An Electrical Resistivity Model of the Southeast Australian Lithosphere and Asthenosphere

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Kate Robertson

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ABSTRACT

A combination of magnetotelluric and geomagnetic depth sounding data were used to attempt to image the electrical resistivity structure of southeast Australia, to investigate the physical state of the crust and upper mantle. A 3D forward model of southeast Australia comprised of regional sets of broadband and long-period magnetotelluric and geomagnetic depth sounding data, over an area of 440 x 300 km², was used to map broad-scale lithospheric properties. Model results show an order of magnitude decrease in resistivity from the depleted continental mantle lithosphere of the Delamerian Orogen in the west, to the more conducting oceanic mantle of the Lachlan Orogen in the east. The decrease in resistivity in conjunction with a 0.1 km/s decrease in P-wave velocity at depths of 50-250 km, suggest a change in temperature ($\Delta T \sim 200^\circ C$) due to lithospheric thinning toward the east as the likely cause, in conjunction with a change in geochemistry and/or hydration. A high resolution two-dimensional inversion using data from 37 new and 39 existing broadband magnetotelluric stations mapped crustal heterogeneity beneath the Delamerian Orogen in much greater detail. Lateral changes in resistivity from 10-10,000 $\Omega m$ occur over the space of a few kilometres. Low resistivity ($\sim 10 \Omega m$) regions occur at depths of 10-40 km. Narrow paths of low resistivity extend to the surface, coinciding with locations of crustal faults from seismic interpretations. Movement of mantle fluids up these faults, during periods of extension prior to the Delamerian Orogen, may have produced a carbon-rich, low resistivity lower crust, leaving a resistive upper mantle, depleted of volatiles.

KEYWORDS

electrical conductivity, resistivity, magnetotellurics, lithosphere, crust, upper mantle, Delamerian Orogen, Lachlan Orogen, southeast Australia
Table of Contents

Introduction 7

Lithospheric Resistivity 9

Previous Results From Seismic Studies 10

MT Theory 11

MT Distortion and Phase Tensor Theory 13

Geomagnetic Depth Sounding 15

Geological Background 16

Observations and Results 18

3D Regional Modelling 18

3D Forward Model 18

Crustal scale MT 22

Data Acquisition 22

Data Processing 23

Responses, Strike and Dimensionality 23

Phase Tensor Analysis 25

2D Inversion 27

Occam 2D Model with Seismic Interpretation 27

Sensitivity analysis 29

Occam 2D Response Curves 29

Discussion 30

3D Forward Model 30

2D Inversion 31

Comparison of MT with Seismic Interpretation 31

Possible Causes of Mid-Lower Crustal Low Resistivities 31
List of Figures

1. Elliptical representation of the phase tensor ........................................... 15
2. Geological map of southeast Australia with MT and GDS station locations .......................................................... 16
3. Cross-section of P-wave velocity structure beneath the Delamerian and Lachlan Orogens, with electrical resistivities for forward model included ........................................... 19
4. Induction arrows of two forward models at 0.01 Hz; One with a west-east decrease in resistivity across velocity boundary between Delamerian and Lachlan Orogens, one unchanging across boundary .......... 21
5. Pseudosection of TE and TM modes, indicating data quality and static shift effects ........................................... 24
6. Geoelectric strike .................................................................................. 25
7. Phase tensor pseudosection ................................................................ 26
8. Phase tensor ellipses over a total magnetic intensity map ................. 27
9. Occam smooth 2D inversion of all data points of the Southern Delamerian transect, SD01’ ........................................... 28
10. Phases and apparent resistivity values for stations SD01, DB24C, DB30C and SD38 ......................................................... 30
11. Appendix A-Station information spreadsheet .................................... 39
12. Appendix B-Processing information spreadsheet .............................. 40
INTRODUCTION

A better understanding of the electrical resistivity structure underlying southeast Australia would improve the current knowledge regarding the physical state of the crust and upper mantle. Uncertainties involving this physical state include: hydration of the upper mantle, rheological controls defining the depth and nature of the lithosphere-asthenosphere boundary, formation and deformation history of the sub-continental lithospheric mantle, and broad-scale heterogeneity of the continental lithosphere and asthenosphere to the transition zone.

With the exception of xenolith/xenocryst data (O’Reilly & Griffin 1985), geophysical surveys are the only means of investigating depths associated with the mid to lower crust and upper mantle. Gravity, magnetic, electromagnetic (EM) and reflection seismic (Brewer & Oliver 1980; Korsch et al. 2002) surveys can all image crustal depths. Reflection seismics show major faults and structural boundaries within the crust, which preserves the brittle and ductile footprints of events (Brewer & Oliver 1980). Magnetotellurics (MT) (Dennis et al. 2012; Hamilton et al. 2006; Heinson & White 2005; Jones 1992) and seismic tomography (Graeber et al. 2002; Rawlinson & Kennett 2008; Rawlinson & Fishwick 2011) are the only methods in common use capable of investigating sub-Moho depths, with the latter commonly used for estimating elasticity (a function of temperature and composition) in the upper mantle to asthenosphere. Seismic tomography is limited in its ability to resolve crustal and even some upper mantle structures (Rawlinson & Fishwick 2011).

