electric and magnetic field components. Modeling all four components ($Z_{xx}$, $Z_{xy}$, $Z_{yy}$, and $Z_{yx}$) not only requires no assumptions to be made about strike directions, but
perhaps more importantly provides added information about the electrical resistivity structure.

[26] Second, recorded MT data often show that the validity of the 2-D assumptions fails due to the influence of 3-D subsurface structures. Depending on the inductive scale length of the target, this can lead to incorrect results if the data are modeled in 2-D. Third, the data coverage of SAMTEX stations (Figure 3) provides a reasonable spatial block that can be inverted in 3-D using a newly developed modular EM code.

[27] In this section, two different inversion schemes were applied to model our data in 3-D. We used the parallel implementation of the WSINV3DMT code of Siripunvaraporn et al. [2005], and the newly developed modular electromagnetic modeling and inversion ModEM scheme of Egbert and Kelbert [2012]. The WSINV3DMT is based on the 2-D data space Occam’s inversion of Siripunvaraporn and Egbert [2000] that seeks the smoothest model to fit the data at the specified level of misfit. The ModEM inversion scheme is, as the name implies, based on a modular design where the Jacobian and sensitivity matrices are factored into several components (i.e., data functionals, forward and adjoint solvers, and model parameter mappings). The computation of the sensitivity matrix this way allows for faster computation times. Not only is ModEM suitable for large inversions (such as in our study) but also consistently resulted in models with low RMS and with geologically reasonable geometry of features. As such, we will only show the ModEM results here but the models we obtained with WSINV3DMT contained the same features. As with the 2-D models, various parameters were tested to obtain geologically plausible models. The ModEM algorithm has provision to invert up to six combinations of impedance tensor elements ($Z_{xx}$, $Z_{xy}$, $Z_{yy}$, and $Z_{yx}$) and vertical transfer functions ($H_t$ tippers).

[28] A total of 88 stations were carefully chosen from a complete data set of 173, and inconsistent responses were edited and bad stations removed. The MT responses for each site were then decimated to 4 periods per decade with a total of 18 periods in the range 0.1–5000 s. For sites that had vertical transfer functions, a constant absolute error floor in $H_t$ was set at 0.01.

[29] The mesh was discretized to be fairly regular in the center, with cell widths of approximately 10 km. The mesh comprised a total of 64 cells in the $X$ direction, 78 cells in the $Y$ direction, and 52 cells in the $Z$ direction, with most stations preferentially adjusted to be located at the center of cells.

[30] Several starting models, including layered Earth models, different resistivity half-space models (100 $\Delta\Omega\text{m}$ and 200 $\Delta\Omega\text{m}$), and various smoothing parameters ($\tau$), were all tested and compared. A 100 $\Omega\text{m}$ prior half-space model consistently resulted in geologically consistent models with...
relatively low RMS. The final model, derived by inverting all four impedance tensor elements including the tippers, from a 100 Ωm half-space, with 10% and 15% error floors in diagonal and off-diagonal impedance elements, respectively, converged to an average RMS of 3.66 (Figures 8 and 9) after 69 h of computation using four nodes on the STOKES (SGI Altix ICE 8200EX) cluster of the Irish Centre of High Performance Computing (ICHEC, www.ichec.ie).

7. Results

7.1. Crustal Structure

[31] The 2-D inversion model (Figure 8) of the ETO-KIM profile reveals a complex crustal structure dominated by three main features that correlate relatively well with known structural and geological features: (1) the upper crustal resistive structures, (2) the middle to lower crustal conductive zone, and (3) the high resistivity Kamanjab Inlier to the north.

[32] The upper crustal resistive features are mapping the syntectonic and posttectonic granitic intrusions that form a significant component of the Damara belt outcrops, particularly in the Central Zone.

[33] Also evident in the MT model is the presence of the lower crustal conductive feature described previously [van Zijl and de Beer, 1983; Ritter et al., 2003]. The 3-D inversion results suggest that the DGC conductor is extensive along strike of the Damara orogen, as is evident on the depth slices (Figure 10). Spatial comparison with geological maps suggests that this conductor resides mainly in the Central Zone of the DGC, the top of which is at approximately 20 km. The depth to the base of the conductive feature is generally poorly resolved due to the inductive nature of the MT technique and the limited penetrative ability of the EM fields below conductors.

