

Resistivity structure of the Precambrian Grenville Province, Canada

by

Ademola Quadri Adetunji

A thesis submitted to the Faculty of Graduate Studies of
The University of Manitoba
in partial fulfilment of the requirements of the degree of

DOCTOR OF PHILOSOPHY

Department of Geological Sciences
University of Manitoba
Winnipeg

Copyright © 2014 by Ademola Quadri Adetunji

Abstract

As part of the southern Ontario POLARIS project, this thesis uses magnetotelluric methods to investigate the lithospheric architecture of the Proterozoic Grenville Province and its margin with the Archean Superior Province. The first multi-dimensional crustal and lithospheric resistivity images for this region are presented.

The resistivity structure of the Phanerozoic sedimentary rocks in the lower Great Lakes region was determined using 1-D methods. The responses are strongly affected by a 20-23 S conductive layer within the sedimentary rocks, interpreted to be associated with Upper Ordovician shale units. This layer excludes resolution of resistivity structure of underlying crust.

The resistivity structure of the Precambrian crust and lithosphere was determined using 2-D methods. Different strike azimuths were determined for the crust, the upper lithospheric mantle and the deeper mantle layer. The crustal resistivity model for a profile from 50°N79°W to 43°N76°W images resistive Laurentian margin rocks dipping southeast to the base of the crust, bounded by the Grenville Front and the Central Metasedimentary Belt Boundary Zone. In a 2-D model of the mantle lithosphere for the same profile, a conductor at 70-150 km depth, located along-strike from the Mesozoic Kirkland Lake and Cobalt kimberlite fields, is interpreted to be due to mantle re-fertilization. Results from multiple MT profiles indicate conductive ($<10 \text{ } \Omega\cdot\text{m}$) lithospheric mantle beneath the Central Metasedimentary Belt and show that the northwestern Grenville Province is characterized by large-scale, resistive lithosphere ($>10,000 \text{ } \Omega\cdot\text{m}$) extending for about 300 km beneath the Grenville Province and 800 km along strike. Lithospheric thickness is interpreted to be 280 km; local decreases in this

depth are attributed to refertilization of the lower mantle lithosphere by fluids associated with Cretaceous kimberlite magmatism.

Anisotropic 2-D modeling reveals minimal electrical anisotropy ($<10\%$) at mantle depths in contrast to the factor of 15 anisotropy determined in earlier 1-D studies. This result suggests that observed MT response anisotropy is caused by large-scale structures. Strike direction in the upper lithospheric mantle is interpreted to be related to the Archean fabric of the Superior craton and in the deeper, conductive, mantle it is interpreted to have been established in the Cretaceous.

Acknowledgements

I would like to thank my supervisor, Professor Ian James Ferguson, for his guidance, inspiration, advice and continuous encouragement throughout my studies. Despite his very tight schedule, he would always found time for patient and thorough explanations. I am confident that the education and experience that I have gained while working with Ian has prepared me well for a future career in Geophysics.

I am indebted to the POLARIS southern Ontario MT team (J. Wenham, M. Serzu, X. Ma, E. Gowan, H. Sealey, T. Pacha, B. Eade, J. McCutcheon, I. Asudeh, C. Andrews) for data collection and landowners in the region for site access. The inspiration for the current work was drawn from M. Mareschal's important earlier work on the Abitibi-Grenville LITHORPOBE MT data. I acknowledge the financial support provided by University of Manitoba and Clayton H. Riddell Faculty of Environment, Earth, and Natural Resources in 2012. This work was supported by the Natural Sciences and Engineering Research Council of Canada's funding for Ian J. Ferguson.

I wish to also thank the members of my advisory and examination committees, Professors A. Frederiksen, A. Camacho, A. Rajapakse and M. Unsworth for their insight on this research and helpful comments on my thesis. In addition, I benefited greatly from comments and discussions with Professor Alan Jones. I thank him for his help with running the anisotropic inversion, providing his codes and reviewing the manuscripts.

Similarly, I am certain that everyone in the Geological Sciences Department at University of Manitoba has helped me in some way during my studies, for which I am very grateful. Particularly, I would like to thank Andrew Frederiksen, Brenda Miller and Mulu Serzu who were always there for me. I also appreciate the technical assistance

given by Yasir Abdu, Kunle Ola, Joe McLeod, Wasiu Omotoso, Ganiyu Asuni and Abiola Mekanjuola.

My parents deserve special mention for their wonderful support and prayers. I am particularly indebted to my mother who raised me with love and fear of Allah. She shaped the fundamentals of my learning characters and showed me the joy of scholarly quest. I am also thankful to my siblings, numerous families and friends for their support, prayers, and for always being my cheering section.

Finally, I am very grateful to my lovely wife, Sekinat Omolola, for her support and sacrifices during this phase of our incredible journey. She has been an outstanding source of strength throughout my academic career. Thank you for being there always and never giving up on me. To my kids – Muhammad Adedoyin and Manar Adesewa – who have shown me endless unconditional love. Thank you for being my source of joy and for understanding why I could not be there sometimes. I will always love you guys.

Dedication

To Sekinat, Muhammad and Manar

Table of Contents

Abstract	ii
Acknowledgements	iv
Dedication	vi
Table of Contents	vii
List of Tables	xii
List of Figures	xiii
List of Copyrighted Material for which Permission was Obtained	xxviii
Chapter 1: Introduction	1
1.1 General Introduction	2
1.2 General Objectives	4
1.3 Specific Objectives	5
1.4 Research Significance	6
1.5 Structure of the Thesis	7
1.6 References Cited	9
Chapter 2: The Magnetotelluric Method	11
2.1 Introduction	12
2.2 Source of MT Signals	13
2.3 Sources of Enhanced Conductivity in the Earth	16
2.4 Background Equations of MT method	20
2.5 The MT Transfer Functions	23
2.5.1 Impedance Tensor	23
2.5.2 Induction Vector	24
2.6 Apparent resistivity, Phase and Skin Depth	25
2.7 Dimensionality	27
2.7.1 1-D Earth	27
2.7.2 2-D Earth	30
2.7.3 3-D Earth	33
2.7.4 Electrical Anisotropy	35
2.8 Data Acquisition and Processing	38
2.8.1 MT Data Acquisition	38

2.8.2	Data Processing.....	43
2.9	Data Analysis	46
2.9.1	Distortion	46
2.9.2	Groom Bailey Tensor Decomposition	50
2.9.3	MT Phase Tensor	53
2.10	Modeling and Inversion	56
2.10.1	Forward modeling.....	57
2.10.2	1-D Inversion	58
2.10.3	2-D Inversion	59
2.10.4	3-D Inversion	60
2.11	Applications of MT	61
2.12	References Cited	63
Chapter 3: Geology, Geophysics and Tectonics		75
3.1	Introduction	76
3.2	The Superior Province.....	76
3.2.1	Abitibi subprovince.....	78
3.2.2	Pontiac subprovince	79
3.3	The Southern Province	80
3.4	The Grenville Province	81
3.4.2	The Central Metasedimentary Belt	84
3.4.3	The Grenville Front Tectonic Zone	87
3.4.4	The Central Metasedimentary Belt Boundary Zone	88
3.5	The Phanerozoic Sedimentary Rocks.....	88
3.6	Ottawa-Bonnechere Graben	94
3.7	Geophysics	94
3.7.1	Electromagnetics.....	95
3.7.2	Seismic Reflection and Refraction.....	98
3.7.3	Teleseismic Studies.....	103
3.7.4	Heat Flow.....	107
3.7.5	Gravity and Magnetics.....	110
3.8	Tectonic Evolution of the Grenville Province.....	114

3.9	Reference Cited	121
Chapter 4: Magnetotelluric Data.....		130
4.1	Introduction	131
4.2	LITHOPROBE MT data	131
4.3	POLARIS MT data.....	133
4.3.1	Data Acquisition	133
4.3.2	Data Processing.....	134
4.4	References Cited	138
Chapter 5: Crustal and lithospheric scale structures of the Precambrian Superior-Grenville margin		140
5.1	Abstract	141
5.2	Introduction	142
5.3	Geological and Tectonic Setting	145
5.4	MT Data	150
5.5	Geoelectric dimensionality and strike analysis	151
5.6	Two-dimensional modeling and inversion	164
5.6.2	Lithospheric mantle resistivity model.....	170
5.7	Comparison of resistivity models with seismic results	173
5.7.1	Crustal model	173
5.7.2	Lithospheric mantle model	179
5.8	Interpretation and discussion of geoelectric strike	183
5.9	Interpretation of crustal resistivity structures.....	185
5.9.1	The Abitibi Subprovince.....	185
5.9.2	The Pontiac Subprovince	187
5.9.3	The Grenville Front.....	188
5.9.4	The Central Gneiss Belt	190
5.9.5	The Central Metasedimentary Belt Boundary Zone	191
5.9.6	The Central Metasedimentary Belt	192
5.9.7	The middle and lower crustal resistivity	193
5.10	Interpretation and discussion of lithospheric mantle resistivity structure.....	196
5.10.1	High-resistivity zone beneath the northwest Grenville Province	196

5.10.2	Deep, laterally-extensive, horizontal conductive anomaly	200
5.10.3	Deep, subvertical conductivity anomaly	203
5.11	Conclusions	206
5.12	Acknowledgments	208
5.13	References Cited	210
Chapter 6: Imaging the lithosphere of the Grenville Province: large-scale structures and electrical anisotropy		219
6.1	Summary	220
6.2	Introduction	222
6.2.1	Electrical anisotropy	223
6.2.2	Previous MT studies of the Grenville Front	227
6.2.3	Objectives of the present study	232
6.3	Tectonic Summary	234
6.4	Depth-Transformed MT Response Anisotropy	236
6.4.1	Phase tensor results	237
6.4.2	Groom-Bailey tensor decomposition results	240
6.4.3	Comparison of MT response anisotropy with other geophysical responses 245	
6.5	Induction Arrows	251
6.6	Determination of 2-D Inversion Azimuths	255
6.6.1	Methodology	255
6.6.2	Common strike azimuth for the upper lithospheric mantle	258
6.6.2.1	Profile 1	258
6.6.2.2	Profile 2	263
6.6.2.3	Profile 3	263
6.6.3	Common strike azimuth for the deeper mantle layer	264
6.7	Isotropic and Anisotropic Two-Dimensional Inversion	266
6.7.1	Preparation of data sets for inversion	266
6.7.2	Inspection of pseudosections for qualitative information	267
6.7.3	Isotropic inversions	271
6.7.4	Anisotropic Inversions	277

6.8	Discussion and Interpretation.....	284
6.8.1	Large scale resistivity structure	284
6.8.2	Geoelectric dimensionality and directionality	291
6.8.3	Resistivity Anisotropy	295
6.9	Conclusions	297
6.10	Acknowledgments.....	301
6.11	References Cited	302
Chapter 7: Magnetotelluric imaging of the lower Great Lakes region, Canada		311
7.1	Abstract	312
7.2	Introduction	313
7.3	Geologic and Tectonic Settings.....	316
7.4	Depth of Investigation.....	321
7.5	Dimensionality and Strike Analysis.....	323
7.5.1	MT Phase tensor	323
7.5.2	Tensor Decomposition Analysis and Geoelectric Strike	324
7.5.3	Pseudosections	328
7.6	Modeling and Inversion	328
7.6.1	Resistivity structure of Paleozoic sedimentary rocks	328
7.6.2	Resistivity structure of the Precambrian lithosphere	335
7.7	Interpretation and discussion.....	337
7.8	Conclusions	345
7.9	Acknowledgements	346
7.10	References Cited	347
Chapter 8: Summary and Future Work.....		353
8.1	Summary	354
8.2	Future work	362
8.3	References Cited	364
Appendix A.....		365
Appendix B		371
Appendix C		380

List of Tables

Table 4.1. Location of POLARIS Southern Ontario (PSO) MT Recordings sites 135

Table 6.1. Summary of the geoelectric strike azimuths. Method 1 is the azimuth derived from the mutli-frequency multi-site GB-MJ decomposition, Method 3 is the azimuth based on the minimum mean absolute error at each site and Method 4 is the azimuth based on the median absolute error at each site. The table also shows the azimuth adopted for 2-D inversions used to obtain the lithopsheric mantle resistivity structure. All azimuths are defined clockwise from true north..... 259

Table 6.2. Summary of parameters and results of isotropic and anisotropic inversion for Profile 1. Figure 18 shows the location of the parts of the models averaged. The results for the resistor are for a 45 to 149 km depth range and for the conductor they are for a 118 to 270 km depth range..... 280

List of Figures

Figure 1.1. The tectonic map of North America (modified from Hoffman, 1989) showing the initial deployments of POLARIS MT sites in Canada. The first four POLARIS deployments were in the Cordillera, Slave craton, Superior craton-Trans Hudson Orogen margin, and the Grenville Province. MCR=Mid-continent rift.	3
Figure 2.1. Ranges in electrical conductivity of some common Earth materials (modified from Palacky, 1988).	18
Figure 2.2. Representation of a 1-D resistivity model with conductivity varying with depth.	29
Figure 2.3. Schematic description of a 2-D model consisting of two quarter spaces with different resistivities. E_y is discontinuous across the boundary due to the conservation of current. For the 2-D case, the EM fields can be decoupled into the TE and TM modes. Redrawn and modified from Simpson and Bahr (2005).	30
Figure 2.4. Schematic description of a 3-D resistivity model. At sites above 3-D structures, all the impedance tensor components are independent of each other.	34
Figure 2.5. Schematic description of the basic parameters of anisotropy. An x -directed dyke is transformed into a more general position by successively applying the three elementary Cartesian rotations angles α_S , α_D and α_L (Taken from Pek and Santos, 2002).	36
Figure 2.6. A 10-minute recorded time series for MT site PSO 001 showing E_x , E_y , B_x , B_y , and B_z components. The site is located at Picton in Ontario.	39
Figure 2.7. Schematic description of a typical broadband MT site layout with pictures of some parts of the Phoenix Geophysics MTU5A field instrument. The magnetic field components are recorded by two horizontal N-S and E-W oriented coils and one vertical coil. The electric field components are recorded by four electrodes oriented N-S and E-W and buried typically 100m apart and a fifth ground electrode (located at the center). The GPS antenna provides the recording unit with the site geographic coordinates and a continuous time signal. All the components are connected to a Phoenix Geophysics MTU5A recording unit which is usually powered by car batteries. Pictures were taken during POLARIS MT data acquisition.	41
Figure 2.8. Description of regional 3-D structure overlain by near-surface heterogeneities. For long-period signals, the regional structure falls into the inductive scale length whereas the near surface small-scale heterogeneities cause galvanic distortion.	48
Figure 2.9. Schematic illustration of twist, shear and anisotropy. A group of unit vectors (left) with two reference vectors (shown in blue and red) which have twist, shear and anisotropy applied to them (right). The black dashed lines indicate the original position and length of the now deformed reference vectors (Redrawn from Groom and Bailey, 1989).	52

Figure 2.10. Graphical illustration of the phase tensor. Coordinate invariants $\Phi_{\max}$ and $\Phi_{\min}$ define the magnitudes of the major and minor axes. The skew angle β measures the tensor's asymmetry and the angle α indicates the tensor's dependence on the coordinate system (x, y). The dashed line is the axis of symmetry and the direction of the major axis $\Phi_{\max}$ is defined by $\alpha - \beta$ (Redrawn from Caldwell et al., 2004).	55
Figure 3.1: The map of the exposed Superior Province and its subprovinces (redrawn from Card and Poulsen, 1998). Shapefile obtained from Geological Survey of Canada. 77	77
Figure 3.2. Subdivisions the Grenville Provinces by (a) Wyne-Edwards (1972) and Easton (1992) and (b) Carr et al., 2000. . GF=Grenville Front, GFTZ=Grenville Front Tectonic Zone (modified from Carr et al., 2000).....	83
Figure 3.3. Geological map of the lower Great Lakes region showing the Michigan Basin, the Allegheny Basins, Algonquin Arch, GFTZ and the CMBBZ (redrawn from Mazurek, 2004). Shapefile obtained from Geological Survey of Canada	91
Figure 3.4. The depositional cycle of Algonquin Arch showing the sedimentary rock formations and groups present within the lower Great lakes region of southern Ontario. The same colour is used to represent different formations that were deposited during the same period. Shapefile obtained from Geological Survey of Canada.	93
Figure 3.5. The effective phase pseudosection (top) and 1-D inversion model (bottom) of the western Quebec MT study (Taken from Kellett et al., 1994).	97
Figure 3.6. MT response anisotropy in the southern subprovinces of the Superior Province (Wawa, Abitibi, and Pontiac subprovinces) and the northern Grenville Province (Taken from Mareschal et al., 1995). DPDZ=Destor-Porcupine deformation zone; CLDZ=Cadillac-Larder deformation zone; GF=Grenville Front. The length of the line at each MT site is proportional to the phase difference between the off-principal MT impedances and the direction is the direction of maximum phase. Sites with erratic apparent resistivity and phase responses as a function of azimuth are shown as dots....	100
Figure 3.7. Schematic cross section showing the interpretation of the crustal architecture of the southwest Grenville Province (After White et al., 2000).	101
Figure 3.8. Reflection seismic data interpretation (a) and its simplified cartoon (b) showing the major thrust boundary (CMBBZ) as a crustal scale-shear zone along which NW-directed movement occurred during the Grenville orogeny (Modified from Roy and Mereu, 2000).....	102
Figure 3.9. Migrated seismic section plotted with no vertical exaggeration for a crustal velocity of 6.0 km/s. CGB=Central Gneiss Belt, CMB=Central Metasedimentary Belt, CMBbtz=Central Metasedimentary Belt boundary thrust zone (Taken from O'Dowd et al., 2004).	103
Figure 3.10. Traveltime inversion model showing the west-northwest-striking low-velocity corridor crossing the teleseismic array at latitude 46°N (Taken from Rondenay et al., 2000).	105

Figure 3.11. The result of traveltime inversion presented by two vertical sections for P-wave (left) and S-wave (right). The arrows on the left panel indicate a linear NE-striking high-velocity anomaly, imaged as a steeply-dipping slab-like feature. The arrow on the right panel indicates a patchy discontinuous, NW-trending low-velocity anomaly, imaged as an inward-dipping ribbon-like feature. The color scale gives percent velocity anomaly with respect to IASP91. Regions of poor ray coverage are masked in black (Taken from Aktas and Eaton, 2006).....	106
Figure 3.12. Heat flow map of the Canadian Shield with white dots indicating heat flow sites. Archean provinces: Superior (Sup), Hearne and Rae; Proterozoic Provinces: Trans-Hudson Orogen (THO), Grenville (Gre); Paleozoic Province: Appalachians (App); Paleozoic Basins: Williston (Will), Hudson Bay (H.B.)(Taken from Mareschal and Jaupart, 2004).....	109
Figure 3.13. Map of heat flow in southeastern Canada with white dots indicating heat flow sites. The center of the Adirondacks is indicated as Ad while A–A’ and B–B’ are the two profiles examined by the authors in their study (Taken from Marescahl et al., 2000)	110
Figure 3.14. Long-wavelength (> 100 km) Bouguer gravity anomaly map of eastern Canada. Ad indicates the center of the Adirondacks while A-A’ and B-B’ show the profiles examined by the authors (Taken from Mareschal et al., 2000). KSz = Kapuskasing Structural Zone.....	112
Figure 3.15. (a) Cartoon showing the major thrust boundaries within the CMB (b) the simplified density distribution within the CMB and (c) depth section obtained from the inversion of the gravity data which shows that the CMB is denser than the CGB and that density variation occurs within the CMB itself (modified from Roy and Mereu, 2000).	113
Figure 3.16. Block diagram showing comparison between the total magnetic field (on top) and seismic section (front face) obtained from southern Ontario (Taken from O’Dowd et al., 2004).	114
Figure 3.17. Schematic plate tectonic model for the formation of the Grenville Province (After Carr et al., 2000).....	116
Figure 3.18. Schematic cross section showing the tectonic architecture of the Grenville Orogen in southern Ontario based on geophysical and geological data (modified from White et al., 2000 and Ludden and Hynes, 2000, Taken from Hynes and Rivers, 2010).	120
Figure 4.1. Map of MT data sites and profiles defined for this research. It shows the POLARIS (red pentagons) and LITHOPROBE Abitibi-Grenville-Transect (black triangles) MT sites. GF = Grenville Front, CMBBZ = Central Metasedimentary Belt Boundary Zone, CGB = Central Gneiss Belt, CMB = Central Metasedimentary Belt, FAB = Frontenac-Adirondack belt. Shapefile obtained from Geological Survey of Canada.	132
Figure 5.1. Initial deployments of POLARIS MT sites in Canada. The first four POLARIS deployments were in the Cordillera, Slave craton, Superior craton-Trans	

Hudson Orogen margin, and the Grenville Province. Tectonic background modified from Hoffman, (1989). MCR=Mid-continent rift.....	143
Figure 5.2. A simplified geotectonic map of the study area showing the location of the MT sites. Triangles represent the LITHOPROBE Abitibi-Grenville transect MT sites and polygons represent the POLARIS MT sites. Sub-units of the Superior and Grenville Provinces are labeled. GF=Grenville Front, GFTZ=Grenville Front Tectonic Zone, OBG=Ottawa Bonnechere Graben; CMBBZ=Central Metasedimentary Belt Boundary Zone.	146
Figure 5.3. The Niblett-Bostick penetration depth for the 40 sites on the profile. Each of the symbols represents the response at a particular period. Red color represents penetration depth for TE mode and blue color represents penetration depth for TM mode. GF=Grenville Front; CMBBZ=Central Metasedimentary Belt Boundary Zone.....	152
Figure 5.4. Distribution of the absolute value of the shear and twist angles for the crustal (a and c) and the lithospheric mantle (b and d) depths. The distribution of the average RMS misfits for the crustal (e) and lithospheric mantle (f) depths are also shown.....	155
Figure 5.5. Rose diagrams of the regional geoelectric strike for (a) crustal depth, (b) lithospheric depth, (c) asthenospheric depth and (d) the absolute plate motion for the HS3_Nuvel 1A hotspot model defined by Gripp and Gordon (2002) and for the GPS based model defined by Larson et al. (1997). The bin size in all cases is 15° and N denotes number of data points (sites) used for each plot.	157
Figure 5.6. The crustal (a), lithospheric mantle (b) and the asthenospheric mantle (c) strike azimuths obtained from the STRIKE program with 90° ambiguity correction are plotted on the regional magnetic map (data obtained from Geological Survey of Canada). The longer arrow, APM-1, shows the HS3-Nuvel-1A absolute plate motion direction defined by the hotspot based model of Gripp and Gordon (2002) and the arrow APM-2 shows the absolute plate motion of the GPS based model of Larson et al. (1997).....	158
Figure 5.7. The apparent resistivity and phase pseudosections of TE and TM responses for the study area. The data are the regional impedances derived from the GB decomposition for the crustal strike azimuth of N45°E. Each panel shows the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.	162
Figure 5.8. The apparent resistivity and phase pseudosections of TE and TM responses for the study area. The data are the regional impedances derived from the GB decomposition for the lithospheric mantle strike azimuth of N85°E. Each panel shows the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.	163
Figure 5.9. Comparison of the observed and model TE and TM apparent resistivity and phases as well as the tipper responses at six sites along the MT profile. The results are for the inversion of the lithospheric mantle data set using the strike azimuth of N85°E. The	

model response shown is obtained from TE+TM+Tz data and the RMS misfit for each site is shown above it. Sites ABI001, ABI002 and ABI004 are located in the Pontiac Subprovince, ABIM14 is located in the CGB, PSO009 is located on the CMBBZ and PSO001 is the southeasternmost site. 167

Figure 5.10. 2-D resistivity section of the crust obtained by joint inversion of TE, TM and tipper responses for the crustal data set (V.E = 5). The RMS values of the TE (triangles), TM (stars) and joint TE+TM+Tz (squares) inversions at individual sites are plotted above the inversion model. 169

Figure 5.11. Resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set (V.E = 1). The RMS values of the TE (triangles), TM (stars) and joint TE+TM+Tz (squares) inversions at individual sites are plotted above the inversion model. The crustal section (upper 48 km) of the model is not shown. The lower section of the model in which there is no data resolution (below the lithospheric mantle conductor C1) has been blanked. The 123 Ω .m contour is shown in order to better define the depth extent of the more resistive part of the mantle. 171

Figure 5.12. Map of the study area showing the relationship between this study and previous seismic and tomography studies. The dashed white line indicates the Lithoprobe Abitibi-Grenville transect wide angle seismic refraction profile. Locations of Lithoprobe reflections seismic lines 12 and 16 (Calvert and Ludden, 1999), line 16A (Benn et al., 1994), line 15 (Kellett et al., 1994), and lines 32, 33 and 71 (White et al., 2000) are also shown. The black lines illustrate the Rondenay et al. (2000), Chen and Li (2012) seismic tomography studies as well as the surface track of Great meteor hotspot track defined by Crough (1981). The locations of the Kirkland Lake and Cobalt kimberlite fields are also indicated. 176

Figure 5.13. Previous seismic reflection and refraction (wide-angle reflection) models of the study area. The seismic refraction wide-angle reflection model (top) is from White et al., (2000). The middle panel shows a composite seismic reflection image from Lithoprobe seismic reflection lines 12 and 16 (Calvert and Ludden, 1999), line 16A (Benn et al., 1994), line 15 (Kellett et al., 1994), and lines 32 and 33 (White et al., 2000) through the boundary zones. The bottom panel represents the crustal seismic interpretation from Ludden and Hynes (2000) and White et al. (2000). 177

Figure 5.14. The crustal resistivity model (V.E = 2.5) is overlain by seismic reflection profiles from Lithoprobe lines 12, 16, 16A, 15, 32, and 33. The Algonquin (in CGB), Sharbot and Frontenac (in CMB) terranes are not covered by the seismic studies. The black continuous lines show the interpretation of Kellett et al. (1994) for line 15, Benn et al. (1994) for line 16A, Calvert and Ludden (1999) for lines 12, and 16, as well as White et al. (2000) for lines 32 and 33. The dashed lines show the additions and modifications of the interpretation based on the resistivity data. 178

Figure 5.15. Comparison of seismic body-wave tomography results of Rondenay et al., (2000) with resistivity model. Left panel shows the location of the seismic tomography

profile and upper right panel shows the percentage P-wave slowness anomaly. Arrows define the low velocity zone. Lower right panel shows the lithospheric mantle resistivity model at the same horizontal and vertical scale as the seismic image..... 180

Figure 5.16. Seismic surface-wave tomography model of Chen and Li, (2012) for NW-SE profile through the present study area. See Figure 5.12 for profile location: the present study area extends over the distance interval 0 to 3. The image shows the absolute S-wave velocity. 181

Figure 5.17. Interpretation of features in lithospheric mantle resistivity model. The figure shows the 2-D lithospheric resistivity model ($V.E = 1$), interpretation of the resistive block in the shallow mantle lithosphere as being modified Archean lithosphere, interpretation of the conductive zone at 160 km depth as the LAB beneath the southeastern Superior province, and interpretation of a localized conductor as a lithospheric mantle scar refertilized in the Cretaceous by passage of the Great Meteor hotspot. The correct model (for the strike azimuth appropriate at crustal signal penetration) is shown above the lithospheric model. The arrows show the part of the model used to define the averaged resistivity profile shown in Figure 5.18. 199

Figure 5.18. Comparison of electrical resistivity depth profiles for the northern Grenville Province (GRV; this study), with the profiles for the Ghanzi-Chobe/Damara Block (DMB), Western Kimberley Block (KBW), and Eastern Kimberley Block (KBE; Muller et al., 2009). The average profile for the Grenville Province is computed for the area shown in Figure 5.17. The theoretical resistivity-depth profiles calculated by Muller et al. (2009) for mantle geotherms of different lithospheric thicknesses are also shown in black. The laboratory measurements of the electrical-conductivity against pressure and temperature for dry olivine and pyroxene are used to compute the profiles (Constable et al., 1992; Xu and Shankland, 1999; Xu et al., 2000). The points of inflection on these curves coincide with the intersection between the adiabat and the conductive mantle geotherm (Muller et al., 2009). 201

Figure 6.1. MT response anisotropy in the southern subprovinces of the Superior Province (Wawa, Abitibi, and Pontiac subprovinces) and the northern Grenville Province (Boerner et al., 2000). DPDZ=Destor-Porcupine deformation zone; CLDZ=Cadillac-Larder deformation zone; GF=Grenville Front. The length of the line at each MT site is proportional to the phase difference between the off-principal MT impedances and the direction is the direction of maximum phase. Sites with erratic apparent resistivity and phase responses as a function of azimuth are shown as dots..... 228

Figure 6.2. Crustal and lithospheric mantle resistivity model for northeast-southwest profile crossing the Grenville Front ($V.E = 1$) (Adetunji et al., 2014). The crustal model is based on a geoelectric strike of $N45^{\circ}E$ and the lithospheric model is based on a strike of $N85^{\circ}E$. The sites used in the study are from the NNW-SSE profile centred on $79^{\circ} W$ in Fig. 1. Labels show the interpretations of Adetunji et al. (2014) of lithospheric mantle

resistivity structures. GFTZ=Grenville Front Tectonic Zone, CMBBZ= Central Metasedimentary Belt Boundary Zone. LAB=Lithospheric-Asthenospheric Boundary.231

Figure 6.3. Map of the study area showing the locations of MT sites considered in the present study. The black triangles indicate MT sites from the Lithoprobe-Abitibi Grenville transect and the wine-coloured pentagons indicate POLARIS MT sites. Profile 1 forms part of the data analyzed by Adetunji et al. (2014). The tectonic elements include the Grenville Front (GF), Central Gneiss belt (CGB), Central Metasedimentary Belt (CMB), Central Metasedimentary Belt Boundary Zone (CMBBZ) and Frontenac-Adirondack Belt (FAB)..... 233

Figure 6.4. Phase tensor response at different Niblett Bostick depths. The phase tensors are plotted with the major axis having a uniform length on the plot and the minor axis proportional to the phase of the lower phase component of the impedance. Circular symbols correspond to 1-D responses and more elliptical symbols to 2-D or 3-D responses. The phase tensor symbols are coloured by the phase skew angle: blue colours correspond to 1-D and 2-D responses and yellow and red symbols to strongly 3-D responses. MT responses with large uncertainties are excluded..... 241

Figure 6.5. Results of GB strike determination for ~50 km thick depth ranges extending from 45 to 350 km. The arrows show the azimuth of the GB regional impedance with the highest phase. The length of the arrow is proportional to the average phase difference, between the maximum and minimum phase values over the depth band. The results are superimposed on the regional Bouguer gravity anomaly map (data from Geological Survey of Canada)..... 244

Figure 6.6. Comparison of MT response anisotropy with seismic anisotropy. The black arrows show the azimuth of the GB regional impedance with the highest phase determined for depth ranges of 45 to 150 km (upper lithospheric mantle) and >200 km (deeper mantle layer) scaled by the phase difference. The dashed purple line on the upper panel shows the southeast limit of the region with consistent conductive direction in the upper lithospheric mantle. The red lines show the SKS fast direction defined by Senechal et al. (1996), Eaton et al. (2004), Rondenay et al. (2000), Evans et al. (2006), and Frederiksen et al. (2006, 2007) scaled by the time split. The plots are superimposed on the regional Bouguer gravity anomaly map (data obtained from Geological Survey of Canada). The white arrow, APM, shows the GPS-based absolute plate motion direction defined by Larson et al. (1997). 248

Figure 6.7. The induction arrow responses at 40 s, 100 s, 320 s and 500 s, for sites along profiles 1 and 3. The real induction arrows (plotted using the Parkinson convention) are denoted by red arrows and the imaginary arrows are denoted by blue arrows. Note that responses with large uncertainties are excluded. 254

Figure 6.8. RMS misfit for single sites versus strike azimuth for a GB fit to the upper lithospheric depth band (45–150 km). Results are plotted in order of the position of each site on the profile but the inter-site distance is not preserved. The red dashed line is the

regional strike azimuth of the profile based on the GB-MJ multisite multifrequency analysis and the white dashed line is based on the minimum misfit of the two more robust misfit measures. The gap in Profile 1 indicates the absence of data for this depth range at that site.	260
Figure 6.9. RMS misfit for single sites versus strike azimuth for a GB fit to the deeper mantle depth band (>200 km). The red dashed line is the regional strike azimuth of the profile based on the GB-MJ multisite multifrequency analysis and the white dashed line is based on the minimum misfit of the two more robust misfit measures. The gaps in the results indicate the absence of data for this depth range at the corresponding sites.	261
Figure 6.10. Misfit versus azimuth plots for robust misfit estimates. The plots show, as a function of regional strike azimuth, the minimum of the mean (top) and median (bottom) of the absolute value of the misfit determined in GB decompositions at individual sites. The final results are based on only those sites with RMS misfit of <2.0.	262
Figure 6.11. The apparent resistivity and phase pseudosections of TE and TM responses for Profile 2. The observed data are the regional MT responses derived from the GB decomposition using the upper lithospheric mantle strike azimuth of N85°E. Panels shows the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.	269
Figure 6.12. The apparent resistivity and phase pseudosections of TE and TM responses for Profile 3. The observed data are the regional MT responses derived from the GB decomposition using the upper lithospheric mantle strike azimuth of N85°E. Panels show the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.	270
Figure 6.13. Resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The RMS misfits of the joint TE and TM (purple squares) inversions at individual sites are plotted above the inversion model. The crustal section (upper 48 km) of the model is not shown as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. This figure is reproduced with the inclusion contour lines in Appendix B.9	274
Figure 6.14. Resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set (V.E = 0.5). The RMS misfits of the joint TE and TM (purple squares) inversions at individual sites are plotted above the inversion model. The crustal section (upper 48 km) of the model is not shown as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. This figure is reproduced with a the inclusion of contour lines in Appendix B.15	276

Figure 6.15. Isotropic and anisotropic inversion (using $\tau_{iso} = 1$) models for Profile 1. The anisotropic inversion results are for the along-strike ρ_{xx} , cross-strike, ρ_{yy} , and vertical ρ_{zz} component of the resistivity. The crustal section (upper 48 km) of the model is not shown.	279
Figure 6.16. Resistivity profiles through the mantle resistor and mantle conductor for the isotropic and anisotropic models. Each panel compares the results from the isotropic model with the three components of resistivity from the anisotropic model: along-strike= ρ_{xx} , cross-strike= ρ_{yy} , and vertical= ρ_{zz} . The left panel is for $\tau_{iso} = 1$ and right panel is for $\tau_{iso} = 0.1$	280
Figure 6.17. The along strike component of the anisotropic inversion models for Profile 1 with different value of τ_{iso} . The crustal section (upper 48 km) of the model is not shown. The rectangles show the parts of the model used to calculate the results shown in Table 6.2.	282
Figure 6.18. Resistivity models of profiles 1 (middle), 2 (east) and 3 (west) projected onto the map of the study area. The models, with V.E=0.5 and depth extent of 350 km, are aligned with the MT data sites. GF=Grenville Front, CMBBZ= Central Metasedimentary Belt Boundary Zone. CGB=Central Gneiss Belt, CMB=Central Metasedimentary belt, FAB=Frontinac-Adirondack Belt. The crustal section (upper 48 km) of all the models is not shown.	285
Figure 6.19. Comparison of electrical resistivity at 180 km depth with shear wave velocity perturbation (Chen and Li, 2012). In order to produce the resistivity results the models for profiles 1 to 3 were reprojected back to a line through the corresponding sites. Resistivity values extracted from the models at a depth close to 180 km were interpolated using a natural neighbor algorithm and then the results were contoured. The resistivity results show good spatial correlation with the S-wave velocity model with the lowest resistivity values at 180 km depth corresponding to $\sim -1.5\%$ velocity perturbation and the highest resistivity values at this depth corresponding to $\sim +1.5\%$ velocity perturbation.	289
Figure 7.1. Geological map of the study area showing depositional cycles and the location of the POLARIS MT sites. Shapefile was obtained from Geological Survey of Canada.	315
Figure 7.2. The upper panel (a) shows the tectonic elements of the study area (modified from Mazurek, 2004). The depths shown are the basin isopachs. The lower panel (b) shows composite stratigraphic succession of the study area (modified from Russell and Gale, 1982) for cross-section AA' shown in the upper panel. The white unit at the top represents the Quaternary sediments.	319
Figure 7.3. Stratigraphic section of Sedimentary rocks showing all the formations (redrawn from Mazurek, 2004) and the locations of the POLARIS MT sites, Mareschal et al. (1991) MT sites and Gomez-Trevino and Edwards (1983) CSEM sites.	320

Figure 7.4. Niblett-Bostick penetration depth for all the sites in the study area. Red bars represent the XY mode while the blue bars represent the YX mode of the impedance response.....	322
Figure 7.5. Phase tensor of the sites at different mantle depths colored by skew angle. CMBBZ=Central Metasedimentary Belt Boundary Zone.	325
Figure 7.6. RMS misfit of the single site geoelectric strike analysis for lithospheric depth (45 – 150 km). The stepwise strike azimuth is plotted against the resulting RMS misfit for each site along the profile. The white dashed line is the regional strike azimuth of the profile based on multisite multifrequency analysis of the profile using the STRIKE program.	327
Figure 7.7. Pseudosections of the TE and TM responses of the 2-D model generated from the GB decomposed dataset compared to the observed responses. Each panel shows the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.....	329
Figure 7.8. 1-D models of the high frequency MT components of four representative sites. The upper panel shows the results of the 1-D MT inversion in which models were fitted to the invariant MT response which averages the two off-diagonal impedance components. The lower panel compares the 1-D models with the models produced at similar locations by Gomez-Trevino and Edwards (1983). The level of editing at sites PSO016 and PSO019 is required because of the high level of noise in the area.....	332
Figure 7.9. Resistivity model derived by joint inversion of TE and TM responses (V.E = 1). The RMS misfits of the joint TE and TM (purple squares) inversions at individual sites are plotted above the inversion model. The crustal section (upper 40 km) of the model is not shown.	338
Figure 7.10. Resistivity models of Profiles 1 (Adetunji et al., 2014a), profiles 2 and 3 (Adetunji et al., 2014b), and Profile 4 (this study) projected along the MT data profiles on the map of the study area. Each profile is truncated at 150 km depth with a V.E of 0.5. GF=Grenville Front, CMBBZ= Central Metasedimentary Belt Boundary Zone. CGB=Central Gneiss Belt, CMB=Central Metasedimentary belt. The crustal section (upper 48 km) of all the models is not shown.....	341
Figure 7.11. A horizontal slice, at 120 km depth, through the resistivity models of profiles 1 to 4. The resistivity values at this depth, interpolated with natural neighbor algorithm, were projected onto the MT data acquisition profile and contoured. The result shows lateral extension of Archean lithosphere throughout the study area. It also shows a lithospheric conductor that extends from profiles 1 to 3. CMBBZ = Central Metasedimentary Belt Boundary Zone	344
Figure A.1. L-curve for 2-D inversion inversion models of Profile 1 showing RMS misfit versus a range of tau (τ) values.	366
Figure A.2. 2-D resistivity section of the crust obtained by inversion of only TM responses for the crustal data set along Profile 1 (V.E = 5). The regularization parameter	

tau of 6 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion.	367
Figure A.3. 2-D resistivity section of the crust obtained by joint inversion of only TM responses for the crustal data set along Profile 1 (V.E = 5). The regularization parameter tau of 6 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion.	367
Figure A.4. 2-D resistivity model derived by joint inversion of TE and TM responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 3 was used to derive this model. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.	367
Figure A.5. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 1 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.....	368
Figure A.6. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 2 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.....	368
Figure A.7. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 3 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.....	369
Figure A.8. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 5 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.....	369

Figure A.9. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 8 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.....	370
Figure A.10. The pseudosection of the observed and calculated tipper magnitude for the crustal model along Profile 1.	370
Figure B.1. RMS misfit for single sites versus strike azimuth for a GB fit to the crustal depth band (1–38 km). Results are plotted in order of the position of each site on the profile but the inter-site distance is not preserved.	372
Figure B.2. L-curve for 2-D inversion inversion models of Profile 2 showing RMS misfit versus a range of tau (τ) values.	373
Figure B.3. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 1 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.	373
Figure B.4. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 2 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.	374
Figure B.5. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 4 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.	374
Figure B.6. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 5 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.	375
Figure B.7. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 7	

was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 375

Figure B.8. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 10 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 376

Figure B.9. Gridded resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set ($V.E = 0.5$). The model is the same as Figure 6.13 but it displays contour lines that help to better delineate resistivity variations in the model 376

Figure B.10. L-curve for 2-D inversion inversion models of Profile 3 showing RMS misfit versus a range of tau (τ) values..... 377

Figure B.11. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 1 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 377

Figure B.12. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 2 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 378

Figure B.13. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 3 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 378

Figure B.14. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 7 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 379

Figure B.15. Gridded resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set ($V.E = 0.5$). The model is the same as Figure 6.14 but it displays contour lines that help to better delineate resistivity variations in the model..... 379

Figure C.1. L-curve for 2-D inversion inversion models of Profile 4 showing RMS misfit versus a range of tau (τ) values. 381

Figure C.2. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 1 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 382

Figure C.3. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 3 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 382

Figure C.4. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 5 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 383

Figure C.5. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 7 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 383

Figure C.6. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 8 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 384

Figure C.7. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 10 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the

model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. 384

List of Copyrighted Material for which Permission was Obtained

Figure 2.5 After Pek and Santos, 2000	36
Figure 3.2 After Carr et al., 2000	83
Figure 3.5 After Kellett et al., 1994	97
Figure 3.6 After Marescahl et al., 1995	100
Figure 3.7 After White et al., 2000	101
Figure 3.8 After Roy and Mereu, 2000	102
Figure 3.9 After O'Dowd et al., 2004	103
Figure 3.10 After Rondenay et al., 2000	105
Figure 3.11 After Aktas and Eaton, 2006	106
Figure 3.12 After Marescahl and Jaupart, 2004	109
Figure 3.13 After Marescahl et al., 2000	110
Figure 3.14 After Marescahl et al., 2000	112
Figure 3.15 After Roy and Mereu, 2000	113
Figure 3.16 After O'Dowd et al., 2004	114
Figure 3.17 After Carr et al., 2000	116
Figure 3.18 After Hynes and Rivers, 2010	120
Figure 5.16 After Chen and Li, 2012.....	181
Figure 6.19 After Chen and Li, 2012.....	289

Chapter 1

Introduction

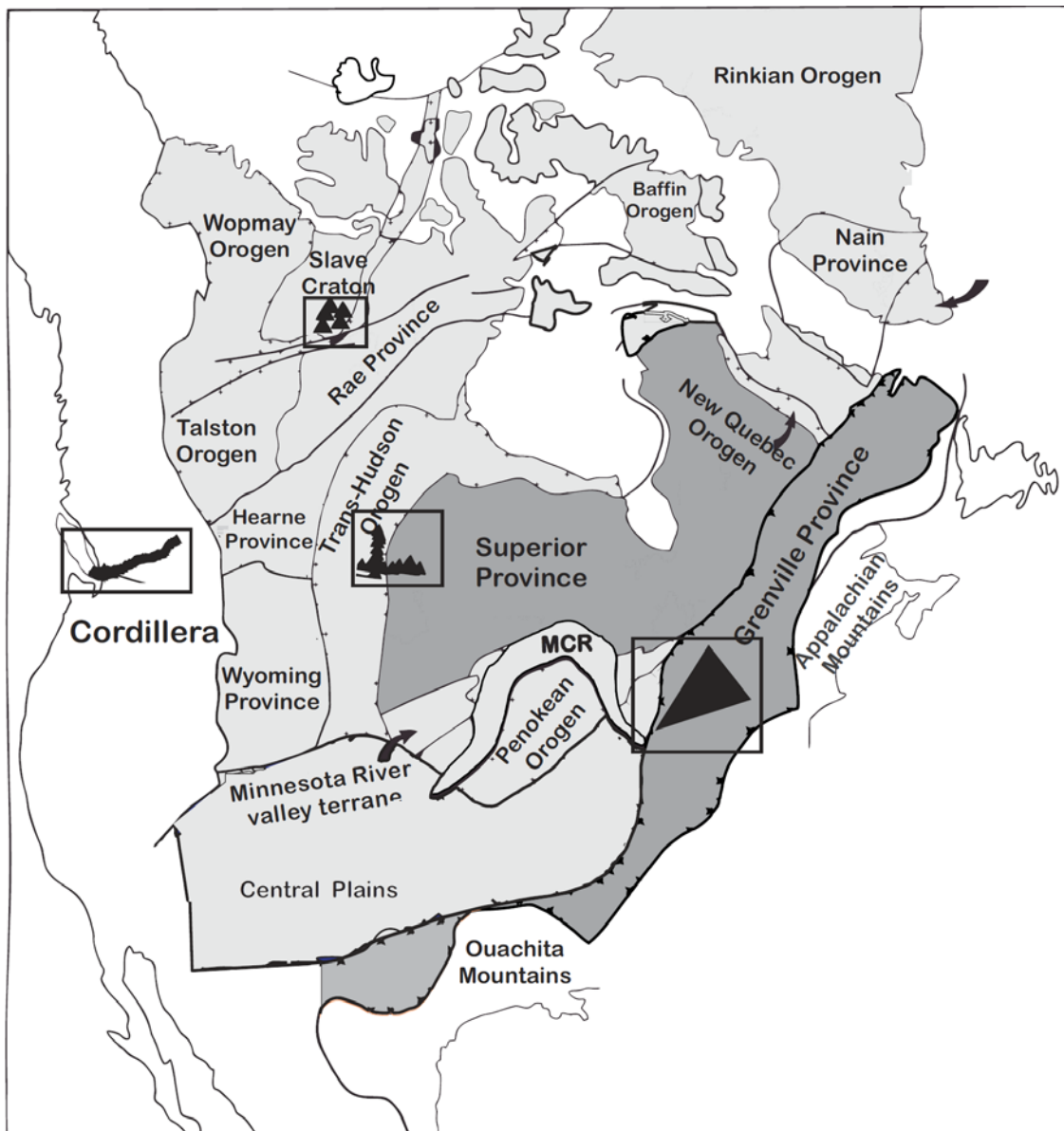
1.1 General Introduction

POLARIS (Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity) is a multi-institutional geophysical consortium focused on investigation of the structure and dynamics of the Earth's lithosphere, as well as assessment of earthquake, and geomagnetic hazards across Canada, using the magnetotelluric (MT) method and earthquake seismology (Eaton et al., 2005). As part of this project, seismic and MT data were collected between 2002 and 2005 in a number of study areas across Canada, including southern Ontario (Fig. 1.1). MT results from the POLARIS projects in British Columbia, Manitoba, and southern Ontario (geoelectric strike) have been presented by Soyer and Unsworth (2006), Gowan et al. (2009) and Frederiksen et al. (2006), respectively. Seismic results for the southern Ontario POLARIS project have been presented by Aktas and Eaton (2006), Frederiksen et al. (2006), Darbyshire et al. (2007), Chen and Li (2012), and Zhang and Frederiksen (2013). This research presents the MT results from southern Ontario.

The study area includes the Archean Superior and Proterozoic Grenville provinces (Fig. 1.1). The Superior Province, the largest Archean province on Earth, shares a tectonic boundary with the Grenville Province (Hoffman, 1989). The Superior Province contains small areas of pre-3.5 Ga crust but generally comprises rocks formed between 2.8 and 2.7 Ga (Hoffman, 1989). The Grenville Province, a product of the middle to Late Mesoproterozoic Grenville orogeny, extends across North America from Mexico to Labrador (Hoffman, 1989). The scale of this orogen is believed to be comparable with the Himalayas and formed between ca 1.2 – 1.0 Ga (Hanmer, 1988; Easton, 1992; Rivers, 1997; Carr et al., 2000; Rivers et al., 2012). The Grenville Province comprises reworked

polycyclic Proterozoic rocks derived from Laurentia (proto-North American landmass) as well as monocyclic rocks accreted just prior to the Grenvillian orogeny (Rivers et al., 1993; Carr et al., 2000; Rivers et al., 2012).

Figure 1.1. The tectonic map of North America (modified from Hoffman, 1989) showing the initial deployments of POLARIS MT sites in Canada. The first four POLARIS deployments were in the Cordillera, Slave craton, Superior craton-Trans Hudson Orogen margin, and the Grenville Province. MCR=Mid-continent rift.



Electromagnetic (EM) images of continental lithosphere help us understand the processes of formation, preservation and deformation of continents. MT is an EM method used to map the spatial variation of the Earth's conductivity by measuring naturally occurring electric and magnetic fields at the Earth's surface. The resolution of an MT survey is limited by the diffusive nature of EM propagation in the earth; it is usually on the order of hundreds of meters to kilometers. But the MT method can probe the Earth to depths of several hundreds of kilometers.

This research uses the MT method and data, integrated with other geophysical and geological information, to answer questions concerning the crustal and lithospheric structure as well as tectonic history of the Grenville Province. This is done by defining the regional conductivity structure of the crust and the lithospheric mantle in this area.

1.2 General Objectives

The POLARIS project focuses on exploring the Earth's crust and upper mantle in order to understand its evolution and dynamics. The research uses geophysical studies to understand the processes that have formed, preserved, and modified the crust and the lithosphere. The Grenville project, a sub-project of POLARIS, aims to produce a high resolution three-dimensional image of the lithospheric architecture to reveal tectonic evolution of the Grenville orogen.

The MT data will be used to image the regional resistivity structure of the Grenville Province lithosphere and the overlying Phanerozoic rocks. The results will improve the resolution of the lithospheric structures. The resistivity images produced by MT will also be used in conjunction with other geological and geophysical results to

produce geological and tectonic interpretations. In addition, the extent of lithospheric electrical resistivity anisotropy at the Grenville Front and other boundaries within the Grenville Province will be defined.

1.3 Specific Objectives

In order to achieve the stated general objectives, this research contributes as follows:

- Extraction of optimal quality MT responses from the recorded time series using robust spectral analysis.
- Geoelectric dimensionality, distortion and strike analysis.
- Definition of crustal and lithospheric mantle resistivity structure of the Grenville Province and its margin with the Superior Province along a main profile (extending work of Kellett et al., 1992).
- Definition of the resistivity structure of the lithosphere of the Grenville Province on three additional profiles.
- Definition of the resistivity structure of the sedimentary rocks of the Michigan and Allegheny basins overlying the Grenville Province.
- Interpretation of crustal and lithospheric mantle tectonic structures.
- Definition of electrical resistivity anisotropy at lithospheric depth south of the Grenville Front.
- Use MT results to establish the depth of the electrical anisotropy and define the sources of the anisotropy.
- Comparison of seismic fast directions and geoelectric strike azimuths within the Grenville Province.

- Comparison of electrical resistivity results with seismic reflection and refraction (White et al., 2000), and seismic tomography results (Rondenay et al., 2000; Aktas and Eaton, 2006; Faure et al., 2011; Chen and Li, 2012; Zhang and Frederiksen, 2013).

1.4 Research Significance

The research involves all aspects of MT work (except data acquisition) including time series processing, data editing and analysis, inversion, and geophysical as well as geological interpretation. The research focuses on areas further south of Grenville Front, within southern Ontario, that have no published record of MT studies. It draws its uniqueness from the following:

- Analysis of new MT data and re-analysis of older data using new methods.
- Comprehensive study of the resistivity structure of the Grenville Province in southern Ontario.
- Application of detailed depth-based dimensionality and strike analysis
- Integration of the resistivity results with seismic and geological information.

In addition, the conclusions from this study provide important results for geomagnetic hazard study and resource exploration in southern Ontario. Pipe-to-soil potentials (PSPs) show enhanced variations around electrically resistive zones (Fernberg et al., 2007). Also, during magnetic storms, geomagnetically induced currents (GICs) can be induced in technological networks such as railroads, power transmission lines, and pipelines (Lanzerotti and Gregori, 1986; Boteler et al., 1998). The Earth's resistivity

information produced by this work can be combined with other data, such as the geometry and location of technological networks, to model PSP and GIC responses to particular magnetic events, and define areas that are prone to hazards caused by PSPs and GICs. Such information can then be used to design hazard assessments and alleviation measures. The findings of this research are also beneficial to the mineral industry. The lithospheric thickness around the Archean craton margin is defined and the ability of MT to image such regions is useful for diamond exploration, as espoused by Jones and Craven (2004) and Jones et al. (2009).

1.5 Structure of the Thesis

This thesis is a comprehensive record of the research undertaken to define the resistivity structure of the Grenville Province and documents new tectonic interpretations of the area. The chapters are written as self-contained manuscripts, each with its own reference list. The thesis starts with a literature review and a description of the MT methodology in Chapter 2. This is followed by a review of the geology, geophysics and tectonic history of the study area in Chapter 3. Detailed information about the POLARIS and LITHOPROBE MT datasets used in this research is provided in Chapter 4. The dataset are divided into four profiles (see Fig. 4.1). Chapter 5 describes the crustal and lithospheric resistivity structure beneath the margin of the Archean Superior and Proterozoic Grenville provinces. This region was investigated with a 650-km-long northwest-southeast oriented profile (Profile 1) that contains 40 MT stations. This paper has been published in *Tectonophysics* (Adetunji et al., 2014a). The results derived from Profile 1 define an extremely resistive, southeast dipping slab of Archean lithosphere

extending southeast from the Superior-Grenville margin. It also defines a sub-vertical conductor that is along strike with the Kirkland Lake and Cobalt kimberlite fields. This feature is interpreted to represent re-fertilization of an older mantle scar. On the basis of these results, three northeast-southwest oriented profiles (1, 2, and 3) with 56 MT sites, with broader period range and extended area of coverage were used to define the extent of these structures into other areas within the Grenville Province. The results of this study are described in Chapter 6. This chapter also seeks to determine whether azimuthal variation in the MT responses, defined by earlier studies (e.g., Mareschal et al., 1995; Frederiksen et al., 2006), is related to anisotropy or to large-scale geological structure. This manuscript has been submitted to *Geophysical Journal International* for review (Adetunji et al., 2014b). The resistivity structure of the Phanerozoic sedimentary rocks as well as the lithosphere beneath the lower Great Lakes region is discussed in Chapter 7. The lithospheric results presented in this chapter complement those in Chapter 6. However, the work is presented separately because the MT data from this area have several distinct features. They were collected in a thick sedimentary basin where previous studies (Gomez-Trevino and Edwards, 1983; Marescahl et al., 1991) showed that the sedimentary rocks have a strong effect on the EM responses. The signal to noise ratio is also lower for the MT data in this region. The MT data available for this study (Profile 4) thus provide limited resolution of the crust and penetration into the deeper mantle is also limited. The results of this work have been discussed in a manuscript that has been submitted to *Geophysical Journal International* for review (Adetunji et al., 2014c). Finally, the conclusions and summary of this research are provided in Chapter 8. Additional models and L-curves are provided in Appendices A, B and C.

1.6 References Cited

- Adetunji, A.Q., Ferguson, I.J., Jones, A.G., 2014a. Crustal and lithospheric scale structures of the Precambrian Superior-Grenville margin. *Tectonophysics* 614, 146-169, <http://dx.doi.org/10.1016/j.tecto.2013.12.008>.
- Adetunji A.Q, Ferguson, I.J., Jones, A.G., 2014b. Imaging the mantle lithosphere of the Grenville Province: large-scale structures and electrical anisotropy. *Geophys. J. Int.* (submitted in Jan., 2014).
- Adetunji, A.Q., Ferguson, I. J., Jones, A.G., 2014c. Magnetotelluric imaging of the lower Great Lakes region. *Geophys. J. Int.* (submitted in Jan., 2014)
- Aktas, K., Eaton, D. W., 2006. Upper-mantle velocity structure of the lower Great Lakes region. *Tectonophysics* 420, 267–281.
- Boteler, D. H., Pirjola, R. J., Nevanlinna, H., 1998. The effects of geomagnetic disturbances on electrical systems at the earth's surface, *Adv. Space Res.* 22, 17–27.
- Chen, C. -W., Li, A., 2012. Shear wave structure in the Grenville Province beneath the lower Great Lakes region from Rayleigh wave tomography. *J. Geophys. Res.* 117, B01303.
- Eaton, D. W., Adams, J., Asudeh, I., Atkinson, G. M., Bostock, M. G., Cassidy, J. F., Ferguson, I. J., Samson, C., Snyder, D. B., Tiampo, K. F., Unsworth, M. J., 2005. Geophysical arrays to investigate lithosphere and earthquake hazards in Canada. *EOS Trans AGU* 86, 169-173.
- Faure S., Godey, S., Fallara, F., Trepanier, S., 2011. Seismic architecture of the Archean North American mantle and its relationship to diamondiferous kimberlite fields. *Econ. Geol.* 106, 223-240.
- Fernberg, P. A., Samson, C., Boteler, D. H., Trichtchenko, L., Larocca, P., 2007. Earth conductivity structures and their effects on geomagnetic induction in pipelines. *Annales Geophysicae* 25, 207 – 218.
- Frederiksen, A.W., Ferguson, I. J., Eaton, D., Miong, S. K., Gowan, E., 2006. Mantle fabric at multiple scales across an Archean-Proterozoic boundary, Grenville Front, Canada. *Phys. Earth Planet. Int.* 158, 240-263.
- Gomez-Trevino, E., Edwards, R. N., 1983. Electromagnetic soundings in the sedimentary basins of southern Ontario- A case history. *Geophysics* 48, 311-330.
- Hoffman, P. F., 1989. Precambrian geology and tectonic history of North America, in: Bally, A. W., Palmer, A. R. (Eds.), *The Geology of North America - An Overview*, The Geological Society of America, vol. A, pp. 447-512.
- Jones, A. G., Craven, J. A., 2004. Area selection for diamond exploration using deep-probing electromagnetic surveying. *Lithos* 77, 765-782.
- Jones, A. G., Evans, R. L., Muller, M. R., Hamilton, M. P., Miensoopust, M. P., Garcia, X., Cole, P., Ngwisanyi, T., Hutchins, D., Fourie, C. J. S., Jelsma, H., Evans, S., Aravanis, T., Pettit, W., Webb, S., Wasborg, J., The SAMTEX, 2009. Area selection for diamonds using magnetotellurics: Examples from southern Africa. *Lithos* 112, 83-92.
- Kellett, R.L., Mareschal, M., Kurtz, R. D., 1992. A model of lower crustal electrical anisotropy for the Pontiac Subprovince of the Canadian Shield. *Geophys. J. Int.* 111, 141-150.

- Lanzerotti, L. J., Gregori, G. P., 1986. Telluric currents: The natural environment and interactions with man-made systems, in: *The Earth's electrical environment*, National Academy Press, pp.232–257.
- Mareschal, M., Kurtz, R.D., Chouteau, M., Chakridi, R., 1991. A magnetotelluric survey on Manitoulin Island and Bruce Peninsula along GLIMPCE seismic line J: black shales mask the Grenville Front. *Geophys. J. Int.* 105, 173–183.
- Mareschal, M., Kellett, R. L., Kurtz, R. D., Ludden, J. N., Ji, S., Bailey, R. C., 1995. Archean cratonic roots mantle shear zones and deep electrical anisotropy. *Nature London* 375, 134-137.
- Rivers, T., Culshaw, N., Hynes, A., Indares, A., Jamieson, R., Martignole, J., 2012. The Grenville orogen- A Post-Lithoprobe Perspective in: Percival, J.A., Cook, F.A., Clowes, R. M. (Eds.), *Tectonic Styles in Canada: The Lithoprobe Perspective*, Geological Association of Canada, Special paper 49, pp 97-236.
- Rondenay, S., Bostock, M. G., Hearn, T. M., White, D. J., Ellis, R. M., 2000. Lithospheric assembly and modification of SE Canadian Shield: Abitibi-Grenville teleseismic experiment. *J. Geophys. Res.* 105, 13735-13754.
- White, D.J., Forsyth, D. A., Asudeh, I. A., Carr, S. D., Wu, H., Easton, R. M., Mereu, R. F., 2000. A seismic-based cross-section of the Grenville Orogen in Southern Ontario and western Quebec. *Can. J. Earth Sci.* 37, 183-192.
- Zhang, J., Frederiksen, A.W., 2013. 3-D crust and mantle structure in Southern Ontario, Canada via receiver function imaging. *Tectonophysics* 608, 700-712

Chapter 2

The Magnetotelluric Method

2.1 Introduction

The magnetotelluric (MT) method is a passive, frequency-domain electromagnetic (EM) method that uses temporal fluctuations of the Earth's EM field to map the spatial variation of the Earth's electrical resistivity. Natural geomagnetic field variations induce electric currents in the Earth and these subsurface currents generate secondary EM fields. The magnetic and electric signals that are detected at the Earth's surface originate from these fields and have periods of between less than 1 millisecond and several years. The strength of the electric field is dependent on the conductivity of the medium. By measuring the time variations of the magnetic and electric fields at the Earth's surface, the apparent resistivity, a volume-averaged resistivity over the penetration range of the signals, can be calculated as a function of period. Since signals with long periods (low frequency) penetrate deeper into the Earth, it is possible to determine the variation of resistivity with depth and distance (Tikhonov, 1950; Cagniard, 1953; Jones, 1999).

Tikhonov (1950) and Cagniard (1953) independently discovered this exploration method. Cagniard (1953), in the mathematical development of this method assumed a planar wave and this assumption was later shown to be valid for periods less than 10^5 s by Madden and Nelson (1964). Since its development, the MT method has undergone several improvements and is now considered to be a useful and reliable exploration method. These improvements were aided by the development of modern instrumentation, sophisticated recorders, low-noise sensors, improved modern robust processing methods, advanced analysis techniques, and multi-dimensional modeling and inversion (Chave and Jones, 1997; Jones, 2012).

Today, MT finds application in activities such as geothermal exploration and hydrocarbon exploration, particularly in sub-salt environments. The method can also be used to address problems related to hydrogeology and environmental geophysics as well as tectonic and lithospheric studies. The main advantages of MT method are its unique capability for exploration from very shallow depths (tens of meters) to very great depths (hundreds of kilometers) without artificial sources, with its negligible environmental impact, and its low cost.

2.2 Source of MT Signals

The source used by the MT method is the Earth's naturally fluctuating magnetic field. This time-varying magnetic field is generated from sources ranging from the core of the Earth to distant galaxies and signals are produced by a large number of different processes (McPherron, 2005; Viljanen, 2012). The geomagnetic fluctuations range between frequencies of 10^3 Hz and 10^{-5} Hz (or between periods of 10^{-3} s and 10^5 s) depending on their origin (Vozoff, 1991).

Worldwide thunderstorm activities are associated with lightning discharges, which radiate EM fields propagating to great distances and are the main source of MT signals above 1 Hz (short-period signal, <1 s). These signals are known as sferics and contain a broad range of EM frequencies. For the whole globe, the frequency of lightning flashes is estimated to be between 100 and 1000 per second (Vozoff, 1991). The most significant lightning discharges are the ones from the highly disturbed equatorial regions. Some large sferics propagate around the world in a waveguide mode between the Earth's surface and the ionosphere. Because the height of the ionosphere changes during the day

(about 60 km in daytime and 90 km at night); the waveguide dimension changes with time (Garcia and Jones, 2002a; Simpson and Bahr, 2005; Viljanen, 2012). At a particular site, the MT signals (sferics) depend on the strength and frequency of occurrence of lightning flashes, and on the distance from the point of lightning discharge. Man-made EM noise also constitutes a (usually undesirable) source of frequencies above 1 Hz. The most important noise originates from power lines and is generally avoided by choosing MT sites far away from them and by the use of notch filters incorporated in the equipment (Simpson and Bahr, 2005; Ferguson, 2012). Local lightning discharges may saturate the instrumentation amplifiers and will appear as spikes in the data (Miensoop, 2010).

The dead band is the band in which natural signals are typically weak. For the MT dead band, there is relatively low signal level in the transition between different sources of the MT field. It is a frequency range spanning approximately 0.5 – 5 Hz in which natural EM fluctuations are of low intensity compared with adjacent frequency ranges in the EM power spectrum. It occurs in the transition between the meteorological- and the solar-caused source processes. It is usually revealed in MT sounding curves by a reduction in data quality (Nichols et al., 1988; Simpson and Bahr, 2005). The frequency range of the audiomagnetotellurics (AMT) dead band is 1-5 kHz (Garcia and Jones, 2002a).

The MT fields at frequencies below 1 Hz (longer-period signals, >1 s) that penetrate deeper (100 km or more) into the Earth are caused by the complex physical interaction of solar plasma with the Earth's magnetosphere. These signals are generally known as micropulsations (Vozoff, 1991; Simpson and Bahr, 2005). Micropulsations

arise from EM energy that is generated by large current loops in the ionosphere and are transmitted to the Earth's surface through the atmosphere. These current loops are considered to have their origin in hydromagnetic or Alfvén waves that result from the interaction between the solar wind and the Earth's main geomagnetic field. Micropulsation amplitudes range from a fraction of a nanotesla to a few hundred nanoteslas, which indicates that their generation probably involves two or more different physical processes (McPherron, 2005; Kaufman and Keller, 1981; Viljanen, 2012). Micropulsations can be divided into regular pulsations called Pc (pulsation continuous) and irregular pulsations called Pi (pulsation irregular), with periods that vary between 0.2 and 600 seconds. The nature of the micropulsations observed at a particular time and site depends on the time of the day, the state of the local ionosphere, the level of geomagnetic activity (which is essentially controlled by solar activity), the time in the solar cycle, and seasonal effects (McPherron, 2005; Rokityansky, 1982). On records of the variations of magnetic and electric field components, regular pulsations are the most prominent and irregular pulsations generally appear as damped trains of waves that follow each other in an irregular sequence. Rokityansky (1982) described a third type referred to as "pearl" pulsations (PP), which occur at sunrise and sunset, and are amplitude-modulated with periods of 20 to 30 seconds.

The plane wave assumption can be considered valid (e.g. Cagniard, 1953; Madden and Nelson, 1964), for the purpose of MT analysis, when the distance to the source is much larger than the depth of interest and when the horizontal wavelength of the source is much larger than the horizontal scale of the area under study. In addition, depending on the distance from the equatorial and auroral electrojets, the uniform source

assumption is not valid at high and low latitude regions while it is most valid at mid-latitude regions (Miensoopust, 2010; Viljanen, 2012).

2.3 Sources of Enhanced Conductivity in the Earth

MT is the most viable geophysical method for the determination of the deep electrical properties of the Earth. In MT prospecting, the physical parameter of interest is the electrical resistivity or conductivity of the geological formations (Palacky, 1988; Jones, 1999; Evans, 2012). The electrical conductivity measures the ability of a material to conduct an electric current; different rock types have conductivities which vary over several orders of magnitude (Fig. 2.1). The geophysical importance of the resistivity results from its relationship with other physical, compositional, and structural parameters. These parameters depend on other factors such as temperature, pressure, porosity, fluid content, amount of melt, interconnectivity of the conductive minerals etc. Most rock matrices are electrical insulators, and therefore the bulk electrical resistivity of a rock is sensitive to very small changes in these factors (Palacky, 1988).

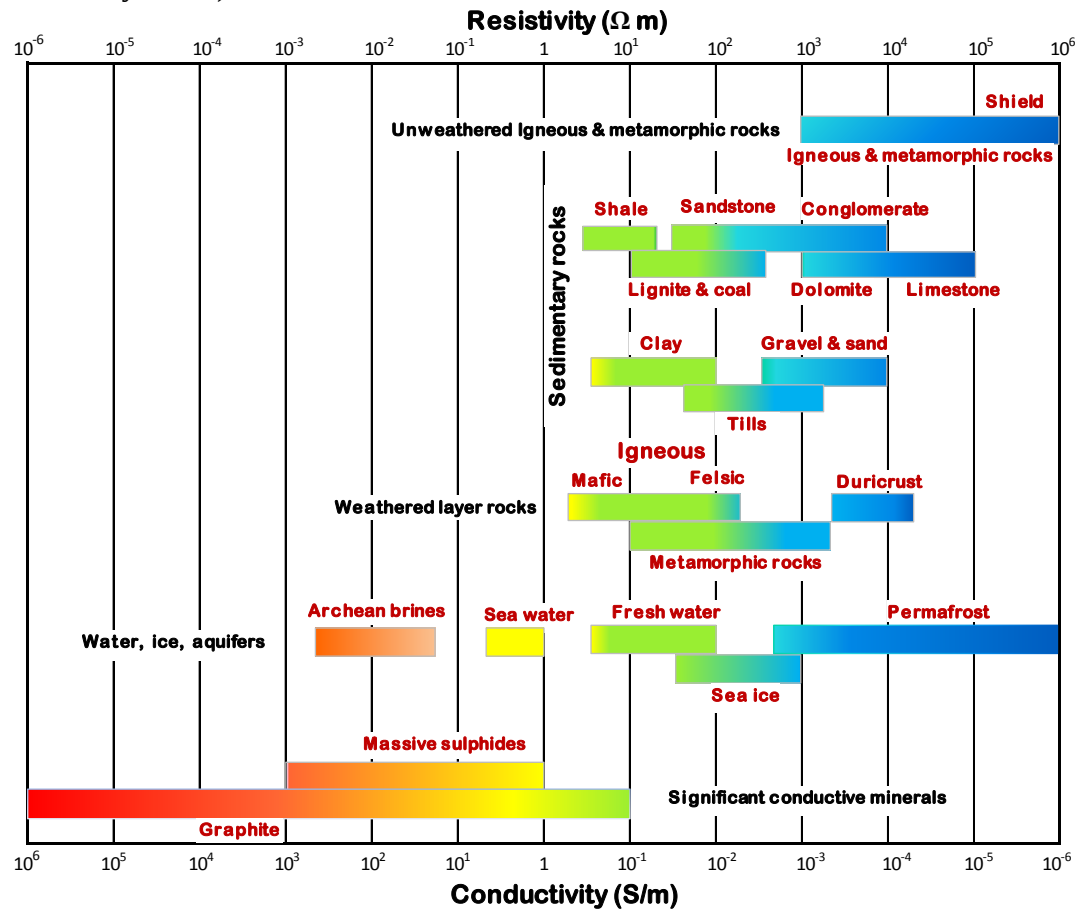
Ionic or electrolytic conduction and metallic or electronic conduction are the two main mechanisms of conduction in rocks. With ionic conduction, ions move in the fluids existing in the pore spaces of the rocks. With metallic conduction, the charge carriers are the electrons, occurring in metallic mineral deposits or graphitic rocks. Ionic conduction through pore fluids is the most important mechanism in most near-surface and crustal rocks (Telford et al., 1990; Evans, 2012). The conductivity of the middle crust is controlled by both electrolytic and electronic conduction while the lower crust is

controlled by electronic mineral conduction. The electrical conductivity of the upper mantle is controlled by semi-conductor conduction (Glover and Vine, 1995).

Fluids, graphite, interconnected grains of metallic oxides and sulphide are sources of the enhanced conductivity of mid-lower crust (e.g Jones and Craven, 1990; Katsube and Mareschal, 1993; Korja et al., 1996; ELEKTG Group, 1997; Li et al., 2003). MT has revealed that the middle and lower crust are zones of high electrical conductivity and provided evidence that the crustal structure is layered and varies laterally with depth (Haak and Hutton, 1986; Campbell, 1987). The mid to lower continental crust is generally much lower in resistivity than the upper crust; the decrease in resistivity is roughly coincident with the brittle-ductile transition. Some researchers suggest this is due to interconnected, saline fluids in the deeper (Marquis and Hyndman, 1992). Others argue that this is petrologically impossible, and that carbon films are responsible for the low resistivity (Yardley and Valley, 1997). The integrated conductance of the middle to lower crust in Phanerozoic areas has a mean value of order 200-2,000 S compared with values of 2-20 S observed in a number of Precambrian shield locations (Jones, 2013). The relatively low resistivity zones generally correlate with zones of high seismic reflection and relatively low seismic velocities (Hyndman and Shearer, 1989).

The enhanced electrical conductivity within the mantle has been attributed to the presence of graphite, hydrogen, partial melt, water and temperature variation (e.g., Karato, 1990; Mibe et al., 1998; Jones, 1999; Hirth et al., 2000; Ducea and Park, 2000; Constable, 2006; Korja, 2007; Yoshino et al., 2008, Muller et al., 2009; Evans, 2012; Selway, 2014). The continental mantle-lithosphere (upper mantle) is generally more resistive than lower lithosphere, but the resistivity is difficult to image. This is because it

Figure .2.1. Ranges in electrical conductivity of some common Earth materials (modified from Palacky, 1988).



is usually below a conductor (the lower crust) (Jones, 1999). The Moho is a seismically-defined discontinuity that cannot be imaged using MT (Jones and Ferguson, 2001; Jones, 2013). Most of the available information on the properties of the upper mantle is indirect and based on the variation of seismic velocities and resistivity with depth, combined with studies of mineral behavior at high temperature and pressures (Heinson and Constable, 1992; Jones et al., 2009; Fulla et al., 2011). Though we cannot usually know the actual resistivity of the continental lithospheric mantle, its conductance can however be estimated.

The asthenosphere represents the location in the mantle where the melting point (solidus) is approached and sometimes intersected. It represents a mechanical boundary between more rigid regions above and below (Heinson, 1999) and it allows mass flow associated with isostatic adjustment (Korja, 2007). It is typically seen at 70-200 km depth (Heinson, 1999) as a low resistivity zone, believed due to the presence of small amounts of interconnected partial melt which should be gravitationally stable below 200 km. Estimates of the resistivity of the asthenosphere are usually around 5 – 20 Ω .m (Jones, 1999). When asthenosphere is shallower than 200 km, the water dissolved in a depleted, anisotropic, olivine mantle enhances the conductivity by ionic conduction (Karato, 1990). A thin conductive asthenosphere may produce the same MT response as a thicker asthenosphere with lower conductivity (Heinson, 1999).

Generally, explanations for enhanced conductivity in a particular location may vary from graphite/sulphides/oxides to fluids, to partial melt, to elevated temperature; hence the interpretation must always be based on local geological and physical conditions (Schwarz, 1990; Korja, 2007).

Everett (2005) stated that in order for electrical anisotropy to occur, mineral phases must be at least partially-interconnected over distances of the order of the inductive scale length. Electrical anisotropy in the near-surface or upper crust can be caused by interconnected saline fluid-filled cracks, lithologic layering and geological structures like dykes and sills. Lower crustal anisotropy is caused by fluid or graphite bearing shear zones (Mareschal et al., 1995; Wannamaker, 2005). Schock et al. (1989), Mackwell and Kohlstedt (1990), Karato (1990), Ji et al. (1996), Hirth et al. (2000) and Wannamaker (2005) proposed that hydrogen diffusion in mantle minerals, such as olivine

or mineral with similar crystal alignment, can cause electrical anisotropy in the mantle. Olivine is electrically anisotropic and it is most abundant mineral in the upper mantle. Its crystals align preferentially when under strain and thus, many crystals with similar alignment would contribute to the overall anisotropic effect. Key (2012) suggested that electrical anisotropy in marine settings could be due to mineral alignment or interbedding of layers of different resistivities. The causes of electrical anisotropy differ slightly from one geological region to another and the anisotropy is usually the result of a combination of several factors that may only be valid for certain depths (Hamilton, 2006).

2.4 Background Equations of MT method

The propagation of EM fields in the Earth is governed by Maxwell's equations. These are four fundamental partial differential equations that describe the behavior of both the electrical and magnetic fields and their interactions. Maxwell's equations in their traditional form are:

$$\nabla \times \mathbf{E} = - \frac{\partial \mathbf{B}}{\partial t} \quad (\text{Faraday's Law}) \quad 2.1$$

$$\nabla \times \mathbf{H} = \mathbf{J} + \frac{\partial \mathbf{D}}{\partial t} \quad (\text{Ampere's Law}) \quad 2.2$$

$$\nabla \cdot \mathbf{B} = 0 \quad 2.3$$

$$\nabla \cdot \mathbf{D} = \rho_f \quad 2.4$$

where $\mathbf{E}$ is the electric field, $\mathbf{H}$ is the magnetic intensity, $\mathbf{B}$ is the magnetic flux density, $\mathbf{D}$ is the electric displacement, $\mathbf{J}$ is the conduction current density and ρ_f is the electric charge density. Maxwell's equations can be coupled using the frequency domain constitutive relations that define the conductive, dielectric, and magnetization responses of media to the imposed electric and magnetic fields. For a linear, isotropic, and

homogeneous material, the three constitutive relationships are defined as Ward and Hohmann (1988) are:

$$\mathbf{B} = \mu \mathbf{H} \quad 2.5$$

$$\mathbf{D} = \epsilon \mathbf{E} \quad 2.6$$

$$\mathbf{J} = \sigma \mathbf{E} \quad 2.7$$

where σ is the electrical conductivity, ϵ is the electric permittivity, and μ is the magnetic permeability of the medium.

Faraday's Law (equation 2.1) indicates that a time-varying magnetic field creates a perpendicular electric field. Ampère's Law (equation 2.2), indicates that both a displacement current and a free electric current will create a perpendicular magnetic field. The displacement current is related to a time-varying electric field. Equation 2.3 (Gauss' Law for magnetism) indicates that in a region containing no sources, the magnetic field is divergence free. This situation occurs because unlike the electric charge case, it is not possible to have isolated magnetic charges of opposite sign (Wangness, 1986). Gauss' Law (equation 2.4) indicates that in certain regions, there can be a non-zero net charges created, a situation that can arise because both positive and negative charges exist (Wangness, 1986). In any region of non-zero conductivity, the charge density in a homogenous region will reach an equilibrium value in an extremely short time ($t \approx \epsilon/\sigma$), so charge does not accumulate significantly during the flow of current (Grant and West, 1965).

Taking the curl of equation 2.1 and reversing the order of differentiation we have:

$$\nabla \times \nabla \times \mathbf{E} = - \frac{\partial}{\partial t} (\nabla \times \mathbf{B}) \quad 2.8$$

Using the constitutive relation for the magnetic field:

$$\nabla \times \nabla \times \mathbf{E} = - \frac{\partial}{\partial t} (\nabla \times \mu \mathbf{H}) \quad 2.9$$

Since the medium is uniform we can move the permeability to the left of the curl operator and substitute equation 2.2:

$$\nabla \times \nabla \times \mathbf{E} = - \mu \frac{\partial \mathbf{J}}{\partial t} - \mu \frac{\partial^2 \mathbf{D}}{\partial t^2} \quad 2.10$$

We next substitute the constitutive relations for the electric field to obtain:

$$\nabla \times \nabla \times \mathbf{E} = - \mu \sigma \frac{\partial \mathbf{E}}{\partial t} - \mu \varepsilon \frac{\partial^2 \mathbf{E}}{\partial t^2} \quad 2.11$$

Using the vector identity:

$$\nabla \times \nabla \times \mathbf{E} = \nabla (\nabla \cdot \mathbf{E}) - \nabla^2 \mathbf{E} \quad 2.12$$

(Stratton, 1941), using approximation $\nabla \cdot \mathbf{D} \sim 0$ (Grant and West, 1965), means that the first term on the right hand sides of the vector identity (2.12) is approximately zero. Thus, we have:

$$\nabla^2 \mathbf{E} - \mu \sigma \frac{\partial \mathbf{E}}{\partial t} - \mu \varepsilon \frac{\partial^2 \mathbf{E}}{\partial t^2} = \mathbf{0} \quad 2.13$$

Starting with the curl of Ampere's law (equation 2.2), we can obtain a similar equation for the magnetic field as:

$$\nabla^2 \mathbf{H} - \mu \sigma \frac{\partial \mathbf{H}}{\partial t} - \mu \varepsilon \frac{\partial^2 \mathbf{H}}{\partial t^2} = \mathbf{0} \quad 2.14$$

Assuming the natural electric and magnetic fields have harmonic time dependence ($e^{+i\omega t}$), where ω is the angular frequency, equations 2.13 and 2.14 can be expressed in the frequency domain as:

$$\nabla^2 \mathbf{E} - i\omega \mu \sigma \mathbf{E} + \omega^2 \mu \varepsilon \mathbf{E} = \mathbf{0} \quad 2.15$$

and

$$\nabla^2 \mathbf{H} - i\omega \mu \sigma \mathbf{H} + \omega^2 \mu \varepsilon \mathbf{H} = \mathbf{0} \quad 2.16$$

In the air, the EM signal travels as a wave but in the Earth it travels by diffusion and the Earth dissipates or absorbs the signals. At frequencies lower than 10^5 Hz, the displacement currents can be neglected in the Earth as they will be very small compared to the conduction currents. Therefore, equations 2.15 and 2.16 become:

$$\nabla^2 \mathbf{E} - i\omega\mu\sigma\mathbf{E} = \mathbf{0} \quad 2.17$$

and

$$\nabla^2 \mathbf{H} - i\omega\mu\sigma\mathbf{H} = \mathbf{0} \quad 2.18$$

Equations 2.17 and 2.18 are termed the Helmholtz equations (Ward and Hohmann, 1990) and they describe quasi-stationary EM diffusion in terms of electric and magnetic fields respectively.

2.5 The MT Transfer Functions

An MT transfer function is defined as a function relating the measured EM fields at given frequencies. This function depends on the electrical properties of the material but not on the nature of the source. MT transfer functions are the impedance tensors and the tipper vectors. In this thesis, I adopt a coordinate system with z vertical downwards, x to the north and y to the east for defining the responses.

2.5.1 Impedance Tensor

The MT impedance is the ratio of the orthogonal horizontal electric (E_x , E_y) and magnetic fields (H_x , H_y) and is a function of frequency. It contains information about the electrical resistivity of the subsurface (Simpson and Bahr, 2005) and it is a complex second rank tensor given as:

$$\mathbf{E} = \mathbf{ZH} \quad 2.19$$

or

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix} \quad 2.20$$

Since the impedance tensor $\mathbf{Z}$ is complex, each component of the matrix, Z_{ij} , contains real and imaginary parts, i.e, they have both magnitude and phase.

2.5.2 Induction Vector

The tipper function (or geomagnetic transfer function), describes the ratios of the vertical to horizontal magnetic fields by:

$$[H_z] = \begin{bmatrix} T_x & T_y \end{bmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix} \quad 2.21$$

and its amplitude is given by:

$$|\mathbf{T}| = \sqrt{|T_x|^2 + |T_y|^2}. \quad 2.22$$

It tips the horizontal magnetic fields into the vertical plane (Parkinson, 1959; Wiese, 1962).

The tipper magnitude is typically between 0.1 to 0.5 and rarely greater than 1 (Vozoff, 1991).

The tipper function is a complex frequency-dependent vector that is often graphically represented by the induction vectors (or induction arrows). The induction vectors are two real, dimensionless vectors that represent the real (equation 2.23) and imaginary (equation 2.24) parts of the tipper function on the xy plane:

$$\mathbf{T}_{Re} = (ReT_x, ReT_y) \quad 2.23$$

and

$$\mathbf{T}_{\text{Im}} = (\text{Im} \mathbf{T}_x, \text{Im} \mathbf{T}_y). \quad 2.24$$

The induction arrows are sensitive to lateral conductivity variations, so can provide insight on the dimensionality of the Earth (Jones and Price, 1970; Jones, 1986; Simpson and Bahr, 2005). The tipper responses, when plotted as induction vectors, point towards (in the Parkinson convention) and away from (in the Wiese convention) areas of anomalous current concentrations.

2.6 Apparent resistivity, Phase and Skin Depth

Apparent resistivity is an average resistivity for the volume of Earth sounded by a particular MT sounding period. For a homogeneous Earth, the apparent resistivity represents the actual resistivity, while for a multi-dimensional Earth; it represents the average resistivity of an equivalent uniform half space (Simpson and Bahr, 2005). Measuring the time variations of the EM fields allows for the calculation of the apparent resistivity as a function of period.

The apparent resistivity is generally related to a component of the MT impedance e.g. $\mathbf{Z}_{xy}$ by:

$$\rho_a = \frac{1}{\omega \mu} |\mathbf{Z}_{xy}|^2 \quad 2.25$$

where ρ_a is the apparent resistivity, and ω is the angular frequency. The apparent resistivity in equation 2.23 represents an average resistivity of the Earth from the surface to a depth related to the skin depth, which is a frequency-dependent measure of the distance that EM fields diffuse into the Earth.

The phase difference between the measured electric and magnetic field components is influenced by the variations in the subsurface resistivity. In a conductive medium, there will be a phase-shift between the electric and magnetic field components of the EM signals. This phase shift is generally given by:

$$\phi = \tan^{-1} \left(\frac{\text{Im } Z_{xy}}{\text{Re } Z_{xy}} \right) \quad 2.26$$

The impedance phase for a homogeneous Earth, a half-space of constant resistivity becomes 45° . It is independent of period and the electric field signals lead the magnetic field signals. In a layered earth, the impedance phase is sensitive to changes in resistivity with depth; it increases above 45° when the resistivity is decreasing with depth and it decreases below 45° when the resistivity is increasing with depth (Vozoff, 1991; Jones and Craven, 2004).

Skin depth (δ) is the depth at which the amplitude of the EM field decays to $1/e$ ($\sim 37\%$) of the surface value (Telford et al., 1990; Simpson and Bahr, 2005). It is given by:

$$\delta = \sqrt{\frac{2}{\omega\mu\sigma}} = \sqrt{\frac{T\rho}{\pi\mu}} \quad 2.27$$

Thus, penetration depends only on the period (T) of measurement, and resistivity (ρ) of the subsurface.

The magnitude of externally-sourced magnetic fields undergoes exponential attenuation as the field penetrates into a body. For material in which $\omega\mu\sigma$ is large, the skin depth is small and the field is attenuated close to the surface of the body. If we regard the external magnetic field as the primary field, the attenuation of the primary

field results from the cancellation caused by secondary fields associated with the currents induced in the medium. For a material with a small skin depth, the currents are confined to the region near the surface. For materials in which $\omega\mu\sigma$ is small; the skin depth is large, the primary field will extend to great depth in the body, and the secondary currents will also be distributed over a large depth but will be comparatively small (e.g., Grant and West, 1965).

The concept of skin depth can be extended in a general sense to three-dimensional bodies. Consider a conductor of typical dimension l . If $\omega\mu\sigma l^2 \gg 1$, the external fields will not penetrate to the middle of the body and the currents will be confined to the centre. If $\omega\mu\sigma l^2 \ll 1$, the conductor will have little secondary current created in it and the primary field will be almost unaltered by the body. Such a body will be difficult to detect inductively.

2.7 Dimensionality

The Earth resistivity model, which could be 1-D, 2-D, or a more complex 3-D, is constructed using the electrical properties derived from the MT impedance response (Vozoff, 1991). The MT transfer functions have a specific form depending on the spatial distribution of the conductivity (Vozoff, 1991; Simpson and Bahr, 2005).

2.7.1 1-D Earth

For a 1-D (uniformly layered Earth), the conductivity is independent of the x- and y-directions, it changes only with depth (Fig. 2.2). The diagonal components of the impedance tensor $\mathbf{Z}$, i.e Z_{xx} and Z_{yy} , which relate the parallel electric and magnetic

fields, are zero. The off-diagonal components, i.e Z_{xy} and Z_{yx} , are equal in magnitude, since there are no lateral variations in conductivity, but they have opposite sign in order to preserve the right hand rule. Therefore, for the case of 1-D Earth, the impedance is given as:

$$Z_{1D} = \begin{pmatrix} 0 & Z_{xy} \\ -Z_{yx} & 0 \end{pmatrix} \quad 2.28$$

The MT impedance for a layered Earth can be calculated using the Wait (1954) recursion formula given as:

$$Z_n = \frac{\omega\mu}{k_n} \left[\frac{k_n Z_{n+1} + \omega\mu \tanh(k_n h_n)}{\omega\mu + k_n Z_{n+1} \tanh(k_n h_n)} \right] \quad 2.29$$

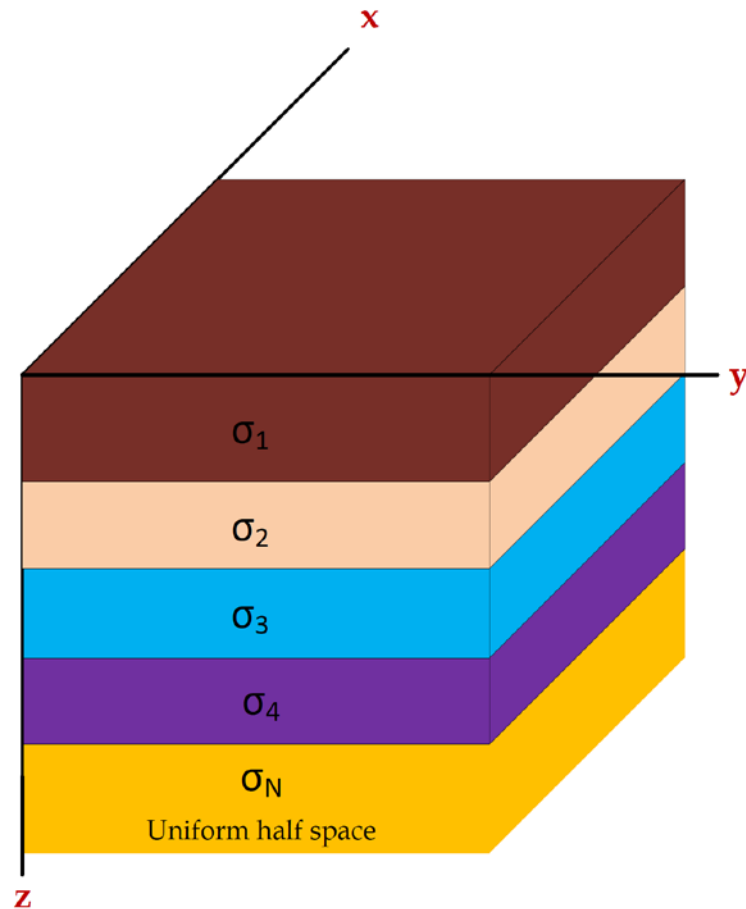
where h_n is the layer thickness and where:

$$k_n = (1-i) \sqrt{\frac{\omega\mu\sigma_n}{2}} \quad 2.30$$

This equation is solved recursively starting from the lowermost layer, which is a homogeneous half space.

In a two-layered Earth, when the period is sufficiently short, the skin depth in the top layer will be less than the thickness of the layer, so the apparent resistivity value will approach that of the top layer and the phase will be approximately 45° . At periods at which the skin depth is approximately equal to the thickness of the first layer, the phase will either be greater than or less than 45° depending on the resistivity of contrast. For periods at which the skin depth is greater than the thickness of the first layer, the resistivity should approach the resistivity of the lower layer while the phase will be approximately 45° (Simpson and Bahr, 2005).

Figure 2.2. Representation of a 1-D resistivity model with conductivity varying with depth

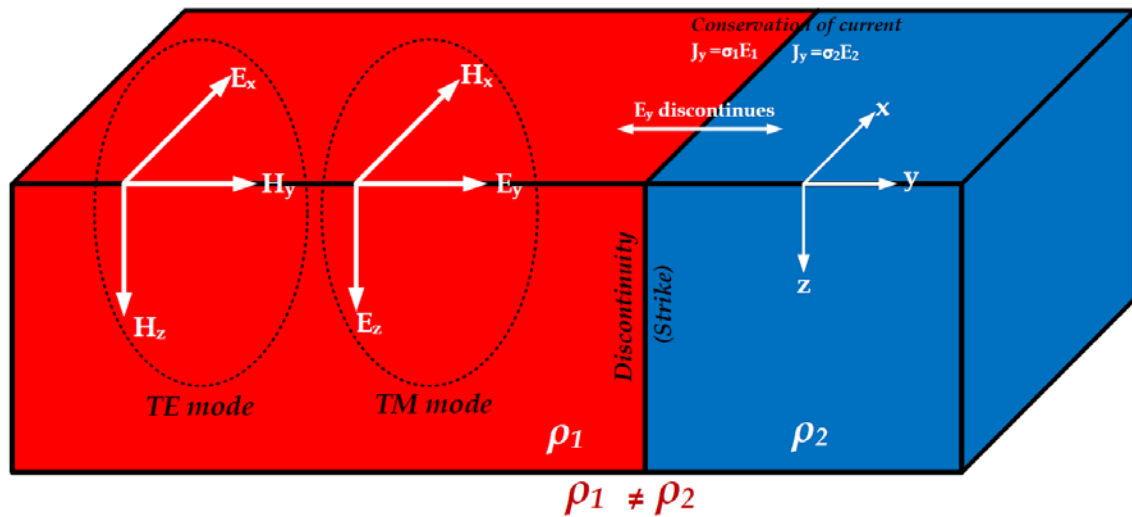


Also, in a 1-D resistivity model, it is almost impossible to resolve a conductive layer with conductance less than that of the overlying layer. MT can detect a conductive layer at depth but it cannot be used to detect relatively thin resistive layers. In addition, thin layers with the same integrated conductance but with different conductivity and thickness will have similar MT response (Simpson and Bahr, 2005).

2.7.2 2-D Earth

In a 2-D Earth, the conductivity varies along one horizontal direction as well as with depth, for example a fault or dyke (Fig. 2.3). The lateral direction (defined here as x) along which there is no field variation (constant conductivity) is referred to as the geoelectric strike or geoelectric strike direction (Simpson and Bahr, 2005).

Figure 2.3. Schematic description of a 2-D model consisting of two quarter spaces with different resistivities. E_y is discontinuous across the boundary due to the conservation of current. For the 2-D case, the EM fields can be decoupled into the TE and TM modes. Redrawn and modified from Simpson and Bahr (2005).



Whenever there is current flow perpendicular to a conductivity discontinuity, charges will build up at the boundary. This is a direct consequence of Ohm's law (equation 2.7) and the boundary condition which requires the electric current density, J_y , on both sides of the interface to be conserved (continuous). In addition, the electric field

(E_y) perpendicular to the the interface must be discontinuous and have different amplitude on either side of the boundary (Figure 2.3). All components of magnetic field, $\mathbf{H}$, and the tangential components of the electric field, E_x and E_z , are continuous across the boundary (Jones and Price, 1970).

For a 2-D structure, Maxwell's equations simplify, and can be decoupled into the two modes, with each mode relating three parts of the electric and magnetic fields. One mode, which describes the current flowing parallel to the geoelectric strike direction (i.e, in x -direction), is known as the Transverse Electric (TE) mode or E-polarisation (Simpson and Bahr, 2005), and is composed of E_x , B_y , and B_z . The equations relating these terms are:

$$\left. \begin{aligned} \frac{\partial E_x}{\partial y} &= \frac{\partial B_z}{\partial t} = i\omega B_z, \\ \frac{\partial E_x}{\partial z} &= \frac{\partial B_y}{\partial t} = -i\omega B_y, \\ \frac{\partial B_z}{\partial y} - \frac{\partial B_y}{\partial z} &= \mu_0 \sigma E_x, \end{aligned} \right\} \quad \text{TE mode} \quad 2.31$$

(Simpson and Bahr, 2005)

The response of this mode is continuous across the boundary, i.e, the estimates of the apparent resistivity (ρ_{xy}) and phase (ϕ_{xy}) are spatially continuous. This mode is capable of resolving a thin conductor but depth resolution is limited (Simpson and Bahr, 2005). This mode has associated vertical magnetic fields which define a non-zero tipper response and give enhanced sensitivity to conductive features in the subsurface.

The second mode is known as the Transverse Magnetic (TM) mode or B-polarisation, and is composed of B_x , E_y , and E_z . It describes the current flowing perpendicular to the geoelectric strike and the magnetic field parallel to the geoelectric

strike. The equations relating these terms are:

$$\left. \begin{aligned} \frac{\partial B_x}{\partial y} &= \mu_0 \sigma E_z \\ \frac{-\partial B_x}{\partial z} &= \mu_0 \sigma E_y, \\ \frac{\partial E_z}{\partial y} - \frac{\partial E_y}{\partial z} &= -i\omega B_x, \end{aligned} \right\} \text{TM mode} \quad 2.32$$

(Simpson and Bahr, 2005)

The MT response of the TM mode is discontinuous across the boundary, i.e, the apparent resistivity (ρ_{yx}) and phase (ϕ_{yx}) are spatially discontinuous. This mode is therefore more sensitive to lateral variations in conductivity than the TE mode.

For a 2-D structure, the measured impedance tensor is dependent on the azimuth of the electric and magnetic field measurements relative to the geoelectric strike. When the measurement direction of the impedance is aligned with the strike coordinates, the diagonal component of the impedance tensor will be zero because the electric components are only related to only the orthogonal magnetic components. The off-diagonal components (Z_{xy} and Z_{yx}) represent the TE and TM modes respectively. These MT impedances are normally not equal, are of opposite signs, and yield different values of apparent resistivity and phase for each mode.

$$Z_{2D} = \begin{pmatrix} 0 & Z_{xy}(\omega) \\ Z_{yx}(\omega) & 0 \end{pmatrix} = \begin{pmatrix} 0 & Z_{TE}(\omega) \\ Z_{TM}(\omega) & 0 \end{pmatrix} \quad 2.33$$

Generally, the geoelectric strike direction is unknown prior to MT survey, and thus, it rarely occurs that MT instruments are aligned with the strike. As a result, the measured impedance tensor will have four non-zero components and the modes will be mixed together in the measured tensors. In noise-free data the corresponding 2-D response, Z_{2D} ,

can be obtained by mathematically rotating the coordinate system in which the observed impedance tensor, $\mathbf{Z}_{\text{obs}}$, is defined through a defined geoelectric strike angle, θ , around a vertical axis using a Cartesian rotation matrix, $\mathbf{R}$, and its transpose, $\mathbf{R}^T$, until the diagonal components are zero:

$$\mathbf{R} = \begin{pmatrix} \cos\theta & \sin\theta \\ -\sin\theta & \cos\theta \end{pmatrix} \quad \text{and} \quad \mathbf{R}^T = \begin{pmatrix} \cos\theta & -\sin\theta \\ \sin\theta & \cos\theta \end{pmatrix}. \quad 2.34$$

The resulting $\mathbf{Z}_{2D}$ is given as:

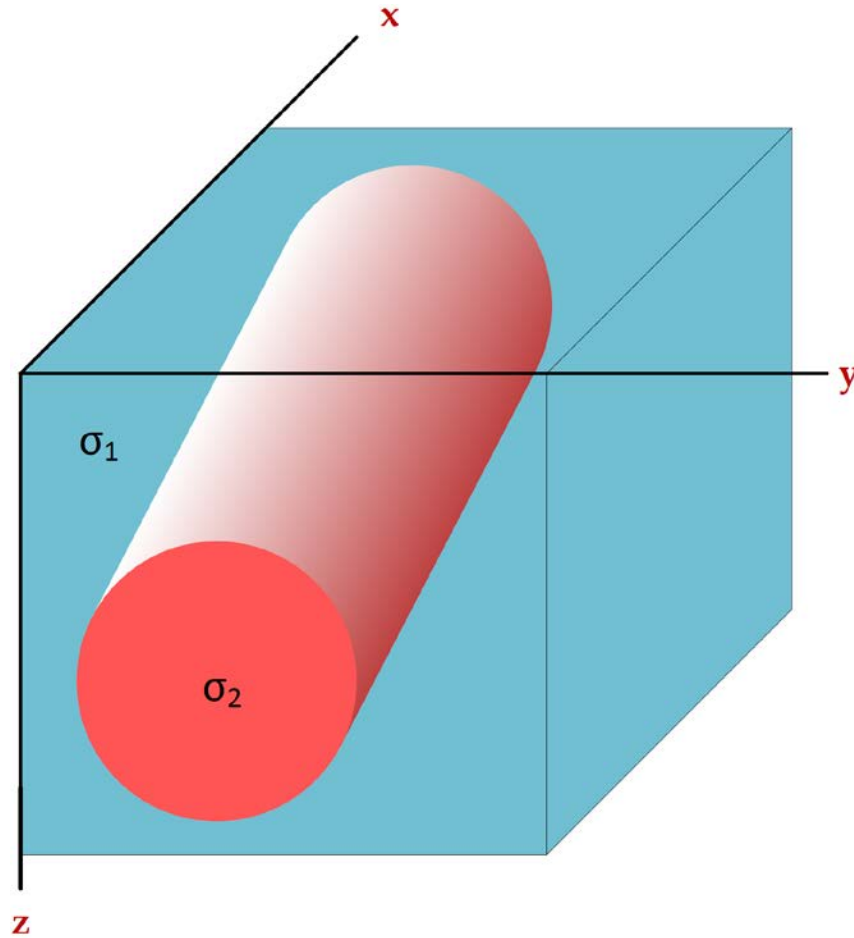
$$\mathbf{Z}_{2D} = \mathbf{R} \mathbf{Z}_{\text{obs}} \mathbf{R}^T \quad 2.35$$

There are two rotations that yield the off-diagonal form; one corresponds to the Z_{xy} term having electric field parallel to strike and the other corresponds to the Z_{yx} term, having electric field parallel to strike. There is thus a 90° ambiguity in the strike determination.

2.7.3 3-D Earth

In reality, measured MT data are never exactly 1-D or 2-D, and the diagonal elements of the impedance tensor are always non-zero. Deviation from ideal 2-D behavior occurs when the subsurface is 3-D (Fig. 2.4), or when the MT data are noisy or are distorted by near-surface structure. For a 3-D Earth, the conductivity distribution varies in all three directions and there is no single rotation angle that will zero the diagonal elements of the impedance tensor. Furthermore, the decoupling of the Maxwell's equations into two separate modes is no longer possible. Determination of the resistivity structure of a 3-D Earth may require the modeling of a full impedance tensor with four complex elements for each frequency (Simpson and Bahr, 2005).

Figure 2.4. Schematic description of a 3-D resistivity model. At sites above 3-D structures, all the impedance tensor components are independent of each other.



The dimensionality and distortion of an MT response in the presence of a burial heterogeneity depends on the length extent of the heterogeneity (L) and the skin depth (δ_h) of the host rock (Simpson and Bahr, 2005). When a 3-D body with conductivity σ_1 is buried in a homogeneous half space with conductivity σ_2 , a 1-D MT response will be produced if the frequencies are sufficiently high and the skin depth is small compared to the depth of the body. It will occur with a surface heterogeneity if the skin depth is small compared to the shortest dimension of the body. If the length of the body is much greater

than the skin depth ($L/\delta_h \gg 1$), for the frequency of interest, then a 2-D interpretation should give, approximately, the correct conductivity structure. If the length of the body is short (small L) or the skin depth is large (low frequency) such that $L/\delta_h \ll 1$, then a 3-D interpretation is necessary. For very low frequencies (long periods), where the skin depth is much larger than the dimensions of the body, the anomaly caused by the body becomes weak, but the frequency independent galvanic distortion remains.

2.7.4 Electrical Anisotropy

Sheriff (1999) defined geoelectric or electrical anisotropy as the variation of electrical property in a material based on the direction of measurement. It depends largely on the interconnectivity of mineral phases responsible for the anisotropy. Electric anisotropy may also occur at a range of scales. An anisotropic rock will have a resistivity structure in which the resistivity in the three-orthogonal directions is represented by a second rank, full 3-D tensor where the resistivity tensor is symmetric ($\rho_{ij} = \rho_{ji}$) and positive definite (Pek and Verner, 1997; Weidelt, 1999):

$$\boldsymbol{\rho}(x, y, z) = \begin{pmatrix} \rho_{xx} & \rho_{xy} & \rho_{xz} \\ \rho_{yx} & \rho_{yy} & \rho_{yz} \\ \rho_{zx} & \rho_{zy} & \rho_{zz} \end{pmatrix} \quad 2.36$$

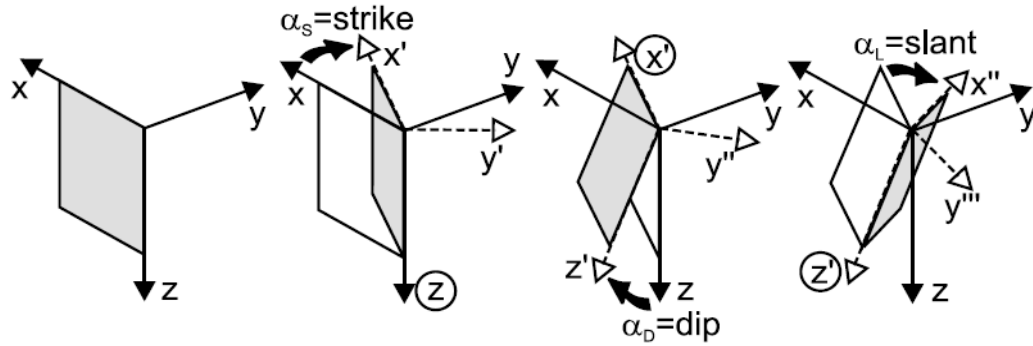
With these properties, the tensor can be diagonalized and expressed by the three principal (non-zero) resistivities, ρ_1, ρ_2, ρ_3 , and three rotation angles, which relate the orientation of the tensor's principal axes (x', y', z') to the reference frame (x, y, z) (Heise et al., 2006). Following the rotation of Pek and Santos (2006), the three rotation angles

used are Euler angles corresponding to the strike (α_S), dip (α_D), and slant (α_L) of the anisotropy. The rotation is:

$$\rho(x,y,z) = \underbrace{R_z^T(\alpha_S)R_x^T(\alpha_D)R_{z'}^T(\alpha_L)}_{\rho(x',y',z')} \begin{pmatrix} \rho_1 & 0 & 0 \\ 0 & \rho_2 & 0 \\ 0 & 0 & \rho_3 \end{pmatrix} R_{z'}(\alpha_L)R_x(\alpha_D)R_z(\alpha_S) \quad 2.37$$

where $\mathbf{R}$ is a 3-D Cartesian rotation matrix. The subscripts z and x specify the axes around which the rotation occurs. The Euler rotation angles α_S , α_D , and α_L are applied successively around the z -axis, x -axis and the most recent z -axis (z') respectively (Fig. 2.5; Pek and Verner, 1997; Pek and Santos, 2002, 2006).

Figure 2.5. Schematic description of the basic parameters of anisotropy. An x -directed dyke is transformed into a more general position by successively applying the three elementary Cartesian rotations angles α_S , α_D and α_L (Taken from Pek and Santos, 2002).



For layered resistivity structures, the presence of anisotropy can create responses in the phase and apparent resistivity resembling those associated with 2-D isotropic structures (e.g., Heise et al., 2006). The anisotropy creates a variation in apparent resistivity with maxima and minima in directions corresponding to the projection of the

anisotropy ellipsoid to the horizontal plane. For anisotropy occurring in a uniformly resistive medium there is no corresponding azimuthal phase variation. Heise et al. (2006) show that such phase variation depends on the resistivity gradient with depth, so an azimuthal phase variation will occur only when the anisotropic layer is bounded by other layers. Layered resistivity structures containing anisotropy will not generate a tipper response (Martí, 2013).

In 2-D structures, the effects of anisotropy can be quite complex and depend on the resistivity structures, the distribution and form of anisotropy, the site location, and the period (e.g., Heise and Pous, 2001). There are a number of effects that may be observed including tipper responses and induction arrows that are misaligned with the 2-D structure, divergence of MT strike directions from the true strike, MT phases lying outside the usual quadrant for 2-D responses, and the appearance of unrealistic structures in 2-D resistivity models fitted to the MT data (Pek and Verner, 1997; Heise and Pous, 2001, 2003; Wannamaker, 2005; Martí, 2013). In general, these effects can also be modeled in terms of 3-D structures although in some cases a 2-D anisotropic model provides a more geologically reasonable explanation. Discriminating the presence of anisotropy in 3-D resistivity structures is even more complex (Wannamaker, 2005; Martí, 2013). Weidelt (1999) showed that the response from any 3-D model containing anisotropy can equivalently be explained by a 3-D isotropic model of arbitrary complexity (Wannamaker, 2005).

2.8 Data Acquisition and Processing

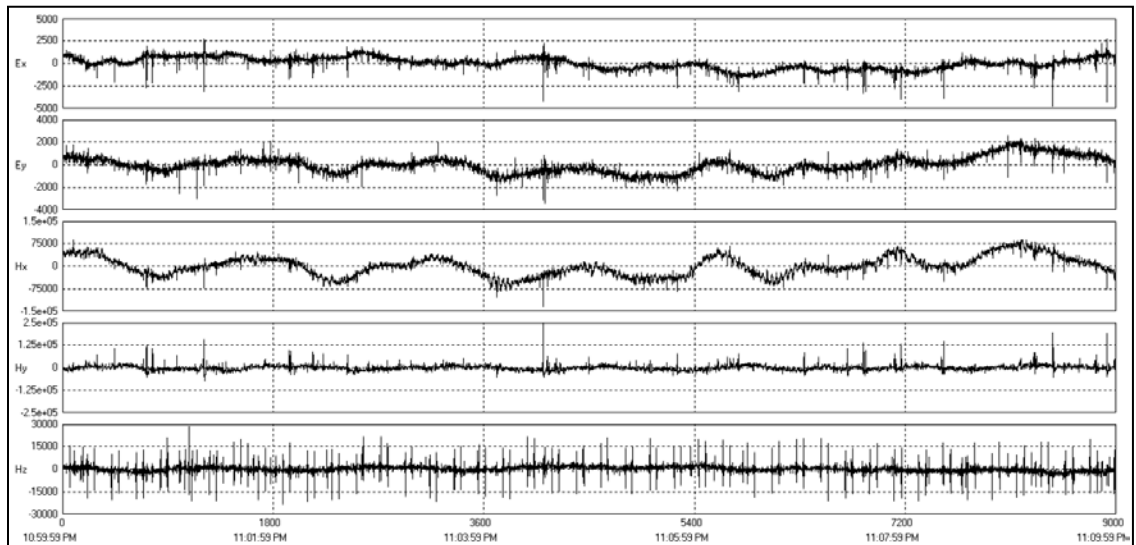
2.8.1 MT Data Acquisition

MT soundings for crustal and upper mantle studies are usually carried out using simultaneous recordings of time variations in the natural electric and magnetic fields. Five field components (Fig 2.6) are usually recorded as a function of time, including three components (two horizontal orthogonal and one vertical) of the magnetic field; [$B_x(t)$, $B_y(t)$, $B_z(t)$] and two components of the horizontal orthogonal electric field [$E_x(t)$, $E_y(t)$]. Remote-reference horizontal magnetic field components are often measured typically at a distance of several hundred meters to several tens of kilometers from the main site (Kaufman and Keller, 1981; Vozoff, 1991; Telford et al., 1990; Simpson and Bahr, 2005; Ferguson, 2012). The response functions can then be computed from recorded time series. In most MT surveys it is important to acquire high-quality data over a broad range of periods and in order to attain this purpose, high quality modern MT instruments, selection of appropriate sites, proper instrument installation and data filtering are necessary. A typical broadband MT station layout is shown in Figure 2.7.