The signature of the geological processes that determine the physical state of the lithosphere are substantially different at mantle and crustal scales. MT has played an increasing role at both mantle and crustal scales (Bedrosian 2007), with different approaches often used for deep lithosphere (grid array of MT sites) and crustal structures (transects of MT sites), however these approaches are interchangeable. MT is
a passive EM technique that records the electrical resistivity response of the Earth as a function of different inducing frequencies (Cagniard 1953; Tikhonov 1950). Low resistivity (i.e. high conductivity) could be due to changes in one or more of the following: temperature, melt fractions, hydration and trace mineralogy, making it a diverse method for determining the physical state of the Earth (Selway 2012). For maximum benefit, combined interpretation of MT in conjunction with other geophysical data sets is recommended (Jones 1987).

In this study, a series of 3D forward models were used to determine the regional-scale variability of the southeast Australian electrical resistivity structure. Two forward models were constructed to investigate if a change in electrical resistivity occurred at the location of a change in velocity of 0.1 km/s where the Delamerian and Lachlan Orogens meet (Graeber et al. 2002; Rawlinson & Fishwick 2011). New and existing broadband (0.0064s-81.92s) MT data of the Southern Delamerian transect were then used to determine the electrical resistivity structure of the Delamerian Orogen subsurface in more detail.

Both of these new models of electrical resistivity, combined with existing passive seismic velocity models, geometric boundaries from reflection seismology and additional geochronological restraints were used to better constrain the physical state of the lithosphere in southeastern Australia. Particular items addressed in this study include; the length and resistivity-scale over which changes in the lithosphere occur, the temperature and hydration state of the lithosphere from electrical resistivity and seismic tomography, comparisons of mantle and crustal heterogeneity, and how crustal heterogeneity relates to crustal deformation imaged by reflection seismics and mapped by geology and geochemistry.
LITHOSPHERIC RESISTIVITY

Low resistivity regions in the upper mantle are caused by higher temperatures, interconnected conducting phases, hydrated minerals (Karato 1990) and iron content (Jones 1999). Lilley et al. (1981) have found that at depths greater than 150 km, electrical resistivity from central Australia to southeast Australia decreases by an order of magnitude. The suggested explanation for this, at the time of the study, was the presence of a small fraction (approximately 5%) of basaltic melt. Studies by O’Reilly & Griffin (1985) also suggested elevated temperatures in southeast Australia when compared with normal continental geotherms.

The possible existence of a partial-melt zone under southeast Australia is supported by EM measurements, seismic studies and thermal models which involve crustal intrusion from a sub-lithospheric magma source. Gaillard et al. (2008) proposed that carbonatite melt in the top of the asthenosphere is capable of producing conductive anomalies present in the upper asthenosphere. A study by Yoshino (2010) suggests at pressures of 3 GPa, carbonatite melt bearing olivine aggregates show an order of magnitude lower resistivity than silicate melt bearing olivine aggregates. Yaxley et al. (1998) provide evidence for the presence of carbonatite in melts in southeast Australia, offering an alternate explanation to basaltic melt (Lilley et al. 1981) causing the lower resistivity to the east.

Low electrical resistivity (i.e. high conductivity) regions within the crust are commonly explained by highly interconnected, conductive phases in grain boundary areas (e.g. Wei 2001). Yang (2011) states that this mechanism for low resistivity is not always supported by geochemical and geophysical data. He suggests that in parts of the lower continental crust (particularly beneath stable areas), low resistivity can often be accounted for by solid-state conduction of the main constituents, due to chemical composition (including some major elements and water content),
temperature, and mineral fabrics.

The South African MT Experiment (Hamilton et al. 2006), found that the crustal MT results obtained in South Africa are strongly related to large-scale geological structures within the Kapvaal Craton and the surrounding terranes. Resistivity shows much greater variability at crustal depths than for upper mantle depths, with only some crustal structures still distinct at mantle depth.

The use of 3D forward modelling and 2D inversions of MT and geomagnetic depth sounding (GDS) data will determine if crustal resistivity is heterogeneous at smaller scale-lengths than upper mantle depths in southeast Australia.

**PREVIOUS RESULTS FROM SEISMIC STUDIES**

Rawlinson and Fishwick (2011) constructed a high-resolution, 3D P-wave velocity model of the southeast Australian upper mantle, to a depth of 300 km. A distinct feature that the model resolved was a higher velocity region that they associated with the Delamerian Orogen, extending beneath the lower velocity Lachlan Orogen separated by the presence of an easterly dipping velocity transition zone, most prominent at depths of 150-200 km. The transition zone is marked by an increase in velocity at these dpeths of $\sim 0.1$

The sudden variation in velocity of 0.1 km/s across the boundary could be a result of elevated temperatures. Higher temperature to the west could arise from lithospheric thinning towards the east as a result of the break-up of Australia and Antarctica and the opening of the Tasman Sea. The velocity model supports a change from Proterozoic mantle lithosphere of continental origin to Phanerozoic mantle lithosphere of oceanic origin, and is consistent with the changes in both
composition (Cammarano et al. 2003; Griffin et al. 1998) and temperature (Faul & Jackson 2005) required for these two mantle types.

**MT THEORY**

The MT method was first suggested by Tikhonov (1950) and later expanded upon by Cagniard (1953). A summary of the theory of MT will be presented here. For a comprehensive theory of MT see Simpson (2005) and Chave (2012). Unless stated otherwise these books (and references therein) are the source of information for this MT Theory section.