[34] The Kamanjab Inlier, which forms the basement upon which Damara sediments were deposited circa 770 Ma, is imaged as a very resistive feature. Compositionally, the Kamanjab inlier comprises granotoids and metarmorphic sequences [Goscombe et al., 2003], which accounts for its high resistivity (over 10,000 Ωm). The surface boundary of the inlier is traced on the geological map between sites ETO011 and ETO012; however, the MT model reveals that this boundary could be extended even further south between ETO009 and ETO010. Furthermore, the boundary of the Kamanjab Inlier at depth appears to have a gentle dip to the south, terminating beneath the Damara north Central Zone.

[35] The Autseib Lineament (AL) is one of the most prominent magnetic anomalies in the Southern Africa [Corner, 2008]. It separates the Damara Central Zone from the Northern Zone and extends northwestward into Botswana (see Figure 2). The AL, a former extensional fault reactivated as recently as the Mesozoic [Clemenson et al., 1997; Corner and Swart, 1997], is now a thrust fault and
appears on the resistivity model as an upper crustal conductor, steeply dipping to the south. Evidence for modern reactivation along the preexisting feature was presented by Raab [2002] using apatite fission results.

7.2. Mantle Lithosphere

Figure 11 shows the (a) isotropic and (b–c) anisotropic lithospheric scale 2-D models derived for the ETO-KIM profile. The smooth electrical resistivity models reveal a heterogeneous electrical resistivity for the mantle lithosphere. The Damara belt crustal conductor remains evident in the lithospheric model, as well as the resistive crustal granitoids. The most striking feature on the MT models is the resistive (i.e., colder), thicker lithosphere interpreted to be the electrical manifestation of the southern Congo craton. The resistivity of the cratonic segment ranges from 5000 $\Omega$m to over 30,000 $\Omega$m at 100 km depths, consistent with laboratory-derived dry olivine resistivities [Constable et al., 1992; Jones et al., 2009; Fullea et al., 2011]. The southernmost extent of the Congo craton lithosphere is prominent and at a depth of about 150 km directly beneath the Autseib lineament, where there is significant change in resistivity from 30,000 $\Omega$m to about a 1000 $\Omega$m.

The anisotropic models suggest elevated mantle conductivities in directions parallel to the profile ($R_{yy}$), particularly for the lowermost mantle beneath the Congo craton. As mentioned above, various candidates for mantle anisotropy include (but are not limited to) strain-induced crystal preferred orientation (CPO) of anisotropic minerals, hydrogen diffusivity in mantle minerals, and/or the presence of partial melt [Wannamaker, 2005; Yoshino et al., 2006]. Due to the lack of mantle xenoliths samples and...
Figure 16. Variation in RMS for all the components of the impedance tensor, depicting the confidence level across the 3-D block. Most stations returned a relatively good fit with a low RMS of less than 2, particularly the diagonal elements (Z_{xy} and Z_{yx}) and their corresponding phase responses.

...
geometry derived from modeling MT data. Major Archean cratonic segments (Kaapvaal and Congo) are delineated from Proterozoic (Rehoboth Terrain) and Neoproterozoic (Damara belt) regions by their thickness and resistivities, confirming again the general correlation of older lithosphere being thick and resistive, whereas younger lithosphere is thin and shows elevated conductivity.

[40] Due to the extensive regional size of the study area, it is important to estimate the depth of investigation at all the stations in order to gain confidence in the derived 2-D and 3-D models. Figure 13a shows the maximum Niblett-Bostick penetration depths interpolated across the 3-D data space, with the Central Damara conductor being most prominent as it results in limited penetration of EM fields. Generally, for most stations, penetration depths were in excess of 200 km (Figure 13b). It must be recognized that the penetration depths are only from a 1-D approach but give a qualitatively accurate view.

[40] Figure 14 shows a perspective view of the 3-D inversion result as N-S profiles across the model block. The thick resistive Congo craton (CC) lithosphere is clearly mapped, including the Central Damara belt conductor (DGCc). In order to obtain an indication of the variation in geometry of the tectonic features across the region, 2-D profiles were extracted out of the 3-D inversion model, and the results are shown in Figure 15 for the ETO-KIM, DMB, NEN, and OKA-WIN profiles from west to east. The TSUM S wave receiver function site is also shown on the DMB profile.