Horizontal orthogonal electric field components, usually magnetic north-south and east-west, are measured by recording the voltage variation versus time between a pair of grounded electrodes. The common electrodes used in MT surveys are non-polarizing or porous pot Cu-CuSO₄, Pb-PbCl₂ and Ag-AgCl electrodes (Vozoff, 1991; Ferguson, 2012). The electrode consists of a porous base casing with metallic conductor immersed in a saturated solution (electrolyte) of a salt in order to avoid unwanted voltages generated by direct contact of the metal with the soil. This electrolyte acts as a liquid connection between the wire and the soil. Contact with the ground occurs by infiltration

of the solution through the porous base (Petiau and Dupis, 1980; Kaufman and Keller, 1981; Ferguson, 2012). The advantage of using non-polarizing electrodes is that the voltage difference between a pair of electrode is relatively stable. If a metal electrode such as copper or steel is used, relatively large time-varying potential differences resulting from electrochemical reactions at the metal surface can be present (Petiau and Dupis, 1980; Kaufman and Keller, 1981; Ferguson, 2012).

Figure 2.6. A 10-minute recorded time series for MT site PSO 001 showing E_x , E_y , B_x , B_y , and B_z components. The site is located at Picton in Ontario.



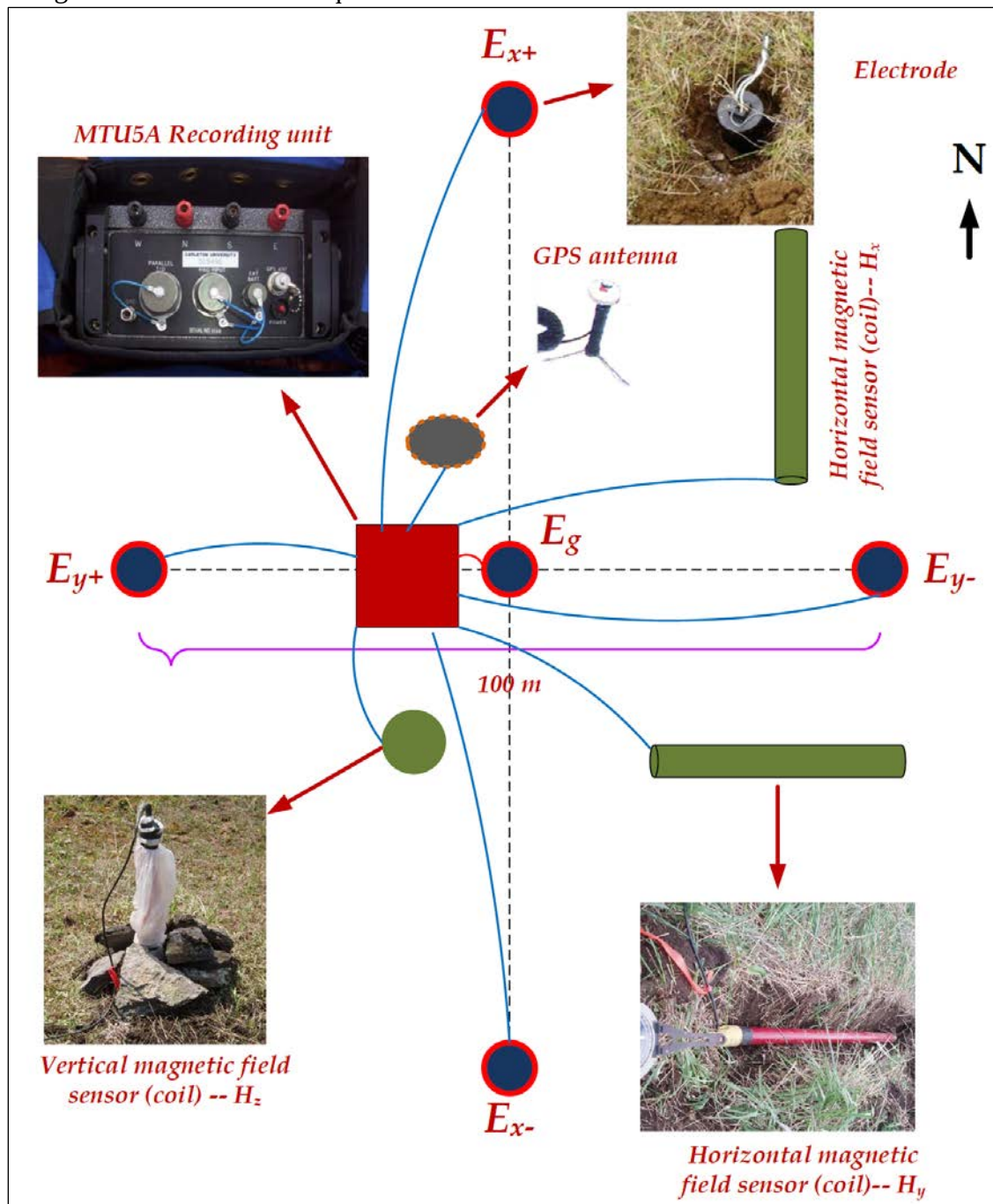
The main source of noise in an electric field measurement comes from the electrodes. This noise may be generated in the electrodes themselves by electrochemical reactions at the metal interface or in the soil around the electrodes. For periods larger than 100 s, the noise coming from the electronic amplifier can be made negligible (Petiau and Dupis, 1980). As most of the electric field noise comes from the electrodes, the most practical way to increase the signal-to-noise ratio is by using larger dipole lengths

(Vozoff, 1991; Simpson and Bahr, 2005; Ferguson, 2012). The electrodes are typically placed 100 m (Fig. 2.7) apart in order to give a measurable signal which must be much higher than noise levels.

Magnetic field components are commonly recorded by induction coils or fluxgate sensors. At high frequencies (above 0.001 Hz) an induction coil is used and for lower frequencies (less than 0.1 Hz), a three-component fluxgate magnetometer is used (Kaufman and Keller, 1981; Vozoff, 1991; Telford et al., 1990; Simpson and Bahr, 2005; Ferguson, 2012). Fluxgate magnetometers are suitable for measuring long-period magnetic field variations, which have high amplitudes; for periods shorter than a threshold period, the amplitude of the natural signal becomes weaker than the noise of the sensor (Simpson and Bahr, 2005).

Fluxgate magnetometers generally consist of two cores of easily saturable, high-permeability material, oppositely wound with coaxial excitation coils relying on the principle of hysteresis (Telford et al., 1990; Simpson and Bahr, 2005). The ambient Earth's magnetic field alters the point at which magnetic saturation occurs in the output signal time-sequence. Many of the fluxgate sensors used in MT are ring-core fluxgates (Kaufman and Keller, 1981; Ferguson, 2012) in which the sensor consists of circular metallic ribbons wound by an excitation coil and located within an external detector coil. The component of the magnetic field in the plane of the ring and parallel to the axis of the detector coil is measured (Burger, 1972). There is always the possibility of increasing the gain of magnetometers by configuring them as null instruments and this can be achieved by using a compensating solenoid to provide a negative feedback that will offset the baseline level of the Earth's field in the measurement direction (Andersen et al., 1988;

Figure 2.7. Schematic description of a typical broadband MT site layout with pictures of some parts of the Phoenix Geophysics MTU5A field instrument. The magnetic field components are recorded by two horizontal N-S and E-W oriented coils and one vertical coil. The electric field components are recorded by four electrodes oriented N-S and E-W and buried typically 100m apart and a fifth ground electrode (located at the center). The GPS antenna provides the recording unit with the site geographic coordinates and a continuous time signal. All the components are connected to a Phoenix Geophysics MTU5A recording unit which is usually powered by car batteries. Pictures were taken during POLARIS MT data acquisition.



Narod and Bennest, 1990; Ferguson, 2012). The measure of the magnetic field is provided by the current in the compensating solenoid provided the output of the magnetometer is successfully nulled.

Fundamental source of errors in fluxgate magnetometer measurements include inherent unbalance in the two cores, thermal and shock noise in cores, drift in the biasing circuit and temperature sensitivity (Narod and Bennest, 1990). Advantages of this instrument include direct readout, arbitrary azimuth of orientation, rather coarse leveling requirements, light weight, small size and reasonable sensitivity. Fluxgate magnetometers can measure any component of the magnetic field (Telford et al., 1990; Ferguson, 2012).

An induction coil is a cylinder with millions of turns of copper wire wound onto a high permeability core sealed within a shock resistant casing (Vozoff, 1991; Simpson and Bahr, 2005). Its signal output depends on the rate of change of flux density, dB/dt , so requires the integration of the output signal to determine the magnetic field (Ferguson, 2012). Since coil sensitivity depends on area of the coil and number of turns, there is a trade-off between transportability and sensitivity of coils designed for MT fieldwork. Induction coils with ferromagnetic core have higher sensitivity than those with air core because the ferromagnetic material acts as a flux concentration inside the coil. However, the introduction of ferromagnetic core also introduces some nonlinear factors which depends on temperature, frequency, flux density etc. These factors decrease the resolution of the sensor and alter the distribution of the investigated magnetic field. By using a large coil sensor, low frequency (mHz) magnetic fields can be investigated.

In an MT survey, magnetometer sensors, both induction coil and fluxgate sensors, are usually buried to minimize motion and give thermal stability (Kaufman and Keller, 1981; Simpson and Bahr, 2005; Tumanski, 2007; Ferguson, 2012).

2.8.2 Data Processing

Time series processing involves the estimation of the EM impedance transfer functions, in the frequency domain, from the recorded time series data. Generally, this process involves the estimation of a small number of transfer functions from time series with a very large number of data points. A number of mathematical methods have been developed and applied in MT impedance estimation. These include least-squares methods, admittance and remote reference methods, parametric methods with and without impedance dispersion constraints, time domain methods, wavelet methods, and robust methods. Jones et al. (1989) and Chave (2012) reviewed some of these methods and compared their performance in the estimates of impedance from geomagnetically quiet periods and from geomagnetically active periods.

The first step in transfer function estimation involves the subdivision of the time series into segments. The higher the number of segments, the better the statistical result but the lower the maximum period recovered. The segments are windowed (e.g., using a Hanning or Parzen window) and sometimes pre-whitened (Egbert and Booker, 1986).

The next stage is the conversion of the (pre-whitened) segmented data from time domain to frequency domain using Fourier Transforms. Once converted to the frequency domain, averaging of adjacent frequency estimates will help to improve the measured signal and reduce the number of data points. The averaging procedures make use of the

auto-spectra and cross-spectra calculated from the raw power spectra, A_k , for each time segment for each channel. Vozoff (1991) indicated that the auto-spectral density can be calculated for a single time series segment using:

$$\langle A(f_j) \rangle = \sqrt{\frac{1}{2m+1} \sum_{k=j-m}^{j+m} A_k A_k^*} = \sqrt{A_j A_j^*} . \quad 2.38$$

where $2m+1$ is the number of spectral estimates averaged and the asterisk indicates conjugation. The square of the right hand side of equation 2.40 is called the autopower spectral density at frequency f_j . The crosspower density of two channels is:

$$\langle A(f_j), B(f_j) \rangle = \frac{1}{2m+1} \sum_{k=j-m}^{j+m} A_k B_k^* = \langle A_j B_j^* \rangle . \quad 2.39$$

The final spectral estimates involve averaging the auto and cross powers over the multiple segments. This step is followed by the calculation of the impedance tensor and tipper vector using autopower and crosspower stored in the spectral matrix. As given by Vozoff (1991) and Simpson and Bahr (2005), these elements are estimated by equations such as:

$$Z_{xx} = \frac{\langle E_x H_x^* \rangle \langle H_y H_y^* \rangle - \langle H_y H_x^* \rangle \langle E_x H_y^* \rangle}{\langle E_x H_x^* \rangle \langle H_y H_y^* \rangle - \langle E_y H_x^* \rangle \langle H_x H_y^* \rangle} . \quad 2.40$$

Transfer function estimates are affected by statistical errors which are due to quantitative accuracy of the measurements and other sources of noise (Jones et al., 1989). The normal least-squares forms of the impedance estimates are biased upward or downward due to noise.

The remote reference method (Gamble et al., 1979; Chave, 2012) requires the simultaneous recording of the EM fields $R_x(t)$ and $R_y(t)$ at the main MT sites and an electromagnetically quiet remote location. Horizontal magnetic remote references are preferred to their electric counterpart because the magnetic fields are generally less contaminated by noise and are more homogeneous in areas around lateral inhomogeneities. The remote fields eliminate uncorrelated noise between the site and the remote fields but cannot remove the correlated noise. Using this method, the impedance elements and the tipper vector can be estimated as follows:

$$Z_{xx} = \frac{\langle E_x R_x^* \rangle \langle H_y R_y^* \rangle - \langle E_x R_y^* \rangle \langle H_y R_x^* \rangle}{\langle H_x R_x^* \rangle \langle H_y R_y^* \rangle - \langle H_x R_y^* \rangle \langle H_y R_x^* \rangle}, \quad 2.41$$

$$Z_{xy} = \frac{\langle E_x R_y^* \rangle \langle H_x R_x^* \rangle - \langle E_x R_x^* \rangle \langle H_x R_y^* \rangle}{\langle H_x R_x^* \rangle \langle H_y R_y^* \rangle - \langle H_x R_y^* \rangle \langle H_y R_x^* \rangle}, \quad 2.42$$

$$Z_{yx} = \frac{\langle E_y R_x^* \rangle \langle H_y R_y^* \rangle - \langle E_y R_y^* \rangle \langle H_y R_x^* \rangle}{\langle H_x R_x^* \rangle \langle H_y R_y^* \rangle - \langle H_x R_y^* \rangle \langle H_y R_x^* \rangle}, \quad 2.43$$

$$Z_{yy} = \frac{\langle E_y R_y^* \rangle \langle H_x R_x^* \rangle - \langle E_y R_x^* \rangle \langle H_x R_y^* \rangle}{\langle H_x R_x^* \rangle \langle H_y R_y^* \rangle - \langle H_x R_y^* \rangle \langle H_y R_x^* \rangle}, \quad 2.44$$

$$T_x = \frac{\langle H_z R_x^* \rangle \langle H_y R_y^* \rangle - \langle H_z R_y^* \rangle \langle H_y R_x^* \rangle}{\langle H_x R_x^* \rangle \langle H_y R_y^* \rangle - \langle H_x R_y^* \rangle \langle H_y R_x^* \rangle}, \quad 2.45$$

$$T_y = \frac{\langle H_z R_y^* \rangle \langle H_x R_x^* \rangle - \langle H_z R_x^* \rangle \langle H_x R_y^* \rangle}{\langle H_x R_x^* \rangle \langle H_y R_y^* \rangle - \langle H_x R_y^* \rangle \langle H_y R_x^* \rangle}, \quad 2.46$$

Although the development of the remote reference processing method marked a milestone in the production of high quality MT data, this method works well only if the noise is stationary and Gaussian in nature. The presence of non-Gaussian non-stationary noise makes the method fail. A number of researchers (e.g., Egbert and Booker, 1986; Chave and Thompson, 1989) have developed statistically robust processing schemes which iteratively re-weight the contribution from different time segments to the transfer function estimates. These methods can effectively produce stable estimates of the transfer functions with reasonable errors (Chave and Thomson, 1989; Chave, 2012). Simple robust schemes using weighting based on the coherence or variance of components in individual time segments have also been developed. Jones et al. (1989) compares the performance of several robust methods.

2.9 Data Analysis

2.9.1 Distortion

Distortions are the perturbation of MT responses by small-scale localized conductivity heterogeneities in the subsurface (Jones, 2012). They can be inductive or galvanic in nature. The inductive effect typically occurs at high frequencies whereas the galvanic effect occurs at low frequencies. Typically, MT data superimpose the galvanic distortion from any localized near-surface heterogeneities on the inductive signature of the regional structure (Fig. 2.8; Wannamaker et al., 1984; Jiracek, 1990; Chave and Jones, 1997; Jones, 2012).

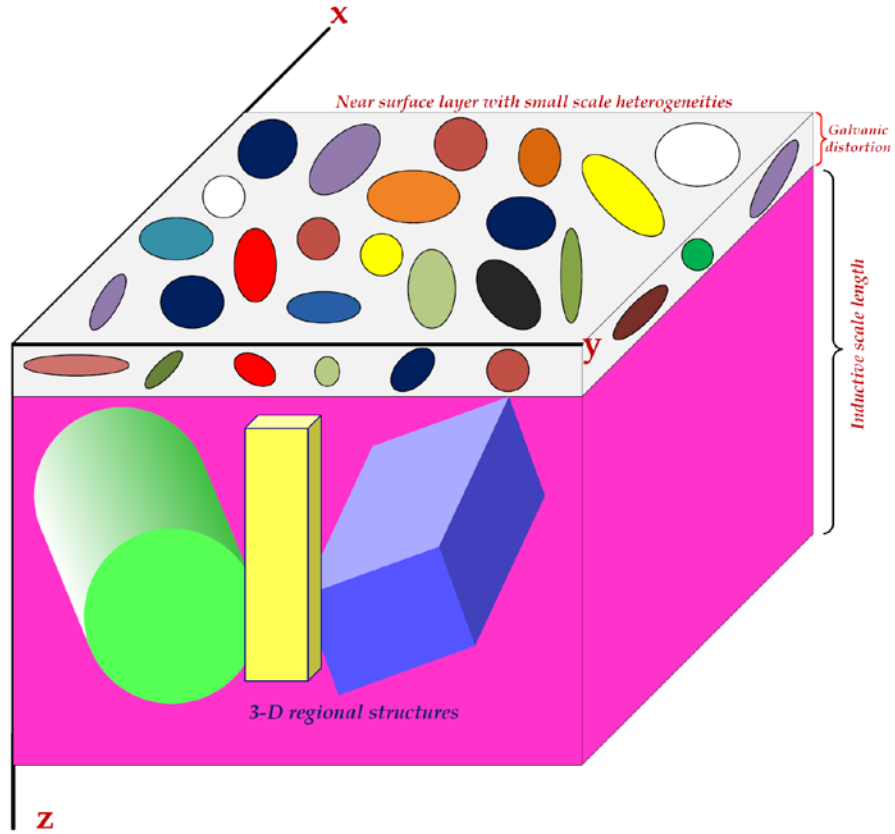
As described by Jiracek (1990) and Groom and Bahr (1992), the time-varying magnetic field of the Earth induces electric current flowing in closed loops, and this

current will in turn induce a secondary magnetic field which adds to the primary magnetic field. Perturbations of the induced secondary magnetic field cause the inductive distortion effect. Inductive distortions are frequency-dependent and they can be described as complex tensors that are characterized by amplitude and phase variations (Smith, 1997; McNeice and Jones, 2001). The distortions generally increase to saturation with increasing frequency. In MT, at sufficiently long periods, the inductive distortion effect can usually be ignored because its importance vanishes with increasing frequency (Chave and Jones, 1997; Jones, 2012). As indicated by Simpson and Bahr (2005), a body immersed in a half space (Fig. 2.4) will act as a local near-surface inhomogeneity when the period is sufficiently long and the skin depth is very large. This will result in a frequency-independent distortion, a non-inductive (galvanic) response.

Galvanic distortion is very important in the MT method and cannot be ignored. It occurs when the multi-dimensional near surface inhomogeneities have dimensions and depths that are much less than the skin depth of the EM signals in both the inhomogeneity and the host (McNeice and Jones, 2001). In contrast to inductive distortion, galvanic distortion increases to saturation with decreasing frequency. Its main effect is on the electric field but it can also affect the magnetic fields (magnetic galvanic distortion). In the case of electric distortion, the primary electric fields, which produce electrical charges at conductivity boundaries, become distorted because of the excess charges. Magnetic distortion occurs when the secondary magnetic fields associated with the currents deflected by electric distortion becomes significant (Chave and Smith, 1994; Chave and Jones, 1997; Smith, 1997; Jones, 2012). In the frequency range at which distortion is

purely galvanic, the effects are frequency independent and the effect can be described as real distortion tensors that affect only the amplitude of the MT impedance data.

Figure 2.8. Description of regional 3-D structure overlain by near-surface heterogeneities. For long-period signals, the regional structure falls into the inductive scale length whereas the near surface small-scale heterogeneities cause galvanic distortion.



The electric galvanic distortion can be represented by a matrix $\mathbf{C}$, whose elements are real and frequency-independent (Chave and Smith, 1994):

$$\mathbf{C} = \begin{pmatrix} C_{xx} & C_{xy} \\ C_{yx} & C_{yy} \end{pmatrix} \quad 2.47$$

The measured impedance, $\mathbf{Z}_{\text{obs}}$, in which the measurement is aligned with the regional

resistivity structure, is related to the regional impedance, $\mathbf{Z}_{2D}$, by:

$$\mathbf{Z}_{\text{obs}} = \mathbf{RCZ}_{2D}\mathbf{R}^T, \quad 2.48$$

The galvanic distortion of the magnetic fields will only be significant at high frequency or if the resistivity is high; it disappears at low frequencies and thus usually neglected in the MT method. The measured impedance in the presence of both electric and magnetic galvanic distortion is given as:

$$\mathbf{Z}_{\text{obs}} = \mathbf{RCZ}_{2D} (\mathbf{I} + \mathbf{DZ}_{2D})^{-1} \mathbf{R}^T \quad 2.49$$

where $\mathbf{I}$ is the identity matrix and $\mathbf{D}$ is the magnetic galvanic distortion matrix which is defined in a similar manner to $\mathbf{C}$.

The distortion complicates the MT response and appropriate analysis is required to separate these effects from the response of the regional structure. In order to remove distortion effects and gain better understanding of the geoelectric dimensionality, we carry out distortion analysis using tensor decomposition. Several approaches have been used to solve distortion problems, including the Groom-Bailey (GB) and Bahr methods, eigenvalue and singular value methods, graphical methods, tensor invariant method and the phase tensor method (Jones, 2012). The common elements of these various approaches are regional strike estimation, and measurement and description of galvanic distortion. It is important to note that static shift is a component of galvanic distortion that cannot be removed by tensor decomposition. Static shift is a shift (upward or downward) of the apparent resistivity curve occurs over the whole period range, in which the shape of the apparent resistivity curve, as well as the phase curve is unaffected. Static shifts is more pronounced in resistive than conductive mediums (Jiracek, 1990; Smith, 1997). The

Groom-Bailey tensor decomposition (Groom and Bailey, 1989) and phase tensor analysis (Caldwell et al., 2004) are described in the following two subsections.

2.9.2 Groom Bailey Tensor Decomposition

The Groom-Bailey (GB) decomposition method (Groom and Bailey, 1989, 1991) is one of the most commonly used decomposition techniques for determining the regional geoelectric strike direction and distortion-free impedance tensor from measured MT data. This method provides a model for the galvanic distortion of a regional 1-D or 2-D impedance. It deals only with the real and frequency-independent electric galvanic distortion matrix. The main focus of GB method is to factorize the measured impedance, $\mathbf{Z}_{\text{obs}}$, into a rotation matrix, $\mathbf{R}$, the distortion matrix, $\mathbf{C}$, and a scaled 2-D regional impedance tensor $\mathbf{Z}_{2D}$. This approach allows separation of the 3-D distortion from the regional 1-D or 2-D response. The telluric distortion tensor $\mathbf{C}$ is parameterized as a product of four effects which are separable into determinable and indeterminable parts as follows:

$$\mathbf{C} = \mathbf{gTSA}, \quad 2.50$$

Equation 2.48 thus becomes:

$$\mathbf{Z}_{\text{obs}} = \mathbf{RgTSAZ}_{2D}\mathbf{R}^T \quad 2.51$$

The matrix $\mathbf{A}$ with its normalizing factors is the local anisotropy or splitting operator and it is given in normalized form as:

$$\mathbf{A} = \frac{1}{\sqrt{1+s^2}} \begin{pmatrix} 1+s & 0 \\ 0 & 1-s \end{pmatrix}. \quad 2.52$$

The matrix $\mathbf{T}$, with its normalizing factor t is a twisting operator given as:

$$\mathbf{T} = \frac{1}{\sqrt{1+t^2}} \begin{pmatrix} 1 & -t \\ t & 1 \end{pmatrix}, \quad \text{where } t = \tan\phi_t, \quad 2.53$$

and the matrix $\mathbf{S}$, with its normalizing term e is a shearing operator given as:

$$\mathbf{S} = \frac{1}{\sqrt{1+e^2}} \begin{pmatrix} 1 & e \\ e & 1 \end{pmatrix}, \quad \text{where } s = \tan\phi_s. \quad 2.54$$

The effects of twist, shear and anisotropy when they are applied to a group of unit vectors is shown in Figure 2.9. The twist tensor rotates the electric field component through a clockwise angle ϕ_t . The shear polarizes the vectors to $\pm 45^\circ$ azimuths. The shear deflects the blue vector in clockwise direction by angle ϕ_s , and the red vector is deflected by the same angle in the anti-clockwise direction. The anisotropy stretches the field components by different factors and imposes anisotropy on the responses (Groom and Bailey, 1989).

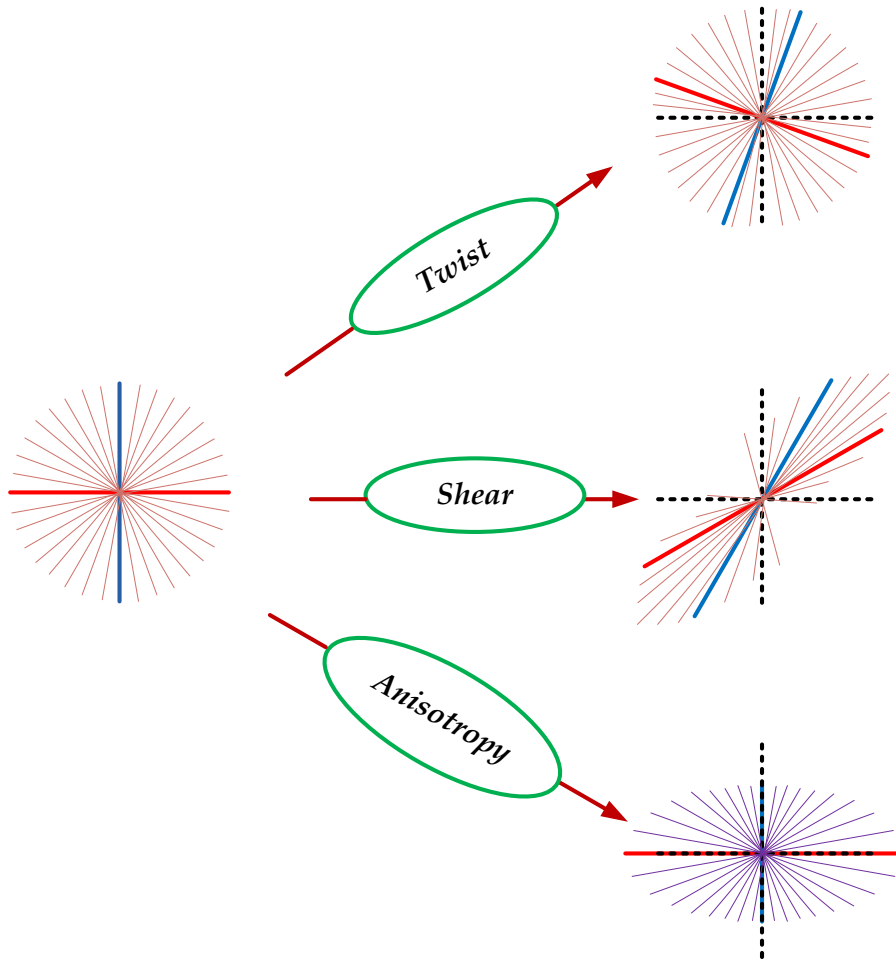
The term g is a real frequency-independent scale factor called site gain that increases or decreases all tensor terms by a constant factor. The anisotropy matrix $\mathbf{A}$ and the gain factor g cannot be determined separately from the regional impedance and are therefore absorbed into impedance to create a static shift (Groom and Bailey, 1989). The model of the distorted impedance, in the measurement coordinate, under GB factorization becomes:

$$\mathbf{Z}_{\text{obs}} = \mathbf{R} \mathbf{T} \mathbf{S} \mathbf{Z}_{2D} \mathbf{R}^T, \quad 2.55$$

$$\mathbf{Z}_{\text{obs}} = \begin{bmatrix} \cos\theta & \sin\theta \\ -\sin\theta & \cos\theta \end{bmatrix} \begin{bmatrix} 1 & -t \\ t & 1 \end{bmatrix} \begin{bmatrix} 1 & e \\ e & 1 \end{bmatrix} \begin{bmatrix} 0 & A \\ -B & 0 \end{bmatrix} \begin{bmatrix} \cos\theta & -\sin\theta \\ \sin\theta & \cos\theta \end{bmatrix} \quad 2.56$$

where A and B are regional TE and TM or TM and TE responses depending on 90° ambiguity in strike angle.

Figure 2.9. Schematic illustration of twist, shear and anisotropy. A group of unit vectors (left) with two reference vectors (shown in blue and red) which have twist, shear and anisotropy applied to them (right). The black dashed lines indicate the original position and length of the now deformed reference vectors (Redrawn from Groom and Bailey, 1989).



When the measured impedance obeys this model, at a particular frequency, there are seven real unknown parameters (twist, shear, strike and the four parameters contained in the two complex impedances A and B). We have eight data contributed by the four complex impedances. The seven unknown parameters are determined by fitting them to the eight data of the measured impedance tensor. This is an overdetermined problem. A

least-squares fitting procedure is used to fit the measured data with a model and the accuracy of the fit is measured by χ^2 misfit, which is a misfit parameter normalized by the data errors (Groom and Bailey, 1989). Groom and Bailey (1991) extend this analysis to fitting responses to multiple frequencies using the assumption that the distortion parameters and regional strike are independent of frequency.

McNeice and Jones (2001) decomposition extends the Groom-Bailey decomposition to multiple sites using the assumption that the regional strike direction is independent of both frequency and position. In this decomposition, the shear and the twist are independent of frequency but are allowed to vary between sites. The authors seek a global minimum to determine a regional geoelectric strike direction and distortion parameters for a range of frequencies and a set of sites. The method is useful for finding the optimal strike direction for a profile to be modeled in 2-D.

2.9.3 MT Phase Tensor

The MT phase tensor method is a distortion characterization procedure which does not require presumptions about the underlying dimensionality, and remains valid when both the heterogeneity and regional conductivity structures are 3-D (Caldwell et al., 2004). The main assumption of this method is that, provided the distortion is purely galvanic, only the amplitudes of the observed electric field are distorted by near-surface heterogeneities and that the phase relationship between the horizontal electric and magnetic field vectors is unaffected.

Separating the impedance tensor (**Z**) into the real (**X**) and imaginary components (**Y**);

$$\mathbf{Z} = \mathbf{X} + i\mathbf{Y} \quad 2.57$$

the second rank MT phase tensor, Φ , is defined as the ratio of the real to the imaginary parts of the impedance tensor and this is given as:

$$\Phi = \mathbf{X}^{-1} \mathbf{Y}. \quad 2.58$$

For a regional impedance tensor ($\mathbf{Z}_R = \mathbf{X}_R + i\mathbf{Y}_R$), galvanic distortion of the EM field scales both the real and imaginary parts of the impedance by $\mathbf{C}$ i.e. $\mathbf{X} = \mathbf{C}\mathbf{X}_R$ and $\mathbf{Y} = \mathbf{C}\mathbf{Y}_R$. The phase tensor can be expressed in terms of the regional field:

$$\Phi = \mathbf{X}^{-1} \mathbf{Y} = (\mathbf{C}\mathbf{X}_R)^{-1} (\mathbf{C}\mathbf{Y}_R) = \mathbf{X}_R^{-1} \mathbf{C}^{-1} \mathbf{C} \mathbf{Y}_R = \mathbf{X}_R^{-1} \mathbf{Y}_R = \Phi_R \quad 2.59$$

and it is evident that it is independent of the galvanic distortion.

The four elements of the phase tensor represent distortion free parts of the regional impedance. An ellipse defined by these four quantities is used to graphically represent the phase tensor for each frequency (Fig. 2.10). This ellipse is characterized by one direction α (which expresses the tensor's dependence on the coordinate), and three coordinate invariants, $\Phi_{\max}$ (maximum phase tensor), $\Phi_{\min}$ (minimum phase tensor), and β (phase tensor skew angle, a measure of asymmetry) (Caldwell et al., 2004; Heise et al., 2007; Hill et al., 2009).

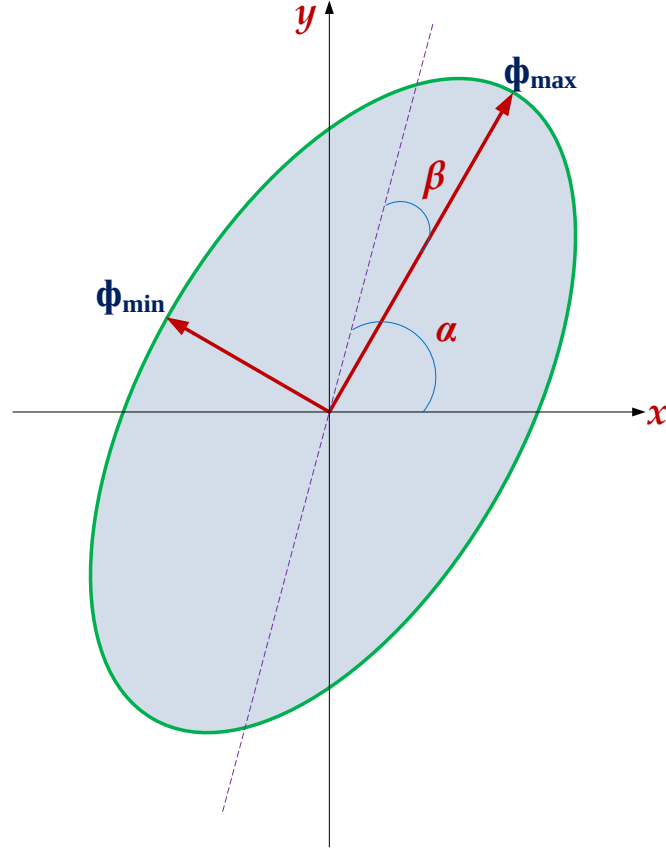
The phase tensor may be represented as the product of three matrices:

$$\Phi = \mathbf{R}^T (\alpha - \beta) \begin{pmatrix} \Phi_{\max} & 0 \\ 0 & \Phi_{\min} \end{pmatrix} \mathbf{R} (\alpha + \beta) \quad 2.60$$

The angles α and β are given as:

$$\alpha = \frac{1}{2} \tan^{-1} \left(\frac{\Phi_{12} - \Phi_{21}}{\Phi_{11} + \Phi_{22}} \right) \quad \text{and} \quad \beta = \frac{1}{2} \tan^{-1} \left(\frac{\Phi_{12} - \Phi_{21}}{\Phi_{11} + \Phi_{22}} \right). \quad 2.61$$

Figure 2.10. Graphical illustration of the phase tensor. Coordinate invariants $\Phi_{\max}$ and $\Phi_{\min}$ define the magnitudes of the major and minor axes. The skew angle β measures the tensor's asymmetry and the angle α indicates the tensor's dependence on the coordinate system (x, y) . The dashed line is the axis of symmetry and the direction of the major axis $\Phi_{\max}$ is defined by $\alpha - \beta$ (Redrawn from Caldwell et al., 2004).



The tensor is defined by the three coordinate invariants as well as the direction:

$$\Phi_{\max} = \left(\Phi_1^2 + \Phi_3^2 \right)^{1/2} + \left(\Phi_1^2 + \Phi_3^2 - \Phi_2^2 \right)^{1/2}, \quad 2.62$$

and

$$\Phi_{\min} = \left(\Phi_1^2 + \Phi_3^2 \right)^{1/2} - \left(\Phi_1^2 + \Phi_3^2 - \Phi_2^2 \right)^{1/2}, \quad 2.63$$

where $\Phi_1 = \frac{(\Phi_{11} + \Phi_{22})}{2}$, $\Phi_2 = [\det(\Phi)]^{1/2}$, and $\Phi_3 = \frac{(\Phi_{12} - \Phi_{21})}{2}$.

The orientation of the major axis of the tensor is given by the angle α , which defines the relationship between reference frame and the tensor. If $\beta=0$, the tensor is symmetric.

When the Earth is 1-D, $\Phi_{\max} = \Phi_{\min} = 0$, and therefore, the phase tensor ellipse will be represented by a circle. The phase tensor skew angle β will be zero and the direction α is undefined (i.e. geoelectric strike is not defined for 1-D structure). In the case of a regional 2-D resistivity structure, the skew angle β will be zero and one principal axis of the ellipse aligns with the geoelectric strike. In this case, the phase tensor is symmetric. In the case of a regional 3-D structure, the phase tensor is not symmetric and all the invariants are non-zero. The magnitude of β measures deviation from the 2-D structure, where a large value of the phase tensor skew angle indicates a 3-D conductivity structure. However, a small value does not necessarily mean that the structure is not 3-D. The spatial variation of the direction of the major axis of the tensor ellipse provides a better discrimination between 2-D or 3-D structures (Caldwell et al., 2004).

2.10 Modeling and Inversion

The electrical properties derived from the MT impedance responses are used to construct the Earth model (Vozoff, 1991; Chave and Jones, 2012). Inversion is an automatic method of generating resistivity models that statistically fit the observed data. It aims to obtain, through the use of theoretical relationships, model parameters from the actual data which produce responses that best fit those data. It automatically optimizes the generated model between different forward solutions by iteratively updating the model parameters. Many 1-D, 2-D and 3-D inversion schemes exist and are used widely

in MT studies (Jupp and Vozoff, 1975; Vozoff, 1991; Whittall and Oldenburg, 1992; Siripunvaraporn, 2012; Miensopust et al., 2013).

2.10.1 Forward modeling

Forward modeling is a trial-and-error method by which different plausible models are used to calculate response functions which are compared with the real data. In EM, the process involves calculating the theoretical response of geoelectric models. This approach, while open ended, is a means of establishing a group of geoelectric models that is most consistent with the observed data and other available information (Hohmann, 1988; Whittall and Oldenburg, 1992).

Forward modeling can be approached in 1-D, 2-D, and 3-D fashions depending on the complexity of the geological/geoelectric structure. 1-D forward modeling is the simplest and most well understood method since it is based on a recursive calculation of the MT impedance at the surface of a layered conductivity structure using equation 2.29. For 2-D and 3-D structures, the aim of forward modeling is to solve Maxwell's equations in piece-wise uniform media based on integral equation (IE; Hohmann, 1983; Wannamaker et al., 1984a), finite element (FE; Wannamaker et al., 1987; Best et al., 1985), finite difference (FD; Mackie et al., 1993) or thin sheet (Dawson and Weaver, 1979; Dawson et al., 1982; Dawson, 1983; McKirdy and Weaver, 1984, 1985) approaches. Wannamaker et al. (1984b) describe an IE algorithm for MT problems that can handle a 3-D body in an arbitrarily layered 1-D Earth. A 2-D FE code has also been developed by Wannamaker et al. (1987) and hybrid methods involving both IE and FE codes are also in use (Best et al., 1985). Thin sheet modeling has become a common

means for treating 3-D surface conductors such as oceans. It should be noted that both approximate and the more complex full solutions to EM problems are useful (West and Edwards, 1985).

2-D models are appropriate for MT data that exhibit two-dimensionality in their responses, while 3-D models must be used for data that cannot be validly modeled using 2-D. As suggested by Jones and Craven (2004), 2-D modeling should follow dimensionality tests within a statistical framework that considers the effects of local electric field distortion and also noise in the data, and tensor decomposition providing the regional 2-D responses (e.g., Bahr, 1988; Groom and Bailey, 1989; McNeice and Jones, 2001; Caldwell et al., 2004; Jones, 2012) or 3-D (e.g., Garcia and Jones, 2002b) responses.

2.10.2 1-D Inversion

Objective inversion methods, such as the Marquardt or Occam algorithm, can be used to fit MT data with 1-D models. The models may use either a minimum number of layers (e.g., Fischer and Le Quang, 1982) or a minimum resistivity change between many layers in an over-parameterized problem (Constable et al., 1987; Smith and Booker, 1988). Early 1-D inversion algorithms were based on small number of layers and linearized inversion for minimum structure (e.g., Marquardt algorithm) (Jupp and Vozoff, 1975). Later algorithms involved regularized overparameterized response. Constable et al. (1987) introduced Occam's inversion which is a practical algorithm for generating smooth 1-D models from MT sounding data. The method generates the smoothest model possible by fitting the data to within an expected tolerance (e.g., as defined by the D^+

algorithm by Parker, 1980). This method was later extended into the inversion of MT data for smooth 2-D models (deGroot-Hedlin and Constable, 1990).

2.10.3 2-D Inversion

In 2-D modeling, the responses of interest are the transverse electric (TE) mode, with electric fields parallel to the geoelectric strike, the transverse magnetic (TM) mode, with electric fields perpendicular to strike, and the transfer function (tipper response), which defines the relationship between the vertical and horizontal magnetic field components (Vozoff, 1991). The TE mode is more sensitive to the conductance of conductive regions of the model than TM, but provides lower resolution of the position of structures. The TM response is more sensitive to lateral boundaries than the TE mode. The tipper is usually sensitive to the local structure but can also be strongly affected by conductivity boundaries at the edges or outside of the region modeled. Inversion of only TE data resolves only the depth and the conductivity-thickness product of the conducting regions whereas inversion of only TM data resolves their horizontal extent and the presence of thin resistive layers (Vozoff, 1991; Agarwal et al., 1993; Simpson and Bahr, 2005). Using both modes in the inversion allows for a better evaluation of the resistivity structure over a large area. When the conductive features in a model have a finite strike length, the TE response is more distorted than the TM response and this necessitates the weighting of the 2-D modeling heavily towards the TM response (Wannamaker, 1999; Ledo et al., 2002).

A rapid 2-D MT inversion for minimum structure, called the Rapid Relaxation Inversion (RRI) has been developed by Smith and Booker (1991). This method represents

the structure as a series of 1-D models with a correction for lateral field gradients, but uses the full 2-D forward model to calculate the response. Several other 2-D inversion algorithms have been developed (e.g., Oldenburg and Ellis, 1993; Uchida, 1993; Siripunvaraporn and Egbert, 2000; Rodi and Mackie, 2001). Mackie et al. (1988) forward modeling and Rodi and Mackie (2001) 2-D inversion codes implemented in the WinGLinK software are used for the data analyses in this thesis. The forward modeling is based on a finite difference approximation of the Maxwell equations. It discretizes the model into piecewise rectangular grids, calculates the unknown EM fields at the points of space discretization using finite differences, and obtains discrete approximations to partial derivatives (Salazar-Palma et al., 1998). The inversion code is based on non-linear conjugate gradients (NLCG) for direct iterative minimization of the Tikhonov regularization function. The data misfit criterion used is a normalized χ^2 misfit. The mesh defining the model used in forward modeling is also used as the regularization grid in the inversion.

2.10.4 3-D Inversion

The MT impedance response over 3-D structures is more complex than 2-D, and several studies have implemented 3-D inversion approaches (Tuncer et al., 2006; Sasaki and Meju, 2006; Mackie and Watts, 2007; Farquharson and Craven, 2008; Patro and Egbert, 2008; Heiss et al., 2008; Jones et al., 2008;). Many 3-D inversion algorithms have been developed for MT data (e.g., Wannamaker, 1991; Xiong, 1992; Mackie and Madden, 1993; Sasaki, 2001; Farquharson et al., 2002; Mitsuhashi and Uchida, 2004; Siripunvaraporn et al., 2005; Sasaki and Meju, 2006; Nam et al., 2007; Han et al., 2008;

Siripunvaraporn and Egbert 2009; Miensoopust et al., 2013). All of these algorithms require large amounts of time and computer memory, and are therefore difficult to use. Siripunvaraporn et al. (2005) is a 3-D inversion algorithm that is available to the academic community and it is based on a data space approach which is designed to reduce the amount of computer time required.

2.11 Applications of MT

MT finds application in the determination of the subsurface resistivity structures, by detecting regions of enhanced conductivity at crustal and lithospheric mantle depths (e.g., Jones and Craven, 1990; Schwarz, 1990; Nover et al., 1998; Jones, 1999; Jones et al., 2005a; Evans et al., 2005; Spratt et al., 2009; Miensoopust et al., 2011; Evans et al., 2011; Ferguson et al., 2012). The method has been used widely in the investigation of lithospheric structures and thicknesses (e.g., Schultz et al., 1993; Jones et al., 2003; Korja, 2007; Heinson and White, 2005; Muller et al., 2009, Evans et al., 2011). MT can also be used to investigate geothermal energy and mineral resources (e.g., Meju, 2002; Jones and Garcia, 2003; Jones and Craven, 2004; Jones et al., 2009; Tuncer et al., 2009; Spichak and Manzella, 2009).

Several authors have used MT to define electrical resistivity anisotropy within the crust and the lithospheric mantle (e.g., Kellett et al., 1992; Eisel and Bahr, 1993; Kurtz et al., 1993; Ji et al., 1996; Eisel and Haak, 1999; Mareschal et al., 1995; Bahr and Simpson, 2002; Eaton et al., 2004; Heinson and White, 2005; Wannamaker et al., 2004; Baba et al., 2006; Frederiksen et al., 2006; Hamilton et al., 2006; Padilha et al., 2006; Wannamaker et

al., 2008; Brasse et al., 2009; Miensopust and Jones, 2011; LePape et al., 2012). Most of these studies have been reviewed by Wannamaker (2005) and Marti (2012).

In Canada, MT has been used to define major conductivity anomalies such as the North American Central Plains (NACP) anomaly, the Athapapuskow Lake conductivity anomaly (ALCA) and the Thompson Belt (TOBE) anomaly (Jones and Savage, 1986; Jones and Craven, 1990; Jones et al., 1993; Ferguson et al., 1999; Ferguson et al., 2005b; Jones et al., 2005a). Several authors have described important crustal and lithospheric resistivity structures obtained using MT (e.g., Mareschal et al., 1994; Jones and Gough, 1995; Boerner et al., 1999; Boerner et al., 2000a; Boerner et al., 2000b; Garcia and Jones, 2005; Jones et al., 2005b; Ferguson et al., 2005a). A complete review of the Canada's lithospheric resistivity is provided in Jones et al. (2013).

2.12 References Cited

- Agarwal, A. K., Poll, H. E., Weaver, J. T., 1993. One and two-dimensional inversion of MT data in continental regions. *Phys. Earth Planet. Inter.* 81, 155-176.
- Andersen, F., Boerner, B. B., Harding, K., Jones, A. G., Kurtz, R. D., Parmelee, J., Trigg, D., 1988. LIMS: Long period intelligent magnetotelluric system. 9th Workshop on EM Induction in the Earth, Sochi, USSR.
- Baba, K., Chave, A. D., Evans, R. L., Hirth, G., Mackie, R. L., 2006. Mantle dynamics beneath the East Pacific Rise at 17°S: Insights from the Mantle Electromagnetic and Tomography (MELT) experiment. *J. Geophys. Res.* 111, B02101.
- Bahr, K., 1988. Interpretation of the magnetotelluric impedance tensor: regional induction and local telluric distortion. *J. Geophys.* 62, 119–127.
- Bahr, K., Simpson, F., 2002. Electrical anisotropy below slow- and fast-moving plates: Paleoflow in the upper mantle? *Science* 295, 1270–1272.
- Best, M.E., Duncan, P., Jacobs, F. J., Scheen, W. L., 1985. Numerical modeling of the electromagnetic response of three-dimensional conductors in a layered earth. *Geophysics* 50, 665-676.
- Boerner, D. E., Kurtz, R. D., Craven, J. A., 2000a. A summary of electromagnetic studies on the Abitibi-Grenville transect. *Can. J. Earth Sci.* 37, 427-437.
- Boerner, D.E., Kurtz, R. D., Craven, J. A., Ross, G. M., Jones, F. W., 2000b. A synthesis of electromagnetic studies in the Lithoprobe Alberta Basement Transect: constraints on Paleoproterozoic indentation tectonics. *Can. J. Earth Sci.* 37, 1509-1534.
- Boerner, D.E., Kurtz, R.D., Craven, J.A., Ross, G.M., Jones, F.W., Davis, W. J., 1999. Electrical conductivity in the Precambrian lithosphere of Western Canada. *Science* 283, 668– 670.
- Brasse, H., Kapinos, G., Li, Y., Mtschard, L., Soyer, W., Eydam, D., 2009. Structural electrical anisotropy in the crust at the South-Central Chilean continental margin as inferred from geomagnetic transfer functions. *J. Phys. Earth Planet. Int.* 173, 7 – 16,
- Burger, J.R., 1972. The theoretical output of a ring core fluxgate sensor. *IEEE Trans. On Magnetics* Mag-8, 791-796.
- Cagniard, L., 1953. Basic theory of the magneto-telluric method of geophysical prospecting. *Geophysics* 18, 605 – 635.
- Caldwell, T.G., H.M. Bibby and C. Brown, 2004 The magnetotelluric phase tensor: *Geophys. J. Int.* 158, 457-469.
- Campbell, W. H., 1987. Introduction to Electrical Properties of the Earth's mantle. *Pure Appl. Geophys.* 125, 193-504.
- Chave, A. D., 2012. Estimation of magnetotelluric response function, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 165-218.
- Chave, A. D., Jones, A. G., 1997. Electric and magnetic field galvanic distortion decomposition of BC87 data. *J. Geomag. Geoelectr.* 49, 767 – 789.
- Chave, A. D., Jones, A.G. (Eds.), 2012. *The Magnetotelluric Method - Theory and Practice*, Cambridge University Press, New York, 570pp.

- Chave, A. D., Smith, J. T., 1994. On electric and magnetic galvanic distortion tensor decompositions. *J. Geophys. Res.* 99, 4669 – 4682.
- Chave, A. D., Thomson, D. J., 1989. Some comments on magnetotelluric response function estimation, *J. Geophys. Res.* 94, 14215-14225.
- Constable S. C., Parker, R. L., Constable, C. G., 1987. Occam's inversion: a practical algorithm for generating smooth models from electromagnetic sounding data. *Geophysics* 52, 289 – 300.
- Constable, S. C., 2006. SEO3: a new model of olivine electrical conductivity. *Geophys. J. Int.* 166, 435-437.
- Dawson, T.W., 1983. H-polarization induction in two thin half-sheets. *Geophys. J. R. Astr. Soc.* 73, 83-107.
- Dawson, T.W., Weaver, J. T., 1979. Three-dimensional induction in a non-uniform thin sheet at the surface of a uniformly conduction Earth. *Geophys. J. R. Astr. Soc.* 59, 45-462.
- Dawson, T.W., Weaver, J. T., Raval, U., 1982. B-polarization induction in two generalized thin sheets at the surface of a conducting half-space. *Geophys. J. R. Astr. Soc.* 69, 209-234.
- deGroot-Hedlin, C., Constable, S. C., 1990. Occam's Inversion to generate smooth, two-dimensional models from magnetotelluric data. *Geophysics* 55, 1613-1624.
- Ducea, M. N, Park, S. K., 2000. Enhanced mantle conductivity from sulfide minerals, southern Sierra Nevada, California. *Geophys. Res. Lett.* 27, 2405-2408.
- Eaton, D., Frederiksen, A., Miong, S.K., 2004. Shear-wave splitting observations in the lower Great Lakes region: Evidence for regional anisotropic domains and keel modified asthenospheric flow. *Geophys. Res. Lett.* 31, L07610.
- Egbert, G. D., 1997. Robust multiple-station magnetotelluric data processing. *Geophys. J. Int.* 130, 475–496.
- Egbert, G. D., Booker, J. R., 1986. Robust estimation of geomagnetic transfer functions. *Geophys. J. R. Astro. Soc.* 87, 173–194.
- Eisel, M., Bahr, K., 1993. Electrical anisotropy in the lower crust of British-Columbia - an interpretation of a magnetotelluric profile after tensor decomposition. *J. Geomag. Geoelectr.* 45, 1115 – 1126.
- Eisel, M., Haak, V., 1999. Macro-anisotropy of the electrical conductivity of the crust: a Magnetotelluric study of the German Continental Deep drilling site (KTB), *Geophys. J. Int.* 136, 109– 122.
- ELEKTG Group, 1997. KTB and the electrical conductivity of the crust, *J. Geophys. Res.* 102, 18289-18306.
- Evans R. L., 2012. Earth's electromagnetic environment, in: Chave, A. D., Jones, A. G. (Eds.), 2012. *The Magnetotelluric Method - Theory and Practice*, Cambridge University Press, New York, pp. 50-95.
- Evans, R. L., Hirth, G., Baba, K., Forsyth, D., Chave, A., Mackie, R., 2005. Geophysical controls from the MELT area for compositional controls on oceanic plates. *Nature* 437, 249–252.
- Evans, R. L., Jones, A. G., Garcia, X., Muller, M., Hamilton, M., Evans, S., Fourie, C. J. S., Spratt, J., Webb, S., Jelsma, H., Hutchins, D., 2011. Electrical lithosphere beneath the Kaapvaal craton, southern Africa. *J. Geophys. Res.* 116, B04105.

- Everett, M. E., 2005. What do electromagnetic induction responses measure? *The Leading Edge* 24, 154 – 157.
- Farquharson, C. G., Oldenburg, D. W., Haber, E., Shekhtman, R., 2002. An algorithm for the three-dimensional inversion of magnetotelluric data. SEG Int'l Exposition and 72nd Annual Meeting, Salt Lake City, Utah.
- Farquharson, C.G., Craven, J.A., 2008. Three-dimensional inversion of Magnetotelluric data for mineral exploration: an example from the McArthur River uranium deposit, Saskatchewan, Canada. *J. Appl. Geophys.* 68, 450–458
- Ferguson, I. J., 2012. Instrumentation and field procedures, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 421-479.
- Ferguson, I. J., Jones, A. G., Chave, A. D., 2012. Case histories and geological applications, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 480-544.
- Ferguson, I.J., Craven, J.A., Kurtz, R. D., Boerner, D. E., Bailey, R. C., Wu, X., Orellana, M. R., Spratt, J., Wennberg, G., Norton, M., 2005a. Geoelectric responses of Archean lithosphere in the western Superior Province, central Canada. *Phys. Earth Planet. Int.* 150, 123–143.
- Ferguson, I.J., Jones, A.G., Yu, S., Wu, X., Shiozaki, I., 1999. Geoelectric response and crustal electrical-conductivity structure of the Flin Flon Belt, Trans-Hudson Orogen, Canada. *Can. J. Earth Sci.* 36, 1917–1938.
- Ferguson, I.J., Stevens, K., Jones, A.G., 2005b. Electrical resistivity imaging of the central Trans-Hudson Orogen. *Can. J. Earth Sci.* 42, 495–515. doi:10. 1139/e05-017.
- Fischer, G., LeQuang, B. V., 1982. Parameter trade-off in one-dimensional magnetotelluric modeling. *J. Geophys.* 51, 206-215.
- Frederiksen, A. W., Ferguson, I.J., Eaton, D., Miong, S.K., Gowan, E., 2006. Mantle fabric at multiple scales across an Archean-Proterozoic boundary, Grenville Front, Canada. *Phys. Earth Planet. Int.* 158, 240-263.
- Fullea, J., Muller, M.R., Jones, A.G., 2011. Electrical conductivity of continental lithospheric mantle from integrated geophysical and petrological modelling: Application to the Kaapvaal Craton and Rehoboth Terrane, southern Africa. *J. Geophys. Res.* 116, B10202, doi: 10.1029/2011jb008544.
- Gamble, T. D., Goubau, W. M., Clarke, J., 1979. Magnetotellurics with a remote magnetic reference. *Geophysiscs* 44, 53 – 68.
- Garcia, X. and A. G. Jones, 2002b. Decomposition of three-dimensional magnetotelluric data. in: M. S. Zhdanov, M.S., Wannamaker, P. E. (Eds.), *Three-Dimensional Electromagnetics*, volume 35 of *Methods in Geochemistry and Geophy.*, pp. 235 – 250. Elsevier. ISBN 0-444-50429-X.
- Garcia, X., Jones, A. G., 2002a. Atmospheric sources for audio-magnetotelluric (AMT) sounding. *Geophysics* 67, 448 – 458.
- Garcia, X., Jones, A. G., 2005. Electromagnetic image of the Trans-Hudson orogen — THO94 transect, *Can. J. Earth Sci.* 42, 479-493.
- Garcia, X., Ledo, J., Queralt, P., 1999. 2D inversion of 3D magnetotelluric data: The Kayabe dataset. *Earth Planets Space* 51, 1135 – 1143.

- Glover, P.W. J., Vine, F.J., 1995. Beyond KTB - Electrical conductivity of the deep continental crust, *Surveys in Geophysics* 16, 5 - 36.
- Grant, F.S., West, G. F., 1965. *Interpretation Theory in Applied Geophysics*, McGraw-Hill, New York, 583 pp.
- Groom, R. W., Bahr, K., 1992. Corrections for near surface effects: Decomposition of the magnetotelluric impedance tensor and scaling corrections for regional resistivities: A tutorial. *Surveys in Geophysics* 13, 341–379.
- Groom, R. W., Bailey, R. C., 1989. Decomposition of magnetotelluric impedance tensors in the presence of local three-dimensional galvanic distortion. *J. Geophys. Res.* 94, 1913-1925.
- Groom, R. W., Bailey, R. C., 1991. Analytic investigations of the effects of near-surface three-dimensional galvanic scatterers on MT tensor decompositions. *Geophysics* 56, 496 – 518.
- Haak, V., Hutton, V.R.S., 1986. Electrical Resistivity in the Continental Lower Crust, in: Dawson, J.B, Carswell, D.A., Hall, J., Wedepol, K.H. (Eds.), *The Nature of the Lower Continental Crust*, Spec. Pub. Geol. Soc., London, 24, 35–49.
- Hamilton, M. P., Jones, A. G., Evans, R. L., Evans, S., Fourie, C. J. S., Garcia, X., Mountford, A., Spratt, J. E., the SAMTEX Team, 2006. Electrical anisotropy of South African lithosphere compared with seismic anisotropy from shear-wave splitting analysis. *Phys. Earth Planet. Int.* 158, 226–239.
- Han N, Nam MJ, Kim HJ, Lee TJ, Song Y, Suh JH (2008) Efficient three-dimensional inversion of Magnetotelluric data using approximate sensitivities. *Geophys J Inter* 175:477–485
- Heinson, G., 1999. Electromagnetic studies of lithosphere and asthenosphere. *Surveys in Geophysics* 20, 229 – 255.
- Heinson, G., Constable, S., 1992. The electrical conductivity of the oceanic upper mantle. *Geophys. J. Int.* 110, 159 - 179.
- Heinson, G., White, A., 2005. Electrical resistivity of the Northern Australian lithosphere: Crustal anisotropy or mantle heterogeneity? *Earth Planet Sci. Lett.* 232, 157-170. doi:10.1016/j.epsl.2004.12.029.
- Heinson, G., White, A., 2005. Electrical resistivity of the Northern Australian lithosphere: crustal anisotropy or mantle heterogeneity? *Earth and Planetary Science Letters* 232, 157–170.
- Heise, W., Bibby, H.M., Caldwell, T.G., Bannister, S.C., Ogawa, Y., Takakura, S., Uchida, T., 2007. Melt distribution beneath a young continental rift: The Taupo volcanic zone, New Zealand: *Geophys. Res. Lett.* 34, L14313.
- Heise, W., Caldwell, T.G., Bibby, H.M., Bannister, S.C. 2008. Three-dimensional modelling of magnetotelluric data from the Rotokawa geothermal field, Taupo Volcanic Zone, New Zealand. *Geophys. J. Int.* 173, 740–750.
- Heise, W., Caldwell, T.G., Bibby, H.M., Brown, C., 2006. Anisotropy and phase splits in magnetotellurics. *Phys. Earth Planet. Int.* 158, 107- 121.
- Heise, W., Pous, J., 2001. Effects of anisotropy on the two-dimensional inversion procedure. *Geophys. J. Int.* 147, 610-621. doi:10.1046/j.0956-540x.2001.01560.x
- Heise, W., Pous, J., 2003. Anomalous phases exceeding 90° in magnetotellurics: anisotropic model studies and a field example. *Geophys. J. Int.* 155, 308-318. doi:10.1046/j.1365-246X.2003.02050.x

- Hill, G.J, Caldwell, T.G., Heise, W., Chertkoff, D.G., Bibby, H.M., Burgess, M.K., Cull, J.P., Cas, R.A.F., 2009. Distribution of melt beneath Mount St Helens and Mount Adams inferred from magnetotelluric data. *Nature Geoscience* 2, 785- 789.
- Hirth, G., Evans, R. L., Chave, A. D., 2000. Comparison of continental and oceanic mantle electrical conductivity: Is the Archean lithosphere dry? *Geochem. Geophys. Geosyst.* 1, 1030, doi:10.1029/2000GC000048.
- Hohmann, G. W., 1983. Three-dimensional EM modeling. *Geophys. Surv.* 6, 27-53.
- Hohmann, G.W., 1988. Numerical modelling for electromagnetic methods of Geophysics, in: , Nabighlan, M.N. (Ed.), *Electromagnetic Methods in Applied Geophysics*, Vol. 1, Theory, Investigations in Geophysics no. 3, SEG, Tulsa, Oklahoma.
- Hyndman R.D., Shearer, P.M., 1989. Water in the lower continental crust: modelling magnetotelluric and seismic reflection results, *Geophys. J. Int.* 98, 343-365.
- Ji, S., Rondenay, S., Mareschal, M., Senechal, G., 1996. Obliquity between seismic and electrical anisotropies as a potential indicator of movement sense for ductile shear zones in the upper mantle. *Geology* 24, 1033-1036.
- Jiracek, G. R., 1990. Near-surface and topographic distortions in electromagnetic induction. *Surveys in Geophysics* 11, 163–203.
- Jones, A. G, Chave, A. D., Egbert, G., Auld, D., Bahr, K., 1989. A comparison of techniques for magnetotelluric response function estimation. *J. Geophys. Res.* 94, 14,201-14,213.
- Jones, A. G., 1986. Parkinson's pointers' potential perfidy! *Geophys. J. Royal Astr. Soc.* 87, 1215 – 1224.
- Jones, A. G., 1999. Imaging the continental upper mantle using electromagnetic methods. *Lithos* 48, 57-80.
- Jones, A. G., 2012. Distortion of magnetotelluric data: its identification and removal, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 219-295.
- Jones, A. G., 2013. Imaging and observing the Electric Moho. *Tectonophysics* (In press), doi: 10.1016/j.tecto.2013.02.025
- Jones, A. G., Craven, J. A., 1990. The North American Central Plains conductivity anomaly and its correlation with gravity, magnetic, seismic, and heat flow data in Saskatchewan, Canada. *Phys. Earth Planet. Int.* 60, 169 - 194.
- Jones, A. G., Craven, J. A., 2004. Area selection for diamond exploration using deep-probing electromagnetic surveying. *Lithos* 77, 765-782.
- Jones, A. G., Evans, R. L., Muller, M. R., Hamilton, M. P., Miensopust, M. P., Garcia, X., Cole, P., Ngwisany, T., Hutchins, D., Fourie, C. J. S., Jelsma, H., Evans, S., Aravanis, T., Pettit, W., Webb, S., Wasborg, J., The SAMTEX, 2009. Area selection for diamonds using magnetotellurics: Examples from southern Africa. *Lithos* 112, 83-92.
- Jones, A. G., Ferguson, I. J., 2001. The electric Moho. *Nature* 409, 331-333.
- Jones, A. G., Ledo, J., Ferguson, I. J., Craven, J. A., Unsworth, M. J., Chouteau, M. and Spratt, J. E. 2013. The electrical resistivity of Canada's lithosphere and correlation with other parameters: Contributions from LITHOPROBE and other programmes. *Can. J. Earth Sci.* submitted, August 2013.

- Jones, A. G., Ledo, J., Ferguson, I. J., Farquharson, C., Garcia, X., Grant, N., McNeice, G., Roberts, B., Spratt, J., Wennberg, G., Wolyneć, L., Wu, X., 2005b. The electrical resistivity structure of Archean to Tertiary lithosphere along 3200 km of SNORCLE profiles, northwestern Canada. *Can. J. Earth Sci.* 42, 1257-1275.
- Jones, A.G., Craven, J.A., McNeice, G.W., Ferguson, I.J., Boyce, T.T., Farquharson, C., Ellis, R., 1993. North American Central Plains conductivity anomaly within the Trans-hudson orogen in northern Saskatchewan, Canada. *Geology* 21, 1027–1030.
- Jones, A.G., Evans, R.L., Eaton, D.W., 2009a. Velocity–conductivity relationships for mantle mineral assemblages in Archean cratonic lithosphere based on a review of laboratory data and Hashin–Shtrikman extremal bounds. *Lithos* 131–143.
- Jones, A.G., Garcia, X., 2003. The Okak Bay MT dataset case study: a lesson in dimensionality and scale. *Geophysics* 68, 70–91.
- Jones, A.G., Gough, D.I., 1995. Electromagnetic studies in southern and central Canadian Cordillera. *Can. J. Earth Sci.* 32, 1541–1563.
- Jones, A.G., Ledo, J., and Ferguson, I.J. 2005a. Electromagnetic images of the Trans-Hudson orogen: the North American Central Plains (NACP) anomaly revealed. *Can. J. Earth Sci.* 42, 457–478.
- Jones, A.G., Lezaeta, P., Ferguson, I.J., Chave, A.D., Evans, R.L., Garcia, X., Spratt, J., 2003. The electrical structure of the Slave craton. *Lithos* 71, 505–527.
- Jones, A.G., Savage, P. J., 1986. North American Central Plains conductivity anomaly goes east. *Geophys. Res. Lett.* 13, 685–688.
- Jones, F. W., Price, A. T., 1970. The perturbations of alternating geomagnetic fields by conductivity anomalies. *Geophys. J. R. Astr. Soc.* 20, 317 – 334.
- Jones, K.A., Ingham, M.R., Bibby, H.M., 2008. The hydrothermal vent system of Mount Ruapehu, New Zealand—a high frequency MT survey of the summit plateau. *J. Volcanol. Geotherm. Res.* 176, 591–600.
- Jupp, D.L.B., VOZOFF, K., 1975. Stable iterative methods for inversion of geophysical data. *Geophys. J. R. Astron. Soc.* 42, 957-976
- Karato, S., 1990. The role of hydrogen in the electrical conductivity of the upper mantle. *Nature* 347, 272–273.
- Katsube, T.J., Mareschal, M., 1993. Petrophysical model of deep electrical conductors: graphite lining as a source and its disconnection due to uplift. *J. Geophys. Res.* 98, 8019 - 8083.
- Kaufman, A. A., Keller, G., 1981. *The Magnetotelluric Sounding Method*. Elsevier, Amsterdam, 595 pp.
- Kellett, R.L., Mareschal, M., Kurtz, R.D., 1992. A model of lower crustal electrical anisotropy for the Pontiac Subprovince of the Canadian Shield, *Geophys. J. Int.* 111, 141-150.
- Key, K., 2012. Marine electromagnetic studies of seafloor resources and tectonics. *Surv. Geophys.* 33, 135-167. doi:10.1007/s10712-011-9139-x
- Korja, T., 2007. How is the European Lithosphere Imaged by Magnetotellurics? *Surv. Geophys.* 28, 239 – 272.
- Korja, T., Tuisku, P., Pernu, T., Karhu, J., 1996. Lapland Granulite Belt– implications for properties and evolution of deep continental crust. *Terra Nova* 8, 48–58.

- Kurtz, R.D., Craven, J.A., Niblett, E.R., Stevens, R.A., 1993. The conductivity of the crust and mantle beneath the Kapuskasing Uplift: electrical anisotropy in the upper mantle. *Geophys. J. Int.* 113, 483–498.
- Le Pape, F., Jones A.G., Vozar, J., Wei, W., 2012. Penetration of crustal melt beyond the Kunlun Fault into northern Tibet. *Nature Geoscience* 5, 330–335. doi:10.1038/NGEO1449.
- Ledo, J., Queralt, P., Marti, A., Jones, A.G., 2002. Two-dimensional interpretation of three-dimensional magnetotelluric data: an example of limitations and resolution. *Geophys. J. Int.* 150, 127–139.
- Li, S., Unsworth, M. J., Booker, J. B., Wei, W., Tan, H., Jones, A. G., 2003. Partial melt or aqueous fluid in the mid-crust of Southern Tibet? Constraints from INDEPTH Magnetotelluric data. *Geophysics* 153, 289–304.
- Mackie, R. L., Madden, T. R., 1993. Three-dimensional magnetotelluric inversion using conjugate gradients. *Geophys. J. Int.* 115, 215 – 229.
- Mackie, R. L., Madden, T. R., Wannamaker, P. E., 1993. Three-dimensional magnetotelluric modeling using difference equations - Theory and comparisons to integral equation solutions. *Geophysics* 58, 215 – 226.
- Mackie, R., Watts, M.D., 2007. Joint 3D inversion of marine CSEM and MT data, SEG, San Antonio 2007 annual meeting, pp 574–578
- Mackie, R.L., B. R. Bennett and T. R. Madden, 1988. Long-period magnetotelluric measurements near the central California coast: A land-locked view of the conductivity structure under the Pazific Ocean. *Geophys. J. Int.* 95, 181 – 194.
- Mackwell, S. J., Kohlstedt, D. L., 1990. Diffusion of Hydrogen in Olivine: Implications for Water in the Mantle. *J. Geophys. Res.* 95, 5079–5088.
- Madden, T., Nelson, P., 1964. A defense of Cagniard's magnetotelluric method. *Geophys. Lab. ONR proj. NR-371-401*, MIT, Cambridge, Massachusetts. Reprinted 1986 in: Vozoff, K., (Ed), *Magnetotelluric Methods*, Tusla, OK, Soc. Expl. Geophys, pp 89 -95.
- Mareschal, M., Kellett, R. L., Kurtz, R. D., Ludden, J. N., Ji, S., Bailey, R. C., 1995. Archean cratonic roots mantle shear zones and deep electrical anisotropy. *Nature London* 375, 134–137.
- Mareschal, M., Kurtz, R. D., Bailey, R. C., 1994. A review of electromagnetic investigations in the Kapuskasing uplift and surrounding regions: electrical properties of key rocks. *Can. J. Earth Sci.* 31, 1042–1051.
- Marquis, G., Hyndman, R.D., 1992. Geophysical support for aqueous fluids in the deep crust: seismic and electrical relationships. *Geophys. J. Int.* 110, 91–105.
- Marti, A., 2013. The role of electrical anisotropy in magnetotelluric responses: From modelling and dimesionality analysis to inversion and interpretation. *Surv. Geophys.*, doi: 10.1007/s10712-013-9233-3.
- McKirdy D. McA., Weaver, J. T., 1984. Induction in a thin sheet of variable conductance at the surface of a stratified earth - I. Two-dimensional theory. *Geophys. J. R. Astr. Soc.* 78, 93– 103.
- McKirdy D. McA., Weaver, J. T., 1985, Induction in a thin sheet of variable conductance at the surface of a stratified earth - II. Three-dimensional theory. *Geophys. J. R. Astr. Soc.* 80, 177 - 194.

- McNeice, G. W., Jones, A. G., 2001. Multisite, multifrequency tensor decomposition of magnetotelluric data. *Geophysics* 56, 158-173.
- McPherron, R.L., 2005. Magnetic pulsations: their sources and relation to solar wind and geomagnetic activity. *Surv. Geophys.* 26, 545–592.
- Meju, M., 2002. Geoelectromagnetic exploration for natural resources: models, case studies and challenges. *Surv. Geophys.* 23, 133– 206.
- Mibe, K., Fujii, T., Yasuda, A., 1998. Connectivity of aqueous fluid in the Earth's upper mantle. *Geophys. Res. Lett.* 25, 1233–1236.
- Miensopust, M. P., 2010. Multidimensional Magnetotellurics, a 2d case study and a 3d approach to simultaneously invert for resistivity structure and distortion parameters. Unpublished PhD thesis, Dublin Institute for Advanced Studies, Dublin, Ireland. 353p.
- Miensopust, M. P., Jones, A. G., 2011. Artefacts of isotropic inversion applied to magnetotelluric data from an anisotropic Earth. *Geophys. J. Int.* 187, 677-689.
- Miensopust, M. P., Jones, A. G., Muller, M. R., Garcia, X., Evans, R. L., 2011. Lithospheric structures and Precambrian terrane boundaries in northeastern Botswana revealed through magnetotelluric profiling as part of the Southern African Magnetotelluric Experiment. *J. Geophys. Res.* 116, B02401.
- Miensopust, M.P., Queralt, P., Jones, A.G., the 3D MT modellers, 2013. Magnetotelluric 3-D inversion—a review of two successful workshops on forward and inversion code testing and comparison. *Geophys. J. Int.* 193, 1261-1238. doi: 10.1093/gji/ggt066
- Mitsuhata, Y., Uchida, T., 2004. 3D magnetotelluric modeling using the T- finite-element method. *Geophysics* 69, 108 – 119.
- Muller, M. R., Jones, A. G., Evans, R. L., Gru¨tter, H. S., Hatton, C., Garcia, X., Hamilton, M. P., Miensopust, M. P., Cole, P., Ngwisanyi, T., Hutchins, D., Fourie, C. J., Jelsma, H. A., Evans, S. F., Aravanis, T., Pettit, W., Webb, S. J., Wasborg, J., The SAMTEX Team, 2009. Lithospheric structure, evolution and diamond prospectivity of the Rehoboth Terrane and the western Kaapvaal Craton, southern Africa: Constraints from broadband magnetotellurics. *Lithos* 112S, 93-105.
- Nam, M. J., Kim, H. J., Song, Y., Lee, T. J., Son, J. –S., Suh, J. H., 2007. 3D magnetotelluric modelling including surface topography. *Geophysical Prospecting* 55, 277 – 287.
- Narod, B.B., Bennest, J. R., 1990. Ring-core fluxgate magnetometers for use as observatory variometers. *Phys. Earth Planet. Int.* 59, 23-28.
- Nichols, E., Morrison, H. F., Clarke, J., 1988. Signals and noise in magnetotellurics. *J. Geophys. Res.* 93B, 13743–13154.
- Nover, G., Heikamp, S., Meurer, H. J., Freund, D., 1998. In-situ electrical conductivity and permeability of mid-crustal rocks from the KTB drilling: consequences for high conductive layers in the earth crust. *Surv. Geophys.* 19, 73 – 85.
- Oldenburg, D.W., Ellis, R.G., 1993. Efficient inversion of Magnetotelluric data in two dimensions. *Phys. Earth Planet. Int.* 81, 177-200.
- Padilha, A.L, Vitorello, I., Pádua, M. B., Bologna, M. S., 2006. Lithospheric and sublithospheric anisotropy beneath central-southeastern Brazil constrained by long period magnetotelluric data. *Phys. Earth Planet Int.* 158,190-209.

- Palacky, G.J., 1988. Resistivity characteristics of geological targets, in: Nabighian, M.N. (Ed), *Electromagnetic Methods in Applied Geophysics - vol. I Theory*, Society of Exploration Geophysicists, Tulsa, Ok, pp. 56- 130.
- Parker, R. L., 1980. The inverse problem of electromagnetic induction: existence and construction of solutions based on incomplete data. *J. Geophys. Res.* 85, 4421-4425.
- Parker, R. L., Booker, J. R., 1996. Optimal one-dimensional inversion and bounding of magnetotelluric apparent resistivity and phase measurements. *Phys. Earth Planet. Int.* 98, 269-282.
- Parkinson, W. D., 1959. Directions of rapid geomagnetic fluctuations. *Geophysical J.R. Astr. Soc.* 2, 1 – 14.
- Patro, P. K., Egbert, G.D., 2008. Regional conductivity structure of Cascadia: preliminary results from 3D inversion of USArray transportable array Magnetotelluric data. *Geophys. Res. Lett.* 35, art. no. L20311
- Pek, J., Santos, F. A. M., 2002. Magnetotelluric impedances and parametric sensitivities for 1-D anisotropic layered media. *Computers and Geosciences* 28, 939 – 950.
- Pek, J., Santos, F. A. M., 2006. Magnetotelluric inversion for anisotropic conductivity in layered media. *Phys. Earth Planet. Int.* 158, 139-158.
- Pek, J., Verner, T., 1997. Finite-difference modelling of magnetotelluric fields in two dimensional anisotropic media. *Geophys. J. Int.* 128, 505 – 521.
- Petiau, G., Dupis, A., 1980. Noise, temperature coefficient, and long time stability of electrodes for telluric observations. *Geophysical Prospecting* 28, 792-804.
- Rodi, W. L., Mackie, R. L., 2001. Nonlinear conjugate gradients algorithm for 2D magnetotelluric inversion. *Geophysics* 66, 174-187.
- Rodi, W. L., Mackie, R. L., 2012. The inverse problem, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 347-414.
- Rokityansky, I. I., 1982. *Geoelectromagnetic investigation of the Earth's crust and mantle*, Springer-Verlag, Berlin, 381 pp.
- Salazar-Palma, M., Sarkar, T. K., Garcia-Castille, L., Roy, T., Djordjevic, A., 1998. Iterative and self-adaptive finite-elements in electromagnetic modelling. *Artech House*.
- Sasaki, Y., 2001. Full 3-D inversion of electromagnetic data on PC. *Journal of Applied Geophysics* 46, 45 – 54.
- Sasaki, Y., Meju, M.A., 2006. Three-dimensional joint inversion for Magnetotelluric resistivity and static shift distributions in complex media. *J. Geophys. Res.* 111:art. no. B05101
- Schock, R. N., Duba, A. G., Shankland, T. J., 1989. Electrical conduction in olivine. *J. Geophys. Res.* 94, 5829 – 5839
- Schultz, A., Kurtz, R.D., Chave, A.D., Jones, A.G., 1993. Conductivity discontinuities in the upper mantle beneath a stable craton. *Geophys. Res. Lett.* 20, 2941–2944.
- Schwarz, G., 1990. Electrical conductivity of the earth's crust and upper mantle. *Surv. Geophys.* 11, 133 – 161.
- Selway, K. 2014. On the causes of electrical conductivity anomalies in tectonically stable lithosphere, *Surv. Geophys.*, 57, 219-257, doi 10.1007/s10712-013-9235-1

- Sheriff, R.E., 1999. Encyclopedic Dictionary of Exploration Geophysics. 3rd edn. Geophysical References Series 1, in: Society of Exploration Geophysicists.
- Simpson, F., Bahr, K., 2005. Practical Magnetotellurics. Cambridge University Press, Cambridge, United Kingdom, 254 pp
- Sims, W. E., Bostick Jr., F. X., Smith, H. W., 1971. The estimation of magnetotelluric impedance tensor elements from measured data. *Geophysics* 36, 938 – 942.
- Siripunvaraporn, W., 2012. Three-Dimensional Magnetotelluric Inversion: An Introductory Guide for Developers and Users. *Surv. Geophys.* 33, 5–27. Doi: 10.1007/s10712-011-9122-6
- Siripunvaraporn, W., Egbert, G., 2000. An efficient data-subspace inversion method for 2-D magnetotelluric data, *Geophysics* 65, 791 – 803.
- Siripunvaraporn, W., Egbert, G., 2009. WSINV3DMT: vertical magnetic field transfer function inversion and parallel implementation. *Phys. Earth planet. Inter.* 173, 317–329.
- Siripunvaraporn, W., Egbert, G., Lenbury, Y., Uyeshima, M., 2005. Three-dimensional magnetotelluric inversion: data-space method. *Phys. Earth Planet. Int.* 150, 3–14.
- Smith, J. T., 1997. Estimating galvanic-distortion magnetic fields in magnetotellurics. *Geophys. J. Int.* 130, 65 – 72.
- Smith, J. T., Booker, J. R., 1988. Magnetotelluric inversion for minimum structure. *Geophysics* 53, 1565 – 1576.
- Smith, J. T., Booker, J. R., 1991. Rapid nversion of two- and three-dimensional magnetotelluric data. *J. Geophys. Res.* 96, 3905 – 3922.
- Spichak, V., Manzella, A., 2009. Electromagnetic sounding of geothermal zones. *Journal of Applied Geophysics* 68, 459-478.
- Spratt, J. E., Jones, A. G., Jackson, V.A., Collins, L., Avdeeva, A., 2009. Lithospheric geometry of the Wopmay orogen from a Slave craton to Bear Province magnetotelluric transect. *J. Geophys. Res.* 114, B01101.
- Stratton, J.A., 1941. *Electromagnetic Theory*: McGraw-Hill Book Company Inc., 631 p.
- Telford, W. M., Geldart, L. P., Sheriff, R. E., 1990. *Applied Geophysics*. Cambridge University Press, Cambridge, UK, 770p
- Tikhonov, A. N., 1950. The determination of the electrical properties of the deep layers of the Earth's crust. *Dokl. Acad. Nauk. SSR*, 73, pp. 295 – 297. (in Russian).
- Tuncer, V., Unsworth, M.J., Siripunvaraporn, W., Craven, J. A., 2006. Exploration for unconformity-type uranium deposits with audiomagnetotelluric data: a case study from the McArthur River Mine, Saskatchewan, Canada. *Geophysics* 71, B201–B209.
- Uchida, T., 1993. Smooth 2-D inversion for Magnetotelluric data based on statistical criterion ABIC. *J. Geomag. Geoelec.* 45, 841-858.
- Viljanen, A., 2012. Description of the magnetospheric/ionospheric sources, in: Chave, A. D., Jones, A. G. (Eds.), 2012. *The Magnetotelluric Method - Theory and Practice*, Cambridge University Press, New York, pp. 96-121.
- Vozoff, K., 1991. The magnetotelluric method, in: Nabighian, M.N. (Ed.), *Electromagnetic Methods in Applied Geophysics*, Vol. 2 Applications, Society of Exploration Geophysicists, Tulsa, 641-711.
- Wait, J.R., 1954. On the relation between telluric currents and the Earth's magnetic field: *Geophysics* 19, 281-289.

- Wangness, R.K., 1986. *Electromagnetic Fields*, John Wiley and Sons, New York, 587 pp.
- Wannamaker P. E., 1999. Affordable magnetotellurics: interpretation in natural environments, in: Oristaglio, M., Spies, B. (Eds.), *Three-dimensional Electromagnetics*, Geophysical development series vol. 7, Society of Exploration Geophysicists, 349–379.
- Wannamaker, P. E., 1991. Advances in three-dimensional magnetotelluric modeling using integral equations. *Geophysics* 56, 1716 – 1728
- Wannamaker, P. E., 2000. Anisotropy versus heterogeneity in continental solid earth electromagnetic studies: fundamental response characteristics and implications for physicochemical state. *Surv. Geophys.* 26, 733 – 765.
- Wannamaker, P. E., Caldwell, T. G., Doerner, W. M., Jiracek, G. R., 2004. Fault zone fluids and seismicity in compressional and extensional environments from electrical conductivity: the New Zealand Southern Alps and U.S. Great Basin. *Earth Planets Space* 56, 1171 – 1176.
- Wannamaker, P. E., Hasterok, D. P., Johnston, J.M., Stodt, J.A., Hall, D.B., Sodergren, T.L., Pellerin, L., Maris, V., Doerner, W.M., Groenewold, K.A., Unsworth, M.J., 2008. Lithospheric dismemberment and magmatic processes of the Great Basin–Colorado Plateau transition, Utah, implied from magnetotellurics. *Geochem. Geophys. Geosyst.* 9, Q05019. doi:10.1029/2007GC001886
- Wannamaker, P. E., Hohmann, G.W., Ward, S. H., 1984b. Magnetotelluric responses of 3-D bodies in layered earths using integral equations. *Geophysics* 49, 1517 – 1533.
- Wannamaker, P.E., 2005. Anisotropy versus heterogeneity in continental solid earth electromagnetic studies: fundamental response characteristics and implications for physicochemical state. *Surv. Geophys.* 26, 733 – 765.
- Wannamaker, P.E., Hohmann, G. W., SanFilipo, W. A., 1984a. Electromagnetic modeling of three-dimensional bodies in layered earths using integral equations. *Geophysics* 49, 60 - 74.
- Wannamaker, P.E., Stodt, J.A., Rijo, L., 1987. A stable finite element solution for two dimensional magnetotelluric modelling. *Geophys. J. R. Astr. Soc.* 88, 277-296.
- Ward, S.H, Hohmann, G. W., 1988. *Electromagnetic Theory for Geophysical Applications*, in: Nabighian, M.N. (Ed.), *Electromagnetic Methods in Applied Geophysics*, vol. I, Theory, Society of Exploration Geophysicists, Tulsa, Ok, 131-311.
- Weidelt, P., 1999. Three-dimensional conductivity models: implications for electrical anisotropy, in: Oristaglio, M., Spies, B. (Eds.), *Three-dimensional Electromagnetics*, Geophysical development series vol. 7, Society of Exploration Geophysicists, 119–137.
- Weidelt, P., Chave, A. D., 2012. The magnetotelluric response function, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 122-162.
- West, G.F., Edwards, R. N., 1985. A simple parametric analysis for the electromagnetic response of an anomalous body in a host medium. *Geophysics* 50, 2542 - 2557.
- Whittall, K.P., Oldenburg, D. W., 1992. *Inversion of Magnetotelluric Data for a one-Dimensional Conductivity*, Geophysical Monograph Series no. 5, Society of Exploration Geophysicists, Tulsa.

- Wiese, H., 1962. Geomagnetische Tiefentellurik. II. Die Streichrichtung der Untergrundstrukturen des elektrischen Widerstandes, erschlossen aus geomagnetischen Variationen. *Geofis. Pura Appl.*, 52, pp. 83 – 103. (in German).
- Xiong, Z., 1992. Electromagnetic modeling of 3-D structures by the method of system iteration using integral equations. *Geophysics* 57, 1556 – 1561.
- Yardley, B., Valley, J., 1997. The petrologic case for a dry lower crust. *J. Geophys. Res.* 102, 12,173 -12,185.
- Yoshino, T., Manthilake, G., Matsuzaki, T., Katsura, T., 2008. Dry mantle transition zone inferred from the conductivity of wadsleyite and ringwoodite. *Nature* 451, 326 – 329.

Chapter 3

Geology, Geophysics and Tectonics

3.1 Introduction

The geologic history of Southern Ontario dates back to the Archean. The rocks present in this region represent remnants of ocean floor, ancient mountain chains, and deep sea sediments as well as ancient glacial deposits. These rocks are divided into three geologic provinces which are overlain by Paleozoic sedimentary cover. Southern Ontario's portion of the Canadian Shield (Fig. 1.1) is made up of the Superior, Southern and Grenville provinces. These provinces were amalgamated at different times during the Precambrian. The Paleozoic sedimentary cover is characterized by the sedimentary rocks of the Michigan and Allegheny basins. Additional information on the geology in the study area is provided in the introductory sections of chapters 5, 6 and 7.

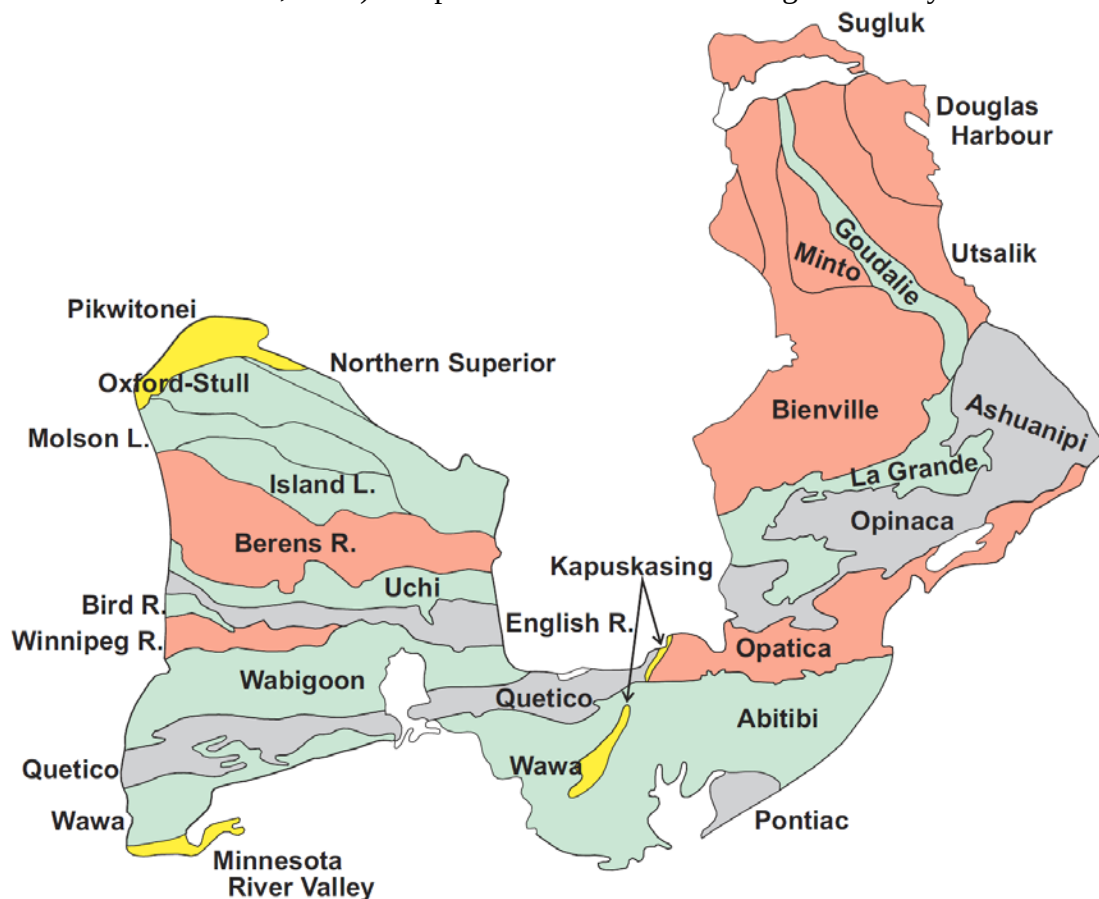
3.2 The Superior Province

The Superior Province (Fig. 3.1) is the largest Archean craton on Earth (Card and Ciesielski, 1986; Hoffman, 1989; Ludden and Hynes, 2000). This province has been stable since the end of a major Late Archean orogeny and is a remnant of a previously more extensive craton that is surrounded and truncated by Proterozoic orogens. Its boundary with the Grenville Province is fault-bounded, while its contact with the Penokean orogen is a zone where Paleoproterozoic supracrustal sequences overly and were thrust upon the Superior craton (Fig. 1.1; Card and Poulsen, 1998).

The Superior Province mainly comprises rocks that formed between 2.8 and 2.7 Ga and small areas of pre- 3.5 Ga crust (Hoffman, 1989). It is composed of about 75 percent gneissic rocks and 25 percent low to medium-grade supracrustal rocks that form distinct sub-parallel east-northeast trending belts of contrasting lithology, age, and/or

metamorphic grade. The belt types that are common in this province include granite-greenstone, metasedimentary belts, plutonic complexes, and high-grade gneiss complexes (Card and Ciesielski, 1986; Percival, 1989; Card, 1990). Boundaries separating subprovinces (Fig. 3.1) are usually zones of structural and metamorphic contrast, many of which are marked by east-west faults that extend throughout much of the Superior Province (Hoffman, 1989; Card, 1990; Card and Poulsen, 1998). Structural trends are mainly east-west in the southern Superior Province, northwesterly in the northwest and northerly in the northeast (Fig. 3.1, Card and Poulsen, 1998).

Figure 3.1: The map of the exposed Superior Province and its subprovinces (redrawn from Card and Poulsen, 1998). Shapefile obtained from Geological Survey of Canada.



The southernmost island arc subprovinces are composed mainly of juvenile crust while the northern subprovinces conserve generations of arc assembly dated back to 3.0 Ga (Hoffman, 1989; Card, 1990). The high-grade gneiss subprovinces contain upper amphibolite facies gneiss of supracrustal and plutonic origin and are characterized by positive gravity and magnetic anomalies. The granite-greenstone subprovinces are characterized by low-to-medium grade volcanic-dominated supracrustal sequences. The metasedimentary subprovinces are made up of deformed and metamorphosed turbidite units and anatectic granite bodies (Card and Poulsen, 1998). The subprovinces that are closest to the southeast margin of the Superior Province and the Grenville Front in Southern Ontario are the Pontiac and the Abitibi subprovinces (Fig. 3.2).

3.2.1 Abitibi subprovince

The Abitibi subprovince is located in the southeastern part of the Superior province and is the world's largest exposed Archean greenstone terrane (Card and Poulsen, 1998). The Abitibi greenstone belt (Ludden et al., 1986; Corfu, 1993) is bounded on the north by the Opatica and on the south by the Pontiac subprovinces (Fig. 3.1). On the east and southeast, the belt is bordered by the Grenville Province. To the west, it is separated from the Quetico and Wawa subprovinces by the high-grade metamorphic rocks of the Kapuskasing structural zone (Percival, 1989). The Abitibi subprovince represents a remnant from a larger Neoproterozoic granite-greenstone terrane which has a record of rapid crustal growth (Card, 1990; Percival et al., 2004). It is characterized by polycyclic volcanic rocks overlain by a late-stage sequence of turbiditic sedimentary rocks and finally synorogenic clastic rocks (Card and Poulsen, 1998).

The Abitibi subprovince is divided into three zones based on lithologic variations and age differences marked by major faults or shear zones. The northern zone is characterized by massive basaltic plain sequences, large synvolcanic gabbro-anorthosite and mafic-ultramafic intrusions as well as diverse multicyclic mafic-felsic sequences. The central zone is dominated by plutonic rocks with linear fault-banded belts of basalt and volcanogenic clastic sedimentary rocks. The southern domain has massive komatiitic rocks, part of thick mafic plain sequences and prevalent multicyclic mafic-felsic volcanic complexes (Dimroth et al., 1982; Ludden et al., 1986). Structurally the northern zone thrusts over the Opatica subprovince and the southern zone thrusts over Pontiac subprovince (Dimroth et al., 1983).

3.2.2 Pontiac subprovince

The Pontiac subprovince is an Archean metasedimentary and plutonic terrane (Card and Cieselski 1986). It is separated from the Abitibi greenstone belt by the Cadillac Larder-Lake Fault Zone (CLLFZ) in the north, and in the south its boundary with the Grenville Province is marked by the Grenville Front Tectonic Zone (GFTZ; Fig. 3.2). The Pontiac subprovince is characterized by structural patterns dominated by southward-verging thrusts and nappes as well as low-grade metamorphism (Card and Poulsen, 1998).

The Pontiac is subdivided into a northern part, dominated by an east-west band of low-to-medium grade turbiditic metasedimentary rocks of the Pontiac Group; a central complex of granitic intrusions, pegmatite units and paragneiss; and a southern region with low-grade volcanic and sedimentary rocks (metagreywackes) (Ludden et al. 1986;

Rive et al., 1990). The sedimentary rocks of the Pontiac group consist of turbiditic wacke, pelite and minor conglomerate, low-grade iron-formation and graphitic schist (Davis 1992; Mortensen and Card 1993; Card and Poulsen, 1998). The plutonic rocks in this province have been grouped into early, pre- to syntectonic plutons of tonalitic composition and late, syn- to post-tectonic intrusions of monzonitic and granitic composition (Rive et al., 1990; Feng et al., 1993).

3.3 The Southern Province

The Southern Province (Fig. 3.2) evolved during Proterozoic times. Only a small part of this province lies within Canada, as most of it occurs in the states of Michigan, Minnesota and Wisconsin (Bennett et al., 1991). This province consists of Paleoproterozoic metasedimentary and metavolcanic rocks of the Huronian Supergroup, mafic intrusions, Nipissing diabase suite, the Sudbury Igneous Complex, and the Whitewater Group. Also included in the Southern Province are Mesoproterozoic plutonic and minor volcanic rocks of the Killarney Magmatic Belt, which is cut by the Grenville Front causing these rocks to be recognized in both the Southern and Grenville provinces (Easton, 1992).

The Huronian Supergroup (Cobalt Embayment and Penokean Fold Belt) unconformably overlies 2.8 - 2.5 Ga Archean plutonic and supracrustal rocks of the Superior Province (Bennett et al., 1991). The supergroup was affected by a poorly documented, <2.2 Ga deformation and the 1.8 Ga compressional Penokean Orogeny (Hoffman, 1989; Thurston, 1991). The 1.85 Ga Sudbury Igneous Complex (SIC) and associated Whitewater Group sedimentary rocks lie along the northern margin of the

Penokean Fold Belt, east of Lake Huron. The origin of the SIC and the associated Whitewater Group breccias and sedimentary sequence is controversial. Some support a meteorite impact origin while others advocate an endogenetic petrogenetic model (Pye et al., 1984).

3.4 The Grenville Province

The Grenville Province is part of the Middle to Late Mesoproterozoic Grenville orogen, extending across the North America from Mexico to Labrador. The scale of this orogen is similar to the present front Himalayas front and formed between ca 1.2 – 1.0 Ga (Hanmer, 1988; Easton, 1992; Rivers, 1997; Carr et al., 2000; Rivers et al., 2012). The province is composed of reworked polycyclic Archean, Paleo- and Meso-Proterozoic rocks derived from Laurentia (proto-North American landmass) as well as monocyclic rocks accreted just prior to the Grenvillian orogeny. The Grenville Province is separated from older blocks units such as the Superior and Southern provinces, the Killarney Magmatic Belt, the Eastern Granite-Rhyolite province and the Midcontinent Rift by the Grenville Front (Williams et al., 1992). The province is largely made up of high grade metamorphic rocks (Hoffman, 1989; Rivers et al., 1993; Carr et al., 2000; Rivers et al. 2012).

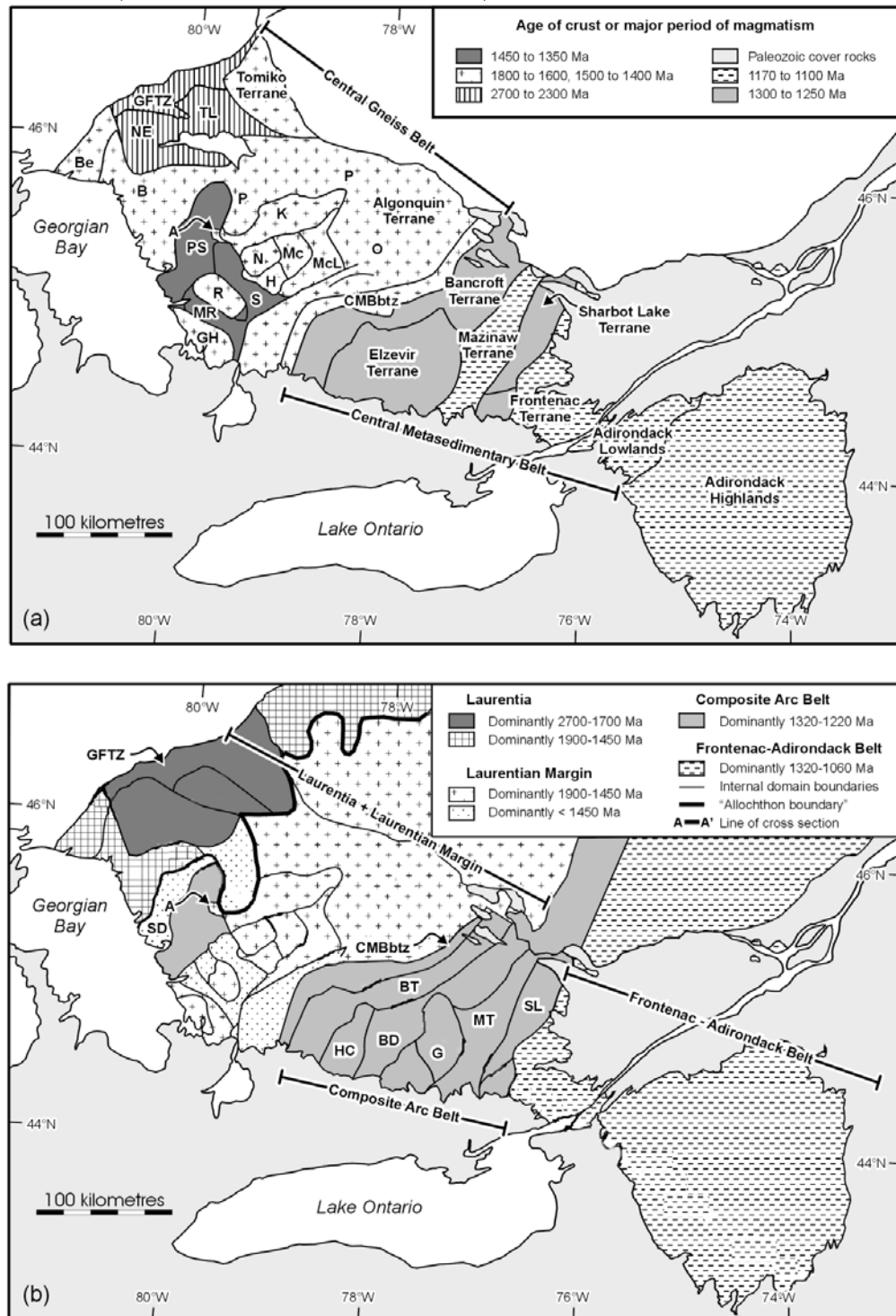
In southern Ontario, the Grenville Province (Fig. 3.2) is bounded by the GFTZ to the northwest, where it overlies the Southern and Superior provinces. It is subdivided into the Central Gneiss Belt (CGB), the Central Metasedimentary Belt (CMB) (Fig. 3.2a; Wynne-Edwards, 1972; Easton. 1992). Carr et al. (2000) subdivide this region into CGB, Composite Arc Belt (CAB), and Frontenac-Adirondack Belt (FAB; Fig. 3.2b). in this

thesis, the terminologies of Carr et al. (2000) will be used to describe the tectonic evolution of the study area. As shown in Figure 3.2, the tectonic contact between the CGB and CMB is a NE-SW trending set of ductile shear zones called the Central Metasedimentary Belt Boundary Zone (CMBBZ; Wynne-Edwards, 1972; Easton, 1992).

3.4.1 The Central Gneiss Belt

The CGB represents part of the reworked margin of the pre-Grenvillian Laurentian craton. It is characterized by felsic gneisses and amphibolites with upper amphibolite to granulite facies plutonic rocks as well as related supracrustal rocks (Easton, 1992). Easton (1986) reported that a variety of Archean to Mesoproterozoic crustal assemblages within the CGB have been affected by the Grenvillian orogeny. Evidence for high grade regional metamorphism at 1.4 Ga with the onset of syn-metamorphic thrusting between 1.6 and 1.12 Ga has also been reported (Tucillo et al., 1992). Ductile shear zones that separate distinctive lithotectonic domains and sub-domains, in the CGB, indicate northwest-directed, low-angle thrusting (Davidson, 1986). Studies have subdivided the CGB into a four domains, Tomiko, Nipissing, Algonquin and Parry Sound (Fig. 3.2) based on their contrasting lithologies, metamorphic grade and deformation histories. These domains represent imbricated panels of relatively low strain rocks bounded by anastomosing ductile shear zones (e.g., Davidson and Morgan, 1981; Davidson et al., 1982; Culshaw et al., 1983; Hanmer and Ciesielski, 1984).

Figure 3.2. Subdivisions the Grenville Provinces by (a) Wyne-Edwards (1972) and Easton (1992) and (b) Carr et al., 2000. . GF=Grenville Front, GFTZ=Grenville Front Tectonic Zone (modified from Carr et al., 2000).



The four domains within the CGB consist of rocks of three different ages (Easton, 1986). The Nipissing domain is made up of Archean and Paleoproterozoic gneisses intruded by Mesoproterozoic plutonic (mainly granitic and monzonitic) rocks. The Algonquin and Tomiko domains consist of Mesoproterozoic gneisses invaded by plutonic rocks that may represent an extension of the eastern granite-rhyolite province across the Grenville Front. The Tomiko terrane is allochthonous to the adjacent Nipissing terrane and the Algonquin domain is probably parautochthonous (Rivers et al., 1989). The Parry Sound domain consists of Mesoproterozoic mafic to intermediate rocks derived from the mantle during the Mesoproterozoic and therefore consists entirely of juvenile crust (Dickin and McNutt 1989; Easton, 1992).

3.4.2 The Central Metasedimentary Belt

The CMB lies to the southeast of the CGB and both are separated by the CMBBZ (Fig. 3.2). Kinematic indicators in this zone indicate that thrusting of the CMB occurred northwestwardly over the CGB (Davidson, 1991; Easton, 1992; Davidson, 1998). The CMB represents a major Mesoproterozoic accretion of supracrustal rocks and consists of marble, quartzites and mafic to volcanic rocks metamorphosed at greenschist to granulite facies during the Grenvillian orogeny (Easton, 1992). These rocks were intruded by a variety of gabbroic, tonalitic, granitic and syenitic plutons (Davidson, 1986).

The rocks of the CMB are dominated by polyfolded domains with complex geometry, and are considered to have been at middle to upper crustal levels during the Grenville orogeny (White et al., 2000; Carr et al., 2000). The CMB rocks are younger than those in the CGB, and they predominantly fall in the age range of 1300-1250 Ma,

although ages as young as 1060 Ma have been obtained for plutonic rocks (Corriveau et al., 1990; Easton, 1992; Rivers, 1997; Davidson, 1998). Based on different lithology, metamorphic, deformation and geological histories, the CMB is subdivided into five terranes (Fig. 3.2) - the Bancroft, Elzevir, Mazinaw, Sharbot Lake and Frontenac. Along with the CMBBZ, the CMB constitutes the Allochthonous Monocyclic Belt (Easton, 1992; Davidson, 1998; Rivers et al., 1989, 2012).

The Bancroft terrane, which was metamorphosed to between middle and upper amphibolite facies, is made up largely of deformed marble with minor intercalations of volcanic rocks intruded by nepheline syenites and syenites between 1.27 to 1.22 Ga (Miller, 1983). The terrane is characterized by marble tectonite, syenitic gneiss and minor pelite, quartzite and amphibolite, as well as thrust slices of various types of metaplutonic rocks (Hanmer and McEachern, 1992).

The Elzevir terrane lies to the southeast of the Bancroft terrane, and predominantly consists of mafic volcanic rocks (1.3 and 1.25 Ga). These rocks are associated with calc-alkaline plutons ranging from gabbro through tonalite to monzogranite (Davidson, 1998). This Elzevir terrane is thought to represent a volcanic arc or back-arc basin that was separated from other domains in the metasedimentary belt and tectonically emplaced in its current position (Condie and Moore, 1977). The domain was subjected to plutonism and metamorphism at 1.25 to 1.23 Ga and at 1.13 to 1.07 Ga (Easton, 1992).

The Mazinaw terrane consists of the folded Grenville supergroup overlain by a younger sedimentary succession of metamorphosed conglomerate, quartz sandstone, argillaceous sediment and carbonates of the Flinton Group. These younger sedimentary

rocks are believed to be derived partially from weathering of the Frontenac terrane plutonic and metamorphic rocks (Moore and Tompson, 1980). The Flinton Group is preserved only in narrow synclines and was metamorphosed, along with the underlying supracrustal and plutonic rocks, at 1.03 Ga (Corfu and Easton, 1995). Rocks present in this terrane include marbles, calc-alkalic metavolcanic and clastic metasedimentary rocks, which were deposited on a basaltic-tonalitic basement. The terrane has an intense and prolonged structural and metamorphic history which is similar to that present in the Frontenac terrane (Kinsman and Parrish, 1990).

The Sharbot Lake terrane, located between the Mazinaw and Frontenac terranes, is a lithologically and structurally distinct member of the CMB (Carr et al., 2000). Rocks of this terrane mainly consist of marble and mafic metavolcanics interlayered with minor siliciclastic metasedimentary rocks, intruded by gabbros and granodiorite plutons (Easton, 1992; Easton and Davidson, 1997; Davidson, 1998). Deposition of the carbonate and clastic sediments was coeval with the emplacement of mafic to felsic volcanic rocks and occurred between 1.3 and 1.25 Ga (Carr et al., 2000). Plutonic rocks of the terrane include gabbro, granodiorite, monzonite and granite of the Elzevir and Frontenac suites (Davidson and van Breemen, 2000). The overall metamorphic grade is greenschist- to lower amphibolite-facies (Easton, 1992)

The Frontenac terrane consists of a succession of multiply deformed, relatively low-pressure granulite facies quartzofeldspathic and polytropic gneiss, quartzite and marble intruded by plutonic rocks and subjected to metamorphism at about 1.17 to 1.16 Ga (Wynne-Edwards, 1967; Easton, 1992; Easton and Davidson, 1997). Wynne-Edwards (1967) proposed a stratigraphy consisting of two stacked sets of tripartite units composed

from bottom to top: gneiss, quartzite and marble. This terrane is different from other terranes within the CMB in that it lacks volcanic rocks. It is preserved at granulite facies, in contrast to the lower grade Bancroft and Elzevir terranes to the north, and the Adirondack lowlands to the south (Easton, 1992).

3.4.3 The Grenville Front Tectonic Zone

The GFTZ is a 32 km wide prominent southeast-dipping crustal-scale shear zone of amphibolite-facies rocks. The front has boundaries that are gradational and are deformed by cataclastic zones, which extend into the adjacent Southern and Superior provinces (Fig. 3.2; Easton, 1992). It marks the northwestern boundary of the Grenville orogen and thrusting along this shear zone occurred late in the Grenvillian orogeny, at ca. 990 Ma (Haggart et al., 1993; Krogh, 1994). Grenvillian structures and metamorphism of the GFTZ are superimposed on older protoliths that likely represent lithologies corresponding to Southern and Superior provinces (Lumbers, 1978; Rivers et al., 2012).