MT is a passive EM technique capable of determining the electrical resistivity response of the Earth down to depths of more than 600 km. MT uses naturally occurring geomagnetic variations external to the Earth as a power source for EM induction. The fluctuations in the natural magnetic fields and induced electric fields are measured in orthogonal directions at the Earth’s surface. The distance over which EM fields diffuse is dependent on the frequency of the variations and the resistivity of the Earth, allowing MT to yield both lateral and vertical constraints. The periods of these variations that are naturally occurring are of the order of $10^{-3}$ to $10^{5}$ s (frequencies $10^{3}$ to $10^{-5}$ Hz). At frequencies of 0.5-5 Hz, the natural source signal has low amplitude, known as the *dead-band*, resulting in a low signal-to-noise ratio in the measured data.

In the frequency domain, complex ratios known as impedances can be derived, that describe the natural variations of the electric ($\mathbf{E}$) and magnetic fields ($\mathbf{H}$). The
impedance tensor $\mathbf{Z}$ is defined as follows:

$$
\begin{pmatrix}
E_x \\
E_y
\end{pmatrix} =
\begin{pmatrix}
Z_{xx} & Z_{xy} \\
Z_{yx} & Z_{yy}
\end{pmatrix}
\begin{pmatrix}
H_x \\
H_y
\end{pmatrix}
$$

(1)

For a 1D resistivity structure, the diagonal components $Z_{xy} = -Z_{yx} = Z$, $Z_{xx} = Z_{yy} = 0$. For a purely 2D structure, the impedance tensor can be rotated to eliminate $Z_{xx}$ and $Z_{yy}$. The TE mode has the electric field parallel to the strike, and the TM mode has the magnetic field parallel to the strike.

$\mathbf{Z}$ is complex and can be separated into real ($\mathbf{X}$) and imaginary ($\mathbf{Y}$) parts:

$$
\mathbf{Z} = \mathbf{X} + i\mathbf{Y}
$$

(2)

The components of the matrix $\mathbf{Z}$ have both magnitude and phase;

The apparent resistivity ($\rho_{a_{ij}}(\omega)$ - the magnitude of $\mathbf{Z}$, where $\omega = 2\pi f$ is the angular frequency), is the average resistivity for the volume of Earth penetrated by a particular MT sounding period, $T$, where $T = \frac{1}{f}$, and is given by the equation:

$$
\rho_{a_{ij}}(\omega) = \frac{1}{\mu_0 \omega} |Z_{ij}(\omega)|^2
$$

(3)

. If the Earth were homogeneous, the apparent resistivity would be the actual resistivity. However for the true, multi-dimensional Earth, the apparent resistivity is actually the average resistivity represented by a uniform half-space (a volume bounded only by an infinite plane).

The phase (component of $\mathbf{Z}$, $\Phi_{ij}$) represents the phase difference between the electric and magnetic fields that are used to calculate the MT impedance, and is equal to:

$$
\Phi_{ij} = \tan^{-1}\frac{\Im\{Z_{ij}\}}{\Re\{Z_{ij}\}}
$$

(4)
For a uniform half-space, impedance phases are $45^\circ$. Impedance phases $>45^\circ$ are indicative of resistivity decreasing with depth, and impedance phases less than $45^\circ$ means resistivity increases with depth.

Penetration depths depend on the EM sounding period and the Earth’s resistivity structure. The depth of penetration of which resolvable signal is obtained approximately equals one skin depth (the depth at which EM fields are attenuated to $e^{-1}$ of their surface amplitude), where a skin depth $\rho(T)$ in simplified form is given by:

$$\rho(T) = 500\sqrt{T\rho}$$  \hfill (5)

From this it can be seen that at longer periods, waves are attenuated more slowly than shorter periods, and therefore penetrate deeper into the Earth.

**MT Distortion and Phase Tensor Theory**

Heterogeneities in the near-surface resistivity cause galvanic effects, which distort the regional MT response, and static shift, which distorts the depth and resolution of underlying structures (Jones 1988). The phase tensor is extremely useful for strike and dimensionality interpretation, because it preserves the regional phase information, and is the component of the impedance tensor that is unaffected by galvanic distortion or static shift (Caldwell et al. 2004; Bibby et al. 2005). Use of the phase tensor is suitable even if both heterogeneity and resistivity structures are 3D, as no assumptions about the resistivity dimensionality are required to calculate it. The phase tensor, $\Phi$ is represented by the equation

$$\Phi = X^{-1}Y$$  \hfill (6)
where $X$ and $Y$ were defined in Equation 2, as the real and imaginary components of $Z$. The coordinate invariants of the phase tensor are the minimum and maximum phases and the skew angle. For 1D regional electrical resistivity structures, the phase tensor is represented by a single coordinate invariant phase equal to the 1D impedance tensor phase. For 2D regional electrical resistivity, the phase tensor is symmetric with one of its principal axes aligned parallel to the regional strike. The geoelectric strike is the preferred direction of current flow and is given by $\alpha$ or $\alpha + 90$ (as there is ambiguity between the two principal axes (Caldwell et al. 2004)), where:

$$\alpha = \frac{1}{2} \tan^{-1} \frac{\Phi_{xy} + \Phi_{yx}}{\Phi_{xx} - \Phi_{yy}}$$  \hspace{1cm} (7)