[41] The nature of 3-D inversion is such that more data (i.e., full impedance tensor) are modeled; as a result, one expects the 2-D and 3-D results to differ (Figure 15). Additionally, variable grid size, lack of data, misfit distribution, and trade-off parameter will introduce variations in the models. The geometric variations of cratonic lithosphere observed between the profiles could be the result of this very fact. That being the case, the general features of the model that are robust are the cratonic lithosphere to the north and the presence of the Central Damara conductor, which were also observed in the 2-D models. The modeled responses fit the observed data well, as illustrated in Figure 16, which shows the RMS difference for all impedance tensor elements of the modeled responses and the resulting 3-D inversion model.

8. Discussion

8.1. Crustal Model

[42] The resistive granitic plutons comprise a significant percentage of the geological outcrop of the Central Zone and reflect massive amounts of volcanism [Gray et al., 2007]. There are three plausible mechanisms that could result in huge voluminous melt generation specific to the Damara Belt Central Zone: (1) volcanism generated during the initial 750 Ma rifting phase, (2) decompression melting of subducted Khomas oceanic lithosphere (560 Ma), and/or (3) melt derived from crustal recycling of turbiditic material during final 540–520 Ma collision [Gray et al., 2007]. Isotopic studies on granites [McDermott et al., 2000; Jung and Mezger, 2003], however, indicate that the granites in the Central Zone lack positive mantle signatures (i.e., positive εNd, low δ18O). S-Type granites, which are particularly most significant in volume in the Central Zone, have similar chemical and isotopic signatures to the Neoproterozoic turbidites in the southern zone, indicating a fertile “turbiditic” crustal source at midcrustal levels by dehydration melting [Gray et al., 2007].

[43] Jung and Mezger [2003] and McDermott et al. [2000] suggest that the Damara Central Zone, which has high thermal gradients (30–50°C/km) and high-temperature/moderate-pressure metamorphic conditions, underwent significant middle to lower crustal low-pressure melting that generated the voluminous posttec tonic granites imaged by the MT model. The elevated gradients correlate with the observed conductivity structure of the Damara Central Zone. More subdued geothermal gradients are observed in the Barrovian metamorphism-dominated marginal zones of the north and southern Damara belt.

[44] The detection of the well-known crustal conductor in the Damara belt still requires explanation, as its understanding and characterization has tectonic implications. Issues that still need attention regarding this conductor include its possible causes and, most importantly, its apparent spatial along-strike extent. These will be addressed now.

[45] Ritter et al. [2003] attributed graphite along fossilized shear zones as the cause of the observed conductor, based on field evidence of graphite-bearing marble units. Geological exposures indicate the presence of accessory sulfides and oxides (Ben Goscombe, personal communication, 2012) which could explain the presence of conductive anomalies in the DGC upper crust. Massive sulfide mineralization (Cu-Zn-Au) is also found, for example, in the Matchless Amphibolite belt in the Damara Southern Zone; the latter being thought to represent a possible relic oceanic lithosphere [Barnes and Sawyer, 1980]. This observation suggests that sulfides could have been introduced in the Damara crust in the same manner (subduction) as for the longest-known crustal conductivity anomaly, namely, the North American Central Plains (NACP) conductor [Jones et al., 1993, 2005b]. Furthermore, if the Matchless Amphibolite represents fossil oceanic lithosphere, it is reasonable to expect the presence of carbon in graphite form, which would elevate the conductivity significantly as it would provide an efficient conduction mechanism. It is worth noting that high heat flow is also associated with crustal conductors in the Wopmay orogen [Wu et al., 2005] and in the Trans-Hudson Orogen [Jones et al., 1993].

[46] While hydrothermal crustal fluids could be present in the middle to lower crust, it is highly unlikely to give rise to the high conductivities observed due to their relatively short residence times in the crust [Bailey, 1990] and lack of supporting field evidence of alteration. Thus, a combination of graphite and sulfides offer the most plausible explanation for the increased conductivity in the Damara Central Zone.