Easton (1992) recognized three main segments within the GFTZ in Ontario. In Segment 1, Killarney-Sudbury area, the dominant rock types are orthogneisses and minor paragneisses most likely derived from the Killarney Magmatic Belt. In Segment 2, which stretches from Sudbury to Ottawa River Valley, the predominant rock types are para- and ortho-gneisses, as well as meta-gabbroic and meta-anorthositic bodies. Segment 3 stretches from Ottawa River Valley to the Ottawa River and consists mainly of ortho- and paragneisses (Lumbers, 1975, 1978).

3.4.4 The Central Metasedimentary Belt Boundary Zone

The crustal-scale CMBBZ separates the CGB from the CMB (Fig. 3.2) and is a regionally extensive 7–12 km thick composite shear zone that consists of gneissic tectonites derived from both the CGB and the CMB (Hanmer, 1988; Easton, 1992; Hanmer and McEachern, 1992). This regionally extensive structure has long been recognized as the structure along which the rocks of the CMB were tectonically juxtaposed with those of the CGB (e.g., Davidson et al., 1982; Culshaw et al., 1983; Davidson, 1984).

Lithologically, the CMBBZ consists primarily of grey tonalite gneiss thrust sheets bound by quartzofeldspathic and porphyroclastic gneiss, interleaved with marble melange, aluminous paragneiss and rare meta-anorthosite (Hanmer et al., 1985; Hanmer, 1988; Hanmer and McEachern, 1992). Rheological contrasts between the competent metagabbroic bodies in the upper part of the CMBBZ and the weak ferro-aluminous pelites and marbles occurring locally at the base of the CMBBZ, are interpreted to have played an important mechanical role in the localization and transport of thrust slices in this zone (Hanmer and Ciesielski, 1984; Hanmer et al., 1985; Hanmer and McEachern, 1992).

3.5 The Phanerozoic Sedimentary Rocks

The Phanerozoic sedimentary rocks of southern Ontario were deposited unconformably over the eroded Proterozoic Grenville Province and in two adjacent paleo-sedimentary basins. These basins are the Allegheny and the Michigan basins and

the Algonquin Arch is the basement topographic feature that separates the two basins (Fig. 3.3).

The Michigan Basin is a roughly circular, carbonate-dominated, evaporite-bearing intracratonic basin formed due to craton margin tectonic processes. It subsided rapidly from Cambrian to Silurian time (Sloss, 1988) and contains inward dipping Paleozoic strata and a veneer of Jurassic sedimentary rocks (Fisher et al., 1988). The Allegheny basin is a foreland basin formed by orogenic processes at the craton margin and filled mainly with orogen-derived clastic sediments, carbonates and evaporites. Only the Upper Cambrian to Upper Devonian strata of this basin occur in southern Ontario (Sloss, 1988). Detailed description of the tectonic processes responsible for the formation of Michigan and Allegheny basins have been discussed by Sleep, 1971; Sleep and Snell, 1976; Sleep and Nunn, 1980; Nunn and Sleep, 1984; Klein and Hsui, 1987; and Howell and van der Pluijm, 1990, 1999. Johnson et al. (1992) suggested that the two tectonic processes responsible for the sedimentation patterns within these basins were controlled by the rate of basin subsidence and movement of the arch separating them.

The St. Lawrence Platform originated as a passive continental margin due to continental separation in early Cambrian (Kumarapeli, 1985). The platform extends from Great Lakes of Ontario and the United States to Western Newfoundland and it is covered by Paleozoic and Mesozoic sedimentary rocks underlain by the Precambrian rocks of the Canadian Shield. It is bounded by the Appalachian orogen to the south (Chapman and Putnam, 1984; Sanford, 1993). The rocks present within the platform range from carbonates, shales, and evaporites to sandstones which were deposited during Upper Precambrian to Carboniferous times. These sedimentary deposits dip southwards at low

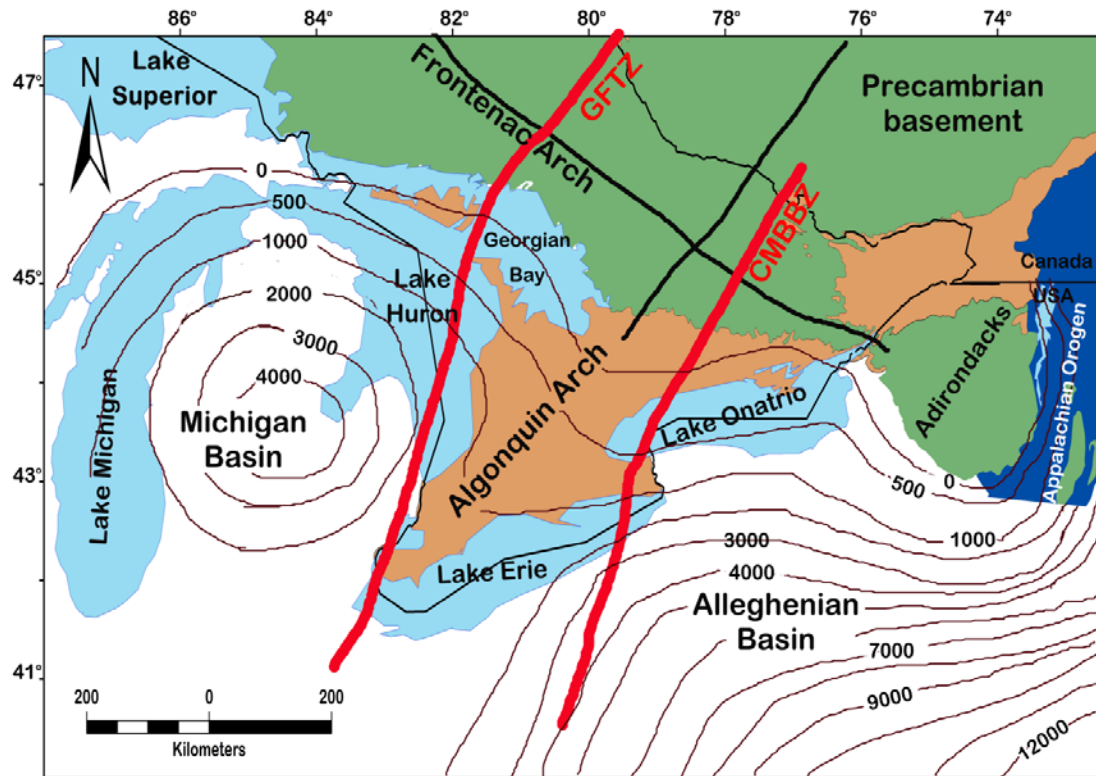
angles into the Michigan Basin or into the Allegheny, Quebec and Anticosti basins that lie parallel to the Appalachian structural front (Sanford et al., 1979; Sanford, 1993). The platform strata are generally undeformed but are locally folded and faulted. However, the rocks around the margins have been deformed by different kinds of Paleozoic orogeny (Sanford, 1993). This platform is subdivided into three segments: the Eastern, Central and Western St. Lawrence Platforms, with only the Western segment present in the study area.

The Algonquin Arch (Fig. 3.3), sometimes referred to as the Western St. Lawrence Platform, is linked with the Paleozoic plate-tectonic events in eastern North America. The stratigraphic succession in this platform is made up of sediments deposited in shallow epicontinental seas (Johnson et al., 1992; Sanford, 1993) with the foreland Allegheny Basin to the south and the intracratonic Michigan Basin to the northwest. The Paleozoic sedimentary rocks in this platform are Cambrian to Devonian (Fig. 3.3) in age and are approximately 1500 – 2000 m thick succession of sandstones, shales, evaporites and carbonates (Fisher et al., 1988; Sanford, 1993). These sedimentary deposits dip at low angles into the Michigan and Allegheny basins, with a thickness of more than 4.5 km at the center of the Michigan Basin, and up to 7 km in the Allegheny basin (Fisher et al., 1988; Sanford, 1993).

Based on regional facies analysis, field mapping, outcrop and borehole data analysis, the Paleozoic stratigraphy within the study area (Fig. 3.4) is described as relatively simple, flat lying and continuous over large distances (Intera, 2009). There is a general agreement about the thickness ranges, lithologies and interpreted facies of this

region (e.g., Johnson et al., 1992; Mazurek, 2004; Armstrong and Carter, 2006; Intera, 2008, 2009; Gartner, 2008).

Figure 3.3. Geological map of the lower Great Lakes region showing the Michigan Basin, the Allegheny Basins, Algonquin Arch, GFTZ and the CMBBZ (redrawn from Mazurek, 2004). Shapefile obtained from Geological Survey of Canada



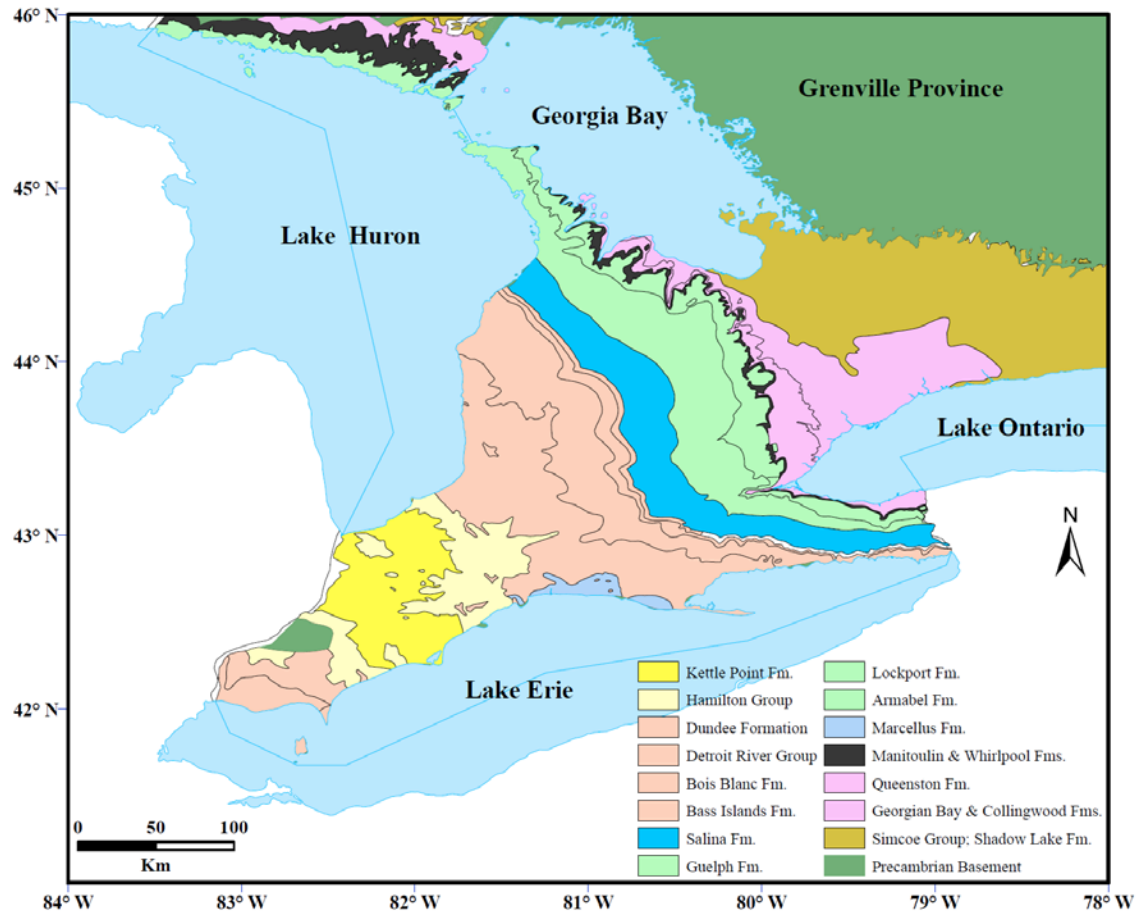
The oldest unit is the Upper Cambrian Shadow Lake Formation which consists mainly of sandstone and dolostones. This formation is unconformably deposited on the altered Precambrian granitic gneiss basement rocks (Mazurek, 2004; Armstrong and Carter 2006; Intera, 2009). The Middle Ordovician carbonates of the Simcoe Group are dominantly limestone and argillaceous limestones (Johnson et al., 1992; Armstrong and

Carter, 2006). Within the study area, the Georgian Bay, Blue Mountain, and Queenston formations constitute the Upper Ordovician units. These units are characterized by blue-grey, non-calcareous shale with minor limestone, sandstone interbeds and red/maroon-green calcareous to non-calcareous shales with limestone interbeds (Johnson et al., 1992; Armstrong and Carter, 2006).

The Lower Silurian is typified with rocks of Cabot Head and Manitoulin formations. Dolostone with minor non-calcareous shale dominates the Cabot Head formation while the Manitoulin Formation is typified by non-calcareous shale with minor dolostone (Armstrong and Carter, 2006; Gartner, 2008). The Middle Silurian carbonate units are represented by rocks of Guelph, Goat Island, Gasport, Rochester, and Reynales formations. Dolostone and fossiliferous dolostone units are predominant carbonate rocks. These units occupy an inter-reef position with respect to the Silurian rocks. The Upper Silurian rocks of Salina and Bass Island formations consist of alternating carbonate, shale and evaporates (Armstrong and Carter, 2006; Gartner, 2008).

The Lower Devonian Bois Blanc Formation is primarily composed of cherty and fossiliferous limestone/dolostone (Johnson et al., 1992). The Middle Devonian Dundee Formation, Detroit River and Hamilton groups are described by Intera (2008) as fossiliferous dolostone, dolostone/limestone with abundant reef building corals and shale units. The Upper Devonian Kettle Point Formation Hamilton group is the youngest layer; it is characterized by black shale (Johnson et al., 1992; Sanford; 1993; Armstrong and Carter, 2006).

Figure 3.4. The depositional cycle of Algonquin Arch showing the sedimentary rock formations and groups present within the lower Great lakes region of southern Ontario. The same colour is used to represent different formations that were deposited during the same period. Shapefile obtained from Geological Survey of Canada.



As shown in Fig. 3.3, the GFTZ and the CMBBZ are traceable from their northeast surface exposure to beneath the sedimentary rocks (Easton, 1992; Forsyth et al., 1994). These sedimentary rocks, in southwestern Ontario, are mainly underlain by rocks of the CGB of the Grenville Province.

3.6 Ottawa-Bonnechere Graben

The Ottawa-Bonnechere Graben (OBG; Fig. 3.2) is a 60 km wide graben that lies along the border between southern Ontario and Quebec. It is bounded by nearly vertical faults trending along an east-west to west-northwest direction (Easton, 1992; Kumarapeli, 1985). The graben has been interpreted as a failed arm of a triple junction of the St. Lawrence rift system. The system was formed during the late Precambrian to early Paleozoic event that led to the formation of the Iapetus ocean. This interpretation is based on geomorphological, topographical, and structural evidence, as well as on the association of intrusive rocks of alkaline affinity (Rankin, 1976; Kumarapeli, 1985; Philpotts, 1978).

The geology of the OBG consists of Paleozoic sedimentary rocks of the Ottawa-St. Lawrence Lowland, inliers and uplifted blocks of Precambrian basement consisting of highly deformed metasedimentary and metavolcanic rocks intruded by felsic to mafic plutons. The sedimentary rocks are composed of coarse to fine clastics as well as pure and impure carbonates that are commonly interbedded with fine clastics (Williams et al., 1992). The graben is spatially associated with Mesozoic lamprophyric dikes and Paleozoic alkali-carbonatite intrusions (Easton, 1992).

3.7 Geophysics

In the past few decades, multiple geophysical methods have been used to study the crustal and lithospheric architecture of the Grenville orogen in southern Ontario and the surrounding geological provinces. These methods include reflection and refraction seismic, teleseismic studies (surface and body wave tomography, shear-wave splitting

and receiver function analysis), gravity and magnetic studies, heat flow and magnetotelluric methods. Some of the results of these studies are summarized in the following sections.

3.7.1 Electromagnetics

The magnetotelluric (MT) method has been used to study much of the Superior Province, its boundary with the Grenville Province and certain parts of the Grenville Province. The electrical resistivity structure of the tectonic units in some sub-provinces of the Superior Province and areas around Grenville Front have been defined using the MT method. Over 100 MT stations were deployed during the LITHOPROBE Abitibi–Grenville Transect to study the electrical resistivity structure of the crust and upper mantle. Studies of the southeastern Superior Province and around the Grenville Front indicate that the geoelectric strike in this area is generally east-west and parallel to the lineations defined by the major deformation zones that affected the area during Precambrian time (Kellett et al., 1994; Sénéchal et al., 1996).

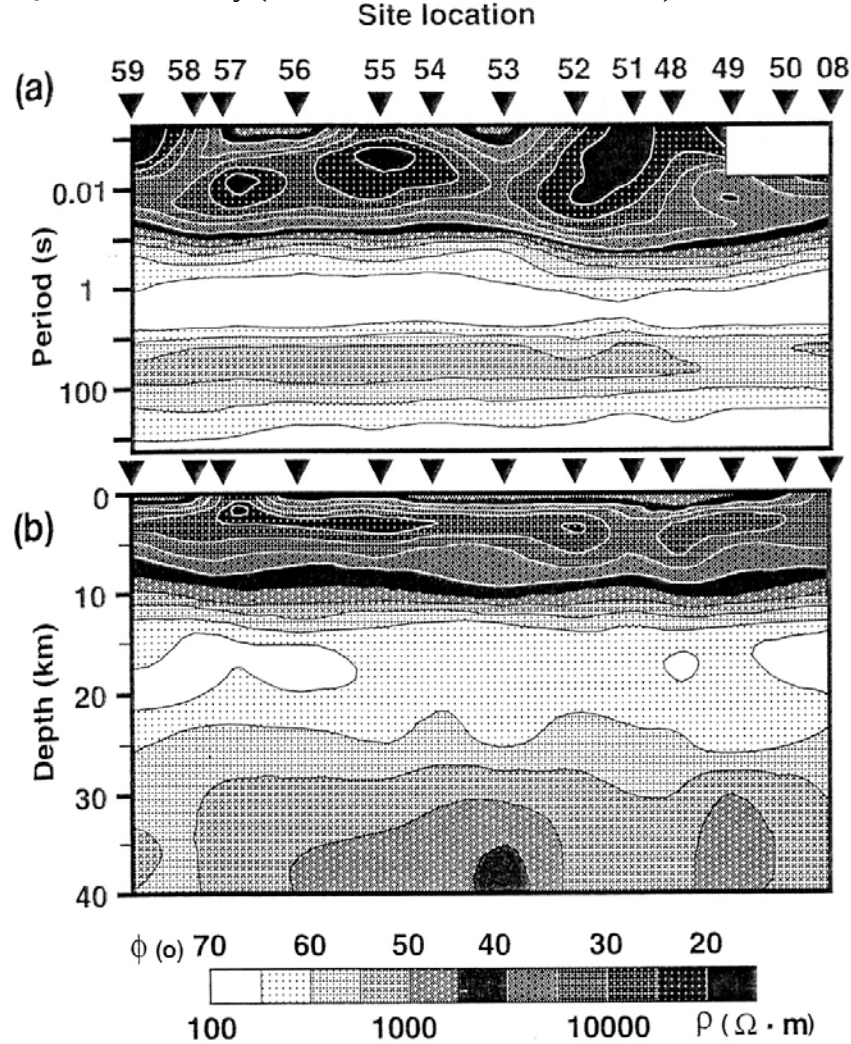
Using ultra long period MT data in the Kapuskasing structural zone within the Superior Province, Schultz et al. (1993) concluded that the upper mantle is resistive and the lower mantle is relatively conductive. In southern Ontario, the upper crust in the Precambrian Shield area is generally resistive and the mid to lower crust is relatively conductive, the uppermost mantle is resistive while the Phanerozoic sedimentary cover is generally conductive (Kurtz, 1976; Kurtz, 1982; Kellett et al., 1992; Zhang et al., 1995). Kellett et al. (1994) also show a similar result for an MT study of the crust at the boundary between the southern Ontario and western Quebec (Fig. 3.5). Boerner et al. (2000) in

their review paper concluded that there is no large-scale enhanced conductivity associated with the metasedimentary rocks in the Abitibi and Pontiac sub-provinces of the Superior Province as well as the Grenville Front area. Mareschal et al. (1991) identified Ordovician black shale at the base of the Whitby Formation with conductivity large enough to mask the resolution of the crustal structures. However, there have been few MT studies in areas south of Grenville Front.

The earlier electromagnetic studies have also included interpretations of deep-seated anisotropic electrical resistivity structure within the mantle of this area (e.g., Kurtz et al., 1988; Kellett et al., 1992; Mareschal et al., 1995; Ji et al., 1996; Frederiksen et al., 2006). The earliest study identifying anisotropy in the southern Superior craton and Grenville Front area was in the area of the Kapuskasing Uplift Structure by Kurtz et al. (1988) who interpreted anisotropy oriented at N65°E at depths exceeding 70 km (Boerner et al., 2000). This anisotropic structure was first discovered below the Pontiac subprovince by Kellett et al. (1992) and it was interpreted as a deep crustal feature. This response has been observed as far as 100 km south of the Grenville Front but it is not observed south of the front further to the east (Boerner et al., 2000). Mareschal et al. (1995) extended the interpretation of electrical anisotropy to Pontiac, Abitibi, Kapuskasing Structural Zone, Wawa subprovinces of the Superior Province and into the Grenville Province (Fig. 3.6).

Mareschal et al. (1995) provided a synthesis and geological interpretation of resistivity anisotropy in the southern Superior and Grenville Front region. They constrained the anisotropy to lie between 50 and 150 km depth with a maximum ratio of horizontal resistivities within the anisotropic layer of 1:15 and a conductive direction

Figure 3.5. The effective phase pseudosection (top) and 1-D inversion model (bottom) of the western Quebec MT study (Taken from Kellett et al., 1994).



azimuth of N80°E (Fig. 3.6). The observed degree of electrical anisotropy cannot be explained by lattice preferred orientation of mantle minerals and was therefore interpreted as being due to an interconnected conducting phase such as graphite. Mareschal et al. (1995) also suggest that the anisotropy may be due to the concentration of graphite in mantle shear zones correlating with observed crustal sutures. The MT responses could be explained by anisotropic resistivity in 50 km wide mantle shear zones,

separated by ~100 km. Ji et al. (1996) suggest that shape-preferred orientation (lineation) of the mantle mineral is responsible for the anisotropic electrical resistivity structure. Obliquity between the electrical anisotropy and seismic anisotropy defined by SKS results has been interpreted as a difference between grain and lattice preferred-orientation of mantle minerals and used as a kinematic indicator of shear in the mantle (Ji et al., 1996; Senechal et al., 1996). Frederiksen et al. (2006) later examined the relationship between electrical and seismic anisotropy over a larger scale. Their result confirms the obliquity between seismic and electrical anisotropy in the vicinity of the Grenville Front.

3.7.2 Seismic Reflection and Refraction

Seismic methods have been extensively used to study the Grenville and the surrounding geological provinces. The Canadian Consortium for Crustal Reconnaissance using Seismic Techniques (COCRUST) used seismic refraction and wide angle seismic reflection method as tools to investigate the crustal structure across the Ottawa-Bonnechere graben and the Grenville Front. Using the results from COCRUST experiment, Mereu et al. (1986) concluded that the crust is at least 5 km thicker at the GFTZ than it is at both the CGB and the CMB. The boundary between the CGB and CMB is deep seated and the CGB is also 3 km thicker than the CMB. These results support the tectonic interpretation that the Grenville Front was formed due to continent-continent collision, and that the CMB was northwestwardly thrust on the CGB. They also concluded that the Moho beneath the Ottawa-Bonnechere graben may have been disrupted by upper mantle materials, which suggests that the graben may be part of a larger ancient rift system.

Using Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE) reflection seismic data, Green et al. (1989) discovered 32 km wide east-dipping reflectors (stacked crustal sheets) that coincide with the GFTZ and extend throughout the crust. These reflectors are attributed to shear zones. The Moho in this region is shown to be between 40 and 46 km depth, except beneath the GFTZ, where it extends to about 55 km due to crustal thickening (Green et al., 1989).

White et al. (2000), using LITHOPROBE seismic reflection and wide angle seismic refraction data together with other previous seismic results, concluded that the thickness of the crust within the Grenville Province in southern Ontario varies from 30 to 44 km (Fig. 3.7). The crustal thickness beneath the Superior and Southern provinces is ~33 km and ~44 km beneath the Grenville Front. Using the reflection patterns, White et al. (2000) subdivided the crust to the upper (0 to 15 km), middle (~15 to 30 km) and lower (~30 to 44 km). The authors argued that the Grenville Front and the GFTZ extend to the mid crustal depth of 25-30 km depth rather than penetrating the entire crust (as indicated by Green et al., 1989). White et al. (2000) also suggest that the Archean Superior Province extends southeast beneath the Grenville Province for several hundred kilometers.

Hughes and Leutgert (1992), using seismic refraction/wide angle reflection data, defined the boundary separating the CMB and the Frontenac-Adirondack Belt (the Carthage Colton mylonite zone) as a 110 km long lineament characterized by intense ductile shear and igneous intrusion that is limited to the upper 2 to 3 km of the crust. This study concluded that the upper mantle of the southeastern Grenville Province is typified by anomalous seismic velocity which forms an eastward dipping layer beneath this area.

This is attributed to the presence of eclogite or anisotropic alignment of olivine in the mantle.

Figure 3.6. MT response anisotropy in the southern subprovinces of the Superior Province (Wawa, Abitibi, and Pontiac subprovinces) and the northern Grenville Province (Taken from Mareschal et al., 1995). DPDZ=Destor-Porcupine deformation zone; CLDZ=Cadillac-Larder deformation zone; GF=Grenville Front. The length of the line at each MT site is proportional to the phase difference between the off-principal MT impedances and the direction is the direction of maximum phase. Sites with erratic apparent resistivity and phase responses as a function of azimuth are shown as dots.

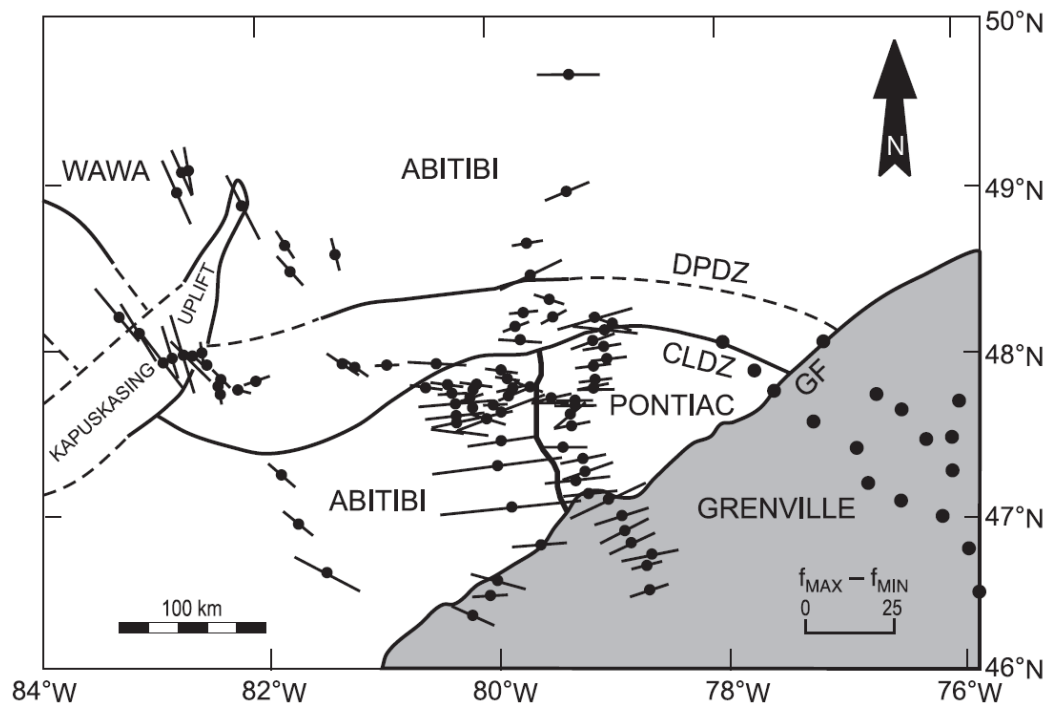
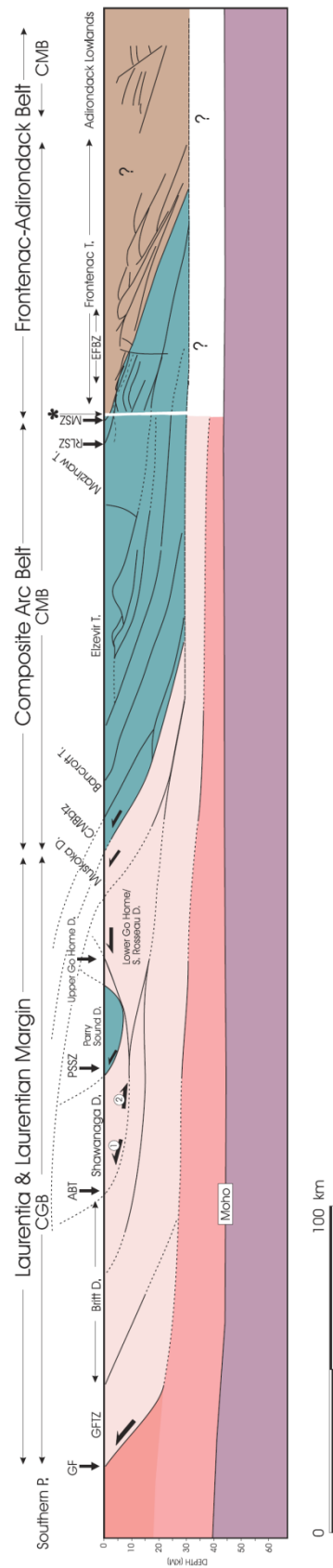


Figure 3.7. Schematic cross section showing the interpretation of the crustal architecture of the southwest Grenville Province (After White et al., 2000).



Using Abitibi-Grenville Lithoprobe transect seismic reflection data, Roy and Mereu (2000) showed that the CMBBZ is a crustal-scale shear zone characterized by bands of southeast-dipping shallow reflectors extending to mid-crustal depths between 25-30 km (Fig. 3.8). In their interpretation, the shear zone, which is similar to the GFTZ, does not penetrate the Moho and accommodated the NW-directed thrusting of the CMB towards the Laurentian margin. The results of this study are consistent with those of White et al. (2000). O'Dowd et al. (2004) processed and interpreted Southern Ontario Seismic Survey (SOSS) reflection seismic data that traversed the CMBBZ (Fig. 3.9). They defined the CMBBZ as a 10 – 12 km wide zone of intense deformation that dips moderately towards the southeast and extends to mid-crustal depth. The CMB was thrust over the pre-Grenvillian Laurentia and its margin along this zone and the thrusting continues in the subsurface beneath undeformed Paleozoic rocks in southern Ontario.

Figure 3.8. Reflection seismic data interpretation (a) and its simplified cartoon (b) showing the major thrust boundary (CMBBZ) as a crustal scale-shear zone along which NW-directed movement occurred during the Grenville orogeny (Modified from Roy and Mereu, 2000).

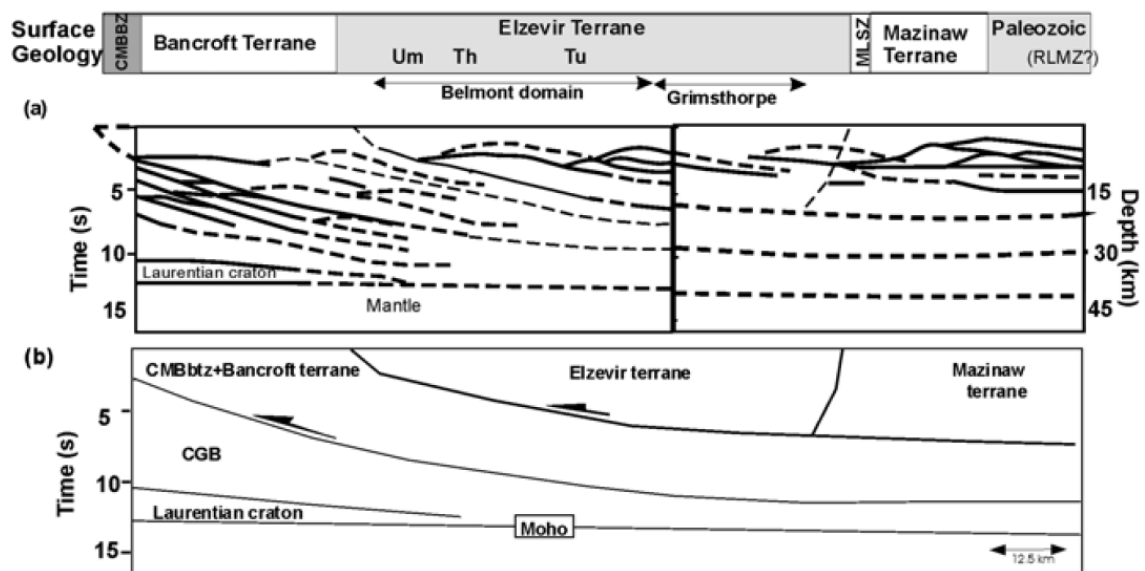
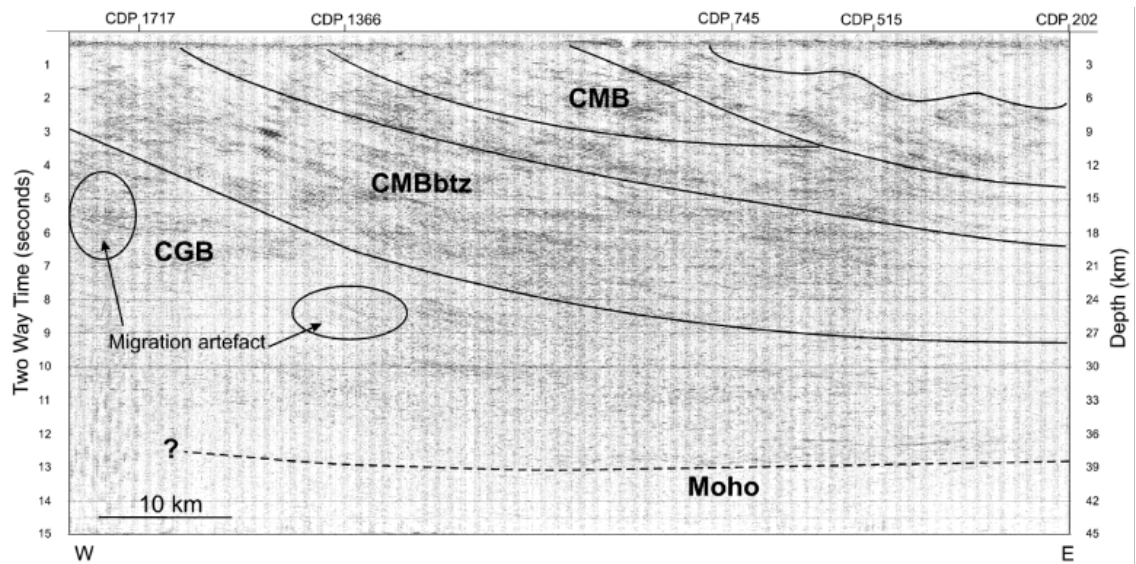


Figure 3.9. Migrated seismic section plotted with no vertical exaggeration for a crustal velocity of 6.0 km/s. CGB=Central Gneiss Belt, CMB=Central Metasedimentary Belt, CMBbtz=Central Metasedimentary Belt boundary thrust zone (Taken from O'Dowd et al., 2004).



3.7.3 Teleseismic Studies

Using POLARIS teleseismic data, Eaton et al. (2006) estimated crustal thickness in the Grenville orogen based on a receiver function approach. The model generated by their study indicates that the crustal thickness varies between 34 km and 52.4 km within the Grenville orogen. The crust is thinnest (34.5 – 37 km) around the Ottawa-Bonnechere graben while it is thickest (~52.4 km) at the Mid-Continental Rift. The crust is 40 – 44 km thick within the CGB while it is 37 – 39 km within the CMB. The scattering tomography results of Zhang and Frederiksen (2013) defined a depth range of 35 to 50 km for the Moho depth in southern Ontario.

van der Lee and Nolet (1997) obtained a model of the upper mantle *S* velocity structure of North America using regional surface wave inversion. They concluded that

the Canadian Shield has a thick lithospheric root which extends to ~250 km beneath the Grenville Province. Rondenay et al. (2000), in their review of the teleseismic studies of the LITHOPROBE Abitibi-Grenville Transect showed, from the *P*-wave velocity model, the presence of a low-velocity corridor (Fig. 3.10). This feature is well defined between 50 and 300 km depths and it is flanked on both sides by regions of relatively high seismic velocity. The anomaly was interpreted as a hot spot track by Rondenay et al. (2000) and Eaton and Frederiksen (2007) and it was described as a lithospheric divot in the North American lithospheric root by Fouch et al. (2000) and Frederiksen et al. (2006). They speculated that this feature may be associated with the process responsible for the emplacement of the Cretaceous Montereign -White Mountains- New England Seamounts magmatic series. Aktas and Eaton (2006) used the LITHOPROBE Abitibi-Grenville and POLARIS teleseismic datasets to study the upper mantle structure beneath the Great Lakes region in southern Ontario, Canada. The *P*- and *S*-wave traveltimes inversion results (Fig. 3.11) show a steeply southeast-dipping slab-like high velocity anomaly that extends up to ~350 km depth and strikes parallel to tectonic belts within the Grenville Province. The feature was interpreted as a relict slab associated with subduction beneath the CMB at 1.25 Ga. In addition, the *S*-wave model also shows a southwest-plunging, ribbon-like low-velocity anomaly that may be associated with the Cretaceous passage of North America over the Greta Meteor hotspot.

Rondenay et al. (2000), using shear-wave splitting analysis, confirmed the presence of seismic anisotropy in the region covered by the dataset (Fig. 3.10). They suggest that the *S*-wave azimuthal anisotropy may be depth dependent and estimated a thickness of 200 km for the anisotropic layer. The teleseismic study of Eaton et al.

(2004), using shear-wave splitting of the POLARIS data from lower Great Lakes region of southern Ontario, suggests the presence of regional anisotropic domains in this area. One of the domains is localized along a narrow belt that coincides with a late Precambrian failed rift that formed during the opening of the Iapetus ocean basin. The spatial variability in the direction and magnitude of splitting is consistent with a flow regime influenced by basal topography of the lithospheric keel (Eaton et al., 2004). The results of the anisotropy studies of Frederiksen et al. (2006) also suggest that the Grenville Province may also be characterized by a seismically anisotropic upper mantle as observed in the Superior Province.

Figure 3.10. Traveltime inversion model showing the west-northwest-striking low-velocity corridor crossing the teleseismic array at latitude 46°N (Taken from Rondenay et al., 2000).

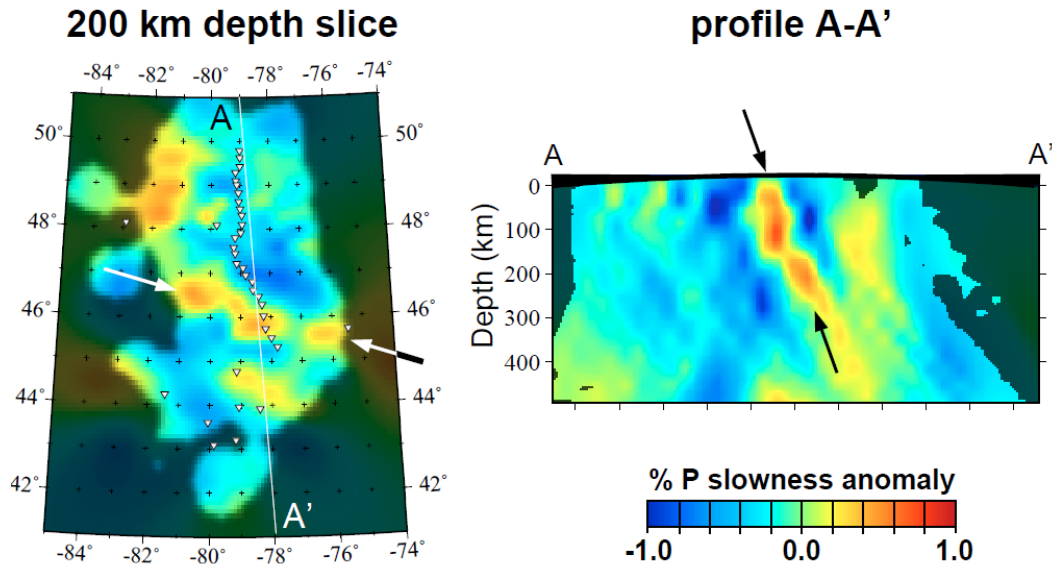
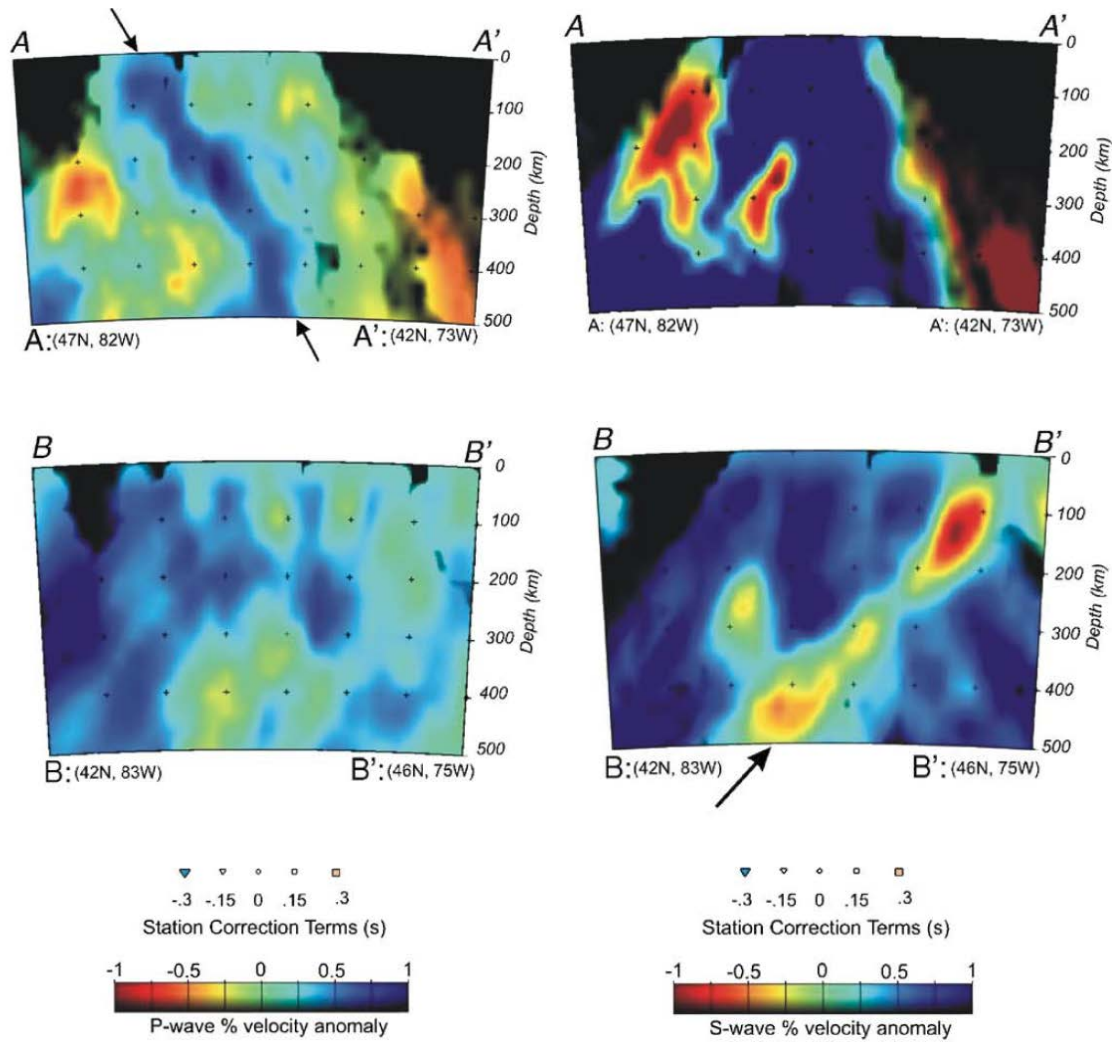


Figure 3.11. The result of traveltime inversion presented by two vertical sections for P-wave (left) and S-wave (right). The arrows on the left panel indicate a linear NE-striking high-velocity anomaly, imaged as a steeply-dipping slab-like feature. The arrow on the right panel indicates a patchy discontinuous, NW-trending low-velocity anomaly, imaged as an inward-dipping ribbon-like feature. The color scale gives percent velocity anomaly with respect to IASP91. Regions of poor ray coverage are masked in black (Taken from Aktas and Eaton, 2006).



The large scale receiver function and SKS studies of Abt et al. (2010), Yuan and Romanowicz (2010), and Yuan et al. (2011) suggest the existence of a two-layer lithosphere and a mid-lithospheric discontinuity (MLD) in areas across North America, especially within Superior Province. This structural boundary is interpreted to exist at depth of approximately 150 km, which is generally shallower than the lithospheric-asthenospheric boundary across the region. In the seismic studies, the layer above the MLD is interpreted to be chemically depleted lithosphere and the lower layer to be a younger, less depleted, thermal boundary layer (Yuan and Romanowicz, 2010; Yuan et al., 2011). More recently, a high resolution 3-D shear wave velocity model of the upper mantle beneath the Grenville Province in the lower Great Lakes region was constructed by Chen and Li (2012). Their result shows a low S-wave velocity anomaly that follows the surface hotspot track and lies in the lithosphere above 100 km. It was suggested that this is an extension of the anomaly imaged by Rondenay et al. (2000). The authors also show NW-SE trending low velocity anomalies beneath the lower Great Lakes region and suggest that the interaction between the plume and the lithosphere is responsible for all these low velocity anomalies. Zhang and Frederiksen (2013) provide scattering tomography P and S velocity models for the crust in the study area. Their model shows break in structure associated with the Ottawa-Bonnechere Graben that cut the whole crust while the CMBBZ is defined as a non-crustal scale structure.

3.7.4 Heat Flow

The heat flow is the sum of the heat produced by the radioactive decay of Uranium (U), Thorium (Th) and Potassium (K) in the crust, and the heat flow at the

Moho, at steady state conditions. Therefore, heat flow is affected by the composition and structure of the crust (Pinet et al., 1991). The heat flow map of the Canadian Shield shows that heat flow contours cross province boundaries but are related to subprovinces with relatively specific geological and petrological characteristics (Fig. 3.12; Mareschal and Jaupart, 2004). The thermal structure of the Canadian Shield is highly heterogeneous (Mareschal et al., 2000).

There are no significant spatial and temporal heat flow variations in the eastern Canadian Shield, thus this area is considered as a single heat flow province. Regionally, there is no clear change in the heat flow across the Grenville Front but there is a clear increase in the surface heat flow in the Appalachians (Fig. 3.13; Mareschal et al., 2000). The low average heat production and flow in the Grenville Province is interpreted to be a result of intense erosion, mafic intrusions and post orogenic extensions (Guillou-Fronttier et al., 1995). The mantle heat flux is interpreted to be relatively constant across different provinces and subprovinces of the southeastern Canadian Shield ($10\text{--}14 \text{ mW.m}^{-2}$; Guillou et al., 1994; Mareschal et al., 2000). This is an indication that the depth to the 450°C isotherm, which defines the effective elastic thickness of the lithosphere, is very sensitive to crustal heat production (Pinet et al., 1991).

Pinet et al. (1991) in their investigation of the thermal structure of the continental lithosphere beneath the Superior (Abitibi greenstone belt) and Grenville provinces show that the average heat flow in the Grenville Province ($41 \pm 10 \text{ mW.m}^{-2}$) is the same as that of the Superior Province ($41 \pm 9 \text{ mW.m}^{-2}$). This result suggests that the crusts of both provinces are similar in composition. In contrast, the heat flow in the Appalachian Province is higher (55 mW.m^{-2}) and results from granitic intrusions, with high heat

producing elements, into its uppermost crustal layers (Pinet et al., 1991; Guillou-Fonttner et al., 1995; Mareschal et al., 2000).

Figure 3.12. Heat flow map of the Canadian Shield with white dots indicating heat flow sites. Archean provinces: Superior (Sup), Hearne and Rae; Proterozoic Provinces: Trans-Hudson Orogen (THO), Grenville (Gre); Paleozoic Province: Appalachians (App); Paleozoic Basins: Williston (Will), Hudson Bay (H.B.) (Taken from Mareschal and Jaupart, 2004).

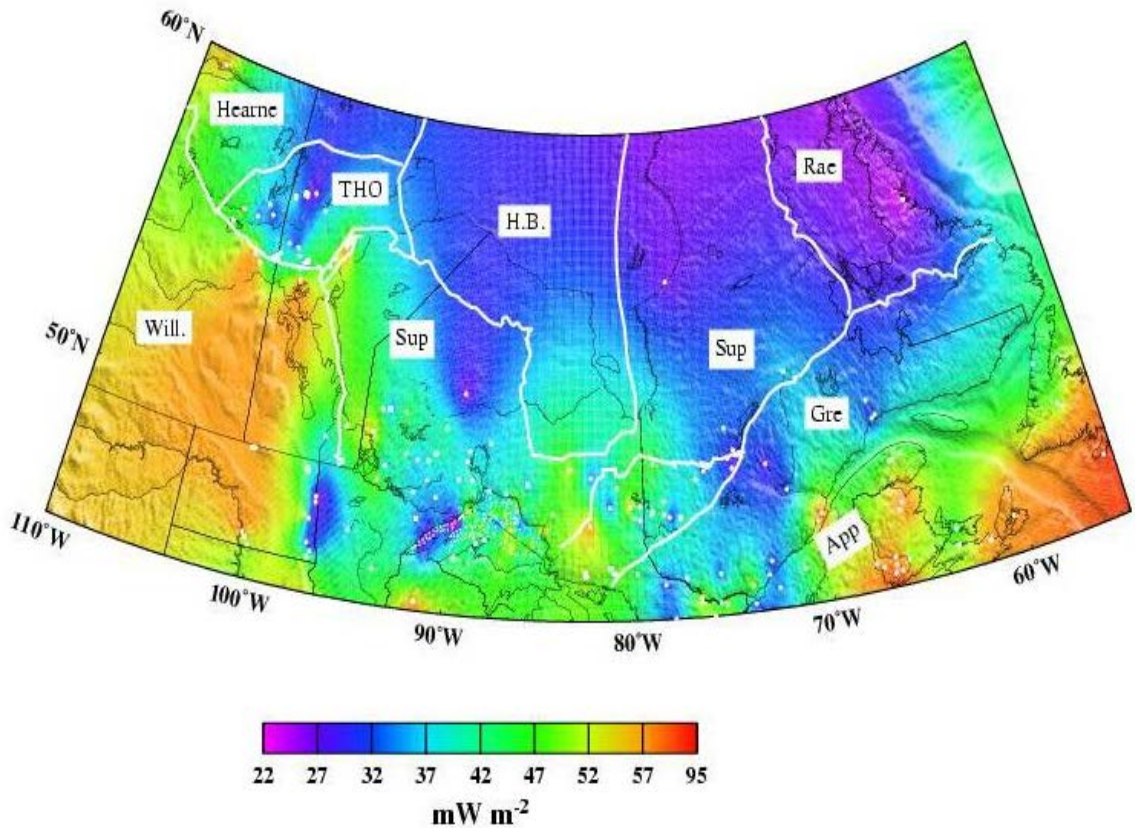
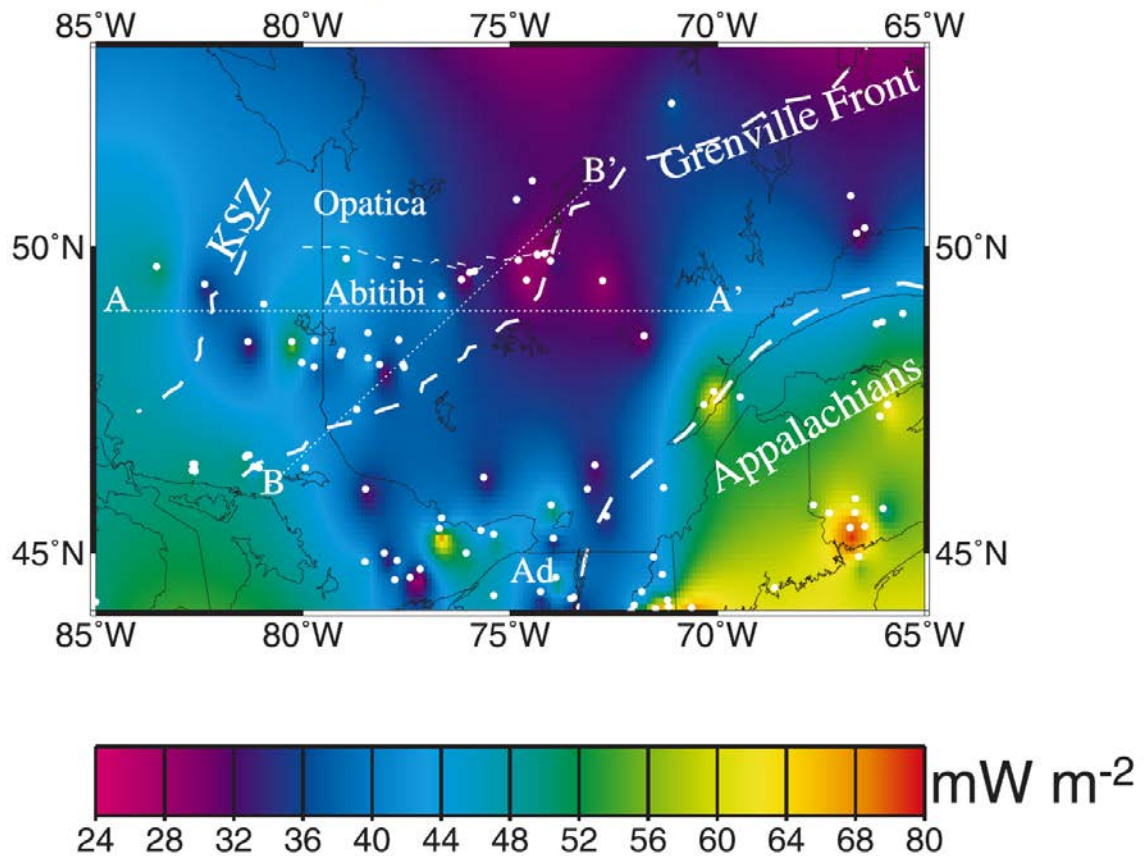


Figure 3.13. Map of heat flow in southeastern Canada with white dots indicating heat flow sites. The center of the Adirondacks is indicated as Ad while A–A' and B–B' are the two profiles examined by the authors in their study (Taken from Marescahl et al., 2000)



3.7.5 Gravity and Magnetics

Some authors have also used gravity and magnetics geophysical methods to determine the crustal and lithospheric structures of southern Ontario. Gravity survey results (Fig. 3.14) shows a general increase of the Bouguer gravity anomaly from the Grenville Province to the Superior Province. Mareschal et al. (2000) used the long wavelength Bouguer anomaly to construct a gravity map which indicates that a gravity low (negative Bouguer anomaly) associated with the eastern part of the Grenville Front

disappears towards the west (southern Ontario). This gravity low is associated with crustal thickening in the area. It is also clear that there is no crustal thickening at the border between the Appalachian and Grenville Province since there is no gravity anomaly associated with this area (Mareschal et al., 2000). Easton (1992) showed that the Grenville Front is characterized by a low gravity anomaly.

Roy and Mereu (2000) investigated the Grenville Province with a 2.5-D modeling of gravity data (Fig. 3.15) and they concluded that rocks of the CMB have higher density (high gravity anomaly) than those of the CGB (low gravity anomaly). They suggest that a crustal-scale process is responsible for the accretion of CGB to the CMB as well as the NW-directed tectonic transport that occurred during Grenville orogeny. Their result also shows that, within the CMB, the rocks of Bancroft terrane are less dense than those of Elzevir and Mazinaw terranes. Bank et al. (2005), in a separate Bouguer gravity study, explained that the lower crust of the CGB is much denser than the lower crust of the CMB. Easton (1992) showed that the most part of the CGB, particularly the Parry Sound and Nepewassi domains, have high gravity anomaly.

O'Dowd et al. (2004) show that the CGB is characterized by high magnetic anomalies, the CMBBZ corresponds to a magnetic low, and the CMB is typified by linear, NE-trending and sub-circular magnetic highs. Figure 3.16 shows the excellent relationship between the total magnetic field model and the seismic section obtained in this region.

Figure 3.14. Long-wavelength (> 100 km) Bouguer gravity anomaly map of eastern Canada. Ad indicates the center of the Adirondacks while A-A' and B-B' show the profiles examined by the authors (Taken from Mareschal et al., 2000). KSz = Kapuskasing Structural Zone

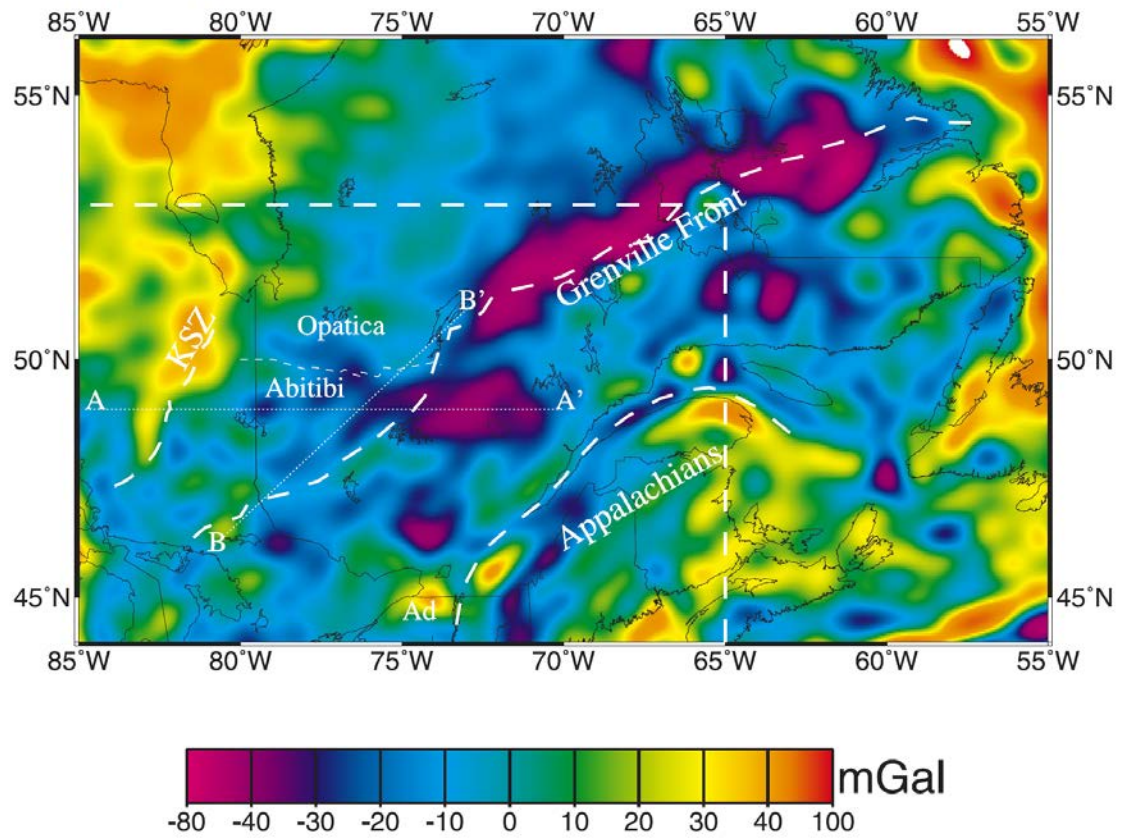


Figure 3.15. (a) Cartoon showing the major thrust boundaries within the CMB (b) the simplified density distribution within the CMB and (c) depth section obtained from the inversion of the gravity data which shows that the CMB is denser than the CGB and that density variation occurs within the CMB itself (modified from Roy and Mereu, 2000).

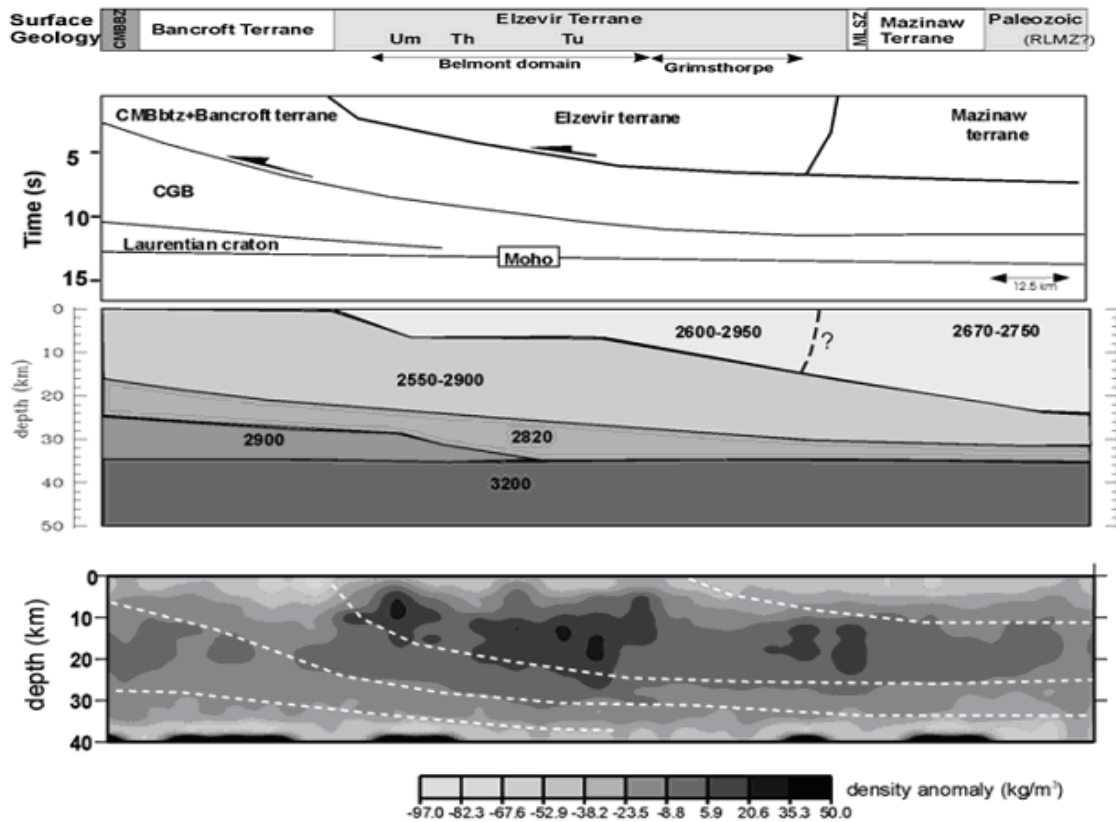
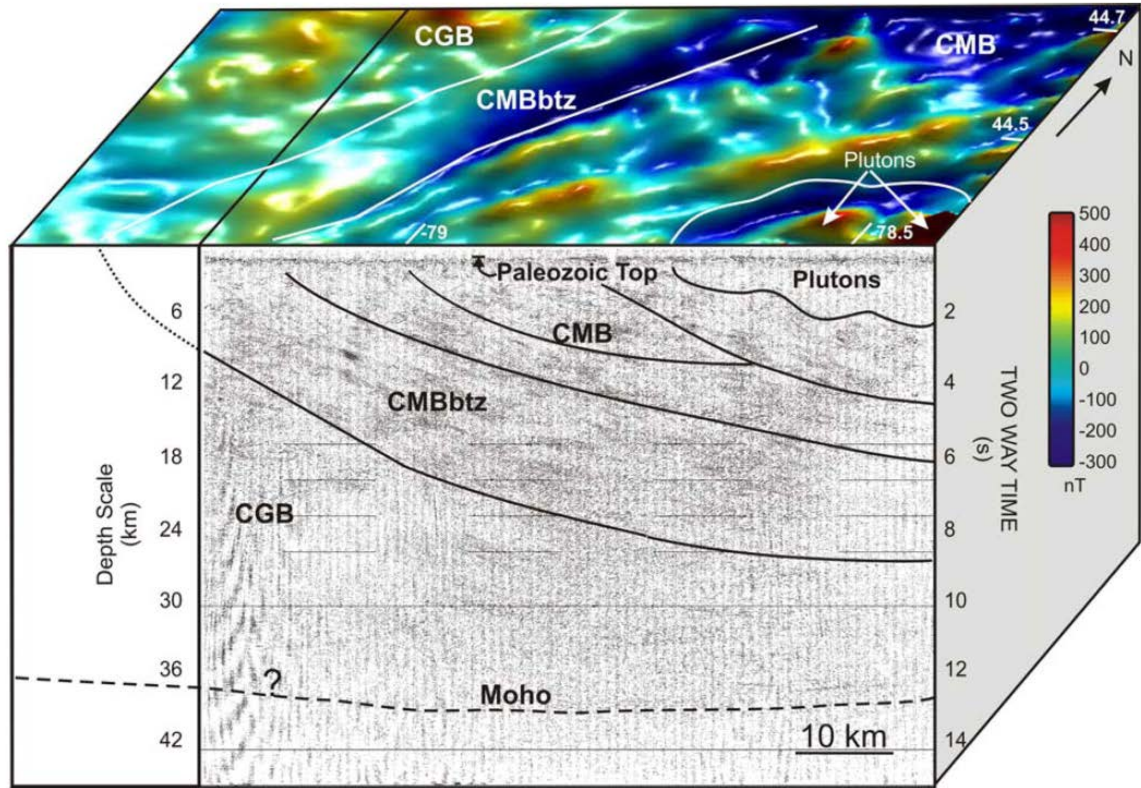


Figure 3.16. Block diagram showing comparison between the total magnetic field (on top) and seismic section (front face) obtained from southern Ontario (Taken from O'Dowd et al., 2004).



3.8 Tectonic Evolution of the Grenville Province

The Grenville orogen is interpreted to have been the result of continent-continent and/or continent-arc collisions (Slagstad, 2003). The Grenville Province was formed by middle to Late Mesoproterozoic orogenic activity (Hanmer, 1988; Hoffman, 1989; Easton, 1992). The orogeny followed a period of pre-Grenvillian continental-margin arc magmatism, accretionary, and collisional tectonic events (Carr et al., 2000). The Grenville Province was formed between 1.2 and 1.0 Ga as a result of Himalayan scale

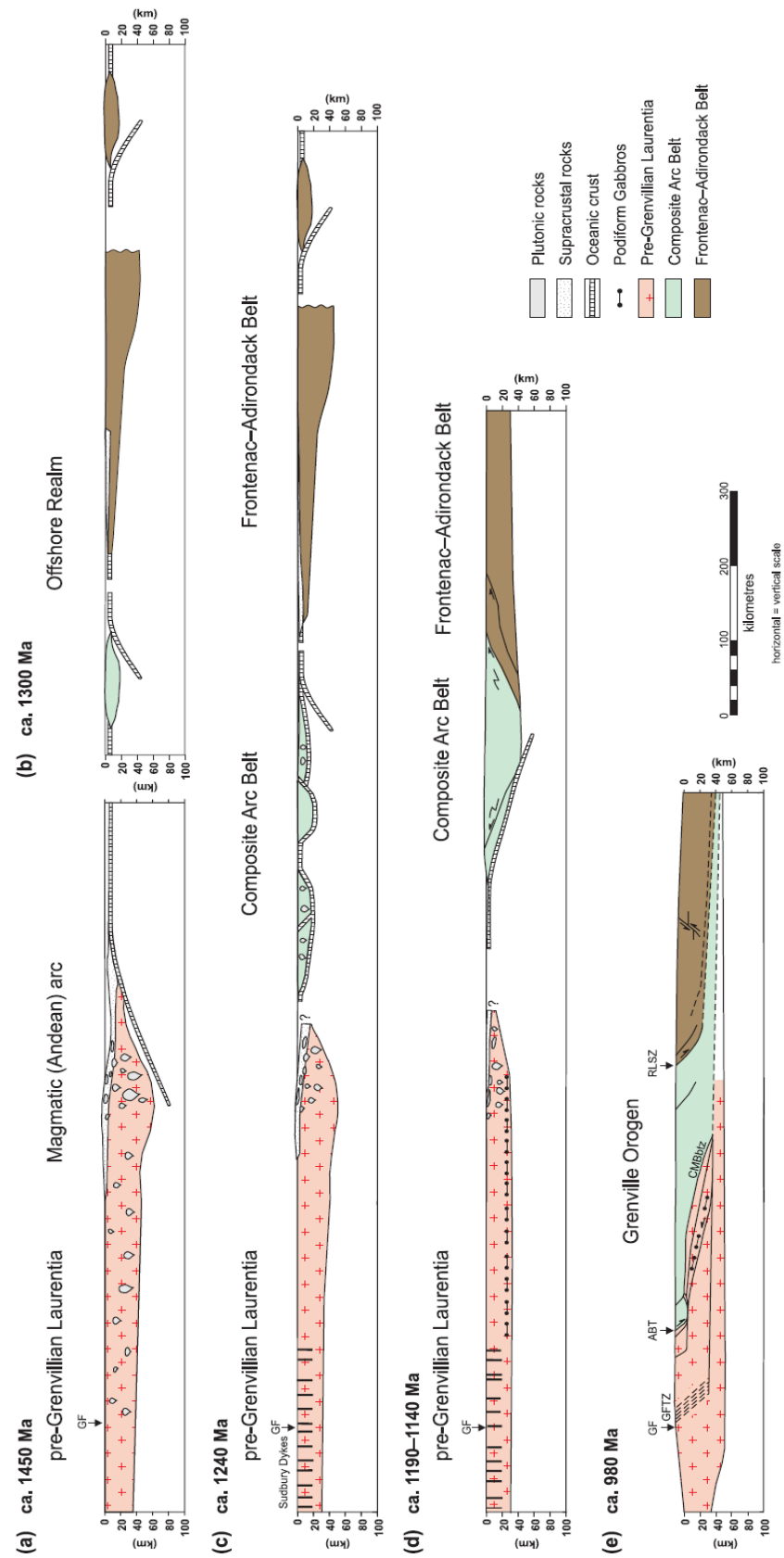
orogeny and produced a northwest-trending thrusting of a composite belt of continental crust and magmatic rocks onto the margin of the Laurentian craton (Carr et al., 2000).

Three episodes of collisional orogeny have been recognized on the basis of their time/space distributions within the 200 Ma period of formation of the Grenville Province (Fig. 3.17; Wynne-Edwards, 1972; Rivers et al., 1989; Carr et al., 2000).

The first orogen-wide phase of high grade metamorphism and convergent deformation associated with the Grenvillian orogenic cycle occurred between 1190 and 1140 Ma. This event is attributed to collision between Laurentia and continental terranes and/or magmatic arcs to the southeast (Rivers et al., 1989, Davidson, 1995; Corrigan and van Breemen, 1997). The pattern of tectonism, with two belts of deformed rocks on either side of the relatively unaffected CAB rocks, suggests a doubly vergent compressional system in which the previously amalgamated, thickened CAB may have collided with the FAB and perhaps the distal margin of Laurentia (Carr et al., 2000).

During this first phase of convergence, the CGB was intruded by gabbro (Ketchum and Davidson, 2000); and the CAB and FAB were affected by high grade metamorphism and deformation. The deformation was accommodated along the CMBBZ area and this led to the initial formation of the CMBBZ (McEachern and van Breemen, 1993). The Sharbot Lake terrane, Frontenac terrane and Adirondack lowlands were penetratively deformed and metamorphosed at this time (Corfu and Easton, 1997). Widespread AMCG magmatism occurred in the CGB and CAB at this time (Rivers, 1997).

Figure 3.17. Schematic plate tectonic model for the formation of the Grenville Province (After Carr et al., 2000).



The initial encounter between the Laurentian margin and the CAB may have been represented by the 1120 Ma tectonism that affected both the Parry Sound domain and rocks interpreted to have formed on the margin of Laurentia (Wodicka et al., 1996; Culshaw et al., 1997). This event is followed by the emplacement of the Parry Sound allochthon over the Laurentia suggesting that convergent orogenesis began at or near the end of this time (Wodicka et al., 1996; Culshaw et al., 1997; Carr et al., 2000).

The most significant tectonic event to affect Laurentia between 1140 and 1090 Ma was the development of the Midcontinental rift system to the west. Rates of flood basalt volcanism and subsidence were greatest at the onset of the rifting and volcanism ceased by 1087 Ma. The predominance of terrigenous sedimentary rocks in the rift system indicates that the region was above sea level. The age and scale of the Midcontinental rift system suggest that its evolution may have been influenced by and/or had influence on Grenville orogeny (Carr et al., 2000).

The second phase, the Ottawa orogeny, differs from the older one in that northwest-directed thrusting occurred along the entire length of the orogen started shortly after 1090 Ma (Fig. 3.17). It involved widespread deformation and metamorphism that affected the Laurentian margin, parts of CAB and the Adirondack highlands. The tectonic setting was dominantly convergent for 100 Ma and regionally significant extensional structures occurred (Rivers et al., 1989; Carr et al., 2000).

The Ottawa orogeny involved the emplacement of previously amalgamated CAB and FAB allochthons over Laurentia (Fig. 3.17). Upper amphibolite facies metamorphism of Laurentian rocks accompanied or immediately postdated this thrusting and high pressure metamorphism preceded it. The cessation of Midcontinental rift volcanism at

1087 Ma may have been linked to this stage of convergence. There is evidence for deep burial and partial subduction of the southeast edge of the Laurentia at the beginning of convergence (Culshaw et al., 1997; Carr et al., 2000).

The propagation of the orogen onto Laurentia and relative waning of deformation in the Adirondack highlands could be explained by the forward migration of the underlying detachment point. If so, deformation and metamorphism of the Mazinaw terrane rocks within the unaffected CAB might indicate temporary re-establishment of the retro-shear accompanying this migration (Carr et al., 2000).

The final stage of the northwest-directed convergence at 1000 Ma affected the GFTZ along the length of the orogen (Carr et al., 2000). The GFTZ has been attributed to exhumation of the retro-shear zone separating the orogen from its foreland (Haggart et al., 1993). Available data indicate that some form of post-orogenic magmatic activity continued in the interior of the orogen for about 40 Ma after the GFTZ stage of convergence. The thickness of the orogen is 60-65 km as inferred from metamorphic pressures obtained from rocks re-crystallized in the hanging wall of the extensional Shawanaga shear zone and the current thickness of underlying crust (40 km; Carr et al., 2000).

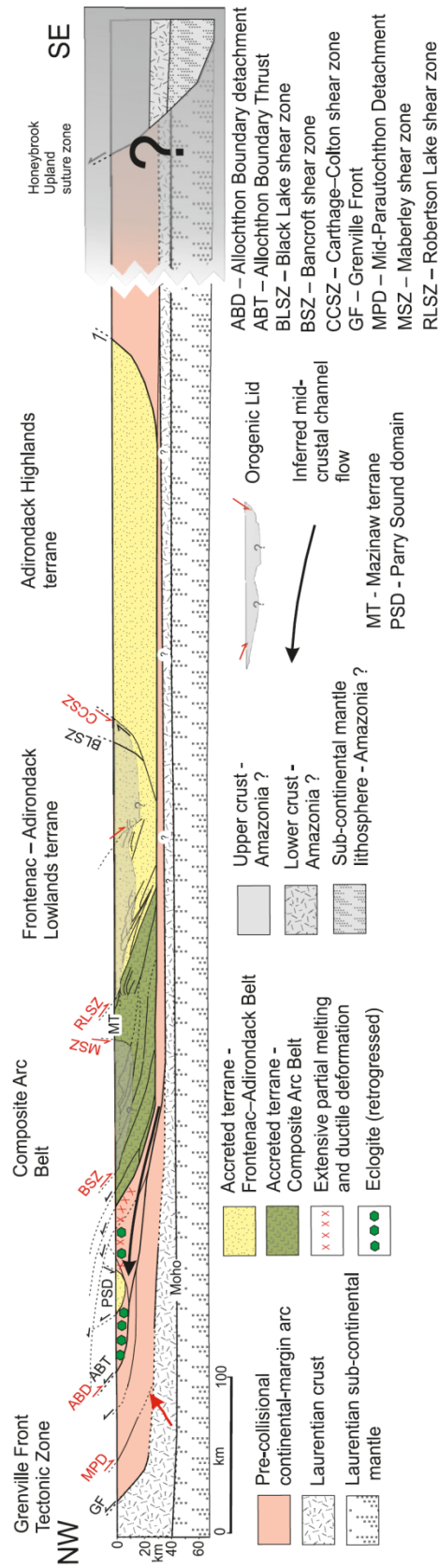
Syn-orogenic extension varies in timing and character and therefore, probably in cause both across and along strike. In the Laurentian margin, pervasive ductile normal sense fabrics and generally shallow dipping structures have been interpreted to have formed during late orogenic extension (Culshaw et al., 1997). In the CAB (Fig. 3.18), moderately dipping to steep brittle-ductile normal faults contrast in age and style with both observed in the northwest. These faults started during the waning stages of

convergence and their prolonged evolution is due to post-orogenic regional cooling and subsidence (Carr et al., 2000).

Rocks from the northwestern flank of Grenville orogen recorded slow regional cooling, presumably due to erosional unroofing after 980 Ma. In the CAB, post-orogenic cooling was accompanied by denudation on normal faults (Carr et al., 2000). Little activity is recorded in the orogen or its Laurentian foreland until 600 Ma, when rifting of North America resulted in the development of the Ottawa-Bonnechere Graben, and emplacement of the Grenville diabase dyke swarm (Kumarapeli, 1985; Sage, 1991; Carr et al., 2000). The Ottawa-Bonnechere Graben is interpreted to be a failed arm of a triple junction bounded by nearly vertical faults trending along an east-west to west-northwest direction (Easton, 1992; Kumarapeli, 1985). Reactivation of these faults is associated with the Mesozoic opening of the Atlantic Ocean (Easton, 1992).

The youngest tectonic event associated with the study area is the Great Meteor hotspot (Crough, 1981; Sleep, 1990). The track of this hotspot was interpreted to be responsible for the emplacement of the Cretaceous Monteregian-White Mountain-New England Seamounts Igneous Province (Crough, 1981; Sleep, 1990). It is also interpreted to be associated with the Mesozoic kimberlite magmatism at Rapide des Quinze (Ji et al., 1996) and Kirkland Lake (Meyer et al., 1994; Ji et al., 1996) areas of the Pontiac subprovince (Griffin et al., 2004; Faure et al., 2011).

Figure 3.18. Schematic cross section showing the tectonic architecture of the Grenville Orogen in southern Ontario based on geophysical and geological data (modified from White et al., 2000 and Ludden and Hynes, 2000, Taken from Hynes and Rivers, 2010).



3.9 Reference Cited

- Abt, D.L., Fischer, K.M., French, S.W., Ford, H.A., Huaiyu, Y., Romanowicz, B., 2010. North American lithospheric discontinuity structure imaged by Ps and Sp receiver functions. *J. Geophys. Res.* 115, B09301, doi:10.1029/2009JB006914.
- Aktas, K., Eaton, D. W., 2006. Upper-mantle velocity structure of the lower Great Lakes region. *Tectonophysics* 420, 267–281.
- Armstrong, D.K., Carter, T. R., 2006. An updated guide to the subsurface Paleozoic stratigraphy of southern Ontario. Ontario Geological Survey, Open File Report 6191.
- Bank, C., Eaton, D., Aktas, K., 2005. Moho topography and lower crustal density in Southern Ontario from linearized gravity inversion. AGU spring meeting 23-27 May, New Orleans, T43D-04.
- Bennett G., Dressler, B. O., Robertson, J. A., 1991. The Huronian Supergroup and associated intrusive rocks, in: Thurston, P.C., Williams, H.R., Sutcliffe, R.H., Stott, G.M. (Eds.), *Geology of Ontario*, Ontario Geological Survey, Special Vol. 4, Part 2, 713 – 904.
- Boerner, D. E., Kurtz, R. D., Craven, J. A., 2000. A summary of electromagnetic studies on the Abitibi-Grenville transect. *Can. J. Earth Sci.* 37, 427–437.
- Card, K. D., Poulsen, K. H., 1998. Geology and mineral deposits of the Superior Province of the Canadian Shield, in: Lucas, S.B. (Eds.), *Geology of the Precambrian Superior and Grenville Provinces and Precambrian Fossils in North America*, Geological Survey of Canada, *Geology of Canada* 7, 13-194.
- Card, K.D., 1990. A review of the Superior Province of the Canadian Shield, a product of Archean accretion, *Precambrian Res.* 48, 99-156.
- Card, K.D., Ciesielski, A., 1986. Subdivisions of the Superior Province of the Canadian Shield. *Geoscience Canada* 13, 5-13.
- Carr, S. D., Easton, R. M., Jamieson, R. A., Culshaw, N. G., 2000. Geologic transect across the Grenville orogen of Ontario and New York: *Can. J. Earth Sci.* 37, 193-216.
- Chapman, I. J., Putman, D. J., 1984. *The physiography of southern Ontario*, third edition; Ontario geological Survey, Special Volume 2, 270p.
- Chen, C. -W., Li, A., 2012. Shear wave structure in the Grenville Province beneath the lower Great Lakes region from Rayleigh wave tomography. *J. Geophys. Res.* 117, B01303.
- Condie, K. C., Moore, J. M., 1977. Geochemistry of Proterozoic volcanic rocks from the Grenville Province, eastern Ontario, in: Baragar, W. R. A., Coleman, L. C., Hall, J. M., (Eds.), *Volcanic regimes in Canada*, Geological Association of Canada Special Paper 16, 149-168.
- Corfu, F., Easton, R. M., 1995. U-Pb geochronology of the Mazinaw terrane, an imbricate segment of the Central Metasedimentary Belt, Grenville Province, Ontario. *Can. J. Earth Sci.* 32, 959-976.
- Corfu, F., Easton, R. M., 1997. Sharbot Lake terrane and its relationships to Frontenac terrane, Central Metasedimentary Belt, Grenville Province: new insights from U–Pb geochronology. *Can. J. Earth Sci.* 34, 1239–1257.
- Corfu, F., 1993. The evolution of the southern Abitibi greenstone belt in light of precise U-Pb geochronology. *Economic Geology* 88, 1323–1341.

- Corrigan, D., van Breemen, O., 1997. U–Pb age constraints for the lithotectonic evolution of the Grenville Province along the Mauricie transect, Quebec. *Can. J. Earth Sci.* 34, 299–316.
- Corriveau, L., Heaman, L. M., Marcantonio, F., van Breeman, O., 1990. 1.1 Ga K-rich alkaline plutonism in the SW Grenville Province, U-Pb constraints for the timing of subduction related magmatism. *Contributions to Mineralogy and Petrology* 105, 473-485
- Crough, S.T., 1981. Mesozoic hot spot epeirogeny in eastern North America. *Geology* 9, 2-6.
- Culshaw, N.G., Davidson, A., Nadeau, L., 1983. Structural subdivisions of the Grenville Province in the Parry Sound-Algonquin region, Ontario. In *Current Research, Part B*, Geological Survey of Canada. Paper 83-1B, 243- 252.
- Culshaw, N.G., Jamieson, R. A., Ketchum, J. W. F., Wodicka, N., Corrigan, D., Reynolds, P. H., 1997. Transect across the northwestern Grenville orogen, Georgian Bay, Ontario: Polystage convergence and extension in the lower orogenic crust. *Tectonics* 16, 966-982.
- Davidson, A., 1991. Metamorphism and tectonic setting of gabbroic and related rocks in the Central Gneiss Belt, Grenville Province, Ontario, GAC-MAC-SEG, Toronto, Guidebook. Field Trip A2.
- Davidson, A., van Breemen, O., 2000. Age and extent of the Frontenac plutonic suite in the Central Metasedimentary Belt, Grenville Province, southeastern Ontario. *Geol. Survey of Canada, Current Research 2000-F4: Radiogenic Age and Isotopic Studies*, Report 13, 15p.
- Davidson, A., 1984. Identification of ductile shear zones in the southwestern Grenville Province of the Canadian Shield; in *Precambrian Tectonics Illustrated*, E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, 263-279.
- Davidson, A., 1986. New interpretations in the southwestern Grenville province, in: Moore, J. M., Davidson, A., Baer, A. J. (Eds), *The Grenville Province*, Geological Association of Canada, Special Paper 31, 61-74.
- Davidson, A., 1995. A review of the Grenville orogen in its North America type area, *Journal of Australian Geology and Geophysics* 16, 3-24.
- Davidson, A., Morgan, W. C., 1981. Preliminary notes on the geology east of Georgian Bay, Grenville Structural Province, Ontario. In *Current Research*, Geological Survey of Canada, Paper 81-1A: 291-298.
- Davidson, A., Culshaw, N. G., Nadeau, L., 1982. A tectono-metamorphic framework for part of the Grenville province, Parry Sound region, Ontario. in: *Current Research. Part A*. Geological Survey of Canada. Paper 82-1A. pp. 175-190.
- Davidson, A., 1998. Questions of correlation across the Grenville Front east of Sudbury, Ontario; in *Current Research 1998-C*; Geological Survey of Canada. Paper 98-IC, p. 145-154.
- Davis, D. W., 1992. Age constraints on deposition and provenance of Archean sediments in the southern Abitibi and Pontiac Subprovinces from U-Pb analyses of detrital zircons. *Lithoprobe Abitibi-Grenville Transect. Report 25*, 147-150.
- Dickinson, A. P., McNutt, R. H., 1989. Nd model-age mapping of south-eastern margin of the Archean foreland in the Grenville Province of Ontario, *Geology* 17, 299-303.

- Dimroth, E., L. Imreh, N. Goulet, Rocheleau, M., 1982. Evolution of the south-central segment of the Archean Abitibi Belt, Quebec, Part I: Stratigraphy and paleogeographic model. *Can. J. Earth Sci.* 20, 1355-1373.
- Dimroth, E., Imreh, L., Goulet, N., Rocheleau, M., 1983. Evolution of the south-central segment of the Archean Abitibi Belt, Quebec, Part II: Tectonic evolution and geomechanical model. *Can. J. Earth Sci.*, 20, 1355-1373.
- Easton, R. M., Davidson, A., 1997. Frontenac-Sharbot Lake relationships: evaluation of a provocative hypothesis concerning the tectonics of the Central Metasedimentary Belt, Ontario. *Friends of the Grenville, Annual Field Excursion Guidebook*, pp. 26-28.
- Easton, R. M., 1986, Geochronology of the Grenville Province, in: Moore, J. M., Davidson, A., Baer, A. J. (Eds), *The Grenville Province*, Geological Association of Canada, Special paper 31, 127-173.
- Easton, R. M., 1992. The Grenville Province and the Proterozoic history of Central and Southern Ontario, in: *Geology of Ontario*, Ontario Geological Survey, Special Volume 4, 713 – 904.
- Eaton, D. W., Frederiksen, A., 2007. Seismic evidence for convection driven motion of the North American plate. *Nature* 446, 428–431.
- Eaton, D. W., Dineva, S., Mereu, R., 2006. Crustal thickness and Vp/Vs variations in the Grenville orogen Ontario, Canada from analysis of teleseismic receiver functions. *Tectonophysics* 420, 223-238.
- Eaton, D., A. Frederiksen, and S. K. Miong (2004), Shear-wave splitting observations in the lower Great Lakes region: Evidence for regional anisotropic domains and keel-modified asthenospheric flow, *Geophys. Res. Lett.*, 31, L07610, doi:10.1029/2004GL019438.
- Faure S., Godey, S., Fallara, F., Trepanier, S., 2011. Seismic architecture of the Archean North American mantle and its relationship to diamondiferous kimberlite fields. *Econ. Geol.* 106, 223-240.
- Feng, K., Fan, J., Kerrich, R., 1993. Noble metal abundances and characteristics of six granitic magma series, Archean Abitibi Belt, Pontiac Subprovince; relationship to metallogeny and overprinting of mesothermal gold deposits. *Economic Geology* 88, 1376 – 1401.
- Fisher, J. H., Barratt, M. W., Droste, J. B., Shaver, R. H., 1988. Michigan Basin, in: Sloss, L.L., (Ed.), *Sedimentary cover—North American craton*, v. D-2 of *The geology of North America*: Boulder, Colo., Geological Society of America, pp. 361-382.
- Forsyth, D.A., Milkereit, B., Davidson, A., Hanmer, S., 1994. Seismic images of a tectonic subdivision of the Grenville Orogen beneath lakes Ontario and Erie. *Can. J. Earth Sci.* 31, 229-242.
- Fouch, M. J., Fischer, K.M., Parmentier, E.M., Wyssession, M.E., Clarke, T.J., 2000. Shear wave splitting, continental keels, and patterns of mantle flow. *J. Geophys. Res.* 105, 6255-6275.
- Frederiksen, A.W., Ferguson, I. J., Eaton, D., Miong, S. K., Gowan, E., 2006. Mantle fabric at multiple scales across an Archean–Proterozoic boundary, Grenville Front, Canada. *Phys. Earth Planet. Int.* 158, 240–263.

- Gartner Lee Limited, 2008. Phase 1 Regional Geology, Southern Ontario, OPG's Deep Geologic Repository for Low and Intermediate Level Waste. OPG Report # 00216-REP-01300-00007-R00.
- Green, A.G., Cannon, W.F., Milkereit, B., Hutchinson, D.R, Davidson, A, Behrendt, J.C., Spencer, C., Lee, M.W., Morel-a-l'Huissier, P. and Agena, W.F., 1989. A "Glimpce" of the deep crust beneath the Great Lakes; in *Properties and Processes of Earth's Lower Crust*, AGU Monograph 51,65-80.
- Griffin, W. L., O'Reilly, S. Y., Doyle, B. J., Pearson, N. J., Coopersmith, H., Kivi, K., Malkovets, V., Pokhilenko, N., 2004. Lithosphere mapping beneath the North American plate. *Lithos* 77, 873-922.
- Guillou, L., Mareschal, J-C., Jaupart, C., Gariépy, C., Bienfait, G., Lapointe, R., 1994. Heat flow gravity and structure of the Abitibi belt, Superior Province, Canada: implications for mantle heat flow. *Earth Plan. Sci. Lett.* 122, 103-123.
- Guillou-Frottier, L., Mareschal, J-C., Jaupart, C., Gariépy, C., Lapointe, R., Bienfait, G., 1995. Heat flow variations in the Grenville Province, Canada. *Earth Plan. Sci. Lett.* 136, 447-460.
- Haggart, M. J., Jemieson, R. A., Reynold, P. H., Krogh, T. E., Beaumont, C., Culshaw, N. G., 1993. Last gasp of the Grenville Orogeny: thermochronology of the Grenville Front Tectonic Zone near Killarney, Ontario. *J. Geology* 101, 575-589.
- Hanmer, S., McEachern, S. J., 1992. Kinematical and rheological evolution of a crustal-scale ductile thrust zone, Central Metasedimentary Belt, Grenville Orogen, Canada. *Can. J. Earth Sci.* 29, 1779-1790.
- Hanmer, S. 1988. Ductile thrusting at mid-crustal level, southwestern Grenville Province, Canada. *Can. J. Earth Sci.* 25, 1049-1059.
- Hanmer, S., Ciesielski, A., 1984. A structural reconnaissance of the northwest boundary of the Central Metasedimentary Belt, Grenville Province, Ontario and Quebec, in: *Current Research Part B, Geological Survey of Canada, Paper 84-18*, pp. 121-131.
- Hanmer, S., Thivierge, R. H., Henderson, J. R., 1985. Anatomy of a ductile thrust zone: part of the northwest boundary of the Central Metasedimentary Belt, Grenville Province, Ontario, in: *Current Research Part B, Geological Survey of Canada, Paper 85-18*, pp. 1-5.
- Hildebrand, R. S., Easton, R. M., 1995. A 1161 Ma suture in the Frontenac terrane, Ontario segment of the Grenville orogeny. *Geology* 23, 917-920.
- Hoffman, P. F., 1989. Precambrian geology and tectonic history of North America, in: Bally, A. W., Palmer, A. R. (Eds.), *The Geology of North America - An Overview*, The Geological Society of America, vol. A, pp. 447-512,
- Howell, P.D., van der Pluijm, B.A. 1990. Early history of the Michigan basin: subsidence and Appalachian tectonics. *Geology* 18, 1195-1198.
- Howell, P.D., van der Pluijm, B.A. 1999. Structural sequences and styles of subsidence in the Michigan basin. *GSA Bulletin* 111, 974-991.
- Hughes, S., Luetgert, J. H., 1992. Crustal structure of the southeastern Grenville Province, northern New York State and eastern Ontario. *J. Geophys. Res.* 97, 17,455-17,479, doi:10.1029/92JB01793.
- Hynes, A., Rivers, T., 2010. Protracted continental collision - evidence from the Grenville Orogen. *Can. J. Earth Sci.* 47, 591-620, doi:10.1139/E10-003

- Intera Engineering Limited 2008. Phase 2 Geoscientific Site Characterization Plan, OPG's Deep Geologic Repository for Low and Intermediate Level Waste. Report, OPG Report # 00216-REP-03902-00006-R00.
- Intera Engineering Limited 2009. Regional and site geological frameworks –proposed Deep Geologic Repository, Bruce County, Ontario. GeoHalifax2009.
- Ji, S., Rondenay, S., Mareschal, M., Senechal, G., 1996. Obliquity between seismic and electrical anisotropies as a potential indicator of movement sense for ductile shear zones in the upper mantle. *Geology* 24, 1033-1036.
- Johnson, M. D., Armstrong, D. K., Sanford, B. V., Telford, P. G., Rutka, M. A., 1992. Paleozoic and Mesozoic Geology of Ontario, in: Thurston, P.C., Williams, H.R., Sutcliffe, R.H., Stott, G.M. (Eds.), *Geology of Ontario*, Ontario Geological Survey , Special Volume 4, Part 2, pp. 907 – 1008.
- Kellett, R.L., Barnes, A. E., Rive, M., 1994. The deep structure of the Grenville Front: a new perspective from western Quebec. *Can. J. Earth Sci.* 31, 282–292.
- Kellett, R.L., Mareschal, M., Kurtz, R. D., 1992. A model of lower crustal electrical anisotropy for the Pontiac Subprovince of the Canadian Shield. *Geophys. J. Int.* 111, 141-150.
- Ketchum, I. W. F, Davidson, A., 2000. Crustal architecture and tectonic assembly of the Central Gneiss Belt, southwestern Grenville Province, Canada: A new interpretation. *Can. J. Earth Sci.* 37, 217-234.
- Kinsman, A., Parrish, R. R., 1990. U-Pb age of detrital zircons from five quartzites in the Central Metasedimentary Belt of the Grenville Province: in Program with abstracts, Geological Association of Canada-Mineralogical Association of Canada, 15, A70.
- Klein, G. deV., Hsui, A.T., 1987. Origin of Cratonic Basins. *Geology* 15, 094-1098.
- Krogh, T. E., 1994. Precise U-Pb ages for Grenvillian and pre-Grenvillian thrusting of Proterozoic and Archean metamorphic assemblages in the Grenville Front Tectonic Zone, Canada. *Tectonics* 13, 963-982.
- Kumarapeli, P. S., 1985. Vestiges of Iapetan rifting in the craton west of Appalachians. *Geoscience Canada* 12, 54 – 59.
- Kurtz, R. D., Garland, G. D., 1976. Magnetotelluric Measurements in Eastern Canada. *Geophys. J. R. Astr. Soc.* 45, 321-347.
- Kurtz, R. D., 1982. Magnetotelluric interpretation of crustal and mantle structure in the Grenville Province. *Geophys. J. R. Astr. Soc.* 70, 373-397.
- Kurtz, R.D., Niblett, E.R., Craven, J.A., Stevens, R.A., Macnae, J.C., 1988. Electromagnetic studies over the Kapuskasing Structural Zone, in: Lithoprobe Kapuskasing Transect, pp. 167–176, Lithoprobe Report 4.
- Ludden, J., Hynes, A., 2000. The Lithoprobe Abitibi-Grenville transect: two billion years of crust formation and recycling in the Precambrian Shield of Canada. *Can. J. Earth Sci.* 37, 459-476.
- Ludden, J., Hubert, C., Gariépy, C., 1986. The tectonic evolution of the Abitibi greenstone belt of Canada. *Geological Magazine* 123, 153–166.
- Lumbers, S.B., 1975. Geology of the Bruwash area: Ontario division of mines. Geological Report 116, 160.
- Lumbers, S.B., 1978. Geology of the Grenville front tectonic zone in Ontario: in Toronto '78, Field Trips Guidebook, Geological Society of America - Geological Association of Canada - Mineralogical Association of Canada, 347-361.

- Mareschal J.C., Jaupart, C., 2004. Variations of surface heat flow and lithospheric thermal structure beneath the North American craton. *Earth Planet. Sci. Lett.* 223, 65 – 77.
- Mareschal, J.C., Jaupart, C., Gariépy, C., Cheng, L.Z., Guillou-Frottier, L., Bienfait, G., Lapointe, R., 2000. Heat flow and deep thermal structure near the southeast edge of the Canadian Shield. *Can. J. Earth Sci.* 37, 399-414.
- Mareschal, M., Kellett, R. L., Kurtz, R. D., Ludden, J. N., Ji, S., Bailey, R. C., 1995. Archean cratonic roots mantle shear zones and deep electrical anisotropy. *Nature London* 375, 134-137.
- Mareschal, M., Kurtz, R.D., Chouteau, M., Chakridi, R., 1991. A magnetotelluric survey on Manitoulin Island and Bruce Peninsula along GLIMPCE seismic line J: black shales mask the Grenville Front. *Geophys. J. Int.* 105, 173–183.
- Mazurek, M., 2004. Long-term used nuclear fuel waste Management - geoscientific review of the sedimentary sequence in southern Ontario. Institute of Geological Sciences University of Bern, Switzerland Technical Report TR 04-01.
- McEachern, S. I., van Breemen, O., 1993. Age of deformation within the Central Metasedimentary Belt boundary thrust zone, southwest Grenville orogen: Constraints on the collision of the Mid-Proterozoic Elzevir terrane. *Can. J. of Earth Sci.* 30, 1155-1165.
- Mereu, R. F., Wang, D., Kuhn, O., Forsyth, D.A., Green, A. G., Morel, P., Buchbinder, G.G.R., Crossley, D., Schwarz, E., duBerger, R., Brooks, C., Clowes, R., 1986. The 1982 COCRUST seismic experiment across the Ottawa-Bonnechere graben and Grenville Front in Ontario and Quebec. *Geophys. J. R. Astron. Soc.*, 84, 491–514, doi:10.1111/j.1365-246X.1986.tb04368.x.
- Meyer, H. O. A., Waldman, M. A., Garwood, B. L., 1994. Mantle xenoliths from kimberlite near Kirkland Lake, Ontario. *Can. Min.* 32, 295-306.
- Miller, R.R., 1983. Age and petrological relationships of some igneous-textured and gneissic alkaline rocks in the Haliburton-Bancroft area: Unpublished PhD thesis, University of Toronto, Toronto, Ontario, 345p.
- Moore, J. M., Thompson, P. H., 1980. The Flinton Group: a late Precambrian metasedimentary succession in the Grenville Province of eastern Ontario, *Can. J. Earth Sci.* 17, 1685-1707.
- Mortensen, J. K., Card, K. D., 1993. U-Pb age constraints for the magmatic and tectonic evolution of the Pontiac Subprovince, Quebec. *Can J. Earth Sci.*, 30, 1970-1980.
- Nunn, J.A., Sleep, N.H., 1984. Thermal contraction and flexure of intracratonic basins: a three-dimensional study of the Michigan Basin. *R. Astr. Soc. Geophys. J.* 76, 587-635.
- O'Dowd, C.R., Eaton, D., Forsyth, D., Asmis, H.W.. 2004. Structural fabric of the Central Metasedimentary Belt of southern Ontario, Canada, from deep seismic profiling. *Tectonophysics* 388, 145– 159.
- Percival, J. A., 1989. A regional perspective of the Quetico metasedimentary belt, Superior Province, Canada. *Can. J. Earth Sci.* 26, 677-693.
- Percival, J. A., Bleeker, W., Cook, F. A., Rivers, T., Ross, G., van Staal, C., 2004. PanLITHOPROBE Workshop IV: intra-orogen correlations and comparative orogenic anatomy. *Geoscience Canada* 31, 23-39.

- Philpotts, A. R., 1978. Rift associated igneous activity in eastern North America, in: Neuman, E. R., Ramberg, I. B. (Eds.), *Petrology and Geochemistry of Continental Rifts*, Reidel, PP. 133- 154.
- Pinet, C., Jaupart, C., Mareschal, J. -C., Gariépy, C., Bienfait, G., Lapointe, R., 1991. Heat flow and structure of the lithosphere in the eastern Canadian Shield. *J. Geophys. Res.* 96, 19,941 – 19,963.
- Pye, E. G., Naldrett, A. J., Giblin, P. E., 1984. Geology and ore deposits of the Sudbury structure. Ontario Geological Survey, Special Volume 1, 603p.
- Rankin, D., 1976. Appalachian salients and recesses: Late precambrian continental breakup and the opening of the Iapetus Ocean. *J. Geophys. Res.* 81, 5605-5619.
- Rive, M., Pintson, H., Ludden, J. N., 1990. Characteristics of late Archean plutonic rocks from the Abitibi and Pontiac subprovinces, Superior province, Canada, in: Rive, M., Verpaerst, P., Gagnon, Y., et al. (Eds.), *The northwestern Quebec polymetallic belt: a summary of 60 years of mining exploration*. Canadian Institute of Mining, Special Volume 43, pp. 65-91.
- Rivers, T., 1997. Lithotectonic elements of the Grenville Province: review and tectonic implications. *Precambrian Res.* 86, 117-154.
- Rivers, T., Corrigan, D., 2000. Convergent Margin on southeastern Laurentia during the Mesoproterozoic: tectonic implications. *Can. J. Earth Sci.* 37, 359-383.
- Rivers, T., Culshaw, N., Hynes, A., Indares, A., Jamieson, R., Martignole, J., 2012. The Grenville orogen- A Post-Lithoprobe Perspective in: Percival, J.A., Cook, F.A., Clowes, R. M. (Eds.) *Tectonic Styles in Canada: The Lithoprobe Perspective*, Geological Association of Canada, Special paper 49, pp 97-236.
- Rivers, T., Martignole, J., Grower, C. P., Davidson, A., 1989. New tectonic subdivisions of the Grenville Province, southeast Canadian Shield. *Tectonics* 8, 63- 84.
- Rivers, T., van Gool, J.A.M., Connelly, J. N., 1993. Contrasting tectonic styles in the northern Grenville Province: implications for dynamics of orogenic fronts. *Geology* 21, 1127-1130.
- Rondenay, S., Bostock, M. G., Hearn, T. M., White, D. J., Ellis, R. M., 2000. Lithospheric assembly and modification of SE Canadian Shield: Abitibi-Grenville teleseismic experiment. *J. Geophys. Res.* 105, 13735-13754.
- Roy, B., Mereu, R. F., 2000. Applications of seismic pattern recognition and gravity inversion techniques to obtain enhanced subsurface images of the Earth's crust under the Central Metasedimentary Belt, Grenville Province, Ontario. *Geophys. J. Int.* 143, 735-751.
- Sage, R. P., 1991. Alkaline rocks, carbonatite and kimberlite complexes of Ontario, Superior Province, in: Thurston, P.C., Williams, H.R., Sutcliffe, R.H., Stott, G.M. (Eds.), *Geology of Ontario*, Ontario Geological Survey, Special Volume 4, Part 1, pp. 683–709.
- Sanford, B. V., Grant, A. C. , Wade, J. A., Barss, M. S., 1979. Geology, eastern Canada and adjacent areas, Geological Survey of Canada, Map 1401A (4 sheets), scale 1 : 2 000 000.
- Sanford, B.V., 1993. St. Lawrence Platform—Geology, in: Stott, D.F., Aitkin, J.D. (Eds.), *Sedimentary cover of the craton of Canada*, v. D–1 of *The geology of North America: Boulder, Colo., Geological Society of America*, pp. 723–786.

- Schultz, A., Kurtz, R.D., Chave, A.D., Jones, A.G., 1993. Conductivity discontinuities in the upper mantle beneath a stable craton. *Geophys. Res. Lett.* 20, 2941–2944.
- Sénéchal, G., Rondenay, S., Mareschal, M., Guilbert, J., Poupinet, G., 1996. Seismic and electrical anisotropies in the lithosphere across the Grenville Front, Canada. *Geophys. Res. Lett.* 23, 2255–2258.
- Slagstad, T., 2003. Muskoka and Shawanaga domains, Central Gneiss Belt, Grenville Province, Ontario: Geochemical and geochronological constraints on pre-Grenvillian and Grenvillian geological evolution. Unpublished PhD thesis, Dalhousie University, Halifax, Nova Scotia, 336p.
- Sleep, N. H., 1990. Montereyan hotspot track: a long-lived mantle plume. *J. Geophys. Res.* 95, 21,983–21,990.
- Sleep, N.H., 1971. Thermal Effects of the Formation of Atlantic Continental Margins by Continental Break up. *Geophys. J. R. Astr. Soc.* 24, 325–350.
- Sleep, N.H. and Nunn, J.A. 1980. Flexure of a rheologically stratified lithosphere, in: Nemat-Nasser, S. (Ed.), *Solid Earth Geophysics and Geotechnology*, Applied Mechanics Division, American Society of Mechanical Engineers 42, pp. 53–60.
- Sleep, N.H., Snell, N.S., 1976. Thermal contraction and flexure of mid-continent and Atlantic marginal basins. *Geophys. J. R. Astr. Soc.* 45, 125–154.
- Sloss, L. L., 1988. Tectonic evolution of the craton in Phanerozoic time, in: Sloss, L.L. (Ed.), *Sedimentary cover—North American craton*, v. D–2 of *The geology of North America: Boulder, Colo.*, Geological Society of America, pp. 25–51.
- Thurston, P.C., 1991. Archean Geology of Ontario: Introduction; in: Thurston, P.C., Williams, H.R., Sutcliffe, R.H., Stott, G.M. (Eds.), *Geology of Ontario*, Ontario Geological Survey, Special Volume 4, Part I, pp. 73 - 81.
- Tuccillo, M. E., Mezger, K., Essene, E. J., van der Pluijm, B. A., 1992. Thermobarometry, geochronology and the interpretation of P-T-t data in the Britt domain, Ontario Grenville Orogen, Canada. *J. Petrology* 33, 1225–1259.
- van Breemen, O., Davidson, A., Loveridge, W. D., Sullivan, R.W., 1986. U-Pb zircon geochronology of Grenville tectonites, granulites and igneous precursors, Parry Sound, Ontario, in: Baer, A. J. (Ed.), *The Grenville Province*, Geological Association of Canada Special Paper 31, pp. 192–207.
- van der Lee, S., Nolet, G., 1997. Upper mantle S velocity structure of North America, *J. Geophys. Res.* 102, 22,815–22,838, doi:10.1029/97JB01168.
- White, D.J., Forsyth, D. A., Asudeh, I. A., Carr, S. D., Wu, H., Easton, R. M., Mereu, R. F., 2000. A seismic-based cross-section of the Grenville Orogen in Southern Ontario and western Quebec. *Can. J. Earth Sci.* 37, 183–192.
- Whitmeyer, S.J., Karlstrom, K.E., 2007. Tectonic model for the Proterozoic growth of North America. *Geosphere* 3, 220–259. doi: 10.1130/GES00055.1
- Williams, H.R., Stott, G.M. , Thurston, P. C., Sutcliffe, R. H., Bennett, G., Easton, R. M., Armstrong, D. K., 1992. Tectonic evolution of Ontario: summary and synthesis, *Geology of Ontario*, 25, 1255–1346.
- Wodicka, N., Parrish, R. R., Jamieson, R. A., 1996. The Parry Sound domain: A far travelled allochthon? New evidence from U-Pb zircon geochronology. *Can. J. Earth Sci.* 33, 1087–1104.
- Wynne-Edwards, H. R., 1967. Westport map area with special emphasis on the Precambrian rocks. Geological Survey of Canada, Memoir 346, 142 p.

- Wynne-Edwards, H. R., 1972. The Grenville Province, in Price, R.A. and Douglas, R.J.W., eds., Variations in tectonic styles in Canada: Geological Association of Canada Special Paper 11, 263-334.
- Yuan, H., Romanowicz, B., 2010. Lithospheric layering in the North American craton. *Nature*, 466, 1063-1069, doi:10.1038/nature09332.
- Yuan, H., Romanowicz, B., Fischer, K.M., Abt, D., 2011. 3-D shear wave radially and azimuthally anisotropic velocity model of the North American upper mantle. *Geophys. J. Int.*, 184, 1237-1260, doi:10.1111/j.1365-246X.2010.04901.x
- Zhang, J., Frederiksen, A.W., 2013. 3-D crust and mantle structure in Southern Ontario, Canada via receiver function imaging. *Tectonophysics* 608, 700-712.
- Zhang, P., Chouteau, M., Mareschal, M., Kurtz, R. D., Hubert, C., 1995. High frequency magnetotelluric investigation of crustal structure in north-central Abitibi, Quebec, Canada. *Geophys. J. Int.* 120, 406-418.