. In the 3D case the phase tensor is non-symmetric and a third coordinate invariant, the skew ($\beta$), is introduced. The skew is a measure of the asymmetry of the regional MT response, and is given by:

$$\beta = \frac{1}{2} \tan^{-1} \frac{\Phi_{xy} - \Phi_{yx}}{\Phi_{xx} + \Phi_{yy}}$$  \hspace{1cm} (8)

. The phase tensor can be represented graphically by an ellipse (Caldwell et al. 2004) (Figure 1), which gives a visual indication of resistivity-dimensionality and strike direction. The major and minor axes of the ellipse represent the principal axes of the tensor. The orientations of the phase tensor principal axes reflect lateral variations (gradients) in the regional resistivity structure. Phase tensor pseudosections (see Figure 7) are a convenient way of visualising these gradients.
Figure 1: Elliptical representation of the phase tensor. The ellipse axes represent the principal axes of the phase tensor and \( \alpha \) represents the geoelectric strike. If the phase tensor is non-symmetric (i.e. the resistivity structure is 3D), a third coordinate invariant known as the skew is needed (represented by \( \beta \)).

**Geomagnetic Depth Sounding**

Geomagnetic depth sounding was developed when Schuster (1889) showed the existence of magnetovariational fields arising from induction. Like MT, GDS determines the electrical resistivity structure of the subsurface. However, in MT, where either 4 or 5 components are measured \( (E_x, E_y, H_x, H_y, \pm H_z) \), only the three magnetic components are measured \( (H_x, H_y, H_z) \) in GDS. The advantage of not measuring the electric fields in GDS, is immunity from the effects of local distortions of the electric field, manifest as static shift (Jones 1988) and galvanic distortion (Bibby et al. 2005). A reference station outside of the survey area is used to improve signal to noise ratios (Gamble et al. 1979). GDS can probe over scale-lengths both laterally and vertically of a few hundred kilometres, as it detects lateral variations in resistivity.

In most studies, induction arrows are plotted for graphical representations of these lateral heterogeneities in the resistivity structure. These induction arrows
are vector representations of the complex ratios of vertical to horizontal magnetic field components. The vertical magnetic fields are generated by lateral resistivity gradients, so induction arrows can be used to infer the presence or absence of lateral variation in resistivity (Whellams 1989). The Parkinson convention for induction arrows is most commonly used (Parkinson 1959), where the arrows point towards conductive bodies. The length of the arrow is determined by a combination of the magnitude of the anomaly and its distance from the MT site, with range increasing for longer periods.

GEOLOGICAL BACKGROUND

Figure 2: Simplified geological map with approximate locations of younger cover sequences. The locations of all MT and GDS stations used in this study have been included. The east-west (SD01') and north-south (SD02') Southern Delamerian MT transects are highlighted with a blue box around it. The inset shows the region in the blue box zoomed in, with stations used in Figure 10, labelled. Note the transect crosses the Glenelg Zone, the Yarramyljup Fault and the Grampian Stavely Zone. Adapted from Rawlinson and Fishwick (2011).
The southeast Australian lithosphere formed during a Palaeozoic Orogeny that commenced in the Early Cambrian and ceased in the Middle Triassic (Glen 2005). The new and reworked lithosphere from this orogeny, covering the eastern third of Australia is collectively known as the Tasmanides (Glen 2005). Orogens comprising the Tasmanides include the Delamerian Orogen, the Lachlan Orogen to the east and the New England Orogen along the east coast of Australia (Foster 2000).

The Delamerian Orogeny began 514 ± 3 Ma (Foden et al. 2006), lasting for approximately 24 m.y., ending 490 Ma with a period of rapid uplift, cooling and extension. The Delamerian Orogen consists of Precambrian and Early Cambrian rocks that experienced deformation and metamorphism in the Cambrian (Foden et al. 2006). The Australian Precambrian cratons are separated from the younger Palaeoico to Mesozoic orogenic belts of eastern Australia by the Delamerian Orogen. Westward-verging folds and thrust faults are abundant in the region (Jenkins & Keene 1992). The orogen comprises the Adelaide Fold belt in South Australia, the Glenelg Complex in western Victoria and parts of western New South Wales and Tasmania. The MT transect, SD01’, of which the second phase of data collection was part of this study, runs east-west through the Adelaide Fold Belt and across the Glenelg and Grampian-Stavely Zones (see Figure 2).

The Ordovician-Devonian Lachlan Fold Belt in western Victoria developed from about 485 to 340 Ma (Kemp et al. 2005). The geodynamic plate tectonic setting for the Lachlan fold belt is still unresolved but may include an accretionary orogen system (Foster 2000), with numerous complications such as the inclusion of rifted Proterozoic continental fragments into the region during deformation (Cayley et al. 2002; Cayley & Taylor 1997).
The location of the boundary between the Delamerian Orogen and the Lachlan Orogen has been a topic of debate (Vandenberg 1978; Glen 1992; Baille 1985). A more recent study by (Miller et al. 2005) deduced that the boundary is likely to be complex and transitional across the vicinity of the Stawell Zone in western Victoria (Miller et al. 2005).

In addition to this Palaeozoic history, this region lay just to the north of the breakup of Australia and Antarctica. This resulted in the formation the Otway Basin and the opening of the Tasman Sea, between 80-90 Ma, causing lithospheric thinning near the present passive margin of southeast Australia. Mafic underplating may be associated with the formation of the passive margin (Rawlinson & Fishwick 2011). The Newer Volcanic Provinces, located in western and central Victoria, consist of substantial Quaternary basalt cover that originated from hot-spot volcanism (Price et al. 1997).