[47] The apparent large strike extent of the Central Zone conductor could be a result of regional metamorphic processes or structural placement of interconnected graphite/sulfides and speaks to coherent along-strike tectonic processes, as for the NACP above. The DGC Central Zone is a high-temperature/low-pressure metamorphic zone, bounded to the north and south by zones of intermediate temperature and pressure metamorphism, which define regions of structural thickening [Maloof, 2000; Gray et al., 2007]. It is possible that during periods of
accretion and marked shortening thickening, graphite was preferentially emplacement along grain boundaries, and deep shear movement facilitated graphite interconnectivity, an argument that was also invoked by Ritter et al. [2003]. This argument is supported by Kisters et al. [2004], who observed that during peak high-temperature/low-pressure metamorphism (circa 538–505 Ma), the central Damara zone underwent pure shear deformation with lateral-parallel orogen stretch. Furthermore, microscopic investigations [Walter, 2004] in the DGC Central Zone reveal that graphite shows a strong lattice-preferred orientation in the mylonitic core zones. This strain-induced graphite, which might have been concentrated during collision of Congo and Kalahari cratons, would explain the large lateral (along strike) extent of the observed conductivity anomalies.

[8] In summary, combining the electrical and metamorphic observations, the abundance of resistive S-type granites in the high-temperature Central Zone resulted from remelting of turbidites, facilitated by magmatic under-plating at lower crust levels, during the final collisional phase of the Congo and Kalahari cratons. The conductivity in the middle to lower crust appears to have multiple possible causal sources, with graphite along deep-seated shear zones being the most probable and sulfide mineralization being more dominant in the upper crust.

8.2. Mantle Lithosphere

[40] Mantle electrical resistivity is more sensitive to temperature variations than it is to composition [Jones et al., 2009]; thus, “colder” and thicker cratonic lithosphere are generally well-imaged with the MT method and can be distinguished from “warmer,” generally thinner, mobile belts. The top of conductive zones overlain by resistive features, such as the boundary between cratonic lithosphere and asthenosphere, can be efficiently imaged to within 10% with high-quality MT data [Jones, 1999]. For smoother models, such as those derived in our case, the lithosphere-asthenosphere transition is not sharply defined, but rather appears as a broad gradational boundary as consequence of smooth regularization.

[50] There are several definitions of what the LAB is (see, for example, Eaton et al. [2009]). Mantle adiabatic temperature and composition will determine the electrical conductivity of the asthenosphere [Evans et al., 2011], but for a completely dry peridotitic rigid lithosphere, the onset of the transition to asthenospheric conditions tends to occur at conductivities in the order of 10⁻² S/m and greater (global averages are for conductivities of order 0.04–0.2 S/m for the asthenosphere) [Jones, 1999]. While this transition is not sharply imaged for the central Damara belt, due to the shielding effect of the conductive mid-lower crust, for sites outside this conductive zones, where depth of penetration is assured, the boundary appears to be at depths of around 160 km. Similar depth estimates were found by Muller et al. [2009]. Fishwick [2010] estimated, from surface wave tomographic models, lithospheric thickness in the range 160–180 km from the DGC. The DGC’s signal for the lithosphere-asthenosphere boundary (LAB) from the SRFs at station TSUM is not well resolved, due to high 2σ error bounds, but is nevertheless estimated at 180 ± 20 km (see Figure 15 for the DMB profile). Thus, all three methods, MT, surface wave tomography, and SRF, give an LAB of 160–180 km. Forward modeling was carried out to test the resolution of the features in the model. This was done by various ways including locking some cells of features in the model and noticing appreciable changes in the RMS. Furthermore, some features, like cratonic lithosphere, were painted with 100 Ωm and the result inverted. The return of the feature in the inversion proved that cratonic segments are required by the model and depth estimates remained the same. Mantle xenoliths, showing lherzolitic and harzburgitic composition, found in the Cretaceous nephelinite plug within the Central Zone of the DGC, 17 km east of Swakopmund, suggest an elevated geotherm (90 mW/m² from Whitehead et al. [2002]) which is significantly higher than the Rehoboth Terrane (45 mW/m² from Muller et al. [2009]). This composition is typical for Proterozoic mantle. The elevated geothermal gradient in the Damara belt is possibly related to the postcollisional tectono-thermal events, evidence of which is widespread and in the form of granites particularly in the Central Zone. Furthermore, Whitehead et al. [2002] infers a heterogeneous, oxidized, and fertile-to-refractory mantle beneath Damara Central Zone. The xenolith geotherm is, however, based on relatively few samples and, if representative at all, reflects upper lithospheric mantle temperatures at the time of kimberlite eruption (75 Ma) suggesting lateral temperature and lithospheric thickness variations from the Rehoboth (180 km) to the DGC (160 km). As geochronical and age constraints for the DGC lithospheric mantle are absent, our data and results provide no constraints on whether the DGC lithospheric mantle might consist of a younger component, created and stabilized coeval with Pan-African orogenesis.