Chapter 4

Magnetotelluric Data

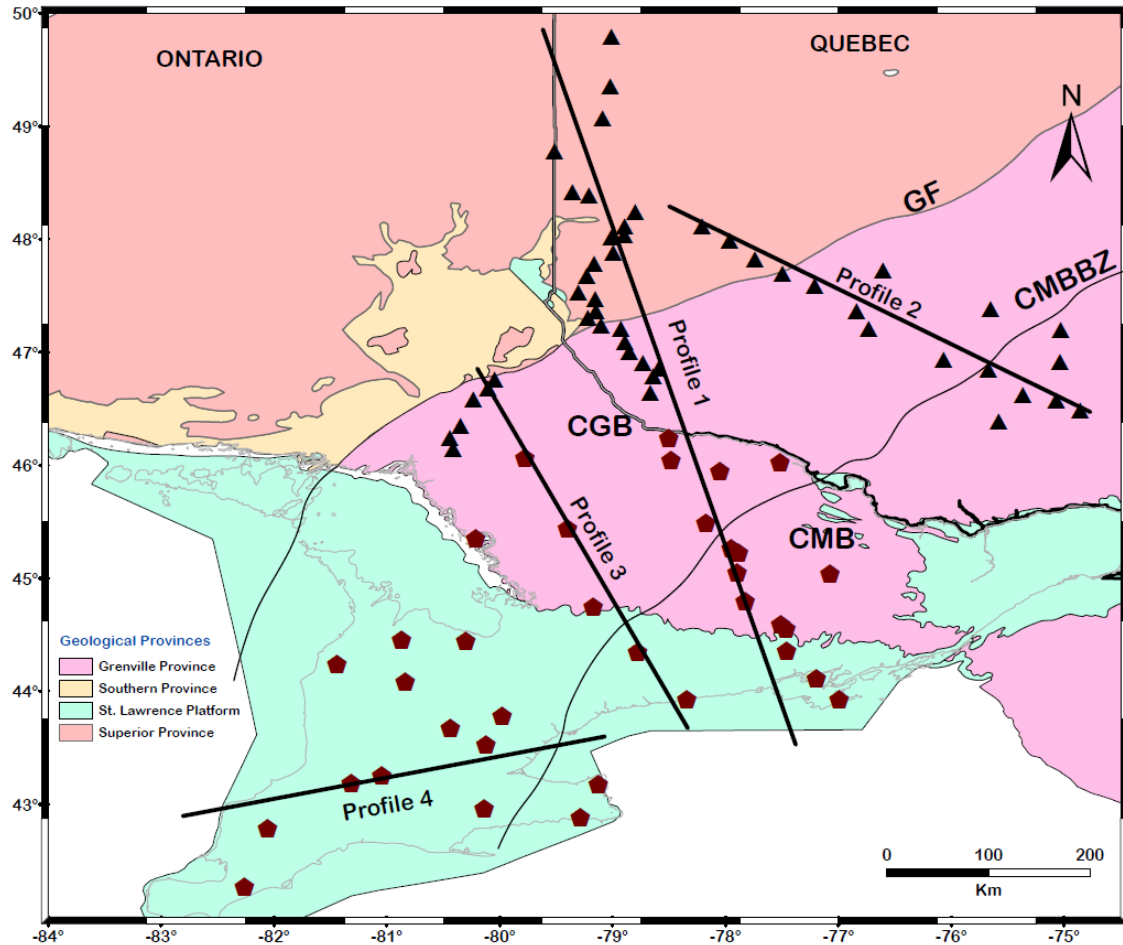
4.1 Introduction

A total of 89 MT sites distributed across the southern Ontario, are analysed in this study. These sites, collected in the LITHOPROBE Abitibi-Grenville transect and POLARIS southern Ontario array, are subdivided into four profiles (Fig. 4.1). Profiles 1 and 3 combine data from both projects; Profile 2 contains mainly LITHOPROBE sites whereas Profile 4 includes only POLARIS MT sites. Profiles 1, 2 and 3 are located on the exposed part of the Canadian Shield whereas sites along Profile 4 are located on the Phanerozoic sedimentary cover.

4.2 LITHOPROBE MT data

The LITHOPROBE Abitibi-Grenville transect, part of Canada's National Geoscience Project, crossed the margin between the Archean Superior and Proterozoic Grenville provinces. It was a regional geological and geophysical study that focused on understanding the growth and stabilization of the Precambrian Shield of Canada (Ludden and Hynes, 2000). As part of this study, between 1990 and 1992, remote-referenced broadband MT recordings, with durations of at least 24 hours, were made at a total of 163 sites (Kellett et al., 1992; Boerner et al., 2000). The data collection and analysis have been described by Kurtz et al. (1993), Schultz et al. (1993), Kellett et al. (1992, 1994), Boerner et al. (1994), Mareschal et al. (1995), Zhang et al. (1995), Ji et al. (1996), Senechal et al. (1996) and Frederiksen et al. (2006). A summary of the EM results from this transect is provided in Boerner et al. (2000).

Figure 4.1. Map of MT data sites and profiles defined for this research. It shows the POLARIS (red pentagons) and LITHOPROBE Abitibi-Grenville-Transect (black triangles) MT sites. GF = Grenville Front, CMBBZ = Central Metasedimentary Belt Boundary Zone, CGB = Central Gneiss Belt, CMB = Central Metasedimentary Belt, FAB = Frontenac-Adirondack belt. Shapefile obtained from Geological Survey of Canada.



In order to extend the area of coverage of the POLARIS MT data and to understand the tectonic structures from the Grenville Province through the Grenville Front into the Superior Province, a total of 58 LITHOPROBE broadband MT sites (black triangles in Fig. 4.1) are being combined with the POLARIS sites in this study. This addition extends the study area of this research into the Abitibi subprovince of the Superior Province and the Western Quebec section of the Grenville Province (Fig. 4.1).

In this study, modern depth-based geoelectric strike analysis of the MT responses is employed. Two-dimensional modeling, isotropic and anisotropic inversion models of these MT data are also being produced for the first time.

4.3 POLARIS MT data

In the POLARIS project, MT data were collected at a total of 41 sites within the southern Ontario POLARIS array between 2002 and 2005 (Eaton et al., 2005). These sites include 20 broadband MT (BBMT; 10^{-3} - $>10^3$ s), 7 long period MT (LMT; 1 - $>10^4$ s) and 14 combined LMT+BBMT sites distributed across the Grenville Province and overlying Phanerozoic rocks in southern Ontario (red pentagons in Fig. 4.1). The locations of the sites, type of data collected and site names are provided in Table 4.1.

4.3.1 Data Acquisition

The LMT dataset collected during the 2002 POLARIS southern Ontario survey was described in Ferguson et al. (2003). Most of the sites were located as close as possible to POLARIS seismograph stations. The data were recorded using the LiMS instruments belonging to the Geological Survey of Canada. Recordings were made using a maximum sampling rate of 1 s and over a two-week recording durations. The resulting data allow the determination of the MT response for the period range of 4 s to ~ 3000 s. Initial examination of the MT responses indicates that the responses are best defined for the period range from 10 s to 1000 s.

The 2005 LMT survey was carried out at POLARIS seismograph sites not included in the 2002 measurements and at additional sites. The survey was aimed at obtaining MT responses with frequencies sufficiently low to image the lithospheric

mantle. POLARIS NIMS (Narod Intelligent Magnetotelluric Systems) MT field equipment was used to record the long period data with durations varying from ten to twenty days. Although the noise level is relatively high in this region, the resulting data produced MT responses for the period range of 10 s to ~3000 s. Preliminary data analysis indicated that good MT responses at a period range of 40 s to 2000 s were obtained at most sites.

Recordings at the BBMT sites were made using POLARIS Phoenix Geophysics MTU-5A equipment, which utilizes induction coil magnetometers and non-polarizing Pb-PbCl₂ electrodes. 100 m electric field dipole lengths were used. Recordings were made for two nights at each site. At most sites, both BBMT and audiofrequency MT (AMT) data were collected. MTC50 MT coils were used to record the BBMT signals and AMTC30 AMT coils were used for the AMT recordings. All five electric and magnetic field components were recorded at each site and data recorded simultaneously at nearby sites were used for the remote-reference noise reduction (Gamble et al., 1979). Cultural noise levels are high throughout the relatively populous study area, and the MT responses at some sites are relatively noisy, particularly for the more conductive MT components.

4.3.2 Data Processing

All the LITHOPROBE broadband MT data were processed as part of that project and are available as electronic data interchange (edi) files containing either spectral or impedance data. The edi files are available in the Geological Survey of Canada archive. However, at the start of the present work, the POLARIS data were available only as time series and a robust spectral analysis was conducted as part of the study. The data were

Table 4.1. Location of POLARIS Southern Ontario (PSO) MT Recordings sites

MT Site	Location of MT Site	Type/Year	Town or Location
PSO001	43°56'04" 76°59'36"	BBMT (05), LMT (02)	Picton
PSO002	44°21'11" 78°46'47"	BBMT (05), LMT (02)	Lindsay
PSO003	45°03'21" 77°53'59"	BBMT (05), LMT (02)	Bancroft
PSO004	44°31'28" 77°38'06"	LMT (02)	Deloro
PSO005	43°56'01" 78°20'39"	BBMT (05), LMT (02)	Wesleyville
PSO006	44°44'46" 79°10'20"	BBMT (05), LMT (02)	Sadowa
PSO007	45°14'10" 76°58'11"	LMT (02)	Centennial Lake
PSO008	45°17'36" 79°11'48"	LMT (02)	Huntsville
PSO009	45°29'19" 78°10'34"	BBMT (05), LMT (02)	Whitney
PSO010	45°39'06" 77°30'27"	LMT (02)	Foy Provincial Park
PSO011	44°46'13" 76°15'54"	LMT (02)	Muphy's Point Provincial Park
PSO012	45°16'07" 77°56'56"	BBMT (05)	Maynooth
PSO013	45°02'24" 77°04'31"	BBMT (05)	Plevna
PSO014	44°35'52" 77°30'44"	BBMT (05)	El Dorado
PSO015	45° 26'34" 79°24'00"	BBMT (05)	Buck Lake
PSO016	43° 46'58" 79°58'38"	BBMT (05)	Terra Cotta
PSO017	43° 40'36" 80°26'11"	BBMT (05)	Elora Gorge
PSO018	42°53'05" 79°17'11"	BBMT (05)	Port Colbourne
PSO019	43°11'05" 79°07'39"	BBMT (05), LMT (05)	Niagara
PSO020	43°11'32" 81°19'01"	BBMT (05)	Elginfield (London)
PSO021	44°14'36" 81°26'34"	BBMT (05)	Tiverton
PSO022	45°57'16" 78°03'03"	BBMT (05), LMT (05)	Algonquin park
PSO023	46°14'52" 78°30'06"	BBMT (05), LMT (05)	Deux Riviere
PSO024	44°26'59" 80°18'04"	LMT (05)	Collingwood
PSO025	45°21'27" 80°12'53"	BBMT (05), LMT (05)	Killbear Park
PSO026	46°03'59" 79°46'36"	BBMT (05), LMT (05)	Restoule Park
PSO027	46°01'56" 77°31'00"	BBMT (05)	Chalk River
PSO028	46°02'42" 78°28'56"	BBMT (05)	Brent
PSO029	43°31'43" 80°07'18"	LMT (05)	Guelph
PSO031	42°58'12" 80°08'23"	BBMT (05), LMT (05)	Hagersville
PSO032	42°16'12" 82°15'40"	BBMT (05)	Chatham
PSO033	41°45'13" 82°37'46"	BBMT (05), LMT (05)	Pelee Island
PSO034	42°47'40" 82°03'25"	BBMT (05), LMT (05)	Oil Springs
PSO035	43°15'36" 81°02'39"	BBMT (05)	Stratford
PSO036	44°05'12" 80°50'13"	BBMT (05)	Durham
PSO037	44°27'40" 80°52'14"	BBMT (05)	Chatsworth
PSO038	45°13'36" 77°52'57"	BBMT (05)	Lake St Peter
PSO039	44°48'10" 77°49'36"	BBMT (05)	Coe Hill
PSO040	44°33'12" 77°27'48"	BBMT (05)	Bannockburn
PSO041	44°21'34" 77°27'29"	BBMT (05)	West Huntington
PSO042	44°07'11" 77°11'43"	BBMT (05)	North Port

processed and analyzed using Phoenix processing software (SSMT 2000) for the BBMT and AMT responses. Estimation of the LMT responses was done with the robust processing algorithm developed by Egbert (1997). There were some challenges involved in the processing of the data due to large sources of artificial noise in the study area which affected the data quality.

The SSMT 2000 is a robust remote-reference processing algorithm that is based on a coherence-sorted cascade decimation method (Wight and Bostick, 1980) and method 6 of Jones et al. (1989). The algorithm is robust cascade decimation, which uses a jackknife approach to determine the harmonic that produces a minimum in variances of the off-diagonal impedances when it is omitted. Processing is repeated using fewer harmonics each time until the variance can not be reduced further. The resulting ministacks from each of the iterations are then combined using a jackknife approach to minimize variances of the final estimates of the off-diagonal impedances. Standard statistical methods with Gaussian assumptions were used to estimate the errors. Responses were obtained to a maximum period of at least 1000 s for most sites, and 2000 s for some sites.

In terms of the the robust parameters used for data processing, the minimum rho variance level was set to 0.8 while the maximum fraction of crosspower estimates was set to 0.2. The minimum coherence level was set at 0.75 and the maximum fraction of crosspower estimates to be rejected was set to 0.35. The rho variance weighting function was used because it applies more weight to data with smaller error bars. The MT-Editor (hosted within SSMT 2000) was used to edit, merge (e.g., BBMT+LMT) and export the data as edi files.

The robust processing code is an automated, statistically robust scheme for time series processing. It uses an iteratively re-weighted least squares algorithm to scale the field components by their errors. Multivariate statistical methods are used to identify and remove coherent and incoherent noise. It does not allow the impedance estimates to be contaminated by outliers. The automatic procedure employed by this method produces stable estimates of the transfer function with reasonable errors (Egbert and Booker, 1986; Egbert, 1997).

4.4 References Cited

- Boerner, D. E., Kurtz, R. D., Craven, J. A., 2000. A summary of electromagnetic studies on the Abitibi-Grenville transect. *Can. J. Earth Sci.* 37, 427-437.
- Boerner, D.E., Kellett, R., Mareschal, M., 1994. Inductive source EM sounding of the Sudbury Structure. *Geophys. Res. Lett.* 21, 943-946.
- Eaton, D. W., Adams, J., Asudeh, I., Atkinson, G. M., Bostock, M. G., Cassidy, J. F., Ferguson, I. J., Samson, C., Snyder, D. B., Tiampo, K. F., Unsworth, M. J., 2005. Geophysical arrays to investigate lithosphere and earthquake hazards in Canada. *EOS Trans AGU* 86, 169-173.
- Egbert, G.D., Booker, J. R., 1986. Robust estimation of geomagnetic transfer functions: *Geophys. J. R. Astron. Soc.* 87, 173-194.
- Egbert, G.D., 1997. Robust multiple-station Magnetotelluric data processing. *Geophys. J. Int.* 130, 475-496.
- Ferguson, I., Wenham, J., Ma, X., Jones, A., Spratt, J., Evans, S., 2003. POLARIS 2002 Magnetotelluric Survey in Southern Ontario: Report of Field-work and Preliminary Data Examination, 18 pp.
- Frederiksen, A.W., Ferguson, I. J., Eaton, D., Miong, S. K., Gowan, E., 2006. Mantle fabric at multiple scales across an Archean-Proterozoic boundary, Grenville Front, Canada. *Phys. Earth Planet. Int.* 158, 240-263.
- Gamble, T.D., Goubau, W. M., Clarke, J., 1979. Magnetotellurics with a remote magnetic reference. *Geophysics* 44, 53- 68. doi:10.1190/1.1440923.
- Ji, S., Rondenay, S., Mareschal, M., Senechal, G., 1996. Obliquity between seismic and electrical anisotropies as a potential indicator of movement sense for ductile shear zones in the upper mantle. *Geology* 24, 1033-1036.
- Jones, A. G, Chave, A. D., Egbert, G., Auld, D., Bahr, K., 1989. A Comparison of Techniques for Magnetotelluric Response Function Estimation. *J. Geophys. Res.* 94, 14,201-14,213.
- Kellett, R.L., Barnes, A. E., Rive, M., 1994. The deep structure of the Grenville Front: a new perspective from western Quebec. *Can. J. Earth Sci.* 31, 282-292.
- Kellett, R.L., Mareschal, M., Kurtz, R. D., 1992. A model of lower crustal electrical anisotropy for the Pontiac Subprovince of the Canadian Shield. *Geophys. J. Int.* 111, 141-150.
- Kurtz, R.D., Craven, J.A., Niblett, E.R., Stevens, R.A., 1993. The conductivity of the crust and mantle beneath the Kapuskasing Uplift: electrical anisotropy in the upper mantle. *Geophys. J. Int.* 113, 483 - 498.
- Ludden, J., Hynes, A., 2000. The Lithoprobe Abitibi-Grenville transect: two billion years of crust formation and recycling in the Precambrian Shield of Canada. *Can. J. Earth Sci.* 37, 459-476.
- Mareschal, M., Kellett, R. L., Kurtz, R. D., Ludden, J. N., Ji, S., Bailey, R. C., 1995. Archean cratonic roots mantle shear zones and deep electrical anisotropy. *Nature* 375, 134-137.
- Schultz, A., Kurtz, R.D., Chave, A.D., Jones, A.G., 1993. Conductivity discontinuities in the upper mantle beneath a stable craton. *Geophys. Res. Lett.* 20, 2941-2944.

- Sénéchal, G., Rondenay, S., Mareschal, M., Guilbert, J., Poupinet, G., 1996. Seismic and electrical anisotropies in the lithosphere across the Grenville Front, Canada. *Geophys. Res. Lett.* 23, 2255-2258.
- Wight, D.E., Bostick, F.X., 1980. Cascade decimation- A technique for real time estimation of power spectra: *Proc. IEEE Conf. Acoustic, Speech, Signal Processing* 5,626-629.
- Zhang, P., Chouteau, M., Mareschal, M., Kurtz, R. D., Hubert, C., 1995. High frequency magnetotelluric investigation of crustal structure in north-central Abitibi, Quebec, Canada. *Geophys. J. Int.* 120, 406-418.

Chapter 5

Crustal and lithospheric scale structures of the Precambrian Superior-Grenville margin

This chapter has been published in the *Tectonophysics* journal in co-authorship with a research colleague (A. G. Jones) and my advisor (I. J. Ferguson). I was responsible for all the data processing, analysis, modeling as well as the writing of this first draft of the manuscript while the co-authors reviewed it. Alan Jones provided the STRIKE code used in this study.

Adetunji, A.Q., Ferguson, I.J. and Jones, A.G., 2014. Crustal and lithospheric scale structures of the Precambrian Superior-Grenville margin, *Tectonophysics* 614, 146-169, <http://dx.doi.org/10.1016/j.tecto2013.12.008>.

5.1 Abstract

The lithosphere beneath the margin of the Archean Superior and Proterozoic Grenville provinces was investigated with a northwest-southeast oriented, 650-km-long profile of 40 magnetotelluric stations. Dominant geoelectric strike azimuths of N45°E and N85°E were defined for the crust and the lithospheric mantle respectively. A 2-D isotropic resistivity model derived using the crustal strike images resistive Laurentian margin rocks dipping southeast to the base of the crust, bounded to the northwest by the Grenville Front, and to the southeast by the Central Metasedimentary Belt Boundary Zone. The observation is in contrast to conductive mid to lower crust elsewhere in the region. A 2-D isotropic resistivity model determined using the lithospheric mantle strike azimuth reveals an extremely resistive region in the upper 100 km of the mantle lithosphere of the northern Grenville Province. The geometry of this body, which includes a well-defined base and southeast dip, suggests that it is Superior lithosphere. A sub-vertical conductor, located approximately 50 km along strike from the Mesozoic Kirkland Lake and Cobalt kimberlite fields, is interpreted to be due to re-fertilization of an older mantle scar. The resistivity model includes a horizontal conductor at 160 km depth beneath the southern Superior Province that is possibly the resistivity signature of the lithospheric-asthenospheric boundary.

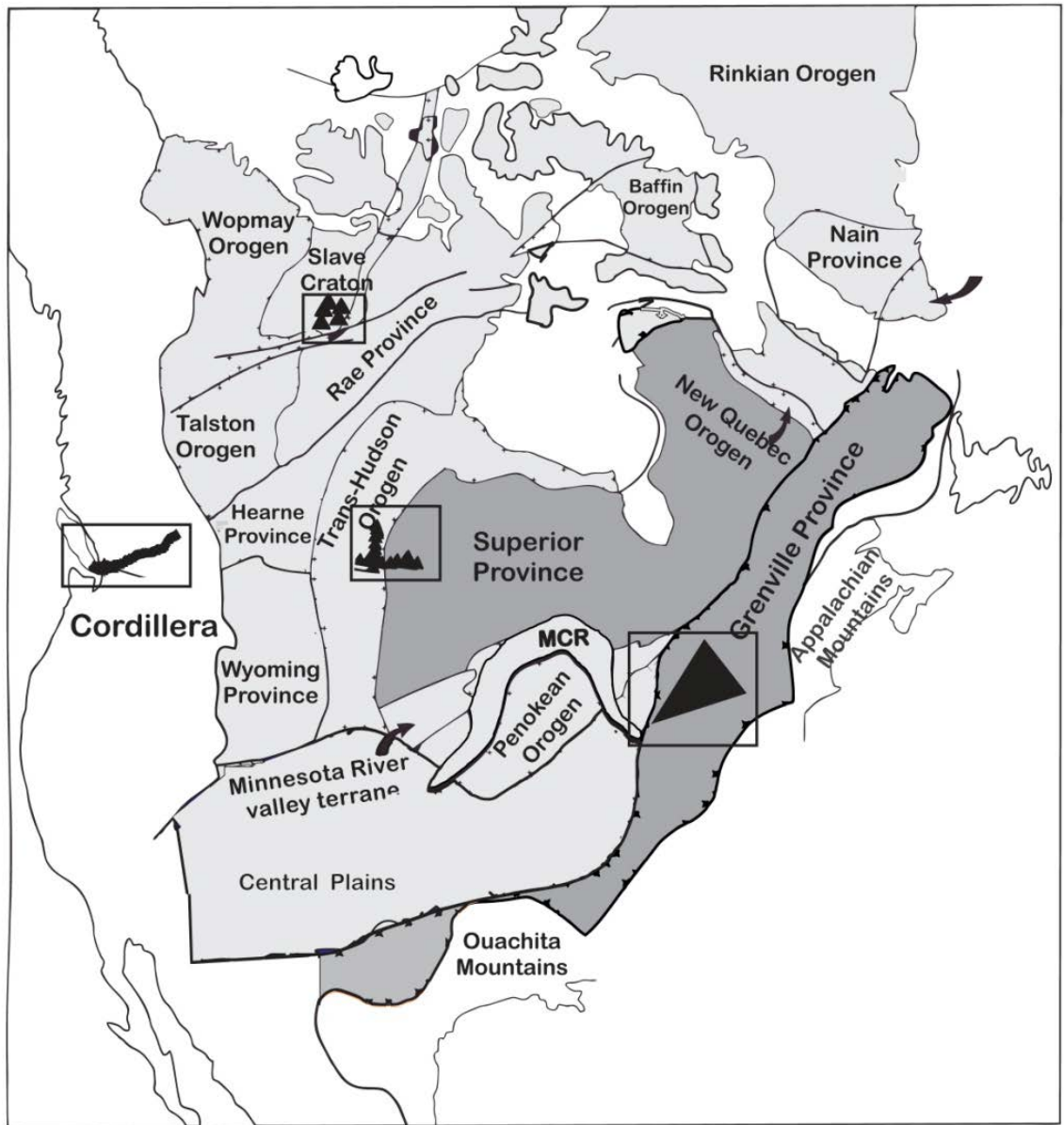
Keywords: Grenville Province, Superior Province, Magnetotellurics, Geoelectric Strike, Conductor, Lithosphere, Asthenosphere

5.2 Introduction

POLARIS (Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity) is a multi-institutional geophysical consortium focused on investigation of the structure and dynamics of the Earth's lithosphere, assessment of earthquake, and geomagnetic hazards across Canada, using the magnetotelluric (MT) method and earthquake seismology (Eaton et al., 2005). Between 2002 and 2005 seismic and MT data were collected in POLARIS study areas across Canada, including southern Ontario (Fig. 5.1). The focus of the Ontario MT study is to use deep electrical resistivity imaging to derive new geophysical and geological information on crustal and upper mantle structures and processes in the Grenville Province. MT results from the POLARIS projects in British Columbia, Manitoba, and southern Ontario have been presented by Soyer and Unsworth (2006), Gowan et al. (2009) and Frederiksen et al. (2006) respectively.

The MT method involves measuring the time-varying fluctuations of the natural electric and magnetic fields of the Earth and analyzing these data to map the spatial variation of the Earth's subsurface electrical properties. Signals with long periods (low frequencies) penetrate deeper into the Earth allowing the variation of resistivity with depth to be determined (Tikhonov, 1950; Cagniard, 1953). The subsurface resistivity structure can provide information on the processes that have formed and deformed a region (e.g., Jones and Craven, 1990; Nover et al., 1998; Jones, 1999; Spratt et al., 2009; Miensopust et al., 2011; Evans et al., 2011; Ferguson et al., 2012) and on the source of major conductivity anomalies (e.g., Jones and Savage, 1986; Ferguson et al., 2005). Enhanced electrical conductivity (reduced resistivity) in the crust is often linked to the

Figure 5.1. Initial deployments of POLARIS MT sites in Canada. The first four POLARIS deployments were in the Cordillera, Slave craton, Superior craton-Trans Hudson Orogen margin, and the Grenville Province. Tectonic background modified from Hoffman, (1989). MCR=Mid-continent rift.



presence and mobility of aqueous fluids (e.g., Hyndman et al., 1993; Glover and Vine, 1995). Graphite, interconnected grains of metallic oxides and sulfides are also important sources (e.g., Jones and Craven, 1990; Katsube and Mareschal, 1993; Li et al., 2003). There are several sources of enhanced electrical conductivity in the mantle, including graphite, hydrogen, partial melt, water and temperature variation (e.g., Karato, 1990; Mibe et al., 1998; Jones, 1999; Hirth et al., 2000; Ducea and Park, 2000; Constable, 2006; Korja, 2007; Yoshino et al., 2008; Muller et al., 2009; Selway 2014).

The electrical resistivity structure of the Grenville Front (GF) region and the southern subprovinces of the Superior craton has been examined in earlier MT studies (Boerner et al., 2000) but there have been very few MT studies in areas south of the GF (e.g., Kurtz, 1982). Previous results for the GF area were derived mainly from 1-D inversion and 2-D forward modeling. They showed that the upper crust is generally resistive, the middle to lower crust is relatively conductive, and the uppermost mantle is resistive (Kurtz, 1982; Kellett et al., 1992; Zhang et al., 1995).

Results from the previous MT studies have been interpreted as indicating anisotropic electrical resistivity within the mantle in the GF area (e.g., Kellett et al., 1992; Mareschal et al., 1995; Ji et al., 1996; Frederiksen et al., 2006). The interpretation of the observed azimuthally-dependent MT responses in terms of anisotropy was based on the absence of induction arrow responses that would be indicative of 2-D or 3-D structures. The anisotropy was constrained to lie between 45 and 150 km depth (Mareschal et al., 1995) and interpreted to be associated with southward extension of Archean Superior rocks beneath the Grenville Province (Kellett et al., 1992; Boerner et al., 2000). Ji et al.

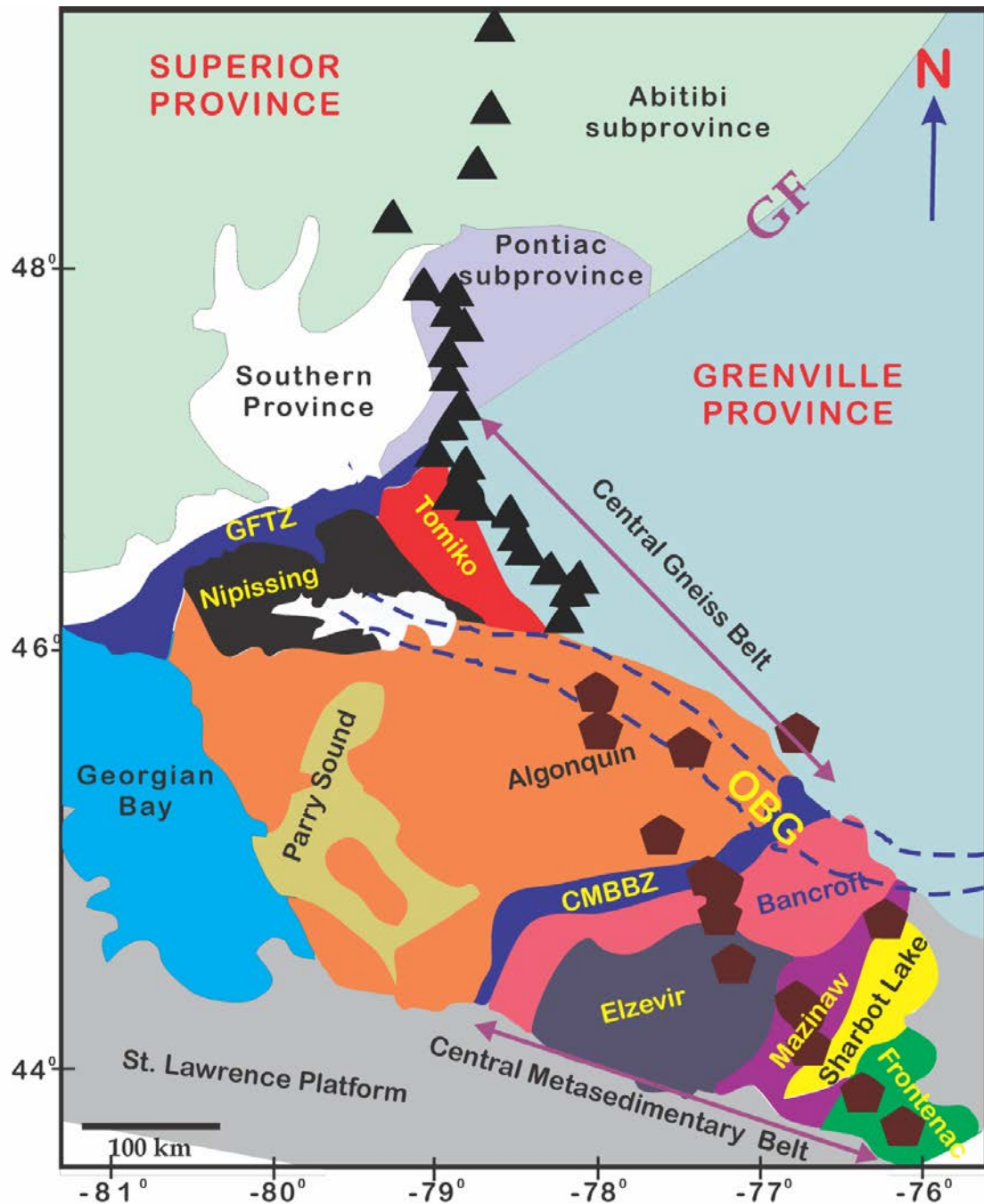
(1996) suggested that the anisotropy is caused by shape-preferred orientation of mantle minerals.

The present study area includes the Abitibi and Pontiac Subprovinces of the Archean Superior Province and the Central Gneiss Belt (CGB) as well as the Central Metasedimentary Belt (CMB) of the Proterozoic Grenville Province (Fig. 5.2). It describes the processing and interpretation of MT data along a profile oriented in a northwest-southeast direction that extends the earlier LITHOPROBE Abitibi-Grenville Transect MT acquisition (Boerner et al., 2000) with POLARIS MT (Eaton et al., 2005) stations. It contrasts with the prior MT studies of the Superior-Grenville margin (e.g., Kellett et al., 1992; Mareschal et al., 1995; Ji et al., 1996) by using full 2-D inversion methods to model the MT data. The profile of MT stations examined in this study provides the highest available density of MT sites crossing the Superior-Grenville boundary. However, additional MT data have been collected in the Superior-Grenville area in the Lithoprobe and POLARIS projects. Results from a larger-scale study area are examined in Adetunji et al. (in preparation) which also include a more detailed examination of local and regional geoelectric strike variation, tipper responses, and MT tensor distortion in the Superior-Grenville area.

5.3 Geological and Tectonic Setting

The Archean Superior Province mostly comprises rocks formed between 2.8 and 2.7 Ga (Hoffman, 1989). The Abitibi Subprovince represents a remnant from a larger Neoproterozoic granite–greenstone terrain that has a record of rapid crustal growth during the Neoproterozoic (Card, 1990; Percival et al., 2004). It is characterized by polycyclic

Figure 5.2. A simplified geotectonic map of the study area showing the location of the MT sites. Triangles represent the LITHOPROBE Abitibi-Grenville transect MT sites and polygons represent the POLARIS MT sites. Sub-units of the Superior and Grenville Provinces are labeled. GF=Grenville Front, GFTZ=Grenville Front Tectonic Zone, OBG=Ottawa Bonnechere Graben; CMBBZ=Central Metasedimentary Belt Boundary Zone.



volcanic stratigraphy overlain by a late-stage sequence of turbiditic sedimentary rocks and synorogenic clastic rocks. Numerous granitoid plutons intruded and deformed these rocks in several phases of folding and faulting (Card and Poulsen, 1998).

The Pontiac Subprovince is an Archean metasedimentary and plutonic terrane separated from the Abitibi Greenstone Belt to the north by the Cadillac Larder-Lake Fault Zone (CLLFZ); while to the south its boundary with the Grenville Province is marked by the southeast-dipping Grenville Front Tectonic Zone (GFTZ) (Card and Cieselski, 1986). It is characterized by supracrustal, metagraywacke and metasedimentary rocks together with a small metavolcanic belt (Dimroth et al., 1982; Card, 1990; Card and Poulsen, 1998).

The Grenville Province was formed during middle to Late Mesoproterozoic orogenic activity (Hanmer, 1988; Hoffman, 1989; Easton, 1992). The orogen followed a period of pre-Grenvillian continental-margin arc magmatism, accretionary, and collisional tectonic events (Carr et al., 2000). This collision and thrusting resulted in a first phase of high grade metamorphism and convergent deformation, which occurred between 1190 and 1140 Ma (Rivers et al., 1989; Corrigan and van Breemen, 1997). In southwestern Ontario, the Grenville orogen produced northwest-trending thrusting of a composite belt of continental crustal and magmatic rocks onto the margin of the Laurentian craton (Rivers, 1997; Carr et al., 2000). The collisional tectonics resulted in crustal thickening, and local over-thickening due to the ductile nature of the crust (Ludden and Hynes, 2000). This was followed by orogenic collapse and emplacement of mantle- and lower-crustal-derived magmas.

The GFTZ marks the northwestern boundary of the Grenville orogen and thrusting along this shear zone occurred late in the Grenvillian orogeny, at ca. 990 Ma (Haggart et al., 1993; Krogh, 1994). The GFTZ is a prominent southeast-dipping crustal-scale shear zone of amphibolite-facies rocks. Its boundaries are gradational and are deformed by cataclastic zones that extend into the adjacent Superior Province (Easton, 1992).

Division of the Grenville Province into the CGB in the north and the CMB to the south is based on the contrasting lithologies of these terranes (Easton, 1992) and we use these terminologies herein. More recently, the CGB has been referred to as part of pre-Grenvillian Laurentia and its margin; the constituent units of the CMB, with the exception of the Frontenac terrane, have been collectively termed the Composite Arc Belt (e.g., Carr et al., 2000).

The CGB represents the reworked margin of the pre-Grenvillian Laurentian craton (Carr et al., 2000). It is characterized by layered felsic gneiss and amphibolites, and upper amphibolite to granulite facies plutonic and supracrustal rocks (Easton, 1992). Several ductile shear zones that separate distinctive lithotectonic domains and sub-domains in the CGB show an indication of a northwest-directed low angle ductile thrusting (Davidson, 1986). This belt is subdivided into four (Nipissing, Algonquin, Tomiko and Parry Sound) domains (Fig. 5.2) based on contrasting lithologies, metamorphic grade and deformation histories. These domains represent imbricated panels of relatively low strain rocks bounded by anastomosing ductile shear zones (e.g., Davidson and Morgan, 1981; Davidson et al., 1982; Culshaw et al., 1983; Hanmer and Ciesielski, 1984).

The CMB lies to the southeast of the CGB from which it is separated by the Central Metasedimentary Boundary Belt Zone (CMBBZ). Shear indicators in this zone indicate thrusting of the CMB northwestwardly against the CGB (Easton, 1992; Davidson, 1998). The CMBBZ is characterized by alternating crystalline thrust sheets enveloped by anastomosing higher strain zones of porphyroclastic to migmatitic tectonites and marble tectonic melange (Hanmer et al., 1985; Hanmer and McEachern, 1992).

The CMB contains a major accumulation of carbonate, siliciclastic, plutonic, and volcanic rocks metamorphosed into greenschist to granulite facies during the Grenvillian orogeny (Easton, 1992; White et al., 2000). The rocks of the CMB are dominated by structural polyfolded domains with complex geometry (White et al., 2000). Based on different lithologies, metamorphic grade, deformation and geochronological histories, the CMB is subdivided into five domains (Fig. 5.2): the Bancroft, Elzevir, Mazinaw, Sharbot Lake and Frontenac terranes.

The Ottawa Bonnechere Graben (OBG) is a 60 km wide graben (Mereu et al., 1986) that lies along the border between Ontario and Quebec (Fig. 5.2). This graben has been interpreted as a failed arm of a triple junction formed during a late Precambrian to early Paleozoic event during which the Iapetus ocean opened (Kumarapeli, 1985; Kamo et al., 1995). It consists of deformed metasedimentary and metavolcanic rocks intruded by felsic to mafic plutons (Easton, 1992).

The youngest tectonic event associated with the study area is the interpreted Great Meteor hotspot (Crough, 1981; Sleep, 1990). The track of this hotspot was interpreted to be responsible for the emplacement of the Cretaceous Monteregian-White Mountain-New

England Seamounts Igneous Province (Crough, 1981; Sleep, 1990). It is also interpreted to be associated with the Mesozoic lamprophyric dykes and kimberlite magmatism at Rapide des Quinze (Ji et al., 1996) and Kirkland Lake (Meyer et al., 1994; Ji et al., 1996) areas of Pontiac Subprovince (Griffin et al., 2004; Faure et al., 2011).

5.4 MT Data

In the POLARIS project, MT data were collected at a total of 41 sites within the southern Ontario POLARIS array between 2002 and 2005 (Eaton et al., 2005). These sites include 20 broadband MT (BBMT), 7 long period MT (LMT) and 14 LMT+BBMT sites distributed across the Grenville Province in southern Ontario. The MT data examined in this study are from 40 Lithoprobe and POLARIS broadband MT sites on a northwest-southeast, 650-km-long profile (Fig. 5.2). Twenty-five Lithoprobe stations (triangles in Fig. 5.2) extend from the Abitibi Subprovince to the CGB while 15 POLARIS stations (pentagons in Fig. 5.2) cross part of the CGB and the entire CMB. The POLARIS sites also cross the Ottawa Bonnechere Graben. The data from the Lithoprobe Abitibi–Grenville Transect studies have been described in Boerner et al. (2000), Ji et al. (1996), Kellett et al. (1992), Mareschal et al. (1995), Senechal et al. (1996), and Zhang et al. (1995). The POLARIS data are described in Adetunji (2013) and Frederiksen et al. (2006).

A resistivity structure may be 1-D or horizontally-layered, 2-D with a geoelectric strike in which the resistivity is invariant or it may have a more complex 3-D form. The primary responses derived from the MT data are the apparent resistivity and impedance phase; both responses provide information on the underlying resistivity structure (Vozoff,

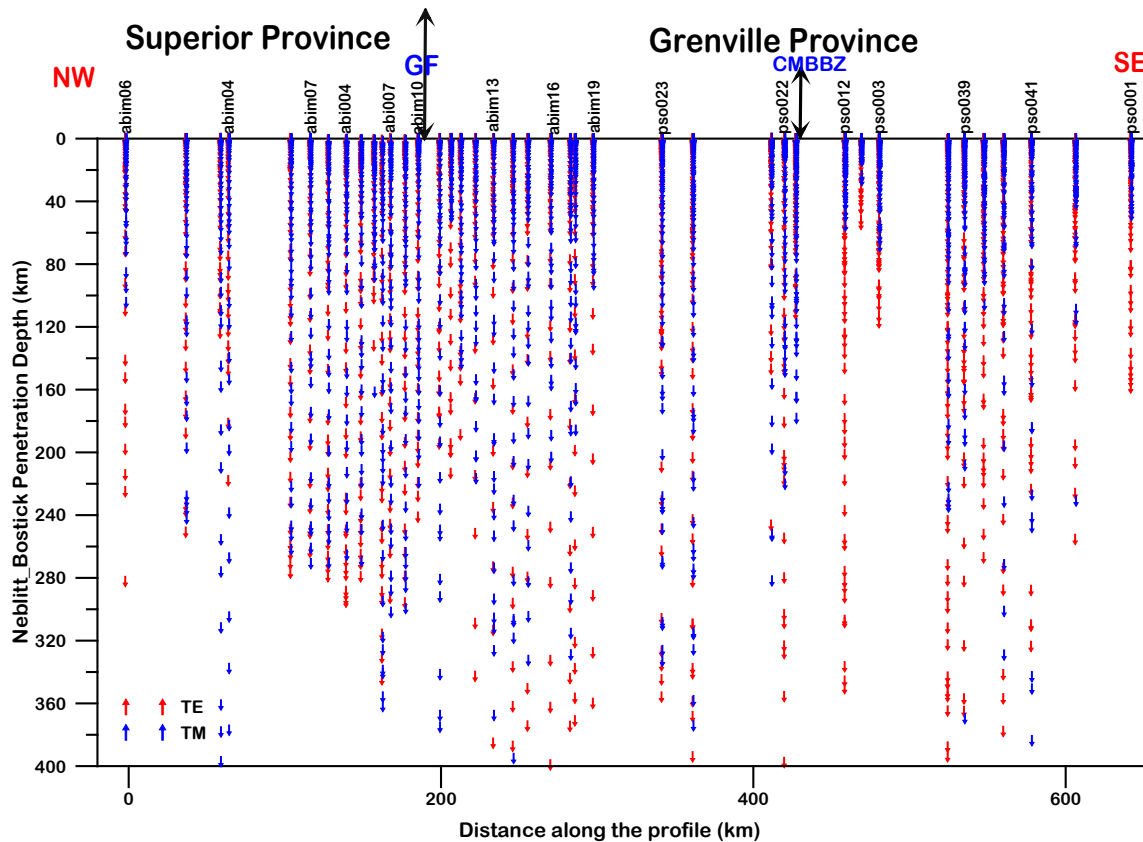
1991; Chave and Jones, 2012). The MT responses measured over a 2-D structure are the transverse electric (TE) mode in which the electric field is parallel to the geoelectric strike, and the transverse magnetic (TM) modes in which the electric field is perpendicular to the strike. The depth of penetration of MT signals depends on both the period of the signal and the local resistivity. The maximum depth of penetration of the responses in the present study was estimated for each site with the Niblett-Bostick depth transform (Niblett and Sayn-Wittgenstein, 1960; Jones, 1983) of the TE and TM modes of best-fitting regional 2-D responses. Although the estimated depths might be affected to some degree by 3-D effects, the Niblett-Bostick approximation provides a superior method for comparing related responses than a period-based data analyses (Jones, 2006). At all of the MT sites considered in the present study, the maximum period of responses is 1820 s and depending on the local resistivity, this period corresponds to a depth of between 100 km and 400 km (Fig. 5.3). The penetration depth differs at the same site for different modes, an effect discussed in Jones (2006).

5.5 Geoelectric dimensionality and strike analysis

Galvanic distortion of the electric field by near-surface inhomogeneities can mask the true dimensionality of the regional resistivity structure unless tensor decomposition methods are applied to remove its effects (e.g., Bahr, 1988, Groom and Bailey, 1989; McNeice and Jones, 2001; Caldwell et al., 2004; Jones, 2012). In this study the galvanic distortion and geoelectric strike were determined using the single-site STRIKE algorithm of McNeice and Jones (2001) that is based on the Groom and Bailey (GB) tensor

decomposition (Groom and Bailey, 1989, 1991). The GB method fits a regional 2-D MT response in the presence of in-phase frequency-independent electric field distortion.

Figure 5.3. The Niblett-Bostick penetration depth for the 40 sites on the profile. Each of the symbols represents the response at a particular period. Red color represents penetration depth for TE mode and blue color represents penetration depth for TM mode. GF=Grenville Front; CMBBZ=Central Metasedimentary Belt Boundary Zone.



McNeice and Jones (2001) extended the GB method to multi-site and multi-frequency fitting which allows for the determination of the optimal strike for a group of sites and/or a selected range of frequencies. A more recent modification involves the determination of strike azimuth for a specified depth range using Niblett-Bostick depth approximation

(e.g; Hamilton et al., 2006; Muller et al., 2009; Miensopust et al., 2011). This approach greatly reduces the mixing of strikes from different depths, which, because of the resistivity dependent depth of MT signal penetration, can occur in frequency-based studies (Jones, 2006). As indicated in Figure 5.3, there is a strong variation in the penetration depth of the MT responses along the profile, and this supports the use of depth-bands in the geoelectric strike analyses.

In the present study, depth is divided into nominal crustal (1–40 km), lithospheric mantle (40–200 km) and asthenospheric mantle (200–400 km) ranges. These depth bands are based on *a priori* information rather than MT results. The crustal depth range was defined so as to exclude near-surface structures, and the crust-mantle depth was based on seismic estimates of the depth to the Moho (Eaton et al., 2006; White et al., 2000). The lithospheric mantle depth range is based on typical lithospheric thickness defined by Artemieva (2009), while the asthenospheric depth range includes all deeper-penetrating periods. The results of the geoelectric strike estimation, presented in this paper, were based on an impedance error floor of 3.5%, which corresponds to a 7% error for the apparent resistivity and 2° for phase.

In GB tensor decomposition, galvanic distortion is parameterized in terms of shear that defines the polarization of the electric field and twist that defines its rotation (Groom and Bailey, 1989, 1991). The root mean square (RMS) misfit of the GB model provides a measure of the fit to the observed response of the GB model. The shear and twist angles, as well as the RMS misfits, for each site are shown in Figure 5.4. The results indicate that there is a moderate level of distortion in the study area. For the crustal depth range, the mean of the absolute value of the shear angle is 19° and the mean of the

absolute value of the twist is 12° (with a standard deviation of 12° in both cases). The absolute values of the shear and twist angles are predominantly $<25^\circ$ and $<15^\circ$ respectively. For the lithospheric mantle depth range (Figs. 5.4b and 5.4d), the mean of the corresponding absolute values of shear and twist values are 24° and 18° (with standard deviation of 14° in both cases). Variogram analysis of the results indicates that the shear and twist responses are dominated by local rather than regional effects (Frederiksen et al., 2006).

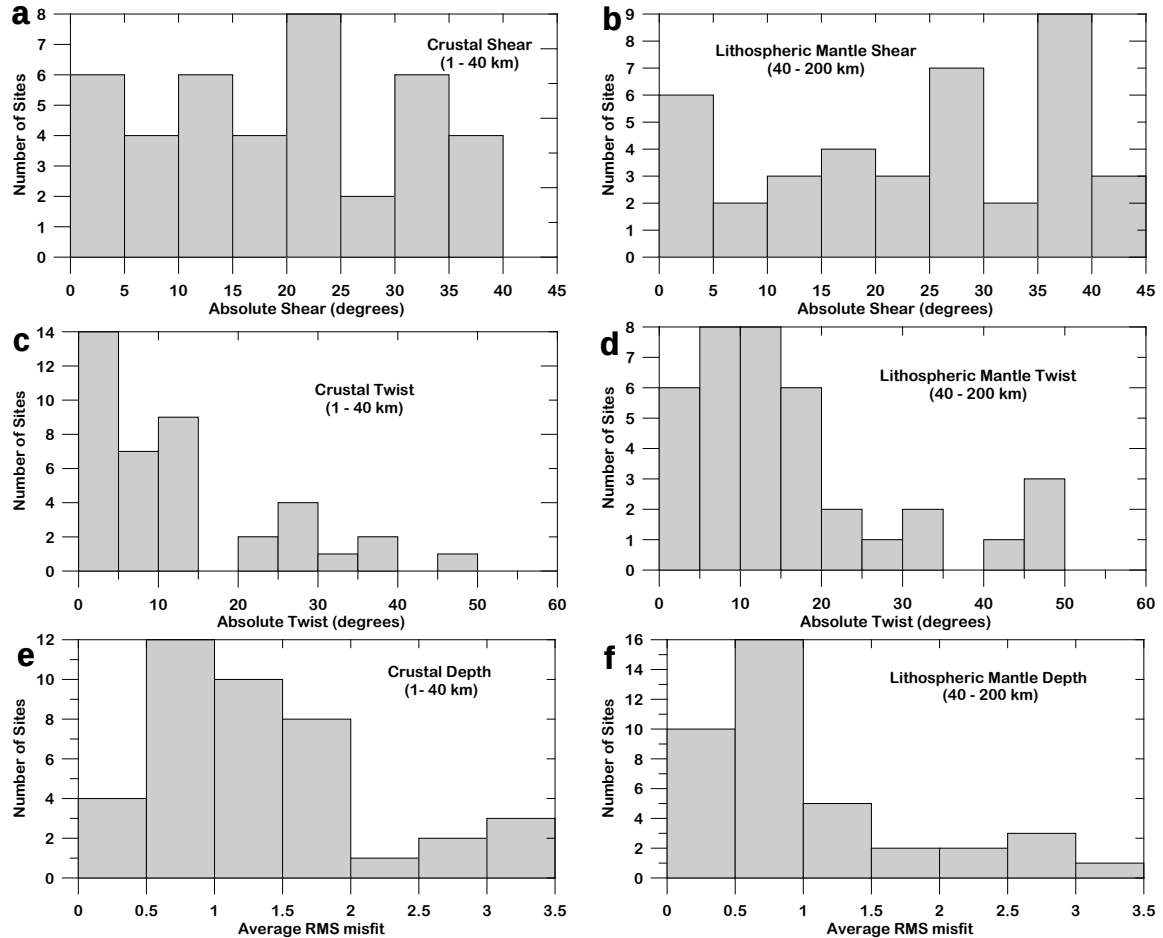
The distribution of the RMS misfits for the crustal (Fig. 5.4e) and the lithospheric mantle (Fig. 5.4f) depths shows that most of the sites have an average RMS misfit of less than 2.0 (which is considered to be an acceptable value) (Jones, 2012) with a peak mode of ~ 0.75 for both the crust and the mantle lithosphere. The sites with high RMS misfit are mainly located near major shear zones (the GF and CMBBZ).

Figure 5.5 shows Rose diagrams of geoelectric strike azimuths for crustal, lithospheric mantle and asthenospheric mantle depths. The 90° ambiguity that normally affects geoelectric strike determination (e.g., Hamilton et al., 2006) was addressed by plotting results in a single quadrant (azimuth of 10° - 100°) that includes the surface geological strike. The dominant geoelectric strikes are $N45^\circ E$ for the crust (Fig. 5.5a), $N85^\circ E$ for the lithospheric mantle (Fig. 5.5b) and $N68^\circ E$ for the asthenospheric mantle (Fig. 5.5c).

At lithospheric mantle depths, most of the strike azimuths lie between $N75^\circ E$ and $N90^\circ E$ with some local deviations within the Pontiac Subprovince and near the CMB. The regional lithospheric mantle strike determined in this study generally agrees with the $N80^\circ E \pm 6^\circ$ obtained by Ji et al. (1996) and the results of Frederiksen et al. (2006). The

differences are due to variation in the number of data used and the method of computation; this study is based on depth-transformed data whereas previous studies were based on

Figure 5.4. Distribution of the absolute value of the shear and twist angles for the crustal (a and c) and the lithospheric mantle (b and d) depths. The distribution of the average RMS misfits for the crustal (e) and lithospheric mantle (f) depths are also shown.



periods. Data from only seventeen sites within the study area provide penetration to mantle depths exceeding 200 km. They show geoelectric strike direction that ranges from N60°E to N75°E (Fig. 5.5c).

Figure 5.6 shows the individual single-site GB strike azimuths for the crust (Fig. 5.6a) and the lithospheric mantle (Fig. 5.6b). At crustal depths, the majority of the sites have strike azimuths that are parallel to, or sub-parallel to, regional geological strike, e.g., as defined by the GF and CMBBZ. As noted in Frederiksen et al. (2006), strike azimuth at some sites is parallel to the local total magnetic field anomalies. However, at other sites, particularly within the CMB, the azimuth is not parallel to either the regional geological strike or local magnetic field anomalies.

At lithospheric mantle depths, except for a couple of local exceptions, the sites show a very consistent east-west strike direction. Over the whole length of the profile this azimuth is consistently 30° to 40° clockwise from the crustal azimuth. As previously observed by Mareschal et al. (1995) and Ji et al. (1996), the lithospheric mantle strike azimuth is very uniform in a region crossing the GF and extending in the CGB.

Figure 5.6c shows the relationship between the individual single-site GB strike azimuths for the asthenospheric mantle depth. There is excellent agreement between the strike azimuths at most sites. The discrepancy at the remaining two sites may be due to the effects of noise or residual distortion in the MT data at these sites.

The consistency of geoelectric strike in the crust and the lithospheric mantle, along with the relatively low RMS misfit values, suggests that 2-D modeling is a reasonable approach for the determination of crustal and lithospheric mantle resistivity structures. However, the different average strike azimuths of $N45^{\circ}E$ (crust) and $N85^{\circ}E$ (lithospheric mantle) indicate the need for separate modeling at the two depths. Data sets were prepared for 2-D inversion using the regional responses from the GB

decompositions with the regional geoelectric strike azimuth constrained to either the crust or lithospheric mantle value. The data set included the tipper

Figure 5.5. Rose diagrams of the regional geoelectric strike for (a) crustal depth, (b) lithospheric depth, (c) asthenospheric depth and (d) the absolute plate motion for the HS3_Nuvel 1A hotspot model defined by Gripp and Gordon (2002) and for the GPS based model defined by Larson et al. (1997). The bin size in all cases is 15° and N denotes number of data points (sites) used for each plot.

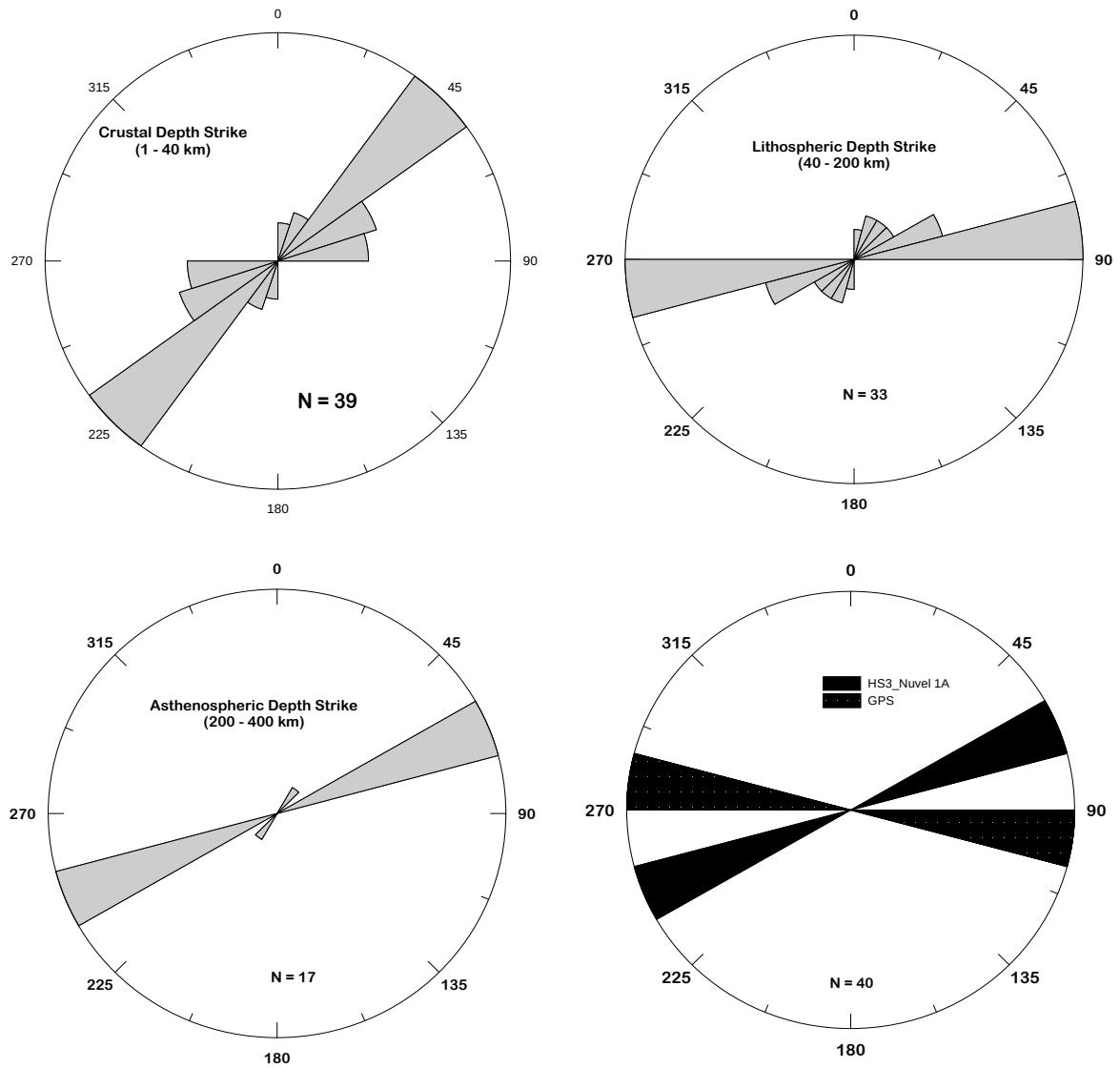
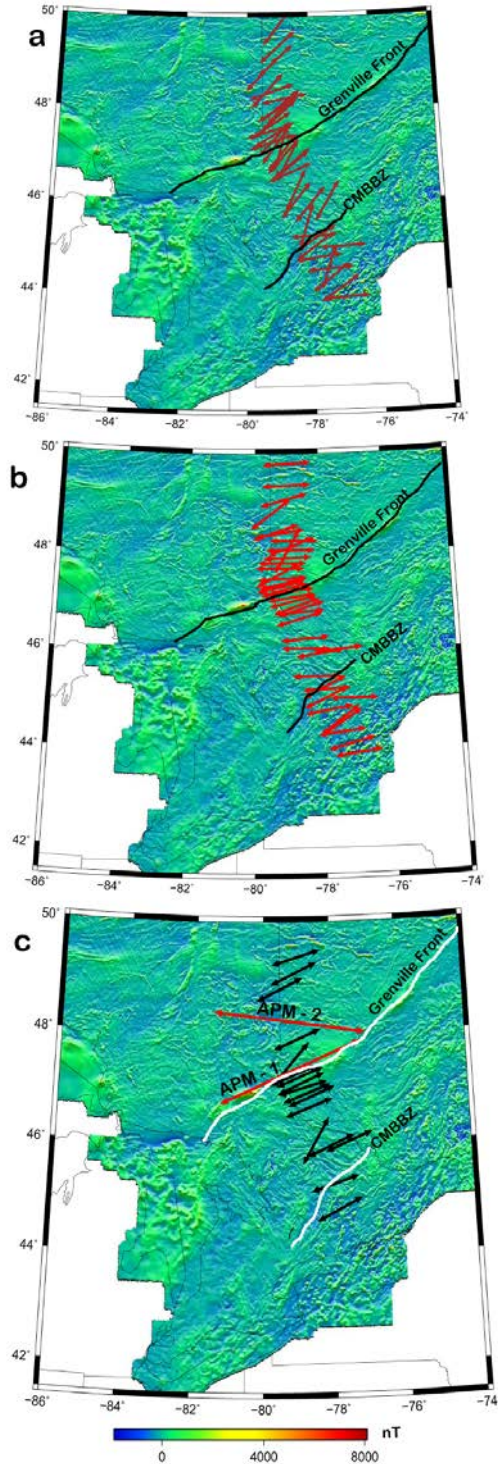


Figure 5.6. The crustal (a), lithospheric mantle (b) and the asthenospheric mantle (c) strike azimuths obtained from the STRIKE program with 90° ambiguity correction are plotted on the regional magnetic map (data obtained from Geological Survey of Canada). The longer arrow, APM-1, shows the HS3-Nuvel-1A absolute plate motion direction defined by the hotspot based model of Gripp and Gordon (2002) and the arrow APM-2 shows the absolute plate motion of the GPS based model of Larson et al. (1997).



response (T_z) which defines the relationship between the vertical and horizontal magnetic field components and forms part of the TE mode response (Vozoff, 1991). Prior to 2-D inversion, the data sets were re-edited and examined using TE and TM pseudosections and tipper pseudosections. The D+ approach (Parker, 1980), which performs 1-D modeling of the admittance, was used to check for consistency between phase and apparent resistivity responses and to aid in identifying unreliable data points that were subsequently removed prior to modeling/inversion.

For 1-D and 2-D structures the impedance phase responses are related to the gradient of apparent resistivity with log period (e.g., Parker and Booker, 1996; Weidelt and Chave, 2012). Increasing phase values greater than 45° indicate increase in conductivity with depth, values close to 45° indicate relatively uniform conductivity with depth while decreasing values less than 45° indicate decrease in conductivity with depth. The phase response is a more robust response than the apparent resistivity and for 1-D and 2-D structures it can be recovered exactly from distorted responses using tensor decomposition methods. In contrast, after tensor decomposition the apparent resistivity may still be affected by static shift, a frequency-independent multiplicative factor (Jones, 1988).

The MT apparent resistivity and phase pseudosections for both the crustal and lithospheric mantle data sets are presented in figures 5.7 and 5.8 respectively. Apparent resistivity and phase responses for both TE and TM modes are examined in both cases and the results indicate that significant differences exist between the TE and TM responses, which indicate the presence of 2-D structures. The TE mode is sensitive to current flow, which is dictated by the conductance of conductive regions of the sub-

surface, but has lower sensitivity to the lateral position of structures than TM mode. The TM response is sensitive to charges on conductivity contrasts or gradients of lateral boundaries. The pseudosections for the crustal and lithospheric mantle strike directions are generally very similar but there are some differences evident for the responses in the different strike directions.

The apparent resistivity response for both pseudosections is dominated along most of the profile by high values ($>2000 \Omega.m$) over an intermediate period band between 10^{-3} and 1 s. This response extends from the northwestern end of the profile southeast to the middle of CMB where the resistivity in the band decreases to values of $<20 \Omega.m$. At shorter periods the response is generally more conductive ($<1000 \Omega.m$), although in the CGB, resistive responses are observed to extend to the shortest periods. The transition from the conductive to the resistive response, at the shortest to the intermediate periods (10^{-4} to 10^{-2} s), is reflected in the phase response by a region of low impedance phase ($<30^\circ$). The apparent resistivity at longer periods (>1 s) is typically $<500 \Omega.m$. The transition from high apparent resistivity at intermediate periods to lower apparent resistivity at long periods is reflected in a period band of high phase responses ($>60^\circ$) centered on 1 s extending along most of the profile. These observations provide evidence for a structure extending over the profile that is relatively conductive near the surface, resistive at intermediate depths, and conductive at depth. Lateral variations in these responses, for example, increased TE mode apparent resistivity at intermediate periods in the CGB, indicate the presence of 2-D structures. In the CMB there are some differences in the form of the responses between the crust and lithospheric mantle pseudosections indicating the presence of strong 2-D structures.

The pseudosections provide evidence for several more conductive areas. In the northwestern part of the profile, there is a relatively conductive zone observed at short periods ($\sim 10^{-4}$ to 10^{-2} s). This zone extends from the Abitibi Subprovince into the middle of the CGB (Fig. 5.7 and 5.8). At the southeastern end of the profile, beneath the first site (PSO001), both pseudosections include a conductive zone defined by low apparent resistivity values in the TE and TM apparent resistivity responses at short periods. The lithospheric mantle apparent resistivity pseudosection (Fig. 5.8) includes a second conductive feature at longer periods at this location. Finally, there is a conductive response observed at long periods in the northwest of the profile in the lithospheric mantle strike pseudosection (Fig. 5.8). The response includes increased TM mode phase values at ~ 1000 s over the northwestern half of the profile. The phase values are locally higher in the part of the profile beneath the Pontiac Subprovince and at this location; there is an additional high-phase response in the TE component and a zone of decreased TM mode apparent resistivity.

The tipper data (not shown) were rotated to the 2-D modeling coordinate system, and the component parallel to the profile was edited carefully prior to inclusion in 2-D inversions. The tipper is usually sensitive to the local structures but can also be strongly affected by conductivity boundaries at the edges or outside of the region modeled. Based on visual inspection, the tipper data were heavily edited to exclude power-line effects and erratic points characterized by jumps in tipper magnitudes, tipper magnitudes exceeding 1, and/or large error estimates. In addition, based on the study of Zhang et al. (1993), short period (< 0.1 s) tipper data were excluded because of the possible

contamination by magnetic effects of galvanic distortion. Following the editing, the tipper responses include several large-scale spatial variations.

Figure 5.7. The apparent resistivity and phase pseudosections of TE and TM responses for the study area. The data are the regional impedances derived from the GB decomposition for the crustal strike azimuth of N45°E. Each panel shows the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.

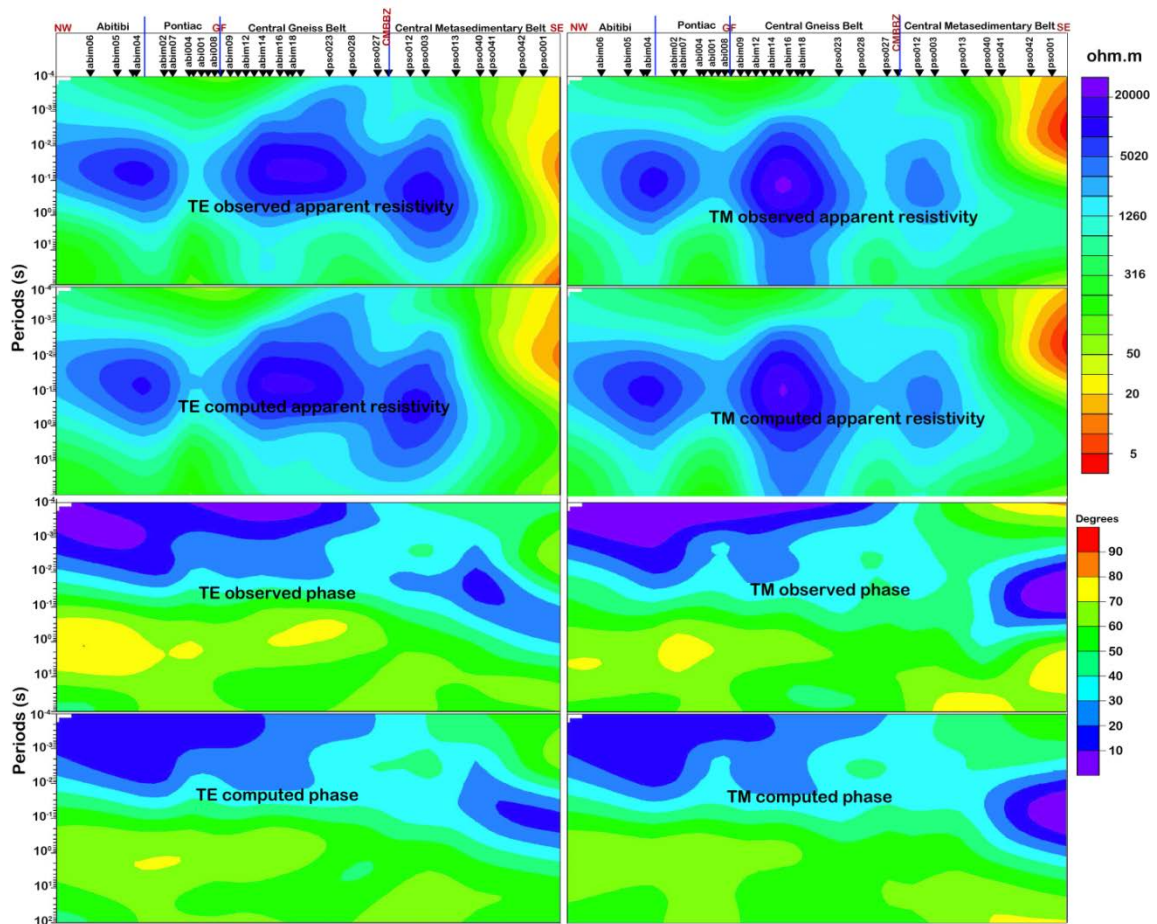
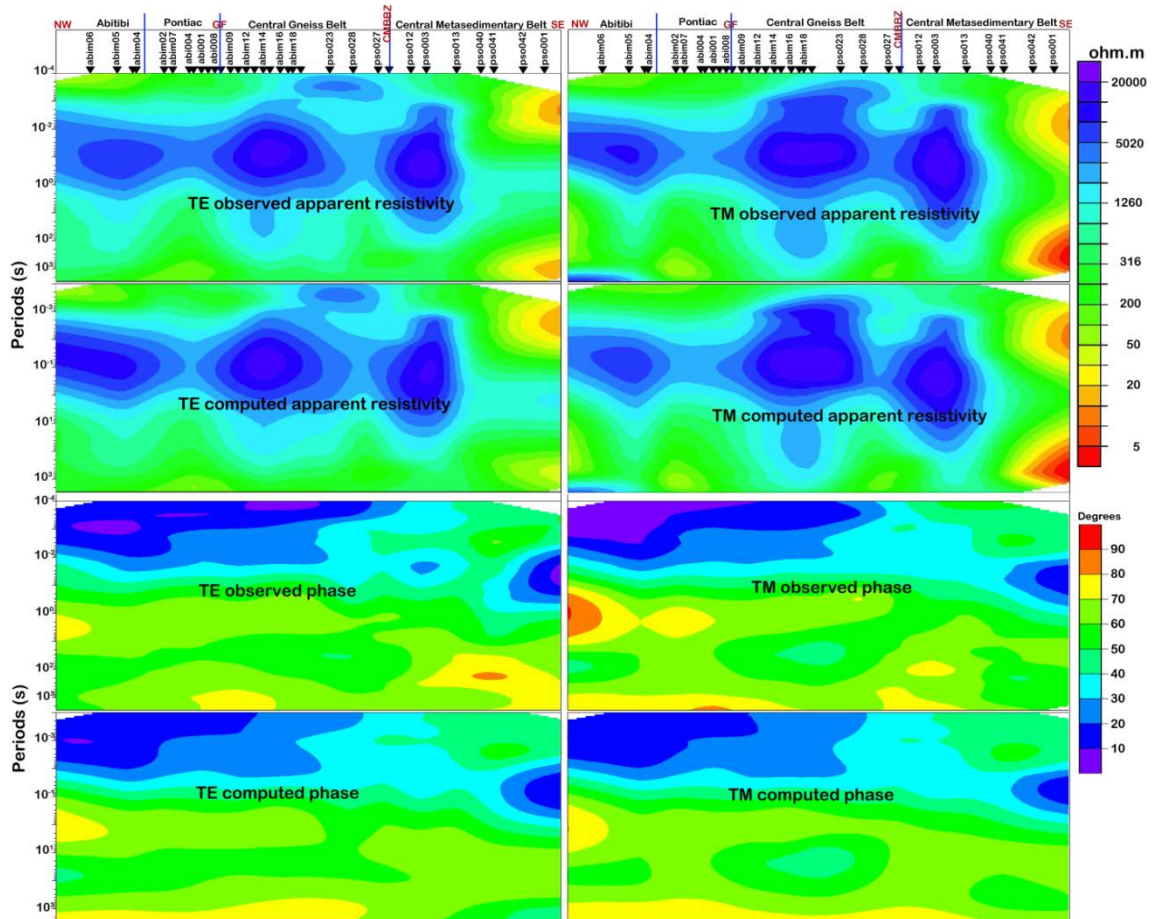


Figure 5.8. The apparent resistivity and phase pseudosections of TE and TM responses for the study area. The data are the regional impedances derived from the GB decomposition for the lithospheric mantle strike azimuth of N85°E. Each panel shows the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.



5.6 Two-dimensional modeling and inversion

The crust and lithospheric mantle in the study area were separately modeled using 2-D isotropic inversion of MT data decomposed to the appropriate strike directions for the two depth ranges. We used the 2-D isotropic modeling and inversion algorithm of Rodi and Mackie (2001), as implemented in Geosystem's (now Western Geco/Schlumberger) WinGLinK software. The 2-D forward modeling uses a finite difference numerical method for calculation. The inversion code applies the non-linear conjugate gradient (NLCG) method for direct iterative minimization of an objective function that penalizes data residuals and second spatial derivatives with respect to resistivity structures (Rodi and Mackie, 2001, 2012).

The inversion models were obtained by solving for the smoothest model using uniform-grid Laplacian regularization and minimizing the integral of the Laplacian (Rodi and Mackie, 2012). Many 2-D inversions, using different combinations of inversion parameters and data, were run to explore the range of possible models that could fit the data. Inversion of TE data alone resolves mainly the depth and the conductivity-thickness product of the conducting regions, whereas inversion of the TM data alone resolves their horizontal extent and the presence of resistive layers (Vozoff, 1991; Agarwal et al., 1993; Simpson and Bahr, 2005). Inclusion of both modes in the inversion allows for superior determination of the subsurface resistivity structure. All the 2-D models were derived using a 100 Ωm half space as the starting model. Initial models were obtained by fitting the TE, TM and vertical magnetic transfer function for all 40 BBMT sites. Some highly distorted sites, with extreme RMS-misfits (> 4.0), were later excluded from the final set of inversions.

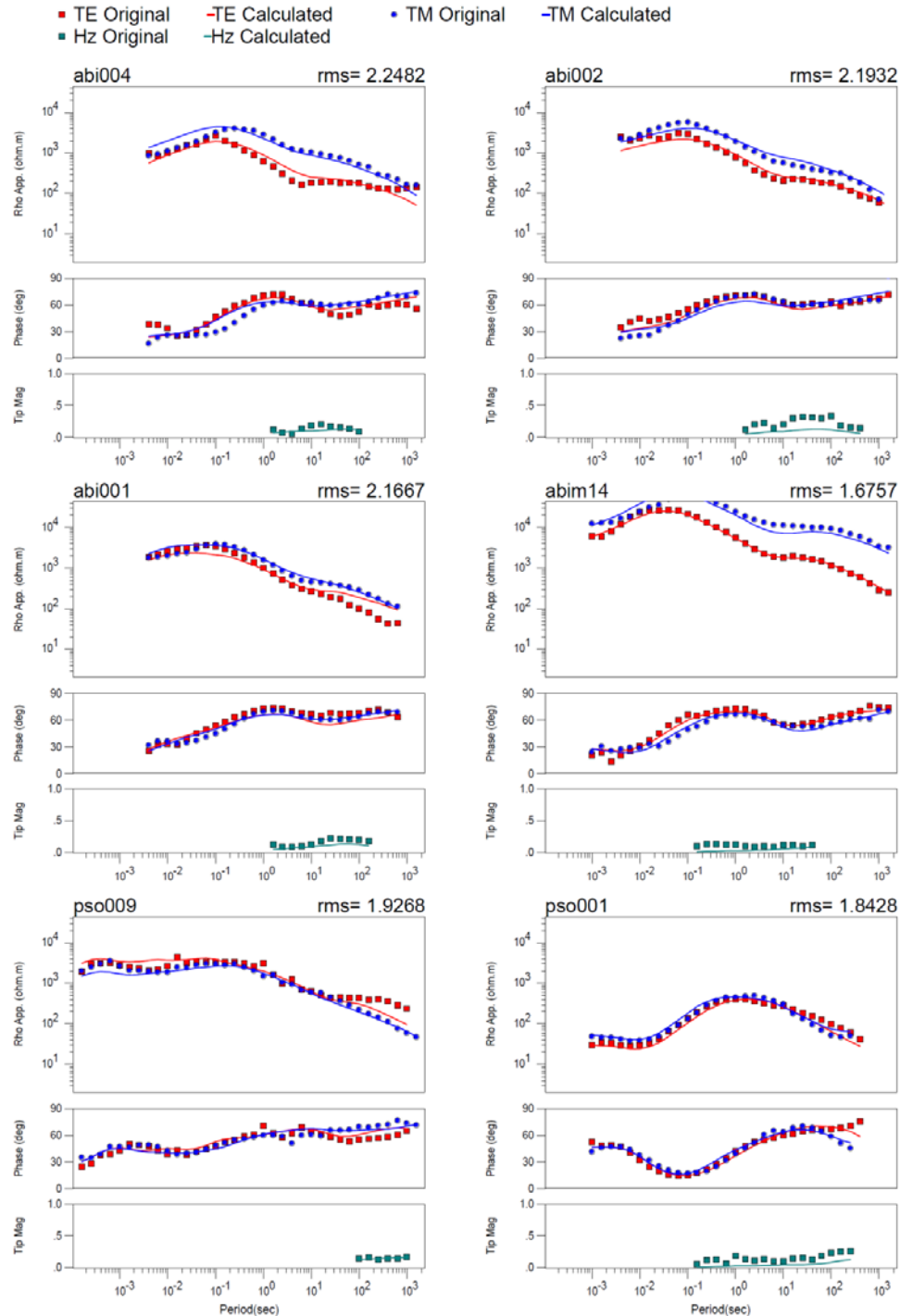
The uncertainty estimate on each data point was set to either the calculated error value or to an error floor if the calculated error was smaller than the value of the error floor. When 2-D inversion methods are applied to data containing weak 3-D effects, error floors must be set appropriately. For example, when the conductive features in a model have a finite strike length, the TE response is generally more distorted than the TM response (Ledo et al., 2002). However, there are some geometries for which the TM mode is more affected by 3-D effects than the TE mode (e.g., Park and Mackie, 1997). In this study, for the initial set of inversions, error floors of 50% were set for the apparent resistivity of both modes, the error floor for TE phase was set at 25% and the error floor for TM phase at 5%. The apparent resistivity was down weighted in order to reduce the effect of static shift. All of the error floors were subsequently reduced to 16% and 12% for TE and TM apparent resistivity and 4% ($\sim 1.2^\circ$) and 3% ($\sim 1^\circ$) for TE and TM phases respectively. The vertical magnetic transfer function (tipper) error was set at 0.02. The apparent resistivity data for the final set of models were corrected for static shift effects. Every new inversion was restarted at least once after its initial termination so that the final model would represent a deeper minimum of the objective function. The typical number of iterations per sequence was 400.

A range of weighting functions and regularization parameters were examined to explore for the optimal model (e.g., Mackie et al., 1997; Spratt et al., 2009; Schäfer et al., 2011). The model term in the objective function includes a multiplication factor α that is applied to the horizontal derivative and controls the relative horizontal smoothness of the inversion model. Another parameter, β , is an exponential factor that increases the weighting function with depth. Additional parameters that regulate the spatial smoothing

are the minimum block dimension in the horizontal and vertical directions, H and V respectively. The Tikhonov regularization parameter (τ) controls the trade-off between RMS-misfit and model-roughness. Large τ results in smooth models with high RMS-misfit whereas small τ produces a good fit to the data but rough structures are required. The optimal value of τ was determined by plotting a trade-off (L) curve between the regularization parameter and the RMS-misfit. The inversions used for this purpose were mainly based on the phase responses. For both the crustal and lithospheric mantle data sets, $\tau = 6$ gave results in the elbow of the trade-off curve providing a relatively good data fit and a relatively smooth model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversions. These values were chosen after a series of initial inversions that were performed to determine the best smoothing and regularization parameters.

Figure 5.8 compares the model response pseudosections with the data pseudosections for the best-fitting model of lithospheric mantle data set, and Figure 5.9 compares the observed TE, TM and tipper responses with the modeled ones at representative sites in the study area. The data fit is considered acceptable for all the sites shown, with individual RMS values of <2 . The misfit at other sites is shown in figures 5.10 and 5.11 for the crustal and lithospheric models respectively. The overall fit to the tipper response is poorer than the fit to the MT apparent resistivity and phase responses, but examination of the real and imaginary pseudosections of the tipper (not shown) shows that the model reproduces the large-scale features of the observed data.

Figure 5.9. Comparison of the observed and model TE and TM apparent resistivity and phases as well as the tipper responses at six sites along the MT profile. The results are for the inversion of the lithospheric mantle data set using the strike azimuth of N85°E. The model response shown is obtained from TE+TM+Tz data and the RMS misfit for each site is shown above it. Sites ABI001, ABI002 and ABI004 are located in the Pontiac Subprovince, ABIM14 is located in the CGB, PSO009 is located on the CMBBZ and PSO001 is the southeasternmost site.



5.6.1 Crustal resistivity model

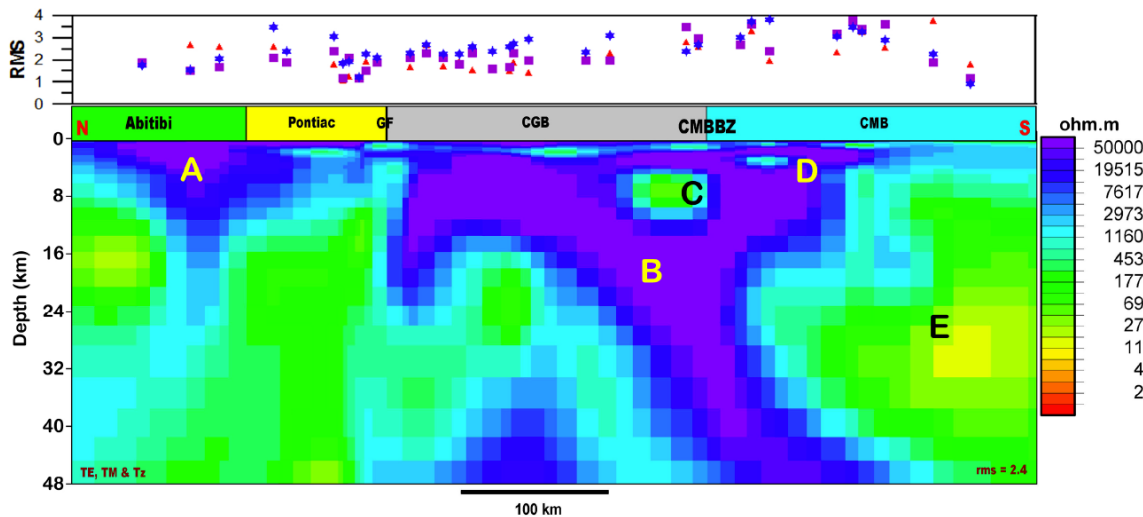
Figure 5.10 shows the final crustal 2-D resistivity model from simultaneously inverting the TE+TM+Tz data. The RMS-misfits of the separate TE (red), TM (blue) and joint TE+TM+Tz (purple) inversions, for each of the sites, are shown at the top of the model. The data fit is relatively good to the northwest of the CMBBZ, where the average misfit for the joint inversion is around 2. The misfit is particularly poor at some sites within the CMB, which is attributed in part to the structurally complex metasedimentary rocks that occur in this region. However, the global RMS misfit value of 2.4 is considered acceptable in light of the distribution of the misfit, and inspection of the data versus model pseudosections indicates that no major data feature is being misfit.

The resistivity structures defined by the joint and individual inversions reveal some differences caused by the different sensitivities of the different modes. Generally, the crustal model shows that there is a resistive upper crust below a relatively thin conductive layer near the surface. The resistive crust extends to 8-10 km depth beneath the Abitibi and Pontiac subprovinces (Feature A in Fig. 5.10). The Superior Province is characterized by very resistive upper crust ($>10,000 \text{ } \Omega\cdot\text{m}$), with resistivity decreasing with increasing depth. The resistive zone extends to 16-20 km depth beneath the CGB and northwestern part of the CMB (Feature D). A well-resolved resistive zone (Feature B), including resistivity exceeding $10,000 \text{ } \Omega\cdot\text{m}$, dips southeast to the base of crust beneath the CMBBZ. The resistive zone is $<5 \text{ km}$ thick in the southeastern part of the CMB and is less resistive than further to the northwest.

Except for the resistor beneath the CMBBZ, the middle to lower crust is relatively conductive, $<500 \text{ } \Omega\cdot\text{m}$, along the whole profile. A poorly-resolved localized conductive

zone (Feature C) ($\sim 30 \Omega.m$) is imaged in the upper crust beneath the CMBBZ. The southern end of the profile is characterized by a fairly well resolved conductive zone in the lower crust (Feature E) ($< 10 \Omega.m$) that is defined by the data at the two southeasternmost sites (PSO042 and PSO001). This zone extends from a depth of around 24 km to 40 km and has a lateral extent of about 80 km. It is poorly resolved because of the strong local variation in the geoelectric strike of this region. However, it is not considered a spurious feature caused by localized 3-D responses in the data or 3-D structures at the southern end of the profile because it appears in both the joint and separate inversions of the individual modes and it is associated with a zone of lower apparent resistivity in the pseudosections (Fig. 5.7).

Figure 5.10. 2-D resistivity section of the crust obtained by joint inversion of TE, TM and tipper responses for the crustal data set (V.E = 5). The RMS values of the TE (triangles), TM (stars) and joint TE+TM+Tz (squares) inversions at individual sites are plotted above the inversion model.



5.6.2 Lithospheric mantle resistivity model

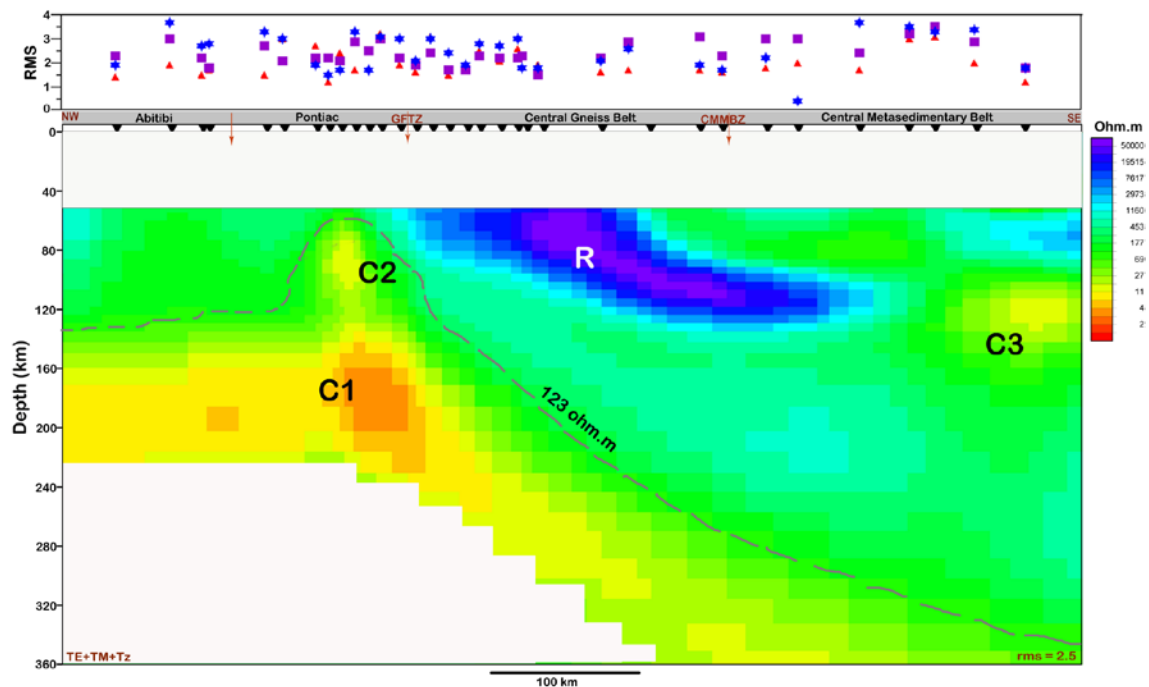
Figure 5.11 shows the resistivity model and data misfit for the lithospheric mantle. In this model the top 48 km, i.e., the crust, has been blanked out as the crustal data are in an inappropriate strike direction for the inversion. Data from a number of sites (ABIM06, ABI005, ABI006, PSO014, PSO022, PSO038, and PSO039) were excluded from the final sets of inversions because these sites exhibited high RMS misfit (>4) in earlier exploratory inversions. The global RMS error of the final model is 2.50. The misfit results show that it is possible to fit the TE mode data to a lower level of relative misfit than the TM mode data. The fit of the joint inversion model is found to be reasonably spatially uniform, except for some sites at the southern end of the profile. As shown by the pseudosection responses (Fig. 5.8), the large-scale features in the observed data are all replicated by the model response.

Figure 5.11 shows that only the northwest and southeast ends, as well as the very deep parts of the lithospheric mantle, along the profile, are relatively conductive ($<200 \Omega.m$). Beneath the CGB, there is a well resolved southeast-dipping highly resistive zone (labeled R in Fig. 5.11) that extends from the base of the crust to about 140 km depth. The upper part of the resistive feature is also visible in the crustal model beneath the CGB (Fig. 5.10). It extends laterally from the GF to the middle of the CMB. The resistivity of the feature is 2,000-8,000 $\Omega.m$. The lithospheric resistor corresponds to a region of interpreted resistive and anisotropic rocks defined in earlier studies (Mareschal et al., 1995; Senechal et al., 1996; Ji et al., 1996). The resistor overlies a less resistive region, with resistivity of ~ 250 to 2,000 $\Omega.m$, that is still more resistive than adjacent

parts of the lithosphere. This zone extends to a depth of around 280 km, and its southeastern extent is similar to that of the overlying more resistive zone.

The lithospheric resistivity model contains two isolated conductors. There is a conductor at approximately 80 km beneath the Pontiac Subprovince (labeled C2 in Fig. 5.11), and at the southern part of the profile, a localized conductor (labeled C3) with conductivity between 10 and 20 $\Omega.m$, is also resolved. This conductor lies between 100 and 150 km depth.

Figure 5.11. Resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set (V.E = 1). The RMS values of the TE (triangles), TM (stars) and joint TE+TM+Tz (squares) inversions at individual sites are plotted above the inversion model. The crustal section (upper 48 km) of the model is not shown. The lower section of the model in which there is no data resolution (below the lithospheric mantle conductor C1) has been blanked. The 123 $\Omega.m$ contour is shown in order to better define the depth extent of the more resistive part of the mantle.



A prominent feature of the mantle lithosphere resistivity model is a deep conductor (labeled C1 in Fig. 5.11) beneath the southwest Superior Province. The top of the conductor is approximately horizontal and lies at a depth of about 150 km. Its resistivity is $\sim 10 \text{ } \Omega\cdot\text{m}$. The conductivity of the feature increases beneath the Pontiac Subprovince (resistivity $< 5 \text{ } \Omega\cdot\text{m}$), where it is overlain by conductor C2. The form of the contour shown in Figure 5.11 suggests that C2 is connected with a zone of enhanced conductivity within the horizontal part of conductor C1. Further to the southeast, the conductive feature appears to dip downwards beneath the Grenville Province but this part of the conductor is not properly resolved by the inversion and the apparent geometry is an artifact of the inversion.

Further data examination was carried out to confirm that the major resistivity structures in the final 2-D resistivity models are supported by the data. The key features in the model occur at depths at which the MT data from the overlying sites provide adequate penetration (shown in Fig. 5.3). Conductor C3 is primarily defined by the TE mode data, whereas resistor R and conductors C1 and C2 are defined by the penetration of both TE and TM modes, as well as the T_z transfer functions. The conductors are associated with increase in phase or decrease in apparent resistivity at longer periods in the data. Conductors C2 and C3 can be related to conductive responses in the TE apparent resistivity response at periods of $\sim 10^3 \text{ s}$. The northwestern part of Conductor C1 appeared in almost all of the various inversions, providing a reasonable fit to the long period MT data. The existence of the conductor is supported by the observation of high phase values (exceeding 70° in places) at periods $> 10^3 \text{ s}$ in the TE and TM phase responses (Fig. 5.8 and 5.9). The absence of decreasing phase values at longer periods

shows that the base of the conductor is not resolved, i.e., the electromagnetic fields at long periods do not penetrate through the conductor. The key features in the resistivity model are also imaged at sites where the data fit is good as shown in Figure 5.11. Sites ABI001, ABI002 and ABI004 are directly above conductors C1 and C2, ABIM14 is above resistor R, while PSO001 is above conductor C2.

The necessity of the features in the resistivity model was examined by hypothesis testing using modeling and inversion. The joint inversion model was edited by removing the conductive and the resistive features (jointly and independently), and the resulting models were taken as starting models for a new series of inversions constrained to find the model closest to the starting model rather than the smoothest model, as is normally the case. The results consistently showed that conductors C1, C2 and C3, as well as the resistor R, are data-supported structures. The complete removal of conductors C1 and C2 or resistor R from the model consistently increased the global RMS misfit of the model by at least a factor of 3 and the RMS misfit of the individual sites also rose significantly. Although the resolution and shape of these features vary to some degree for different inversions, the results consistently suggest that the conductors and the resistor are first-order features existing in the subsurface.

5.7 Comparison of resistivity models with seismic results

5.7.1 Crustal model

Figure 5.12 shows the location of the MT profile relative to seismic results available in the study area. The seismic results include a seismic refraction/wide-angle reflection profile that is approximately coincident with the southeastern part of the MT

profile, as well as seismic reflection lines 12, 16, 16A, 15, 32, 33 and 71 that provide overlap coverage over much of the profile. Figure 5.13 shows the seismic reflection and refraction models of the study area. Calvert and Ludden (1999) and White et al. (2000) used the Lithoprobe depth-migrated, near-vertical incidence, seismic reflection data along with seismic refraction-wide angle reflection velocity model to construct the cross-section of the crust in this region. The seismic profiles used include Lithoprobe lines 12 and 16 (Calvert and Ludden, 1999), line 16A (Benn et al., 1994), line 15 (Kellett et al., 1994), and lines 32 and 33 (White et al., 2000). Zhang and Frederiksen (2013) provide scattering tomography P and S velocity models for the crust in the study area but neither the P- nor S- velocity models show significant correlation with tectonic structures and the interpretation of their results remains uncertain.

Seismic reflection results show that the southern end of the Abitibi Subprovince is characterized by a half graben structure (Calvert and Ludden, 1999; Benn, 2006), and that the Pontiac crust is primarily made up of a mid-crustal duplex, underlain by a deeper thrust and overlain by near surface thrusts (Benn et al., 1994; Kellett et al., 1994; Calvert and Ludden, 1999). To the west of the study area, the GFTZ is 32 km wide (Green et al., 1989). Within the study area, it is a zone of southeast dipping stacked crustal sheets that extend to a maximum depth of 25-30 km (White et al., 2000).

The near-vertical and wide-angle seismic reflection data show dominantly southeast dipping reflectivity throughout the Grenville Province (White et al., 2000). The seismic results are interpreted to show southeast to northwest assembly of allochthonous crustal elements making up the CMB, followed by the deformation of this unit and its northwest transport over the rocks of the Laurentian margin (CGB) and pre-Grenvillian

Laurentia (southeast Superior craton) (White et al., 2000). The reworked pre-Grenvillian Laurentia and Laurentian margin rocks are interpreted to extend at least 350 km south of the GF. Roy and Mereu (2000) showed that the CMBBZ is characterized by bands of southeast-dipping shallow reflectors extending to mid-crustal depth of 25 to 30 km. O'Dowd et al. (2004) defined this zone to be 10 to 12 km wide.

Figure 5.14 overlays the seismic reflection image on the 2-D crustal resistivity model. The figure shows previous interpretations of the seismic reflection data (e.g., Kellett et al., 1994; Benn et al., 1994; Calvert and Ludden, 1999; White et al., 2000) (solid lines) and some modified interpretations (dashed lines) based on the combined resistivity and seismic reflection information. Within the Grenville Province, the strongest dipping reflectors are associated with very resistive regions in the resistivity model. There are prominent packages of reflectors, interpreted in previous studies (White et al., 2000) to correspond to the GF and CMBBZ, near the upper and lower surfaces of the dipping resistor (Feature B in Fig. 5.10). Strong coherent reflectors also occur within the resistive upper crust (Feature D) several tens of kilometers to the southeast of the CMBBZ. Further to the southeast, the reflectors associated with more conductive parts of the crust are more diffuse or sparse, and have irregular dips. There are no reflection data available in the area of the crustal conductors (Features C and E) in the Grenville Province.

Figure 5.12. Map of the study area showing the relationship between this study and previous seismic and tomography studies. The dashed white line indicates the Lithoprobe Abitibi-Grenville transect wide angle seismic refraction profile. Locations of Lithoprobe reflections seismic lines 12 and 16 (Calvert and Ludden, 1999), line 16A (Benn et al., 1994), line 15 (Kellett et al., 1994), and lines 32, 33 and 71 (White et al., 2000) are also shown. The black lines illustrate the Rondenay et al. (2000), Chen and Li (2012) seismic tomography studies as well as the surface track of Great meteor hotspot track defined by Crough (1981). The locations of the Kirkland Lake and Cobalt kimberlite fields are also indicated.

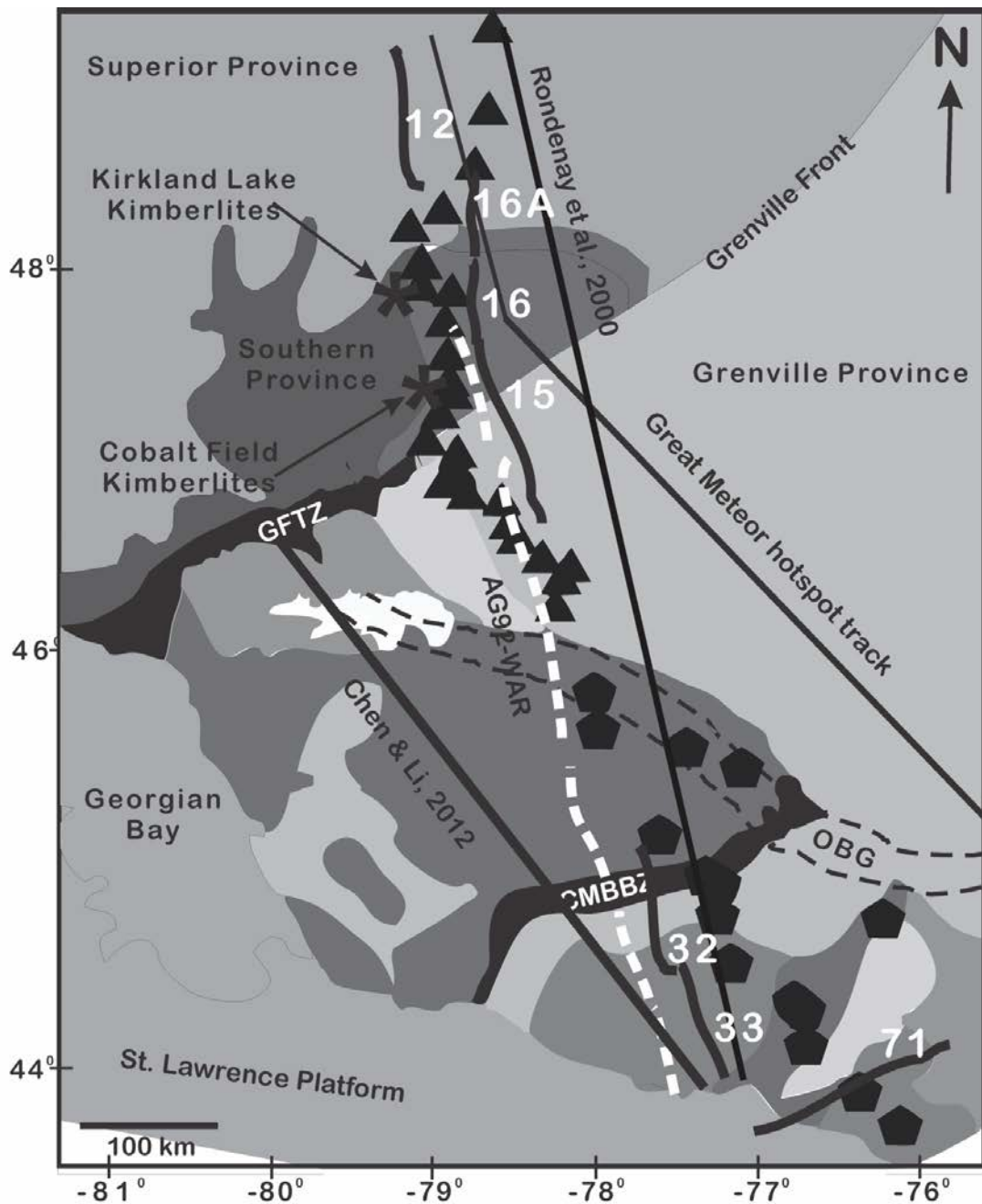


Figure 5.13. Previous seismic reflection and refraction (wide-angle reflection) models of the study area. The seismic refraction wide-angle reflection model (top) is from White et al., (2000). The middle panel shows a composite seismic reflection image from Lithoprobe seismic reflection lines 12 and 16 (Calvert and Ludden, 1999), line 16A (Benn et al., 1994), line 15 (Kellett et al., 1994), and lines 32 and 33 (White et al., 2000) through the boundary zones. The bottom panel represents the crustal seismic interpretation from Ludden and Hynes (2000) and White et al. (2000).

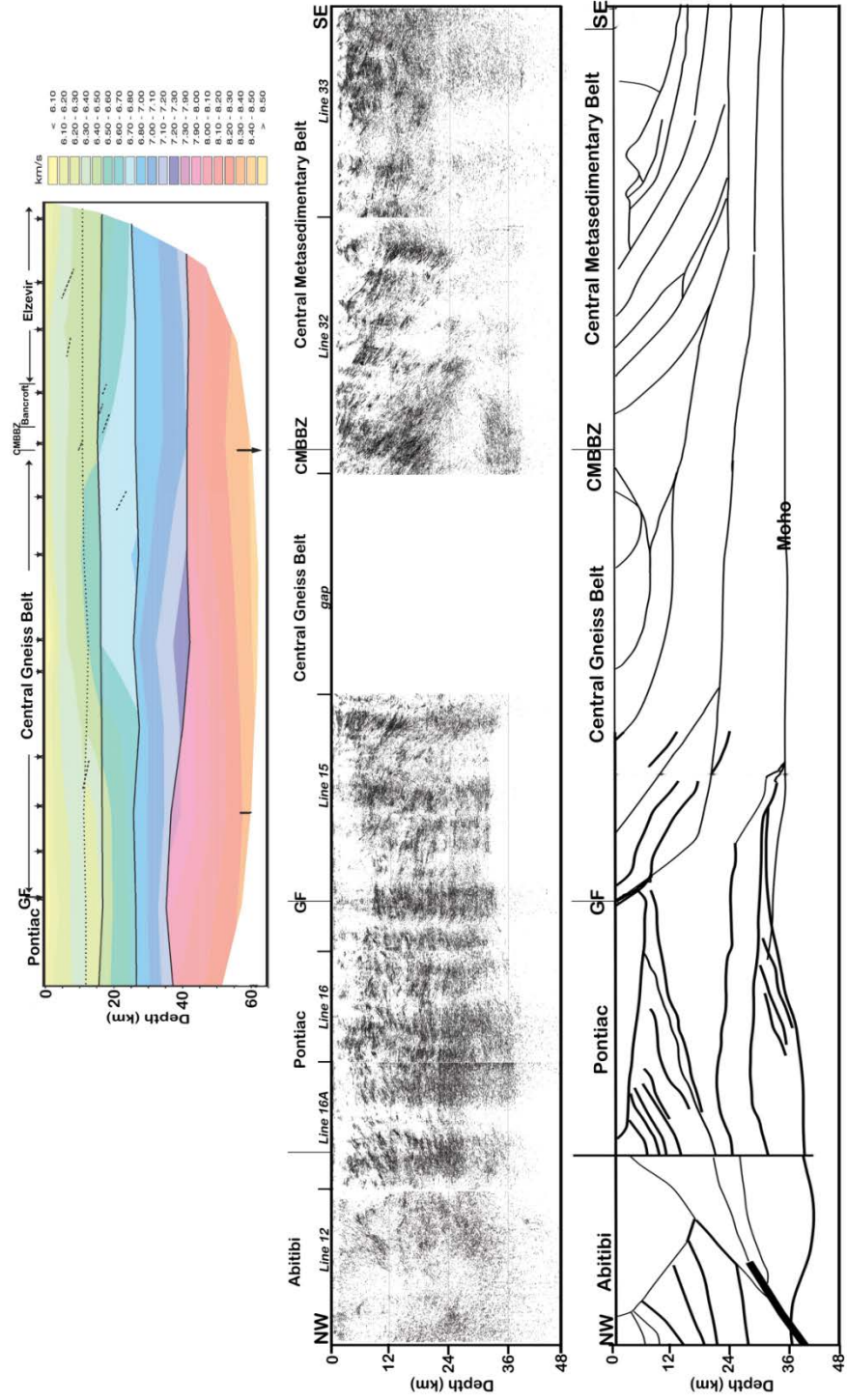
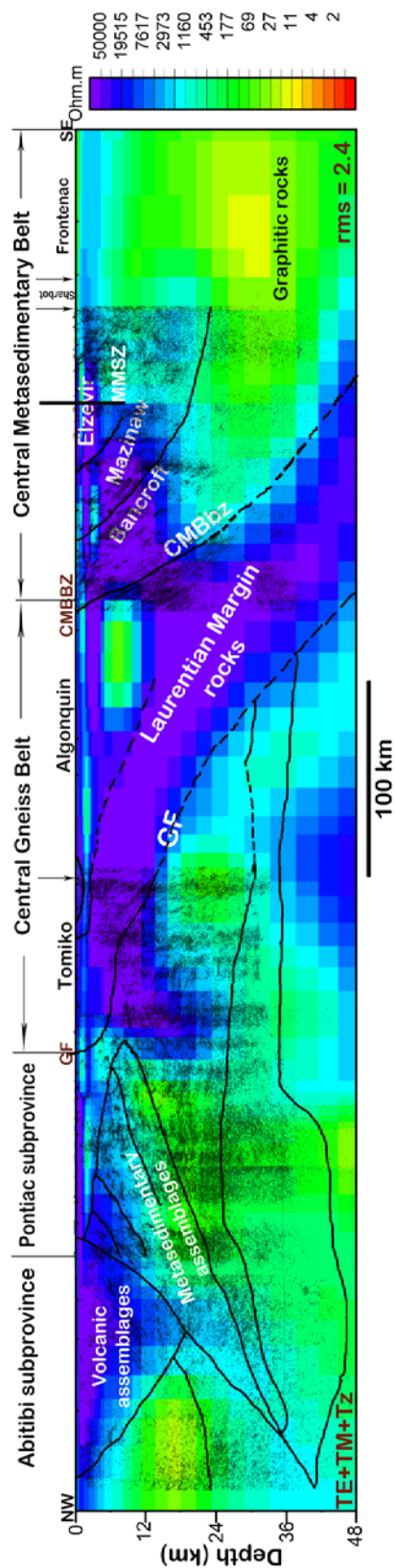


Figure 5.14. The crustal resistivity model ($V.E = 2.5$) is overlain by seismic reflection profiles from Lithoprobe lines 12, 16, 16A, 15, 32, and 33. The Algonquin (in CGB), Sharbot and Frontenac (in CMB) terranes are not covered by the seismic studies. The black continuous lines show the interpretation of Kellett et al. (1994) for line 15, Benn et al. (1994) for line 16A, Calvert and Ludden (1999) for lines 12, and 16, as well as White et al. (2000) for lines 32 and 33. The dashed lines show the additions and modifications of the interpretation based on the resistivity data.



5.7.2 Lithospheric mantle model

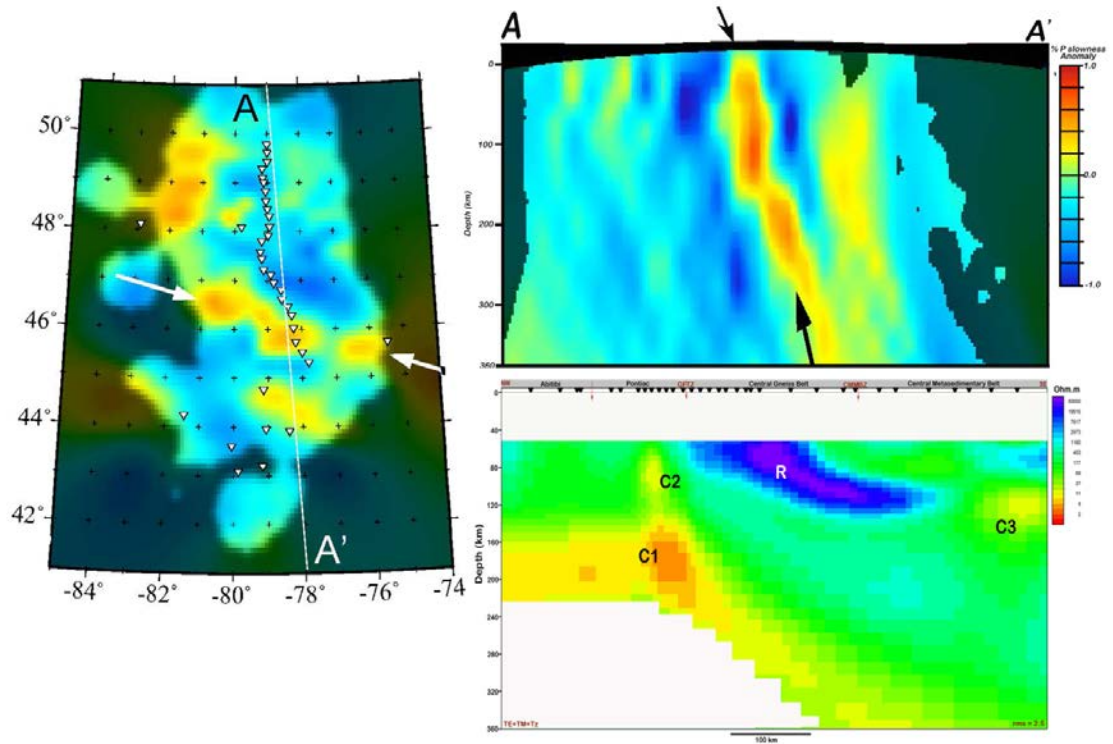
Figure 5.12 shows the location of seismic tomographic images derived in studies by Rondenay et al. (2000) and Chen and Li (2012) and reproduced here in Figures 5.15 and 5.16. Aktas and Eaton (2006) also conducted a tomographic study of the same region and Zhang and Frederiksen (2013) conducted scattering tomography.