**OBSERVATIONS AND RESULTS**

**3D Regional Modelling**

**3D FORWARD MODEL**

WinGLink® (Rodi & Mackie 2001), a non-linear conjugate algorithm, was used to create a database consisting of; two broadband MT transects comprising 14 stations, a grid of 39 long-period MT stations (Aivazpourporgou et al. 2012), a grid of 25 GDS stations (Whellams 1989), in addition to the 39 existing and 37 newly attained (in this study- see subsections 'Data Acquisition' and 'Data Processing' for more information) broadband MT stations of the Southern Delamerian transects (see Figure 2 for all station locations).
Figure 3: A cross section of the P-wave velocity structure ($V_p$) from seismic tomography in southeast Australia, at 36.7° S, 138-150° E. Red indicates regions where P-waves travel slower, purple indicates faster regions. Depths greater than 300 km were not included in the seismic tomography model and are coloured grey. The location of the transitional velocity region at the proposed boundary of the Delamerian/Lachlan Orogens is the east-dipping dashed line at 142° E. The resistivity values marked on the figure indicate the values that were used in the second 3D forward model of MT and GDS data, Model B, to test if the change in velocity across the boundary could be modelled by a change in resistivity. Model A had no change in resistivity across the boundary, with all values set to those of the western half of Model B. The MT transect (SD01') is at approximately 140-142 ° E (not shown on this map). Adapted from Rawlinson and Fishwick (2011).

Construction of two 3D forward models aimed to put the newly acquired Southern Delamerian MT data into a regional framework. Resistivity values were set using findings from Heinson and White (2005). For depths down to 40 km in Model A, the first model, a 1D resistivity structure was assumed (Figure 3). Bathymetry and the Otway Basin sediments (location and depth as indicated by the cross sections from Petkovic (2004), were incorporated into the model. At depths greater than 40 km, resistivities to the west of the apparent seismic boundary between the Delamerian and Lachlan orogens were set one order of magnitude greater than those to the east. For the Newer Volcanic Provinces, which is a low velocity zone, resistivities were set anomalously low ($\sim$45 $\Omega$m at depths of 120-180 km) (Rawlinson & Kennett 2008; Rawlinson & Fishwick 2011). An alternative model, Model B, was
also constructed, with resistivites unchanging across the velocity boundary (all values were set to that of the resistive Delamerian Orogen in Figure 3). Both models extended to depths exceeding 800 km.

Forward modelling results are shown in Figure 4. The induction arrows for each model were plotted for frequencies of 0.1 Hz, 0.01 Hz, 0.001 Hz and 0.0002 Hz to investigate the fit of the two models with the actual data. The closer in magnitude and orientation the induction arrows are, the closer that the model created in WinGLink® is to the actual structure of the Earth. The modelled induction arrows reproduce the observed induction arrows better when the model incorporates a change in resistivity across the Moyston Fault, to correspond with the 0.1 km/s change in P-wave velocity (Figure 3). Figure 4 shows the best fit induction arrows, occurring at a frequency of 0.01 Hz. To the middle of the figure and further east is where the arrows of Model B show particular improvement over those of Model A. The lower resistivity in the east has rotated the modelled data arrows so that they have a more similar orientation to the observed data arrows. Induction arrows at frequencies higher than the 0.01 Hz (at shallower depths) plotted in Figure 4, do not show any trends and point in many different directions, due to local effects of low resistivity crustal regions.

The MT stations of SD01’ and SD02’, do not have induction arrows as only the x and y (i.e. horizontal) components of the magnetic field were recorded; the z-component is also required to create induction arrows. However, three stations from Aivazpourporgou et al. (2012) long-period grid overlap the eastern end of SD01’, with induction arrows for periods greater than 100 s (frequency <0.01 Hz) fitting well in this region, indicating that below the transect, the mantle is predominantly homogeneous.
Figure 4: The real induction arrows of the two different forward models. The arrows point towards the anomalous internals concentrations of current. The red arrows are the induction arrows from the data (note that the Southern Delamerian stations do not have induction arrows). The blue arrows indicate the induction arrows created by the 3D WinGLink® forward model. How well the data fits the model is given by how closely the magnitude and orientations align.

a) The induction arrows for Model A, with no change in resistivity across the boundary.
b) The bottom figure shows the induction arrows for Model B, the model with the order of magnitude decrease in resistivity, west to east, across the boundary.
Crustal scale MT

DATA ACQUISITION

In June 2012, broadband MT data were collected at 37 sites along a 75 km east-west transect in west Victoria, Australia. Locations of the sites can be found in Appendix A. Data collection comprised the second phase of a two-part survey. The first 39 stations were collected in 2009 for Phase 1, along the main transect running east-west labelled SD01' (stations SD01-SD10, SD14, SD21-39), and a much shorter transect, SD02', (stations SD11-SD13, SD15-SD20) running north-south. The survey was run to coincide approximately with the 2009 Southern Delamerian AuScope seismic survey line.