[51] What is immediately clear is that the envisaged boundary of the Congo craton by Corner [2008] and Begg et al. [2009] does not correlate with the mapped extension of the cratonic lithosphere from the electrical resistivity results. These authors have the margin close to MT station NEN022 on the NEN profile (Figure 3). The resistivity model suggests that this margin is approximately 100 km further south, close to station NEN118, and has a steep southward dip up to 80 km depth. The Tsukwe (in Namibia) and Nxa-Nxa kimberlite fields on the border between Namibia and Botswana (see Figure 3) are located on an area of resistivity gradients which spatially correlates with lithospheric thickness change from a thinner mobile belt to a thicker craton (see the NEN model, for example). A similar trend was remarked upon by O’Neill et al. [2005] and Jones et al. [2009] where, in general, diamondiferous kimberlites appear to occur on areas of lateral resistivity and velocity gradients at the edges of cratons, indicating shallowing of lithospheric roots.

[52] The Okavango Dyke Swam (ODS), which crosses the OKA-WIN profile at an area between OKA008 and OKA004, is imaged as a resistive feature on this profile. Compositionally, the ODS is made up of dolerite dykes of varying widths ranging from 0.2 to 67 m, which Miensopust et al. [2011] showed manifest themselves as an anisotropic feature in 2-D models. The ODS is hosted in similarly resistive granite-gneiss host rocks, such as the Kwando complex, the Rooibok Complex (analogous to Matchless Amphibolite in Namibia) and various Paleoproterozoic igneous basements obscured by younger strata (though magnetically distinct). Thus, while its field outcrops is well constrained, it is
not readily distinguishable from its resistive hosts on the
MT model.

8.3. Rifting in the Okavango Delta?
[53] There has been some suggestions that the East
African Rift System extends into northwestern Botswana,
based on persistent seismicity [Reeves, 1972], forward
modeling of magnetic and gravity data [Kinabo, 2007], and
anomalous heat flow [Chapman and Pollack, 1974]. All
geophysical experiments done on the Okavango Rift Zone
(ORZ), before ours, focused on shallow (less than 1 km)
structure [Modisi et al., 2000; Kinabo, 2007; Bufford et al.,
2012] and revealed the ORZ as a half-graben feature. If
indeed continental rifting is taking place, then one might
expect a classic rift signature of thinned lithosphere ac-
companied by a localized deep-seated high conductivity mantle
anomalies, as is observed, for example, in the Rio Grande
Rift [Hermance and Pedersen, 1980; Jiracek et al., 1983;
Hermance and Neumann, 1991]. None of these features is
observed on the electrical resistivity models, particularly
along the OKA-WIN profile. The lack of elevated mantle
conductivity suggests that incipient continental rifting, if
taking place, is generated from preexisting structural fea-
tures in the crust, as proposed by Birt et al. [1997] for the
Kenya Rift, and propagates downward without any mantle
upwelling signature.

8.4. Congo Craton Southern Extent
[54] In general, the MT models suggest that the Congo
craton lithosphere extends significantly southward beneath
the Northern Platform sequences of Damara belt. The
southern and northern DGC passive margin sequences,
which were deposited on a rift margin, appear to be over-
thrust onto both cratonic nuclei during continental amal-
gamation at 495 Ma, supporting studies by Goscombe
[2004], to give a doubly vergent orogen. This result means
that the current tectonic lines mapping the southern edge of
the Congo craton need to be revised and drawn further south.
[55] It is interesting to note the striking similarity between
our MT models and other Archean-Proterozoic margins
around the world in imaging steeply conductive features sep-
arating the two terrains. For example, the Central Australian
Suture Zone, which forms the contact between the North
Australian craton and the Proterozoic Warumpi province
[Selway et al., 2006, 2009], is mapped as a dipping
conductor. Also, in Australia, the Errabiddy Shear Zone
forms a conductive suture between the Yilgarn Craton and
the Glenburgh Terrane [Selway et al., 2009]. The margin
between the Slave craton and the Wopmay orogen [Spratt et al.,
2009] is preserved as steeply dipping conductive struc-
ture. All of these features were subjected to later reactiva-
tion postcollisional collision. This observation suggests that
steeply dipping crustal-scale conductive structures can be
used to infer continental accretion, and perhaps, broad gen-
eralization can be inferred about similar mode of evolution
for Archean/Proterozoic margins.