In a large-scale teleseismic study, van der Lee and Nolet (1997) concluded that the Canadian Shield has a thick lithospheric root that extends to ~250 km depth beneath the Grenville Province. Smaller-scale seismic tomography studies have imaged a high velocity zone, or “seismic lid” in the upper lithosphere of the northern Grenville Province. It is defined in the Rondenay et al. (2000) (Fig. 5.15) and Chen and Li (2012) (Fig. 5.16) models as being between 50 and 100 km thick and lying between depths of 50 and 150 km. The strike direction of this anomaly is roughly parallel to the Grenville orogen tectonic trend (Rondenay et al., 2000; Aktas and Eaton, 2006; Chen and Li, 2012). Aktas and Eaton (2006) interpreted a high velocity anomaly within the Grenville Province as a relict slab associated with subduction beneath the CMB at 1.25 Ga.

There are several differences in how the body is imaged in different seismic studies. In the Rondenay et al. (2000) model, the high velocity zone is separated into two parts by the underlying low velocity anomaly, but in the Chen and Li (2012) model it is continuous across the area of the low velocity zone. Considering the superior depth resolution of the surface-wave based results of Chen and Li (2012), it appears that the break in the high velocity zone is likely an artifact of upward smearing of the low velocity anomaly. In contrast to the body wave studies of Aktas and Eaton (2006) and Rondenay et al. (2000), Chen and Li (2012) show that the high velocity zone extends to

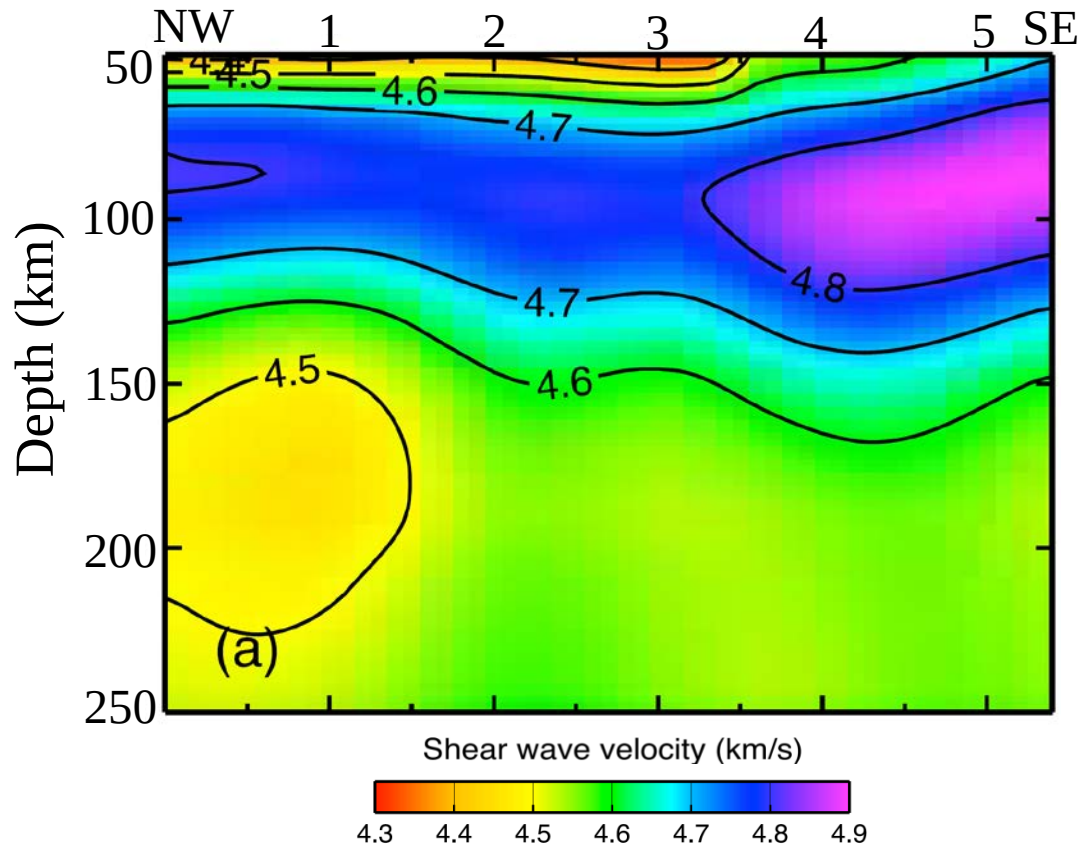
the south of the current study area and that the velocity anomaly becomes stronger near the Appalachian Front. Considering the superior lateral resolution of the travel time inversion, it appears possible that the Chen and Li (2012) model is connecting separate anomalies to the north and south of Lake Ontario.

Figure 5.15. Comparison of seismic body-wave tomography results of Rondenay et al., (2000) with resistivity model. Left panel shows the location of the seismic tomography profile and upper right panel shows the percentage P-wave slowness anomaly. Arrows define the low velocity zone. Lower right panel shows the lithospheric mantle resistivity model at the same horizontal and vertical scale as the seismic image.



The resistor R1 imaged in this study is closely spatially correlated with the high velocity anomaly. The upper and lower surfaces of the resistor are at similar depth to those of the velocity anomaly resolved in the surface wave study of Chen and Li (2012). The northern limit at 47° N is consistent with the northern limit of upper lithospheric (<150 km) high velocities in the Aktas and Eaton (2006) and Rondenay et al. (2000) body-wave tomography models.

Figure 5.16. Seismic surface-wave tomography model of Chen and Li, (2012) for NW-SE profile through the present study area. See Figure 5.12 for profile location: the present study area extends over the distance interval 0 to 3. The image shows the absolute S-wave velocity.



Taken together, the seismic and MT results provide compelling evidence for a coincident high-velocity, high-resistivity zone in the upper lithosphere extending southwards from near the Grenville Front. The southern limit of the feature is less well resolved but the MT and body wave seismic inversions suggest that a maximum southern extent is a point beneath the Frontenac terrane. Based on the resistivity model, the top of the body dips to the southeast from near the Moho at the GF to around 80 km depth beneath the Frontenac terrane. The base of the anomaly increases from around 80 km beneath the GF to around 150 km beneath the Frontenac terrane.

The seismic tomography also identifies a low velocity anomaly in the study area. Rondenay et al. (2000), in their P-wave travel-time inversion results (Fig. 5.15), resolve a near-vertical, WNW-ESE-trending low velocity anomaly. This anomaly extends between 50 and 300 km depth, with a relatively constant width of ~120 km. Aktas and Eaton (2006), also using body wave travel-time tomography, imaged the same anomaly and referred to it as a patchy NW-trending low velocity anomaly. Chen and Li (2012) produced a 3-D shear wave model of this region using a surface-wave two-plane wave inversion method, and their result (Fig. 5.16) shows a near-vertical anomaly similar to those imaged by earlier studies but in their model the top of the anomaly is at 150 km depth.

The low velocity anomaly is interpreted as being spatially correlated with the Great Meteor hotspot track by Aktas and Eaton (2006), Chen and Li (2012), Eaton and Frederiksen (2007), and Rondenay et al. (2000). The geometry and location of this anomaly approximately match an indentation in the North American lithospheric root defined by Fouch et al. (2000). This indentation is parallel to, and located to the

southwest of, the Ottawa Bonnechere Graben. Faure et al. (2011) interpreted this feature as a mantle scar, produced during Neoarchean subduction, which has been reused by several magmatic events.

The low velocity lithospheric anomaly imaged in the seismic tomography study lies to the south of the NW-SE dipping lithospheric conductor C1 imaged in this study (Fig. 5.15). The lateral position of the seismic anomaly is defined accurately in the body wave travel time tomography studies of Aktas and Eaton (2006) and Rondenay et al. (2000) both of which involved relatively high site density. Likewise, the lateral position will be quite well resolved in the MT 2-D inversions as it is defined by both the TE and TM mode data. Although the seismic low velocity and low resistivity anomalies have a somewhat similar geometry, they are clearly distinct features. The resistivity is relatively uniform and resistive in the vicinity of the seismic velocity anomaly. However, there is a weak low velocity response observed in the vicinity of the resistivity anomaly.

5.8 Interpretation and discussion of geoelectric strike

The geoelectric strike results show geographically consistent, but different, strike azimuths for the crust, lithosphere and the asthenosphere of the Grenville Province and adjacent Superior Province in southern Ontario. The defined crustal geoelectric strike (45°) is approximately parallel to the geologic strike of the GF and the CMBBZ in the study area. The strike azimuth for the crust exhibits more variability than the lithospheric and asthenospheric azimuths. These variations are attributed to some sites within the Pontiac Subprovince and the CMB. Within the CMB the variability is explained by the presence of polyfolded domains with complex geometry and by the brittle to brittle-

ductile thrust and normal faults that separate the domains (White et al., 2000). These rocks contrast with the largely flat-lying, amphibolite-granulite facies magmatic gneisses to the north in the CGB.

This study defined a regional lithospheric strike direction of N85°E, which agrees well with the N80°E defined by earlier studies (e.g., Mareschal et al., 1995). The new MT sites in the present study suggest that this strike azimuth extends some distance south into the Grenville province and therefore that there may be additional controls on the strike direction to the deformation in transcurrent shear zones in the mantle as suggested by Ji et al. (1996).

Deeper (>200 km depth) geoelectric strike azimuth, can be compared with estimates of absolute plate motion (APM) e.g., as done by Simpson (2001) (Fig. 5.5d, 5.6c). The APM direction at each of the MT sites was defined using three different APM models: the model of Larson et al. (1997), which is based on Global Positioning System (GPS) results, has a mean direction of 277°; the APM model of Demets et al. (1990), which is based on no-net-rotation reconstruction, has a direction of 279°; and the model of Gripp and Gordon (2002), which is a hotspot-referenced model, has a direction of 248°. The observed asthenospheric (>200 km depth) geoelectric strike azimuth is in excellent agreement with the Gripp and Gordon (2002) direction but it appears that this global APM model provides a poor estimate of the true plate motion in the study area. The Gripp and Gordon (2002) result, which is the estimate of the direction of the trace a hotspot would make over the last 50 Myr, is oblique to the interpreted surface trace of the Great Meteor hotspot for 150 to 115 Myr (Eaton and Frederiksen, 2007). The Great Meteor hotspot trail implies an approximately northwest plate motion in the study area,

closer to direction indicated by the other APM models. Frederiksen et al. (2006) show that, the shear-wave fast direction, in the study area, coincides with the direction of plate motion defined by Larson et al. (1997).

The results of the present study therefore indicate that the deep (>200 km) geoelectric strike is oblique to both the plate motion and seismic fast direction. Similar obliquity has been observed in other Precambrian cratons (e.g., Simpson, 2001; Hamilton et al., 2006). In the Grenville-Superior area, the observed obliquity provides support for interpreted lithospheric deformation in the region (Eaton and Frederiksen, 2007). The deep geoelectric strike has the same sense as an observed offset between the surface trace of the Great Meteor hotspot and the seismic velocity “divot” observed at depths of 200 km that it is interpreted to be associated with the hotspot. The offset is explained by Eaton and Frederiksen (2007) by deformation of the lithospheric keel arising from viscous coupling with asthenospheric flow. Noting that the Niblett-Bostick depth transform used in the study is an approximate transform, and therefore that there may be some contribution to the “asthenospheric” results from lithospheric depths; the MT results are also consistent with such deep lithospheric deformation.

5.9 Interpretation of crustal resistivity structures

5.9.1 The Abitibi Subprovince

Figure 5.14 shows the interpretation of the crustal resistivity model based on previous seismic reflection results and known surface geology of the region. Seismic reflection interpretations suggest that the crust of the Abitibi has a three-layer form (Calvert and Ludden, 1999; Benn, 2006). The upper to middle crust, at the northwestern

end of the resistivity model, corresponding to the upper layer in seismic reflection models, includes a very resistive, synformal structure that extends to about 20 km depth. Our interpretation of the MT results follows the original interpretation of reflection seismic line 12 which suggests that this feature is a half graben developed as a result of large-scale extension (Calvert and Ludden, 1999). The surface geological evidence suggests that this feature consists of stacked volcanic assemblages of the southern volcanic zone of the Abitibi Subprovince (Ludden and Hubert, 1986). The high resistivity in the upper crust of the Abitibi may also be due locally to the granitic plutons intruded into what are now shallow depths (e.g., Benn, 2006).

The high resistivity of the Archean upper crust of the Abitibi Subprovince was noted in earlier MT studies. In Precambrian basement terranes, metasedimentary rocks, and particularly pelitic rocks, are observed to be more conductive than metavolcanic rocks or granites because of their increased content of conducting elements such as graphite or sulfides (e.g., Ferguson et al. 1999; Boerner et al., 2000; Korja et al. 2002). Deformation can further increase the conductivity by increasing the interconnection of these elements (Jones et al., 1997; Gowan et al., 2009).

Boerner et al. (2000) speculated that the lack of enhanced conductivity in the metasedimentary rocks of the southern Superior province may be due to the limited extent, or poor preservation, of anoxic basin sediments. Earlier high resolution MT studies have shown mild enhancement of conductivity at regional scale fault zones in the Abitibi, such as the Destor-Porcupine deformation zone (Zhang et al., 1995; Boerner et al., 2000). However, this enhanced conductivity is not resolved by the subset of MT sites included in the present regional-scale study.

The middle crust of the southern Abitibi has different reflectivity patterns from the upper crust, and was interpreted by Calvert and Ludden (1999) to be of unknown affinity, with the upper crust representing an allochthonous unit. Benn (2006) re-interpreted the seismic reflection data and suggested that the middle crust consists of tonalite-trondhjemite-granodiorite (TTG) gneiss and paragneiss and that the upper crust and middle crust represent a single, differentiated autochthonous unit. The lower crust of the southern Abitibi is interpreted by both Calvert and Ludden (1999) and Benn (2006) to represent older, pre-3000 Ma crust from beneath the Pontiac Subprovince wedged into the Abitibi Subprovince (Fig. 5.14).

The resistivity model (Fig. 5.14) shows that the middle to lower crust of the Abitibi Subprovince is relatively conductive, and provides no differentiation between the lower two seismic zones. In earlier interpretations (Boerner et al., 2000 and references therein), the enhanced conductivity of the middle crust of Archean terranes of the southern Superior was thought to reflect the present-day state of the crust rather than inherent conductivity within the Precambrian rocks. This interpretation is re-examined below in light of the new results from further south in the Grenville Province.

5.9.2 The Pontiac Subprovince

The Pontiac Subprovince was imaged by seismic lines 16, 16A and part of 15. At the northern end of the Pontiac Subprovince, the interpreted seismic reflection packages dip to the north and extend beneath the surface boundary of the Abitibi Subprovince. Southward thinning fan-shaped packages of reflectors are interpreted to represent the expression of the Pontiac metasedimentary rocks at depth (Calvert and Ludden, 1999). In

the same area the resistivity image shows a northwest dipping boundary between a more resistive zone and more conductive rocks at depth. The more conductive rocks are interpreted to correspond to the metasedimentary rocks, with the increase in conductivity with depth occurring as a result of thrusting. The geophysical results, along with the structural studies of Camire and Burg (1993) and Dimroth et al. (1982) which indicate the presence of south-verging folds in the Pontiac, suggest that the southern Abitibi overthrusts the Pontiac Subprovince.

The upper few kilometers of the Pontiac crust along much of the profile are relatively resistive. This is attributed to the presence of granitic rocks of the Décelles batholith that was intruded towards the end of the accretionary process. The southernmost 10 km of the subprovince is more conductive and correlates with a zone of metagraywacke and pink granite (Kellett et al., 1994). The enhanced conductivity is again attributed to the metasedimentary component of the rocks. As observed in the Abitibi Subprovince the middle and lower crusts in the Pontiac Subprovince are relatively conductive, as previously noted by Kellett et al. (1994).

5.9.3 The Grenville Front

At crustal scale, the observed resistivity structure in the region of the GF and the GFTZ consists of a southeast-dipping resistivity contrast extending from the upper crust to the base of the crust (Fig. 5.14). This contrast separates more conductive rocks to the northwest from very resistive rocks to the southeast. At 10 to 15 km depth, the resistivity contrast correlates spatially with very strong southeast-dipping reflectivity that is interpreted to represent the GF (e.g., Kellett et al., 1994; White et al., 2000). The

correlation of the resistivity and seismic reflection results provides an indication that the GF is represented in the 2-D resistivity model by a boundary between more conductive Archean Pontiac rocks to the northwest and significantly more resistive rocks of the CGB to the southeast. Boerner et al. (2000), based on the earlier work of Kellett et al. (1994), interpreted the increase in resistivity to be associated with an increase in metamorphic grade across the GFTZ from the Archean foreland into the allochthonous rocks of the CGB.

Seismic profiles across different parts of the Grenville Province have indicated different depths of penetration for the GF (Martignole and Calvert, 1996; White et al., 2000). Ludden and Hynes (2000) interpret the penetration of the GF, as a normal fault, throughout the crust and into the mantle. The MT results from the present study provide confirmation that the GF extends to at least the base of the crust. The resistivity contrast can be traced clearly to this depth (Fig. 5.14). This interpretation is also consistent with that of the seismic reflection results by White et al. (2000) that suggest that the Archean footwall rocks are preserved along the length of the GF and extend a significant distance beneath the Grenville Province.

At upper-crustal scale, the 2-D resistivity model from this study supports the earlier interpretations that the GF has a relatively steep dip in the upper 6 to 8 km (e.g., Kellett et al., 1994). The resistivity model includes a prominent resistive block extending to around 20 km depth immediately beneath the GF. This feature is well resolved and appears in all inversion models. This part of the crust was interpreted by Kellett et al. (1994) to correspond to metagraywacke and orthogneiss. However, based on the

observation of significantly lower resistivity in the metagraywackes closer to the surface, the MT results suggest that the resistive unit may consist largely of orthogneiss.

The seismic interpretation of Culshaw et al. (1997) and Green et al. (1988) indicates that the GFTZ is the shear zone that penetrates the entire crust and propagates into the mantle. As noted by previous authors (e.g., Boerner et al., 2000) the resistivity models indicate that, unlike some other major fault zones elsewhere, the GFTZ is not characterized by a significantly enhanced conductivity. However, the result is consistent with observations at some other major Precambrian shear zones, e.g., the Great Slave Lake shear zone in northern Canada (Wu et al., 2002).

5.9.4 The Central Gneiss Belt

The 2-D crustal resistivity model shows that the CGB is characterized by a southeast-dipping package of highly resistive rocks that overly the GFTZ and extends to the lower crust beneath the CMB. These rocks have been interpreted as upper amphibolite facies, reworked, and displaced pre-Grenvillian Laurentia and Paleoproterozoic rocks (Rivers et al., 1989). The high resistivity can be attributed to the relatively low sedimentary component, the high metamorphic grade and the ductile thrusting the rocks underwent during the collisional process. Boerner et al. (2000) comment that the observed resistivity supports the surface geological mapping, by suggesting that there is no foreland basin type sediments preserved in the Archean foreland. However, more conductive rocks are observed at depth further to the southeast. The upper crustal conductive feature located in the CGB to the north of the CMBBZ lies within the Algonquin terrane. This anomaly is interpreted to correspond to a sequence of

paragneissic rocks within this terrane. The enhanced conductivity is attributed to a component of graphitic or sulfidic metasedimentary rocks.

5.9.5 The Central Metasedimentary Belt Boundary Zone

In the 2-D resistivity model, the CMBBZ is interpreted to correspond to a strong resistivity contrast between the resistive rocks of the CGB to the northwest and more conductive rocks to the southeast. This feature is located immediately down dip of the seismic reflections and is oriented parallel to them, allowing it to be confidently reinterpreted in this study as the CMBBZ. The strong resistivity contrast can be traced from around 20 km depth to the base of the crust.

Based on this model, we interpret the CMBBZ as a southeast-dipping feature that extends throughout the crust and possibly propagates into the mantle, rather than rooting into a lower crust decollement as suggested by White et al. (2000). This reinterpretation means that accretion of the Laurentian margin rocks onto Laurentia, between 1080 and 1040 Ma, involved a whole crustal section rather than simple telescoping of the margin rocks onto older Laurentian crust. In this model, the lower crust of Laurentia is interpreted to extend only as far as to beneath the Elzevir terrane of the CMB rather than to further south beneath the Frontenac-Adirondack Belt, as suggested by White et al. (2000). The Sm-Nd data set also provides little evidence for Archean crust beneath the Frontenac-Adirondack Belt (e.g., Dickin, 2000). As with the GFTZ, the CMBBZ is characterized by an absence of enhanced conductivity, an observation which is consistent with the high metamorphic grade and ductile deformation that characterize the zone.

5.9.6 The Central Metasedimentary Belt

The resistivity model of the upper crust in the CMB includes very resistive rocks in the Bancroft terrane and northern part of the Elzevir terrane, and relatively conductive rocks in the southern part of the Elzevir terrane, the Sharbot Lake terrane and in the Frontenac-Adirondack terrane to the south. In seismic reflection Lines 32 and 33, the zone of high resistivity in the upper crust is characterized by strong south-dipping reflectivity, whereas the more conductive area to the south is characterized by horizontal or irregular reflections.

The southeast margin of the resistive upper crust lies in the Elzevir terrane, and is interpreted to correspond to the transition between the Belmont domain to the north and the Grimsthorpe domain to the south. The higher conductivity in the Grimsthorpe domain can be attributed to the presence of volcanoclastic sedimentary rocks in that domain and also to the lower grade of metamorphism. The Belmont domain has been metamorphosed to amphibolite facies, whereas the Grimsthorpe domain records mainly greenschist facies metamorphism. Carr et al. (2000) note that the nature of the Belmont-Grimsthorpe boundary remains uncertain. The resistivity model from the present study suggests a near-vertical boundary between these units.

The relatively conductive upper crust in the Frontenac terrane can be attributed to an increased proportion of graphite or sulfides in the predominantly supercrustal rocks which include quartzofeldspathic gneiss, marble, and quartzite (Carr et al., 2000). The boundary between the rocks of the Composite Arc Belt and the Frontenac-Adirondack Belt is interpreted to be a south-dipping shear zone (Robertson Lake shear zone) extending into a regional mid-crustal decollement (White et al., 2000; Carr et al., 2000).

The resistivity model provides no resolution of either enhanced conductivity or a resistivity contrast across this feature.

5.9.7 The middle and lower crustal resistivity

A number of MT studies in the southern Superior craton have mapped relatively conductive middle and lower crusts. The enhanced conductivity has been observed to cross major structural boundaries, e.g., between the Abitibi and Pontiac subprovinces (Senechal et al., 1996), between the Abitibi and Kapuskasing structural zone (Kurtz et al., 1993), and between the Pontiac Subprovince and the northernmost Grenville terrane (Kellett et al., 1994). In the Kapuskasing structural zone, enhanced conductivity cuts across steeply-dipping structural and seismic reflection fabric (Boerner et al., 2000). These observations led to the interpretation that the enhanced conductivity reflects the present day mechanical and fluid state of the crust rather than an inherent feature of the Precambrian rocks that are present (Mareschal et al. 1995; Boerner et al., 2000). Boerner et al. (2000) noted that for the Abitibi-Grenville region, the interpretation is based only on observations from the Superior and northernmost Grenville crust.

As discussed above, the results from the current study define a well-resolved resistive zone, corresponding to the rocks of the CGB, extending into the lower crust. They also define a localized zone of enhanced conductivity beneath the Frontenac terrane. The dipping nature of the resistive rocks of the CGB, and the localized extent of the Frontenac conductor, indicate control on resistivity by the associated Precambrian rocks. This observation suggests a significant difference in controls on middle and lower crustal resistivity in the Grenville and Superior Provinces.

Boerner et al. (2000) discuss the constraints on the timing of the formation of the conductive middle and lower crusts in the southern Superior. Evidence for widespread tectonic activity in the deep crust of the Superior Province includes 2660-2580 Ma growth of metamorphic zircon and 2520-2400 Ma overgrowths on Archean zircons in kimberlite xenoliths. However, these events predate the formation of the ~1900 Ma Kapuskasing Structural Zone in the Paleoproterozoic in response to the Hudsonian orogeny (Kerrick and Ludden, 2000) and cannot explain the continuity of enhanced conductivity across the boundaries of this structure. Fluid events at 1950-1800 Ma associated with the actual formation of the Kapuskasing Structural Zone (Kerrick and Ludden, 2000) provide a more suitable timing provided that a component of the fluid flux occurred late in the tectonic process.

The rocks in the dipping resistor created by the CGB are of Archean to Mesoproterozoic age. The bounding faults of the structure, the CMBBZ and the GF, were active at around 1080-1060 Ma and ca. 1000 Ma respectively (Carr et al., 2000). Late normal faulting on these shear zones may be as young as 850 Ma (Ludden and Hynes, 2000). The observation of structures of this age cutting across the conductive lower crust implies that the pervasive enhanced mid-crustal conductivity to the northwest is either older than these dates or, if it is younger, it occurs in a restricted geographical area. The observations would appear to exclude the suggestion by Mareschal et al. (1995) that the enhanced conductivity is associated with a present-day ductile lower crust.

The lower crustal conductor that exists in the Frontenac domain is not well resolved in the crustal model. The feature is however consistently present in all crustal models and also appears well-resolved in the lithospheric model (Fig. 5.11). Along with

the poorer fits to the GB decompositions for the CMB, the result suggests that the conductive feature may have a strike that lies closer to east-west than the 45° azimuth used in the 2-D inversions of the crustal data set.

Enhanced electrical conductivity in the lower continental crust is usually interpreted to be caused by fluids, graphite, and/or sulfide-bearing rocks of sedimentary sequences that have undergone complex deformation such as thrusting (Schwarz, 1990; Jones, 1992; Korja et al., 1996). The localized geometry of the conductive zone would be very difficult to explain in terms of a fluid source. The Frontenac terrane consists of a succession of multiply deformed, relatively low-pressure, granulite facies, quartzofeldspathic and pelitic gneiss, quartzite and marble intruded by plutonic rocks and subjected to metamorphism (Easton and Davidson, 1997). Easton (1992) suggested that the Frontenac terrane is different from other terranes within the CMB in terms of composition (it lacks volcanic rocks) and in that it is preserved granulite facies rocks in contrast to the lower grade facies to its north (Bancroft and Elzevir) and south (Adirondack lowlands). The hypothetical stratigraphy of this terrane consists of two stacked sets of tripartite units composed from bottom to top, gneiss, quartzite and marble (Hilderbrand and Easton, 1995). The marble in the Frontenac terrane is composed of calcite with varying proportions of graphite, serpentine and calc-silicate minerals (Hilderbrand and Easton, 1995). The lower conductor is thus fairly reliably interpreted to be caused by graphitic rocks of the CMB.

5.10 Interpretation and discussion of lithospheric mantle resistivity structure

5.10.1 High-resistivity zone beneath the northwest Grenville Province

Figure 5.17 shows the 2-D resistivity model of the study area along with our interpretation of the prominent features. An important feature of the lithospheric resistivity model is the southeast-dipping resistive feature that extends from ~40 to 140 km depth. This northwest-striking resistivity anomaly extends laterally from GF into the middle of the CMB at depth and includes resistivity values exceeding 2,000 $\Omega\cdot\text{m}$. As described above there is excellent spatial correlation of the high resistivity anomaly with a seismic high velocity anomaly.

Two explanations for the seismic high velocity zone have been proposed: that it is either associated with depletion of iron in the cratonic lithosphere (Chen and Li, 2012) and/or that it is a relict slab associated with the Early Proterozoic subduction beneath the CMB (Aktas and Eaton, 2006). The high velocity observed in the feature can be explained petrologically by either an increased Mg to Fe ratio in a peridotitic or harzburgitic mantle, or to the presence of eclogite and garnet phases produced by plate subduction (Chen and Li, 2012).

An increased Mg to Fe ratio, for example associated with depletion of the mantle, would also be associated with increased electrical resistivity (e.g., Jones et al., 2009a; Fulla et al., 2011; Selway, 2014). Petrological studies of mantle xenoliths from different parts of the world have confirmed the existence of systematic variations in the composition of the lithospheric mantle due to depletion of the cratonic mantle in basaltic components, primarily iron, aluminum and volatiles, which results in the increase of the

upper mantle resistivity (Artemieva, 2009). The first explanation for the high velocity is therefore in accord with the resistivity results.

The second explanation of the high velocity is less compatible with the MT results in this study as the resistivity model differs significantly from models commonly observed in plate subduction settings. In regions of active subduction, resistivity anomalies are dominated by enhanced conductivity rather than dipping resistors (e.g., Soyer and Unsworth, 2006; Jödicke et al., 2006). The enhanced conductivity, which may form localized features, is interpreted to be caused by fluids released by the subducting slab. The resulting anomalies have been observed to persist for a long duration following the cessation of subduction (e.g., Ledo et al., 2004) and it is generally expected that refertilization of the mantle by the fluids will lead to long-lived enhanced conductivity (Selway, 2014). The relatively conductive uppermost mantle lithosphere observed downdip from the southeastern CGB and CMB in Fig. 5.11 could be interpreted as enhanced conductivity associated with subduction, but the thick resistive slab represents an unusual observation in a subduction setting.

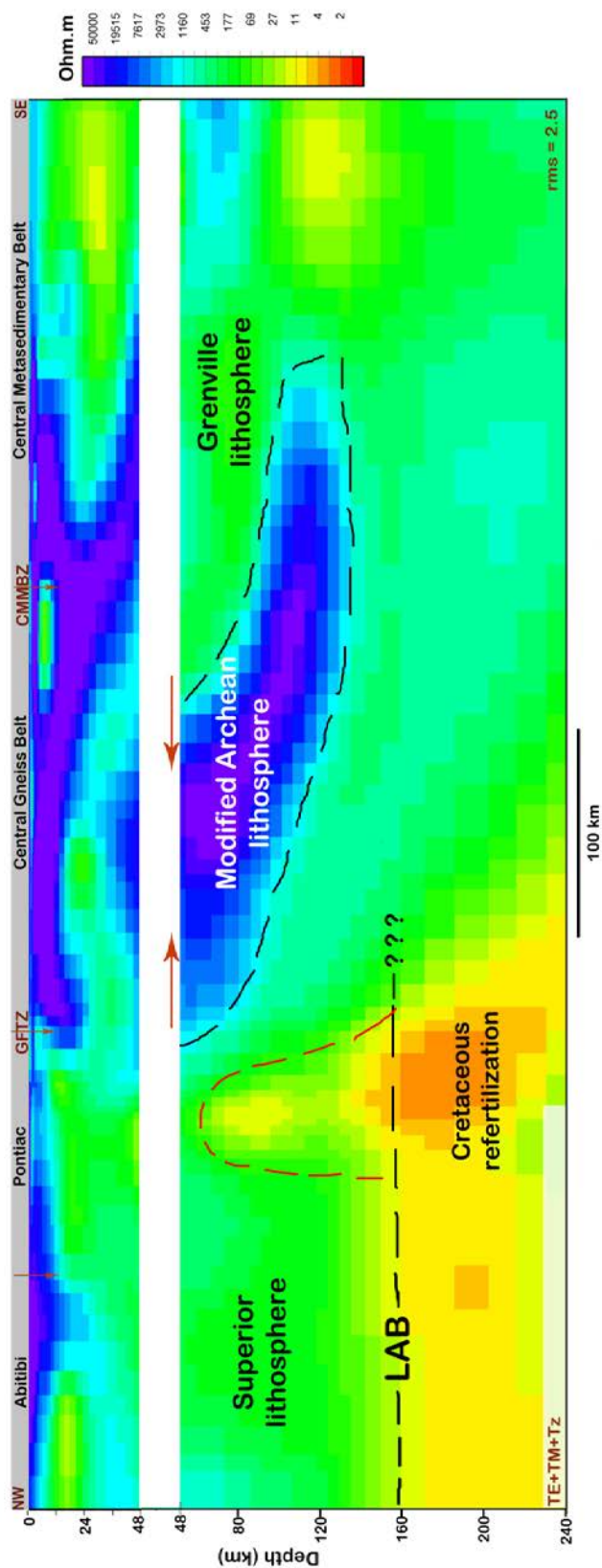
The mantle lithosphere beneath the northern Grenville Province may have formed by one of two mechanisms. If the lithosphere is in situ and the same age as the overlying upper crustal rocks, it must have developed during the series of long-lived orogenies that accreted Andean arc and back arc rocks onto the Laurentian margin between 1.9 and 1.2 Ga. However, there is strong evidence to suggest that the accretion of the Proterozoic rocks before and during the Grenville orogen occurred onto Archean basement of the Superior Province. The observation in both seismic reflection images and electrical resistivity images (this study) of the Laurentian margin rocks extending some distance to

the south in the middle and lower crusts provides support for this interpretation. In other studies, Ludden and Hynes (2000) suggest that an Archean lithospheric root was responsible for providing a stable, cold, and relatively low density basement to the Grenvillian orogen. Hynes and Rivers (2010) inferred that the central Grenville Province is underlain at depth by Archean rocks of the Superior Province. Finally, within the study area there is no major change in the shear wave velocity at 150 km depth at the location of the surface margin of the Superior and Grenville Provinces (Faure et al., 2011) providing further support for the fact that the Grenville Province is underlain by the Archean mantle of the Superior Province.

The values of resistivity observed in the resistor beneath the northern Grenville Province are extremely high. Comparison of the Grenville results with a deep 2-D resistivity model crossing Proterozoic and Archean blocks in southern Africa (Muller et al., 2009) shows that in terms of the resistivity values, the upper 100 km in the Grenville lies between that observed in the East and West Kimberley blocks of the Archean Kaapvaal Craton and is approximately an order of magnitude more resistive than the lithosphere in adjacent Proterozoic terranes, including the Ghanzi-Chobe and Damara belts (Fig. 5.18).

The MT results do not provide definitive support for either of the proposed mechanisms of formation of the high velocity-high resistivity zone. The geometry of the high resistivity feature in the MT model includes a well-resolved dip to both the upper and lower surfaces, as would be expected for an anomaly related to a relict slab feature. The results also show that the base of the feature is relatively well defined (Fig. 5.18). However, as discussed above, it is presently unclear how a relict slab could create such

Figure 5.17. Interpretation of features in lithospheric mantle resistivity model. The figure shows the 2-D lithospheric resistivity model ($V.E = 1$), interpretation of the resistive block in the shallow mantle lithosphere as being modified Archean lithosphere, interpretation of the conductive zone at 160 km depth as the LAB beneath the southeastern Superior province, and interpretation of a localized conductor as a lithospheric mantle scar refertilized in the Cretaceous by passage of the Great Meteor hotspot. The correct model (for the strike azimuth appropriate at crustal signal penetration) is shown above the lithospheric model. The arrows show the part of the model used to define the averaged resistivity profile shown in Figure 5.18.



highly resistive features, especially in the absence of any associated conductive anomalies. In addition, if this feature were to be a relict slab, it should be located down dip from the CMB or further southeast. In contrast to the geometry of the high resistivity zone, the actual values of resistivity observed, provide support for formation of the feature by high levels of mantle depletion.

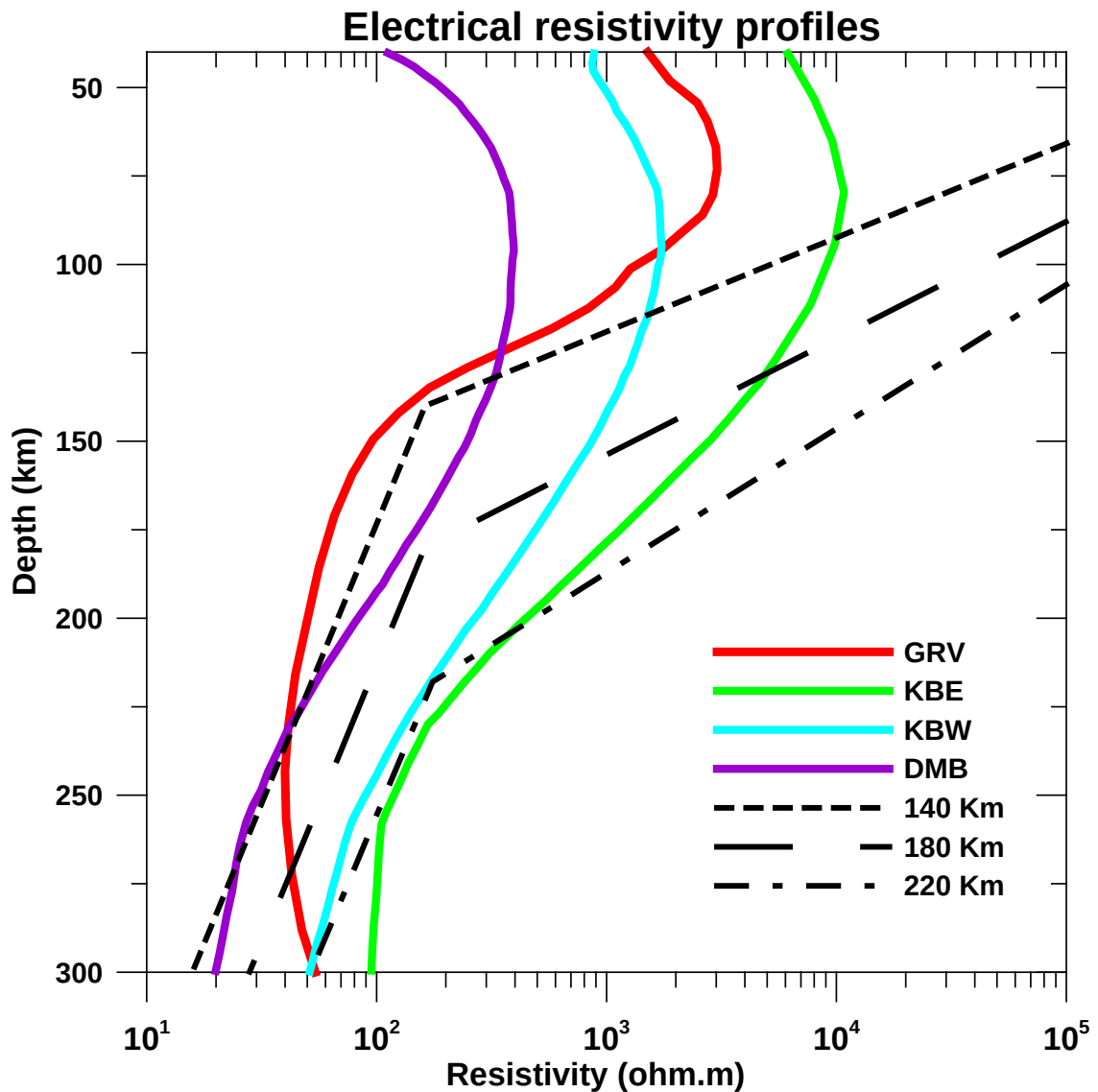
5.10.2 Deep, laterally-extensive, horizontal conductive anomaly

An important feature of the 2-D lithospheric resistivity model is the lithospheric conductor (C1) that extends horizontally beneath the southern Superior Province. The top of this feature is observed at around 160 km depth (Fig. 5.17).

A common interpretation for deep (>150 km) laterally-continuous conductors is that they represent the electrical signature of the lithosphere-asthenosphere boundary (eLAB) (Jones et al., 2010). The LAB depth has been estimated from a sharp change in mantle conductivity that is interpreted to be associated with the onset of partial melt (Artemieva, 2009; Jones et al., 2010) or of enhanced bound water in olivine and pyroxenes (Karato, 1990, 2012). Muller et al. (2009) describe an alternative method for estimating the LAB using comparison of resistivity-depth profiles derived for predicted geotherms with observed resistivity-depth profiles. This approach assumes a dominant thermal control on resistivity.

Fischer et al. (2010), using shear wave velocity, suggested that the seismically-defined LAB depth of the Superior Province ranges from 150 to 220 km. Darbyshire et al. (2007) suggested that the lithospheric thickness of the Superior Province in the study area varies from 100 to 220 km. Within the study area the estimated depth of the

Figure 5.18. Comparison of electrical resistivity depth profiles for the northern Grenville Province (GRV; this study), with the profiles for the Ghanzi-Chobe/Damara Block (DMB), Western Kimberley Block (KBW), and Eastern Kimberley Block (KBE; Muller et al., 2009). The average profile for the Grenville Province is computed for the area shown in Figure 5.17. The theoretical resistivity-depth profiles calculated by Muller et al. (2009) for mantle geotherms of different lithospheric thicknesses are also shown in black. The laboratory measurements of the electrical-conductivity against pressure and temperature for dry olivine and pyroxene are used to compute the profiles (Constable et al., 1992; Xu and Shankland, 1999; Xu et al., 2000). The points of inflection on these curves coincide with the intersection between the adiabat and the conductive mantle geotherm (Muller et al., 2009).



petrologic LAB beneath the southern Superior Province, has been estimated using kimberlites from Cobalt and Kirkland Lake, to be 160 km (Griffin et al., 2004) and 144 to 150 km (Snyder and Griitter, 2010).

We interpret the horizontal portion of the lithospheric conductor in the southern Superior Province to be the eLAB. Our suggestion is supported by the success of electrical LAB depth interpretations elsewhere (Jones et al., 2010) and by the good agreement between the depth of 160 km indicated by the resistivity results and the seismic and petrological results for the southern Superior area. Figure 5.18 shows theoretical resistivity profiles calculated by Muller et al. (2009) based on thermal models for different thicknesses of the lithosphere. For all models, the LAB coincides with resistivity values of 100 to 150 $\Omega\cdot\text{m}$. As shown in Figure 5.11, this value of resistivity is attained at a depth of about 140 km beneath the southeastern part of the Superior craton. As noted in previous studies, the LAB depth is relatively shallow for an Archean craton. Faure et al. (2011) also show that the lithospheric seismic velocity at 150 km depth beneath the southern Superior craton is significantly less than farther to the north in the craton. These results have been attributed to the extensive tectonic processes that have affected this region following formation of the craton, e.g., during the 2.4 Ga events associated with formation of the Huronian basin and the 1.8 Ga events associated with the Trans-Hudson orogen. In addition, the earlier postulated mantle plumes (Faure et al., 2011) and lithospheric subduction-dominated processes, associated with the formation of the Abitibi Greenstone Belt during the Neoproterozoic, may have modified the lithospheric mantle in this area (Daigneault et al., 2004).

5.10.3 Deep, subvertical conductivity anomaly

Another important feature of the lithospheric 2-D resistivity model is the subvertical conductor C2 beneath the Pontiac Subprovince. This conductor overlies enhanced conductivity at the depth of the interpreted LAB (Fig. 5.17).

Previous studies have concluded that the seismic low-velocity anomaly, farther to the south in the lithospheric mantle of this region, was formed as a result of the geochemical alteration of the mantle due to the passage of the Cretaceous Great Meteor hotspot, which is the last tectonothermal event that affected this region (van der Lee and Nolet, 1997; Rondenay et al., 2000; Aktas and Eaton, 2006; Eaton and Frederiksen, 2007; Chen and Li, 2012). The alkaline volcanism of the Great Meteor hotspot is responsible for the emplacement of features along the Montereign-White Mountain-New England seamount track (Crough, 1981). Faure et al. (2011) however suggested that the low velocity zone is produced by tectonothermal events (Early Proterozoic to Cretaceous) that have reused the permanent mantle scar left by the early modification of the mantle during the formation of greenstone belts in the Neoarchean. This zone correlates spatially with the Neoarchean Greenstone Belt of the Abitibi Subprovince and its plume-driven subduction zone. Rondenay et al. (2000) and Faure et al. (2011) relate the low velocity anomaly to the magmatism responsible for the emplacement of the Kirkland Lake and Cobalt (Rapide des Quinze) kimberlite fields.

Selway (2014) reviewed sources of enhanced conductivity in tectonically stable lithosphere. She suggested that the electrical conductivity of stable lithosphere is mainly affected by hydrogen content of nominally anhydrous mantle minerals and grain boundary graphite films. A subduction or plume event may have introduced hydrogen

and carbon into Archean lithosphere as much as several billion years ago and, if no high temperature events have occurred to remove those species, they will still remain to produce a conductive anomaly (Selway, 2014). However, it should be noted that not all regions of the mantle lithosphere in which metasomatism has occurred have decreased resistivity (Jones et al., 2002).

Considering both the controls on mantle resistivity and the results of the earlier seismic studies we interpret the enhanced conductivity in the sub-vertical conductor to have resulted from the re-fertilization of the mantle scar, initially created by an early thermomechanical event responsible for the formation of the Abitibi Greenstone Belt, by the Cretaceous Great Meteor hotspot plume. We propose that the alkaline rich mantle plume, associated with the Great Meteor hotspot, refertilized the mantle by re-introducing grain-boundary graphite films and possibly other incompatible elements by intrusion of melt through a weak zone within the lithosphere, thereby increasing the conductivity of the zone. Following this interpretation, it is concluded that the conductivity structure is of Cretaceous age. Faure et al. (2011) pointed out that plumes that accumulate around the LAB interface form unstable pockets that migrate by excess pressure vertically through sub-vertical zones of weakness. This description is consistent with the observation of enhanced conductivity at the LAB beneath the subvertical conductor and indicates the existence of a complex interaction between the postulated mantle plume and the base of the lithosphere. This interaction is generally controlled by the geometry of the LAB or the lithospheric zones of weakness.

It is of note that the Mesozoic Kirkland Lake and Cobalt kimberlite fields are located about 50 km along strike from the lithospheric conductor (C2). It is therefore

possible that diamondiferous kimberlites passed through the part of the region of lithosphere containing the subvertical conductor. The conductor itself lies above the diamond-graphite stability field at 140 -150 km depth and may therefore be caused by graphite. However, the enhanced conductivity at the LAB lies within the diamond stability field and must therefore be explained in terms of mantle enrichment by other elements.

Jones et al. (2001) imaged a conductor beneath the Slave Craton in northern Canada, the Central Slave Mantle Conductor (CSMC). As observed in the present study, the conductor seems confined to depths of less than 130 km. The CSMC is collocated with an Eocene-aged kimberlite field in the Slave Craton. In earlier interpretations, the CSMC was interpreted to be due to carbon, either as grain boundary film or as graphite or water associated with subduction beneath the Slave craton (Jones et al., 2003). However, in a recent re-interpretation by Selway (2014), the enhanced conductivity is attributed to initial re-fertilization of the lithosphere by subduction and subsequent remobilization of the conducting constituents due to kimberlite emplacement (Selway, 2014). It is possible that such an interpretation also applies in the southern Superior, with the initial introduction of carbon and water into the mantle having occurred in the Archean and/or Proterozoic events with the remobilization occurring in the Cretaceous. In this case, the enhanced conductivity and decreased velocity observed beneath 160 km depth in this region may be attributed to the presence of refertilized lithosphere rather than to asthenosphere.

5.11 Conclusions

The MT data, collected at sites in southern Ontario, during the Lithoprobe-Abitibi Grenville Transect and POLARIS project, are used to image the Proterozoic Grenville Province and its margin with the Archean Superior Province.

The strike results show a different geoelectric strike azimuth for the crust, lithosphere and the asthenosphere of this region. The crustal geoelectric strike (45°) is subparallel to the surface geologic strike of the Grenville Front and the CMBBZ. The lithospheric mantle strike direction of $N85^{\circ}E$, agrees well with the $N80^{\circ}E$ defined by earlier studies, and the new observations extend the region of this strike direction farther south into the Grenville province.

A 2-D resistivity model determined using the crustal strike azimuth delineates a number of crustal structures. The upper crust of the Abitibi Subprovince is resistive and the middle to lower crust is relatively conductive. The upper crust in the Pontiac Subprovince contains some areas of increased conductivity that are attributed to the metasedimentary component of the rocks in the area. The new MT results, from farther south in the Grenville Province, provide confirmation that the GF extends to at least the base of the crust. Based on the resistivity model, the CMBBZ is interpreted as a southeast-dipping feature that extends throughout the crust rather than rooting into a lower crust decollement as suggested by White et al. (2000). Graphite-bearing rocks of the CMB are suggested to be responsible for the lower crustal conductor imaged beneath the Frontenac terrane.

The crustal resistivity model includes very resistive Laurentian margin rocks dipping southeast to the base of the crust, bounded to the northwest by the GF, and to the

southeast by the CMBBZ. The observation is in contrast to the observation elsewhere in the region of conductive mid to lower crust and requires revision of the interpretation that the enhanced conductivity reflects present day geodynamic conditions. If a single event caused the enhanced conductivity, it must be younger than the 1950-1800 Ma formation of the Kapuskasing Structural Zone which is crosscut by the enhanced conductivity, and older than the 1080-1060 Ma age of the CMBBZ and the ca. 1000 Ma age of the GF which crosscut the enhanced conductivity. The Superior Province crust records a 1950-1800 Ma fluid event associated with the formation of the Kapuskasing Structural Zone which may explain the observations as long as a component of the fluid flux postdates the main deformation.

A 2-D resistivity model for the lithospheric mantle was derived using the corresponding strike azimuth. It defines a very resistive lithospheric layer that extends between approximately 40 and 140 km vertically and laterally from the GF southeast into the middle of the CMB. The feature corresponds to a high velocity zone in seismic tomographic models. The resistivity of this feature is comparable to that observed in depleted Archean lithosphere in southern Africa and suggests that the lithosphere has not been re-fertilized by young tectonic process (e.g., Selway, 2014). The geometry of the body suggests that it may be a relict slab of Superior craton lithosphere. As noted by previous authors, the presence of modified Archean lithosphere beneath the Grenville Province would have provided the necessary basement for the Grenvillian orogeny.

A lithospheric conductor that extends horizontally beneath the southern Superior Province with its top at 160 km is interpreted to represent the eLAB beneath that craton. The observed depth of the feature is in agreement with seismic and petrological

information for the area. A sub-vertical conductor observed in the lithospheric mantle beneath the Pontiac Subprovince overlies a region of enhanced conductivity at about 150 km depth. This feature lies well to the north of a sub-vertical low seismic velocity zone delineated in seismic tomography studies. We interpret the enhanced conductivity in the sub-vertical conductor to have resulted from re-fertilization of the mantle scar, initially created by an early thermomechanical event responsible for the formation of the Abitibi Greenstone Belt, by the Cretaceous Great Meteor hotspot plume. The location of the sub-vertical lithospheric conductor, which is along strike with the Kirkland Lake and Cobalt field kimberlite fields, also drives us to suggest that the diamondiferous kimberlites being produced in these fields passed through this part of the lithosphere. The assertion that MT is capable of imaging such enrichment is important for the interpretation of lithospheric structure and evolution. More importantly, the ability of MT to image such regions is useful for diamond exploration, as espoused by Jones and Craven (2004) and Jones et al. (2009b).

5.12 Acknowledgments

The authors thank the POLARIS southern Ontario MT team (J. Wenham, M. Serzu, X. Ma, E. Gowan, H. Sealey, T. Pacha, J. McCutcheon, B. Eade, I. Asudeh, C. Andrews) for MT data collection and landowners in the region for site access. Comments from A. Frederiksen, A. Camacho, A. Rajapakse, and two anonymous reviewers have led to significant improvement of the manuscript. The current study could not have taken place without M. Mareschal's important earlier work on the Abitibi-Grenville LITHOPROBE MT data. Funding for POLARIS has been provided by industry, the

Canada Foundation for Innovation, provincial agencies from Ontario and British Columbia, the Federal Economic Development Initiative for Northern Ontario, Natural Resources Canada, and NSERC (Natural Sciences and Engineering Research Council of Canada). The present research was supported by an NSERC Discovery Grant (105748) to Ian Ferguson.

5.13 References Cited

- Adetunji A.Q, Ferguson, I.J., Jones, A.G., Imaging the mantle lithosphere of the Grenville Province: large-scale structures and electrical anisotropy. (submitted in Jan. 2014)
- Agarwal, A. K., Poll, H. E., Weaver, J. T., 1993. One and two-dimensional inversion of MT data in continental regions. *Phys. Earth Planet. Inter.* 81, 155-176.
- Aktas, K., Eaton, D. W., 2006. Upper-mantle velocity structure of the lower Great Lakes region. *Tectonophysics* 420, 267–281.
- Artemieva, I. M., 2009. The continental lithosphere: Reconciling thermal, seismic, and petrologic data. *Lithos* 109, 23–46.
- Bahr, K., 1988. Interpretation of the magnetotelluric impedance tensor: regional induction and local telluric distortion. *J. Geophys.* 62, 119–127.
- Benn, K., 2006. Tectonic Delamination of the lower crust during Late Archean Collision of the Abitibi-Opatika and Pontiac terranes, Superior Province, Canada. *Archean Geodynamics and Environments, Geophysical Monograph Series* 164, American Geophysical Union 10.1029/164GM17.
- Benn, K., Miles, W., Ghassemi, M. R., Gillett, J., 1994. Crustal structure and kinematic framework of the northwestern Pontiac Subprovince, Quebec: An integrated structural and geophysical study. *Can. J. Earth Sci.* 31, 271-281.
- Boerner, D. E., Kurtz, R. D., Craven, J. A., 2000. A summary of electromagnetic studies on the Abitibi-Grenville transect. *Can. J. Earth Sci.* 37, 427-437.
- Cagniard, L., 1953. Basic theory of the magnetotelluric method of geophysical prospecting. *Geophysics* 18, 605-635.
- Caldwell, T. G., Bibby, H. M., Brown, C., 2004. The magnetotelluric phase tensor. *Geophys. J. Int.* 158, 457–469.
- Calvert, A. J., Ludden, J. N., 1999. Archean continental assembly in the southeastern Superior Province of Canada. *Tectonics* 18, 412-429.
- Camire, G. E., Burg, J. –P., 1993. Late Archean thrusting in the northwestern Pontiac Subprovince, Canadian Shield. *Precambrian Research* 61, 51-66.
- Card, K. D., 1990. A review of the Superior Province of the Canadian Shield, a product of Archean accretion. *Precambrian Research* 48, 99-156.
- Card, K. D., Ciesielski, A., 1986. Subdivisions of the Superior Province of the Canadian Shield. *Geoscience Canada* 13, 5-13.
- Card, K. D., Poulsen, K. H., 1998. Geology and mineral deposits of the Superior Province of the Canadian Shield, in: Lucas, S. B. (Ed.), *Geology of the Precambrian Superior and Grenville Provinces and Precambrian Fossils in North America*, *Geology of Canada Series*, No. 7, Geological Survey of Canada, pp.13-194.
- Carr, S. D., Easton, R. M., Jamieson, R. A., Culshaw, N. G., 2000. Geologic transect across the Grenville orogen of Ontario and New York. *Can. J. Earth Sci.* 37, 193-216.
- Chave, A. D., Jones, A. G. (Eds.), 2012. *The Magnetotelluric Method - Theory and Practice*, Cambridge University Press, New York, 570pp.
- Chen, C. -W., Li, A., 2012. Shear wave structure in the Grenville Province beneath the lower Great Lakes region from Rayleigh wave tomography. *J. Geophys. Res.* 117, B01303.

- Constable, S. C., 2006. SEO3: a new model of olivine electrical conductivity. *Geophys. J. Int.* 166, 435-437.
- Constable, S. C., Shankland, T. J., Duba, A., 1992. The electrical conductivity of an isotropic olivine mantle. *J. Geophys. Res.* 97, 3397-3404.
- Corrigan, D., van Breemen, O., 1997. U–Pb age constraints for the lithotectonic evolution of the Grenville Province along the Mauricie transect, Quebec. *Can. J. Earth Sci.* 34, 299–316.
- Crough, S.T., 1981. Mesozoic hotspot epeirogeny in eastern North America. *Geology* 9, 2-6.
- Culshaw, N. G., Davidson, A., Nadeau, L., 1983. Structural subdivisions of the Grenville Province in the Parry Sound-Algonquin region, Ontario, in: *Current Research Part B*, Geological Survey of Canada, paper 83-1B, pp. 243-252.
- Culshaw, N. G., Jamieson, R. A., Ketchum, J. W. F., Wodicka, N., Corrigan, D., Reynolds, P. H., 1997. Transect across the northwestern Grenville orogen, Georgian Bay, Ontario: Polystage convergence and extension in the lower orogenic crust. *Tectonics* 16, 966-982.
- Daigault, R., Mueller, W. U., Chown, E. H., 2004. Abitibi greenstone belt plate tectonics: the diachronous history of arc development, accretion and collision, in: Eriksson, P., Altermann, W., Nelson, D., Mueller, W., Catuneanu, O. (Eds.), *Tempos of events in Precambrian time*, *Developments in Precambrian Geology*, Elsevier, London, vol. 12, pp. 88–102, ,
- Darbyshire, F.A., Eaton, D. W., Frederiksen, A. W., Ertolahti, L., 2007. New insights into the lithosphere beneath the Superior Province from Rayleigh wave dispersion and receiver function analysis. *Geophys. J. Int.* 169, 1043-1068.
- Davidson, A., 1986. New interpretations in the southwestern Grenville Province, in: Moore, J. M., Davidson, A., Baer, A. J. (Eds.), *The Grenville Province*, Geological Association of Canada, Geological Survey of Canada, Special paper 31, pp. 61-74.
- Davidson, A., 1998. Questions of correlation across the Grenville Front east of Sudbury, Ontario; in: *Current Research 1998-C*; Geological Survey of Canada. Paper 98-IC, pp. 145-154.
- Davidson, A., Culshaw, N. G., Nadeau, L., 1982. A tectono-metamorphic framework for part of the Grenville Province, Parry Sound region, Ontario, in: *Current Research Part A*, Geological Survey of Canada, paper 82-1A, pp. 175-190.
- Davidson, A., Morgan, W. C., 1981. Preliminary notes on the geology east of Georgian Bay, Grenville Structural Province, Ontario, in: *Current Research*, Geological Survey of Canada, paper 81-1A, pp. 291-298.
- Demets, C., Gordon, R., Argus, D., Stein, S., 1990. Current plate motions. *Geophys. J. Int.* 101, 425-478.
- Dickinson, A. P., 2000. Crustal formation in the Grenville: Nd-isotope evidence. *Can. J. Earth Sci.* 37, 165-181.
- Dimroth, E., Imreh, L., Rocheleau, M., Goulet, N., 1982. Evolution of the south-central segment of the Archean Abitibi Belt, Quebec, Part I, Stratigraphy and paleogeographic model. *Can. J. Earth Sci.* 19, 1729-1758.
- Ducea, M. N., Park, S. K., 2000. Enhanced mantle conductivity from sulfide minerals, southern Sierra Nevada, California. *Geophys. Res. Lett.* 27, 2405-2408.

- Easton, R. M., 1992. The Grenville Province and the Proterozoic history of Central and Southern Ontario, in: *Geology of Ontario*, Ontario Geological Survey, Special Volume 4, 713 – 904.
- Easton, R. M., Davidson, A., 1997. Frontenac-Sharbot Lake relationships: evaluation of a provocative hypothesis concerning the tectonics of the Central Metasedimentary Belt, Ontario. *Friends of the Grenville, Annual Field Excursion Guidebook*, pp. 26-28.
- Eaton, D. W., Adams, J., Asudeh, I., Atkinson, G. M., Bostock, M. G., Cassidy, J. F., Ferguson, I. J., Samson, C., Snyder, D. B., Tiampo, K. F., Unsworth, M. J., 2005. Geophysical arrays to investigate lithosphere and earthquake hazards in Canada. *EOS Trans AGU* 86, 169-173.
- Eaton, D. W., Dineva, S., Mereu, R., 2006. Crustal thickness and Vp/Vs variations in the Grenville orogen Ontario, Canada from analysis of teleseismic receiver functions. *Tectonophysics* 420, 223-238.
- Eaton, D. W., Frederiksen, A., 2007. Seismic evidence for convection driven motion of the North American plate. *Nature* 446, 428–431.
- Evans, R. L., Jones, A. G., Garcia, X., Muller, M., Hamilton, M., Evans, S., Fourie, C. J. S., Spratt, J., Webb, S., Jelsma, H., Hutchins, D., 2011. Electrical lithosphere beneath the Kaapvaal craton, southern Africa. *J. Geophys. Res.* 116, B04105.
- Faure S., Godey, S., Fallara, F., Trepanier, S., 2011. Seismic architecture of the Archean North American mantle and its relationship to diamondiferous kimberlite fields. *Econ. Geol.* 106, 223-240.
- Ferguson, I. J., Jones, A. G., Chave, A. D., 2012. Case histories and geological applications, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 480-544.
- Ferguson, I. J., Jones, A. G., Yu, S., Wu, X., Shiozaki, I., 1999. Geoelectric response and crustal electrical-conductivity structure of the Flin Flon Belt, Trans-Hudson Orogen, Canada. *Can. J. Earth Sci.* 36, 1917-1938.
- Ferguson, I.J., Craven, J.A., Kurtz, R. D., Boerner, D. E., Bailey, R. C., Wu, X., Orellana, M. R., Spratt, J., Wennberg, G., Norton, M., 2005. Geoelectric responses of Archean lithosphere in the western Superior Province, central Canada. *Phys. Earth Planet. Int.* 150, 123-143.
- Fischer, K. M., Ford, H. A., Abt, D. L., Rychert, C. A., 2010. The lithosphere-aesthenosphere boundary. *Annu. Rev. Earth Planet. Sci.* 38, 551-575.
- Fouch, M. J., Fischer, K. M., Parmentier, E. M., Wysession, M. E., Clarke, T. J., 2000. Shear wave splitting, continental keels, and patterns of mantle flow. *J. Geophys. Res.* 105, 6255-6275.
- Frederiksen, A.W., Ferguson, I. J., Eaton, D., Miong, S. K., Gowan, E., 2006. Mantle fabric at multiple scales across an Archean-Proterozoic boundary, Grenville Front, Canada. *Phys. Earth Planet. Int.* 158, 240-263.
- Fullea, J., Muller, M. R., Jones, A. G., 2011. Electrical conductivity of continental lithospheric mantle from integrated geophysical and petrological modeling: Application to the Kaapvaal Craton and Rehoboth Terrane, southern Africa. *J. Geophys. Res.* 116, B10202, doi: 10.1029/2011jb008544.
- Glover, P.W. J., Vine, F. J., 1995. Beyond KTB - Electrical conductivity of the deep continental crust. *Surv. Geophys.* 16, 5-36.

- Gowan, E. J., Ferguson, I. J., Jones, A. G., Craven, J. A., 2009. Geoelectric structure of the northeastern Williston basin and underlying Precambrian lithosphere. *Can. J. Earth Sci.* 46, 441-464.
- Green, A. G., Cannon, W. F., Milkereit, B., Hutchinson, D. R., Davidson, A., Behrendt, J. C., Spencer, C., Lee, M. W., Morel-a-l'Huissier, P., Agena, W. F., 1989. A "Glimpce" of the deep crust beneath the Great Lakes, in *Properties and Processes of Earth's Lower Crust*. AGU Monograph 51, pp. 65-80. doi:10.1029/GM051p0065.
- Green, A. G., Milkereit, B., Davidson, A., Spencer, C., Hutchinson, D. R., Cannon, W. F., Lee, M. W., Agena, W. F., Behrendt, J. C., Hinze, W. J., 1988. Crustal structure of the Grenville Front and adjacent terranes. *Geology* 16, 788-792.
- Griffin, W. L., O'Reilly, S. Y., Doyle, B. J., Pearson, N. J., Coopersmith, H., Kivi, K., Malkovets, V., Pokhilenko, N., 2004. Lithosphere mapping beneath the North American plate. *Lithos* 77, 873-922.
- Gripp, A. E., Gordon, R. G., 2002. Young tracks of hotspots and current plate velocities. *Geophys. J. Int.* 150, 321-361.
- Groom, R. W., Bailey, R. C., 1989. Decomposition of magnetotelluric impedance tensors in the presence of local three-dimensional galvanic distortion. *J. Geophys. Res.* 94, 1913-1925.
- Groom, R. W., Bailey, R. C., 1991. Analytic investigations of the effects of near-surface three-dimensional galvanic scatterers on MT tensor decompositions. *Geophysics* 56, 496 – 518.
- Haggart, M. J., Jemieson, R. A., Reynold, P. H., Krogh, T. E., Beaumont, C., Culshaw, N. G., 1993. Last gasp of the Grenville Orogeny: thermochronology of the Grenville Front Tectonic Zone near Killarney, Ontario. *J. Geology* 101, 575-589.
- Hamilton, M. P., Jones, A. G., Evans, R. L., Evans, S., Fourie, C. J. S., Garcia, X., Mountford, A., Spratt, J. E., the SAMTEX Team, 2006. Electrical anisotropy of South African lithosphere compared with seismic anisotropy from shear-wave splitting analysis. *Phys. Earth Planet. Int.* 158, 226-239.
- Hanmer, S., 1988. Ductile thrusting at mid-crustal level, southwestern Grenville Province, Canada. *Can. J. Earth Sci.* 25, 1049-1059.
- Hanmer, S., Ciesielski, A., 1984. A structural reconnaissance of the northwest boundary of the Central Metasedimentary Belt, Grenville Province, Ontario and Quebec, in: *Current Research Part B, Geological Survey of Canada, Paper 84-18*, pp. 121-131.
- Hanmer, S., McEachern, S. J., 1992. Kinematical and rheological evolution of a crustal-scale ductile thrust zone, Central Metasedimentary Belt, Grenville Orogen, Canada. *Can. J. Earth Sci.* 29, 1779-1790.
- Hanmer, S., Thivierge, R. H., Henderson, J. R., 1985. Anatomy of a ductile thrust zone: part of the northwest boundary of the Central Metasedimentary Belt, Grenville Province, Ontario, in: *Current Research Part B, Geological Survey of Canada, Paper 85-18*, pp. 1-5.
- Hildebrand, R. S., Easton, R. M., 1995. A 1161 Ma suture in the Frontenac terrane, Ontario segment of the Grenville orogeny. *Geology* 23, 917-920.
- Hirth, G., Evans, R. L., Chave, A. D., 2000. Comparison of continental and oceanic mantle electrical conductivity: Is the Archean lithosphere dry? *Geochem. Geophys. Geosyst.* 1, 1030, doi:10.1029/2000GC000048.

- Hoffman, P. F., 1989. Precambrian geology and tectonic history of North America, in: Bally, A. W., Palmer, A. R. (Eds.), *The Geology of North America - An Overview*, The Geological Society of America, vol. A, pp. 447-512.
- Hyndman, R. D., Wang, K., Yuan, T., Spence, G. D., 1993. Tectonic sediment thickening, fluid expulsion, and the thermal regime of subduction zone accretionary prisms: The Cascadia margin off Vancouver Island. *J. Geophys. Res.* 98, 21865-21876, doi: 10.1029/93JB02391.
- Hynes, A., Rivers, T., 2010. Protracted continental collision - evidence from the Grenville Orogen. *Can. J. Earth Sci.* 47, 591-620.
- Ji, S., Rondenay, S., Mareschal, M., Senechal, G., 1996. Obliquity between seismic and electrical anisotropies as a potential indicator of movement sense for ductile shear zones in the upper mantle. *Geology* 24, 1033-1036.
- Jödicke, H., Jording, A., Ferrari, L., Arzate, J., Mezger, K., Rüpke, L., 2006. Fluid release from the subducted Cocos plate and partial melting of the crust deduced from magnetotelluric studies in southern Mexico: Implications for the generation of volcanism and subduction dynamics. *J. Geophys. Res.* 111, B08102, doi: 10.1029/2005JB003739.
- Jones, A. G., 1983. On the equivalence of the "Niblett" and "Bostick" transformations in the magnetotelluric method. *J. Geophysics* 53, 72-73.
- Jones, A. G., 1988. Static shift of magnetotelluric data and its removal in a sedimentary basin environment. *Geophysics* 53, 967-978.
- Jones, A. G., 1992. Electrical conductivity of the continental lower crust, in: Fountain, D. M., Arculus, R. J., Kay, R. W. (Eds.), *Continental Lower Crust*, Elsevier, Amsterdam, pp. 81-143.
- Jones, A. G., 1999. Imaging the continental upper mantle using electromagnetic methods. *Lithos* 48, 57-80.
- Jones, A. G., 2006. Electromagnetic interrogation of the anisotropic Earth: Looking into the Earth with polarized spectacles. *Phys. Earth Planet. Int.* 158, 281-291.
- Jones, A. G., 2012. Distortion of magnetotelluric data: its identification and removal, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 219-295.
- Jones, A. G., Chave, A. D., Egbert, G., Auld, D., Bahr, K., 1989. A comparison of techniques for magnetotelluric response function estimation. *J. Geophys. Res.* 94, 14,201-14,213.
- Jones, A. G., Craven, J. A., 1990. The North American Central Plains conductivity anomaly and its correlation with gravity, magnetic, seismic, and heat flow data in Saskatchewan, Canada. *Phys. Earth Planet. Int.* 60, 169 - 194.
- Jones, A. G., Craven, J. A., 2004. Area selection for diamond exploration using deep-probing electromagnetic surveying. *Lithos* 77, 765-782.
- Jones, A. G., Evans, R. L., Eaton, D. W., 2009a. Velocity-conductivity relationships for mantle mineral assemblages in Archean cratonic lithosphere based on a review of laboratory data and Hashin-Shtrikman extremal bounds. *Lithos* 109, 131-143, doi: 10.1016/j.lithos.2008.10.014.
- Jones, A. G., Evans, R. L., Muller, M. R., Hamilton, M. P., Miensopust, M. P., Garcia, X., Cole, P., Ngwisanyi, T., Hutchins, D., Fourie, C. J. S., Jelsma, H., Evans, S., Aravanis, T., Pettit, W., Webb, S., Wasborg, J., The SAMTEX, 2009b. Area

- selection for diamonds using magnetotellurics: Examples from southern Africa. *Lithos* 112, 83-92.
- Jones, A. G., Ferguson, I. J., Chave, A. D., Evans, R. L., McNeice, G. W., 2001. The electric lithosphere of the Slave craton. *Geology* 29, 423-426.
- Jones, A. G., Snyder, D., Hanmer, S., Asudeh, I., White, D., Eaton, D., Clarke, G., 2002. Magnetotelluric and teleseismic study across the Snowbird Tectonic Zone, Canadian Shield: A Neoarchean mantle suture? *Geophys. Res. Lett.* 29, doi: 10.1029/2002GL015359.
- Jones, A.G., Katsube, J., Schwann, P. 1997. The longest conductivity anomaly in the world explained: sulphides in fold hinges causing very high electrical anisotropy. *J. Geomag. Geoelect.* 49, 1619–1629.
- Jones, A.G., Lezaeta, P., Ferguson, I. J., Chave, A. D., Evans, R. L., Garcia, X., Spratt, J., 2003. The electrical structure of the Slave craton. *Lithos* 71, 505-527.
- Jones, A.G., Plomerova, J., Korja, T., Sodoudi, F., Spakman, W., 2010. Europe from the bottom up: A statistical examination of the central and northern European lithosphere-asthenosphere boundary from comparing seismological and electromagnetic observations. *Lithos* 120, 14-29.
- Jones, A.G., Savage, P. J., 1986. North American Central Plains conductivity anomaly goes east. *Geophys. Res. Lett.* 13, 685-688.
- Kamo, S.L., Krogh, T.E., Kumarapeli, P.S., 1995. Age of the Grenville dyke swarm, Ontario–Quebec: implications for the timing of Iapetan rifting. *Can. J. Earth Sci.* 32, 273–280.
- Karato, S., 1990. The role of hydrogen in the electrical conductivity of the upper mantle. *Nature* 347, 272–273.
- Karato, S. I., 2012. On the origin of the asthenosphere, *Earth Planet. Sci. Lett.* 321, 95-103.
- Katsube, T.J., Mareschal, M., 1993. Petrophysical model of deep electrical conductors: graphite lining as a source and its disconnection due to uplift. *J. Geophys. Res.* 98, 8019 - 8083.
- Kellett, R.L., Barnes, A. E., Rive, M., 1994. The deep structure of the Grenville Front: a new perspective from western Quebec. *Can. J. Earth Sci.* 31, 282–292.
- Kellett, R.L., Mareschal, M., Kurtz, R. D., 1992. A model of lower crustal electrical anisotropy for the Pontiac Subprovince of the Canadian Shield. *Geophys. J. Int.* 111, 141-150.
- Kerrick, R., Ludden, J., 2000. The role of fluids during formation and evolution of the southern Superior Province lithosphere: an overview. *Can. J. Earth Sci.* 37, 135-164.
- Korja, T., 2007. How is the European lithosphere imaged by magnetotellurics? *Surv. Geophys.* 28, 239 – 272.
- Korja, T., Engels, M., Zhamaletdinov, A.A., Kovtun, A.A., Palshin, N.A., Smirnov, M.Yu, Tokarev, A., Asming, V.E., Vanyan, L.L., Vardaniants, I.L., the BEAR Working Group, 2002. Crustal conductivity in Fennoscandia – a compilation of a database on crustal conductance in the Fennoscandian Shield. *Earth Planets Space* 54, 535–558.
- Korja, T., Tuisku, P., Pernu, T., Karhu, J., 1996. Lapland Granulite Belt– implications for properties and evolution of deep continental crust. *Terra Nova* 8, 48–58.

- Krogh, T. E., 1994. Precise U-Pb ages for Grenvillian and pre-Grenvillian thrusting of Proterozoic and Archean metamorphic assemblages in the Grenville Front Tectonic Zone, Canada. *Tectonics* 13, 963-982.
- Kumarapeli, P. S., 1985. Vestiges of Iapetan rifting in the craton west of Appalachians. *Geoscience Canada* 12, 54 – 59.
- Kurtz, R. D., 1982. Magnetotelluric interpretation of crustal and mantle structure in the Grenville Province. *Geophys. J. R. Astr. Soc.* 70, 373-397.
- Kurtz, R.D., Craven, J.A., Niblett, E.R., Stevens, R.A., 1993. The conductivity of the crust and mantle beneath the Kapuskasing Uplift: electrical anisotropy in the upper mantle. *Geophys. J. Int.* 113, 483–498.
- Larson, K.M., Freymueller, J. T., Philipson, S., 1997. Global plate velocities from the Global Positioning System. *J. Geophys. Res.* 102, 9961-9981.
- Ledo, J., Jones, A. G., Ferguson, I. J., Wolyne, L., 2004. Lithospheric structure of the Yukon, northern Canadian Cordillera obtained from magnetotelluric data. *J. Geophys. Res.* 109, B04410, doi:10.1029/2003JB002516.
- Ledo, J., Queralt, P., Marti, A., Jones, A. G., 2002. Two-dimensional interpretation of three-dimensional magnetotelluric data: an example of limitations and resolution. *Geophys. J. Int.* 150, 127-139.
- Li, S., Unsworth, M. J., Booker, J. B., Wei, W., Tan, H., Jones, A. G., 2003. Partial melt or aqueous fluid in the mid-crust of Southern Tibet? Constraints from INDEPTH Magnetotelluric data. *Geophysics* 153, 289-304.
- Ludden, J., Hubert, C., 1986. Geologic evolution of the Late Archean Abitibi Greenstone Belt of Canada. *Geology* 14, 707- 711.
- Ludden, J., Hynes, A., 2000. The Lithoprobe Abitibi-Grenville transect: two billion years of crust formation and recycling in the Precambrian Shield of Canada. *Can. J. Earth Sci.* 37, 459-476.
- Mackie, R., Rieven, S., Rodi, W., 1997. Users manual and software documentation for two-dimensional inversion of magnetotelluric data. San Francisco, GSY-USA, 14pp.
- Mareschal, M., Kellett, R. L., Kurtz, R. D., Ludden, J. N., Ji, S., Bailey, R. C., 1995. Archean cratonic roots mantle shear zones and deep electrical anisotropy. *Nature London* 375, 134-137.
- Martignole, J., Calvert, A. J., 1996. Crustal-scale shortening and extension across the Grenville Province of western Quebec. *Tectonics* 15, 376–386.
- McNeice, G. W., Jones, A. G., 2001. Multisite, multifrequency tensor decomposition of magnetotelluric data. *Geophysics* 56, 158-173.
- Meyer, H. O. A., Waldman, M. A., Garwood, B. L., 1994. Mantle xenoliths from kimberlite near Kirkland Lake, Ontario. *Can. Min.* 32, 295-306.
- Mereu, R. F., Wang, D., Kuhm, O., 1986. The 1982 COCRUST seismic experiment across the Ottawa-Bonnechere graben and Grenville Front in Ontario and Quebec. *Geophys. J. R. Astr. Soc.* 84, 491-514.
- Mibe, K., Fuji, T., Yasuda, A., 1998. Connectivity of aqueous fluid in the Earth's upper mantle. *Geophys. Res. Lett.* 25, 1233-1236.
- Miensopust, M. P., Jones, A. G., Muller, M. R., Garcia, X., Evans, R. L., 2011. Lithospheric structures and Precambrian terrane boundaries in northeastern Botswana revealed through magnetotelluric profiling as part of the Southern African Magnetotelluric Experiment. *J. Geophys. Res.* 116, B02401.

- Muller, M. R., Jones, A. G., Evans, R. L., Gru'tter, H. S., Hatton, C., Garcia, X., Hamilton, M. P., Miensopust, M. P., Cole, P., Ngwisanyi, T., Hutchins, D., Fourie, C. J., Jelsma, H. A., Evans, S. F., Aravanis, T., Pettit, W., Webb, S. J., Wasborg, J., The SAMTEX Team, 2009. Lithospheric structure, evolution and diamond prospectivity of the Rehoboth Terrane and the western Kaapvaal Craton, southern Africa: Constraints from broadband magnetotellurics. *Lithos* 112S, 93-105.
- Niblett, E. R., Sayn-Wittgenstein, C., 1960. Variation of electrical conductivity with depth by the magneto-telluric method. *Geophysics* 25, 998-1008.
- Nover, G., Heikamp, S., Meurer, H. J., Freund, D., 1998. In-situ electrical conductivity and permeability of mid-crustal rocks from the KTB drilling: consequences for high conductive layers in the earth crust. *Surv. Geophys.* 19, 73-85.
- O'Dowd, C. R., Eaton, D., Forsyth, D., Asmis, H. W., 2004. Structural fabric of the Central Metasedimentary Belt of southern Ontario, Canada, from deep seismic profiling. *Tectonophysics* 388, 145-159.
- Park, S. K., Mackie, R. J., 1997. Crustal structure at Nanga Parbat, northern Pakistan, from magnetotelluric soundings. *Geophys. Res. Lett.* 24, 2415-2418.
- Parker, R. L., 1980. The inverse problem of electromagnetic induction: existence and construction of solutions based on incomplete data. *J. Geophys. Res.* 85, 4421-4425.
- Parker, R. L., Booker, J. R., 1996. Optimal one-dimensional inversion and bounding of magnetotelluric apparent resistivity and phase measurements. *Phys. Earth Planet. Int.* 98, 269-282.
- Percival, J. A., Bleeker, W., Cook, F. A., Rivers, T., Ross, G., van Staal, C., 2004. PanLITHOPROBE Workshop IV: intra-orogen correlations and comparative orogenic anatomy. *Geoscience Canada* 31, 23-39.
- Rivers, T., 1997. Lithotectonic elements of the Grenville Province: review and tectonic implications, *Precambrian Res.* 86, 117-154.
- Rivers, T., Martignole, J., Gower, C. F., Davidson, A., 1989. New tectonic subdivisions of the Grenville Province, southeast Canadian Shield. *Tectonics* 8, 63-84.
- Rodi, W. L., Mackie, R. L., 2001. Nonlinear conjugate gradients algorithm for 2D magnetotelluric inversion. *Geophysics* 66, 174-187.
- Rodi, W. L., Mackie, R. L., 2012. The inverse problem, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 347-414.
- Rondenay, S., Bostock, M. G., Hearn, T. M., White, D. J., Ellis, R. M., 2000. Lithospheric assembly and modification of SE Canadian Shield: Abitibi-Grenville teleseismic experiment. *J. Geophys. Res.* 105, 13735-13754.
- Roy, B., Mereu, R. F., 2000. Applications of seismic pattern recognition and gravity inversion techniques to obtain enhanced subsurface images of the Earth's crust under the Central Metasedimentary Belt, Grenville Province, Ontario. *Geophys. J. Int.* 143, 735-751.
- Schäfer, A., Hought, L., Brasse, H., Hoffmann, N., EMTESZ Working Group, 2011. The North German Conductivity Anomaly revisited. *Geophys. J. Int.* 187, 85-98.
- Schwarz, G., 1990. Electrical conductivity of the earth's crust and upper mantle. *Surv. Geophys.* 11, 133 - 161.
- Selway, K. 2014. On the causes of electrical conductivity anomalies in tectonically stable lithosphere, *Surv. Geophys.*, 57, 219-257, doi 10.1007/s10712-013-9235-1

- S  n  chal, G., Rondenay, S., Mareschal, M., Guilbert, J., Poupinet, G., 1996. Seismic and electrical anisotropies in the lithosphere across the Grenville Front, Canada. *Geophys. Res. Lett.* 23, 2255-2258.
- Simpson, F., 2001. Resistance to mantle flow inferred from the electromagnetic strike of the Australian upper mantle. *Nature* 412, 632-633.
- Simpson, F., Bahr, K., 2005. *Practical Magnetotellurics*. Cambridge University Press, Cambridge, United Kingdom, 254 pp
- Sleep, N. H., 1990. Monteregean hotspot track: a long-lived mantle plume. *J. Geophys. Res.* 95, 21,983-21,990.
- Snyder, D., Gr  tter, H. S., 2010. Lithoprobe's impact on the Canadian diamond-exploration industry. *Can. J. Earth Sci.* 47, 783-800.
- Soyer, M., Unsworth, M., 2006. Deep electrical structure of the northern Cascadia British Columbia, Canada) subduction zone: Implications for the distribution of fluids. *Geology* 34, 53-56.
- Spratt, J. E., Jones, A. G., Jackson, V. A., Collins, L., Avdeeva, A., 2009. Lithospheric geometry of the Wopmay orogen from a Slave craton to Bear Province magnetotelluric transect. *J. Geophys. Res.* 114, B01101.
- Tikhonov, A. N., 1950. The determination of the electrical properties of the deep layers of the Earth's crust. *Dokl. Acad. Nauk. SSR* 73, 295-297 (in Russian).
- van der Lee, S., Nolet, G., 1997. Upper mantle S velocity structure of North America. *J. Geophys. Res.* 102, 22,815-22,838.
- Vozoff, K., 1991. The magnetotelluric method, in: Nabighian, M. N. Ed.), *Electromagnetic methods in applied geophysics*, Society of Exploration Geophysicist, pp. 641 – 711.
- Weidelt, P., Chave, A. D., 2012. The magnetotelluric response function, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 122-162.
- White, D.J., Forsyth, D. A., Asudeh, I. A., Carr, S. D., Wu, H., Easton, R. M., Mereu, R. F., 2000. A seismic-based cross-section of the Grenville Orogen in Southern Ontario and western Quebec. *Can. J. Earth Sci.* 37, 183-192.
- Wu, X., Ferguson, I. J., Jones, A. G., 2002. Magnetotelluric response and geoelectric structure of the Great Slave Lake Shear Zone. *Earth Planet. Sci. Lett.* 196, 35-50.
- Xu, Y.S., Shankland, T. J., 1999. Electrical conductivity of orthopyroxene and its high pressure phases. *Geophys. Res. Lett.* 26, 2645-2648.
- Xu, Y.S., Shankland, T. J., Duba, A. G., 2000. Pressure effect on electrical conductivity of mantle olivine. *Phys. Earth Planet. Int.* 118, 149-161.
- Yoshino, T., Manthilake, G., Matsuzaki, T., Katsura, T., 2008. Dry mantle transition zone inferred from the conductivity of wadsleyite and ringwoodite. *Nature* 451, 326-329.
- Zhang, J., Frederiksen, A.W., 2013. 3-D crust and mantle structure in Southern Ontario, Canada via receiver function imaging. *Tectonophysics* 608, 700-712
- Zhang, P., Chouteau, M., Mareschal, M., Kurtz, R. D., Hubert, C., 1995. High frequency magnetotelluric investigation of crustal structure in north-central Abitibi, Quebec, Canada. *Geophys. J. Int.* 120, 406-418.
- Zhang, P., Pedersen, L. B., Mareschal, M., Chouteau, M., 1993. Channelling contribution to tipper vectors: a magnetic equivalent to electrical distortion. *Geophys. J. Int.* 113,693-700.

Chapter 6

Imaging the lithosphere of the Grenville Province: large-scale structures and electrical anisotropy

This chapter has been submitted for publication in *Geophysical Journal International* (in January 2014) in co-authorship with a research colleague (A.G. Jones) and my advisor (I. J. Ferguson). I was responsible for all the data processing, analysis, modeling, and isotropic inversions as well as the writing of the first draft. My co-authors reviewed the draft of the manuscript. A. G. Jones provided the STRIKE and the MTMAP codes, and ran the anisotropic inversion for one of the profiles.

Adetunji, A.Q., Ferguson, I.J. and Jones, A.G., 2014. Imaging the lithosphere of the Grenville Province: large-scale structures and electrical anisotropy, *Geophysical Journal International* (Submitted in Jan. 2014)

6.1 Summary

Three northwest-southeast profiles, containing 56 magnetotelluric stations, are used to investigate the lithospheric mantle beneath the Grenville Front and Proterozoic Grenville Province in Ontario, Canada. Dimensionality analysis indicates dominantly 2-D resistivity structure below the Grenville Front and northwestern Grenville Province, with more significant 3-D responses to the southeast. It allows the division of the lithosphere into upper (45-150 km) and deeper (>200 km) mantle layers with regional strike azimuths of N85°E ($\pm 5^\circ$) and N65°E ($\pm 5^\circ$) respectively. Isotropic 2-D inversion of distortion-free MT responses reveals a very resistive (>20,000 $\Omega\cdot\text{m}$) region in the lithospheric mantle beneath the northwest Grenville Province that extends 300 km southeast of the Grenville Front and for at least 600 km along-strike. This feature is interpreted to be Superior Province lithosphere and the corresponding N85°E geoelectric strike to be associated with the fabric of the Superior province. The upper surface of the resistor dips southeast and it is overlain by mostly more conductive rocks, with a geoelectric strike of N45°E, interpreted to be Grenville lithosphere. The base of the resistor reaches depths of 280 km on two of the three MT profiles and this depth is interpreted to be the lithospheric thickness. A large region of enhanced conductivity in the lower lithosphere, spatially correlated with decreased seismic velocity, is bounded to the northwest by a sub-vertical resistivity anomaly located near the Kirkland Lake and Cobalt kimberlite fields. The enhanced conductivity in the lower lithosphere is attributed to refertilization by fluids associated with Cretaceous kimberlite magmatism. It can be explained by a water content in olivine of 50 wt ppm in background areas with higher values in a localized anomaly beneath the kimberlite fields. Based on the interpretation of the enhanced conductivity,

the N65°E strike in the lower lithosphere is interpreted to be of Cretaceous age. Anisotropic 2-D inversion of the MT data shows that the electrical resistivity in the mantle lithosphere beneath the Grenville Front and Grenville province is almost electrically isotropic with a maximum resistivity ratio of less than 10%. Our results support earlier interpretations of the Superior province lithosphere extending beneath the Grenville Front but they significantly revise earlier interpretations, based on 1-D modelling, of the presence of strong electrical anisotropy (anisotropic ratios of 15) in the region. We also suggest the review of earlier interpretation which suggest the existence of dextra shearing in the lithospheric mantle.

Key words: Magnetotelluric; Grenville Province, Anisotropy; Lithospheric root

6.2 Introduction

Electromagnetic (EM) images of the continental lithosphere help elucidate the formation, preservation and deformation of the continents (e.g., Jones, 1999; Davis et al. 2003; Ferguson et al., 2012). The magnetotelluric (MT) method is an EM method that maps spatial variation of electrical conductivity using naturally-occurring time variations of the horizontal components of the electric and magnetic fields that can probe the Earth from a hundred meters to depths of several hundreds of kilometers (Chave and Jones, 2012). MT studies of the Grenville Front, and the adjacent southeastern Superior Province and northwestern Grenville Province, in North America, have provided results that were interpreted in terms of electrical resistivity anisotropy recording past tectonic deformation of the mantle lithosphere (e.g., Kellett et al., 1992; Mareschal et al., 1995; Ji et al., 1996). Obliquity of the resistivity anisotropy and seismic anisotropy in the region led to the interpretation, now internationally renowned (e.g., Wannamaker, 2005), that the resistivity anisotropy in the mantle lithosphere is caused by shape-alignment of the mantle minerals whereas the seismic anisotropy is caused by the crystallographic alignment (Ji et al., 1996; Senechal et al., 1996; Frederiksen et al., 2006). The resistivity anisotropy was modelled (in one-dimension, 1-D) to occur in a resistive zone between 50 and 150 km depth. More recently, Adetunji et al. (2014) show that the MT responses in the Grenville Front area can, in fact, be accurately modelled by two-dimensional (2-D) models without any anisotropy. As with the earlier results, these 2-D models indicate the presence of resistive mantle lithosphere between 50 km and 150 km depth in the Grenville Province.

In the present study we re-examine the evidence for anisotropy in the lithospheric mantle at the Grenville Front, compare isotropic and anisotropic 2-D models of the resistivity structure of the region, define the extent of the resistive mantle lithosphere, and examine the geological cause and tectonic history of the resistive mantle lithosphere.

6.2.1 Electrical anisotropy

Electrical anisotropy, sometimes called inherent anisotropy, is the variation of the electrical conductivity (or its inverse, resistivity) as a function of direction. At the smallest scale, anisotropy is caused by alignment of minerals with different resistivities along the crystallographic axes, where it is called microscopic intracrystalline anisotropy (Wannamaker, 2005; Martí, 2014). At larger scales, the anisotropy is called macroscopic anisotropy and arises from the superposition of individual components (ranging in size from grains to rocks to rock units) with different resistivity. All structures that produce a directionality in the bulk electrical resistivity, but are too small to be resolved by the EM method being used, are usually considered to constitute electrical anisotropy. In order for electrical anisotropy to occur, the more conducting phases must be at least partially-interconnected over a distance of the order of the inductive scale length (Everett, 2005). For the MT method, the resolution of an MT survey is limited predominantly by the diffusive nature of induced electromagnetic propagation, but also by the geometric nature of the galvanic response to charges on lateral conductivity gradients. The inductive scale length is usually on the order of hundreds of meters to kilometers (Vozoff, 1991), so the sources of macroanisotropy may be quite large. In the lithospheric mantle, features of up to tens of kilometers could contribute to macroscopic anisotropy.

Hydrogen diffusion in mantle minerals, such as olivine or minerals with similar crystal alignment has been identified as important cause of electrical anisotropy in the mantle (Schock et al., 1989; Karato, 1990; Mackwell and Kohlstedt, 1990; Ji et al., 1996; Hirth et al., 2000; Wannamaker, 2005; Pommier, 2014). Olivine is mildly electrically anisotropic when dry (Du Frane et al., 2005), but significantly more so when wet (e.g., Poe et al., 2010), and is the most abundant mineral in the upper mantle. Its crystals align preferentially when under strain and thus, many crystals with similar alignment would contribute to the overall anisotropic effect. Jones (1992) suggested that the anisotropy in the mantle could also be caused by preferred interconnection of graphite minerals along a foliation in the uppermost mantle.

The MT tensor defines the surface EM impedance corresponding to electric current flow in different horizontal directions. The MT response may also be directionally-dependent. This directionality is also sometimes loosely called anisotropy, but herein we will refer to this effect as MT response anisotropy to distinguish it from anisotropy related to the material properties. Response anisotropy will occur at MT sites in 2-D or 3-D resistivity structures or over general anisotropic structures. It will be absent only for MT sites over 1-D (horizontally-layered) resistivity structures or over layered anisotropic rocks in which the projection of the anisotropic conductivity ellipsoid onto the horizontal plane gives equal conductivity in the horizontal directions (e.g., Heise et al., 2006).

A number of studies have examined ways of discriminating whether observed anisotropic MT responses are caused by resistivity anisotropy or by 2-D or 3-D resistivity structures. Wannamaker (2005) and Martí (2014) provide reviews of these approaches.

The methods make use of both of the standard MT responses, the impedance phase and the apparent resistivity which is derived from the impedance magnitude, and the tipper response which defines the transfer function between the vertical and horizontal magnetic field components and is often represented graphically in the form of induction arrows (Parkinson, 1962; Weise, 1962). The phase response is a more robust measure of aspects of the dimensionality than apparent resistivity because, even after tensor decomposition to remove the effects of local galvanic distortion on MT responses, the apparent resistivity may still be affected by static shift, a frequency-independent multiplicative factor (Jones, 1988). Thus the phase tensor of Caldwell et al. (2004), recently reviewed by Booker (2014), has merit, although one must be careful when using it given its statistical instability in the presence of noise or significant distortion (Jones, 2012; Chave, 2014).

For layered resistivity structures, the presence of anisotropy can create responses in the phase and apparent resistivity resembling those associated with 2-D isotropic structures (e.g., Heise et al., 2006). The anisotropy creates a variation in apparent resistivity with maxima and minima in directions corresponding to the projection of the anisotropy ellipsoid to the horizontal plane. For anisotropy occurring in a uniformly resistivity medium there is no corresponding azimuthal phase variation. Heise et al. (2006) show that such phase variation depends on the resistivity gradient with depth, so an azimuthal phase variation will occur only when the anisotropic layer is bounded by other layers. Layered resistivity structures containing anisotropy will not generate a tipper response (Martí, 2014).

As discussed in Heise et al. (2006) a number of studies have interpreted relatively widespread consistent azimuthal phase splits as indicative of anisotropy in an underlying layered resistivity structure. This interpretation has usually been supported by observation of a lack of a consistent large-scale tipper responses perpendicular to the structure (Rasmussen, 1988; Kellett et al., 1992; Kurtz et al., 1993; Eisel and Bahr, 1993; Mareschal et al., 1995; Simpson and Tommasi, 2005). However, such interpretations must be made with care; they require that lateral conductivity variations be absent on a scale much larger than the conductivity gradient producing the phase split, and verification that a smaller tipper response is not simply due to 2-D resistivity structures (Heise et al., 2006).

In 2-D structures the effects of anisotropy can be quite complex and depend on the resistivity structures, the distribution and form of anisotropy, the site location, and the period (e.g., Heise and Pous, 2001). There are a number of effects that may be observed, including tipper responses and induction arrows that are misaligned with the 2-D structure, divergence of MT strike directions from the true strike, MT phases lying outside the usual quadrant for 2-D responses, and the appearance of unrealistic structures in 2-D resistivity models fitted to the MT data (Pek and Verner, 1997; Heise and Pous, 2001, 2003; Wannamaker, 2005; Martí, 2014). In general, these effects can also be modelled in terms of 3-D structures, although in some cases a 2-D anisotropic model provides a more geologically reasonable explanation. For example, although out-of-quadrant phases can also be observed in other situations, Heise and Pous (2003) showed that in southwest Iberia, such responses are very plausibly explained by anisotropic graphitic rocks. In some studies, anisotropic 2-D models provide a better fit to the

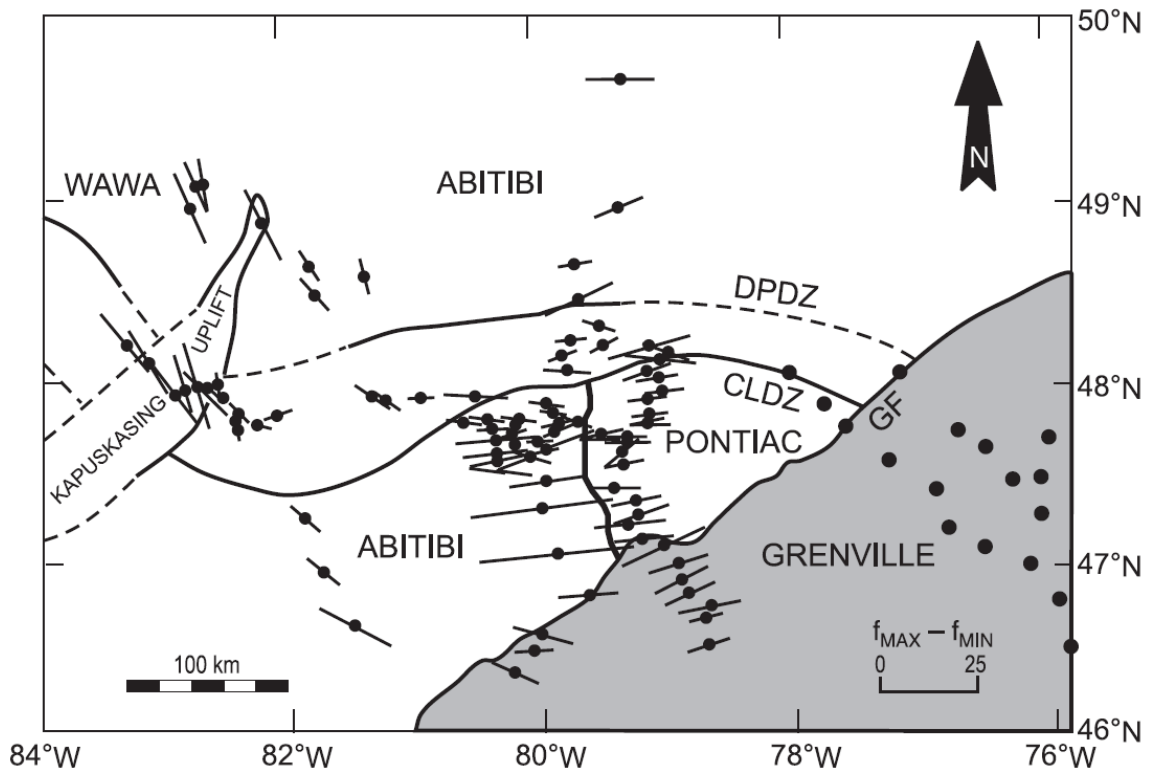
observed data and more geologically plausible models than corresponding isotropic 2-D models (e.g., Baba et al., 2006; Le Pape et al., 2012). In other cases, 3-D models may provide equally plausible explanations of the MT data (Nieuwenhuis et al., 2013). Discriminating the presence of anisotropy in 3-D resistivity structures is even more complex (Wannamaker, 2005; Martí, 2014). Weidelt (1999) showed that the response from any 3-D model containing anisotropy can equivalently be explained by a 3-D isotropic model of arbitrary complexity (Wannamaker, 2005). Finally, 2-D anisotropic modeling of 3-D structures can be more advantageous and yield useful results (e.g., Schmoldt and Jones, 2013).

6.2.2 Previous MT studies of the Grenville Front

The Archean-Proterozoic boundary between the Superior craton and the Grenville Province, the Grenville Front, is regarded as a classic example of lithospheric electrical anisotropy (e.g., Wannamaker, 2005). The earliest study identifying anisotropy in the southern Superior craton and Grenville Front area was in the area of the Kapuskasing Uplift Structure (Fig. 6.1) by Kurtz et al. (1988) who interpreted anisotropy oriented at N65°E at depths exceeding 70 km (Boerner et al., 2000). Subsequently, Kellett et al. (1992) interpreted MT responses in the Pontiac subprovince as being indicative of electrical anisotropy. Their conclusion was based on observation of spatially consistent MT response across the Pontiac subprovince (Fig. 6.1), a lack of consistent induction vector responses in the subprovince, and, based on MT data extending 30 km south of the Grenville Front, an apparent absence of major resistivity structures adjacent to the subprovince. The responses were modelled by an anisotropic layer in the lower crust

and/or upper mantle extending anywhere from 20 km to 100 km depth. Within this layer, the ratio of the horizontal resistivities was in the range of 1:6 to 1:13. Kellett et al. (1994) interpreted similar MT and induction vector responses extending across the Grenville Front in terms of electrical anisotropy, and correlated the anisotropic layer with the southward extension of the Archean rocks of the Superior Province at depth into the Grenville province.

Figure 6.1. MT response anisotropy in the southern subprovinces of the Superior Province (Wawa, Abitibi, and Pontiac subprovinces) and the northern Grenville Province (Boerner et al., 2000). DPDZ=Destor-Porcupine deformation zone; CLDZ=Cadillac-Larder deformation zone; GF=Grenville Front. The length of the line at each MT site is proportional to the phase difference between the off-principal MT impedances and the direction is the direction of maximum phase. Sites with erratic apparent resistivity and phase responses as a function of azimuth are shown as dots.



Mareschal et al. (1995) provided a synthesis and tectonic interpretation of resistivity anisotropy in the southern Superior and Grenville Front region. They constrained the electrical anisotropy to lie between 50 and 150 km depth with a maximum ratio of horizontal resistivities within the anisotropic layer of 1:15 and a conductive direction azimuth of N80°E (Fig. 6.1). Given extant knowledge at the time, the observed degree of electrical anisotropy could not be explained by lattice preferred orientation of dry mantle minerals and was therefore interpreted as being due to an interconnected conducting phase such as graphite. The authors suggested that the anisotropy may be due to the concentration of graphite in mantle shear zones correlating with observed crustal sutures. The MT responses could be explained by anisotropic resistivity in 50 km wide mantle shear zones, separated by ~100 km.

Shear-wave splitting studies by Senechal et al. (1996) along a profile coincident with the MT profile across the Grenville Front revealed spatially consistent seismic anisotropy with a fast direction of N103°E and an average time delays of 1.46 ± 0.21 s. Ji et al. (1996) explained the obliquity between the interpreted seismic and electrical responses as arising because of the seismic response being controlled by lattice preferred anisotropy (LPO) and the MT response by shape preferred anisotropy (SPO). Taken together the results implied dextral shearing in a mantle transcurrent shear zone which is consistent with the inferred motion from crustal shear zones. In a more recent study, Frederiksen et al. (2006) examined the directional dependence of seismic and MT responses at a larger scale across the Grenville Front. Although both seismic and MT responses show significant directional dependence over a large area of the southern Superior and Grenville provinces, the patterns observed at the Grenville Front, e.g., the

obliquity between the MT conductive direction and seismic fast direction, are restricted to the area around the Grenville Front itself. The authors conclude that their analyses were unable to provide a definitive answer as to whether the MT directionality at the Grenville Front could be attributed to electrical anisotropy alone.

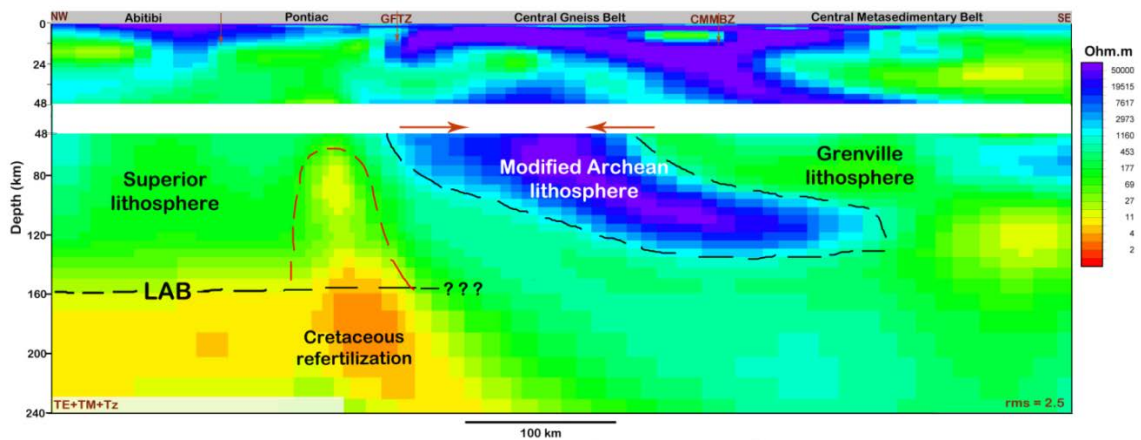
The earlier MT studies of the Grenville Front region were based on 1-D inversion methods and suggested the crust and lithospheric mantle near the Grenville Front possesses a simple layered structure. Kellett et al. (1994) resolved a layered structure extending from the Pontiac subprovince into the Grenville Province. It includes a resistive ($>3000 \Omega.m$) upper crust (<10 km), a more conductive ($<300 \Omega.m$) middle and lower crust, and the thick anisotropic resistive layer in the lowermost crust and uppermost mantle lithosphere.

Adetunji et al. (2014) using new MT data from the Grenville Province, along with the earlier MT data set, conducted extensive 2-D isotropic modelling of the data. Their results indicated a more complex resistivity structure in the crust which included a layer of resistive rocks dipping southwards from near the Grenville Front (Fig. 6.2). Their results also show that the upper 150 km of the mantle lithosphere, south of Grenville Front, is characterized by an extremely resistive, southeast dipping slab. They interpret this slab to be Superior Province lithosphere. Their resistivity model includes a large-scale sub-vertical conductor, beneath the Pontiac subprovince that they interpret to be due to re-fertilization of an older mantle scar.

Studies of the Grenville Front have also been conducted outside the present study area. Mareschal et al. (1991) analyzed data from an MT profile crossing the Grenville Front on Manitoulin Island and the Bruce Peninsular, approximately 300 km to the

southwest of the present study. At this location, resolution of the underlying lithospheric resistivity structure is severely limited by the conductive Phanerozoic rocks at the surface, and the MT data are only able to place a broad lower limit on the underlying resistivity of 500 $\Omega\cdot\text{m}$. Ogawa et al. (1996) and Wannamaker et al. (1996) present 2-D models from the southeastern United States that show the signature of the Grenville orogen. In contrast with the results from Canada, the Grenville suture is imaged as a conductive feature in the uppermost mantle.

Figure 6.2. Crustal and lithospheric mantle resistivity model for northeast-southwest profile crossing the Grenville Front (V.E = 1) (Adetunji et al., 2014). The crustal model is based on a geoelectric strike of $\text{N}45^\circ\text{E}$ and the lithospheric model is based on a strike of $\text{N}85^\circ\text{E}$. The sites used in the study are from the NNW-SSE profile centred on 79°W in Fig. 1. Labels show the interpretations of Adetunji et al. (2014) of lithospheric mantle resistivity structures. GFTZ=Grenville Front Tectonic Zone, CMMBZ= Central Metasedimentary Belt Boundary Zone. LAB=Lithospheric-Asthenospheric Boundary.



6.2.3 Objectives of the present study

In the present work we use a larger MT data set than in earlier studies, along with modern tensor decomposition and isotropic and anisotropic 2-D inversions, to investigate the resistivity structure of the Grenville Front and Grenville Province. We also seek to determine whether the azimuthal variation in the MT responses, defined by the earlier studies, is related to electrical anisotropy or to larger-scale geological structure.

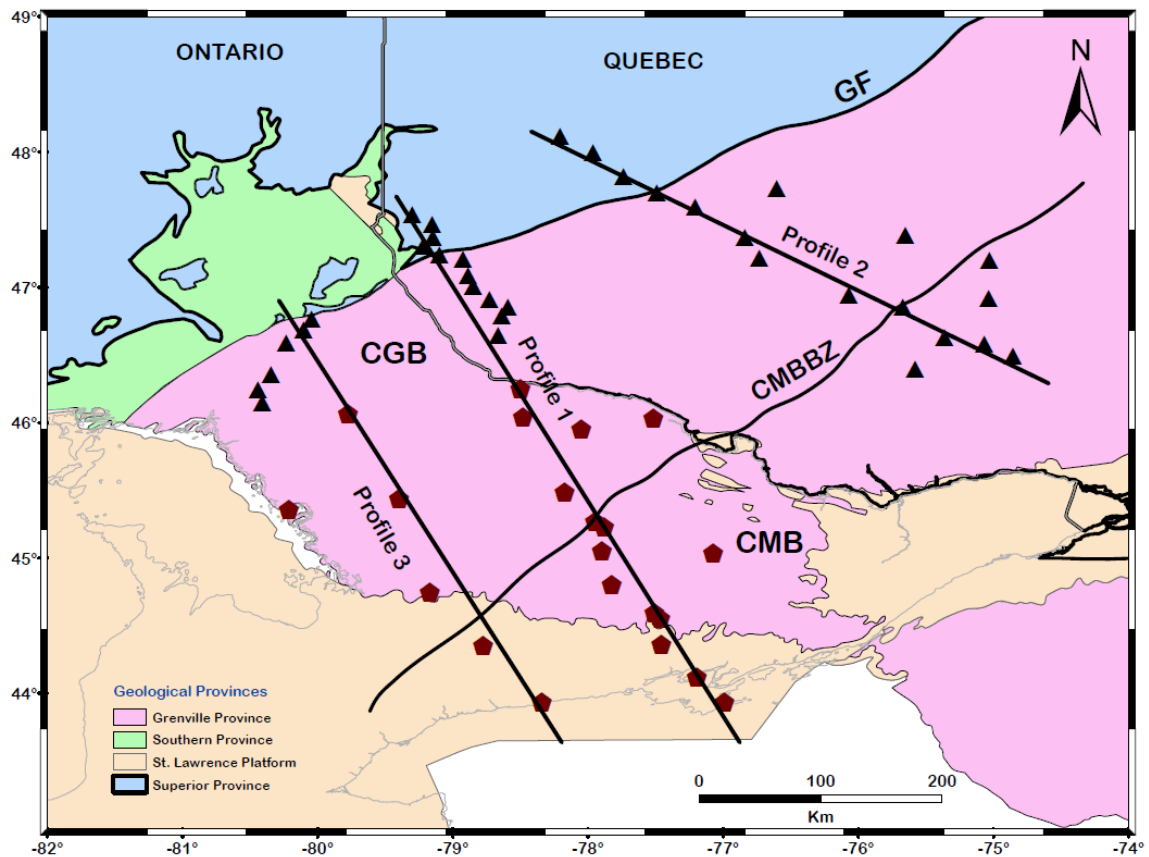
Broadband MT responses from a total of 56 sites are analyzed (Fig. 6.3); these include 35 sites from the Lithoprobe Abitibi-Grenville Transect (Boerner et al., 2000) and 21 sites from the POLARIS project (Eaton et al., 2005). The MT sites form three profiles with northwest-southeast azimuths, and span a north-south distance of 500 km and an east-west distance of 650 km. The data at sites along profiles 1 and 3 have both MT and tipper responses available but those on Profile 2 have only MT responses. The MT data from the central profile (Profile 1) have been examined previously by Kellett et al. (1992), Kellett et al. (1994), Zhang et al. (1995), Mareschal et al. (1995), Ji et al. (1996), Senechal et al. (1996) and Adetunji et al. (2014). Mareschal et al. (1995), Boerner et al. (2000), and Frederiksen et al. (2006) examine the MT response anisotropy for sites in all three profiles. The present study presents the first modelling of responses from Profile 2 which lies to the east of Profile 1, and for Profile 3 which lies to the west.