The stations of Phase 2 were placed along the eastern half of the 140 km SD01' transect. The second survey was conducted because preliminary inversions of the original 39 stations (unpublished) suggested a much more heterogeneous resistivity structure to the east than the west. This corresponds with more complex and steeply dipping geology within this region. An extra 1-3 sites were infilled for each existing site SD24-SD39 (named DB24a, DB24b etc. see Figure 2). Combining the two surveys resulted in a 67 station, 140 km transect, SD01’, with 5 km spacing for stations 1-24, and approximately 1.2 km spacing for sites 24-39. Data were recorded at a frequency of 1000 Hz for a duration of 20-40 hours using AuScope equipment. Each station recorded two horizontal components of the electric field using 40-50 m dipoles (\(E_x, E_y\) with x north-south and y east-west) and two components of the magnetic field (\(H_x, H_y\)). At least 3 stations recorded simultaneously allowing for remote referencing (Gamble et al. 1979).
DATA PROCESSING

The time series data for each site were processed and converted from the time-domain to the frequency-domain, using a robust remote referencing code, BIRRP (Chave & Thomson 2004). EDI files were output from this code, which contain information regarding the impedance tensor in the frequency domain. 50, 150 and 250 Hz noise occurred when stations were situated near powerlines, and were removed using a notch filter. A regional geomagnetic declination of approximately 10° has been corrected for, ensuring that the x and y directions represent geographic north and east. Apparent resistivity and phases were calculated from MT impedance components and plots of these values were used to analyse noise across the processed frequency range. If plots appeared to show noisy data, a different time window was chosen to process. Coherence values comparing the electric response with the magnetic response provided a further indication of the quality of the data. Most stations recorded good quality data.

RESPONSES, STRIKE AND DIMENSIONALITY

The pseudosection in Figure 5 was plotted using the processed data. The modes are quite smooth, with minor evidence of static shift effects, apparent in the vertical shift between stations of the apparent resistivities. The effect of this small shift would be minimal, however, so it was not corrected for prior to inversion (Figure 5).

The regional geological strike is north-west, as evident from TMI images and geological maps. Strike analysis returned geoelectric strike directions of 37° east of North. Prior to inversion, all data were rotated by an angle of 45° east of North.

Dimensionality of the resistivity structure is given by the shape of the ellipse: a circular phase tensor indicates 1D resistivity structure and an elliptical phase tensor could indicate either 2D or 3D resistivity structure. Analysis of the skew and ellipticity for each station, at every period, was undertaken to find 3D effects. Periods where the ellipticity was greater than 0.1 and the skew was outside the range of
Figure 5:

a) The apparent resistivity (Ωm) for the TE mode. Small vertical shifts station-to-station are visible.

b) The phase (°) of the TE mode. The phase is unaffected by static shift and distortion.

c) The apparent resistivity for the TM mode; note that it is much smoother than the TE mode, and static shift effects are less.

d) The phase of the TE mode. There is some evidence of static shift effects, -5° to 5° were said to be 3D. Very few data points had 3D effects and the strike is reasonably invariant with depth; it was determined that 2D modelling of the data were appropriate.
Figure 6: The angle of geoelectric strike plotted against the sampling period, for every station. At short periods, a trend is not obvious, however at longer periods a constant strike of about 40° east of North is evident, indicated by the aqua dashed line.

PHASE TENSOR ANALYSIS

In Figure 7, the minimum phase angle is less than 45° for periods up to 1.28 s, which corresponds to an increase in resistivity with depth. At slightly longer periods (greater than 1.28 s) the minimum phase angle is greater than 45° indicating that resistivity decreases with depth.

The orientation angle of the ellipses across the stations is approximately 45° east of North at longer periods, with little variance. At short periods, ellipses can be grouped into 4 broad regions of similar orientation and shape, which are outlined in black in Figure 7. The western-most group, group 1, has very thin ellipses, indicating possible current channeling with low resistivity in the direction of elongation. For group 2, at periods less than 1.28 s, ellipses are almost circular. This is indicative of a 1D environment which is typical for sediments. Group 3 has very uniform ellipses that have an angle of orientation of approximately 20° east of North. This group corresponds to a very resistive section in the MT inversion (Figure 9. Group 4 is similar to the west-most group and corresponds to a low resistivity region, C3 (see
Figure 7: Elliptical representation of the phase tensors displayed in a pseudosection. The distance from station DB24A to DB38C is approximately 75 km. The direction of the ellipse axes align with the direction of maximum conductivity. The colour is a representation of the invariant minimum phase value relating to resistivity changes with depth (phase $>45^\circ$ implies resistivity decreases with depth, phase $<45^\circ$ implies resistivity increases with depth). The black boxes group trends of ellipses, and are numbered 1-4 in red.

Figure 9).

Figure 8 shows phase tensors at a frequency of 0.1 Hz on top of a total magnetic intensity (TMI) image. The northwest geological strike can be seen. The phase tensors are oriented approximately northeast, with the phase tensors to the east rotating slightly counter-clockwise.
Figure 8: This figure provides another way of viewing the phase tensor ellipses. Ellipses of phase tensors at 0.1 Hz are superimposed on a TMI image. Geological strike of approximately northwest is visible. Phase tensors are all quite similar, with the principal axis aligned approximately northeast.

2D INVERSION

Occam 2D Model with Seismic Interpretation

The data points for all stations of the SD01’ transect were inverted using the smooth 2D Occam inversion code (Constable et al. 1987; de Groot-Hedlin & Constable 1990). A smooth modelling technique was used to reduce overinterpretation, reflecting the actual resolving power of the MT method. Error floors are set to 10% and 20% for the apparent resistivity and phase, respectively (Rodi & Mackie 2001; Siripunvaraporn 2012).