9. Conclusions
[56] The three dominant features evident in the 2-D and
3-D MT models are (i) the Pan-African granitic plutons, (ii)
the mid-lower crustal conductor of the Damara belt, and (iii)
the thick resistive lithosphere of the southern Congo craton.
The DGC is thought to have experienced a full Wilson cycle,
including initial continental rifting (770–746 Ma) leading
to the development of the Khomas sea basin at 770 Ma,
followed by the subduction closing of Khomas ocean. The
presence and spatial extent of Congo cratonic lithosphere
was mapped in northern Namibia for the first time using
magnetotelluric data. The two-dimensional resistivity mod-
els reveal features which correlate with known structural
and lithologic features. The upper crust of the DGC is char-
erized by both resistive and conductive structures, the
former interpreted to be granitic intrusions in the Central
Zone related to the Pan-African magmatic events. A middle
to lower crustal conductor is confirmed, the origin of which
is observed in the DGC which we propose to be related pref-
erential alignment of graphitic/sulfide materials during the
collision of Kalahari and Congo cratons as part of Gondwana
amalgamation. The southward steeply dipping structure (cor-
responding to the Autseib Lineament) we propose is a
feature that is distinct to Archean/Proterozoic margins and
that it should be used to target and distinguish terranes of
different ages.

[57] The MT models constrain the previously unknown
position and geometry of the southern Congo craton at depth
and provide the first electrical resistivity map of the crustal
and mantle lithosphere beneath this previously unknown
boundary. The lack of mantle signature typically associated
with rifts is not observed in the MT models which suggest
that if indeed incipient rifting is taking place in northwestern
Botswana, the process is initiated from surface downward.

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Tectonic model of the Limpopo belt: Constraints from magnetotelluric data

D. Khoza\textsuperscript{a,b,*}, A.G. Jones\textsuperscript{a}, M.R. Muller\textsuperscript{a}, R.L. Evans\textsuperscript{c}, S.J. Webb\textsuperscript{b}, M. Miensopust\textsuperscript{d}, the SAMTEX team

\textsuperscript{a} Dublin Institute for Advanced Studies, 5 Merrion Square, Dublin, Ireland
\textsuperscript{b} University of the Witwatersrand, 1 Jan Smuts Avenue, Johannesburg, South Africa
\textsuperscript{c} Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, MA 02543-1050, USA
\textsuperscript{d} Institut fur Geophysik, Westfälische Wilhelms Universität, Munster, Germany

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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 anatetic 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 Sonnyside-Palala-Tshipise-Sheer 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.

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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...
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 Southern African Magneto Telluric Experiment (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
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 charnockitic 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 Southpansberg basin, which forms a 40 km wide, 300 km long, 7 km thick volcanoclastic-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 $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ 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 $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ 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 crustonic blocks have been central to some of the proposed collisional models of the

### Table 1

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
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<tbody>
<tr>
<td>KC</td>
<td>Kaapvaal craton</td>
</tr>
<tr>
<td>SMZ</td>
<td>South Marginal Zone</td>
</tr>
<tr>
<td>CZ</td>
<td>Central Zone</td>
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<tr>
<td>NMZ</td>
<td>North Marginal Zone</td>
</tr>
<tr>
<td>ZC</td>
<td>Zimbabwe craton</td>
</tr>
<tr>
<td>TSZ</td>
<td>Triangle Shear Zone</td>
</tr>
<tr>
<td>PTSZ</td>
<td>Palala-Tshipise Shear Zone</td>
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<tr>
<td>SPTSS</td>
<td>Sunnyside-Palala-Tshipise Shear System</td>
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<tr>
<td>NLTZ</td>
<td>North Limpopo Thrust Zone</td>
</tr>
<tr>
<td>HRSZ</td>
<td>Hout-River Shear Zone</td>
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<tr>
<td>SSZ</td>
<td>Sunnyside Shear Zone</td>
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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.
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 magnatism) (Eglington, 2004). The very widespread platform sedimentation of the Transvaal Supergroup that overlies the Ventersdorp Supergroup represents exceptional early stability of the Kaapvaal craton. Bushveld complex (Fig. 2), which intruded into the Kaapvaal craton and 1 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 (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 transcurrent 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 Sunny-side-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
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-crust/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; Gwawusa 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 (Ex, Ey) and horizontal and vertical magnetic (Hx, Hy, and Hz) 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 $Z_{xy}(ω) = \frac{E_x(ω)}{H_y(ω)}$ from which the apparent resistivity (i.e., the resistivity of a homogeneous half space)