Specific objectives of the study are as follows:

1. Determination of the geoelectric dimensionality and strike as a function of depth over a broad region of the Grenville Front and northern Grenville Province: Use of an expanded MT data set and modern depth-based dimensionality and strike

determination will provide improved understanding of the resistivity structures present in the lithospheric mantle in the area. The expanded data set will also allow re-assessment of the tipper responses that formed a major justification for the interpretation of electrical anisotropy.

Figure 6.3. Map of the study area showing the locations of MT sites considered in the present study. The black triangles indicate MT sites from the Lithoprobe-Abitibi Grenville transect and the wine-coloured pentagons indicate POLARIS MT sites. Profile 1 forms part of the data analyzed by Adetunji et al. (2014). The tectonic elements include the Grenville Front (GF), Central Gneiss belt (CGB), Central Metasedimentary Belt (CMB), Central Metasedimentary Belt Boundary Zone (CMBBZ) and Frontenac-Adirondack Belt (FAB).



2. Determination of the optimal regional geoelectric strike for 2-D modelling of the MT data from each profile and isotropic and anisotropic 2-D inversion of the data.
3. Definition of the spatial extent of the resistive lithospheric slab identified in the earlier MT studies (Mareschal et al., 1995, Adetunji et al., 2014): The inclusion of profiles 2 and 3 will help define the along-strike extent of this feature.
4. Synthesis of the MT results with other geophysical and geological constraints to provide an updated interpretation of the geological cause and tectonic history of the resistive mantle lithosphere in the northern Grenville Province: Adetunji et al. (2014) note several unresolved questions regarding this feature. Its geometry, with a southeast dip of its upper surface and a well-defined base suggest a history involving subduction. However, the very high resistivity and high seismic velocity within the slab are characteristic of buoyant depleted Archean lithosphere which would not be expected to be involved in subduction.

6.3 Tectonic Summary

The Grenville Province was formed between 1090 and 980 Ma as a result of the Himalayan-scale Grenville orogeny (Rivers et al., 2012). This orogeny produced a northwest-trending (present-day coordinates) thrusting of a composite belt of continental crust and magmatic rocks onto the margin of the Laurentian craton (Carr et al., 2000), creating a collage of tectonically stacked domains separated by ductile shear zones (Davidson et al., 1982; Culshaw et al., 1983; Rivers et al., 1989). The accepted orogenic model incorporates pre-Grenvillian continental-margin arc magmatism and accretionary tectonics followed by repeated episodes of continental collision that resulted in crustal thickening, followed by orogenic collapse and emplacement of mantle- and lower-

crustal-derived magmas (Rivers, 1997; Rivers et al., 2012). The final stage of the convergence at 1000 Ma affected the Grenville Front Tectonic Zone (GFTZ) along the length of the orogen (Haggart et al., 1993).

The tectonic subdivision of the study region includes the Laurentian margin, which is the lowest unit in the orogenic stack. This is overlain to the southeast by the Central Gneiss Belt (CGB), which is characterized by upper amphibolite- to granulite-facies metamorphic assemblages. The CGB is in turn overlain by the Central Metasedimentary Belt (CMB), which consists of several arc and/or back-arc terranes that are characterized by greenschist- to amphibolite-facies assemblages. The CMB is separated from the CGB by the Central Metasedimentary Boundary Belt Zone (CMBBZ). Shear indicators in this zone indicate thrusting of the CMB northwestwardly against the CGB (Fig. 6.3; Rivers et al., 1989; Easton, 1992; Rivers, 1997; Davidson, 1998; Carr et al., 2000; Rivers et al., 2012).

In the Laurentian margin, pervasive ductile fabrics and generally shallow-dipping structures have been interpreted to have formed during late orogenic extension (Culshaw et al., 1997). Within the CMB, the moderately to steeply dipping brittle-ductile normal faults contrast in age and style with those observed in the northwest (Carr et al., 2000; Rivers et al., 2012). The postulated Cretaceous Great Meteor hotspot trail is interpreted to have passed through the Grenville Province's lithosphere (Crough, 1981; Sleep, 1990; Adetunji et al., 2014). This mantle plume has been associated with the Mesozoic kimberlite magmatism at Cobalt and Kirkland Lake fields in southern Ontario (e.g., Meyer et al., 1994; Ji et al., 1996; Griffin et al., 2004; Faure et al., 2011; Adetunji et al., 2014).

6.4 Depth-Transformed MT Response Anisotropy

In this section, we characterize variation of the dimensionality and anisotropy of the MT response with location and with period, or equivalently depth. Spatial variation in the electrical resistivity affects the depth of penetration of MT responses so a given period may correspond to different depths at different sites (e.g., Hamilton et al., 2006; Miensopust et al., 2011; Adetunji et al., 2014). The Niblett-Bostick approximate transformation from period to depth (Niblett and Sayn-Wittgenstein 1960; Bostick 1977; Jones, 1983a) provides a superior method for comparing related responses than period-based data analyses (Jones, 2006; Jones et al., 2013).

The dimensionality and strike determination methods applied are phase tensor analysis (Caldwell et al., 2004) and Groom Bailey (GB) tensor decomposition (Groom and Bailey, 1989, 1991; McNeice and Jones, 2001). Both methods provide a measure of the dimensionality and directionality of the resistivity structures responsible for the MT response. They also both provide a measure of the response anisotropy in terms of the estimated phase of the principle impedance tensor responses. A small phase difference indicates responses close to those of a 1-D structure whereas a large phase difference implies significant 2-D or 3-D regional structures (Hamilton et al., 2006; Miensopust et al., 2011) or anisotropy within the layers of a 1-D structure (Heise et al., 2006).

We characterize the MT response anisotropy in terms of the most conductive direction. For the phase-based phase-tensor and GB decomposition methods this is the direction of largest impedance phase. For sites adjacent to a linear 2-D conductor the conductive direction will correspond to geoelectric strike. However, at a 2-D resistivity contrast between resistive and conductive blocks, the conductive direction will be parallel

to strike in the resistive block but perpendicular to strike in the conductive block. An observed 90° flip in the conductive direction can therefore be used to identify the location of a resistivity boundary (Hamilton et al., 2006; Miensopust et al., 2011). For anisotropic 1-D structures the azimuthal dependence of the phase response is caused by the resistivity contrast at the top and base of the anisotropic layer and the most conductive direction may correspond to either the highest or lowest component of the horizontal projection of the conductivity tensor (Heise et al., 2006).

6.4.1 Phase tensor results

The phase tensor method provides a measure of the dimensionality of the MT response, and for 2-D structures the geoelectric strike that is free from the effects on the electric field of local galvanic distortion (but not those on the magnetic field, e.g., Chave and Jones, 1997). A positive attribute of the method is that it does not require any assumption regarding the dimensionality of the regional resistivity structure, and the results remain valid when both the galvanic distorting local heterogeneity and the regional structures are 3-D (Caldwell et al., 2004). A negative attribute is that it operates on individual frequency responses and the responses are not independent of noise in the MT data. Also, the phase tensor, being a re-representation of the impedance tensor elements and not a model fit, has poor statistics associated with it (Jones, 2012; Chave, 2014). Phase tensors are usually plotted using an elliptical representation with major and minor axes of the ellipse scaled in proportion to the phase response in the corresponding direction. For 2-D responses these directions correspond to azimuths parallel and

perpendicular to the geoelectric strike and for 1-D anisotropic responses they correspond to the horizontal projection of the axes of anisotropy.

The phase tensor skew provides a measure of the three-dimensionality of the response and the phase tensor ellipses are often colored to show this parameter. Booker (2014) suggested that the non-zero phase tensor skew β is the appropriate measure for indicating depth variation of the structural azimuth or anisotropy. Although a large value of the skew angle implies a 3-D regional conductivity structure, a small value of the skew angle is not necessarily a good indication of the closeness of the conductivity structure to 2-D (Caldwell et al., 2004). Constancy of the direction of the principal axes of the phase tensor, with period and with location, provides a more reliable indicator of two-dimensionality (Caldwell et al., 2004).

In the current study the phase tensors are plotted using MTMAP that plots MT parameters using Generic Mapping Tools (Wessel and Smith, 1991). Here we use the Niblett-Bostick approximate depth transformation of the Berdichevsky (arithmetic) average of the off-diagonal elements of the impedance tensor to convert from period to depth. Results were obtained at 50 km depth increments by closest response to the specified depth.

Figure 6.4 shows the phase tensor maps, at different Niblett-Bostick depths, for sites in the study area. Robustness of the phase tensor estimates can be assessed from the similarity of the phase tensors between adjacent stations within the same inductive scale length of each other (given by the depth of investigation). For example, the very high skew ($>40^\circ$, red) and the NW-SE major axis orientation at 50 km at the site on Profile 1 in the SW part of the CMB is in stark contrast to the low skew ($2^\circ - 5^\circ$, light blue) and NE-SW orientation of the site within 10 km of it, whereas the latter site is consistent in skew and

orientation with other nearby sites. Similarly the site in the middle of the CGB on Profile 1 with high skew is contradicted by the nearby sites with low skews. Clearly the high skew sites are erroneous and the phase tensor estimates are driven by poor quality MT responses.

The number of sites with data that penetrate deeper than 150 km is limited, especially on Profile 2 and within the CMB. At all depths, the phase tensor skew β is relatively small ($< 5^\circ$) for sites extending over a northwest-southeast distance of ~ 250 km from northwest of the Grenville Front to the middle of the CGB. This result is consistent with the presence of 1-D or 2-D structures (Caldwell et al., 2004) in this region. Within this zone, the axes of the phase tensor are mostly aligned roughly east-west or north-south on profiles 1 and 2, and WNW to ESE on Profile 3. The consistency of the phase tensor alignment provides further evidence of locally 1-D anisotropic or 2-D structures but with a change of strike azimuth between Profile 2 and Profile 3.

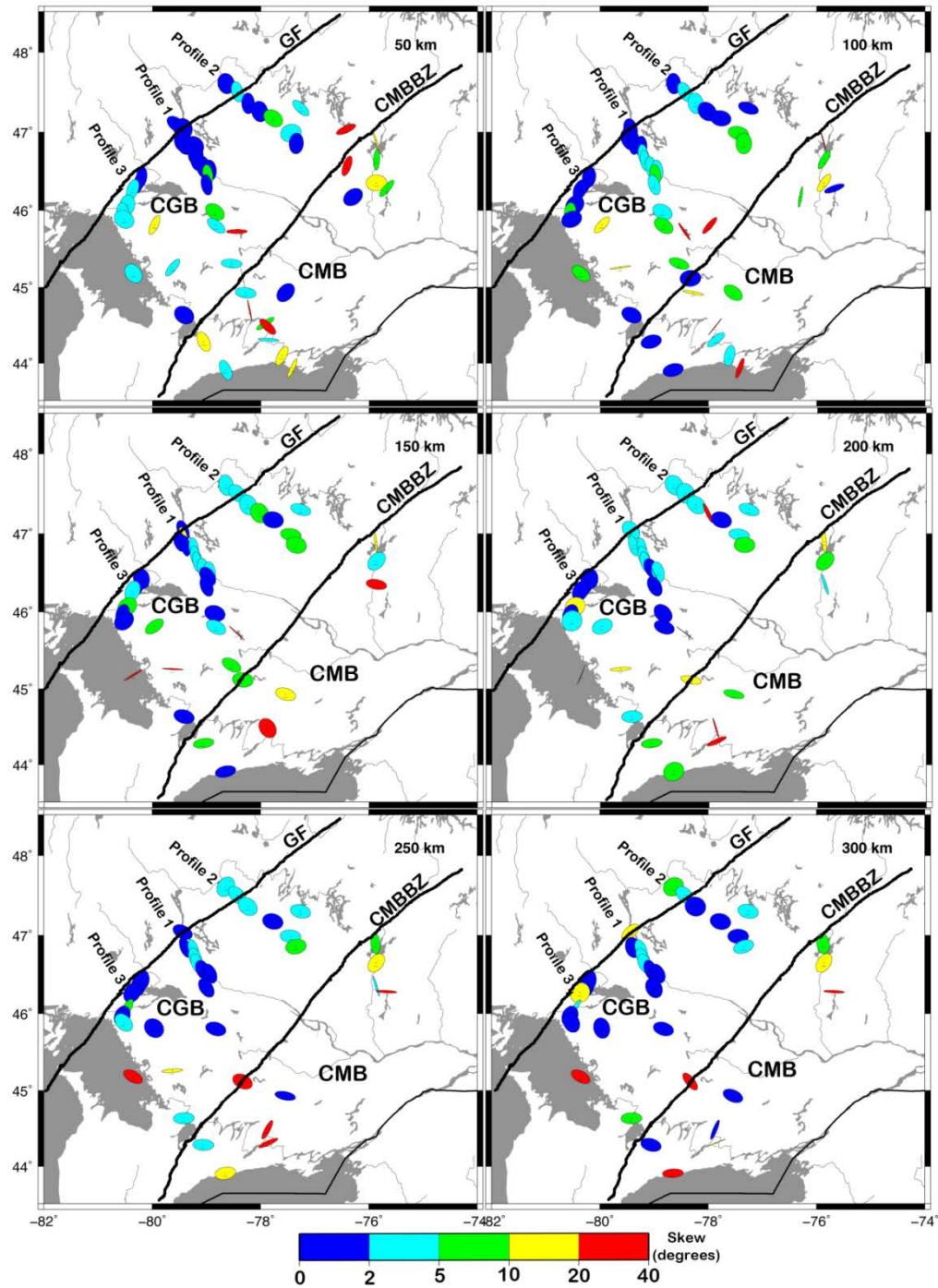
In the southeastern part of the CGB and in the CMB, phase-tensor skews are higher and very strong phase differences are observed. Whether these inconsistent skews are due to higher 3-D heterogeneity or due to less stable, precise and accurate impedance tensors must be examined. For Profile 1, at depths of less than 150 km, the directions of the principle axes of the phase tensors in this zone are somewhat aligned with azimuths between east-west and WNW-ESE. On Profile 2, the alignment of the phase tensors at sites near the CMBBZ appears to be orthogonal to this structure, although a similar response is not observed on the other profiles. In contrast, for Profile 3 at a depth of 50 km, phase tensor ellipses are aligned in a relatively consistent northwest-southeast direction. At depths exceeding 100 km this alignment becomes more east-west and is moderately consistent between sites.

6.4.2 Groom-Bailey tensor decomposition results

Determination of dimensionality and the azimuth of regional response anisotropy can also be done using Groom Bailey (GB) tensor decomposition (Groom and Bailey, 1989, 1991). The procedure employs a least-squares inversion method to fit a model of regional 2-D impedance at a single site and single frequency to observed MT data in the presence of galvanic distortion. The GB method was extended by McNeice and Jones (2001) to find models of 3-D distortion of regional 2-D impedances over multiple sites and multiple frequencies simultaneously. The algorithm applied fits a regional response to a group of frequencies, or their equivalent depths, and sites by minimizing the total squared misfit between observed and predicted impedances (e.g., Hamilton et al., 2006; Muller et al., 2009; Miensoopust et al., 2011). Positive attributes of the GB method are that it uses all of the information in the impedance tensor and that it accommodates errors in the MT data in an appropriate statistical manner. A negative attribute is the required assumption of a regional 2-D resistivity model, although the validity of that assumption is tested statistically. As noted above, even after tensor decomposition is done to remove the effects of galvanic distortion on MT responses, the apparent resistivity may still be affected by “site gain”, a static shift frequency-independent multiplicative factor operating on both apparent resistivity curves (Jones, 1988, 2012).

The quality of the fit of the GB model to an observed data set provides a measure of the three-dimensionality of the regional structure. A large normalized root mean square (rms) misfit provides an indication that the data are unable to be explained by a 1-D or 2-D regional resistivity structure, or that the errors have been improperly estimated and are too small. Chave and Jones (1997) demonstrated that parametric error estimators are

Figure 6.4. Phase tensor response at different Niblett Bostick depths. The phase tensors are plotted with the major axis having a uniform length on the plot and the minor axis proportional to the phase of the lower phase component of the impedance. Circular symbols correspond to 1-D responses and more elliptical symbols to 2-D or 3-D responses. The phase tensor symbols are coloured by the phase skew angle: blue colours correspond to 1-D and 2-D responses and yellow and red symbols to strongly 3-D responses. MT responses with large uncertainties are excluded.



inappropriate and lead to far too small error estimates compared to jackknife error estimates. The GB method provides the strike of the regional MT response. For 2-D responses this direction will correspond to an azimuth parallel and perpendicular to the geoelectric strike and for 1-D anisotropic responses it will correspond to the horizontal projection of the axes of anisotropy. For convenience, the results are discussed here in terms of “strike” which could refer to either the azimuth of macroscopic structure or the horizontal projection of axes of anisotropy.

In the present study the GB decompositions were performed using the STRIKE program (McNeice and Jones, 2001). Decomposition results were obtained for individual sites using least-squares fits over six ~50 km thick depth bands between 45 and 350 km. Since this study is focused on the mantle lithosphere, a depth of 45 km was chosen for the top of the shallowest depth band based on the seismically defined Moho depth for this region (White et al., 2000; Eaton et al., 2006). This depth band extended from 45 to 100 km. Each of the other five bands was 50 km thick. The GB misfits were calculated using the default impedance error floor of 3.5%, which corresponds to a 7% error for the apparent resistivity and 2° for phase. This floor was calculated based on the largest impedance value, and applied in an absolute manner to the other three. For these values, normalized rms misfits exceeding 2.0 are often regarded as indicative of significantly 3-D structures (Jones, 2012).

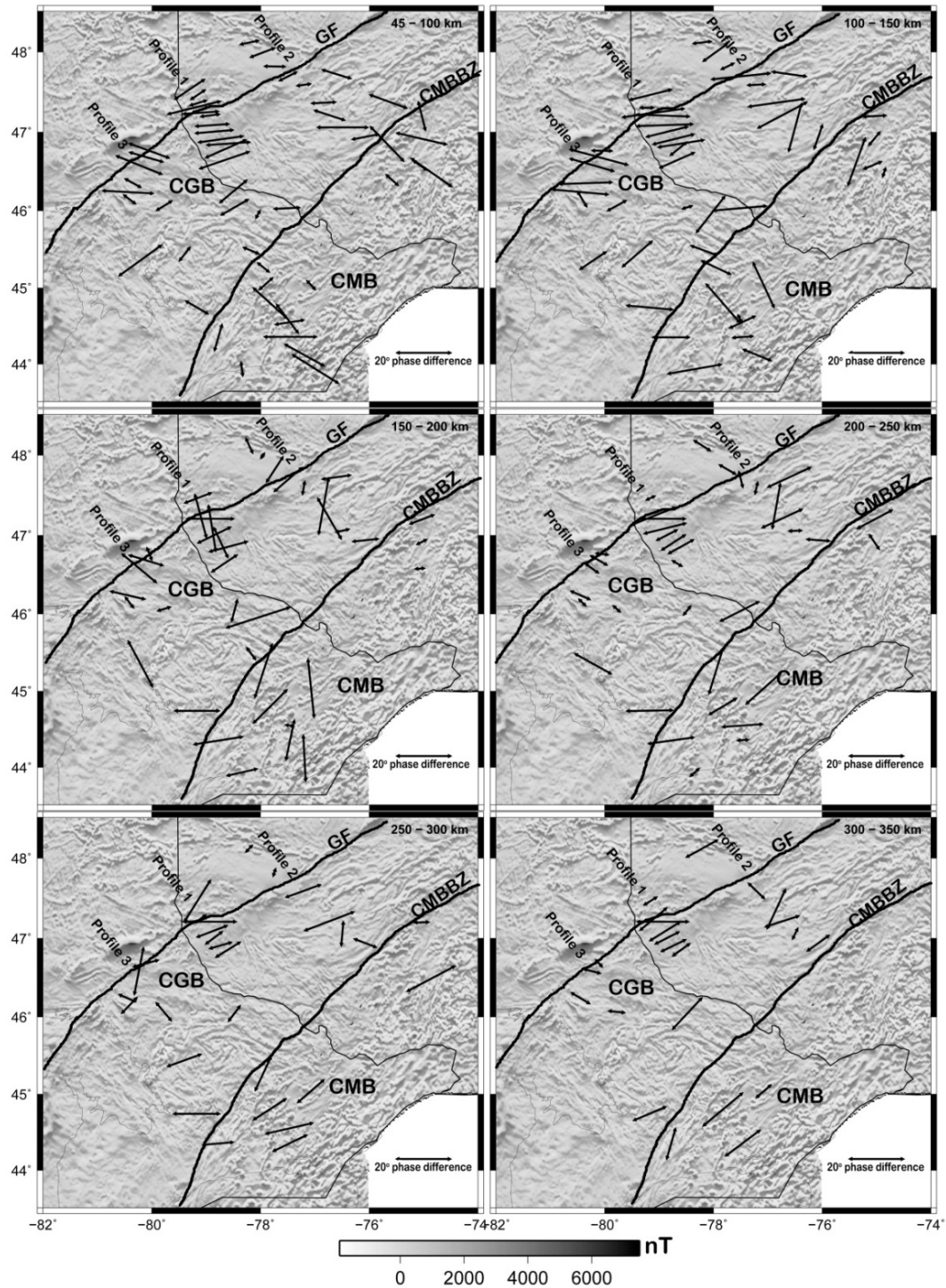
Figure 6.5 shows the results of the GB tensor decompositions based on single site but multiple frequencies (converted into approximate depths). As noted in the phase tensor results, the data at almost all of the sites have Niblett-Bostick depth penetration to a depth of 150 km, beyond which there is penetration at fewer sites. For most of the

sites, the root mean square (RMS) misfit of the GB model for the 50 km bands were less than <1.0 . However, at isolated sites, particularly in the CMB where the phase tensor results indicated the presence of 3-D structures or of noisy data, values of up to 4.0 occur. The GB misfit is examined in more detail in syntheses of the results for larger depth ranges in a later section.

In regions around the Grenville Front and the northwestern part of the CGB the conductive direction at depths of less than 150 km on profiles 1 and 2 has a dominantly N85°E azimuth. In contrast to the results on profiles 1 and 2, the conductive direction in the same region along Profile 3 has an overall E20°S trend. Within this depth range the phase is typically in the range of 5° to 20°. The conductive direction azimuths, at depths of less than 150 km depth in the southeastern CGB and CMB, are more erratic than further to the northwest but display more site-to-site consistency than suggested by the phase tensor results (Fig. 6.4). The greater consistency of the GB results is explained by the greater degree of averaging in the GB analysis and the superior accommodation of noise. Although there is some dispersion, the majority of conductive directions in the CMB are parallel or perpendicular to a southwest-northeast azimuth with phase splits between 10° and 20°.

The spatial consistency of the conductive directions decreases in the 150 to 200 km depth band. On Profile 1 the response near the Grenville Front and in the northwestern CGB is characterized by a number of 90° flips in azimuth. This effect may be explained in part by the penetration of the Niblett-Bostick depth estimates through the base of the layer creating the 2-D response. However, there is also a transition from azimuths aligned in N80°E direction to those with a ~N65°E orientation. Within this

Figure 6.5. Results of GB strike determination for ~50 km thick depth ranges extending from 45 to 350 km. The arrows show the azimuth of the GB regional impedance with the highest phase. The length of the arrow is proportional to the average phase difference, between the maximum and minimum phase values over the depth band. The results are superimposed on the regional Bouguer gravity anomaly map (data from Geological Survey of Canada).



depth band on all three profiles there are flips in the azimuth at locations 100 to 150 km southeast of the Grenville Front suggesting the presence of the southeastern margin of a large resistivity block.

At larger signal penetration (>200 km) there is a reasonable level of coherence in the conductive directions across the study area. The majority of sites show a conductive direction with a southwest to northeast azimuth and phase splits between 10° and 20° . This consistency is again far stronger than that observed in the more disparate phase tensor results (Fig. 6.4), and attests to the superiority of the GB approach.

6.4.3 Comparison of MT response anisotropy with other geophysical responses

In order to facilitate comparison of MT response anisotropy with other geophysical results, we subdivided the full depth range into two broader divisions: the upper lithospheric mantle, represented by a 45 - 150 km depth band, and a deeper zone, represented by depths greater than 200 km, i.e., the upper lithospheric mantle in the first band and the deepest lithosphere and potentially upper asthenosphere in the second band. The division between these zones was based on the observed changes in the phase tensor and GB conductive directions at about 150 km depth, the observation of different strike directions for Profile 1 at depths above and below 200 km by Adetunji et al. (2014), and on the observation of the base of the resistive lithosphere on Profile 1 at a depth of 150 km (Fig. 6.2). Here, we refer to the lower layer as the "deeper mantle layer" as the relative contributions to the depth range from the lower lithosphere and asthenosphere are not yet fully established. The GB analysis was redone for individual sites using these

broader depth bands, and is shown in Fig. 6. For most of the sites the rms misfit of the GB model was similar to the results obtained using 50 km bands i.e., less than <1.0 .

Comparison of the upper lithospheric mantle results (Fig. 6.6) with those for the individual 50 km thick bands (Fig. 6.5) shows that the most conductive direction within the upper lithospheric mantle depth is controlled mainly by the response from 45 and 100 km depth where the greatest phase differences lie. Over most of the study area, the observed azimuths are in close agreement for the 45-100 km and the 100-150 km bands. However, on Profile 2 in the southeastern CGB and CMB there are some 90° flips between the two depth ranges, and the results for the whole upper lithospheric mantle more closely resemble the azimuths in the 45-100 km depth range. The azimuths for the deeper mantle layer are more spatially consistent than those for individual depth ranges. In some parts of the study area, the azimuths for the deeper mantle layer are closer to those in the 200 to 250 km deep range, e.g. in the southeastern CGB on Profile 3, whereas in other areas they are more similar to deeper results, e.g., in the southeastern CGB on Profile 2.

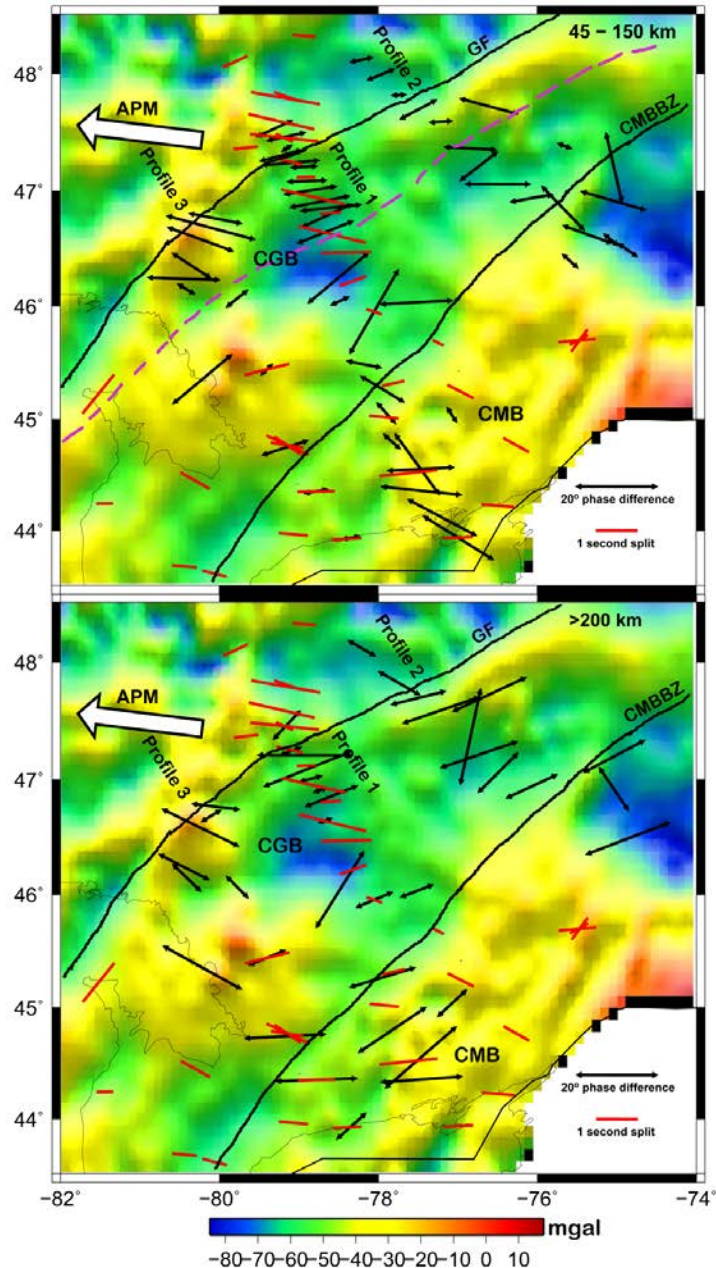
Primary observations that can be drawn from the tensor decompositions for the larger depth ranges (Fig. 6.6) are:

1. Spatially-consistent conductive directions are observed in the upper lithospheric mantle from the southern Superior craton to locations in the CGB that lie within approximately 100 km southeast of the Grenville Front. The southeast limit of this region is marked by the dashed line in Fig. 6.6. On profiles 1 and 2 azimuths in this region are $\sim N85^\circ E$ ($E15^\circ N$) whereas on Profile 3 they are $N110^\circ E$ ($E20^\circ S$).

2. In the southwest CGB and CMB, the conductive directions in the upper lithospheric mantle are more erratic but the majority of sites have conductive directions parallel or perpendicular to a $N45^{\circ}E$ azimuth.
3. Across the whole area, azimuths at deeper lithospheric mantle depths are mostly parallel or perpendicular to a $\sim N45^{\circ}E$ direction. For profiles 1 and 2 the conductive direction is $N45^{\circ}E$ but for sites in the CGB on Profile 3 it is $E45^{\circ}S$.

The conductive direction results can be compared with those from previous MT studies. The upper lithospheric mantle results for Profile 1 are in agreement with the earlier studies of Mareschal et al. (1995) and others (Fig. 6.1) who noted a conductive direction azimuth of $N80^{\circ}E$ and a phase-shift of 20° at sites within about 50 km of the Grenville Front. In contrast to the earlier studies, our results show that a similar response extends several hundred kilometres east to sites on Profile 2. Boerner et al. (2000) concluded that the MT sites on Profile 2 in the northern Grenville Province exhibited erratic azimuths and magnitudes and had a different character from responses along Profile 1. It is probable that this observation was influenced by galvanic distortions of the responses on Profile 2 that were more effectively removed by the GB tensor decomposition in the present study. The new results differ slightly from those of Adetunji et al. (2014), who concluded that in the upper lithospheric mantle the east-west strikes on Profile 1 extended all the way through the southern CGB and into the CMB, rather than only as far as south as the middle of the CGB. The difference reflects the use of a broader depth range of 40 to 200 km in the earlier study that may have led to greater mixing of crustal and deeper mantle layer responses with the upper lithospheric mantle estimate.

Figure 6.6. Comparison of MT response anisotropy with seismic anisotropy. The black arrows show the azimuth of the GB regional impedance with the highest phase determined for depth ranges of 45 to 150 km (upper lithospheric mantle) and >200 km (deeper mantle layer) scaled by the phase difference. The dashed purple line on the upper panel shows the southeast limit of the region with consistent conductive direction in the upper lithospheric mantle. The red lines show the SKS fast direction defined by Senechal et al. (1996), Eaton et al. (2004), Rondenay et al. (2000), Evans et al. (2006), and Frederiksen et al. (2006, 2007) scaled by the time split. The plots are superimposed on the regional Bouguer gravity anomaly map (data obtained from Geological Survey of Canada). The white arrow, APM, shows the GPS-based absolute plate motion direction defined by Larson et al. (1997).



The results in the present study are considered a more robust statement of the conductive directions in the upper lithospheric mantle.

The most conductive directions derived in this analysis are co-plotted with the SKS fast direction and superimposed on the regional Bouguer gravity anomaly map of the study area in Fig. 6.6. The figure also shows the direction (277°) of the present-day absolute plate motion (APM) defined by Larson et al. (1997) based on Global Positioning System (GPS) analyses. The seismic anisotropy information was derived from shear wave splitting measurements by Senechal et al. (1996), Eaton et al. (2004), Rondenay et al. (2000), Evans et al. (2006), and Frederiksen et al. (2006, 2007).

To the northwest of the present study area the Grenville Front is characterized by an ~300 km wide Bouguer gravity low with central values of ~-50 mgal. This low has been modelled in terms of fossil response to lithospheric flexure during the Grenville orogen (Hynes, 1994). Farther to the southwest the anomaly is more moderate, and this observation has been interpreted as being due to a steeper angle of the dipping plate (Hynes, 1994). Easton (1992), Roy and Mereu (2000), Mareschal et al. (2000) and Bank et al. (2005) provide additional analyses of the gravity response in the vicinity of the Grenville Front. Within the study area, the Bouguer anomaly in the CGB is similar to that in the adjacent Superior Province, and is around 30 mgal greater in the CMB (Fig. 6.6). However, there are significant lateral variations within each tectonic unit. The MT sites in the CGB on Profile 3 lie mostly within two gravity highs. The northern high correlates spatially with the Nipissing domain of the Grenville Province and the southern high with the Parry Sound domain and surrounding units e.g., (White et al., 1994).

The MT anisotropy response is compared with the gravity data in order to examine spatial correlation of the MT responses with a measure of the large-scale physical property variation in the upper lithosphere, although it is not expected that the responses are necessarily caused by the same structures. The results show that across the study area there appears to be minimal correlation between spatial variations in the most conductive direction and in the Bouguer anomaly. The spatially-confined zone of E20°S strikes southwest of the Grenville Front on Profile 3 lies within the gravity high. However, beyond this observation few trends are evident.

There is significant obliquity between the electrical conductive direction and the SKS fast direction across most of the study area. As noted using period-based results by Ji et al. (1996) and Frederiksen et al. (2006), there is a very consistent obliquities between the conductive direction at lithospheric depths (N80°E) and the SKS fast direction (N103°E) along Profile 1 in the vicinity of the Grenville Front. The MT results indicate that this relationship persists at least 200 km to the east to Profile 2. Extended GB analysis, described below, shows that the conductive direction azimuth determined for profiles 1 and 2 also provides a reasonable statistical fit to the MT data on Profile 3.

South of 46° latitude there is greater correlation between the conductive directions and the SKS fast directions, particularly for the deeper mantle layer MT responses. There are a number of sites for which the two directions are very closely aligned, whereas at other sites significant obliquity is observed. However, unlike in the region around the Grenville Front this obliquity is localized. Frederiksen et al. (2006) determined that for 18 pairs of closely spaced MT and seismic sites south of 46°, the mean azimuthal difference between the conductive direction (for the 1-1000 s period range of the MT

data) and SKS fast direction is 14.2° . They interpreted this result to indicate the existence of a lithospheric component of anisotropy in this region.

Most of the SKS split directions in this region correspond closely to the present day APM direction, e.g., as defined by Larson et al. (1997) (Fig. 6.6). The MT conductive direction azimuths are also oblique to the APM direction.

6.5 Induction Arrows

An important justification for the interpretation of the presence of laterally-pervasive and extensive resistivity anisotropy in the vicinity of the Grenville Front was the observation of relatively small and inconsistent induction arrows across an area with very consistent MT response anisotropy (Kellett et al., 1992; Mareschal et al., 1995; Boerner et al., 2000). Kellett et al. (1992) show an example of the data on which these interpretations were based, the induction arrows for periods of 11 s and 85 s at sites in the Pontiac subprovince on Profile 1, to the north of the present study area. Heise et al. (2006) discussed the limitations of this interpretational method, and together with the recent results of Adetunji et al. (2014), who were able to fit combined MT and tipper data from Profile 1 to a reasonable statistical level with isotropic 2-D models, raise uncertainty about the validity of the interpretation.

For the present study tipper data are available for sites on profiles 1 and 3, but not on Profile 2, as the vertical magnetic field was not recorded at sites on this profile. As the tipper response is based on the presence of 2-D or 3-D resistivity structures, for a uniform source field (e.g., Jones and Spratt, 2002), it is inappropriate to use the 1-D Niblett-Bostick transform to represent these data. Also, the effects of a particular resistivity

structure may be observed at quite different periods in the MT and tipper responses (e.g., Wu et al., 2005). Figure 6.7 shows maps of the induction arrows at different periods.

The data were edited to exclude power-line effects and erratic points characterized by jumps in tipper magnitudes, tipper magnitudes greatly exceeding unity, and/or large error estimates. Based on the study of Zhang et al. (1993), short period (period <0.1 s) tipper data were excluded because of potential magnetic effects of galvanic distortion.

As noted in previous studies, induction arrows are relatively small in the study area. The typical length of the real induction arrows is 0.1–0.3 at 40 s period, 0.2–0.3 at 100 s period and 0.2–0.5 for periods between 320–500 s. As has been previously noted, the induction arrows appear to be somewhat erratic and spatially inconsistent. However, careful examination reveals a number of large-scale spatial trends in the data particularly at longer periods.

At a period of 40 s along Profile 1 and Profile 2 the real induction arrows exhibit spatially variable responses (Fig. 6.7) that are different from the upper lithospheric mantle geoelectric strike (Fig. 6.6). Kellett et al. (1992) showed similar response in the Pontiac subprovince. The azimuths of the real induction arrows tend to be fairly consistent within small groups of sites over spatial scales of around 50 km or less, so it is possible that this response is influenced by local, shallow or crustal scale structures. For example, the arrows point outwards from the region of E20°S strikes and the gravity high on Profile 3 south of the Grenville Front on Profile 3.

The imaginary induction arrow direction at 40 s period is much more spatially consistent than the real component. At a large majority of sites on profiles 1 and 2 the

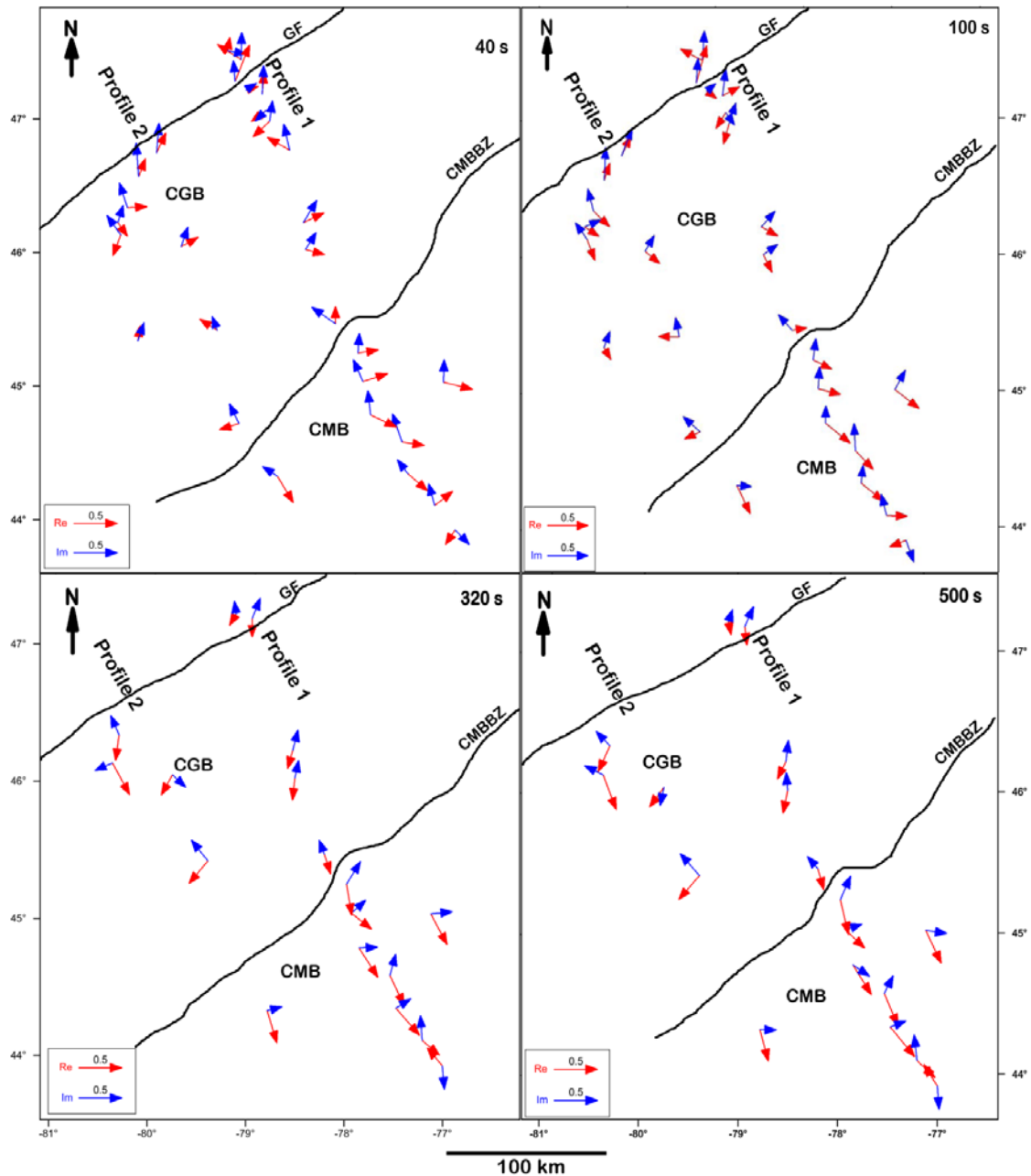
imaginary arrow has a significant northwards component. Kellett et al. (1992) showed that farther to the north, the imaginary component of the 11 s induction arrows is dominantly northwards in the southern Pontiac but reverses to southwards in the northern Pontiac. Examination of the theoretical tipper response of a contact between resistive and conductive quarter spaces shows at a given distance from a contact the effects of the contact are observed at a smaller period in the imaginary response than the real response, particularly on the resistive side of the contact. The observed results are thus consistent with the presence of relatively large-scale resistivity structures.

At longer periods there is a significant amount of spatial consistency in the induction arrow responses. At 320 s and 500 s the real induction arrows point mainly to the SSE (E70°S), although there is a reversal observed at the southernmost site on Profile 2 at an latitude of approximately 44° N. The larger, and thus better resolved, imaginary arrows tend to point in the antiparallel direction to the corresponding real arrows, providing support for the existence of large-scale 2-D resistivity structures (e.g., Booker, 2014).

The re-examination of the induction arrow responses near the Grenville Front using the larger data set available in the current study leads to a revised conclusion concerning their support for anisotropic resistivity. The induction arrow responses for periods longer than 100 s provide a strong indication of the presence of large-scale 2-D (or 3-D) resistivity structures. Even at shorter periods of 10-100 s the arrows appear to indicate the presence of macroscopic resistivity structures. The induction arrow results do not exclude the presence of resistivity anisotropy within the study area, but they do establish the presence of some large-scale resistivity structures. Thus, the conductive

directions established by the phase tensor and GB decomposition are now taken to represent the geoelectric strike of macroscopic resistivity structures.

Figure 6.7. The induction arrow responses at 40 s, 100 s, 320 s and 500 s, for sites along profiles 1 and 3. The real induction arrows (plotted using the Parkinson convention) are denoted by red arrows and the imaginary arrows are denoted by blue arrows. Note that responses with large uncertainties are excluded.



6.6 Determination of 2-D Inversion Azimuths

The results presented above provide justification for the application of 2-D modelling to the data from individual profiles. The upper lithospheric mantle and deeper mantle conductive directions show reasonable azimuthal consistency over each of the profiles (Fig. 6.6), particularly profiles 1 and 2, and the long-period induction arrow response (Fig. 6.7) provides further support for a 2-D modelling approach. The three-dimensionality indicators, i.e., the phase tensor skews and the GB RMS misfits, are also relatively low over the northern half of each profile. In the present section we use GB decomposition methods to establish the optimal geoelectric strike for 2-D modelling the data from each profile.

6.6.1 Methodology

To determine a regional geoelectric strike for the 2-D modelling of each of the profiles, four methods are used to assess the strike and its uncertainty. Each of the methods makes use of the GB misfit for a single depth estimate at a single site. The squared GB misfit for site j for each depth estimate k is defined in terms of re-parameterized impedance terms $\alpha_1 = Z_{xx} + Z_{yy}$, $\alpha_2 = Z_{xy} + Z_{yx}$ and $\alpha_3 = Z_{xy} - Z_{yx}$ as:

$$e_{jk}^2 = \sum_{i=1}^3 \left[\frac{(\text{Re} \alpha_{ik}^{\text{obs}} - \text{Re} \alpha_{ik}^{\text{model}})}{\sigma_{\alpha_{ik}}} \right]^2 + \sum_{i=1}^3 \left[\frac{(\text{Im} \alpha_{ik}^{\text{obs}} - \text{Im} \alpha_{ik}^{\text{model}})}{\sigma_{\alpha_{ik}}} \right]^2 \quad \text{Eq. 1}$$

where the model response depend on the GB distortion parameters and the regional strike θ_r for the depth band under consideration and σ_{α} is the estimated standard deviation for each datum (e.g., Groom and Bailey, 1989; McNeice and Jones, 2001).

A common experience in GB decomposition of MT data from sites in Precambrian crystalline terranes is that the minimization provides statistically reasonable levels of misfit for the data from most sites, but very poor results for the data at a small number of the sites. The observations suggest that the "geological noise" in the GB fit (e.g., local departures of the data from the underlying assumption of a 1-D or 2-D structure) is distributed very unevenly between the sites. The higher levels of noise at some sites may be due to a strong local 3-D inductive response or to other effects. A common approach in this situation is to simply remove those sites and repeat the GB analysis. Here, we examine the possibility of using more robust estimate of the misfit.

The suite of methods we used is as follows:

Method 1: The first estimate is the normal multi-site regional strike provided by the McNeice-Jones extension of the GB method (GB-MJ) implemented in the STRIKE program. This estimate is the regional strike that minimizes the summed squared misfit for the band of depths and for a group of sites:

$$E_1 = \sum_{j=1}^M \sum_{k=1}^N e_{jk}^2 \quad \text{Eq. 2}$$

Method 2: The second method involves calculating and plotting the total squared error $e_j^2(\theta_r)$ for the depth band for individual sites at specified values of the regional strike (e.g., Schmoldt, 2011). This method permits examination of how different sites contribute to the constraint of the regional strike through inspection of the RMS error distribution along the profile.

Method 3: The third method is based on examining a sum of the absolute value of the error for a range of specified values of the regional strike:

$$E_2(\theta_r) = \frac{1}{MN} \sum_{j=1}^M \sum_{k=1}^N |e_{jk}| \quad \text{Eq. 3}$$

The use of an L_1 norm means that this measure will be less sensitive than the GB-MJ measure E_1 to individual sites with very large misfits. Plots of E_2 versus θ_r (e.g., Pous et al., 1997) can be used to obtain a visual estimate of the uncertainty in the strike azimuth; broad minima indicate poorly resolved strikes whereas narrow minima indicate tightly resolved azimuths.

Method 4. In the final method an additional allowance is made for outliers in the noise at different sites by using the median rather than the mean of the depth-band misfits:

$$E_3(\theta_r) = \text{median} \left[\frac{1}{N} \sum_{k=1}^N |e_{jk}| \right] \quad \text{Eq. 4}$$

For all analyses we used depth bands corresponding to the upper lithospheric mantle (45-150 km) and deeper mantle layer (>200 km), as in Section 3. Results were also obtained for crustal depths but are not reported here. The analyses used the default impedance error floor of 3.5%. Methods 2 to 4 were implemented using single-site decomposition and regional strike angles constrained from 0° to 90° with 1° increments. The McNeice and Jones (2001) STRIKE program was treated as a “black box”; we extracted values from the list of misfit values for the different depths provided in its output files to formulate the various error measures.

During the analyses we found that, despite being more robust to sites with higher noise levels, methods 3 and 4 usually failed to produce reasonable results if data from such sites were included. For example, erratic jumps occurred in error versus strike angle when these sites were used. These effects are attributed to the complexity of the GB

misfit in the presence of noisy data. Despite this limitation, the availability of the full suite of strike results enhanced the strike determination process. We used methods 2, 3 and 4 to choose which sites to exclude from the final analyses, method 2 to determine which sites controlled the final strike, and methods 3 and 4 to estimate the confidence in the strike results. We also compared all of the estimates to obtain robust final results. Regional strikes determined using the different methods are listed in Table 6.1; figures 6.8 and 6.9 show the results from Method 2 and Fig. 6.10 shows the results from methods 3 and 4.

6.6.2 Common strike azimuth for the upper lithospheric mantle

6.6.2.1 Profile 1

As shown in Fig. 6.6, the strongest control on the strike direction on Profile 1 come from several sites northwest of the Grenville Front and in the northwestern CGB. These sites all exhibit lowest misfit for strike directions between about N75°E and N90°E. The azimuth is less defined in the southeastern CGB, but the data at most sites is compatible with a similar result to further north. The response at sites in the CMB is far more erratic with several sites having large misfits at all regional strike azimuths (PSO003, PSO013, and PSO042). At sites where the azimuth is better defined, PSO014, PSO040, PSO041, and PSO001, the minimum misfit occurs for a southwest-northeast strike with an azimuth between about N45°E and N65°E. This change in strike angle at upper lithospheric mantle depths was also noted in the phase tensor results and previous GB strike results (Fig. 6.6).

After excluding sites with high rms misfit (PSO001, PSO003, PSO009, PSO013, PSO042 and PSO039) from the analysis, the MJ-GB multisite multi-frequency analysis yields an optimal direction of N85°E for the profile. In Fig. 6.10, the graphs of E_2 and E_3 versus strike angle, obtained after exclusion of the same sites, exhibit broad minima between N70°E and N90°E. Using these graphs we determine an overall strike angle for

Table 6.1. Summary of the geoelectric strike azimuths. Method 1 is the azimuth derived from the mutli-frequency multi-site GB-MJ decomposition, Method 3 is the azimuth based on the minimum mean absolute error at each site and Method 4 is the azimuth based on the median absolute error at each site. The table also shows the azimuth adopted for 2-D inversions used to obtain the lithospheric mantle resistivity structure. All azimuths are defined clockwise from true north.

	Upper Lithospheric Mantle Depth Range			Deeper Mantle Depth Range		
	Profile 2	Profile 1	Profile 3	Profile 2	Profile 1	Profile 3
Mareschal et al. (1995)	-	$80^\circ \pm 6^\circ$	-	-	-	-
Method 1	85°	85°	86°	57°	62°	67°
Method 3	$85^\circ \pm 4^\circ$	$80^\circ \pm 10^\circ$	$84^\circ \pm 5^\circ$	$65^\circ \pm 5^\circ$	$65^\circ \pm 5^\circ$	$62^\circ \pm 5^\circ$
Method 4	$80^\circ \pm 6^\circ$	$80^\circ \pm 10^\circ$	$85^\circ \pm 5^\circ$	All angles	$65^\circ \pm 6^\circ$	$65^\circ \pm 6^\circ$
2-D inversion azimuth	85°	85°	85°			

Figure 6.8. RMS misfit for single sites versus strike azimuth for a GB fit to the upper lithospheric depth band (45–150 km). Results are plotted in order of the position of each site on the profile but the inter-site distance is not preserved. The red dashed line is the regional strike azimuth of the profile based on the GB-MJ multisite multifrequency analysis and the white dashed line is based on the minimum misfit of the two more robust misfit measures. The gap in Profile 1 indicates the absence of data for this depth range at that site.

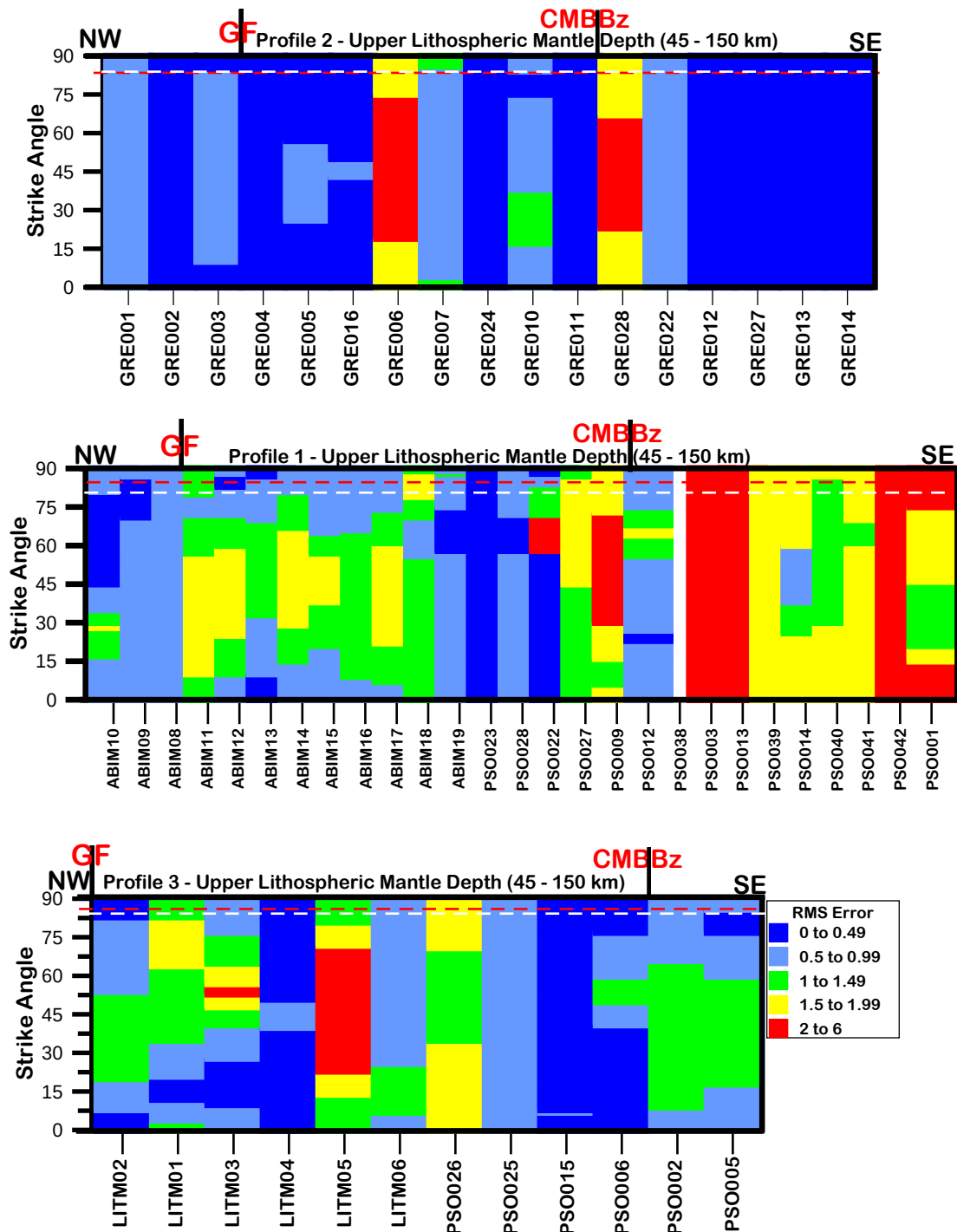


Figure 6.9. RMS misfit for single sites versus strike azimuth for a GB fit to the deeper mantle depth band (>200 km). The red dashed line is the regional strike azimuth of the profile based on the GB-MJ multisite multifrequency analysis and the white dashed line is based on the minimum misfit of the two more robust misfit measures. The gaps in the results indicate the absence of data for this depth range at the corresponding sites.

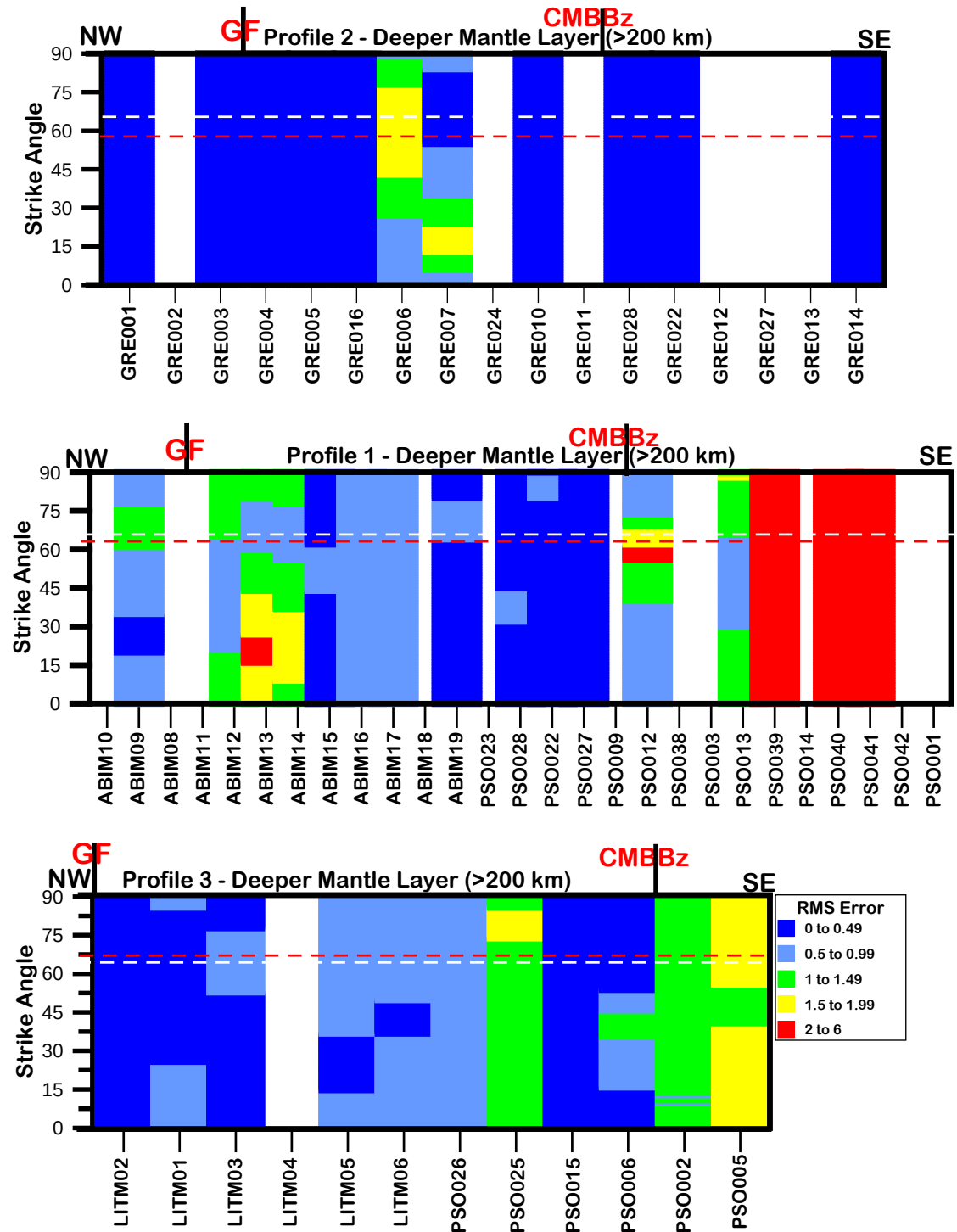
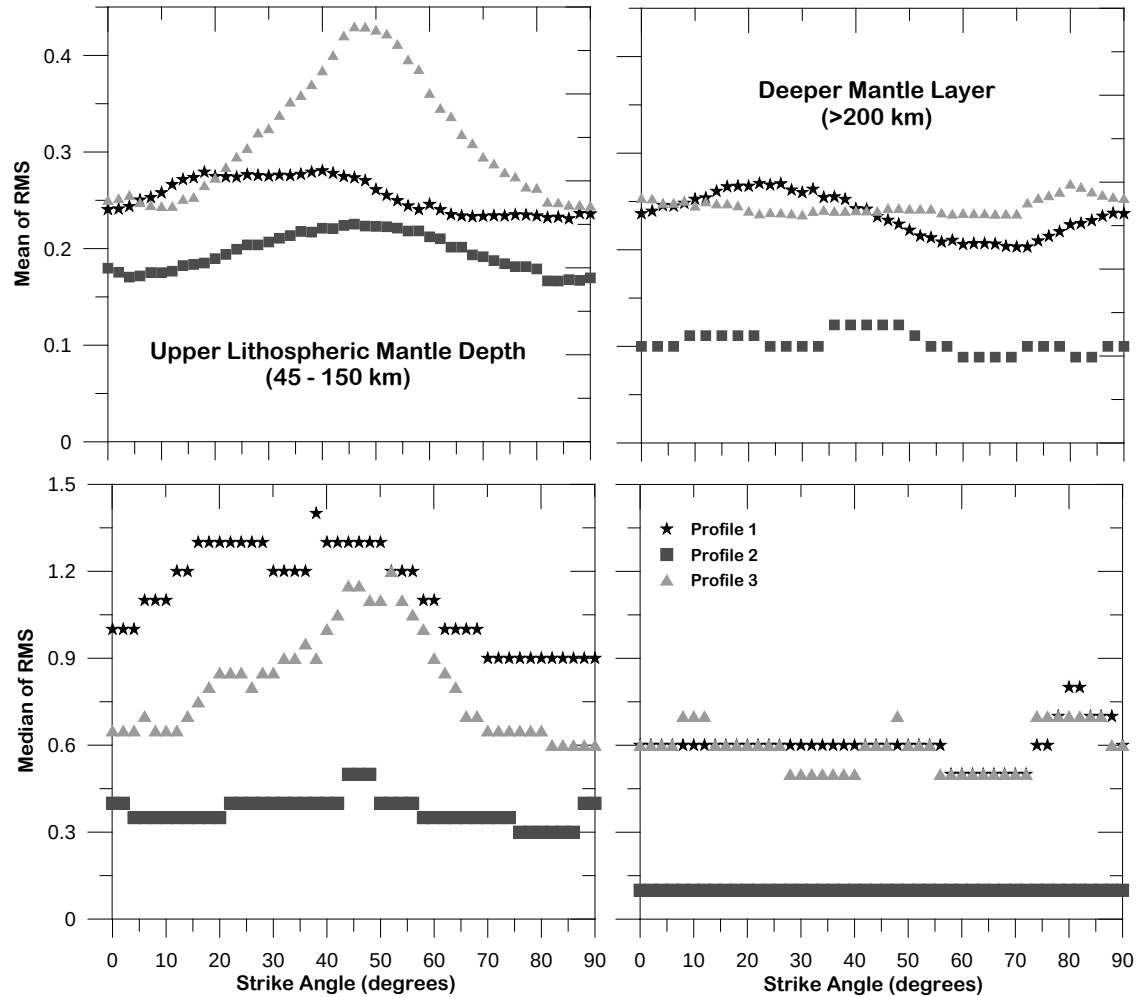


Figure 6.10. Misfit versus azimuth plots for robust misfit estimates. The plots show, as a function of regional strike azimuth, the minimum of the mean (top) and median (bottom) of the absolute value of the misfit determined in GB decompositions at individual sites. The final results are based on only those sites with RMS misfit of <2.0 .



the profile of $N80^{\circ}E \pm 10^{\circ}$. The GB-MJ result lies well inside this range. Taken together, the results for Profile 1 indicate that $N85^{\circ}E$ is the appropriate strike angle for 2-D inversion. However, the part of the inversion model corresponding to the CMB, where the true strike angle is closer to $N55^{\circ}E$, will need to be interpreted cautiously.

6.6.2.2 Profile 2

With the exception of sites GRE006 and GRE028 (which were eliminated from multi-site analyses) (Fig. 6.8) the misfit is relatively low for all sites and all azimuths on Profile 2. The misfit values for this depth range are lower than on the other profiles. Schmoldt (2011) suggests that responses where RMS misfit is low in most directions indicates locally 1-D sub-surface resistivity structure. However, this observation can also occur in 2-D structures for which the TE and TM phase curves cross or take on similar values at the periods corresponding to the depth range under consideration.

The geoelectric strike direction for Profile 2 determined by MJ-GB (N85°E) agrees well the azimuth determined by the minimum of E_2 (N85°E $\pm 4^\circ$) and E_3 (N80°E $\pm 6^\circ$). Misfit curves for Methods 3 and 4 show that the optimal strike direction for Profile 2 is defined more accurately than for Profile 1. The results for Profile 2 indicate that N85°E is the optimal strike angle for 2-D inversion.

6.6.2.3 Profile 3

Profile 3 contains 12 unevenly distributed sites with six closely spaced sites in the northwest and six more sparsely spaced sites in the southeast. Based on the RMS misfits (Fig. 6.8), the optimal geoelectric strike direction on the northwestern half of the profile is around N15°E. Allowing for the 90° ambiguity in the strike determination this result agrees well with the E20°S conductive direction in the earlier determinations (Fig. 6.6). However, the results in Fig. 6.10 show that for these sites a strike between N90°E and N80°E also produces a low misfit. This result is very significant for interpreting the extent of the MT response anisotropy previously observed on Profile 1. It shows that

although the minimum misfit on Profile 3 corresponds to strikes that are oblique to that on Profile 1, it is possible to fit the observations reasonably well with a common strike azimuth. The differences observed in the minimum misfit results may be caused by a subtle difference, such as some residual influence of crustal strike effects in the Profile 3 results. Within the southeastern CGB and CMB the results suggest an upper lithospheric mantle strike angle in the N90°E and N80°E azimuthal range. The misfits are generally high for sites LITM05 and LIT026 when compared with other sites along the profile, and these sites were eliminated from further geoelectric strike analyses.

The MJ-GB strike determined for this profile is N86°E. The E_2 and E_3 misfit curves indicate results of $84^\circ \pm 5^\circ$ and $85^\circ \pm 5^\circ$ respectively in good agreement with the MJ-GB value. The results for Profile 3 indicate that N85°E is an appropriate strike angle for 2-D inversion, meaning that a common azimuth can be used for modelling all three profiles.

6.6.3 Common strike azimuth for the deeper mantle layer

The strike direction in this deeper mantle layer range is determined mainly for the purpose of comparing the geoelectric strike results with absolute plate motion directions. Figure 6.9 shows the individual site misfit versus regional strike for this depth range. The plot includes a number of sites with missing data corresponding to sites in which there are no responses with equivalent Niblett-Bostick depths larger than 200 km.

The RMS misfit distribution of Profile 1, for the deeper mantle depth range (Fig. 6.9), shows that the regional strike is defined mainly by sites in the CGB. The data from the CMB is characterized by high RMS misfits, and there is no distinct strike direction

defined in this region. The MJ-GB multi-site multi-frequency analysis yields an optimal direction of N62°E for the whole profile, and minimization of the more robust measures of misfit yields values of N65°E $\pm 5^\circ$ and N66°E $\pm 5^\circ$ for Methods 3 and 4 respectively.

The data from only 11 sites from Profile 2 penetrate to deeper than 200 km and, as observed at upper lithospheric depths, the corresponding GB misfits are characterized by very low values and low sensitivity to the regional strike azimuth, likely due to large errors on the impedances. However, the results from Methods 2 and 3 do define an overall strike angle that is in good agreement with that found for Profile 1. The MJ-GB multi-site multi-frequency analysis produces an optimal strike direction of N57°E for the whole profile, and Method 3 produces a strike of N65°E with an uncertainty of 5° (Fig. 6.10). For this data set there was no significant dependence of the Method 4 misfit measure with azimuth.

For Profile 3, the RMS misfit of the distortion model is generally low for the CGB section of the profile but exceeds 1.0 for the CMB section. This observation suggests that a wide range of strike azimuths provide adequate GB fits to the CGB section of the profile. The MJ-GB multi-site, multi-frequency analysis yields an optimal strike direction of N67°E for the whole profile. Method 3 yields an azimuth of N62°E $\pm 5^\circ$ and Method 4 yields a result of N65°E $\pm 6^\circ$.

For each profile, the result of the overall strike for deeper mantle depth range is defined with greatest resolution for Profile 1 and with least resolution for Profile 2. The result shows very good agreement across the whole study area and indicates an overall strike azimuth of N65°E. This result confirms the estimate made above from inspection of the GB results for individual sites (Fig. 6.6).

6.7 Isotropic and Anisotropic Two-Dimensional Inversion

6.7.1 Preparation of data sets for inversion

Prior to 2-D modelling and inversion, MT datasets for each profile were prepared by decomposing them to the regional geoelectric strike azimuths. This is done by fitting the regional impedances derived from the GB decomposition with the regional geoelectric strike azimuth constrained to the upper lithospheric mantle value of N85°E. Note that the data were not rotated, but were fit to an appropriate model of distortion with the defined strike angle. Such model fitting yields far superior results than rotation alone (Jones and Groom, 1993; McNeice and Jones, 2001).

In 2-D, the MT responses formally decouple into two independent modes of induction and are termed the transverse electric (TE) and the transverse magnetic (TM) modes. For the TE mode, the electric field is parallel to the geoelectric strike, and for the TM mode it is perpendicular. These terms form the off-diagonal elements of a 2-D impedance tensor aligned with strike, with the TE mode usually in the upper right and the TM mode in the lower left. The TE mode is sensitive to current flow, which is dictated by the conductance of conductive regions of the model, but provides lower resolution of the lateral position of structures than TM. The TM response is primarily sensitive to charges on conductivity contrasts or gradients of lateral boundaries. Inclusion of both modes in the inversion, which is formally a joint inversion due to the decoupling of Maxwell's equations into two independent sets, allows for superior determination of the subsurface resistivity structure. Two-dimensional inversion methods are sometimes applied to MT data containing weak 3-D effects. When the conductive features in a model have a finite strike length, the TE response is generally more distorted than the TM response and this

necessitates the weighting of the 2-D modelling heavily towards the TM response (Jones, 1983b; Wannamaker, 1999; Ledo et al., 2002; Ledo, 2005). There are some structures for which the TM mode is more affected by 3-D effects than the TM mode (e.g., Park and Mackie, 1997), so one must be cautious when only modelling the TM mode.

6.7.2 Inspection of pseudosections for qualitative information

Prior to the actual geophysical inversions the data were edited and examined using TE and TM pseudosections. Consistency between apparent resistivity and phase responses at individual sites was checked using the D+ approach of Parker (1980), as implemented in the WinGLink software. This method performs 1-D modelling of the admittance which identifies unreliable data points that were subsequently removed prior to modelling and inversion.

The MT pseudosection for Profile 1 is presented and discussed in Adetunji et al. (2014) so will not be included here. The MT pseudosections for profiles 2 and 3 data sets are shown in Figs. 6.11 and 6.12 respectively. Examination of the TE and TM apparent resistivity and phase responses shows that there are significant differences between these modes along both profiles. This is an indication that 2-D structures exist in the subsurface, along both profiles.

Along Profile 2 (Fig. 6.11), the apparent resistivity response is dominated by high values (some exceeding 8000 $\Omega.m$) over a period band between 0.01 s and 10 s. This response extends from the northwestern end of the profile southeast to the middle of CMB where the resistivity decreases to <100 $\Omega.m$ for periods greater than 0.1 s. In areas around the Grenville Front low resistivity values are observed at periods longer than 100

s. However, in the central part of the profile, corresponding to the CGB, the response at these periods is resistive, particularly for the TM mode. The long period response is relatively conductive in the CMB. The transition from high apparent resistivity at intermediate periods to lower apparent resistivity at long periods is reflected, as it should be, in a period band of high phase responses ($>50^\circ$) centered on 10 s extending along the entire length of the profile. Increased phase at the longest periods (>100 s) provides an indication of an increase in conductivity at depths corresponding to the maximum penetration of the signal. The pseudosections also provide evidence for a localized conductor at the southeastern end of the profile in the CMB. This feature is defined by low apparent resistivity values in both TE and TM apparent resistivity responses at intermediate periods, and increased TE and TM mode phase values at periods centered around 10 s (Fig. 6.11).

The pseudosection for Profile 3 (Fig. 6.12) shows responses that are similar to those in the Profile 2 pseudosections (Fig. 6.11). For Profile 3 the apparent resistivity responses for both modes, dominated by high values ($>5000 \Omega.m$) over a period band of 0.01 to 100 s from the northwest end of the profile southeast to the middle of CMB where the resistivity decreases to $<2000 \Omega.m$. The high resistivity response extends to longer periods (~ 1.000 s) in the northwestern CGB and in the vicinity of the CMBBZ. A significant difference between Profile 2 and Profile 3 occurs in the central CGB, where the response along Profile 3 is significantly more conductive than along Profile 2, particularly in the TM mode. Another difference between the profiles is that, for sites between the middle of the CGB and the southwest end of the profile along Profile 3, the response at periods shorter than 0.1 s is characterized by $<400 \Omega.m$ resistivity.

Figure 6.11. The apparent resistivity and phase pseudosections of TE and TM responses for Profile 2. The observed data are the regional MT responses derived from the GB decomposition using the upper lithospheric mantle strike azimuth of N85°E. Panels shows the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.

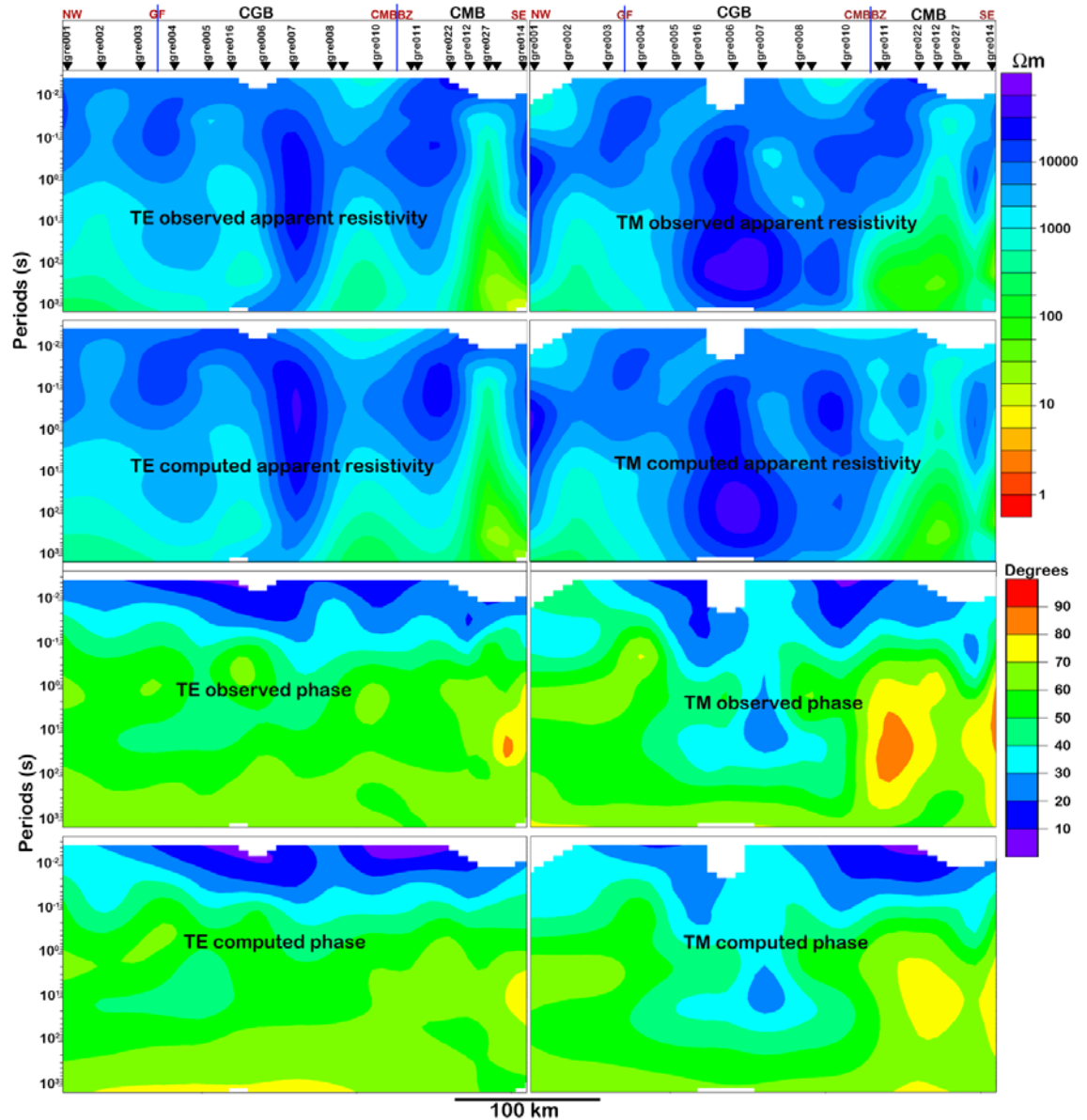
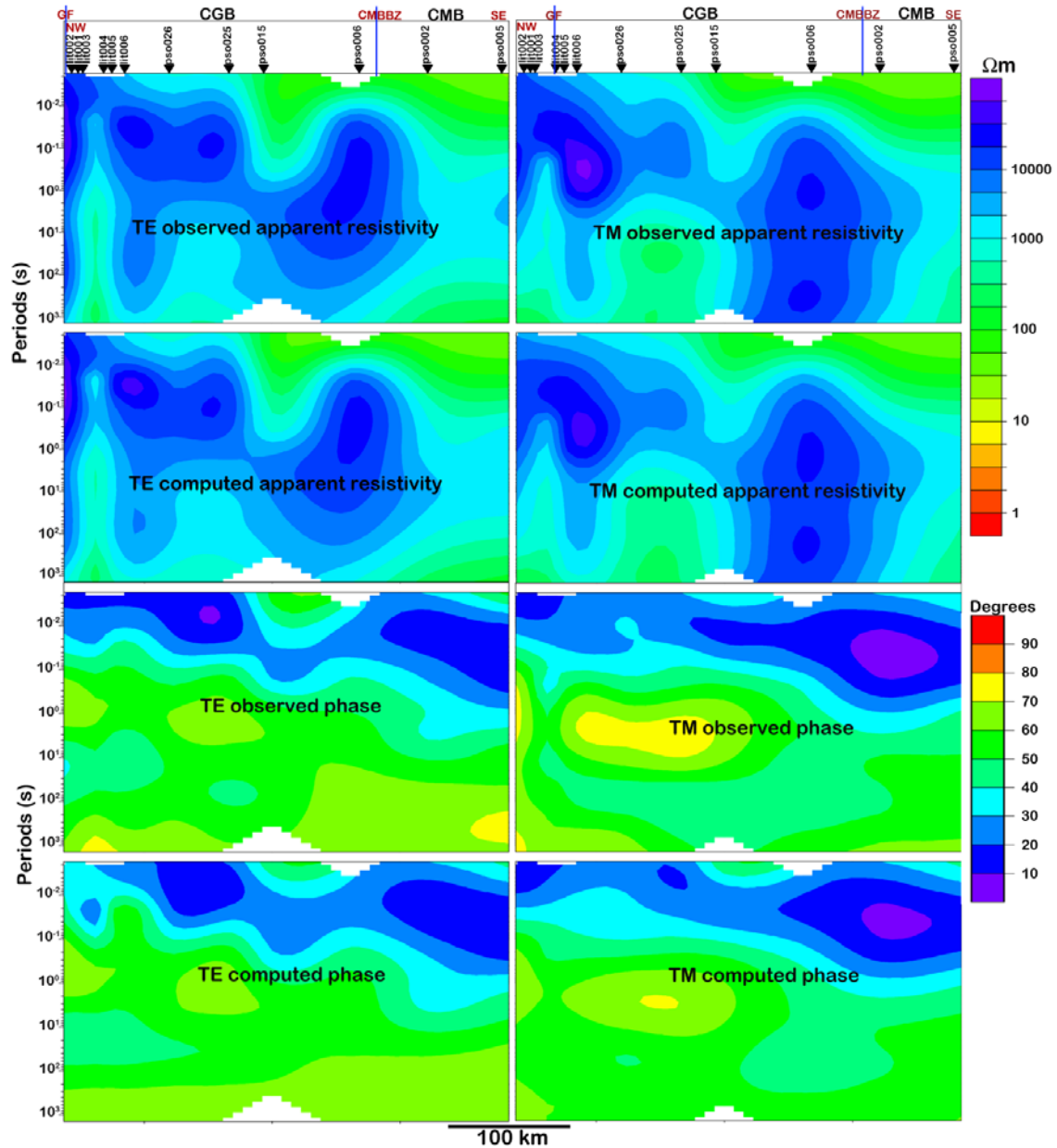


Figure 6.12. The apparent resistivity and phase pseudosections of TE and TM responses for Profile 3. The observed data are the regional MT responses derived from the GB decomposition using the upper lithospheric mantle strike azimuth of N85°E. Panels show the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.



The region of low impedance phase ($<30^\circ$) that extends laterally from the northwest to the southeastern of the profile indicates transition from the conductive response at short periods to resistive responses at intermediate periods. As in Profile 2, the transition from high apparent resistivity at intermediate periods to lower apparent resistivity at longer periods is indicated by high phase responses ($>50^\circ$) that are centred on 1 s. In the centre of the CGB the more conductive long period responses observed on Profile 3 correlate with high phase values ($>70^\circ$ in the TM response).

6.7.3 Isotropic inversions

In this study we used the 2-D isotropic modelling and inversion algorithm of Rodi and Mackie (2001) as implemented in Geosystem's WinGLinK software. The software uses a finite difference numerical method for forward modelling and the non-linear conjugate gradient (NLCG) method for inversion. NLCG implements direct iterative minimization of an objective function that penalizes data residuals and second spatial derivatives with respect to resistivity structures (Rodi and Mackie, 2001, 2012).

The inversion models were obtained by solving for the smoothest model using uniform-grid Laplacian regularization and minimizing the integral of the Laplacian (Rodi and Mackie, 2012). The range of possible models that could fit the data was explored by running several 2-D inversions using different combinations of inversion parameters (both regularization and smoothing parameters) and data. Final 2-D models were obtained by fitting both TE and TM data for all the sites using a $100\ \Omega\text{m}$ half space as the starting model. The standard deviation error on each data point was set to either the calculated error values or to an error floor, whichever one is the larger. For the initial set

of inversions, the apparent resistivity error floors were set at 60% for both modes, the TE phase error floor was set at 30% and the TM phase error floor at 5%. The apparent resistivity was down weighted in order to reduce the effect of static shift in the inversion. These floors emphasize the fit to the TM phase, and yield an appropriate model that served as the start model for the next inversion step. All the error floors were subsequently reduced to 20% and 16% for TE and TM resistivity and 5% ($\sim 1.2^\circ$) and 4% ($\sim 1^\circ$) for TE and TM phases. Thus again the TM data were prioritized over the TE data. At the beginning of every new inversion, the inversions were based on an interpolated data set with five frequencies per decade but the data set was changed to the actual station data, through re-starting, as the inversion progressed. During inversion static shift was also determined and corrected. For a more reliable final model with deeper minimum of the objective function, all inversions were re-started at least once after its initial termination. The typical number of iterations per sequence ranges from 400 to 600, and the number of sequences was 6-10 depending on final models.

Exploring for the optimal model involves the use of different combinations of weighting functions and regularization parameters (e.g., Mackie et al., 1997; Spratt et al., 2009; Matsuno et al., 2010; Schäfer et al., 2011). Parameters such as the Tikhonov regularization parameter (τ), horizontal derivative (α), depth function (β), horizontal smoothing (H) and vertical smoothing (V) factors were examined to explore for the optimal model. Following this examination, for the final inversion models, the weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for both profiles while $\tau = 3$ and $\tau = 6$ were used for profiles 2 and 3 respectively. These values were chosen after a series of initial inversions were performed to determine the best smoothing

and regularization parameters, using an L-curve approach (Hansen, 1992). The L-curves and several initial models are shown in Appendix B.

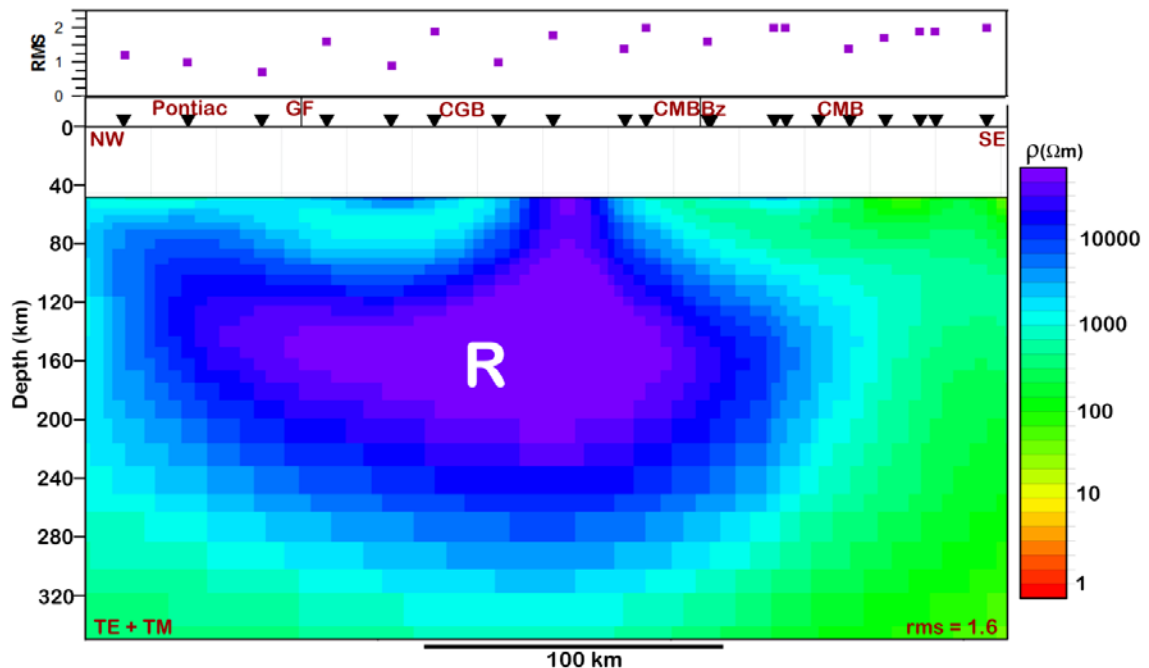
The modelling results for Profile 1 are presented in Adetunji et al. (2014) and are not repeated here. The final model for this profile is shown in Fig. 6.2. An isotropic 2-D model for Profile 1 was also obtained as part of an anisotropic modelling study as discussed below.

Figure 6.13 shows the mantle resistivity model obtained for Profile 2 together with site-by-site data misfits. The crustal part of the resistivity model is not considered here as the optimal crustal model is derived using a different strike angle from the N85°E azimuth that is most appropriate for the upper lithospheric mantle. The northwestern half of the model shows a relatively good fit to the data, with acceptable RMS misfit values that are mostly <2. However, to the west of CMBBZ, the fit becomes poorer as most sites consistently show values closer to 2. The global RMS misfit of the final mantle resistivity model is 1.6. The pseudosection response of the model is shown in Fig. 6.11; it shows a good fit of the model to all of the major data features. The inversion model reproduces all of the large-scale features in the observed data (Fig. 6.11).

The resistivity model for Profile 2 is dominated by a large-scale resistive region, labeled R, which extends laterally from the Pontiac subprovince into the middle of the CMB. This is a very well resolved feature, defined by 13 MT sites, that extends from ~70 km depth to a maximum of ~300 km depth in the centre of the CGB. Its resistivity is 3,000 to 50,000 $\Omega\cdot\text{m}$. The lithospheric resistor lies along strike from a resistive region in the upper mantle lithosphere on Profile 1 (Fig. 6.2). The top of the resistive zone is at a similar depth in both profiles. On Profile 1 it has a clear southeastwards dip and increases

in depth from ~50 km, in the middle of the CGB, to ~120 km at its southeastern extent. For Profile 2 there is also some indication that the top of the feature has a southeastern dip as it again increases in depth from ~40 km, in the CGB, to ~120 km at its southeastern extent. However, further to the northwest the upper 100 km of the resistivity structure is more complex and includes some conductive regions. The resistor extends to much greater depths on Profile 2 than on Profile 1 where its base is at a depth of ~150 km.

Figure 6.13. Resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The RMS misfits of the joint TE and TM (purple squares) inversions at individual sites are plotted above the inversion model. The crustal section (upper 48 km) of the model is not shown as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. This figure is reproduced with the inclusion contour lines in Appendix B.9



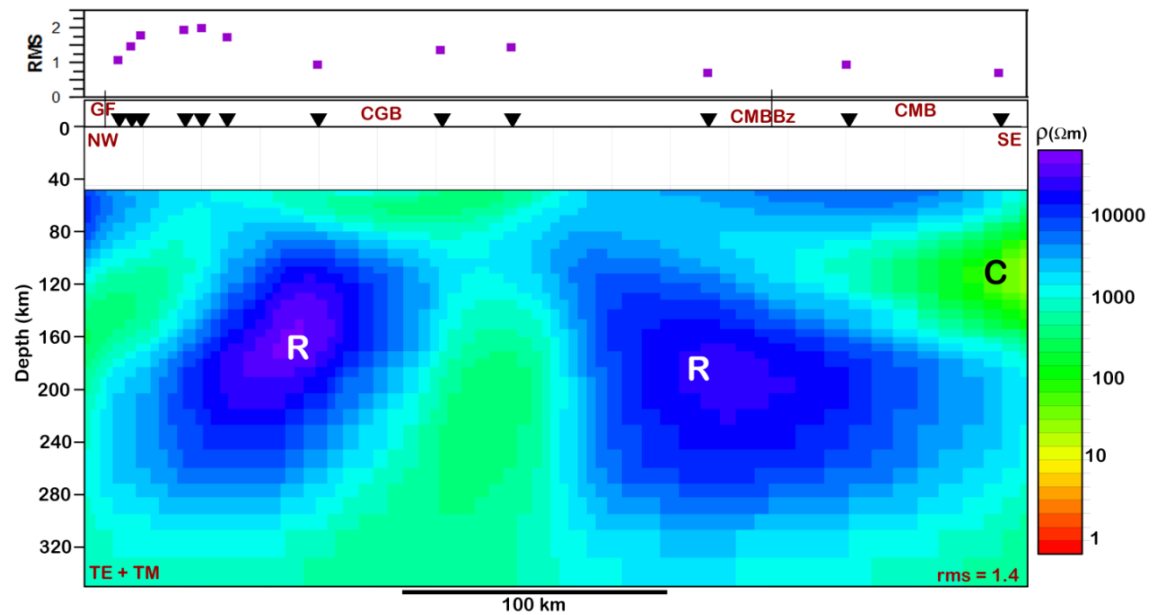
The mantle lithosphere in the southeastern end of Profile 2, beneath the CMB, is relatively conductive (mostly $<100 \Omega.m$). On both profiles 1 and 2, the southern margin of the resistive lithosphere is approximately 100 km southeast of the CMBBZ.

The mantle resistivity model for Profile 3, along with misfits for individual sites, is shown in Fig. 6.14 and the pseudosection response is shown in Fig. 6.12. The average site RMS misfit for sites close to the GF is around 2, and this may be as a result of the closeness of the sites in this region. Other sites along the profile show good fit to the data, with values mostly <1 . The model provides a very good fit to the data with a global RMS misfit of 1.4. Examination of the pseudosection responses shows that all of the main features in the observed responses are reproduced. The model responses slightly underestimate the high phase anomaly observed at intermediate periods in the central CGB.

The resistivity model along Profile 3 is again dominated by a large-scale resistor (labelled R in Fig. 6.14) that extends from 80 to 280 km depth with a resistivity of 3,000 to over 50,000 $\Omega.m$. This feature is defined by the data from a number of the sites along this profile. As on profiles 1 and 2, the southeastern margin of the resistor again occurs at a location about 100 km southeast of the CMBBZ. In contrast, though to the mantle resistivity models for profiles 1 and 2, the resistor is separated into two parts by a conductive (500 $\Omega.m$) region in the central CGB. This feature is clearly supported by the observation of a phase anomaly at ~ 1 s period in this location and by decreased apparent resistivity observed at longer periods. It also appears in early iterations of the inversions, regardless of the parameters or starting model used. The upper 120 km of the Profile 3 resistivity model is more complex than at the same depth range in the models for the

other profiles. It includes a conductive region within the northwest part of the CGB and a resistive zone in the southern CGB and CMB.

Figure 6.14. Resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set (V.E = 0.5). The RMS misfits of the joint TE and TM (purple squares) inversions at individual sites are plotted above the inversion model. The crustal section (upper 48 km) of the model is not shown as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response. This figure is reproduced with the inclusion of contour lines in Appendix B.15



The resolution of the resistivity models, and particularly of the large scale resistor in each model, was assessed through examination of signal penetration and through hypothesis-testing of alternative models. Adetunji et al. (2014) describe a similar analysis for Profile 1. Examination of the equivalent Niblett-Bostick depths for responses on Profiles 2 and 3 (e.g., as shown in Fig. 6.4) demonstrates that there are a number of sites with penetration to the base of the resistors. The resistors are defined by an increase in apparent resistivity, and corresponding decrease in phase, at intermediate to longer

periods (Figs. 6.11 and 6.12). The base of the resistor is defined by an increase in phase at the longest periods (>100 s) and, as noted above, the gap in the middle of the resistor on Profile 2 is defined by a decrease in apparent resistivity and corresponding increase in phase at shorter periods. Finally, inversions were re-run using starting models based on the inversion results, but with the resistive zones replaced by more conductive zones. Even when the minimization norm was set to return the model closest to the starting model, the models obtained still included the resistors, indicating that they are data-supported structures.

6.7.4 Anisotropic Inversions

The MT data for Profile 1 were re-modeled using 2-D anisotropic inversion. Profile 1 was chosen as the first profile for this analysis because it has the largest number of MT sites and the best coverage of the tectonic units in the northern Grenville Province. The approach was not applied to the other profiles after it was determined that there was minimal evidence for resistivity anisotropy on Profile 1.

The anisotropic inversions used the 2-D anisotropic inversion code of Mackie (2002) and Baba et al. (2006). This code fits the observed data using an anisotropic model in which the anisotropy is aligned with the structural directions, along-strike (xx), across-strike (yy), and vertical (zz). The code uses the nonlinear conjugate gradient (NLCG) method to find a solution to the 2-D inverse problem (Mackie, 2002). Tikhonov regularization is implemented in terms of smoothness and anisotropy parameters. An isotropy weighting parameter (τ_{iso}) is used to control the degree of anisotropy (e.g; Evans et al., 2005; Evans et al., 2011; Matsuno et al., 2011; Miensopust and Jones, 2011; Le

Pape et al., 2012; Schmoldt and Jones, 2013). Small values of this parameter will allow for variations between the results of the diagonal of the anisotropic resistivity tensor. Values of τ_{iso} close to zero allow the models to be as anisotropic as possible, i.e., the TE and TM modes are fit independently, while the case is reversed for values of $\tau_{\text{iso}} \geq 100$ (Mackie, 2002).

Figure 6.15 compares the isotropic resistivity model for the mantle with the components of the anisotropic model produced using a smoothing regularization parameter of $\tau = 6$ and $\tau_{\text{iso}} = 1.0$ and Table 6.2 lists the results of the inversion. The isotropic and anisotropic models were obtained using identical grid and inversion parameters. The isotropic model differs slightly from the one from Adetunji et al. (2014) shown in Fig. 6.2, and the differences are attributed to differences in the gridding and smoothing parameters between these models. The final anisotropic model produced a slightly better data fit (rms = 2.05) than the isotropic model (rms = 2.43). The decrease in misfit is relatively small considering the use of an anisotropy regularization factor of 1.0.

The anisotropic resistivity model for the mantle is very similar to the isotropic model; the lithospheric resistor has an identical geometry in all components of the anisotropic model. The mantle conductor at the northwest end of the model also has a similar form in all components of the anisotropic model. Figure 6.16 compares the resistivity on vertical cross-sections through the resistor and the conductor. In general, the anisotropic model provides a smoother model than the isotropic model with lower resistivity in the resistor and higher resistivity in the upper part of the conductor. The results show minimal evidence for anisotropy in the mantle lithosphere with all three components of the resistivity model exhibiting very similar resistivity values.

Figure 6.15. Isotropic and anisotropic inversion (using $\tau_{iso} = 1$) models for Profile 1. The anisotropic inversion results are for the along-strike ρ_{xx} , cross-strike, ρ_{yy} , and vertical ρ_{zz} component of the resistivity. The crustal section (upper 48 km) of the model is not shown.

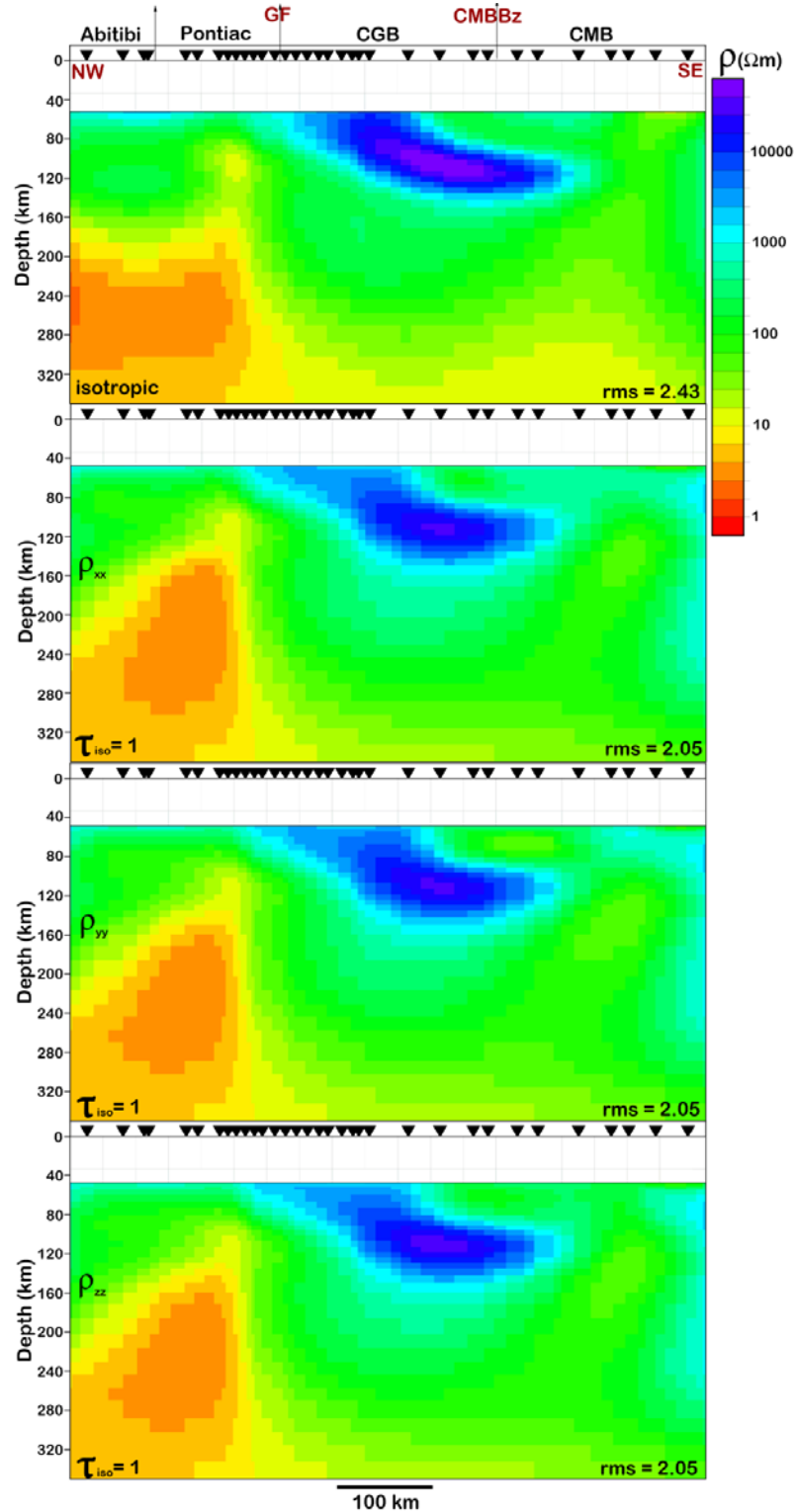
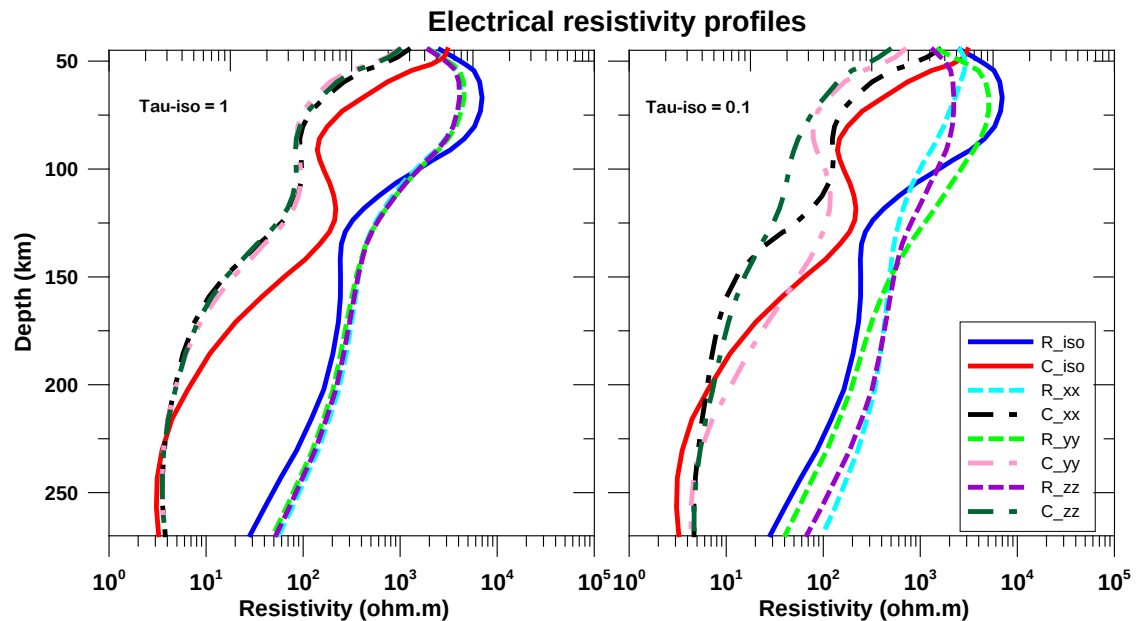


Table 6.2. Summary of parameters and results of isotropic and anisotropic inversion for Profile 1. Figure 18 shows the location of the parts of the models averaged. The results for the resistor are for a 45 to 149 km depth range and for the conductor they are for a 118 to 270 km depth range.

Model	τ	τ_{iso}	rms	Cross-section through resistor				Cross-section through conductor			
				ρ_{yy}/ρ_{xx}		ρ_{zz}/ρ_{xx}		ρ_{yy}/ρ_{xx}		ρ_{zz}/ρ_{xx}	
				mean	median	mean	median	mean	median	mean	median
Isotropic	6	10^6	2.43	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00
Anisotropic	6	1.0	2.05	1.04	1.06	1.01	1.01	1.08	1.06	1.00	0.99
Anisotropic	6	0.1	1.99	1.91	2.02	1.15	1.22	2.28	1.97	1.00	1.05
Anisotropic	6	0.01	1.96	4.15	3.70	0.18	0.17	9.88	5.91	0.94	0.98

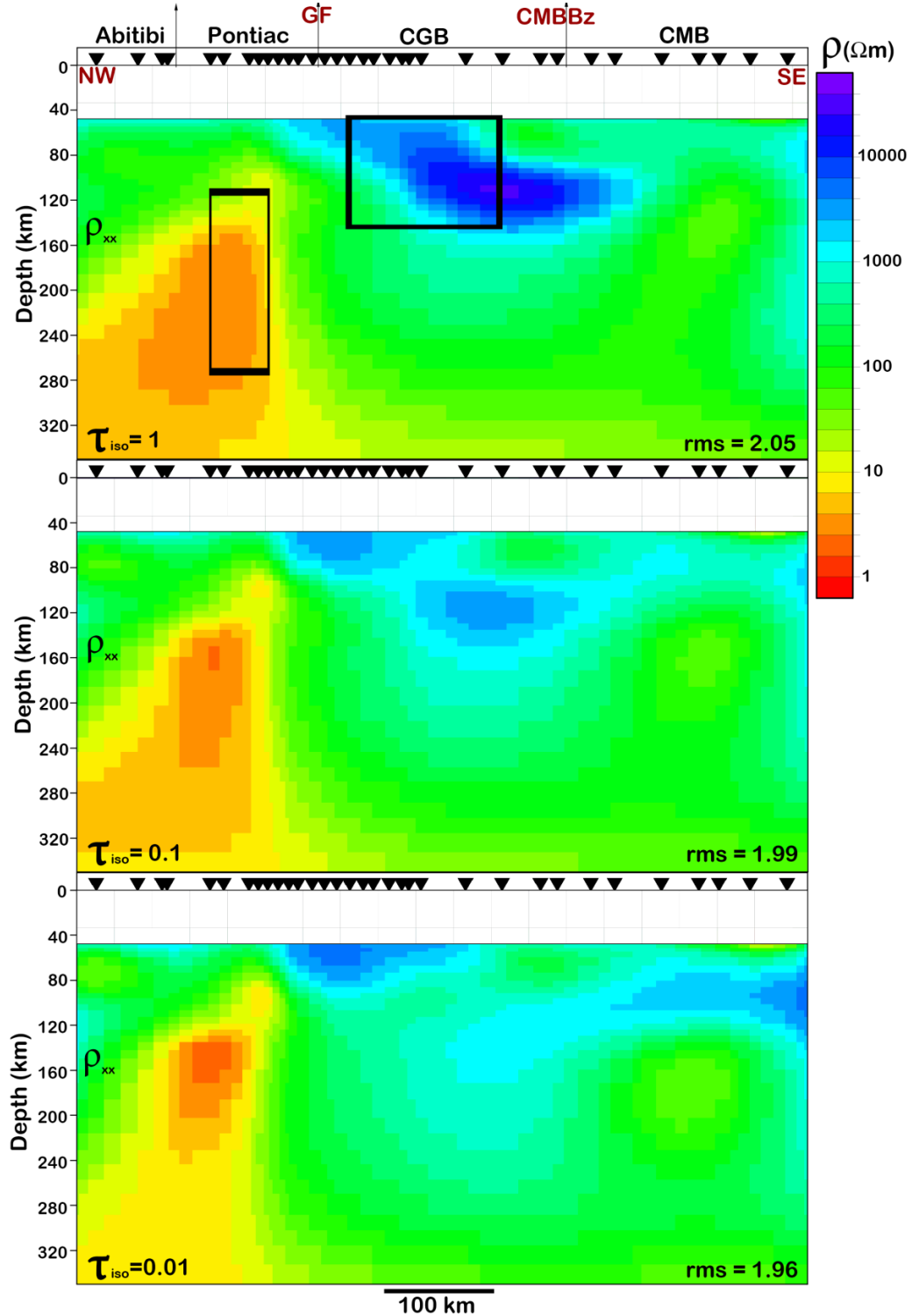
Figure 6.16. Resistivity profiles through the mantle resistor and mantle conductor for the isotropic and anisotropic models. Each panel compares the results from the isotropic model with the three components of resistivity from the anisotropic model: along-strike= ρ_{xx} , cross-strike= ρ_{yy} , and vertical= ρ_{zz} . The left panel is for $\tau_{iso} = 1$ and right panel is for $\tau_{iso} = 0.1$.



The anisotropy in the resistor and in the conductor was quantified using resistivity values extracted from a region of the anisotropic model extending from 45 to 150 km depth in the resistor and from 120 to 270 km depth in the conductor (Fig. 6.17). The resistivity values extending across each averaging volume at each available depth were combined by geometric averaging to provide a representative value for each depth. The ratio of the across-strike to along-strike resistivity ρ_{yy}/ρ_{xx} and the ratio of the vertical to along-strike resistivity ρ_{zz}/ρ_{xx} were computed at each depth. Finally, we computed the mean and median of these ratios as representative values of the anisotropy. The results are tabulated in Table 6.2. Within the resistor, there is an indication of very slight lateral anisotropy with higher conductivity along strike. However, the lateral anisotropy is less than about 5%. There is minimal vertical anisotropy with the mean and median ratios for ρ_{zz}/ρ_{xx} being very close to unity.

The results provide firm confirmation that there is minimal anisotropy in the resistive upper lithosphere at, and to the immediate south of, the Grenville Front. It is of note that the $\tau_{iso} = 1.0$ anisotropic model does indicate more strongly anisotropic zones at shallower depths within conductive zones of the crust of the CMB (not shown in Fig. 6.16). This result is in accord with Adetunji et al. (2014), who interpreted enhanced conductivity in these zones as being associated with graphitic rocks. Studies using the same anisotropy regularization factor of 1.0 in different tectonic settings have revealed significant anisotropy within the mantle in those locations (Baba et al., 2006; Le Pape et al., 2012). We can therefore conclude quite reliably that there is minimal resistivity anisotropy in the resistive lithosphere. The lateral anisotropy is significantly less than 10% or approximately two orders of magnitude smaller than the ratio of 1:15 determined

Figure 6.17. The along strike component of the anisotropic inversion models for Profile 1 with different value of τ_{iso} . The crustal section (upper 48 km) of the model is not shown. The rectangles show the parts of the model used to calculate the results shown in Table 6.2.



previously by Mareschal et al. (1995) using 1-D modelling. The results also show that there is minimal anisotropy within the lithospheric conductor where the anisotropy ratios are also less than 10% different from unity (Table 6.2).