The Occam 2D smooth inversion in Figure 9 shows crustal heterogeneity over small distances, with four orders of magnitude variation in resistivity. The most striking feature is C1, the large (approximately 30 km wide) crustal structure at a depth of 10-40 km, of very low resistivity (<10 Ωm) immediately adjacent to the high resistivity (>10 000 Ωm) region (R1) to the west. Two low resistivity vertical features, C2 and C3, extend up to the surface and appear to be approximately 5 km in width. A combination of the smooth modelling technique and the diffusive
Figure 9: A smooth Occam 2D inversion of broadband (0.0064 s-81.92 s) MT data collected in 2009 and 2012. The data has been rotated 45 ° east of North to align the TE mode with the geoelectric strike. The background resistivity was set to 100 Ωm. The root-mean squared (RMS) value is 4.9 with a roughness of 229. Red indicates highly conductive (i.e. low resistivity) regions and blue, resistive. The seismic interpretation from the 2009 Southern Delamerian AuScope line has been superimposed on the MT, with faults of significance labelled with an F and the Moho with an M. Regions of low or high resistivity anomalies have been labelled with a C or an R respectively for reference purposes. The location of the Glenelg Zone (GZ) and the Grampian-Stavely Zone (GSZ) have also been marked. Two vertical faults (interpreted from MT results) have been added to the interpretation. Note that the western-most of the vertical faults separates the GZ from the GSZ. The star indicates a region where the mantle was particularly close to the surface in the past, during a period of extension.

The nature of MT means that resolution at depth is poor and for this reason these low resistivity zones are very likely to have a smaller width and be much more concentrated along fractures. The mantle beneath the Delamerian Orogen has little variation in resistivity, with an average value of approximately 1000 Ωm.
Sensitivity analysis

To investigate whether conductive anomalies were real or perhaps artefacts of the inversion, the starting background resistivity was set to 10, 100, 1000 and 10 000 $\Omega$m in consecutive inversions. Each inversion imaged almost identical features and the upper mantle resistivity was invariant, indicating confidence in the ability of MT to resolve resistivities at mantle depths and the robustness of the resistivity structure. An inversion using only the TM mode (rather than the common simultaneous inversion of the TE and TM modes) resulted in the same resistivity anomalies being imaged. These sensitivity analysis tests indicate that the features of the model in Figure 9 are very robust and are unlikely to be artefacts of the inversion.

Occam 2D Response Curves

Figure 10 shows the apparent resistivities and phases for various stations along the transect, and how the inversion result fit the data. The target root-mean squared value (RMS) was set to 1 for each inversion. The closer the final RMS is to this target, the better the model fit the data. After the lowest RMS was reached (in this case 4.77), the inversion was repeated, this time with the target RMS set just above this lowest RMS (4.9 for this inversion). This caused Occam to generate a much smoother (i.e. minimal structure) and hence more geologically plausible model (de Groot-Hedlin & Constable 1990), with the roughness decreasing from 565 to 229. Note that in each station there is a general increase in apparent resistivity with period, implying that resistivity is increasing with depth. An exception to this is at a period of approximately 10 s where there is a dip in resistivity. It is likely this dip in resistivity is a result in loss of signal strength associated with the dead-band, but it is possible that this could be a real feature.
DISCUSSION

3D Forward Model

To account for the change in resistivity and velocity (Rawlinson & Fishwick 2011) between the Delamerian and Lachlan Orogens, changes in one or more of the following
must occur: temperature, composition or hydration. Large contrasts in temperature occur between the orogens, due to varying lithospheric thickness (Rawlinson & Fishwick 2011). A model of dry olivine electrical resistivity as a function of temperature indicates that for a change in electrical resistivity from $10^3$ to $10^2$ Ωm (occurring at depths of 80-110 km in the model), a temperature increase of approximately 200°C would be required (Constable 2006).

2D Inversion

COMPARISON OF MT WITH SEISMIC INTERPRETATION

Fluid pathways often form along faults and can be associated with low resistivity regions, either due to fluids present, or because of alteration in mineralogy due to fluid interaction in the past. A major northwest-southeast trending fault (F2) has been highlighted by a white dashed line in Figure 9. This fault is a possible pathway for fluids from the mantle, which would provide an explanation of the highly anomalous, low resistivity C1 region. Although the fault continues past the vertical black dashed line, the subsurface changes drastically from very conductive to very resistive. A possible explanation for this is the timing of geological events—either the fluid averted to the northeast-southwest fault at this point, or the fault extended beyond this point after the fluids had ceased their upward movement.

POSSIBLE CAUSES OF MID-LOWER CRUSTAL LOW RESISTIVITIES

A variety of mechanisms could be responsible for low resistivities observed in the mid-lower crust, as investigated by Haak & Hutton (1986), Jones (1987) and Jones (1992). The presence of carbon causes low resistivity either as: a product of CO$_2$ mantle degassing of carbonate phases, mobilised during low degrees of partial melting in extensional periods, biogenic in the form of graphite (downthrust during collision) (Glover 1996) or slab devolatilisation (Luque et al. 1998). Graphitisation is promoted by shear stress and therefore its presence may be indicative of shear
deformation (Ross & Bustin 1990). Free fluids is another common cause of low resistivity (Wei 2001), but the availability of fluids is very dependent on the region. For instance, a young and active tectonic region might have a ready supply of fluids, however in stable crust the presence of fluids is less likely and other causes of low resistivity may be favoured. The presence of minerals other than carbon in the Earth, that can cause observable low resistivities are magnetite, haematite, pyrite, pyrrhotite and other sulphides. In certain settings, sulphides are of particular importance and are in sufficient abundance and distribution to effect the electrical resistivities significantly (Chouteau et al. 1997; Livelybrooks 1996).