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

and impedance phase

$$\phi_{xy}(ω) = \arctan \left( \frac{E_x(ω)}{H_y(ω)} \right)$$

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) MT5 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
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 southwestern 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 Kaapvaal craton 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 (Zxx, Zxy, Zyy, Zyx) and Hz 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 Zxx, Zxy, Zyy, Zyx were modelled, using a tau value of 3 and 100 ohm-m half-spaces as input).
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 KAPO74 and LOW001 are clustered around the 2.57 Ga Bulai pluton. Compositionally the pluton is charnockitic, 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

![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.](image)

![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.](image)
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 kimberlites 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.

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 located 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
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 (Klem 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 2m) 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-liquids 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 Veneta kimberlite cluster. This structure consists of supracrustal rocks down to 10 km depth, overlaying 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/cm³). 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 Mouru (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 difficult 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 CO₂-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 granites 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 CO₂ 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 CO₂, 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.,
It is therefore reasonable that graphite was precipitated from the CO₂-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 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; Tourret 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 CO₂ and brine rich fluids sourced from devolatilisation 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 PTZS 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 PTZS as a conductive feature and suggest that it in fact represents a fundamental suture between the Kaapvaal and Zimbabwe cratons. The PTZS appears to be sub-vertical on the KAP and LIMSSO profiles and dips slightly to the north on the LOW profile. On a mantle lithospheric scale, the PTZS 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 PTZS 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 PTZS, suggesting a transpressive collisional event. Given that the Soutpansberg basin is located in the SMZ just south of the PTZS, it is possible that the deposition of volcanic-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 aurolacogen 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 intercratonic sutures have played a significant role in the evolution of the African tectonic landscape, by focussing ascending magmas and areas of localized rifting (Reegg et al., 2009). The close spatial proximity of the PTZS 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 Mangodi 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 Neoproterozoic 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 SPTSZ 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; Kamnunz 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 anatexis at this time. Deeper melting of remnant oceanic crust resulted in CO₂-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
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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Quantifying South Africa’s carbon storage potential using geophysics

Along with many other nations, South Africa faces the challenge of curbing carbon emissions, as coal-fired power plants generate 92% of the total electricity used. The result of this generation is the release into the atmosphere of 400 million tonnes of CO₂ annually, which contributes to the greenhouse gases that have a detrimental effect on global climate. Given the length of time needed to implement renewable energy sources, an alternative solution to significantly reduce greenhouse gas emissions is to capture the CO₂ produced by coal-fired power stations and store it in geological formations in the subsurface – a process broadly called carbon capture and storage. In order to successfully achieve this sequestration, a mechanism must exist to monitor the behaviour of the CO₂ injected into the earth.

Accordingly, in 2009 the South African government established the Centre for Carbon Capture and Storage, a division of the South African National Energy Development Institute. The Centre is tasked with the research and technical development of carbon capture and storage. The establishment of the Centre was followed in 2010 by the publication of an atlas which identifies and ranks potential CO₂ storage sites, mostly in Mesozoic basins along the coast (Outeniqua, Orange and Durban/Zululand basins), and to a lesser extent the Karoo Basin (Figure 1). The theoretical study that led to the production of the atlas was based largely on a literature review of all available boreholes and other geological information. What is needed is a more quantitative way of imaging potential storage sites, and I aim to address this need here.