To further test for the existence of anisotropy, models were obtained using even smaller isotropic regularization parameters. The resulting model for $\tau_{iso} = 0.1$ produces only a slight decrease in the misfit with an rms misfit of 1.99 (as opposed to rms = 2.05 for $\tau_{iso} = 1$). This decrease is extremely small considering the increase in the effective number of model parameters produced by the factor of 10 decrease in the anisotropy regularization factor and provides further support for the presence of minimal mantle lithospheric anisotropy in the study area. The results indicate the presence of lateral anisotropy in both the resistor and conductor with a ρ_{yy}/ρ_{xx} ratio of about 2 and minimal vertical anisotropy. However, examination of the resistivity model shows that the major features of the model have become less spatially coherent. In particular, the resistor is subdivided into parts and the conductor includes a small internal zone of increased conductivity. For this reason the model is not considered an optimal inversion solution.

For completeness, a very rough model was also obtained with an extreme value of $\tau_{iso} = 0.01$. This model has an rms misfit of 1.96, with minimal improvement over the $\tau_{iso} = 0.1$ model. In the rough resistivity model the mantle resistor is now divided into three parts and within the conductor there is a focusing of the enhanced conductivity into a small zone. It is not considered a reasonable resistivity model. The anisotropy is characterized by a maximum ρ_{yy}/ρ_{xx} ratio of about 4 in the resistor and minimal vertical anisotropy in the conductor. More significant vertical anisotropy occurs in the resistor

and lateral anisotropy in the conductor. Although the corresponding resistivity model is not considered reasonable, this model allows us to completely exclude the possibility of the presence of the previously interpreted 15:1 resistivity anisotropy within the resistive mantle lithosphere.

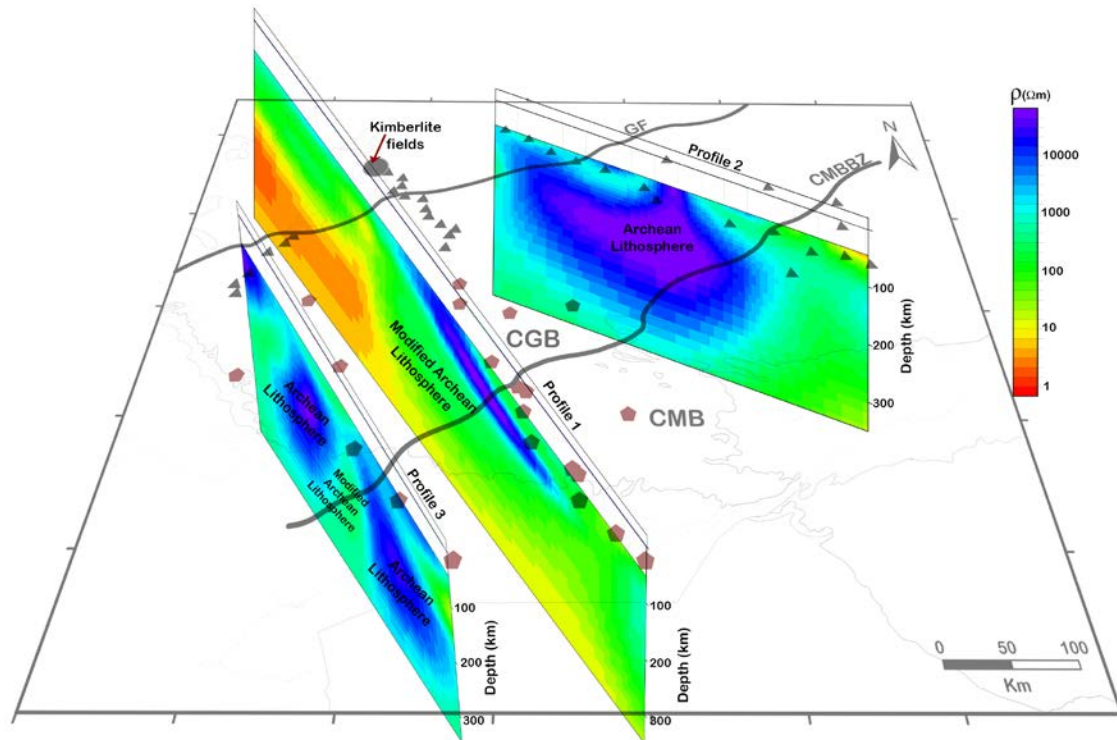
6.8 Discussion and Interpretation

6.8.1 Large scale resistivity structure

Figure 6.18 shows the resistivity model of the study area based on the upper lithospheric geoelectric strike along the three profiles. The profiles are arranged in a way that provides three-dimensional view of the resistivity structures. The most striking feature on the MT models is the southeast-dipping resistive lithosphere that extends from the base of the crust to a depth of ~280 km in profiles 2 and 3 and ~150 km in Profile 1. The resistor extends laterally from the Superior Province into the center of the CMB in the Grenville Province.

In earlier MT studies of Profile 1 (e.g., Mareschal et al., 1995; Adetunji et al., 2014), the resistor was interpreted to be an extension of Archean Superior mantle lithosphere beneath the northeastern Grenville Province. The resistor is associated with a zone of high seismic velocity between 50 and 150 km depth in which the S-wave velocity reaches values up to 10% higher than global models (Chen and Li, 2012) supporting the idea that it represented highly depleted continental lithosphere. Adetunji et al. (2014) compared the resistivity within the feature to 2-D resistivity models from southern Africa and showed that the modelled resistivity is comparable to that in lithosphere beneath the Archean Kapvaal craton.

Figure 6.18. Resistivity models of profiles 1 (middle), 2 (east) and 3 (west) projected onto the map of the study area. The models, with V.E=0.5 and depth extent of 350 km, are aligned with the MT data sites. GF=Grenville Front, CMBBZ= Central Metasedimentary Belt Boundary Zone. CGB=Central Gneiss Belt, CMB=Central Metasedimentary belt, FAB=Frontinac-Adirondack Belt. The crustal section (upper 48 km) of all the models is not shown.



The new 2-D resistivity models for Profile 2 and Profile 3 provided in the present study show the resistor extends along strike for at least 300 km and includes resistivity values exceeding 20,000 $\Omega.m$ over this whole region. The models from all three profiles also show that the resistor extends 300 km southeast beneath the Grenville Province from the Grenville Front to a point about 100 km southeast of the CMBBZ. Results from the present study thus show the resistor observed in Profile 1 results is a major regional feature and therefore provide additional strong support for the interpretation that the

resistor represents Superior lithosphere rather than a local feature within Grenville aged mantle lithosphere. These observations provide support for the idea that the Proterozoic rocks of the Grenville Province accreted onto the Archean basement of the Superior Province during the Grenville orogen and that this Archean lithospheric root provided the stable, cold and low-density basement for the orogen (Ludden and Hynes, 2000).

In all of the resistivity models, the southeast limit of the resistor has a wedge-like form and the southeastern part of the resistor is overlain by more conductive lithosphere. We interpret the conductive lithosphere to be Grenville-aged lithosphere thrust over the margin of the Superior lithosphere. Geoelectric strikes provide support for this interpretation. This geometry is most clearly illustrated on Profile 1 for which the upper surface of the resistor has a clear southeast dip. The resistive lithosphere is characterized by east-west oriented electrically conductive directions observed on all three profiles near the Grenville Front. This pattern extends southeast from the Grenville Front to a point about 100 km southeast of the Grenville Front where there is a transition to more SW-NE strikes characteristic of the southern part of the CGB and CMB (Fig. 6.6). If the east-west strikes are interpreted to indicate the presence of Superior lithosphere and the SW-NE strikes to indicate the presence of Grenville lithosphere they suggest that Grenville lithosphere has overthrust the southeastern 200 km of Superior lithosphere.

The present study has also provided important new information on the geometry of the base of the resistor, which varies significantly in the three profiles. As previously noted, on Profile 1 the base of the resistor occurs at ~150 km depth. On Profile 2, the base occurs at ~280 km along much of the profile. On Profile 3, the bases of each of the two resistive blocks are at around 280 km depth but there is a gap between the blocks, in

which either the base shallows significantly or the blocks are separated by more conductive mantle, whose nature is unknown at this time. The gap between these profiles on Profile 3 appears to be contiguous with the shallowing of the base of the resistor on Profile 1 (Fig. 6.19).

We interpret the observed 280 km depth of the base of the resistor on Profiles 2 and 3 to represent the base on the lithosphere in the region, and the relief in the lower part of the resistor to be caused by enhanced conductivity in the lower lithosphere. Adetunji et al. (2014) showed that the resistivity below the resistor on Profile 1 is less than that in some Proterozoic terranes in southern Africa. The enhanced conductivity beneath the resistor is therefore interpreted as refertilized lithosphere with the processes of refertilization introducing conducting elements into the lithospheric mantle, such as carbon (in various forms), sulphides and water. As shown by Adetunji et al. (2014) on Profile 1, the conductive zone is bounded to the northwest by a vertical anomaly located only a few tens of kilometres along strike from the Kirkland Lake and Cobalt kimberlite fields. We therefore interpret the conductive lower lithosphere on Profile 1 and the middle on Profile 3 as a zone that has been metasomatized by fluids associated with the Cretaceous kimberlitic magmatism. As pointed out by Selway (2014), refertilization of the mantle can lead to long-lived enhanced conductivity.

Figure 6.19 compares a horizontal slice through the resistivity models at ~180 km with the seismic tomography results of Chen and Li (2012) for the same depth. There is a very strong spatial correlation of the S-wave velocity and electrical resistivity over the study area. The maximum decrease in resistivity, of approximately three orders of magnitude is correlated with a 3% reduction in the S-wave velocity. The surface wave

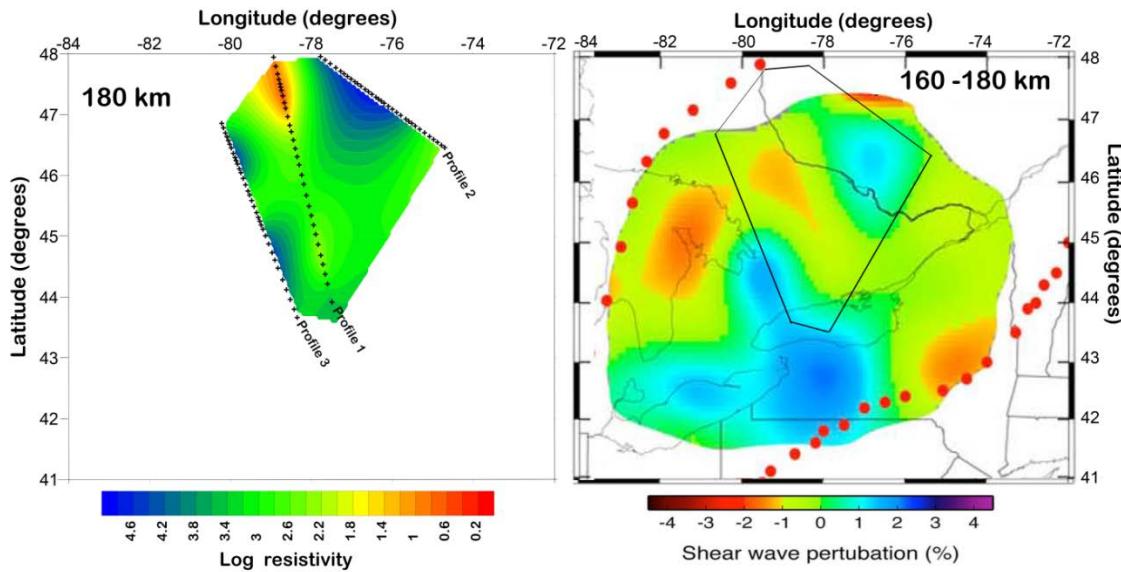
tomography models show similar lateral variations at depths extending from 120 km to 200 km and the resistivity models show similar patterns extending from 150 to 240 km (Fig. 6.18).

At a larger scale, the receiver function and SKS inversion studies of Abt et al. (2010), Yuan and Romanowicz (2010) and Yuan et al. (2011) suggest the existence of a two-layer lithosphere and a mid-lithospheric discontinuities (MLD) in areas near the present study area. The structural boundary occurs at ~150 km depth. In the seismic studies, the layer above the MLD is interpreted to be chemically depleted lithosphere and the lower layer to be a younger, less depleted, thermal boundary layer (Yuan and Romanowicz, 2010; Yuan et al., 2011). The two-layer lithosphere resolved in the resistivity model for Profile 1 is consistent with the seismic models, but the combined results from the three profiles suggests the existence of significant spatial variations in the MLD and that in places it may represent a localized feature.

The geophysical results from the present study provide strong evidence of refertilization of a large volume of the lower lithospheric mantle by the passage of fluids. Conductivity is enhanced over a depth range between 150 km and 280 km and over a horizontal spatial scale of several hundred kilometres. Veeraswamy et al. (2012) make a similar interpretation of relatively conductive lower lithosphere of a large region of Dharwar craton in India. In this case, the craton is interpreted to have been refertilized during its passage over the Reunion hotspot approximately 63 Ma ago.

The enhanced conductivity in the Grenville Province is observed at a depth of more than 150 km, beneath the graphite stability field (Kennedy and Kennedy, 1976). The absence of any surface gravity anomalies or surface topography mirroring the spatial

Figure 6.19. Comparison of electrical resistivity at 180 km depth with shear wave velocity perturbation (Chen and Li, 2012). In order to produce the resistivity results the models for profiles 1 to 3 were reprojected back to a line through the corresponding sites. Resistivity values extracted from the models at a depth close to 180 km were interpolated using a natural neighbor algorithm and then the results were contoured. The resistivity results show good spatial correlation with the S-wave velocity model with the lowest resistivity values at 180 km depth corresponding to $\sim -1.5\%$ velocity perturbation and the highest resistivity values at this depth corresponding to $\sim +1.5\%$ velocity perturbation.



variations in the base of the resistor suggests that the refertilization has produced minimal changes to the density and thermal properties of the lower lithosphere. We therefore interpret the enhanced conductivity to be associated with the presence of water in nominally anhydrous minerals (e.g., Karato, 1990; Jones, 1999; Hirth et al., 2000; Korja, 2007; Muller et al., 2009; Jones et al., 2013; Selway, 2014). It is possible to estimate the amount of water required to explain the observed resistivity. At a depth of 180 km, the average resistivity of the refertilized region at around 45.5° N on Profile 1 is $180 \Omega \cdot \text{m}$ (Fig. 6.19). This value is taken from the resistivity profiles shown in Fig. 6.16 and represents an average result at 180 km depth for the part of the profile beneath the southwest box shown in Fig. 6.17. The average crustal heat flow in the Grenville

Province is $41 \pm 10 \text{ mW.m}^{-2}$ (Pinet et al., 1991; Mareschal et al., 2000) and the corresponding lithospheric geotherm model of Artemieva (2006) indicates a temperature at 180 km depth of 1220°C . Noting the dominance of olivine in the lithospheric composition, and that the laboratory conductivity estimates for olivine lie in the middle of the results for ortho- and clinopyroxene and garnet, the olivine resistivity-temperature relationship is used to convert from resistivity to temperature. Using the relationship of Jones et al. (2012), the resistivity of $180 \text{ }\Omega\text{.m}$ and temperature of 1220°C indicate a water content of 50 wt ppm. There is some uncertainty in this estimate associated with the resolution of the resistivity value in the MT inversions and the spatial variations in the resistivity in the models, as well as with the estimate of the temperature and the resistivity-temperature relationship, but the result provides a reasonably firm indication of a water content of several tens of wt ppm.

Analyses of water in olivine from xenoliths found in kimberlites on the Kaapvaal Craton by Peslier et al. (2010) and Baptiste et al. (2012) show that the amount of water decreases with increasing pressure (depth), from about 100 wt ppm at 100 km to 10 wt ppm (essentially dry) at 200 km. This was taken by Peslier et al. (2010) as explanation for the longevity of cratons. However, very recent work on xenoliths from the Udachnaya kimberlite pipe in Siberia sourced from depths down to 200 km shows high water content in olivine, up to 300 wt ppm, that is depth independent (Doucet et al., 2013).

The estimate of water content at 180 km depth beneath the Grenville Province exceeds the values determined by Peslier et al. (2012) for the lower lithosphere of the Kapvaal Craton and is therefore consistent with the interpretation of refertilization of parts of the lower Grenville lithosphere. Farther to the northwest on Profile 1, the resistivity

decreases to lower values (e.g., as shown by Fig. 18 in Adetunji et al., 2014), and in the anomaly near the Kirkland Lake and Cobalt kimberlite fields it reaches values of $\sim 10 \Omega.m$. Explanation of the localized anomaly by the presence of water in the olivine, would require concentrations of several hundred wt ppm, comparable to those noted by Doucet et al. (2013).

6.8.2 Geoelectric dimensionality and directionality

The present study has extended earlier determinations of geoelectric dimensionality and directionality in the mantle of the study area by using a depth-based approach and by using a broader distribution of MT sites than Adetunji et al. (2014) and older analyses. The regional analysis was based on an *a priori* division of responses into those corresponding to an upper lithospheric layer (45-150 km) and those corresponding to deeper mantle depths (>200 km). Based on the observation of resistive blocks extending to 280 km depth on Profiles 2 and 3, we can now identify the lower zone as corresponding mainly to the lower lithospheric mantle.

In terms of dimensionality, phase tensor skew (Fig. 6.4) and average GB rms misfit values show that on all three profiles and at all depths the MT responses around the Grenville Front and in the northwestern CGB are compatible with the presence of 2-D resistivity structures. The spatial consistency of the most conductive direction at upper lithospheric depths provides further evidence of two-dimensionality in this depth range. These results justify the application of the 2-D inversion approach for modelling. Further to the southeast, in the southeastern part of the CGB and the CMB, higher phase tensor skew and GB misfits exceeding 1.5 suggest the presence of more complex 3-D structures, or of lower quality data.

Within the upper lithospheric mantle depth range, in areas near the Grenville Front and in the northern CGB, the conductive direction is approximately east-west. For profiles 1 and 2 the exact azimuth of the conductive direction is N85°E (e.g., Fig. 6.6). For Profile 3 the direction is E20°S but examination of GB misfit curves (Fig. 6.9) shows that a regional N85°E azimuth is associated with only slightly higher misfit than the E20°S azimuth. The MT data are thus consistent with a common lithospheric conductive direction over a ~300 km (east-west) by 200 km (north-south) region of the northern Grenville Province. This result is in accord with the earlier analysis of Mareschal et al. (1995) who determined a regional azimuth of N80°E along Profile 1 (Fig. 6.1). However, the tensor decomposition applied in the present study shows this azimuth extends significantly farther to the east than interpreted by Mareschal et al. (1995); to at least as far as Profile 2.

Comparison of the depth-based conductive directions (Figs. 6.4, 6.5 and 6.6) and 2-D resistivity models (Figs. 6.2, 6.13 and 6.14) shows that, on all three profiles, the east-west azimuth is associated with resistive mantle lithosphere. For example, the high depth-resolution results in Fig. 6.6 show that for Profile 1 this azimuth is very well defined in the 45-100 km and 100-150 km depth range, but is more variable at depths of 150-200 km i.e., below the resistor. As shown by the anisotropic inversions the conductive direction is caused by large-scale geoelectric structure rather than resistivity anisotropy. Taken together, the observations suggest that the conductive directions near the Grenville Front are caused by structures in and near the northwestern margin of the mantle lithospheric resistor.

There are two possible contributions to the observed strike azimuths. The first contribution is from internal macroscopic structures within the resistive lithosphere. The observed east-west strikes are sub-parallel to the Archean fabric of the Superior craton and so are consistent with this source. The second possible contribution is from the margin of the resistive block itself. For Profile 1, Adetunji et al. (2014) interpreted this margin as being associated with refertilization related to the Mesozoic kimberlitic magmatism in the Kirkland Lake and Cobalt kimberlite fields. However, even in this case, the azimuth of the feature may reflect the original Archean fabric, as the refertilization is interpreted to have occurred through a zone of weakness in the lithosphere associated with the original formation of the Abitibi greenstone belt (e.g., Faure et al., 2011). It is important to note that the Cretaceous activity appears to have not affected the lattice preferred orientation interpreted to be responsible for the SKS fast direction near the GF. However, this result is consistent with Frederiksen et al. (2006) who show that there is a change in the shear wave splitting parameters between the GF and areas farther to the north thus suggesting limited resetting near the GF itself.

The depth-based MT tensor analysis done in the present study has provided new results on the strike azimuth at >200 km depth in the Grenville Province in southern Ontario. The results define a regional strike direction of N65°E ($\pm 5^\circ$) for the whole study area. As noted above, based on the large-scale geoelectric structure, we interpret this depth range to include a significant contribution from the lower mantle lithosphere, so the observed regional strike azimuth may reflect either the lower lithospheric or deeper mantle structures.

As shown in Fig. 6.6 the mean conductive direction is oblique to both the mean seismic anisotropy in the study region, as defined by the SKS fast direction, and to the GPS based APM direction. For sites south of 46°N, there is some local correlation of the SKS fast direction and MT strikes determined in period-based analyses (Frederiksen et al., 2006) but this correlation has been interpreted as being due to lithospheric sources.

Available APM models in this region include the GPS model of Larson et al. (1997) with APM direction 277°, the no-net-rotation reconstruction model of Demets et al. (1990) with direction of 279°, and the hotspot-referenced model of Gripp and Gordon (2002) with direction 248°. The deep geoelectric azimuth, derived from this study (equivalent to 248°) agrees excellently well with the Gripp and Gordon (2002) APM result. However, this APM model appears to provide a poor estimate of the true plate motion in the study area as, unlike the alternative other models, its APM direction is oblique to the Great Meteor hotspot trail. Taken together, the results suggest that the most conductive direction is oblique to the hotspot trail, the APM, and to the SKS fast directions. Similar obliquity has been observed in other Precambrian terrains in the world (e.g., Simpson, 2001; Hamilton et al., 2006).

Based on its obliquity to the present APM, we interpret the deeper strike direction to be associated with lower lithospheric structure. The corresponding MT responses can be fitted with isotropic 2-D models (Figs. 6.2, 6.12, and 6.13) so the strikes may be explained by the deeper features in these models including the geometry of the base of the resistive blocks on profiles 2 and 3. The lower lithosphere in the study area is interpreted to have been metasomatized by fluids associated with the Cretaceous kimberlitic magmatism. We therefore hypothesize that the observed deep MT strike

azimuth was established at the time of this process and potentially records the plate motion at that time.

The results of the large-scale shear wave study of Yuan et al. (2011) indicate a depth-dependent fast axis within the lithosphere across much of North America. For locations near the present study area, the results (Fig. 8 of Yuan et al., 2011) show an approximately east-west SKS fast direction at 70 km depth, a north-south SKS fast-direction at 150 km depth, and a southwest-northeast SKS fast-direction at 250 km depth. The MT strike direction determined for the upper lithosphere is in good agreement with the seismic fast direction in the upper lithosphere. However, the MT strike directions do not show evidence of the north-south azimuth determined by Yuan et al. (2011) for the lower lithosphere. This difference may relate to the localized spatial variations of the two-layer lithosphere suggested by the MT resistivity models.

6.8.3 Resistivity Anisotropy

The results of the present study along with the 2-D isotropic inversion results of Adetunji et al. (2014) indicate that the MT response anisotropy near the Grenville Front is due to large-scale geoelectric structure rather than to resistivity anisotropy, as had previously been interpreted. The availability of induction arrows from a broader region than in the earlier studies (e.g., Kellett et al., 1992) shows that at periods longer than about 30 s there is a dominant alignment of the real and imaginary arrows in a direction at high angle to the geoelectric fast direction. This observation is consistent with the presence of macroscopic resistivity structures and the 2-D inversion models demonstrate the presence of such structures.

The 2-D anisotropic modelling in this study shows that models obtained with $\tau_{\text{iso}} = 1.0$ reveal minimum anisotropy at mantle depths. The results show that within the lithospheric resistor, the resistivity differs in the two horizontal directions by less than 6%. Other studies using the same value of the anisotropic regularization factor have revealed significant mantle resistivity anisotropy (Baba et al., 2006; Le Pape et al., 2012) and in the present study the inversions done using this factor revealed significant anisotropy at shallower depths in the CMB crust. Consideration of results obtained with lower anisotropic regularization factors indicated in the extreme case of $\tau_{\text{iso}} = 0.01$, for which the actual resistivity models become geologically infeasible, the maximum horizontal anisotropy in the resistive lithosphere is a factor of 5, well less than the factor of 15 determined in earlier 1-D modelling studies. We conclude that there is no significant resistivity anisotropy in the upper lithospheric mantle. The results also provide no evidence for vertical anisotropy within the lithospheric conductor.

The presence of shearing in the lithospheric mantle beneath the GF is supported by the seismic anisotropy results and the alignment of the SKS fast direction with the GF (Senechal et al., 1996; Ji et al., 1996; Frederiksen et al., 2006). However, the new MT results show that the Ji et al. (1996) model, in which the obliquity between the MT and seismic strikes is interpreted as being due to dextral shearing, is not valid. If dextral shearing occurred, it appears to have not caused any significant shape preferred electrical resistivity anisotropy as interpreted by Ji et al. (1996). The upper lithospheric mantle MT strikes are interpreted to be related to large-scale structures (rather than to shape-preferred orientation of mantle minerals), that are in turn interpreted to be related to the fabric of the Superior craton. The obliquity between the seismic and MT responses arises

because the Superior fabric is oblique to the Grenville Front. The new MT results provide some support for aspects of the earlier interpretation of Mareschal et al. (1995) who interpreted strike directions in the southern Superior as being related to metasomatism of mantle roots of major Archean shear zones.

The present study does not exclude the presence of resistivity anisotropy farther to the north in the Superior Province (e.g., Kellett et al., 1992, Mareschal et al., 1995; Ji et al., 1996). However, it is clear that the interpretation of resistivity anisotropy in this area should also be reassessed using modern 2-D isotropic and anisotropic inversion methods and that it is inappropriate to now base such an interpretation on 1-D inversion only.

6.9 Conclusions

The structure of the lithospheric mantle beneath the Grenville Front and the Grenville Province in Ontario, Canada is investigated using the MT method. Data from a total of 56 sites on three profiles with northwest-southeast azimuths, spanning a north-south distance of 500 km and an east-west distance of 650 km, are analyzed. Modern depth-based tensor decomposition techniques along with isotropic and anisotropic 2-D inversions are applied. The paper updates the work of Kellett et al. (1992), Kellett et al. (1994), Zhang et al. (1995), Mareschal et al. (1995), Senechal et al. (1996), Ji et al. (1996), Frederiksen et al. (2006), and Adetunji et al. (2014). A major objective of this paper was to determine whether the azimuthal variation in the MT responses, defined by earlier studies, is related to electrical anisotropy or to larger-scale geological structure.

Depth-based dimensionality analysis of the data from all three profiles shows that at lithospheric mantle depths the MT responses from sites at the Grenville Front and in

the northwestern CGB are compatible with the presence of 2-D resistivity structures. This result is further supported by spatial consistency of the geoelectric strike azimuth. Sites in the southeastern part of the CGB and the CMB exhibit more significant 3-D structures. The depth-based analysis of the MT responses suggest that the lithosphere can be subdivided into upper (45-150 km) and deeper (>200 km) lithospheric mantle layers with dominant regional strike azimuths of N85°E ($\pm 5^\circ$) and N65°E ($\pm 5^\circ$) respectively. The result for the upper mantle lithosphere agrees well with the lithospheric strike azimuth of N80°E defined for Grenville Front region in the earlier studies, and provides a larger scale geoelectric strike azimuth that can be used for 2-D modeling the Grenville Province across the study area.

Isotropic 2-D resistivity models of the study were obtained by inversion of distortion-free MT responses from each profile. The resistivity models for all of profiles include a large-scale resistive (>20,000 $\Omega\cdot\text{m}$) region extending to the southeast from the Grenville Front. It extends 300 km southeast beneath the Grenville Province from the Grenville Front to a point about 100 km southeast of the CMBBZ and based on the results for the three profiles it extends for at least 600 km along-strike. The resistor is interpreted to form part of the Superior lithosphere, the Archean lithospheric root which provided the stable, cold and low density basement for the Grenville orogen, which supports the earlier interpretation by Ludden and Hynes (2010). Based on the resistivity models the observed N85°E strike angle in the upper lithospheric layer is interpreted to correspond to the Archean geoelectric fabric of the Superior craton.

The top of the resistor dips to the southeast in the central profile and has an overall southeast-dipping trend in the other two profiles. It is overlain by rocks that are in

most locations significantly more conductive and have an associated N45°E geoelectric strike azimuth. This part of the model is interpreted to be Grenvillian-aged lithosphere thrust over the southeast margin of the Superior lithosphere.

The base of the lithospheric resistor has significant relief. The feature extends to a maximum depth of ~280 km in the east and west of the study area on profiles 2 and 3 respectively. However, on the central Profile 1 it extends to only ~150 km. There is also a gap in the lower part of the resistor in the central part of western Profile 3 that appears to be contiguous with the shallowing of the resistor on Profile 1. We interpret the observed 280 km depth to be the base on the lithosphere in the region, and the relief on the lower surface of the resistor to be caused by enhanced conductivity in the lower lithosphere.

The enhanced conductivity in the lower lithosphere exhibits high spatial correlation with decreased shear wave velocity. On Profile 1 the enhanced conductivity is bounded to the northwest by a sub-vertical resistivity anomaly located near the Kirkland Lake and Cobalt kimberlite fields. The connected region of the conductive mantle along profiles 1 and 3 is therefore interpreted as a zone that has been metasomatized by fluids associated with Cretaceous kimberlite magmatism. The absence of any surface gravity anomalies or surface topography mirroring the spatial variations in the base of the lithospheric resistor suggest that the refertilization has produced minimal changes to the density and thermal properties of the lower lithosphere. The enhanced conductivity is therefore interpreted to be associated with the presence of water in nominally anhydrous minerals. Based on the observed resistivity values of 180 $\Omega\cdot\text{m}$ at 180 km depth, and estimates of the temperature at this depth of 1220 °C, the relationship

between resistivity and water content of Jones et al. (2012) indicates a water content of approximately 50 wt ppm. Higher water content is required to explain the localized resistivity anomaly observed at 180 km depth in the area of the Kirkland Lake and Cobalt kimberlite fields.

A new regional strike direction of N65°E ($\pm 5^\circ$), based on the conductive direction, is defined in this study for the lower lithosphere of the Grenville Province. This conductive direction is oblique to the SKS fast direction and the Great Meteor hotspot trail or the GPS-based APM direction. The obliquity suggests that the strike azimuth is associated with fossil resistivity structures in the lower lithosphere rather than present day plate motion. Based on the interpreted Cretaceous metasomatism of the lower lithosphere, we hypothesize that the measured strike azimuth was established in the Cretaceous and potentially records the plate motion at that time.

Careful examination of the induction arrows at long periods, from an area larger than in previous studies, shows significant alignment of the real and imaginary arrows in a direction at high angle to the geoelectric strike. This observation is consistent with the presence of macroscopic resistivity structures, which is confirmed by the 2-D inversion models. The results of 2-D anisotropic modelling indicate that the mantle lithosphere beneath the Grenville Front and Grenville Province is almost electrically isotropic (i.e., anisotropy is not required to fit the 2-D lithospheric resistivity model). For anisotropic models obtained with $\tau_{iso} = 1.0$, the maximum anisotropy in both resistive and conductive regions of the lithosphere is less than 10%. This result contrasts with the factor of 15 resistivity anisotropy determined in earlier 1-D modelling studies. We can therefore conclude that the MT response anisotropy near the Grenville Front is related to large-

scale geoelectric structure rather than resistivity anisotropy, as interpreted in previous studies. A similar interpretation can be made for the observed azimuthal variation of MT responses further southeast within the Grenville Province. We also suggest that earlier interpretation the existence of dextra shearing in the upper lithospheric mantle should be reviewed. Our results do not exclude the existence of resistivity anisotropy to the northwest within the Superior Province but we suggest a re-examination of this possibility with contemporary 2-D isotropic and anisotropic inversion methods.

6.10 Acknowledgments

The authors thank the POLARIS southern Ontario MT team (J. Wenham, M. Serzu, X. Ma, E. Gowan, H. Sealey, T. Pacha, B. Eade, J. McCutcheon, I. Asudeh, C. Andrews, C. Samson, and P. Fernberg) for MT data collection and landowners in the region for site access. Comments from A. Frederiksen led to significant improvement of the manuscript. Some figures were produced using Wessel and Smith's GMT package. Funding for POLARIS has been provided by industry, the Canada Foundation for Innovation, provincial agencies from Ontario and British Columbia, the Federal Economic Development Initiative for Northern Ontario, Natural Resources Canada, and NSERC (Natural Sciences and Engineering Research Council of Canada). The present research was supported by an NSERC Discovery Grant (105748) to Ian Ferguson.

6.11 References Cited

- Abt, D.L., Fischer, K.M., French, S.W., Ford, H.A., Huaiyu, Y., Romanowicz, B., 2010. North American lithospheric discontinuity structure imaged by Ps and Sp receiver functions. *J. Geophys. Res.* 115, B09301, doi:10.1029/2009JB006914.
- Adetunji, A.Q., Ferguson, I.J., Jones, A.G., 2014. Crustal and lithospheric scale structures of the Precambrian Superior-Grenville margin, *Tectonophysics* 614, 146-169, <http://dx.doi.org/10.1016/j.tecto.2013.12.008>.
- Artemieva, I.M., 2006. Global 1 degrees x 1 degrees thermal model TC1 for the continental lithosphere: Implications for lithosphere secular evolution, *Tectonophysics* 416, 245-277.
- Baba, K., Tarits, P., Chave, A.D., Evans, R.L., Hirth, G., Mackie, R. L., 2006. Electrical structure beneath the northern MELT line on the East Pacific Rise at 1545'S. *Geophys. Res. Lett.* 33, 1-5.
- Bailey, R.C., Groom, R.W., 1987. Decomposition of the magnetotelluric impedance tensor which is useful in the presence of channeling. in 57th Annual international Society of Exploration Geophysicists meeting and exposition, pp. 154-156, Tulsa, OK.
- Bank, C., Eaton, D., Aktas, K., 2005. Moho topography and lower crustal density in Southern Ontario from linearized gravity inversion. AGU, New Orleans, poster, T43D-04.
- Baptiste, V., Tommasi, A., Demouchy, S., 2012. Deformation and hydration of the lithospheric mantle beneath the Kaapvaal craton, South Africa. *Lithos* 149, 31-50.
- Boerner, D. E., Kurtz, R. D., Craven, J. A., 2000. A summary of electromagnetic studies on the Abitibi-Grenville transect. *Can. J. Earth Sciences* 37, 427-437.
- Booker, J.R., 2014. Magnetotelluric Phase Tensor Evolution. *Surv. Geophys.* 35, 7-40, doi: 10.1007/s10712-013-9234-2.
- Bostick, F. X., 1977. A simple almost exact method of MT analysis. Workshop on Electrical Methods in Geothermal Exploration. US. Geol. Surv. Contract No. 14080001-8-359.
- Caldwell, T.G., Bibby, H.M., Brown, C., 2004. The magnetotelluric phase tensor. *Geophys. J. Int.* 158, 457-469.
- Carr, S. D., Easton, R. M., Jamieson, R. A., Culshaw, N. G., 2000. Geologic transect across the Grenville orogen of Ontario and New York, *Can. J. Earth Sci.*, 37,193-216.
- Chave, A. D., 2014. On the statistics of magnetotelluric rotational invariants. *Geophys. J. Int.* 196, 111-130, doi:10.1093/gji/ggt366.
- Chave, A.D., Jones, A.G., 1997. Electric and magnetic field galvanic distortion decomposition of BC87 data. *J. Geomagn. Geoelectr* 49, 767-789.
- Chave, A.D., Jones, A.G. 2012. *The Magnetotelluric Method - Theory and Practice*. Publ. by Cambridge University Press, Cambridge, UK., pp. 570, ISBN: 9780521819275.
- Chen, C. -W., Li, A., 2012. Shear wave structure in the Grenville Province beneath the lower Great Lakes region from Rayleigh wave tomography. *J. Geophys. Res.* 117, B01303.
- Crough, S.T., 1981. Mesozoic hot spot epeirogeny in eastern North America. *Geology* 9, 2-6.

- Culshaw, N. G., Davidson, A., Nadeau, L., 1983. Structural subdivisions of the Grenville Province in the Parry Sound-Algonquin region, Ontario, in: Current Research Part B, Geological Survey of Canada, paper 83-1B, pp. 243-252.
- Culshaw, N. G., Jamieson, R. A., Ketchum, J. W. F., Wodicka, N., Corrigan, D., Reynolds, P. H., 1997. Transect across the northwestern Grenville orogen, Georgian Bay, Ontario: Polystage convergence and extension in the lower orogenic crust. *Tectonics* 16, 966-982.
- Davidson, A., 1998. Questions of correlation across the Grenville Front east of Sudbury, Ontario; in: Current Research 1998-C; Geological Survey of Canada. Paper 98-IC, pp. 145-154.
- Davidson, A., Culshaw, N. G., Nadeau, L., 1982. A tectono-metamorphic framework for part of the Grenville Province, Parry Sound region, Ontario, in: Current Research Part A, Geological Survey of Canada, paper 82-1A, pp. 175-190.
- Davis, W.J., Jones, A.G., Bleeker, W., Grutter, H., 2003. Lithosphere development in the Slave craton: a linked crustal and mantle perspective. *Lithos*, 71, 575-589.
- Demets, C., Gordon, R., Argus, D., Stein, S., 1990. Current plate motions. *Geophys. J. Int.* 101, 425-478.
- Doucet, L.S., Peslier, A.H., Ionov, D.A., Brandon, A.D., Golovin, A.V., Ashchepkov, I.V., 2013. High water contents in the Siberian cratonic mantle: an FTIR study of Udachnaya peridotite xenoliths, in Fall AGU, San Francisco, CA, USA.
- Du Frane, W. L., Roberts, J.J., Toffelmier, D.A., Tyburczy, J.A., 2005. Anisotropy of electrical conductivity in dry olivine, *Geophys. Res. Lett.* 32, doi: 10.1029/2005gl023879.
- Easton, R. M., 1992. The Grenville Province and the Proterozoic history of Central and Southern Ontario, in: *Geology of Ontario*, Ontario Geological Survey, Special Volume 4, 713 – 904.
- Eaton, D. W., Frederiksen, A., 2007. Seismic evidence for convection driven motion of the North American plate. *Nature* 446, 428–431.
- Eaton, D. W., Adams, J., Asudeh, I., Atkinson, G. M., Bostock, M. G., Cassidy, J. F., Ferguson, I. J., Samson, C., Snyder, D. B., Tiampo, K. F., Unsworth, M. J., 2005. Geophysical arrays to investigate lithosphere and earthquake hazards in Canada. *EOS Trans AGU* 86, 169-173.
- Eaton, D. W., Dineva, S., Mereu, R., 2006. Crustal thickness and Vp/Vs variations in the Grenville orogen Ontario, Canada from analysis of teleseismic receiver functions. *Tectonophysics* 420, 223-238.
- Eaton, D., Frederiksen, A.W., Miong, S.K., 2004. Shear-wave splitting observations in the lower Great Lakes region: Evidence for regional anisotropic domains and keel-modified asthenospheric flow. *Geophys. Res. Lett.* 31, L07610
- Eisel, M., Bahr, K., 1993. Electrical anisotropy in the lower crust of British-Columbia - an interpretation of a magnetotelluric profile after tensor decomposition. *J. Geomag. Geoelec.* 45, 1115-1126.
- Evans, M.S., Kendall, J.-M., Willemann, R.J., 2006. Automated SKS splitting and upper-mantle anisotropy beneath Canadian seismic stations. *Geophys. J. int.* 165, 931-942

- Evans, R.L., Hirth, G., Baba, K., Forsyth, D., Chave, A. Mackie, R., 2005. Geophysical controls from the MELT area for compositional controls on oceanic plates. *Nature* 437, 249–252.
- Evans, R. L., Jones, A. G., Garcia, X., Muller, M., Hamilton, M., Evans, S., Fourie, C. J. S., Spratt, J., Webb, S., Jelsma, H., Hutchins, D., 2011. Electrical lithosphere beneath the Kaapvaal craton, southern Africa. *J. Geophys. Res.* 116, B04105.
- Everett, M. E., 2005. What do electromagnetic induction responses measure? *The Leading Edge* 24, 154 – 157.
- Faure S., Godey, S., Fallara, F., Trepanier, S., 2011. Seismic architecture of the Archean North American mantle and its relationship to diamondiferous kimberlite fields. *Econ. Geol.* 106, 223-240.
- Ferguson, I.J., Jones, A.G., Chave, A.D., 2012. Case histories and geological applications, in: *The Magnetotelluric Method*: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 480-544.
- Frederiksen, A.W., Ferguson, I. J., Eaton, D., Miong, S. K., Gowan, E., 2006. Mantle fabric at multiple scales across an Archean-Proterozoic boundary, Grenville Front, Canada. *Phys. Earth Planet. Int.* 158, 240-263.
- Frederiksen, A. W., Miong, S.-K., Darbyshire, F. A., Eaton, D. W., Rondenay, S., Sol, S., 2007. Lithospheric Variations Across the Superior Province, Ontario, Canada: Evidence from Tomography and Shear-Wave Splitting. *J. Geophys. Res.*, 112, doi: 10.1029/2006JB004861.
- Griffin, W. L., O'Reilly, S. Y., Doyle, B. J., Pearson, N. J., Coopersmith, H., Kivi, K., Malkovets, V., Pokhilenko, N., 2004. Lithosphere mapping beneath the North American plate. *Lithos* 77, 873-922.
- Gripp, A.E., Gordon, R. G., 2002. Young tracks of hotspots and current plate velocities. *Geophys. J. Int.* 150, 321-361.
- Groom, R. W., Bailey, R. C., 1989. Decomposition of magnetotelluric impedance tensors in the presence of local three-dimensional galvanic distortion. *J. Geophys. Res.* 94, 1913-1925.
- Groom, R. W., Bailey, R. C., 1991. Analytic investigations of the effects of near-surface three-dimensional galvanic scatterers on MT tensor decompositions. *Geophysics* 56, 496 – 518.
- Haggart, M. J., Jemieson, R. A., Reynold, P. H., Krogh, T. E., Beaumont, C., Culshaw, N. G., 1993. Last gasp of the Grenville Orogeny: thermochronology of the Grenville Front Tectonic Zone near Killarney, Ontario. *J. Geology* 101, 575-589.
- Hamilton, M. P., Jones, A. G., Evans, R. L., Evans, S., Fourie, C. J. S., Garcia, X., Mountford, A., Spratt, J. E., the SAMTEX Team, 2006. Electrical anisotropy of South African lithosphere compared with seismic anisotropy from shear-wave splitting analysis. *Phys. Earth Planet. Int.* 158, 226–239.
- Hansen, P.C., 1992. Analysis of discrete ill-posed problems by means of the L-curve. *Siam Review* 34, 561-580.
- Heise, W., Pous, J., 2001. Effects of anisotropy on the two-dimensional inversion procedure. *Geophys. J. Int.* 147, 610-621. doi:10.1046/j.0956-540x.2001.01560.x

- Heise, W., Pous, J., 2003. Anomalous phases exceeding 90° in magnetotellurics: anisotropic model studies and a field example. *Geophys. J. Int.* 155, 308-318. doi:10.1046/j.1365-246X.2003.02050.x
- Heise, W., Caldwell, T.G., Bibby, H. M., Brown, C., 2006. Anisotropy and phase splits in magnetotellurics. *Phys. Earth Planet. Int.* 158, 107- 121.
- Hirth, G., Evans, R. L., Chave, A. D., 2000. Comparison of continental and oceanic mantle electrical conductivity: Is the Archean lithosphere dry? *Geochem. Geophys. Geosyst.* 1, 1030, doi:10.1029/2000GC000048.
- Hynes, A., 1994. Gravity, flexure, and the deep structure of the Grenville Front, eastern Quebec and Labrador. *Can. J. Earth Sci.* 31, 1002–1011.
- Ji, S., Rondenay, S., Mareschal, M., Senechal, G., 1996. Obliquity between seismic and electrical anisotropies as a potential indicator of movement sense for ductile shear zones in the upper mantle. *Geology* 24, 1033-1036.
- Jones, A. G., 1983a. On the equivalence of the "Niblett" and "Bostick" transformations in the Magnetotelluric Method. *J. Geophysics* 53, 72-73.
- Jones, A.G., 1983b. The problem of current channelling: A critical review. *Surveys in Geophysics* 6, 79-122.
- Jones, A. G., 1988. Static shift of magnetotelluric data and its removal in a sedimentary basin environment. *Geophysics* 53, 967-978.
- Jones, A. G., 1992. Electrical conductivity of the continental lower crust, in: Fountain, D. M., Arculus, R. J., Kay, R. W. (Eds.), *Continental Lower Crust*, Elsevier, Amsterdam, pp. 81-143.
- Jones, A. G., 1999. Imaging the continental upper mantle using electromagnetic methods. *Lithos* 48, 57-80.
- Jones, A. G., 2006. Electromagnetic interrogation of the anisotropic Earth: Looking into the Earth with polarized spectacles. *Phys. Earth Planet. Int.* 158, 281-291.
- Jones, A. G., 2012. Distortion of magnetotelluric data: its identification and removal, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 219-295.
- Jones, A. G., Ledo, J., Ferguson, I.J., Craven, J.A., Unsworth, M.J., Chouteau, M., Spratt, J.E., 2013. The electrical resistivity of Canada's lithosphere and correlation with other parameters: Contributions from LITHOPROBE and other programmes. *Can. J. Earth Sciences* (accepted January 2014).
- Jones, A.G., Spratt, J., 2002. A simple method for deriving the uniform field MT responses in auroral zones. *Earth Planets Space* 54, 443-450.
- Karato, S., 1990. The role of hydrogen in the electrical conductivity of the upper mantle. *Nature* 347, 272–273.
- Kellett, R.L., Barnes, A. E., Rive, M., 1994. The deep structure of the Grenville Front: a new perspective from western Quebec. *Can. J. Earth Sci.* 31, 282–292.
- Kellett, R.L., Mareschal, M., Kurtz, R. D., 1992. A model of lower crustal electrical anisotropy for the Pontiac Subprovince of the Canadian Shield. *Geophys. J. Int.* 111, 141-150.
- Kennedy, C.S., Kennedy, G.C., 1976. Equilibrium boundary between graphite and diamond. *J. Geophys. Res.* 81, 2467-2470.
- Korja, T., 2007. How is the European lithosphere imaged by magnetotellurics? *Surv. Geophys.* 28, 239 – 272.

- Kurtz, R.D., Craven, J.A., Niblett, E.R., Stevens, R.A., 1993. The conductivity of the crust and mantle beneath the Kapuskasing Uplift: electrical anisotropy in the upper mantle. *Geophys. J. Int.* 113, 483–498.
- Kurtz, R.D., Niblett, E.R., Craven, J.A., Stevens, R.A., Macnae, J.C., 1988. Electromagnetic studies over the Kapuskasing Structural Zone, in: *Lithoprobe Kapuskasing Transect*, pp. 167–176, Lithoprobe Report 4.
- Larson, K.M., Freymueller, J. T., Philipson, S., 1997. Global plate velocities from the Global Positioning System. *J. Geophys. Res.* 102, 9961–9981.
- Le Pape, F., Jones, A.G., Vozar, J., Wenbo, W., 2012. Penetration of crustal melt beyond the Kunlun Fault into northern Tibet. *Nature Geoscience* 5, 330–335.
- Ledo, J., Queralt, P., Marti, A., Jones, A. G., 2002. Two-dimensional interpretation of three-dimensional magnetotelluric data: an example of limitations and resolution. *Geophys. J. Int.* 150, 127–139.
- Ludden, J., Hynes, A., 2000. The Lithoprobe Abitibi-Grenville transect: two billion years of crust formation and recycling in the Precambrian Shield of Canada. *Can. J. Earth Sci.* 37, 459–476.
- Mackie, R. L., 2002. User Manual and Software Documentation for Two-Dimensional Inversion of Magnetotelluric data, Anisotropy Version 6.7. GSY-USA, Inc., San Francisco, USA.
- Mackie, R., Rieven, S., Rodi, W., 1997. Users manual and software documentation for two-dimensional inversion of magnetotelluric data. San Francisco, GSY-USA, 14pp.
- Mackwell, S., Kohlstedt, D., 1990. Diffusion of hydrogen in olivine: Implications for water in the mantle. *J. Geophys. Res.* 95, 5079–5088.
- Mareschal, J.C., Jaupart, C., Gariépy, C., Cheng, L.Z., Guillou-Frottier, L., Bienfait, G., Lapointe, R., 2000. Heat flow and deep thermal structure near the southeast edge of the Canadian Shield. *Can. J. Earth Sci.* 37, 399–414.
- Mareschal, M., Kellett, R. L., Kurtz, R. D., Ludden, J. N., Ji, S., Bailey, R. C., 1995. Archean cratonic roots mantle shear zones and deep electrical anisotropy. *Nature London* 375, 134–137.
- Mareschal, M., Kurtz, R.D., Chouteau, M., Chakridi, R., 1991. A magnetotelluric survey on Manitoulin Island and Bruce Peninsula along GLIMPCE seismic line J: black shales mask the Grenville Front. *Geophys. J. Int.* 105, 173–183.
- Marti, A., 2014. The role of electrical anisotropy in magnetotelluric responses: From modelling and dimensionality analysis to inversion and interpretation. *Surv. Geophys.* 57, 179–218, doi: 10.1007/s10712-013-9233-3.
- Matsuno, T., Seama, N., Evans, R.L., Chave, A.D., Baba, K., White, A., Goto, T., Heinson, G., Boren, G., Yoneda, A., Utada, H., 2010. Upper mantle electrical resistivity structure beneath the central Mariana subduction system. *Geochem. Geophys. Geosyst.* 11, Q09003, doi:10.1029/2010GC003101.
- McNeice, G. W., Jones, A. G., 2001. Multisite, multifrequency tensor decomposition of magnetotelluric data. *Geophysics* 56, 158–173.
- Meyer, H. O. A., Waldman, M. A., Garwood, B. L., 1994. Mantle xenoliths from kimberlite near Kirkland Lake, Ontario. *Can. Min.* 32, 295–306.
- Miensopust, M. P., Jones, A. G., Muller, M. R., Garcia, X., Evans, R. L., 2011. Lithospheric structures and Precambrian terrane boundaries in northeastern

- Botswana revealed through magnetotelluric profiling as part of the Southern African Magnetotelluric Experiment. *J. Geophys. Res.* 116, B02401.
- Miensopust, M.P., Jones, A.G., 2011. Artefacts of isotropic inversion applied to magnetotelluric data from an anisotropic Earth. *Geophys. J. Int.* 187, 677-689.
- Muller, M. R., Jones, A. G., Evans, R. L., Gru tter, H. S., Hatton, C., Garcia, X., Hamilton, M. P., Miensopust, M. P., Cole, P., Ngwisanyi, T., Hutchins, D., Fourie, C. J., Jelsma, H. A., Evans, S. F., Aravanis, T., Pettit, W., Webb, S. J., Wasborg, J., The SAMTEX Team, 2009. Lithospheric structure, evolution and diamond prospectivity of the Rehoboth Terrane and the western Kaapvaal Craton, southern Africa: Constraints from broadband magnetotellurics. *Lithos* 112S, 93-105.
- Niblett, E. R., Sayn-Wittgenstein, C., 1960. Variation of electrical conductivity with depth by the magneto-telluric method. *Geophysics* 25, 998-1008.
- Nieuwenhuis, G., Unsworth, M.J., Pana, D., Craven, J.A., 2013. Three dimensional resistivity structure of Southern Alberta, Canada: Implications for Pre-Cambrian tectonics. *Geophys. J. Int.* (submitted).
- Ogawa, Y., Jones, A. G., Unsworth, M. J., Booker, J. R., Lu, X., Craven, J., Roberts, B., Parmalee, J., Farquharson, C., 1996. Deep Electrical Conductivity Structures of the Appalachian Orogen in the Southeastern US. *Geophys. Res. Lett.* 23, 1597–1600.
- Park, S. K., Mackie, R. J., 1997. Crustal structure at Nanga Parbat, northern Pakistan, from magnetotelluric soundings. *Geophys. Res. Lett.* 24, 2415-2418.
- Parker, R.L., 1980. The inverse problem of electromagnetic induction: existence and construction of solutions based on incomplete data. *J. Geophys. Res.* 85, 4421-4425.
- Parkinson, W.D., 1962. The influence of continents and oceans on geomagnetic variations. *Geophys. J.R. Astr. Soc.* 6, 441–449.
- Peslier, A.H., Woodland, A.B., Bell, D.R., Lazarov, M., 2010. Olivine water contents in the continental lithosphere and the longevity of cratons. *Nature* 467, 78-81.
- Pek, J., Verner, T., 1997. Finite-difference modelling of magnetotelluric fields in two-dimensional anisotropic media. *Geophys. J. Int.* 128, 505-521. doi:10.1111/j.1365-246X.1997.tb05314.x
- Pinet, C., Jaupart, C., Mareschal, J-C., Gari py, C., Bienfait, G., Lapointe, R., 1991. Heat flow and structure of the lithosphere in the eastern Canadian Shield. *J. Geophys. Res.* 96, 19941 – 19963.
- Poe, B.T., Romano, C., Nestola, F., Smyth, J.R., 2010. Electrical conductivity anisotropy of dry and hydrous olivine at 8 GPa. *Phys. Earth, Planet. Inter.* 181, 103-111.
- Pommier, A., 2014. Interpretation of magnetotelluric results using laboratory measurements. *Surv. Geophys.* 35, 41-84.
- Pous, J., Ledo, J., Marcuello, A., Queralt, P., 1997. On the Resolution of the Darai Limestones by Two-Dimensional MT Forward Modelling. *J. Geomag. Geoelect.* 49, 817-825.
- Rasmussen, T.M., 1988. Magnetotellurics in Southwestern Sweden: evidence for electrical anisotropy in the lower crust? *J. Geophys. J.R. Astron. Soc.* 89, 799-820.

- Rivers, T., 1997. Lithotectonic elements of the Grenville Province: review and tectonic implications. *Precamb. Res.* 86, 117-154.
- Rivers, T., Culshaw, N., Hynes, A., Indares, A., Jamieson, R., Martignole, J., 2012. The Grenville orogen- A Post-Lithoprobe Perspective in: *Tectonic Styles in Canada: The Lithoprobe Perspective*, pp 97-236, eds. Percival, J.A., Cook, F.A., Clowes, R. M., Geological Association of Canada, Special paper 49.
- Rivers, T., Martignole, J., Gower, C. F., Davidson, A., 1989. New tectonic subdivisions of the Grenville Province, southeast Canadian Shield. *Tectonics* 8, 63-84.
- Rodi, W. L., Mackie, R. L., 2001. Nonlinear conjugate gradients algorithm for 2D magnetotelluric inversion. *Geophysics* 66, 174-187.
- Rodi, W. L., Mackie, R. L., 2012. The inverse problem, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 347-414.
- Rondenay, S., Bostock, M. G., Hearn, T. M., White, D. J., Ellis, R. M., 2000. Lithospheric assembly and modification of SE Canadian Shield: Abitibi-Grenville teleseismic experiment. *J. Geophys. Res.* 105, 13735-13754.
- Roy, B., Mereu, R. F., 2000. Applications of seismic pattern recognition and gravity inversion techniques to obtain enhanced subsurface images of the Earth's crust under the Central Metasedimentary Belt, Grenville Province, Ontario. *Geophys. J. Int.* 143, 735-751.
- Schäfer, A., Hout, L., Brasse, H., Hoffmann, N., EMTESZ Working Group, 2011. The North German Conductivity Anomaly revisited. *Geophys. J. Int.* 187, 85–98.
- Schmoldt, J.-P., 2011. Multidimensional isotropic and anisotropic investigation of the Tajo Basin subsurface: A novel anisotropic inversion approach for subsurface cases with oblique geoelectric strike directions. PhD thesis, National University of Ireland, Galway.
- Schmoldt, J.-P., Jones, A.G., 2013. A novel anisotropic inversion approach for magnetotelluric data from subsurfaces with orthogonal geoelectric strike directions, *Geophys. J. Int.* 195, 1576-1593.
- Schock, R.N., Duba, A.G., Shankland, T.J., 1989. Electrical conduction in olivine. *J. Geophys. Res.* 94, 5829–5839.
- Schwarz, G., 1990. Electrical conductivity of the earth's crust and upper mantle. *Surv. Geophys.* 11, 133 – 161.
- Selway, K. 2014. On the Causes of Electrical Conductivity Anomalies in Tectonically Stable Lithosphere. *Surv. Geophys.* 57, 219-257, doi 10.1007/s10712-013-9235-1
- Sénéchal, G., Rondenay, S., Mareschal, M., Guilbert, J., Poupinet, G., 1996. Seismic and electrical anisotropies in the lithosphere across the Grenville Front, Canada. *Geophys. Res. Lett.* 23, 2255-2258.
- Simpson, F., Tommasi, A., 2005. Hydrogen diffusivity and electrical anisotropy of a peridotite mantle. *Geophys. J. Int.* 160, 1092-1102. doi:10.1111/j.1365-246X.2005.02563.x
- Simpson, F., 2001. Resistance to mantle flow inferred from the electromagnetic strike of the Australian upper mantle. *Nature* 412, 632-633.
- Sleep, N. H., 1990. Montereyan hotspot track: a long-lived mantle plume. *J. Geophys. Res.* 95, 21,983–21,990.

- Spratt, J.E., Jones, A.G., Jackson, V.A., Collins, L., Avdeeva, A., 2009. Lithospheric geometry of the Wopmay orogen from a Slave craton to Bear Province magnetotelluric transect. *J. Geophys. Res.* 114, B01101.
- Veeraswamy, K.K., Abdul Azeez, K.K., Gupta, A.K., Babu, N., Harinarayana, T., 2012. Deep geo-electric structure of the Dhawar craton (India) inferred from magnetotelluric studies. *Electromagnetic Induction Workshop, Darwin*, poster 1.3-22p.
- Vozoff, K., 1991. The magnetotelluric method, in: Nabighian, M. N. (Ed.), *Electromagnetic methods in applied geophysics*, Society of Exploration Geophysicist, pp. 641 – 711.
- Wannamaker P. E., 1999. Affordable magnetotellurics: interpretation in natural environments, in: Oristaglio, M., Spies, B., (Eds.), *Three-dimensional electromagnetics*, pp. 349–379, SEG.
- Wannamaker, P.E., 2005. Anisotropy versus heterogeneity in continental solid earth electromagnetic studies: fundamental response characteristics and implications for physicochemical state. *Surv. Geophys.* 26, 733 – 765.
- Wannamaker, P. E., Chave, A. D., Booker, J. R., Jones, A.G., Filloux, J. H., Ogawa, Y., Unsworth, M., Tarits, P., Evans, R., 1996. Magnetotelluric Experiment Probes Deep Physical State of Southeastern U.S., *EOS Trans. AGU*, 329, 332–333.
- Weidelt, P., 1999. 3D conductivity models: implications of electrical anisotropy, in: Oristaglio, M., Spies, B., (eds.), *Three-dimensional electromagnetics*, pp. 119–137, SEG.
- Wessel, P., Smith, W.H.F., 1991. Free software help map and display data. *EOS Transactions AGU* 72, 441.
- White, D.J., Easton, R.M., Culshaw, N.G., Milkereit, B., Forsyth, D.A., Carr, S., Green, A.G., Davidson, A., 1994. Seismic images of the Grenville Orogen in Ontario. *Can. J. Earth Sci.* 31, 293–307.
- White, D.J., Forsyth, D. A., Asudeh, I. A., Carr, S. D., Wu, H., Easton, R. M., Mereu, R. F., 2000. A seismic-based cross-section of the Grenville Orogen in Southern Ontario and western Quebec. *Can. J. Earth Sci.* 37, 183-192.
- Wiese, H., 1962. Geomagnetische tiefensondierung. Teil II: Die Streichrichtung der Untergrundstrukturen des elektrischen Widerstandes, erschlossen aus geomagnetischen variationen. *Geofis. Pura et Appl.* 52: 83–103.
- Wu, X., Ferguson, I.J., Jones, A.G., 2005. Geoelectric structure of the Proterozoic Wopmay Orogen and adjacent terranes, Northwest Territories, Canada. *Can. J. Earth Sci.* 42, 955-981.
- Yuan, H., Romanowicz, B., 2010. Lithospheric layering in the North American craton. *Nature*, 466, 1063-1069, doi:10.1038/nature09332.
- Yuan, H., Romanowicz, B., Fischer, K.M., Abt, D., 2011. 3-D shear wave radially and azimuthally anisotropic velocity model of the North American upper mantle. *Geophys. J. Int.*, 184, 1237-1260, doi:10.1111/j.1365-246X.2010.04901.x
- Zhang, P., Chouteau, M., Mareschal, M., Kurtz, R. D., Hubert, C., 1995. High frequency magnetotelluric investigation of crustal structure in north-central Abitibi, Quebec, Canada. *Geophys. J. Int.* 120, 406-418.

Zhang, P., Pedersen, L. B., Mareschal, M., Chouteau, M., 1993. Channelling contribution to tipper vectors: a magnetic equivalent to electrical distortion. *Geophys. J. Int.* 113,693-700.

Chapter 7

Magnetotelluric imaging of the lower Great Lakes region, Canada

This chapter has been submitted for publication in the *Geophysical Journal International* in Jan. 2014. I was responsible for all the data processing, analysis, modeling, isotropic inversions as well as the writing of the first draft of the manuscript. My advisor (I. J. Ferguson) reviewed the draft of the manuscript.

Adetunji, A.Q., Ferguson, I.J., Jones, A. G., 2014. Magnetotelluric imaging of the lower Great Lakes region, Canada. *Geophysical Journal International* (Submitted in January 2014)

7.1 Abstract

The electrical resistivity structure beneath the lower Great Lakes region in southern Ontario is investigated with 15 MT sites from the POLARIS project. The MT resolution in the study area is influenced strongly by a 400 m thick sequence of Upper Ordovician shale units with conductance range of 20-23 S within the sedimentary rocks. These rocks severely limit the resolution of the MT method in the crust and the deep mantle lithosphere. A regional geoelectric strike azimuth of N8°W is determined for the upper lithospheric mantle of this region. Two-dimensional inversion reveals a lithospheric resistor that extends from the base of the crust to at least 150 km depth in the lithospheric mantle. The resistor is interpreted to be an extension of the Archean Superior lithosphere beneath the Grenville Province. This feature provided the basement for the Grenville orogenic events. A lithospheric conductor is resolved between 100 and 150 km depth beneath the Central Metasedimentary Belt. It confirms the presence of very conductive lithosphere beneath Central Metasedimentary Belt in the Grenville Province. The enhanced conductivity could either be attributed to grain boundary graphite films, associated with the Cretaceous kimberlitic magmatic process, or water and carbon, introduced into the mantle during the pre-Grenvillian tectonism.

7.2 Introduction

As part of the POLARIS (Portable Observatories for Lithosphere Analysis and Research Investigating Seismicity) project, this study uses the magnetotelluric (MT) method to image the electrical resistivity structure beneath the Lower Great lakes region in southern Ontario. POLARIS is a Canadian university-government-industry geophysical consortium that focuses on the investigation of the structure and dynamics of the Earth's lithosphere, assessment of earthquake, and geomagnetic hazards across Canada, using the MT method and earthquake seismology (Eaton et al., 2005).

Electrical resistivity of geological formations is the major physical parameter of interest in MT studies (Jones 1999). In the crust, this parameter is sensitive to presence and mobility of aqueous fluids, graphite, interconnected grains of metallic oxides and sulphides (e.g., Jones and Craven, 1990; Katsube and Mareschal, 1993; Hyndman et al., 1993; Glover and Vine, 1995; Li et al., 2003). Sources of enhanced conductivity in the mantle include graphite, water, partial melt, temperature and hydrogen (e.g., Karato, 1990; Mibe et al., 1998; Jones, 1999; Hirth et al., 2000; Ducea and Park, 2000; Constable, 2006; Korja, 2007; Yoshino et al., 2008; Muller et al., 2009; Selway, 2014). Interpretation of source of enhanced conductivity must always be based on local geological and physical conditions (Schwarz, 1990; Korja 2007, Selway, 2014).

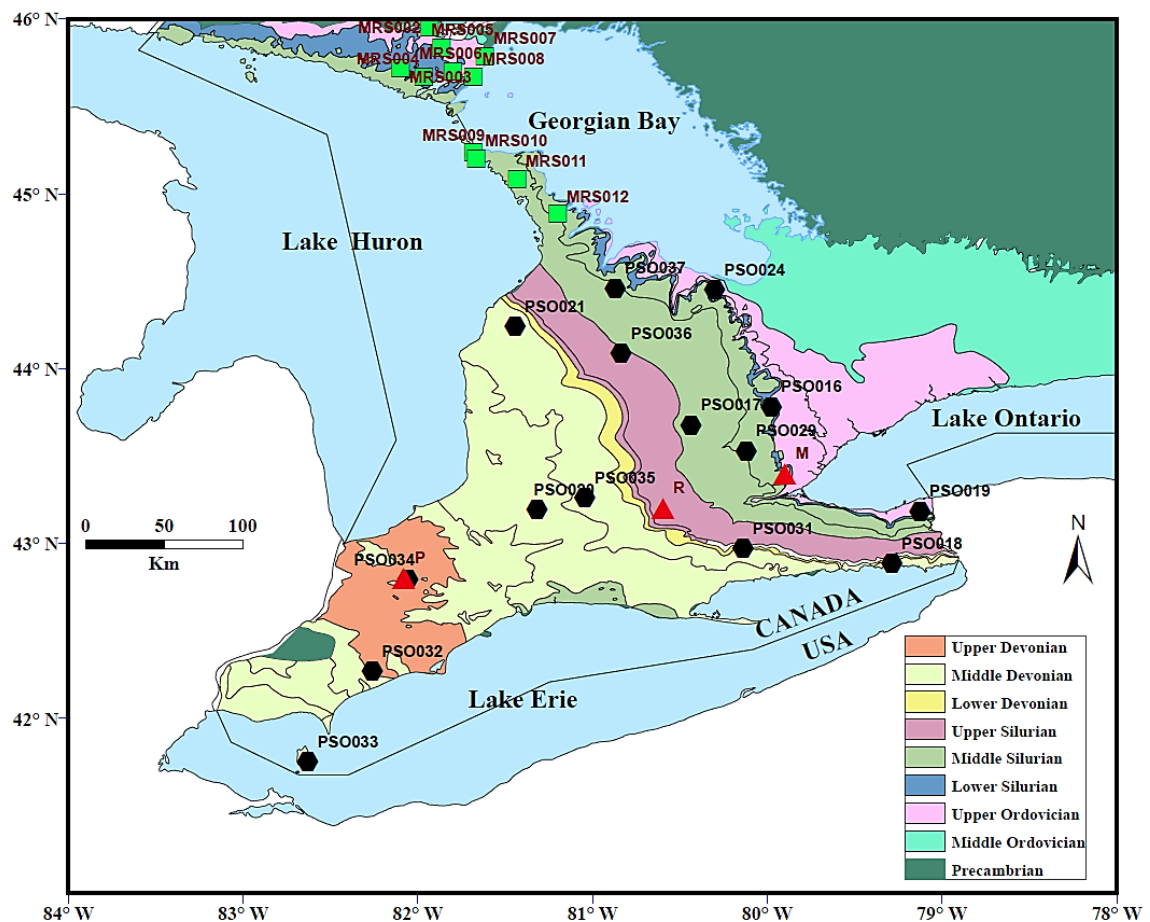
The primary response derived from MT data is the complex impedance tensor. The individual terms of the tensor may be examined in terms of the corresponding apparent resistivity and impedance phase, both of which give insight into underlying resistivity structure (Vozoff, 1991; Chave and Jones, 2012). The depth of penetration of MT signals depends on both the period of the signal and the local resistivity. The

subsurface resistivity structure can be 1-D or horizontally-layered, 2-D with a geoelectric strike in which the resistivity is invariant, or it may have a more complex 3-D structure. For 1-D and 2-D structures, the impedance phase responses are related to the gradient of apparent resistivity with log period (e.g., Parker and Booker, 1996; Weidelt and Chave, 2012). The phase response is more robust than the apparent resistivity and it can be recovered exactly from responses that have been galvanically distorted by near-surface heterogeneities using tensor decomposition methods. The apparent resistivity may still be affected by a period-independent static shift after tensor decomposition (Jones, 1988). Prior to the rotation of the coordinate system, the off-diagonal terms of the MT impedance tensor correspond to electric current flow in the north direction (XY-component) and east-direction (YX-component). For 2-D structures, the impedance tensor may be rotated into coordinates orthogonal to the geoelectric strike and separated into the transverse electric (TE) mode in which the electric field is parallel to the geoelectric strike, and the transverse magnetic (TM) mode in which the electric field is perpendicular to the strike.

The MT data in this study are modeled and interpreted in detail for the first time. A preliminary period-based analysis of the geoelectric strike of data is provided in Frederiksen et al. (2006). The data examined consist of impedance tensor estimates from 15 MT sites distributed systematically across the Paleozoic sedimentary rocks in the study area (Fig. 7.1). Four of these sites contain long period (LMT) data while the rest are broadband MT. The data were collected using relatively short duration recording times (two nights for broadband MT data and one-week for LMT data) with the intent of providing a synoptic characterization of the earth resistivity structure over a large region

within a several month period. Because of the high level of urbanization in the area the MT responses at most of the sites are relatively noisy and required moderate to severe editing prior to their use in subsequent analyses. The MT analysis included examination of phase tensors (Caldwell et al., 2004), application of depth-dependent tensor decomposition (Groom and Bailey, 1989, 1991; McNeice and Jones, 2001), 1-D inversion to define resistivity structures in the sedimentary basin rocks, and 2-D inversion to define resistivity structures in the underlying lithosphere of the Grenville Province.

Figure 7.1. Geological map of the study area showing depositional cycles and the location of the POLARIS MT sites. Shapefile was obtained from Geological Survey of Canada.



Previous crustal scale electromagnetic studies in this area are limited. Mareschal et al. (1991) identified Ordovician black shale at the base of the Whitby Formation (northwest of the study area) with conductivity large enough to mask the resolution of deeper crustal structures. Boerner et al. (2000) review MT surveys conducted within the LITHOPROBE project to the northeast of the present study area and Frederiksen et al. (2006) and Adetunji et al. (2014a, 2014b) describe recent analyses of LITHOPROBE and POLARIS MT data from the same area. Duncan et al. (1980) as well as Gomez-Trevino and Edwards (1983) investigated the electrical resistivity structure of the sedimentary rocks within the study area using controlled source electromagnetic (CSEM) methods.

Based on gravity and magnetic studies of Easton and Carter (1995) and Wallach et al. (1998), the Central Gneiss Belt underlying the study area, is described as being structurally simple and is not associated with any major structural feature. The Central Metasedimentary Belt Boundary Zone has been described as a structurally deformed (O'Dowd et al., 2004) southeast-dipping (White et al., 2000; Adetunji et al., 2014a) feature that extends throughout the crust (Adetunji et al., 2014a). The seismically defined Moho depth for this region is 40-45 km (Cullota et al., 1990; Eaton et al., 2006; Zhang and Frederiksen, 2013).

7.3 Geologic and Tectonic Settings

The Paleozoic sedimentary rocks of the study area consist of a 1500 to 2000 m thick succession of Cambrian to Devonian aged sandstones, shales, evaporites and carbonates (Fisher et al., 1988; Sanford, 1993). The rocks dip at low angles and are generally undeformed but may be locally folded and faulted (Sanford, 1993).

The Phanerozoic sedimentary rocks of southern Ontario were deposited unconformably over the Proterozoic Grenville Province in two sedimentary basins: the Allegheny and Michigan basins (Fig. 7.2a). These basins are separated by a basement topographic feature called the Algonquin Arch. The Michigan Basin is a roughly circular intracratonic basin formed due to cratonic margin tectonic processes (Howell and Van der Pluijm, 1990, 1999; Johnson et al., 1992). The basin contains inward-dipping Paleozoic strata deposited during rapid subsidence in during the Cambrian to Silurian period (Sloss, 1988) and a veneer of Jurassic sedimentary rocks (Fisher et al., 1988). At the center of the basin the maximum thickness of sedimentary rocks is more than 4.5 km thickness. The Allegheny Basin is a foreland basin formed by orogenic processes at the craton margin and filled mainly with upper Cambrian to Upper Devonian sedimentary rocks. It reaches a thickness of 7 km. Numerous authors (e.g.; Sleep, 1971; Sleep and Snell, 1976; Sleep and Nunn, 1980; Nunn and Sleep, 1984; Klein and Hsui, 1987; Howell and Van der Pluijm, 1990, 1999; Johnson et al., 1992) have provided detailed interpretations of the tectonic processes responsible for the formation of these basins. The Algonquin Arch, sometimes referred to as the Western St. Lawrence Platform, originated as a passive continental margin due to continental separation in the early Cambrian (Kumarapeli, 1985). The stratigraphic succession in this platform is made up of sediments deposited in shallow epicontinental seas (Johnson et al., 1992; Sanford, 1993).

The Paleozoic stratigraphy within the study area has been described as relatively simple, flat lying, and continuous over long distances (e.g., Johnson et al., 1992; Mazurek, 2004; Armstrong and Carter, 2006; Intera, 2008, 2009; Gartner, 2008). Fig. 7.2b shows a geological cross-section that extends across the Algonquin Arch and the

stratigraphic section of southern Ontario is shown in Fig. 7.3. The altered Precambrian granitic gneiss basement rocks are unconformably overlain by the sandstone and dolostones of the Cambrian rocks. The middle Ordovician is characterized by shales and carbonate rocks of the Black River, Trenton and Collingwood formations. The Upper Ordovician is typified by shales of the Blue Mountain, Georgian Bay, and Queenston formations. The lower Silurian contains shales of Cataract Formation while the middle Silurian is made up of dolomite and shale of Guelph, Goat Island, Gasport, Rochester, and Reynales formations. The Upper Silurian is typified by alternating carbonate, shale and evaporates of the Salina Group and Bass Islands Formation. The Lower Devonian rocks are mainly dolostones of Bois Blanc Formation, and the Middle Devonian consists of the dolomite, limestone and shale units of Dundee Formation, Detroit River and Hamilton groups. The Upper Devonian is mainly characterized by the shales and sandstone of the Kettle Point, Bedford and Berea formations (Fig. 7.3; Mazurek, 2004).

In the study area, the Proterozoic Grenville Province is interpreted to be bounded to the west by the Proterozoic Mazatzal Province (Whitmeyer and Karlstrom, 2007). In this area the Precambrian basement is covered by the Michigan Basin so there may be some uncertainty in this interpretation. The Mazatzal Province is believed to have formed by in continental margin volcanic arcs and back arc-related supracrustal successions (Whitmeyer and Karlstrom, 2007). As shown in Fig. 7.2a, the Grenville Front Tectonic Zone (GFTZ) and the Central Metasedimentary Belt Boundary Zone (CMBBZ) are traceable from their northeast surface exposure to beneath the sedimentary rocks (Easton, 1992; Easton and Carter, 1995; Hynes and Rivers, 2010). Over most of the study area, the sedimentary rocks are underlain by rocks of the Central Gneiss Belt (CGB) of the

Grenville Province. The CGB represents the reworked margin of the pre-Grenvillian Laurentian craton (Carr et al., 2000; Rivers et al., 2012). It is characterized by layered felsic gneisses and amphibolites with upper amphibolite to granulite facies plutonic rocks as well as related supracrustal rocks (Easton, 1992). The Grenvillian orogeny affected most of the Archean to Mesoproterozoic crustal assemblages within the CGB (Easton, 1986). Several ductile shear zones that separate distinctive lithotectonic domains and

Figure 7.2. The upper panel (a) shows the tectonic elements of the study area (modified from Mazurek, 2004). The depths shown are the basin isopachs. The lower panel (b) shows composite stratigraphic succession of the study area (modified from Russell and Gale, 1982) for cross-section AA' shown in the upper panel. The white unit at the top represents the Quaternary sediments.

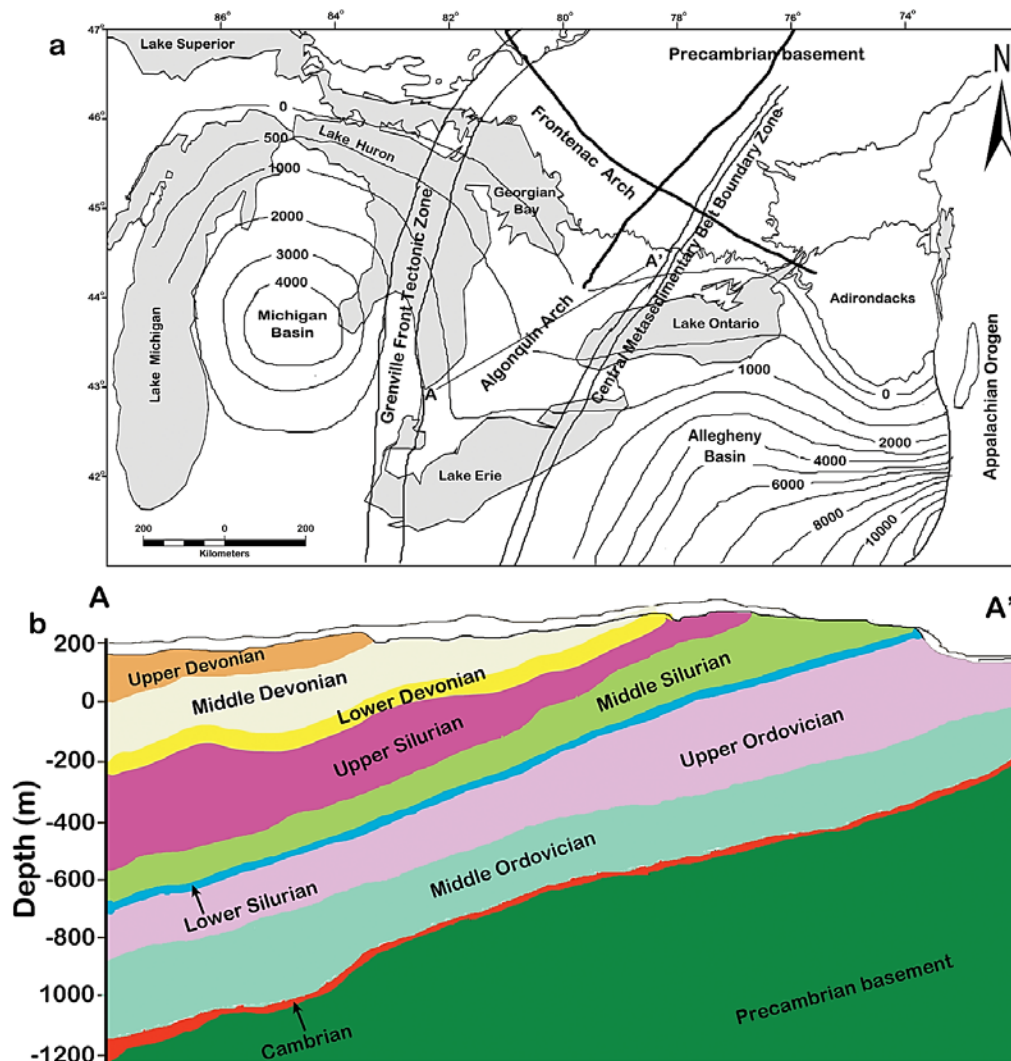
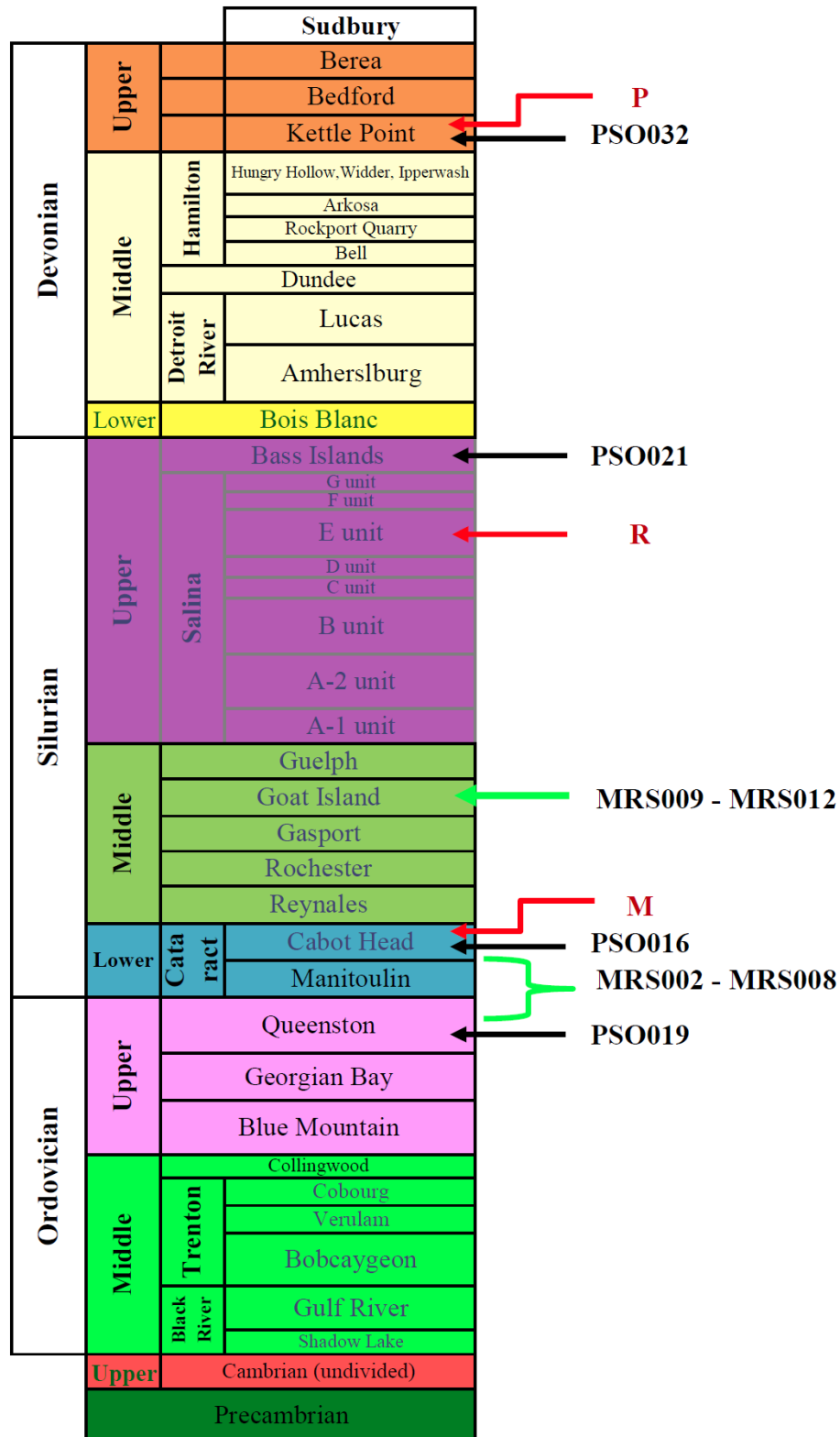


Figure 7.3. Stratigraphic section of Sedimentary rocks showing all the formations (redrawn from Mazurek, 2004) and the locations of the POLARIS MT sites, Mareschal et al. (1991) MT sites and Gomez-Trevino and Edwards (1983) CSEM sites.

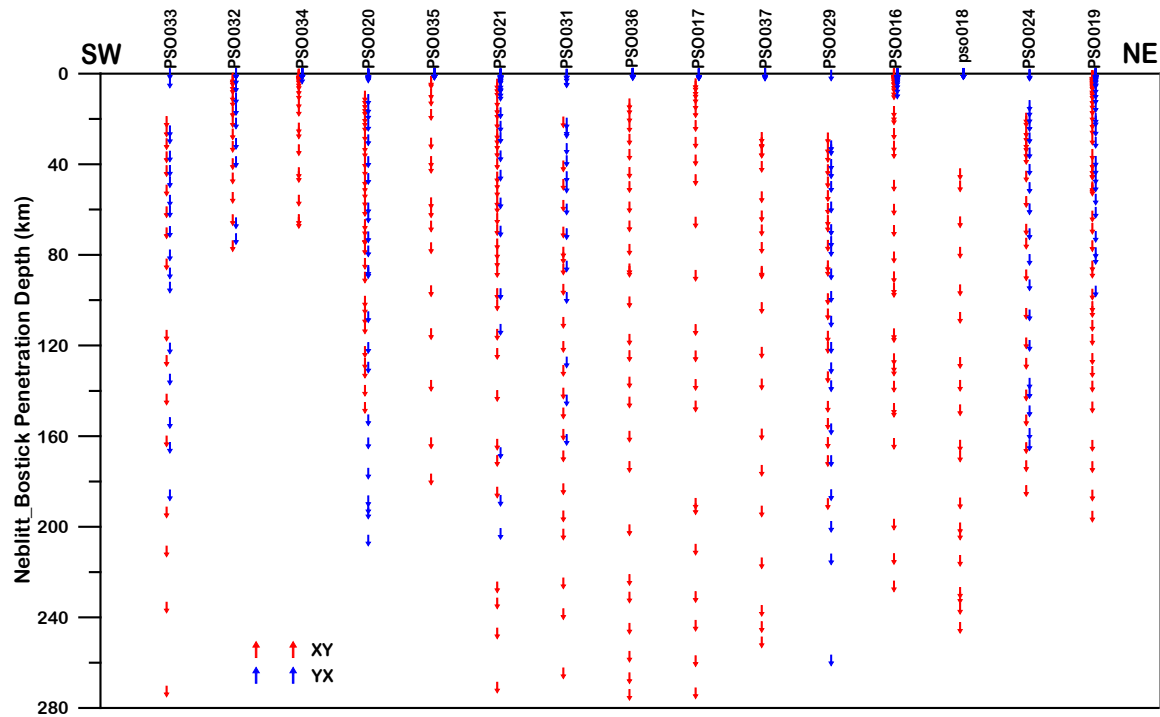


sub-domains in the CGB show an indication of northwest-directed low-angle ductile thrusting (Davidson, 1986). The GFTZ is a southeast-dipping crustal-scale shear zone of amphibolite-facies rocks; its boundaries are gradational and are deformed by visible cataclastic zones, which extend into the adjacent Southern and Superior provinces (Easton, 1992). It marks the northwestern boundary of the Grenville orogen and thrusting along this shear zone occurred late in the Grenvillian orogeny, (Haggart et al., 1993; Krogh, 1994). The CMBBZ, which separates the CGB from the Central Metasedimentary Belt (CMB), is a composite ductile shear zone characterized by alternating crystalline thrust sheets enveloped by anastomosing higher strain zones of porphyroclastic to migmatitic tectonites and marble tectonic mélange (Hanmer and McEachern, 1992; Rivers et al., 2012).

7.4 Depth of Investigation

The MT analysis commenced with careful editing of the MT data, for each site, using WinGLink software and re-export of the data for further analysis using other programs. The depth of penetration of the MT responses at each site, for both the XY and YX impedance responses, was then estimated using the Niblett-Bostick transform (Niblett and Sayn-Wittgenstein, 1960; Bostick, 1977; Jones, 1983) of each response. This depth transformation method provides a more reliable means of comparing related responses than a period-based data analysis (Jones, 2006). Figure 7.4 shows the calculated penetration depths for each of the sites. It is evident from the results that the depth of penetration varies from site to site along the profile; this is an indication that the same period images different depths at different sites. Also, at some of the sites, the

Figure 7.4. Niblett-Bostick penetration depth for all the sites in the study area. Red bars represent the XY mode while the blue bars represent the YX mode of the impedance response.



penetration of the YX mode is limited to the near surface, within the Phanerozoic rocks, while the XY modes at the same sites penetrate deeper into the lithosphere.

The penetration depth results (Fig. 7.4) show that only a few sites have penetration of both MT modes through the sedimentary basin rocks but all of the sites have at least component with deeper penetration. Most of the sites, especially on the northeastern half of the profile, have a complete lack of responses with Niblett-Bostick depths in the deeper crust i.e., between 2 km and 45 km. In addition, most of the sites in this same region show penetration of only the XY mode into the lithospheric mantle (>45 km). Based on these observations, this dataset is not considered suitable for modelling the

resistivity structure of the crust. Modelling of the lithospheric resistivity structure is restricted to depths of between 40 km and 150 km as there are relatively few MT responses with Niblett-Bostick depths deeper than 150 km.

7.5 Dimensionality and Strike Analysis

Understanding the dimensionality of the electrical structure is important for applying the appropriate methods to produce a subsurface resistivity image. The MT impedance tensor response from each site can be analyzed to define the dimensionality (1-D, 2-D or 3-D structure) of the regional resistivity structure, and in the case of 2-D or near 2-D structures, the geoelectric strike direction and galvanic distortion.

7.5.1 MT Phase tensor

The MT phase tensor method is a dimensionality analysis tool that requires no presumptions about the underlying regional structure. It assumes that the phase relationship between the horizontal electric and magnetic field vectors will not be distorted by near-surface heterogeneities (Caldwell et al., 2004). The magnitude of the phase tensor skew provides a measure of the three-dimensionality of the response. Small skew angle values could indicate a 2-D or 3-D conductivity structure but large values provide a definitive indication of 3D structures. The ellipticity and variation in the direction of the major axis of the tensor ellipse are indicators of the strength of two-dimensionality and the azimuth of the geoelectric strike. A consistent alignment of the direction of the principal axis of the phase tensor ellipse also provides a reliable indicator of two-dimensionality (Caldwell et al., 2004).

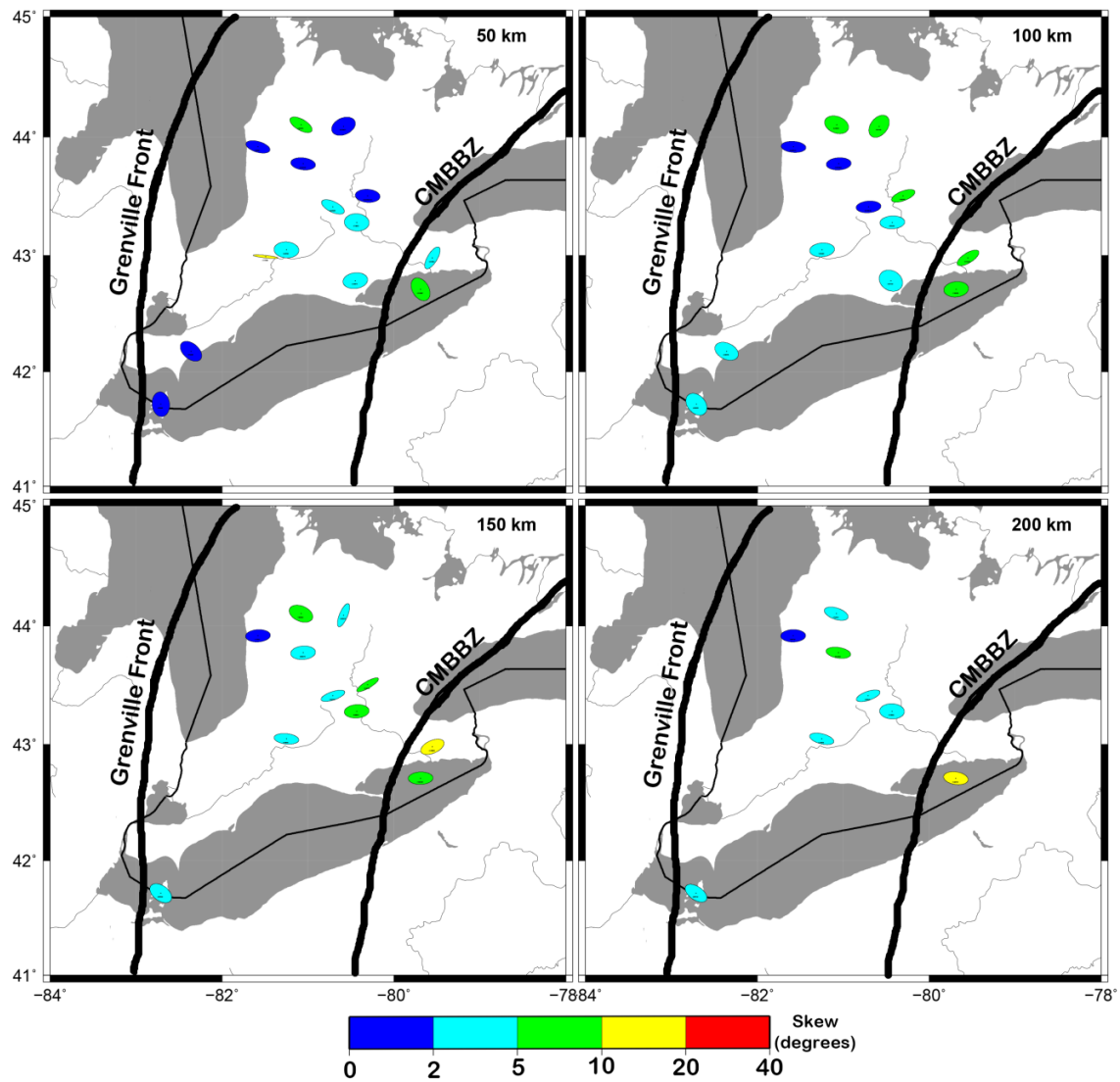
Figure 7.5 shows the phase tensor maps, for 50 km to 200 km depths, for sites in the study area. The phase tensor results are plotted at the estimated Niblett-Bostick depths, which for this analysis are based on the determinant impedance response at each period. Results are not shown for the depth range in the Paleozoic sedimentary basin rocks as these responses are expected to be close to 1-D. Although there are several responses at lithospheric mantle depths with phase tensor skew greater than 10° , the absolute value of skew angle is mostly less than 5° between 50 and 150 km depth. As suggested by Caldwell et al. (2004), these values are not necessarily diagnostic of a 2-D conductivity structure, but indicate that a 2-D structure is possible. The azimuths of the principal axes of the phase tensor ellipses are reasonably constant at each site for all depths. At many sites either the major or minor axes have an approximately east-west or north-south azimuth. This pattern is consistent with a 2-D conductivity distribution and supports the application of a 2-D inversion approach for defining the resistivity structure within the lithospheric mantle.

7.5.2 Tensor Decomposition Analysis and Geoelectric Strike

MT responses can be used to determine the geoelectric strike and its variation with period using the Groom-Bailey (GB) tensor decomposition method (Groom and Bailey, 1989, 1991). This process determines the geoelectric strike simultaneously with a parameterization of the associated galvanic distortion. The GB method was extended by McNeice and Jones (2001) in their STRIKE program to multi-site and multi-period fitting and it was subsequently modified to use depths rather than periods by incorporation of Niblett-Bostick transforms (e.g., Hamilton et al., 2006; Miensopust et al., 2011). The

geolectric strike, in the lithospheric mantle of the lower Great Lakes region, was determined using the McNeice Jones – Groom Bailey (MJ-GB) method implemented in the STRIKE program. It was applied for a depth range extending between 45 and 150 km.

Figure 7.5. Phase tensor of the sites at different mantle depths colored by skew angle. CMBBZ=Central Metasedimentary Belt Boundary Zone.



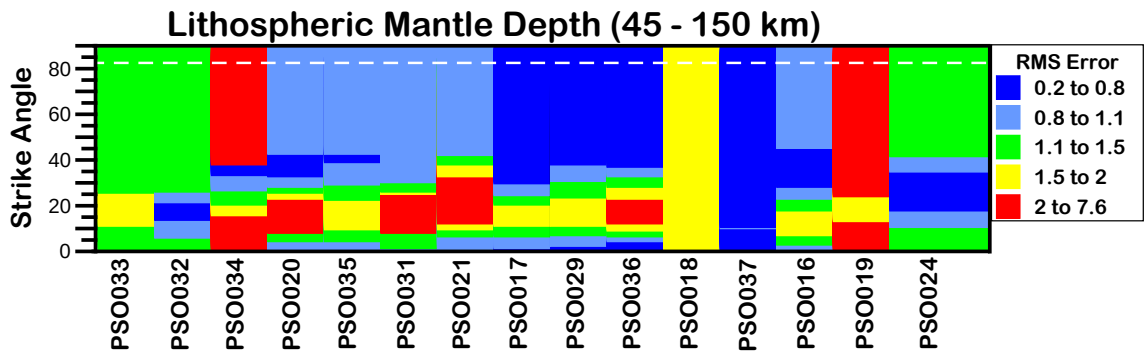
The determination of a suitable regional geoelectric strike azimuth, required for 2-D modeling, was achieved using two methods. Method 1 involves single-site decomposition of the MT data to fixed strike angles from 0° to 90° with a 1° increment. For each site, the angles of decomposition are then plotted against the resulting root mean square (RMS) misfits (e.g., Schmoldt, 2011; Adetunji et al., 2014b). The RMS misfit provides a measure of the fit to the observed impedance of the GB model of a regional 1-D or 2-D response with a frequency-independent distortion over the analyzed band. Method 2 involves determination of a regional strike using the MJ-GB multisite multifrequency strike determination in the STRIKE program.

The RMS misfit for the lithospheric mantle depth band (Fig. 7.6) has a relatively complex form. Several sites show high levels of misfit at all strike azimuths: PSO019 and PSO034 show high (>2) RMS misfit distribution for this depth range and PSO018 is characterized by misfit values of between 1.5 and 2. Many of the remaining sites show a high RMS misfit at azimuths between $N10^{\circ}E$ and $N30^{\circ}E$ indicating that an appropriate regional strike is excluded from this azimuthal range. In contrast, a geoelectric strike azimuth between $N35^{\circ}E$ and $N90^{\circ}E$ seems appropriate for many sites along the profile. Following the elimination of some sites with high RMS misfit and galvanic distortion (PSO019, PSO020, PSO029, and PSO034) the MJ-GB multisite multi-period analysis produced an optimal strike direction of $N8^{\circ}W$ for the whole profile.

The inherent 90° ambiguity in strike azimuth was resolved by choosing the direction closest to the strike of the large-scale geological features in the study area, which is $N8^{\circ}W$. This approach is consistent with observations farther to the northeast. At that location, the large-scale trend of the Grenville Front and CMBBZ is closer to east-

west and strike analyses yield an azimuth close to this value (Adetunji et al., 2014b). It is of note that the phase tensor results for a number of sites in the present study indicate that the azimuth of N8°W corresponds to the minor axis of the phase tensor or the more resistive direction (e.g., Hamilton et al., 2006). Prior to 2-D modeling and inversion, MT datasets were decomposed to the chosen regional strike azimuth by fitting the data to a GB decomposition based on this azimuth. The regional TE impedance response corresponds to electric fields in the N8°W direction and the TM impedance response to fields in the N82°E direction. The sites were also projected perpendicularly onto a modelling profile oriented at N82°E passing through the middle of the study area.

Figure 7.6. RMS misfit of the single site geoelectric strike analysis for lithospheric depth (45 – 150 km). The stepwise strike azimuth is plotted against the resulting RMS misfit for each site along the profile. The white dashed line is the regional strike azimuth of the profile based on multisite multifrequency analysis of the profile using the STRIKE program.



7.5.3 Pseudosections

Figure 7.7 shows MT apparent resistivity and phase pseudosections for the decomposed MT data. The TE mode is sensitive to current flow but has lower sensitivity to the lateral position of the structures than the TM mode. The TM response is more sensitive to lateral conductivity gradients and boundaries. The apparent resistivity response across the whole profile exhibits relative low apparent resistivity ($<300 \Omega.m$) at periods less than 10 s indicating a conductive surface response associated with the sedimentary basin rocks. The transition from the conductive sedimentary rocks to more resistive basement, is reflected in the phase response by a region of low phase ($<40^\circ$) at periods less than 10 s.

The apparent resistivity at longer periods (>10 s) is typically $>500 \Omega.m$. A zone with lower values of apparent resistivity occurs beneath PSO024 and PSO016 and at adjacent sites at the longest periods. Fig. 7.7 shows a second zone with lower apparent resistivity between PSO032 and PSO021 but this feature occurs between the MT sites so is not considered to be resolved. At the longest periods there is an increase in phase to values over 60° on the northeastern end of the profile providing an indication of increased conductivity at depth.

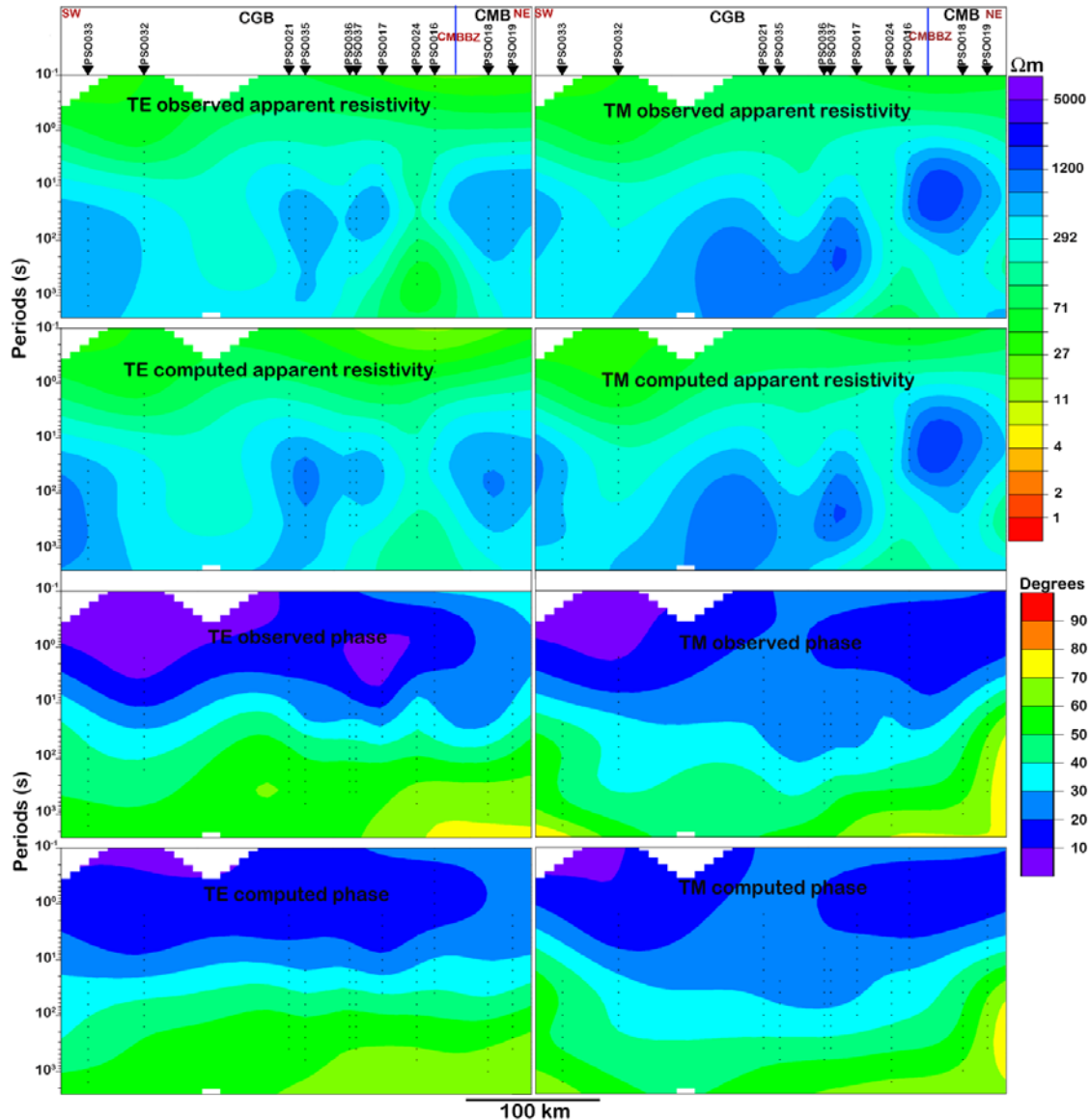
7.6 Modeling and Inversion

7.6.1 Resistivity structure of Paleozoic sedimentary rocks

The short period parts of most of the observed MT responses are consistent with the response of 1-D resistivity structure. Therefore, a 1-D approach was used for the

inversion. A total of 4 sites were selected for this purpose (PSO016, PSO019, PSO021 and PSO032) based on the geographical distribution of the sites, the MT data quality, and the degree of static shift as judged by the offset between the short period XY and YX apparent resistivity responses.

Figure 7.7. Pseudosections of the TE and TM responses of the 2-D model generated from the GB decomposed dataset compared to the observed responses. Each panel shows the observed responses and the responses of the 2-D resistivity model fitted to the observed data. Static shift correction has been applied to the apparent resistivity responses.



The process of 1-D inversion involves use of the MT response from a single site to determine the underlying layered resistivity structure. The 1-D MT response is insensitive to the presence of thin resistive features and its resolution is limited by equivalence in the case of thin conductive layers (Simpson and Bahr, 2005). The response is also relatively insensitive to resistive or moderately conductive features located beneath strong conductors. The Marquardt inversion within the WinGLink software was used to fit the apparent resistivity and phase responses. Models were fitted to the invariant MT response, which averages the two off-diagonal impedance components. In some inversions, the depth and or resistivity of a particular layer was constrained to fixed values. The quality of the fit to the MT data was assessed using root mean square (RMS) misfits. The normalized RMS misfit for the final models at each of the sites varies from 0.05 to 0.08.

In order to address the non-uniqueness inherent in 1-D inversions (Parker, 1980; Constable et al., 1987) the MT data were inverted using models with the smallest number of layers that allowed an adequate fit to the data. An inversion strategy was adopted in which the sites located on the oldest Paleozoic rocks were inverted first in order to establish the resistivity of the lower Paleozoic units. These values were then used in the deeper part of starting models for inversions of data from sites located on younger Paleozoic rocks.

The results of the 1-D inversion are shown in Figure 7.8. At three of the sites, results are compared with resistivity models of Gómez-Treviño and Edwards (1983) which were based on CSEM soundings at nearby locations. This CSEM study used an

electric dipole source which provided signals that provided superior resolution of the resistivity structure of the Paleozoic rocks relative to 1-D MT methods.

Site PSO016 is located in the northeast of the study area on Lower Silurian rocks near the Ordovician-Silurian boundary (Fig. 7.1). The MT data from this site are fitted well by a three-layer model including a 48 m thick upper layer of 324 $\Omega\cdot\text{m}$, a 602 m thick middle layer of 26 $\Omega\cdot\text{m}$ and a resistive, 7000 $\Omega\cdot\text{m}$ basement layer. In Fig. 8 the MT resistivity model for PSO016 is compared with the CSEM resistivity model for site M of Gómez-Treviño and Edwards (1983). Site M is located 40 km to the southwest of PSO016. The two resistivity models agree well. In particular, the conductive middle layer has similar resistivity and thickness in both models. The differences in the parameters of this layer can be attributed to equivalence. The integrated conductance of the layer is more accurately resolved: the CSEM model indicates an integrated conductance of the layer of 24.6 S and the MT model indicates a value of 23.2 S. The conductive layer is interpreted to consist of Upper Ordovician shale units of the Blue Mountain, Georgian Bay and Queenston formations. It may also include a contribution from shale units in the Lower Silurian Cataract Group. The differences in the thin surface layer in the two models layer are likely due to differences in the resistivity of the surficial sediments at the two sites, or in the exact thickness of the Lower Silurian rocks overlying the shale units at the two sites. The lower layer in the MT model is interpreted to represent relatively resistive rocks of the Precambrian basement, Cambrian sandstone and dolostone units, and Middle Ordovician carbonate rocks. The exact resistivity of this part of the model is not well resolved by the MT method, but the results do establish that the rocks are significantly more resistive than the overlying Upper Ordovician units. In

Figure 7.8. 1-D models of the high frequency MT components of four representative sites. The upper panel shows the results of the 1-D MT inversion in which models were fitted to the invariant MT response which averages the two off-diagonal impedance components. The lower panel compares the 1-D models with the models produced at similar locations by Gomez-Trevino and Edwards (1983). The level of editing at sites PSO016 and PSO019 is required because of the high level of noise in the area.



contrast, the CSEM model for site M provides a measure of the resistivity of the uppermost part of this sequence, and shows that the Middle Ordovician carbonate rocks of the Simcoe Group have a resistivity of 100 Ω .m.

MT site PSO019 is located on Upper Ordovician rocks, in the southeast of the study area. Figure 7.8 shows the resistivity model. At this site the main conductive layer has a thickness of only 145 m compared with 602 m at PSO016 so we interpret the site to be lower in the stratigraphic succession (Fig. 7.3). The lower integrated conductance of this layer at this site (6 S) permits better resolution of the underlying units. The MT data are able to resolve an underlying 123 Ω .m unit which, based on the CSEM model for site M, is interpreted to represent the Middle Ordovician carbonate rocks. This layer is underlain by a very resistive layer interpreted to be Precambrian basement and possibly a thin layer of Cambrian sandstone and dolostone units. With this interpretation the MT model indicates a depth to the Precambrian basement at PSO019 is approximately 800 m. This value is in good agreement with the value of 1.0 km estimated from the isopach map in Fig. 7.2a. The thin surface unit at PSO019 has a thickness of 20.5 m and is interpreted to be surficial sediments. There are no CSEM soundings available near this site.

Site PSO021 is located in the north of the study area above the subcropping Upper Silurian Bass Islands Formation (Fig. 7.1). The MT data can be fitted by a three layer model at this site. The MT resistivity model for PSO021 includes two lower layers that are similar to those at PSO016. At this site the conductive layer has an integrated conductance of 20.8 S. The MT data resolves only a single 426 m thick layer above the conductive layer. Based on the stratigraphic position of PSO021 this layer can be interpreted to represent the combined thickness of Silurian rocks.