GEOLOGICAL IMPLICATIONS

The star on Figure 9 indicates where the upper mantle was close to the surface due to rifting before the Delamerian Orogeny compressed the region (Foden et al. 2006). It is possible that during this period of extension, when the crust was at its thinnest, fluids were able to travel up through the Proterozoic crust. However, the low resistivity region extends up into the oceanic material overlying the crust, which was thrust up after the extensional period, complicating the matter. An alternative explanation is that the fluids originated from the compression of the Delamerian Orogeny.

The MT transect is intersected by the Yarramyljup Fault at station SD33, the west-most of the two vertical faults. This fault is thought to be a late Delamerian structure (Morand et al. 2003). The dip of the fault is unknown, but the dominant foliation in the rocks either side of it is close to vertical, suggesting vertical dip for the fault (e.g. Gibson 1992). This fault forms the boundary between the Glenelg Zone to the east, and the Grampians-Stavely Zone to the west (Cayley & Taylor 1998). The conductive region, C2, coincides with the location of the fault. This further
suggests the cause of low resistivity to be fluids. The fault has been interpreted as vertical on Figure 9, however it could be slightly dipping to the east.

**Crustal and Mantle Heterogeneity**

Dennis et al. (2012) MT survey in the southeast Australia region also show a very heterogeneous crust, with low resistivity regions occurring at depths of about 10-70 km, extending slightly deeper than low resistivity regions in this study (see Figure 9). Crustal features as interpreted from the deep seismic section commonly are expressed in the form of low resistivity (e.g., the Yarramyljup Fault), but the sub-Moho lithosphere has no resistivity variations in the location of this major fault. Features that are evident in the crust are either confined to the crust, or do not have a significant enough signature in the upper mantle to resolve using MT. An exception to this is a possible continuation of the Moyston Fault, expressed at depth as a change in resistivity and velocity.
CONCLUSIONS

2D inversions resulted in a heterogeneous crust with lateral variations of the order of $10^4 \, \Omega \text{m}$ over tens of km’s. The lithospheric mantle changes over a much broader scale, order of magnitude decrease in resistivity from the Delamerian to the Lachlan Orogen than no variation at all, as indicated by the fit of induction arrows.

A 0.1 km/s variation in velocity across the apparent boundary between the Delamerian and Lachlan Orogens implies a temperature and/or composition change. A resistivity decrease from the Delamerian to the Lachlan Orogen is a result of increased temperature ($\sim 200^\circ \text{C}$) from west to east and a change in composition from continental to oceanic mantle. A change to a more hydrated mantle may also occur.

Despite the distinct crustal heterogeneity, the sub-Moho lithosphere contains almost no expression of the events that shaped the crust with regards to resistivity. The electrical resistivity and velocity change between the Delamerian and Lachlan Orogen appears to be a deeper expression of the crustal structure, the Moyston Fault, however resolution is not high enough to be sure if this is the case.

Anomalously low resistivity regions in the crust align well with fault structures from seismic interpretations interpreted as fluid pathways. There are low resistivities ($< 10 \, \Omega \text{m}$) at a depth of 10-40 km, with smaller paths of lower resistivity leading to the surface. It is possible that fluid movement from the mantle may have produced a carbon rich lower crust and very low resistivities, leaving the upper mantle in this region to be depleted of volatiles.
ACKNOWLEDGMENTS

I would like to thank David Taylor for his assistance in the field and for helpful discussions regarding the geology of Victoria; Jared Peacock for his assistance in the field, with data processing, and explaining MT; Sebastian Schnaidt for assistance in the field and reviewing my work; Goran Boren for preparing the field equipment; Dr Lars Krieger for technical assistance; my peer-reviewers Simon Carter and Paige Honour; the geophysics honours students, particularly Millicent Crowe, discussions as we learnt the MT method together; Sahereh Aivazpourporgou for providing her MT data for my usage and AuScope for the use of their equipment. I would also like to thank my supervisor Professor Graham Heinson for his wonderful guidance throughout the year and sharing his extensive knowledge of the theory of MT, and also my secondary supervisor, Dr Stephan Thiel, for always being willing to help with problems and providing a different perspective on my project.

REFERENCES


## APPENDIX A

**Figure 11:** Spreadsheet showing information regarding the date of deployment and location of MT stations. The length of electric dipoles at deployment is listed in the columns, length of Ex, and length of Ey. The UTM zone for all stations was 54H.

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<td>150000</td>
<td>004</td>
<td>051000</td>
<td>350</td>
</tr>
<tr>
<td>DB28D</td>
<td>100000</td>
<td>0.90, 9.98</td>
<td>167</td>
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<td>060000</td>
<td>004</td>
<td>051000</td>
<td>350</td>
</tr>
</tbody>
</table>

**Undecimated Data - 100 Hz**

**Decimated Data - 10 Hz**