The main challenge in carbon capture and storage is identifying those localities and geological settings within South Africa which have the greatest potential to store significant volumes of CO₂. The capture and storage of carbon is analogous to how oil and gas are naturally trapped in underground formations. Thus an ideal storage site must comprise a porous and permeable medium (e.g. sandstone) where CO₂ can be injected and stored, overlain by an impermeable cap rock (e.g. shale) that will retain (by dissolution or adsorption) the injected material and prevent it from moving or escaping into the atmosphere. Geological storage options for deep injection of CO₂ are numerous and include both onshore and offshore sites. One such onshore site is the Karoo Basin, where the authors have conducted research using geophysical techniques to image areas of interest for CO₂ storage.

**FIGURE 1:** Map showing the distribution of potential onshore and offshore CO₂ storage sites and the locations of magnetotelluric stations in the Karoo Basin.
CO₂ include depleted oil and gas formations, deep unmineable coal seams and deep saline formations; the latter offering perhaps the most potential in the South African context.

One of the potential storage sites identified is in the Karoo Basin (Figure 1). Given the reported low permeability and porosity of the Ecca Group in the Karoo Basin, the potential for CO₂ storage in the region has been inferred as low; more quantitative work needs to be undertaken to determine if this is the case. To this end, I am using a geophysical remote sensing method, magnetotellurics (MT), to provide quantitative estimates as to the storage potential of the Karoo Basin. The MT technique is a deep imaging geophysical technique whereby naturally occurring electric and magnetic fields (in the frequency range 1000 Hz – 0.001 Hz) are measured on the surface of the earth to determine the resistivity structure of the subsurface (from a few hundred metre to tens of kilometres). The resistivity of a rock formation is a function of four parameters, (1) the porosity of the rock that is occupied by a fluid; (2) the degree of interconnection of the fluid; (3) the resistivity of the host rock; and (4) the salinity of the groundwater. Thus, by knowing the resistivity of a geological formation we can, in principle, determine the rock properties (porosity and permeability) needed for reservoir characterisation using Archie’s Law. This principle is illustrated in Figure 2, which shows a porosity–resistivity–salinity nomogram that can be used to estimate porosity from bulk resistivity measurements. Thus, if one has temperature and salinity measurements (for example from boreholes) and resistivity from MT results to plot on a nomogram, connecting the points with a line of best fit would yield the resistivity of the pore fluid, which could in turn be used to estimate porosity.

In southern Africa we have collected over 750 MT sites as part of the highly successful Southern African Magnetotelluric Experiment, in order to study the crustal and mantle structure of the region. Figure 2 shows an example of a resistivity response from one MT site in the Karoo Basin, which was derived from the processing of recorded electromagnetic responses. The two curves essentially represent apparent resistivity variations as a result of induced electrical current flow in directions parallel (transverse electric) and perpendicular (transverse magnetic) to the north-west to south-east profile. The abscissa represent the period in seconds (the inverse of frequency in Hz) which is a proxy for depth in kilometres. One can use the MT responses like these collected in the Karoo Basin to characterise one of the potential CO₂ storage sites shown in Figure 1. The MT sites are spaced at intervals of approximately 10 km – 15 km along a north-west to south-east profile. The acquisition of MT data is usually done along two-dimensional profiles, and at each site horizontal variations in electric and magnetic fields are recorded, using non-polarising electrodes for the former and magnetometers for the latter. For optimal resolution of geological formations such as in the Karoo Basin, much more detailed data is required. The processing, analysis and modelling of MT data can be done in one, two and three dimensions. It is hoped that upon successful characterisation of onshore storage sites using MT, data acquisition will be extended to offshore basins.

South African activities in carbon capture and storage are still in their infancy, but rapid progress is being made in developing applicable research methodologies that will guide the understanding of carbon capture and storage in the long term. This account is an attempt to explain the geophysical research that is being done to understand and characterise the carefully chosen sites within South Africa that have been identified to have the potential to store significant amounts of CO₂. Given its sensitivity to bulk rock properties and its non-intrusive nature, the magnetotelluric technique provides
by far the most promising way of mapping, modelling and monitoring onshore and offshore sites pre- and post-injection. It is thus recommended that this technique be employed at a finer resolution (a site spacing of 2 km) in addition to employing seismic methods, particularly in areas where near-surface basaltic layers (such as in the Karoo basin or offshore basins) attenuate seismic signals quite rapidly.

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