In Figure 7.8 the MT resistivity model for PSO021 is compared with the CSEM resistivity model site R of Gómez-Treviño and Edwards (1983). This site is located about 100 km to the south of PSO021 but occupies a similar position in the stratigraphic sequence (Fig. 7.3). Unlike the MT sounding, the CSEM method resolves individual layers within the Silurian sequence including a 6000 $\Omega\cdot\text{m}$ layer, composed mainly of evaporites of the Salina Formation, and three poorly-resolved overlying layers with resistivity of about 100 $\Omega\cdot\text{m}$. Despite the good quality of the MT data at PSO021 (Fig. 7.8) the resistive layer is masked in the MT method by the more conducting rocks closer to the surface.

Site PSO032 is located in the southwest of the study area on rocks of the Upper Devonian Kettle Point Formation. As shown in Figure 7.8, the MT model for this site is similar to the model for PSO021. This site is fitted with a three-layer model including a 430 m thick middle conductive layer with integrated conductance of 20 S. Similar to PSO016, it may also include a contribution from shale units in the Lower Silurian Cataract Group. Comparison of the MT result with that of the CSEM method shows that better resolution of younger sequences is provided by the CSEM result. Based on the CSEM model for site P, the conductive layer is interpreted to consist of Upper Ordovician shale units of the Blue Mountain, Georgian Bay and Queenston formations.

The results from the present study can also be compared with those of Mareschal et al. (1991) from the Bruce Peninsular and Manitoulin Island which are located to the north of the current study area. The Mareschal et al. (1991) MT profile crosses the Ordovician-Silurian boundary and has a similar stratigraphic position as PSO016. Mareschal et al. (1991) determined that the shale units had a resistance of 15 $\Omega\cdot\text{m}$ when

located in the subsurface, and 8 $\Omega\cdot\text{m}$ when outcropping. The shale units have an integrated conductance of between 15 S on Manitoulin Island and 25 S on the Bruce Peninsular. The second value agrees well with the 23.2 S determined in the MT inversion at PSO016. Mareschal et al. (1991) also suggest that a significant contribution to the overall conductance of the shale units comes from a relative thin layer of black shale.

7.6.2 Resistivity structure of the Precambrian lithosphere

Based on the strike and dimensionality results described above the resistivity of the lithosphere was determined using a 2-D approach. The 2-D modeling and inversion were done using the algorithm of Rodi and Mackie (2001) implemented in WinGLinK software. The 2-D forward modeling uses a finite difference numerical method for calculation. The inversion code applies the non-linear conjugate gradient (NLCG) method for direct iterative minimization of an objective function that penalizes data residuals and second spatial derivatives with respect to resistivity structures (Rodi and Mackie, 2001, 2012). The inversion models were obtained by solving for the smoothest model using uniform-grid Laplacian regularization and minimizing the integral of the Laplacian (Rodi and Mackie, 2012).

Many 2-D inversions were run to examine different possible models that could fit the data. The final 2-D model was obtained by fitting both TE and TM modes for all the sites using a 100 $\Omega\cdot\text{m}$ half space as the starting model. The data errors on both TE and TM resistivity values were set at 10% and 5% ($\sim 1.5^\circ$) for the phase. The final error floors were set to 16% and 12% for TE and TM resistivity and 4% and 3% for TE and TM phases. New inversions were started using interpolated data with five frequencies per

decade and completed using station data. A Tikhonov regularization parameter ($\tau = 6$), horizontal derivative ($\alpha = 1$), depth function ($\beta = 0.3$), horizontal smoothing ($H = 500$) and vertical smoothing ($V = 500$) factors were used to obtain the optimal model. For a more reliable final model with a deeper minimum of the objective function, all inversions were restarted at least once after its initial termination and the typical number of iterations was 400. The final model was corrected for static shift effects on the apparent resistivity responses.

Figure 7.9 shows the resistivity model and data misfit for the lithospheric mantle. The top 45 km of this model (i.e., the crust) has been blanked out because of the low model resolution at that depth (Fig. 3). Sites PSO020, PSO029, PSO031, and PSO034 were excluded from the final set of inversions because they exhibited high RMS misfit (>4) in earlier sets of inversions. The global RMS error of the final resistivity model is 1.47 and the fits of the inversion model to individual sites are spatially uniform. In addition, the large-scale features in the observed data are all reproduced by the model response as shown by the pseudosection in Figure 7.7.

The model shows that the lithospheric mantle in the northeast part of the profile is relatively conductive with a resistivity of 100 to 200 $\Omega.m$ decreasing to values of $<5 \Omega.m$ in a conductor beneath the CMB (labeled C). There are resistive blocks in the middle and southwest parts of the model. Beneath the northeastern CGB a resistive zone (labeled R), containing resistivity values of $>2,000 \Omega.m$ extends from the base of the crust to depth of at least 150 km. The resistive block to the southwest extends from the base of the crust to about 90 km depth. The resistivity of the lithosphere beneath this block and between the two resistive blocks is between 200 and 500 $\Omega.m$.

The data were examined to ascertain the reliability of the structures in the resistivity model. The results show that, although the resolution and shape of these features vary to some degree for different inversions, the conductor C and the resistor R are first-order features that exist in the subsurface. The resistor is overlain by five MT sites so is considered to be a very reliable aspect of the model. The data from MT sites overlying these features have corresponding Niblett-Bostick depths exceeding the depth to the top of the structures, with conductor C being defined by both TE and TM modes, and resistor R by the TE mode. The conductor is associated with an increase in the TE and TM phase responses at longer periods (Fig. 7.7). Modelling indicates that although the geometry of the individual resistive blocks is not well resolved, the two blocks are definitely separated by more conductive lithosphere. The MT data are unable to resolve the depth to the base of either the resistor R or conductor C.

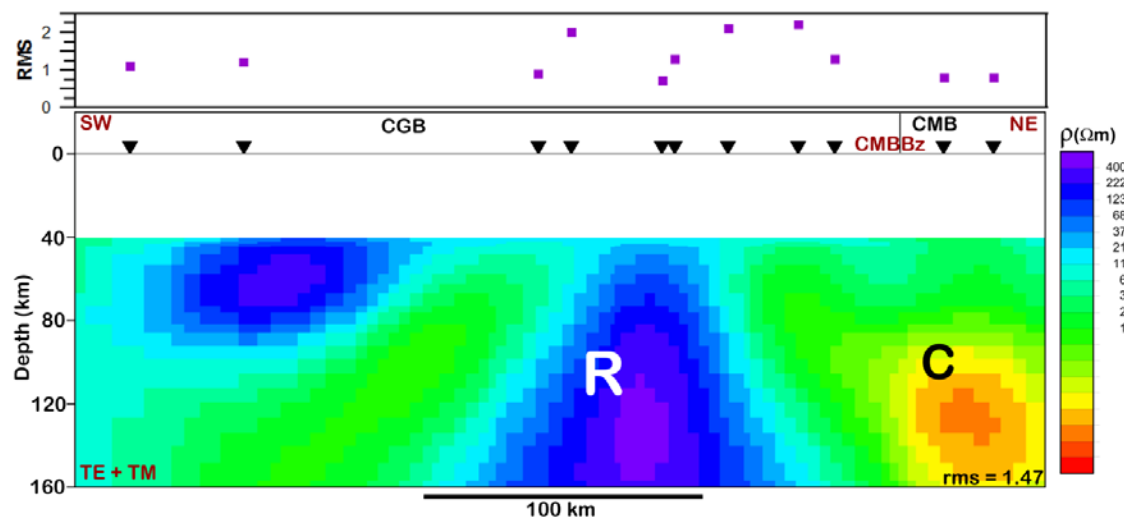
7.7 Interpretation and discussion

The results of the phase tensor and geoelectric strike analysis provide insight into the dimensionality of the lithospheric structure in the study area. Within the lithospheric mantle depth range (45 – 150 km), the direction of the principal axes of the phase tensor ellipse varies significantly across the area. The complexity inherent in this result suggests a regionally 2-D but locally 3-D environment. Phase tensor skew angle values are mostly small and analysis of the average tensor decomposition misfits for this lithospheric depth range shows that the region is characterized by relatively small RMS misfit values (mostly <1.5). These observations indicate that 2-D modeling and inversion is acceptable

for this region. The MJ-GB decomposition results indicate an optimal strike direction of N8°W for this modelling.

The results in the current study have provided an indication of the resolution of the MT method for mapping units within the Michigan and Allegany basins in southwest Ontario. The Upper Ordovician shales that extend across the study area, and have a conductance of 15-25 S, have a major influence on the MT response as initially shown by Mareschal et al. (1991). Resolution of the underlying Middle Ordovician rocks and the approximate depth to basement is possible only at locations beyond the margins of the Upper Ordovician rocks or at stratigraphic positions low within these units. As shown by

Figure 7.9. Resistivity model derived by joint inversion of TE and TM responses ($V.E = 1$). The RMS misfits of the joint TE and TM (purple squares) inversions at individual sites are plotted above the inversion model. The crustal section (upper 40 km) of the model is not shown.



the Niblett-Bostick depths in Fig. 7.4, the shale units effectively preclude MT imaging at crustal depths. The MT method provides some resolution of units lying above the shales, particularly where these units subcrop. However, in layered sequences of resistive carbonates or evaporites and conductive shales, the conductive units preclude resolution of more resistive underlying units. Although the application of MT in stratigraphic studies is limited by these aspects, it still has important applications in the region. For example, the MT responses can be used in geomagnetic hazard applications in the calculation of surface electric fields needed for modelling geomagnetically induced currents on powerlines and pipelines (Lanzerotti and Gregori, 1986; Boteler et al., 1998; Fernberg et al., 2007).

An important and well-resolved feature of the lithospheric resistivity model is the resistive structure (labeled R in Fig. 7.9) that extends from the base of the crust to at least 150 km depth in the lithospheric mantle. The base of this feature cannot be resolved in this study due to lack of adequate data penetration. To the northeast of the current study area Adetunji et al. (2014a, 2014b) imaged a large-scale, highly resistive structure that extends from the base of the crust to a maximum depth of about 280 km beneath Grenville Province. This feature extends from the Grenville Front to a location 100 km south of the CMBBZ, it extends along strike for a distance of 300 km, and it is spatially correlated with high S-wave velocity. The authors interpret this feature to be stable Archean Superior Province lithosphere. This lithosphere would have acted as a platform for the Grenville orogenic events (Ludden and Hynes, 2000; Adetunji et al., 2014a, 2014b).

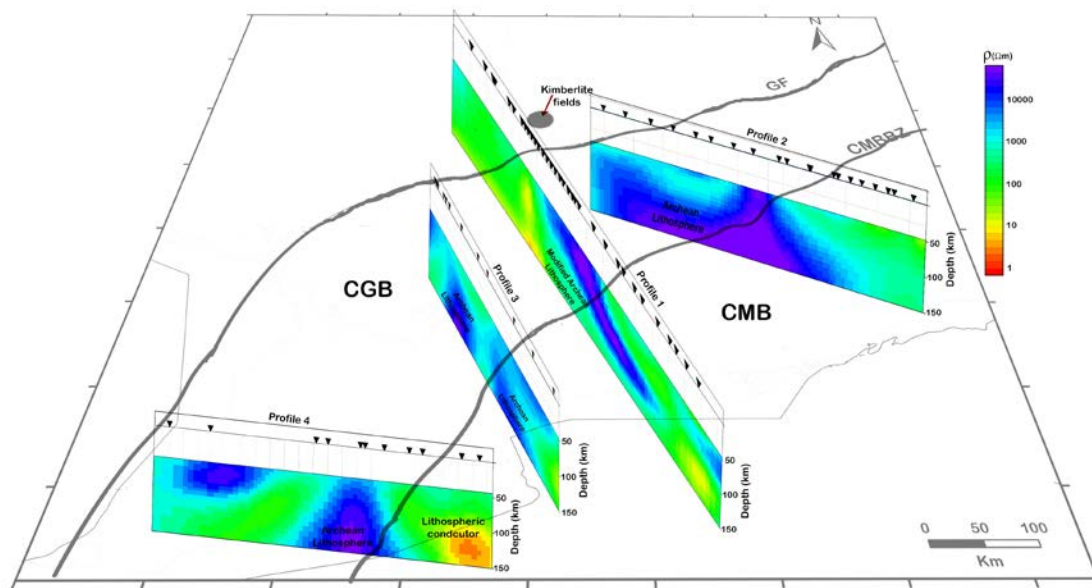
Figures 7.10 and 7.11 shows the spatial relationship of this resistor with the one imaged in the present study. We interpret the lithospheric resistor R as a southwestern extension of the resistor to the northwest. It has a similar range of resistivity values, a similar southern margin, and a similar upper surface to the resistor imaged in the previous studies. The result therefore suggests the presence of resistive lithosphere beneath the CGB of the northwest Grenville Province extending for at least 800 km along strike. Chen and Li (2013) show that the upper 75 km of mantle lithosphere displays relatively high S wave velocities ($>4.7 \text{ km.s}^{-1}$) across this same area.

The observed along-strike extent of the resistor is a factor that must be included in its interpretation. To the northeast, the resistive lithosphere beneath the Grenville Province was interpreted as being associated with the adjacent Superior Province (Adetunji et al., 2014a, 2014b). However, farther to the west the Grenville Province truncates younger rocks of the Early Proterozoic Southern Province, and farther southwest, it truncates progressively younger rocks including those deformed during the Penokean orogeny (e.g., Forsyth et al., 1994), and Mazatzal orogeny (Whitmeyer and Karlstrom, 2007). Based on the similarity of the resistor along strike it is more likely that it represents a coherent block of Archean lithosphere rather than being composed of Archean rocks in the northeast and Proterozoic rocks to the southwest. We therefore interpret the mantle lithosphere beneath the lower Great Lakes region as being part of the Superior lithosphere. This interpretation implies that the Mazatzal mantle lithosphere does not extend into this area. Additional support for this interpretation comes from shear-wave splitting anomalies that show fairly consistent near east-west fast directions

and ~0.5 s time splits within the CGB of the Grenville Province over the whole Great Lakes region.

Further collection and integration of deep penetrating geophysical results from the region will be required in order to fully define the origin of the lithosphere beneath the northwest Grenville Province. The MT data collected in the POLARIS project provide a preliminary resistivity model for the Great Lakes area but because of the high noise levels, limited recording durations, and the masking by the 15-25 S Upper Ordovician shale units in the region, the model does not have sufficient resolution to make definitive tectonic interpretations. In particular, improved resolution of the gap between the two resistive blocks will assist in the interpretation.

Figure 7.10. Resistivity models of Profiles 1 (Adetunji et al., 2014a), profiles 2 and 3 (Adetunji et al., 2014b), and Profile 4 (this study) projected along the MT data profiles on the map of the study area. Each profile is truncated at 150 km depth with a V.E of 0.5. GF=Grenville Front, CMBBZ= Central Metasedimentary Belt Boundary Zone. CGB=Central Gneiss Belt, CMB=Central Metasedimentary belt. The crustal section (upper 48 km) of all the models is not shown.



The resistivity model from the current study, in association with the analysis of POLARIS MT data on profiles to the northeast (Adetunji et al., 2014a, 2014b), shows that, to the south of the resistive lithosphere, beneath the CMB, the lithosphere of the Grenville Province is conductive. In the current study this lithosphere hosts a localized conductor with its top at ~100 km depth and with resistivity values $<5 \Omega.m$. Adetunji et al. (2014a) model a conductor at the southeastern end of their profile, Profile 1, approximately 200 km south of the CMBBZ. This conductor has its top at ~100 km depth and contains resistivity values $<10 \Omega.m$. On Profile 1 the base of the conductor is defined by a decrease in phase responses at periods of several hundred seconds and is at a depth of ~160 km. Figures 7.10 and 7.11 show the geometrical relationship of the conductors in the two studies. The 2-D MT inversions for two additional profiles also suggested the presence of deep conductors beneath the CMB (Adetunji et al., 2014b). However, these features were located beyond the final site in each profile and their resolution was considered inadequate for them to be examined in detail.

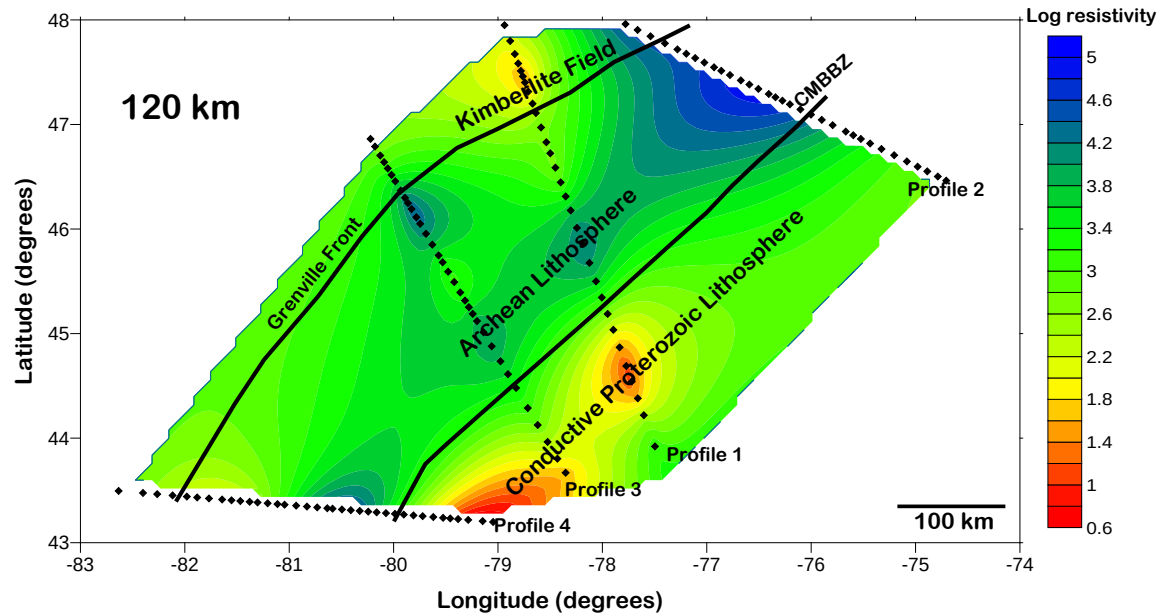
The observations in southern Ontario firmly establish the presence of very conductive lithosphere beneath the CMB in the Grenville Province. As shown in figures 7.10 and 7.11 the conductive lithosphere appears to be continuous along strike for at least 100's of kilometers. The top of localized conductors within this zone is at a depth of around 100 km and based on the results for Profile 1 the base is at a depth of around 160 km. This depth distribution provides an indication that anomalies may be due to the presence of grain boundary graphite films (Selway, 2014). Such anomalies would be expected to have a base at around 150 km depth in stable continental lithosphere at the limit of the graphite stability field.

Adetunji et al. (2014a, 2014b) interpret conductive features northwest of the CMB in the Grenville Province as being associated with the deep effects of the passage of the postulated Great Meteor hotspot (source) and kimberlitic magmatism. A subvertical conductor beneath the Pontiac subprovince occurs only a few tens of kilometers along strike from the diamondiferous kimberlites in the Kirkland Lake and Cobalt kimberlite fields (Adetunji et al., 2014a). Enhanced conductivity (and spatially correlated low S-wave velocity) in the lower lithosphere of the adjacent Grenville Province, between 150 and 280 km depth, is interpreted to be associated with refertilization by the hotspot and kimberlitic magmatic processes (Adetunji et al., 2014b). At these depths the enhanced conductivity must be attributed to the presence of water in nominally anhydrous minerals rather than graphite.

One explanation for the localized anomalies at 100 to 150 km depth in the CMB is that they are also associated with kimberlitic magmatic processes. It has been shown that kimberlite melts are normally rich in carbon and are likely to release carbon dioxide in the uppermost mantle (Hunter and McKenzie, 1989; Pearson et al., 1994; Sleep, 2009). The degassed carbon dioxide could move along grain boundaries (Hunter and McKenzie, 1989) and become converted to graphite, which is stable at depths less than 150 km (Pearson et al., 1994). However, it is unclear whether these processes can occur over a sufficiently wide spatial scale to explain conductive anomalies on profiles separated by 300 km.

It is probable that conductive nature of the deep lithosphere beneath the CMB can be due, at least in part, to the introduction of water and carbon into the mantle during pre-Grenvillian (1450-1740 Ma) collisional events and subduction (Carr et al., 2000). This

Figure 7.11. A horizontal slice, at 120 km depth, through the resistivity models of profiles 1 to 4. The resistivity values at this depth, interpolated with natural neighbor algorithm, were projected onto the MT data acquisition profile and contoured. The result shows lateral extension of Archean lithosphere throughout the study area. It also shows a lithospheric conductor that extends from profiles 1 to 3. CMBBZ = Central Metasedimentary Belt Boundary Zone



mechanism would provide the source for widespread introduction of elements capable of enhancing conductivity. It is of note that in the southern United States the Grenville suture is associated with enhanced lower crustal and upper mantle conductivity (Wannamaker et al., 1996, Ogawa et al., 1996, Wannamaker, 2005) suggesting a very large-scale distribution of conductive lithosphere. Other Proterozoic suture zones are also associated with enhanced conductivity at crustal and/or mantle depth (e.g., Korja and Hjelt, 1993; Boerner et al., 2000; Jones et al., 2005; Heinson et al., 2006; Spratt et al., 2009). The carbon and water introduced during the pre-Grenvillian tectonic activity may

also have been an important source during the Cretaceous kimberlitic activity. As noted above, this process may also have caused the localized conductive anomalies in the lithosphere of the CMB. However, it is also conceivable that these anomalies were formed during the pre-Grenvillean tectonism.

7.8 Conclusions

The electrical resistivity structure beneath the lower Great Lakes region in southern Ontario is investigated using broadband and long-period MT data from POLARIS project. The geology of the region consists of sedimentary rocks of the Michigan and Alleghany basin overlying Proterozoic rocks of the Grenville Province. The MT resolution in the study area is influenced strongly by a 400 m thick sequence of Upper Ordovician shale units with conductance of 21 S within the sedimentary rocks. These rocks severely limit the resolution of the MT method in the crust and the deep mantle lithosphere. As a result the MT analysis is focused on the Paleozoic sedimentary rocks in the upper 2 km and on the lithospheric mantle at 45 to 150 km depth.

A regional geoelectric strike azimuth of N8°W was determined for the lithospheric mantle of this region using MJ-GB multisite multifrequency and RMS misfit analysis. Dimensionality analysis shows results that are consistent with 2-D conductivity distribution. Two-dimensional modeling and inversion of the MT data was done using a profile oriented at N82°E. A well resolved lithospheric resistor, which extends from the base of the crust to at least 150 km depth in the lithospheric mantle, is interpreted to be the southwestern extension of the Archean lithosphere described in Adetunji et al. (2014a, 2014b). It is suggested that this is a part of stable lithosphere which acted as the

platform for the Grenville orogenic events. Also, based on the similarity of the resistor along strike, it is likely that it represents a coherent block of Archean lithosphere.

A localized lithospheric conductor that extends from about 100 to 150 km depth beneath the CMB is imaged in this study. The conductor is along strike with a similar one imaged by Adetunji et al. (2014a) on Profile 1. The observation affirms the presence of very conductive lithosphere beneath the CMB in the Grenville Province. The source of enhanced conductivity in this conductor may be attributed to grain boundary graphite films associated with Cretaceous kimberlitic magmatic process. It could also be as a result of the water and carbon introduced into the mantle during the pre-Grenvillian tectonism.

7.9 Acknowledgements

The authors thank the POLARIS southern Ontario MT team (J. Wenham, M. Serzu, X. Ma, E. Gowan, H. Sealey, T. Pacha, J. McCutcheon, I. Asudeh, C. Andrews, C. Samson, P. Fernberg) for MT data collection and landowners in the region for site access. We also thank A. Frederiksen for his comments and A. Jones for his STRIKE and MTMAP codes. Phase tensor figures were produced using Wessel and Smith's GMT package. Funding for POLARIS has been provided by industry, the Canada Foundation for Innovation, provincial agencies from Ontario and British Columbia, the Federal Economic Development Initiative for Northern Ontario, Natural Resources Canada, and NSERC (Natural Sciences and Engineering Research Council of Canada). The present research was supported by an NSERC Discovery Grant (105748) to Ian Ferguson.

7.10 References Cited

- Adetunji, A.Q., Ferguson, I.J., Jones, A.G., 2014a. Crustal and lithospheric scale structures of the Precambrian Superior-Grenville margin. *Tectonophysics* 614, 146-169, <http://dx.doi.org/10.1016/j.tecto.2013.12.008>.
- Adetunji A.Q, Ferguson, I.J., Jones, A.G., (2014b). Imaging the mantle lithosphere of the Grenville Province: large-scale structures and electrical anisotropy. *Geophys. J. Int.* (submitted in Jan. 2014).
- Armstrong, D.K., Carter, T. R., 2006. An updated guide to the subsurface Paleozoic stratigraphy of southern Ontario. Ontario Geological Survey, Open File Report 6191.
- Boerner, D. E., Kurtz, R. D., Craven, J. A., 2000. A summary of electromagnetic studies on the Abitibi-Grenville transect. *Can. J. Earth Sci.* 37, 427-437.
- Bostick, F. X., 1977. A simple almost exact method of MT analysis. Workshop on Electrical Methods in Geothermal Exploration. US. Geol. Surv. Contract No. 14080001-8-359.
- Boteler, D. H., Pirjola, R. J., Nevanlinna, H., 1998. The effects of geomagnetic disturbances on electrical systems at the earth's surface, *Adv. Space Res.* 22, 17-27.
- Caldwell, T.G., Bibby, H.M., Brown, C., 2004. The magnetotelluric phase tensor. *Geophys. J. Int.* 158, 457-469.
- Carr, S. D., Easton, R. M., Jamieson, R. A., Culshaw, N. G., 2000. Geologic transect across the Grenville orogen of Ontario and New York. *Can. J. Earth Sci.* 37, 193-216.
- Chave, A. D., Jones, A. G. (Eds.), 2012. *The Magnetotelluric Method - Theory and Practice*, Cambridge University Press, New York, 570pp.
- Chen, C. -W., Li, A., 2012. Shear wave structure in the Grenville Province beneath the lower Great Lakes region from Rayleigh wave tomography. *J. Geophys. Res.* 117, B01303.
- Constable S. C., Parker, R. L., Constable, C. G., 1987. Occam's inversion: a practical algorithm for generating smooth models from electromagnetic sounding data. *Geophysics* 52, 289 – 300.
- Constable, S. C., 2006. SEO3: a new model of olivine electrical conductivity. *Geophys. J. Int.* 166, 435-437.
- Culotta, R.C., Pratt, T., Oliver, J., 1990. A tale of two sutures: COCORP's deep seismic surveys of the Grenville province in the eastern U.S. midcontinent. *Geology* 18, 646-649.
- Davidson, A., 1986. New interpretations in the southwestern Grenville Province, in: Moore, J. M., Davidson, A., Baer, A. J. (Eds.), *The Grenville Province*, Geological Association of Canada, Geological Survey of Canada, Special paper 31, pp. 61-74.
- Ducea, M. N., Park, S. K., 2000. Enhanced mantle conductivity from sulfide minerals, southern Sierra Nevada, California. *Geophys. Res. Lett.* 27, 2405-2408.
- Duncan, P. M., Hwang, A., Edwards, R. N., Bailey, R. C., Garland, G. D., 1980. The development and applications of a wide band electromagnetic sounding system using a pseudo-noise source. *Geophysics* 45, 1276-1296.
- Easton, R. M., 1986. Geochronology of the Grenville Province, in: Moore, J. M., Davidson, A., Baer, A. J. (Eds.), *The Grenville Province*, Geological Association of Canada, Geological Survey of Canada, Special paper 31, pp. 127-173.

- Easton, R. M., 1992. The Grenville Province and the Proterozoic history of Central and Southern Ontario, in: Thurston, P.C., Williams, H.R., Sutcliffe, R.H., Stott, G.M. (Eds.), *Geology of Ontario*, Ontario Geological Survey, Special Volume 4, pp. 713 – 904.
- Easton, R.M., Carter, T.R., 1995. Geology of the Precambrian basement beneath the Paleozoic of southwestern Ontario, in: Ojakangas, R. W., Dickas, A.B., Green, J. C. (Eds.), *Basement Tectonics*, Kluwer Academic Publishers, The Netherlands. pp. 221-264.
- Eaton, D. W., Adams, J., Asudeh, I., Atkinson, G. M., Bostock, M. G., Cassidy, J. F., Ferguson, I. J., Samson, C., Snyder, D. B., Tiampo, K. F., Unsworth, M. J., 2005. Geophysical arrays to investigate lithosphere and earthquake hazards in Canada. *EOS Trans AGU* 86, 169-173.
- Eaton, D. W., Dineva, S., Mereu, R., 2006. Crustal thickness and Vp/Vs variations in the Grenville orogen Ontario, Canada from analysis of teleseismic receiver functions. *Tectonophysics* 420, 223-238.
- Fernberg, P. A., Samson, C., Boteler, D. H., Trichtchenko, L., Larocca, P., 2007. Earth conductivity structures and their effects on geomagnetic induction in pipelines. *Annales Geophysicae* 25, 207 – 218.
- Fisher, J. H., Barratt, M. W., Droste, J. B., Shaver, R. H., 1988. Michigan Basin, in: Sloss, L. L. (Ed.), *Sedimentary cover—North American craton*, Geological Society of America, volume D-2, pp. 361-382.
- Forsyth, D.A., Milkereit, B., Davidson, A., Hanmer, S., 1994. Seismic images of a tectonicsubdivision of the Grenville Orogen beneath lakes Ontario and Erie. *Can. J. Earth Sci.* 31, 229-242.
- Frederiksen, A.W., Ferguson, I. J., Eaton, D., Miong, S. K., Gowan, E., 2006. Mantle fabric at multiple scales across an Archean-Proterozoic boundary, Grenville Front, Canada. *Phys. Earth Planet. Int.* 158, 240-263.
- Gartner Lee Limited, 2008. Phase 1 Regional Geology, Southern Ontario, OPG's Deep Geologic Repository for Low and Intermediate Level Waste. OPG Report # 00216-REP-01300-00007-R00.
- Glover, P.W. J., Vine, F. J., 1995. Beyond KTB - Electrical conductivity of the deep continental crust. *Surv. Geophys* 16, 5-36.
- Glover, P.W. J., Vine, F. J., 1995. Beyond KTB - Electrical conductivity of the deep continental crust. *Surv. Geophys.* 16, 5-36.
- Gomez-Trevino, E., Edwards, R. N., 1983. Electromagnetic soundings in the sedimentary basins of southern Ontario- A case history. *Geophysics* 48, 311-330.
- Groom, R. W., Bailey, R. C., 1989. Decomposition of magnetotelluric impedance tensors in the presence of local three-dimensional galvanic distortion. *J. Geophys. Res.* 94, 1913-1925.
- Groom, R. W., Bailey, R. C., 1991. Analytic investigations of the effects of near-surface three-dimensional glavanic scatterers on MT tensor decompositions. *Geophysics* 56, 496 – 518.
- Haggart, M. J., Jemieson, R. A., Reynold, P. H., Krogh, T. E., Beaumont, C., Culshaw, N. G., 1993. Last gasp of the Grenville Orogeny: thermochronology of the Grenville Front Tectonic Zone near Killarney, Ontario. *J. Geology* 101, 575-589.

- Hamilton, M. P., Jones, A. G., Evans, R. L., Evans, S., Fourie, C. J. S., Garcia, X., Mountford, A., Spratt, J. E., the SAMTEX Team, 2006. Electrical anisotropy of South African lithosphere compared with seismic anisotropy from shear-wave splitting analysis. *Phys. Earth Planet. Int.* 158, 226–239.
- Hanmer, S., McEachern, S. J., 1992. Kinematical and rheological evolution of a crustal-scale ductile thrust zone, Central Metasedimentary Belt, Grenville Orogen, Canada. *Can. J. Earth Sci.* 29, 1779–1790.
- Heinson, G., Direen, N.G., Gill, R., 2006. Magnetotelluric evidence for a deep-crustal mineralising system beneath the giant Olympic Dam iron-oxide copper gold deposit, southern Australia. *Geology* 34, 573–576. doi: 10.1130/G22222.1
- Hirth, G., Evans, R. L., Chave, A. D., 2000. Comparison of continental and oceanic mantle electrical conductivity: Is the Archean lithosphere dry? *Geochim. Geophys. Geosyst.* 1, 1030, doi:10.1029/2000GC000048.
- Howell, P.D., van der Pluijm, B.A., 1990. Early history of the Michigan basin: subsidence and Appalachian tectonics. *Geology* 18, 1195–1198.
- Howell, P.D., van der Pluijm, B.A., 1999. Structural sequences and styles of subsidence in the Michigan basin. *GSA Bulletin* 111, 974–991.
- Hunter R.H., McKenzie, D., 1989. The equilibrium geometry of carbonate melts in rocks of mantle composition. *Earth Planet. Sci. Lett.* 92, 347–356.
- Hyndman, R. D., Wang, K., Yuan, T., Spence, G. D., 1993. Tectonic sediment thickening, fluid expulsion, and the thermal regime of subduction zone accretionary prisms: The Cascadia margin off Vancouver Island. *J. Geophys. Res.* 98, 21865–21876, doi: 10.1029/93JB02391.
- Hynes, A., Rivers, T., 2010. Protracted continental collision - evidence from the Grenville Orogen. *Can. J. Earth Sci.* 47, 591–620.
- Intera Engineering Limited, 2008. Phase 2 Geoscientific Site Characterization Plan, OPG's Deep Geologic Repository for Low and Intermediate Level Waste. Report, OPG Report # 00216-REP-03902-00006-R00.
- Intera Engineering Limited, 2009. Regional and site geological frameworks –proposed Deep Geologic Repository, Bruce County, Ontario. GeoHalifax2009.
- Johnson, M. D., Armstrong, D. K., Sanford, B. V., Telford, P. G., Rutka, M. A., 1992. Paleozoic and Mesozoic Geology of Ontario, in: *Geology of Ontario*. Thurston, P.C., Williams, H.R., Sutcliffe, R.H., Stott, G.M. (Eds.), Geology of Ontario, Ontario Geological Survey, Special Volume 4, pp. 907 – 1008.
- Jones, A. G., 1988. Static shift of magnetotelluric data and its removal in a sedimentary basin environment. *Geophysics* 53, 967–978.
- Jones, A. G., 2006. Electromagnetic interrogation of the anisotropic Earth: Looking into the Earth with polarized spectacles. *Phys. Earth Planet. Int.* 158, 281–291.
- Jones, A. G., Craven, J. A., 1990. The North American Central Plains conductivity anomaly and its correlation with gravity, magnetic, seismic, and heat flow data in Saskatchewan, Canada. *Phys. Earth Planet. Int.* 60, 169 - 194.
- Jones, A.G., 1983. On the equivalence of the “Niblett” and “Bostick” transformations in the magnetotelluric method. *Z. Geophys.* 53, 72–73.
- Jones, A.G., 1999. Imaging the continental upper mantle using electromagnetic methods. *Lithos* 48, 57– 80.

- Jones, A.G., Ledo, J., and Ferguson, I.J. 2005. Electromagnetic images of the Trans-Hudson orogen: the North American Central Plains (NACP) anomaly revealed. *Can. J. Earth Sci.* 42, 457–478.
- Karato, S., 1990. The role of hydrogen in the electrical conductivity of the upper mantle. *Nature* 347, 272–273.
- Katsube, T.J., Mareschal, M., 1993. Petrophysical model of deep electrical conductors: graphite lining as a source and its disconnection due to uplift. *J. Geophys. Res.* 98, 8019 - 8083.
- Klein, G. deV., Hsui, A.T., 1987. Origin of Cratonic Basins. *Geology* 15, 094-1098.
- Korja, T., 2007. How is the European Lithosphere Imaged by Magnetotellurics? *Surv. Geophys.* 28, 239 – 272.
- Korja, T., Hjelt, S. –E., 1993. Electromagnetic studies in the Fennoscandian Shield—electrical conductivity of Precambrian crust. *Phys. Earth Planet. Inter.* 81, 107–138.
- Krogh, T.E., 1994. Precise U-Pb ages for Grenvillian and pre-Grenvillian thrusting of Proterozoic and Archean metamorphic assemblages in the Grenville Front Tectonic Zone, Canada. *Tectonics* 13, 963-982.
- Kumarapeli, P. S., 1985, Vestiges of Iapetan rifting in the craton west of Appalachians. *Geoscience Canada* 12, 54 – 59.
- Lanzerotti, L. J., Gregori, G. P., 1986. Telluric currents: The natural environment and interactions with man-made systems, in: *The Earth's electrical environment*, National Academy Press, pp.232–257.
- Li, S., Unsworth, M. J., Booker, J. B., Wei, W., Tan, H., Jones, A. G., 2003. Partial melt or aqueous fluid in the mid-crust of Southern Tibet? Constraints from INDEPTH Magnetotelluric data. *Geophysics* 153, 289-304.
- Ludden, J., Hynes, A., 2000. The Lithoprobe Abitibi-Grenville transect: two billion years of crust formation and recycling in the Precambrian Shield of Canada. *Can. J. Earth Sci.* 37, 459-476.
- Mareschal, M., Kurtz, R.D., Chouteau, M., Chakridi, R., 1991. A magnetotelluric survey on Manitoulin Island and Bruce Peninsula along GLIMPCE seismic line J: black shales mask the Grenville Front. *Geophys. J. Int.* 105, 173–183.
- Mazurek, M., 2004. Long-term used nuclear fuel waste Management - geoscientific review of the sedimentary sequence in southern Ontario. Institute of Geological Sciences University of Bern, Switzerland Technical Report TR 04-01.
- McNeice, G. W., Jones, A. G., 2001. Multisite, multifrequency tensor decomposition of magnetotelluric data. *Geophysics* 56, 158-173.
- Mibe, K., Fuji, T., Yasuda, A., 1998. Connectivity of aqueous fluid in the Earth's upper mantle. *Geophys. Res. Lett.* 25, 1233-1236.
- Miensopust, M. P., Jones, A. G., Muller, M. R., Garcia, X., Evans, R. L., 2011. Lithospheric structures and Precambrian terrane boundaries in northeastern Botswana revealed through magnetotelluric profiling as part of the Southern African Magnetotelluric Experiment. *J. Geophys. Res.* 116, B02401.
- Muller, M. R., Jones, A. G., Evans, R. L., Gru"tter, H. S., Hatton, C., Garcia, X., Hamilton, M. P., Miensopust, M. P., Cole, P., Ngwisanyi, T., Hutchins, D., Fourie, C. J., Jelsma, H. A., Evans, S. F., Aravanis, T., Pettit, W., Webb, S. J., Wasborg, J., The SAMTEX Team, 2009. Lithospheric structure, evolution and diamond

- prospectivity of the Rehoboth Terrane and the western Kaapvaal Craton, southern Africa: Constraints from broadband magnetotellurics. *Lithos* 112S, 93-105.
- Niblett, E. R., Sayn-Wittgenstein, C., 1960. Variation of electrical conductivity with depth by the magneto-telluric method. *Geophysics* 25, 998-1008.
- Nunn, J.A., Sleep, N.H., 1984. Thermal contraction and flexure of intracratonal basins: a three-dimensional study of the Michigan Basin. *J. R. Astr. Soc.* 76, 587-635.
- O'Dowd, C. R., Eaton, D., Forsyth, D., Asmis, H. W., 2004. Structural fabric of the Central Metasedimentary Belt of southern Ontario, Canada, from deep seismic profiling. *Tectonophysics* 388, 145-159.
- Ogawa, Y., Jones, A. G., Unsworth, M. J., Booker, J. R., Lu, X., Craven, J., Roberts, B., Parmalee, J., Farquharson, C., 1996. Deep Electrical Conductivity Structures of the Appalachian Orogen in the Southeastern US. *Geophys. Res. Lett.* 23, 1597-1600.
- Parker, R. L., Booker, J. R., 1996. Optimal one-dimensional inversion and bounding of magnetotelluric apparent resistivity and phase measurements. *Phys. Earth Planet. Int.* 98, 269-282.
- Parker, R.L., 1980. The inverse problem of electromagnetic induction: Existence and construction of solutions based on incomplete data, *J. Geophys. Res.*, **85**, 4421-4428.
- Pearson, D., Boyd, F., Haggerty, S., Pasteris, J., Field, S., Nixon, P., Pokhilenko, N., 1994. The characterisation and origin of graphite in cratonic lithospheric mantle: a petrological carbon isotope and Raman spectroscopic study. *Contrib. Min. Pet.* 115, 449-466.
- Rivers, T., Culshaw, N., Hynes, A., Indares, A., Jamieson, R., Martignole, J., 2012. The Grenville orogen- A Post-Lithoprobe Perspective in: *Tectonic Styles in Canada: The Lithoprobe Perspective*, pp 97-236, eds. Percival, J.A., Cook, F.A., Clowes, R. M., Geological Association of Canada, Special paper 49.
- Rodi, W. L., Mackie, R. L., 2001. Nonlinear conjugate gradients algorithm for 2D magnetotelluric inversion. *Geophysics* 66, 174-187.
- Rodi, W. L., Mackie, R. L., 2012. The inverse problem, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 347-414.
- Russell, D. J., Gale, J. E., 1982. Radioactive waste disposal in sedimentary rocks of southern Ontario. *Geoscience Canada* 9, 200-207.
- Sanford, B.V., 1993. St. Lawrence Platform: geology, in: Stott, D. F., Aitkin, J. D. (Eds.), *Sedimentary cover of the craton of Canada*, Geological Survey of Canada, *Geology of Canada Series* 5, pp. 723-786.
- Schmoldt, J.-P. 2011. Multidimensional isotropic and anisotropic investigation of the Tajo Basin subsurface: A novel anisotropic inversion approach for subsurface cases with oblique geoelectric strike directions. PhD thesis, National University of Ireland, Galway.
- Schwarz, G., 1990. Electrical conductivity of the earth's crust and upper mantle. *Surv. Geophys.* 11, 133 -161.
- Selway, K., 2014. On the causes of electrical conductivity anomalies in tectonically stable lithosphere. *Surv. Geophys.*, 57, 219-257, doi 10.1007/s10712-013-9235-1
- Simpson, F., Bahr, K., 2005. *Practical Magnetotellurics*. Cambridge University Press, Cambridge, United Kingdom, 254 pp

- Sleep, N. H., 2009. Stagnant lid convection and carbonate metasomatism of the deep continental lithosphere. *Geochem. Geophys. Geosyst.* 10, Q11010.
- Sleep, N.H., 1971. Thermal Effects of the Formation of Atlantic Continental Margins by Continental Break up. *J. R. Astr. Soc.* 24, 325-350.
- Sleep, N.H., Nunn, J.A., 1980. Flexure of a rheologically stratified lithosphere, in: Nemat-Nasser. S. (Ed.), *Solid Earth Geophysics and Geotechnology*, Applied Mechanics Division, American Society of Mechanical Engineers 42, pp. 53–60.
- Sleep, N.H., Snell, N.S., 1976. Thermal contraction and flexure of mid-continent and Atlantic marginal basins. *J. R. Astr. Soc.* 45, 125-154.
- Sloss, L. L., 1988. Tectonic evolution of the craton in Phanerozoic time, in: Sloss, L. L. (ed.), *In Sedimentary cover-North American craton.*, Geological Society of America, volume D-2, pp. 25–51.
- Spratt, J. E., Jones, A. G., Jackson, V. A., Collins, L., Avdeeva, A., 2009. Lithospheric geometry of the Wopmay orogen from a Slave craton to Bear Province magnetotelluric transect. *J. Geophys. Res.* 114, B01101.
- Vozoff, K., 1991. The magnetotelluric method, in: Nabighian, M. N. Ed.), *Electromagnetic methods in applied geophysics*, Society of Exploration Geophysicist, pp. 641 – 711.
- Wallach, J.L., Mohajer, A.A., Thomas, R.L., 1998. Linear zones, seismicity, and the possibility of a major earthquake in the intraplate western Lake Ontario area of Eastern North America. *Can. J. Earth Sci.* 35, 762-786.
- Wannamaker, P. E., Chave, A. D., Booker, J. R., Jones, A.G., Filloux, J. H., Ogawa, Y., Unsworth, M., Tarits, P., Evans, R., 1996. Magnetotelluric Experiment Probes Deep Physical State of Southeastern U.S., *EOS Trans. AGU*, 329, 332–333.
- Wannamaker, P.E., 2005. Anisotropy versus heterogeneity in continental solid earth electromagnetic studies: fundamental response characteristics and implications for physicochemical state. *Surv. Geophys.* 26, 733 – 765.
- Weidelt, P., Chave, A. D., 2012. The magnetotelluric response function, in: Chave, A. D., Jones, A. G. (Eds.), *The Magnetotelluric Method: Theory and Practice*, Cambridge University Press, New York, pp. 122-162.
- White, D.J., Forsyth, D. A., Asudeh, I. A., Carr, S. D., Wu, H., Easton, R. M., Mereu, R. F., 2000. A seismic-based cross-section of the Grenville Orogen in Southern Ontario and western Quebec. *Can. J. Earth Sci.* 37, 183-192.
- Whitmeyer, S.J., Karlstrom, K.E., 2007. Tectonic model for the Proterozoic growth of North America. *Geosphere* 3, 220-259. doi: 10.1130/GES00055.1
- Yoshino, T., Manthilake, G., Matsuzaki, T., Katsura, T., 2008. Dry mantle transition zone inferred from the conductivity of wadsleyite and ringwoodite. *Nature* 451, 326-329.
- Zhang, J., Frederiksen, A.W., 2013. 3-D crust and mantle structure in Southern Ontario, Canada via receiver function imaging. *Tectonophysics* 608, 700-712.

Chapter 8

Summary and Future Work

8.1 Summary

As part of the POLARIS project, this work uses the magnetotelluric (MT) method and data, integrated with other geophysical and geological information, to investigate the crustal and lithospheric structure as well as tectonic history of the Grenville Province. A total of 89 broadband and long-period MT sites, distributed across southern Ontario, are subdivided into four profiles and analyzed. Profile 1 to 3 are located to the north of Lake Ontario and Profile 4 is located in southwest Ontario. These sites consist of new MT data from the POLARIS southern Ontario array and earlier MT data sets collected during the LITHOPROBE Abitibi-Grenville transect (Chapter 4). Modern tensor decomposition techniques along with isotropic and anisotropic 2-D inversions were used to revise and significantly update the known resistivity and geological structure of this region. This research presents the first large-scale crustal and lithospheric resistivity images of the Proterozoic Grenville Province and its margin with the Archean Superior Province in southern Ontario. The results of this research have helped to understand the processes that have formed, preserved, and modified the crust and the lithosphere in southern Ontario. Three publications in refereed journals (Adetunji et al., 2014a; 2014b; 2014c) are being used to publish these results.

The dimensionality analysis was based on phase tensor and average Groom-Bailey (GB) rms misfit values. The results show that the MT responses, on all profiles and at all depths, around the Grenville Front (GF) and in the Central Gneiss Belt (CGB) are generally compatible with the presence of 2-D resistivity structures. The reasonable spatial consistency of the most conductive direction, at upper lithospheric depths and deeper mantle layers, provides further evidence of two-dimensionality (Adetunji et al.,

2014b). These results provide justification for the application of a 2-D modelling and inversion approach. In addition, the long-period induction arrow responses (from profiles 1 and 3) provide further support for a 2-D modelling approach. Farther to the southwest, in the southeastern part of the CGB and within the Central Metasedimentary Belt (CMB), higher phase tensor skew and GB misfits suggest the presence of more complex 3-D structures (Chapter 5, 6, and 7; Adetunji et al., 2014a, 2014b; 2014c).

In this study, a depth-based approach was used for the geoelectric strike analysis. Based on *a priori* information, responses were divided into those corresponding to the crust (1-40 km), the upper lithospheric layer (45-150 km) and deeper mantle depths (>200 km). The results show geographically consistent, but different, strike azimuths for the crust, the upper lithospheric mantle and the deeper mantle layer. Along Profile 1, the regional crustal geoelectric strike of N45°E (Chapter 5; Adetunji et al., 2014a) is approximately parallel to the geologic strike of the GF and the Central Metasedimentary Belt Boundary Zone (CMBBZ) in the study area. Local variations in the crustal strike azimuths, especially within the CMB, are attributed to variation in resistivity structure associated with polyfolded domains, and to the brittle to brittle-ductile thrust and normal faults that separate these domains (Chapter 5; Adetunji et al., 2014a).

A regional geoelectric strike azimuth of N85°E is determined for the upper lithospheric mantle depth of the northeastern part of the study area, along profiles 1-3 (Chapter 6; Adetunji et al., 2014b). The result agrees well with the lithospheric strike azimuth of N80°E defined earlier for the GF region (Mareschal et al., 1995). A regional strike azimuth of N8°W is determined for Profile 4 (Chapter 7; Adetunji et al., 2014c). This result is attributed to the change in the geologic strike of the tectonic structure in this

area. The strike results are based on a comprehensive analysis of the depth-based most-conductive direction, and four other measures that make use of the GB misfit for a single depth estimate at a single site.

For the purpose of comparing the geoelectric strike results with the absolute plate motion (APM) and seismic fast direction, the geoelectric strike direction in the deeper mantle layer was examined. Using the same methods as above, a regional geoelectric strike azimuth of N65°E is determined for the deeper mantle layer (Chapter 5 and 6; Adetunji et al., 2014a, 2014b). This result indicates that the most conductive direction of the deeper mantle layer is oblique to the seismic fast direction, Great Meteor hotspot trail and the APM direction. Therefore, the observed regional strike azimuth is interpreted to reflect lower mantle lithospheric structures (Chapter 6; Adetunji et al., 2014b).

In the southeastern part of the study area, the Canadian Shield is covered by the Phanerozoic sedimentary rocks. 1-D inversion models show that the resistivity structure includes more resistive rocks in the Precambrian to Lower Ordovician units overlain by conductive Upper Ordovician rocks. These units are in turn overlain by more resistive Silurian and Devonian rocks (Chapter 7; Adetunji et al., 2014c). The conductive layer resolved by the MT results is interpreted to be associated with the Upper Ordovician shale of the Queenston, Georgian Bay and Blue Mountain formations. It is also possible that the above package is being imaged as a single unit with the overlying Lower Silurian shales of Cabot Head and Manitoulin formations whenever they are present. The average integrated conductance of this conductive layer is 21 S (Chapter 7; Adetunji et al., 2014c). The presence of highly conductive shale layers, in the near surface, excludes resolution of crustal resistivity structure in the region.

The crustal resistivity structure was determined from the MT data along Profile 1 which extends from the Abitibi Subprovince of the Archean Superior Province, through the entire Grenville Province in southern Ontario (Chapter 5; Adetunji et al., 2014a). The two-dimensional resistivity model, which is based on the crustal strike azimuth, delineates several crustal structures. The upper crust of the Abitibi Subprovince is resistive and the middle to lower crust is relatively conductive. The upper crust in the Pontiac Subprovince contains some areas of increased conductivity that are attributed to the metasedimentary component of the rocks in the area. The new MT results, from farther south in the Grenville Province, provide confirmation that the GF extends to at least the base of the crust. Based on the resistivity model, the CMBBZ is interpreted as a southeast-dipping feature that extends throughout the crust rather than rooting into a lower crust decollement as suggested by some earlier studies (e.g., White et al., 2000). Graphite-bearing rocks of the CMB are interpreted to be responsible for a lower crustal conductor imaged beneath the Frontenac terrane. Regionally, the crustal resistivity model includes very resistive Laurentian margin rocks dipping southeast to the base of the crust, bounded to the northwest by the GF, and to the southeast by the CMBBZ. This is in contrast to the observation elsewhere in the region of conductive mid to lower crust and requires revision of the interpretation that the enhanced conductivity reflects present day geodynamic conditions (Chapter 5; Adetunji et al., 2014a).

In Chapter 5 (Adetunji et al., 2014a), the horizontal lithospheric mantle conductor observed at a depth of 160 km beneath the southern Superior Province was interpreted to be the electrical lithospheric-asthenospheric boundary (eLAB). This interpretation is supported by the available seismic and petrological information for the area. However,

Abt et al. (2010), Yuan and Romanowicz (2010), and Yuan et al. (2011) suggest the existence of a MLD at approximately ~150 km depth beneath the Superior craton. In addition, results from Chapter 6 of this study indicate that the regional lithospheric thickness in the Grenville Province is 280 km (Adetunji et al., 2014b). Therefore, it is more reasonable to interpret the 160 km deep horizontal lithospheric conductor, in the southern Superior Province, as a mid-lithospheric feature. Such an interpretation is consistent with that of the increased conductivity in the lower lithosphere of the Grenville Province.

A sub-vertical lithospheric conductor at 70 – 150 km, beneath the Pontiac Subprovince, and along strike with the Kirkland Lake and Cobalt kimberlite fields, is also defined. The conductor lies above the diamond-graphite stability field at 140 -150 km depth and may therefore be caused by graphite. It is attributed to the re-fertilization of the mantle scar, initially created by an early thermomechanical event responsible for the formation of the Abitibi Greenstone Belt, by the alkaline-rich Cretaceous Great Meteor hotspot plume. Our results support the occurrence of a complex interaction between the postulated mantle plume and the base of the lithosphere. Melt could accumulate around the LAB interface to form unstable pockets that migrate vertically through sub-vertical zones of weakness (e.g., Faure et al., 2011). Such interaction is generally controlled by the geometry of the LAB or the lithospheric zones of weakness. It is suggested that the diamondiferous kimberlites being produced in the Kirkland Lake and Cobalt fields passed through this part of the lithosphere (Chapter 5; Adetunji et al., 2014a).

A lithospheric conductor that extends from about 100 to 150 km depth beneath the CMB is imaged on Profile 4 beneath the lower Great Lakes region (Chapter 7;

Adetunji et al., 2014c). The conductor is along strike with a similar one imaged by Adetunji et al. (2014a) on Profile 1. The observation suggests that very conductive lithosphere exists extensively beneath the CMB in the Grenville Province. The source of enhanced conductivity in this conductor is attributed to grain boundary graphite films. The source of the graphite is interpreted to have been either the fluids associated with the passage of the Cretaceous Great Meteor hotspot or carbon introduced during the pre-Grenvillian tectonism. The mobilization and emplacement of the anomaly could have been during Grenvillian orogenies or else associated with the Cretaceous kimberlitic magmatic process (Chapter 7; Adetunji et al., 2014c).

The entire Grenville Province in southern Ontario is characterized by large-scale, laterally extensive, resistive lithosphere (Chapters 5, 6 and 7; Adetunji et al., 2014a, 2014b; 2014c). The top of this feature is at about 45 km depth and it dips southeast along Profile 1. The geometry of its base, on profiles 2 and 3, shows that the true thickness of the lithosphere in this region is about 280 km (Chapter 6; Adetunji et al., 2014b). This resistivity anomaly is interpreted as the Superior Province lithosphere as on Profile 1, it extends for about 300 km southeast of the GF from the exposed Superior Province and along strike for at least 800 km. This Archean lithospheric root provided the stable basement for the Grenville orogen (Ludden and Hynes, 2000). The geoelectric conductive direction measured in the upper lithospheric mantle of the Grenville Province is interpreted to be the Archean fabric of the Superior lithosphere.

It is suggested that the narrow zone of conductive mantle that separates the lithospheric resistors on Profile 4 in southwest Ontario (Chapter 7; Adetunji et al., 2014c) is connected to the more conductive mantle that separates two parts of lithospheric

resistor on Profile 3 north of Lake Superior (Chapter 6; Adetunji et al., 2014b) and to the conductive lower lithosphere on adjacent Profile 1 (Chapter 5; Adetunji et al., 2014a). Extending the interpretation from Profile 1, the connected region of the conductive mantle is therefore attributed to the refertilization of the mantle lithosphere by fluids associated with the Cretaceous kimberlite magmatism. The results provide evidence for refertilization of a large volume of the lower mantle lithosphere by the passage of fluids. The enhanced conductivity is interpreted to be associated with the presence of water in nominally anhydrous minerals. A water content of 50 wt ppm was estimated based on average crustal heat flow, average resistivity at 180 km depth and temperature of 1220°C. The lack of any surface gravity anomalies or surface topography mirroring the spatial variations in the base of the lithospheric resistor suggests that the refertilization has produced minimal changes to the density and thermal properties of the lower lithosphere. It is hypothesized that the strike direction measured in the deeper mantle of the study area was established in the Cretaceous and potentially records the plate motion at that time (Chapter 6; Adetunji et al., 2014b).

Taken as a whole, the MT data set considered in this study shows the presence of several 3-D geoelectric elements in the region. These elements include the different strike directions for the crust, upper lithospheric mantle and deeper mantle layer, the change in the strike direction between the northwest and southeast parts of profiles 1, 2 and 3 at upper lithospheric mantle depth, and the fact that the observed dominant strike direction for profiles 1, 2 and is different from that of Profile 4. The changes between profiles occur over distance scales that are less than the length of each profile indicating, that although the MT responses are locally 2-D, there are 3-D structures present at a very

large scale. In addition, the resistivity models show along-strike changes in the resistivity structure at lithospheric depths of 200 to 300 km, at a comparable scale to the distance between the profiles, again indicating 3-D geometry at large scales. Consequently, it is suggested that regional-scale 3-D modeling and inversion of the MT data be completed as part of ongoing studies of the data set. This work is expected to better delineate all large scale resistivity structures, e.g., improving the resolution of the continuity of resistive blocks and conductive zones as well as improving the resolution of smaller scale lithospheric mantle and crustal structures and confirming the 2-D inversion results.

The Archean-Proterozoic boundary (the GF) has been previously regarded as a classic example of lithospheric electrical anisotropy (e.g., Wannamaker, 2005; Marti, 2013). This interpretation was based on in part on examination of shorter period (<100 s) induction arrows. However, analysis of the induction arrows at long periods, from an area larger than in the previous studies, shows results consistent with the presence of macroscopic resistivity structures. In addition, 2-D anisotropic inversion results show that anisotropy is not required to fit the MT data. The maximum horizontal anisotropy in the resistive lithosphere is less than 10%, much less than the factor of 15 determined in earlier 1-D studies (Mareschal et al., 1995). The anisotropic inversion also indicates minimal vertical anisotropy within the lithospheric conductor to the north of the GF. The results show that the upper lithospheric mantle is devoid of any significant resistivity anisotropy and the observed MT response anisotropy is due to large-scale resistivity structure. We suggest that the Ji et al. (1996) model, in which the obliquity between the MT and seismic strikes is interpreted as being due to dextral shearing, is not valid. The obliquity between MT and seismic responses arises because the large scale structures in

the Superior lithosphere, interpreted to be controlling the geoelectric strike, are oblique to the GF. If dextral shearing occurred, it appears to have not caused any significant shape preferred electrical resistivity anisotropy as interpreted by Ji et al. (1996). The possible existence of resistivity anisotropy, farther north, within the Superior Province, is not excluded by this study. It is however suggested that the presence of such resistivity anisotropy be re-examined with contemporary 2-D isotropic and anisotropic inversion methods (Chapter 6; Adetunji et al., 2014b).

8.2 Future work

Suggested additional work to complete the analysis and interpretation of LITHOPROBE-POLARIS MT data in southern Ontario is as follows:

- Further collection and integration of high quality deep penetrating MT data between, and on, the existing profiles will increase the data density. The new data will improve the resolution gaps between the defined resistive blocks and improve the resolution of smaller scale and crustal structures. The planned collection of new MT data in adjacent areas of United States as part of the USarray program will provide additional constraint for the interpretation of the lithospheric scale resistivity structure.
- 3-D modeling and inversion of the existing and newly collected MT data will help to better delineate all large scale resistivity structures in this area.
- Application of the results from the Paleozoic sedimentary rocks studies to geomagnetic hazards, an important objective of POLARIS project in southern Ontario. This work will involve calculating the surface electric fields needed for

modelling geomagnetically induced currents on powerlines and pipelines using the resistivity models of the sedimentary basin rocks.

- In addition, a robust lithospheric model of this region can be constructed using the LitMod approach which involves the simultaneous inversion of geophysical and petrological data (Afonso et al., 2013). This approach eliminates discrepancies that may be associated with the use of only one or two types of data for lithospheric modeling and interpretation.

8.3 References Cited

- Abt, D.L., Fischer, K.M., French, S.W., Ford, H.A., Huaiyu, Y., Romanowicz, B., 2010. North American lithospheric discontinuity structure imaged by Ps and Sp receiver functions. *J. Geophys. Res.* 115 B09301, doi:10.1029/2009JB006914.
- Adetunji, A.Q., Ferguson, I.J., Jones, A.G., 2014a. Crustal and lithospheric scale structures of the Precambrian Superior-Grenville margin. *Tectonophysics* 614, 146-169, <http://dx.doi.org/10.1016/j.tecto.2013.12.008>.
- Adetunji, A.Q., Ferguson, I.J., Jones, A.G. 2014b. Imaging the lithosphere of the Grenville Province: large-scale structures and electrical anisotropy. *Geophys. J. Int.* (submitted in Jan. 2014).
- Adetunji, A.Q., Ferguson, I.J., Jones, A.G. 2014c. Magnetotelluri imaging of the lower Great Lakes region. *Geophys. J. Int.* (submitted in Jan. 2014.)
- Afonso, J. C., Fulla, J., Griffin, W. L., Yang, Y., Jones, A. G. , Connolly, , J. A. D., O'Reilly, S. Y., 2013. 3-D multiobservable probabilistic inversion for the compositional and thermal structure of the lithosphere and upper mantle. I: a priori petrological information and geophysical observables. *J. Geophys. Res.* 118, doi:10.1002/jgrb.50124.
- Faure S., Godey, S., Fallara, F., Trepanier, S., 2011. Seismic architecture of the Archean North American mantle and its relationship to diamondiferous kimberlite fields. *Econ. Geol.* 106, 223-240.
- Ji, S., Rondenay, S., Mareschal, M., Senechal, G., 1996. Obliquity between seismic and electrical anisotropies as a potential indicator of movement sense for ductile shear zones in the upper mantle. *Geology* 24, 1033-1036.
- Ludden, J., Hynes, A., 2000. The Lithoprobe Abitibi-Grenville transect: two billion years of crust formation and recycling in the Precambrian Shield of Canada. *Can. J. Earth Sci.* 37, 459-476.
- Mareschal, M., Kellett, R. L., Kurtz, R. D., Ludden, J. N., Ji, S., Bailey, R. C., 1995. Archean cratonic roots mantle shear zones and deep electrical anisotropy. *Nature London* 375, 134-137.
- Marti, A., 2013. The role of electrical anisotropy in magnetotelluric responses: From modelling and dimesionality analysis to inversion and interpretation. *Surv. Geophys.*, doi: 10.1007/s10712-013-9233-3.
- Wannamaker, P.E., 2005. Anisotropy versus heterogeneity in continental solid earth electromagnetic studies: fundamental response characteristics and implications for physicochemical state. *Surv. Geophys.* 26, 733 – 765.
- White, D.J., Forsyth, D. A., Asudeh, I. A., Carr, S. D., Wu, H., Easton, R. M., Mereu, R. F., 2000. A seismic-based cross-section of the Grenville Orogen in Southern Ontario and western Quebec. *Can. J. Earth Sci.* 37, 183-192.
- Yuan, H., Romanowicz, B., 2010. Lithospheric layering in the North American craton, *Nature* 466, 1063-1069, doi:10.1038/nature09332.
- Yuan, H., Romanowicz, B., Fischer, K.M., Abt, D., 2011. 3-D shear wave radially and azimuthally anisotropic velocity model of the North American upper mantle. *Geophys. J. Int.* 184, 1237-1260, doi:10.1111/j.1365-246X.2010.04901.x

Appendix A

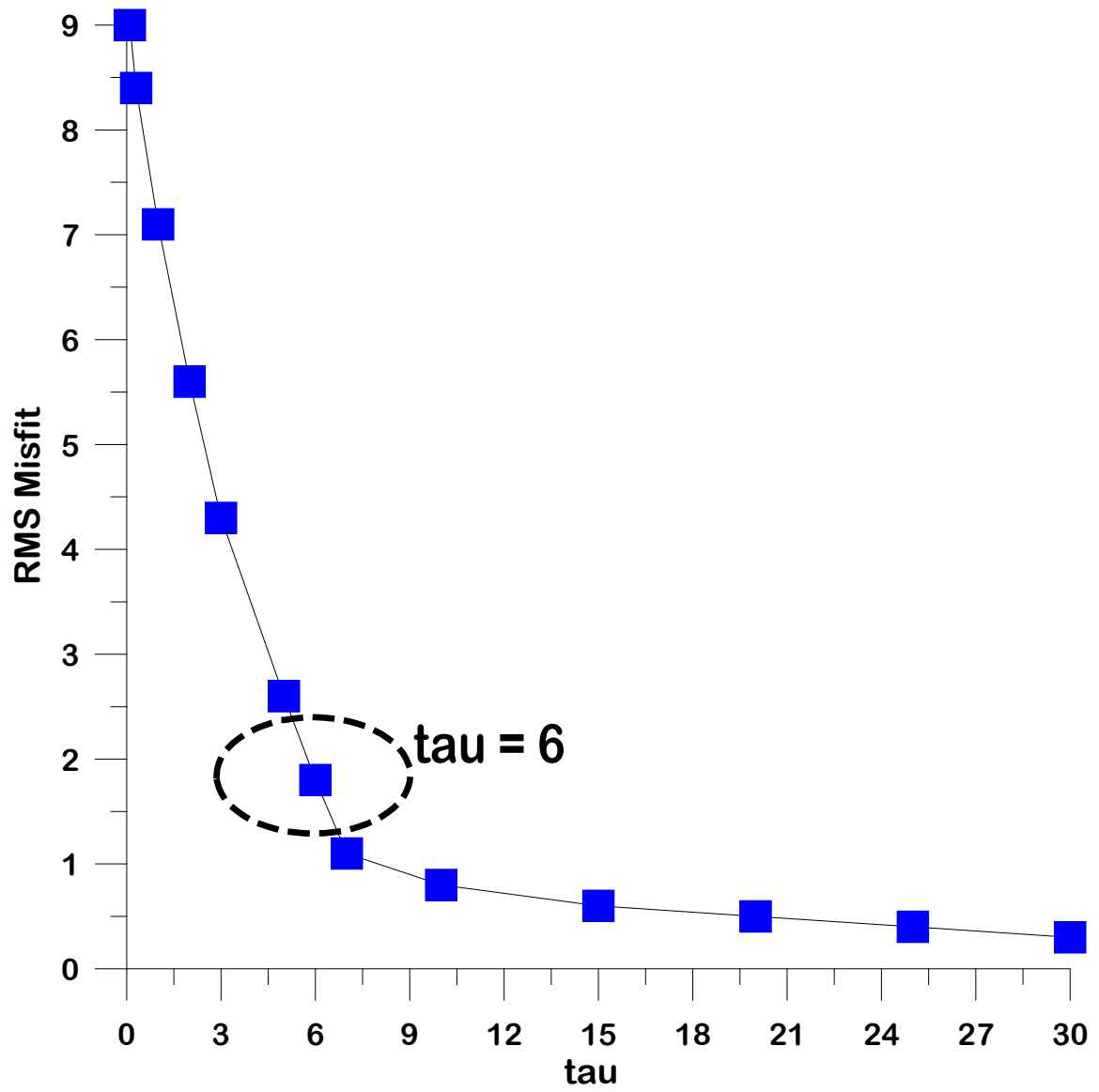


Figure A.1. L-curve for 2-D inversion inversion models of Profile 1 showing RMS misfit versus a range of tau (τ) values.

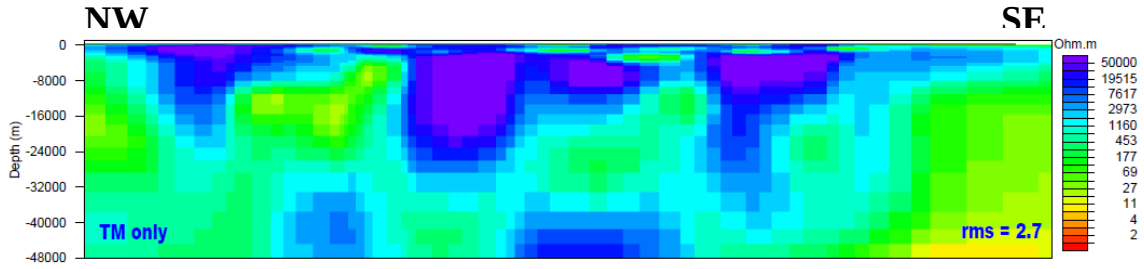


Figure A.2. 2-D resistivity section of the crust obtained by inversion of only TM responses for the crustal data set along Profile 1 (V.E = 5). The regularization parameter tau of 6 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion.

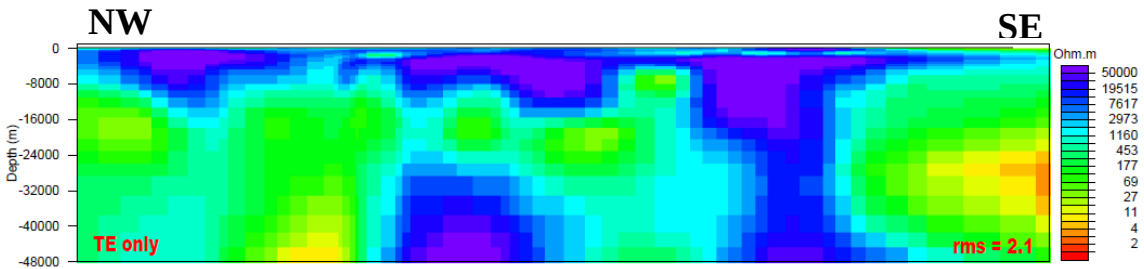


Figure A.3. 2-D resistivity section of the crust obtained by joint inversion of only TM responses for the crustal data set along Profile 1 (V.E = 5). The regularization parameter tau of 6 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion.

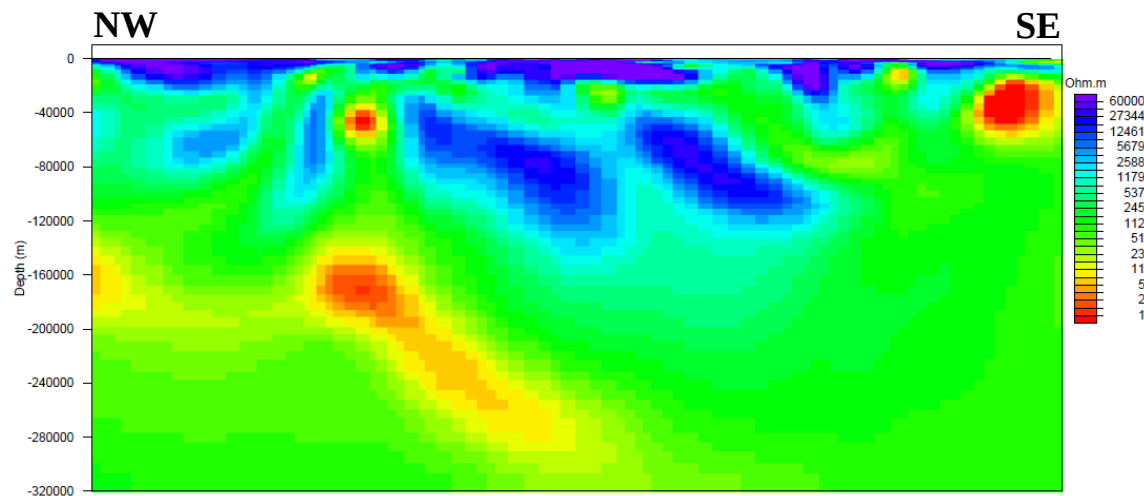


Figure A.4. 2-D resistivity model derived by joint inversion of TE and TM responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 3 was used to derive this model. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

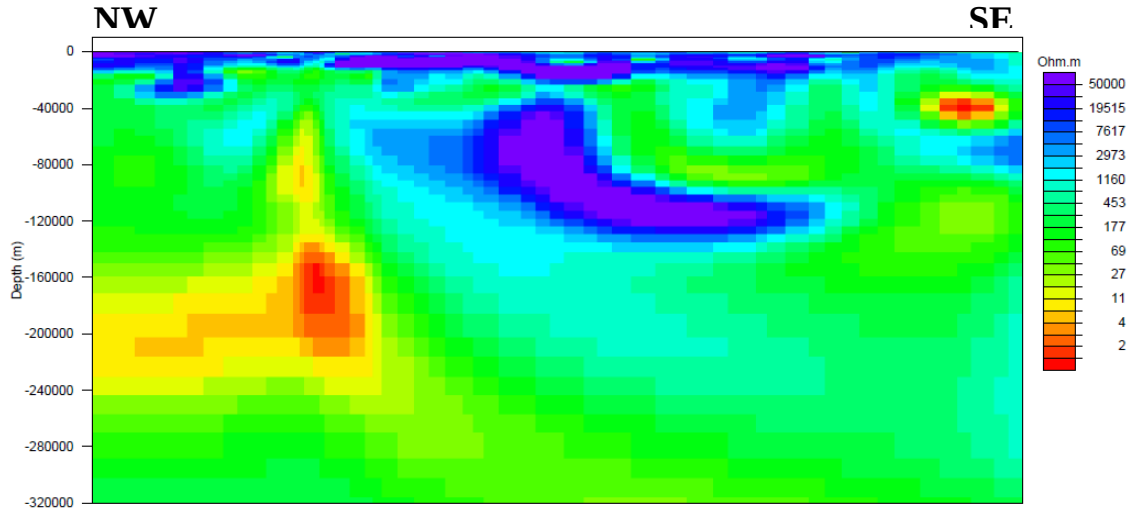


Figure A.5. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 1 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

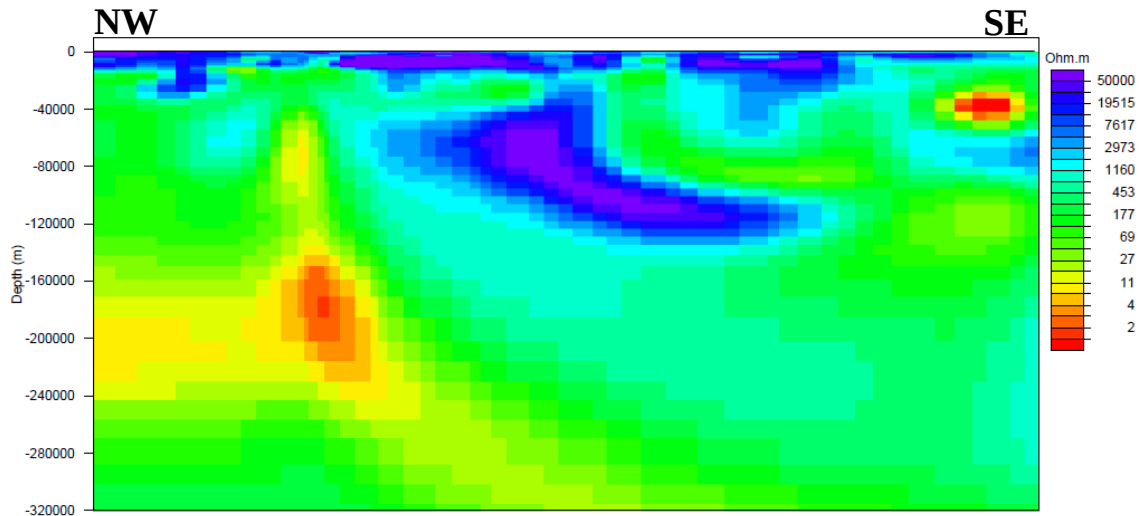


Figure A.6. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 2 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

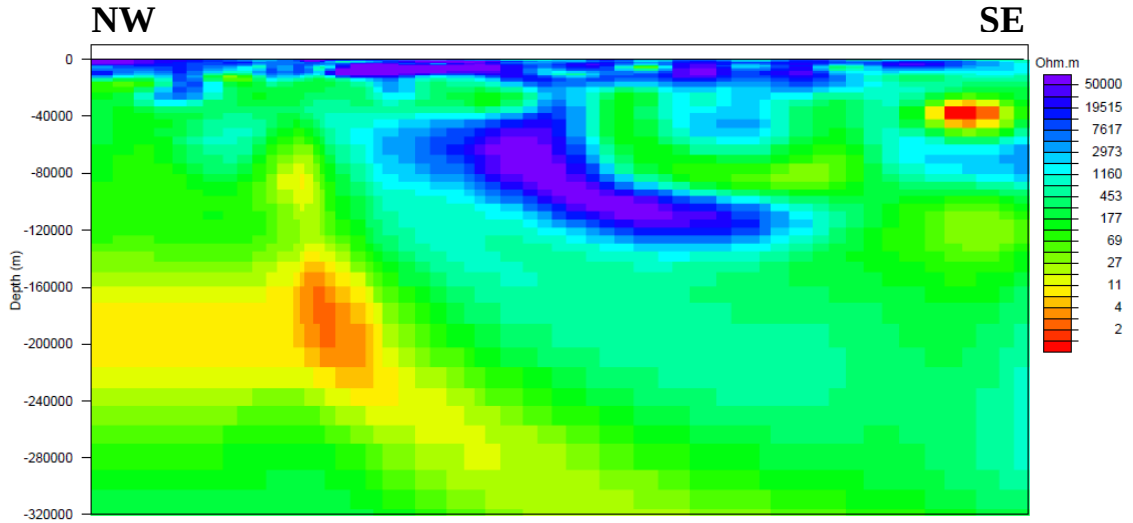


Figure A.7. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 3 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

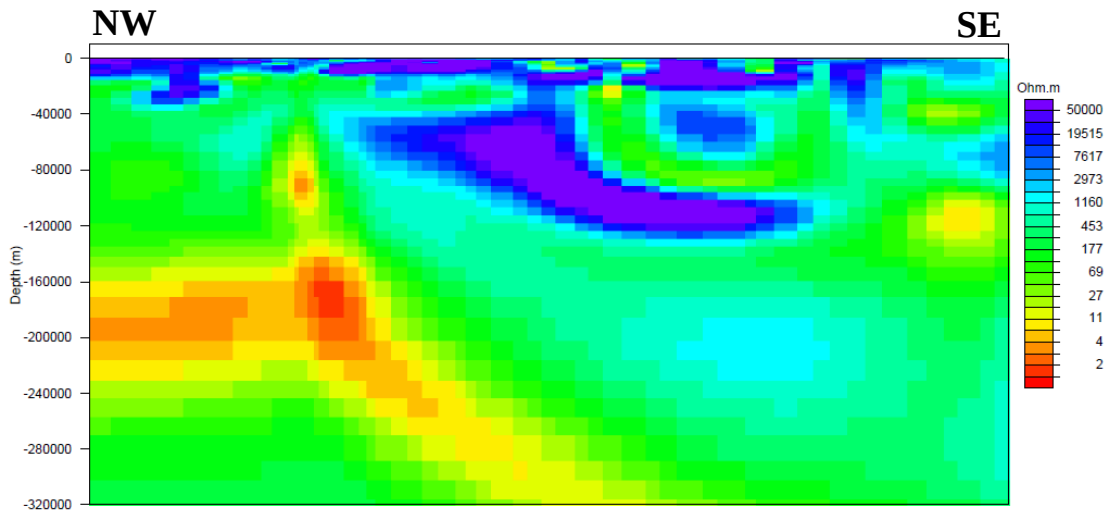


Figure A.8. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 5 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

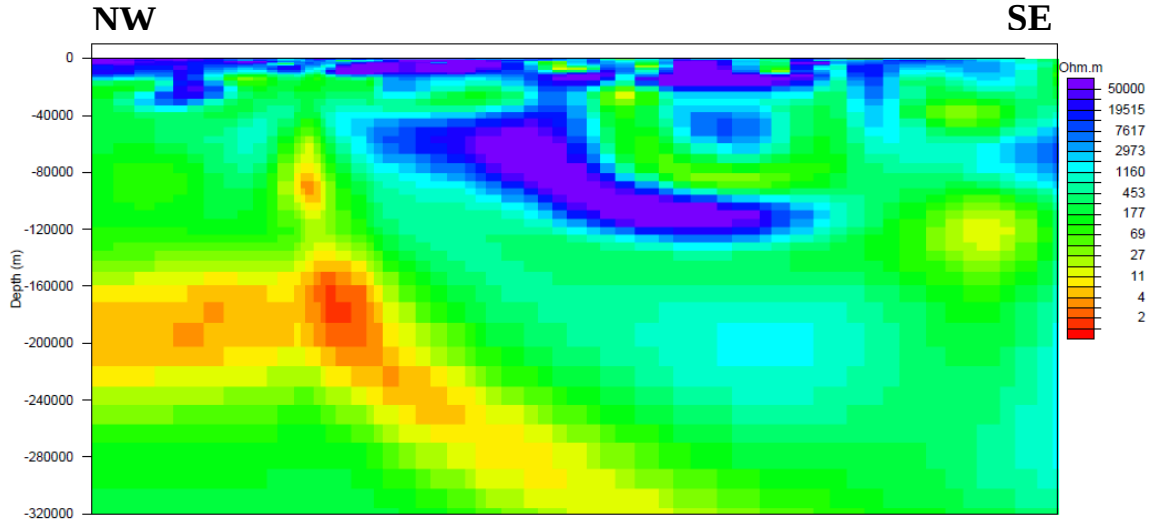


Figure A.9. 2-D resistivity model derived by joint inversion of TE, TM and tipper responses for the lithospheric data set along Profile 1 (V.E = 1). The regularization parameter tau of 8 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

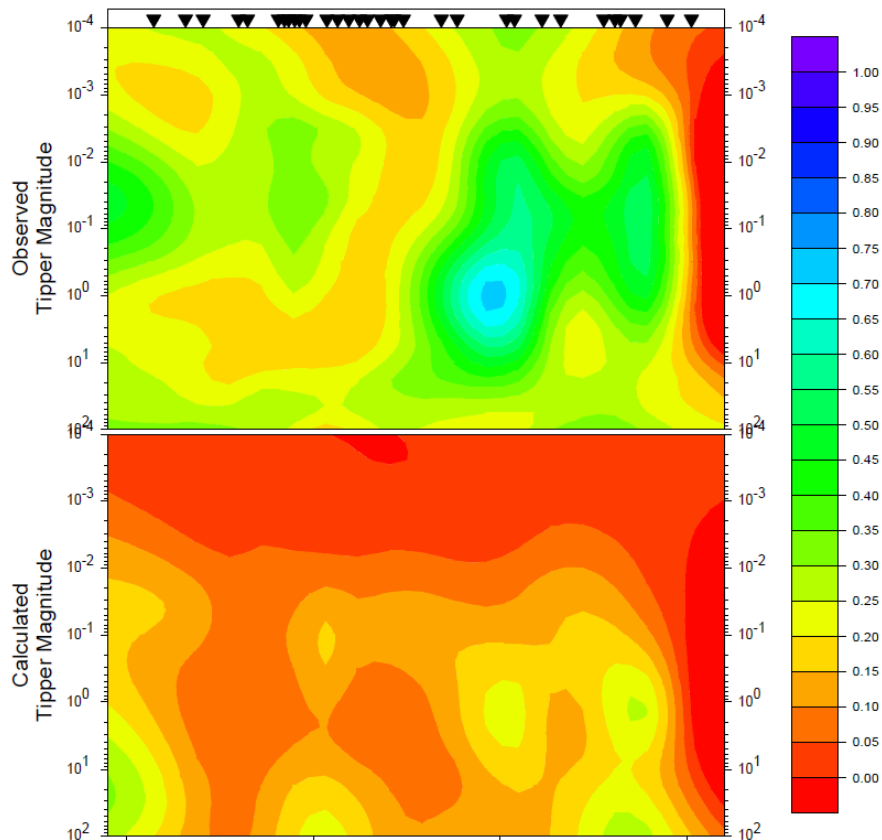


Figure A.10. The pseudosection of the observed and calculated tipper magnitude for the crustal model along Profile 1.

Appendix B

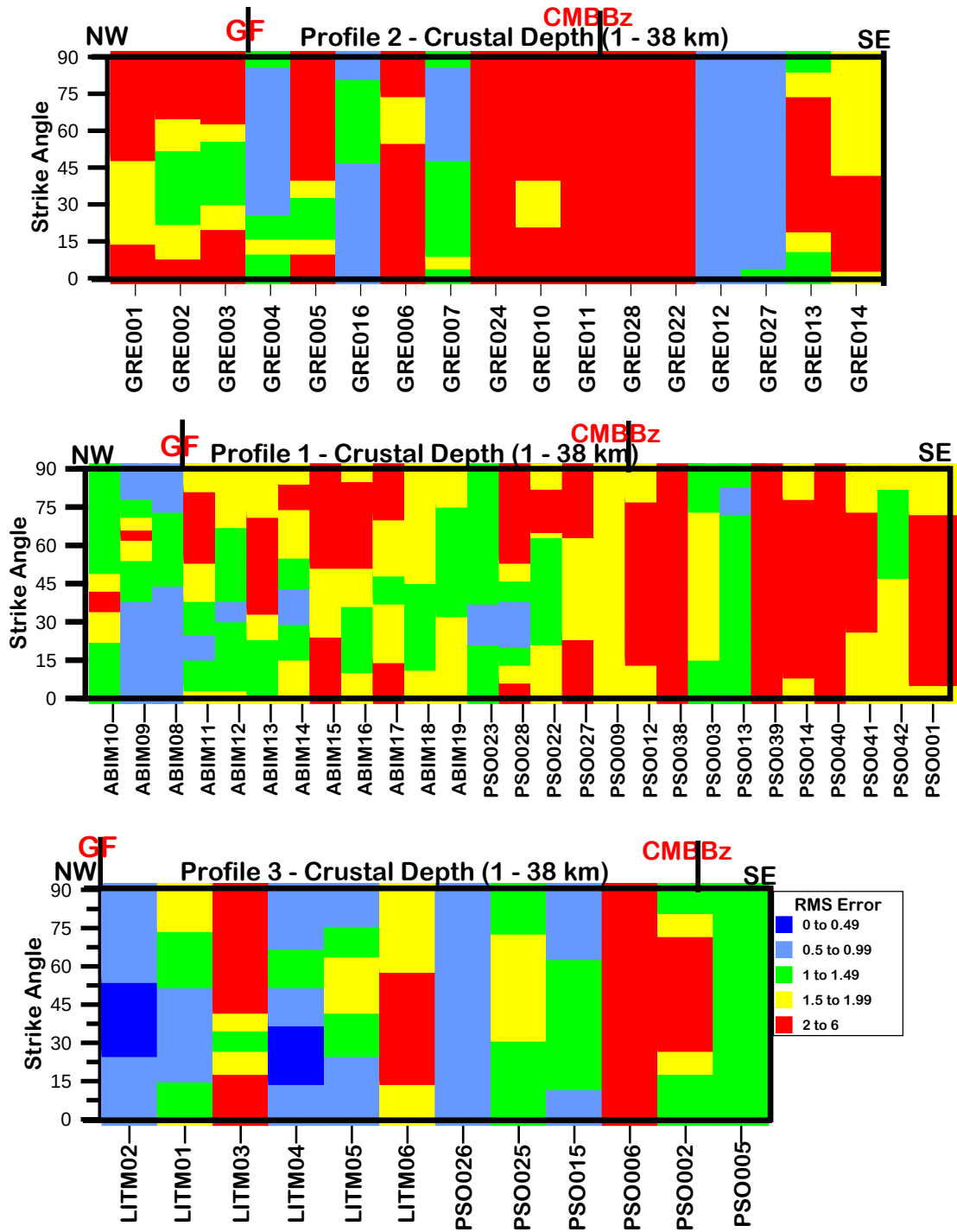


Figure B.1. RMS misfit for single sites versus strike azimuth for a GB fit to the crustal depth band (1–38 km). Results are plotted in order of the position of each site on the profile but the inter-site distance is not preserved.

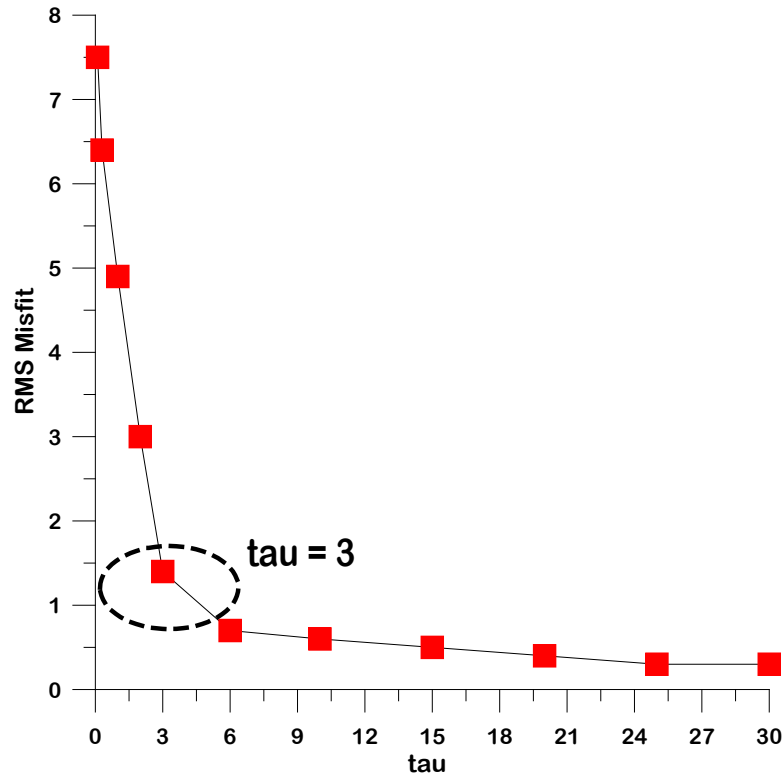


Figure B.2. L-curve for 2-D inversion models of Profile 2 showing RMS misfit versus a range of tau (τ) values.

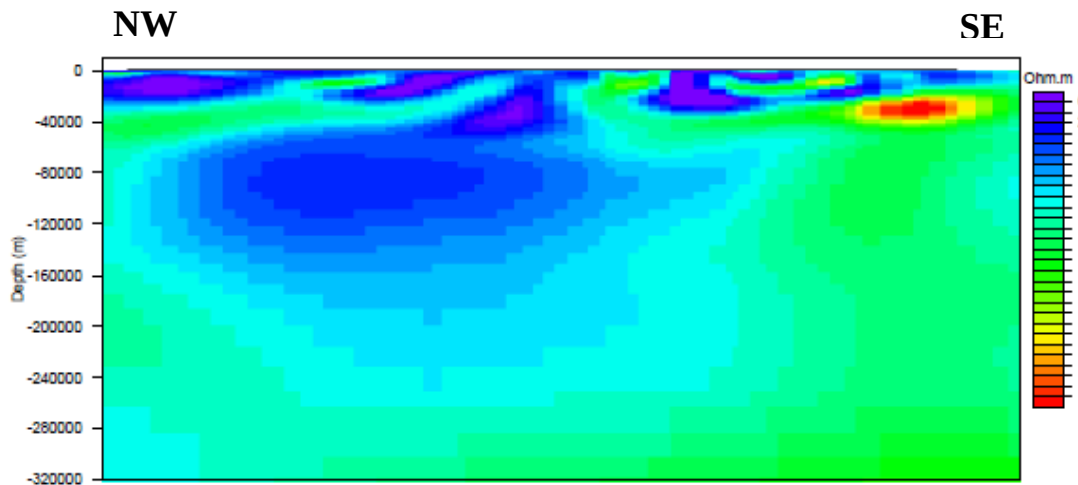


Figure B.3. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set ($V.E = 0.5$). The regularization parameter tau of 1 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

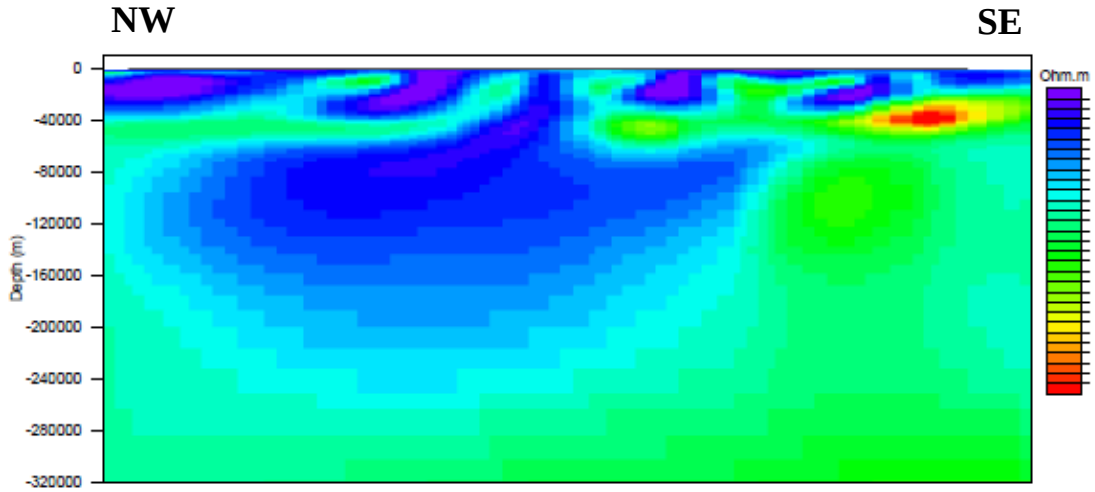


Figure B.4. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 2 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

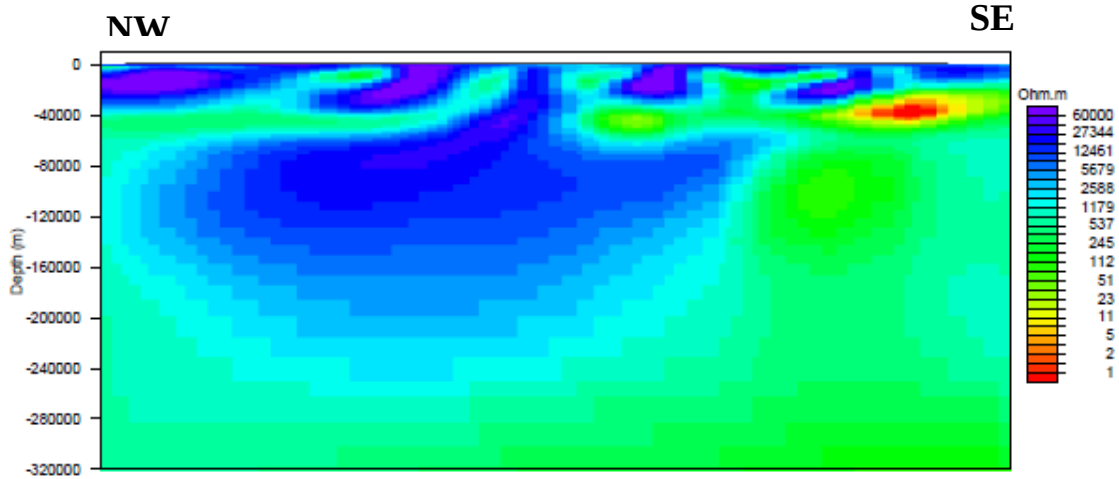


Figure B.5. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 4 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

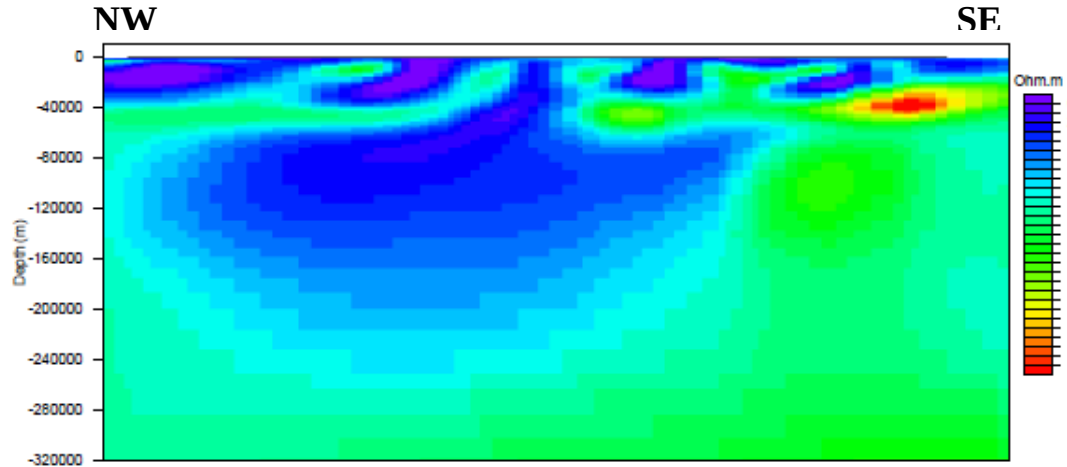


Figure B.6. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 5 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

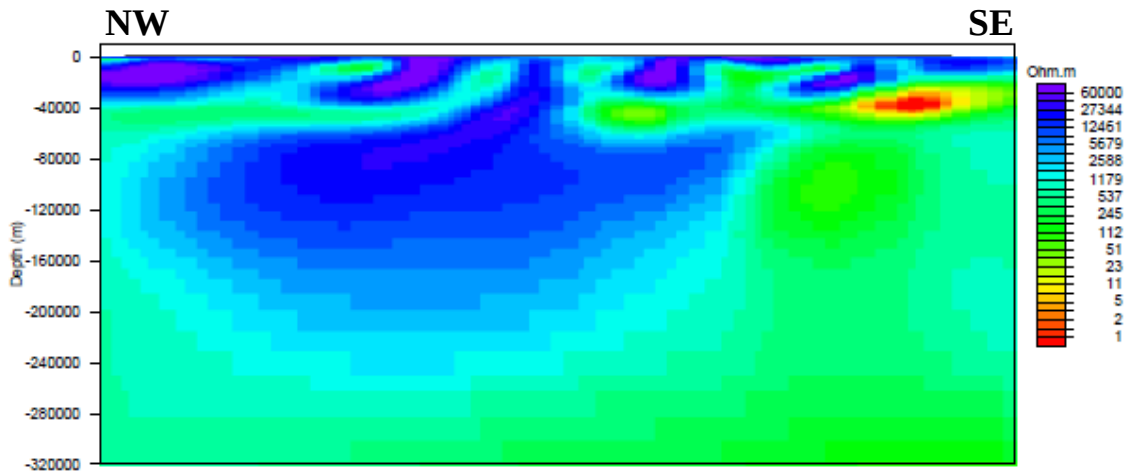


Figure B.7. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 7 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

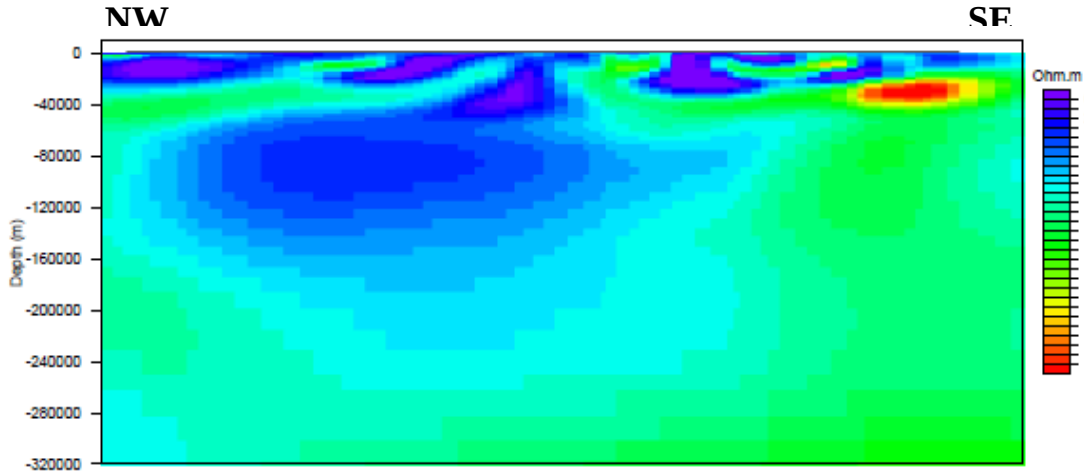


Figure B.8. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 10 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

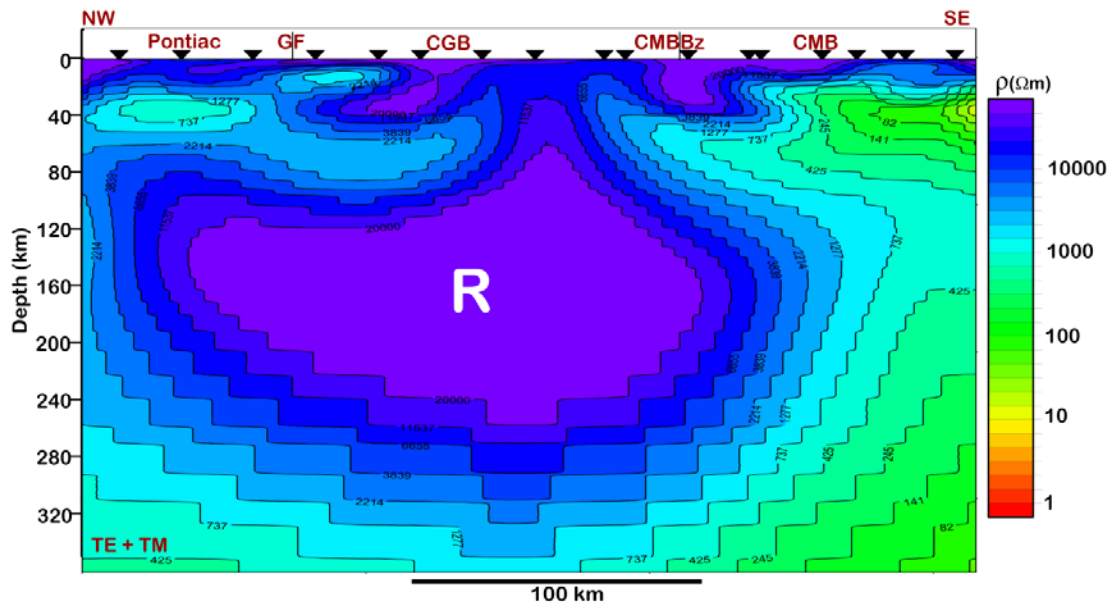


Figure B.9. Gridded resistivity model derived by joint inversion of TE and TM responses of Profile 2 for the lithospheric data set ($V.E = 0.5$). The model is the same as Figure 6.13 but it displays contour lines that help to better delineate resistivity variations in the model

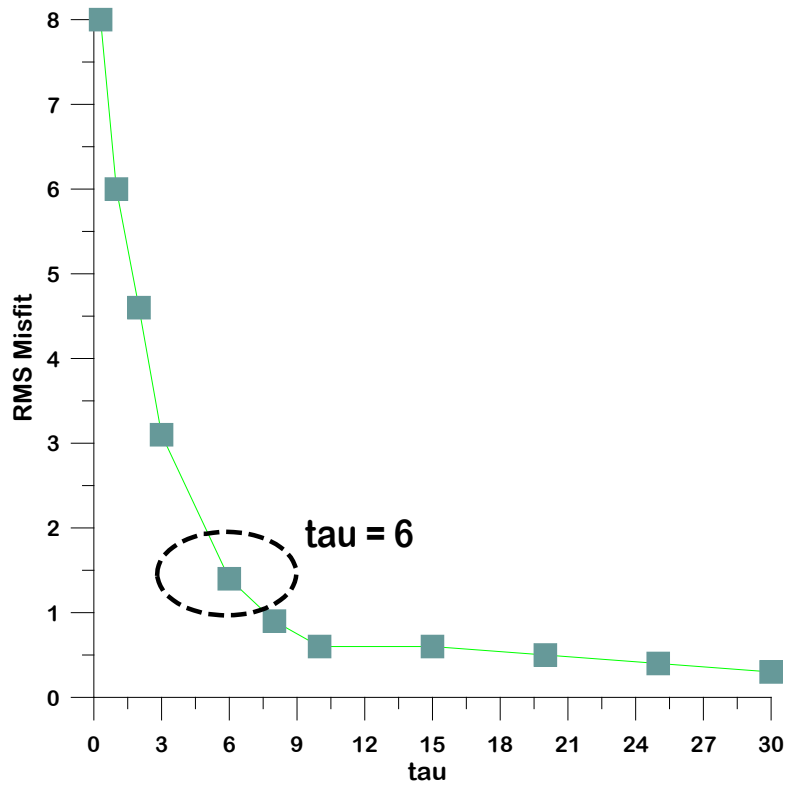


Figure B.10. L-curve for 2-D inversion models of Profile 3 showing RMS misfit versus a range of τ (τ) values.

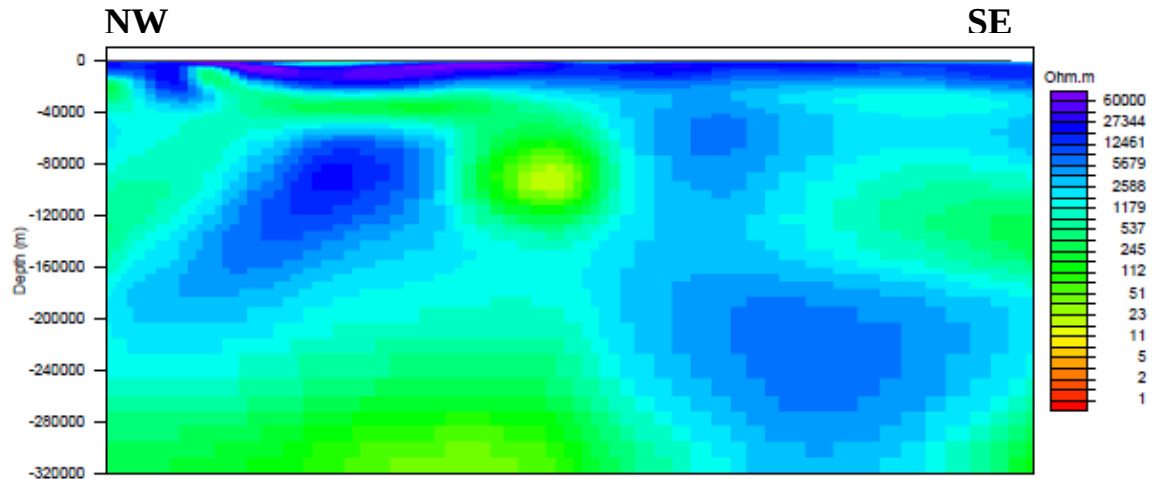


Figure B.11. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 1 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

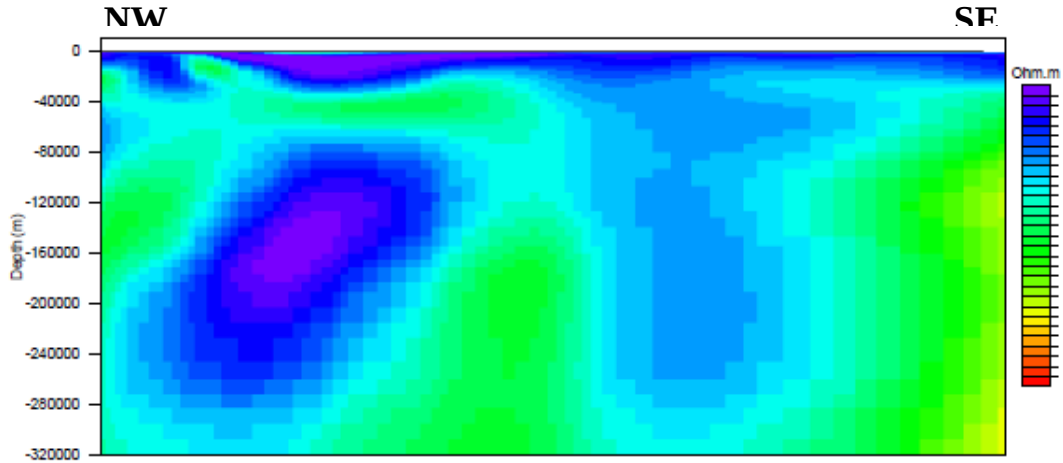


Figure B.12. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 2 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

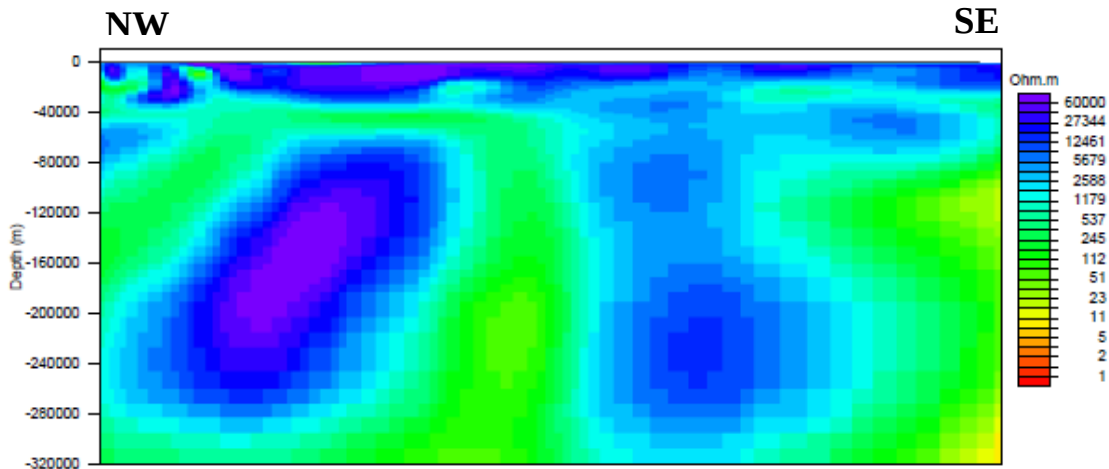


Figure B.13. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set (V.E = 0.5). The regularization parameter tau of 3 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

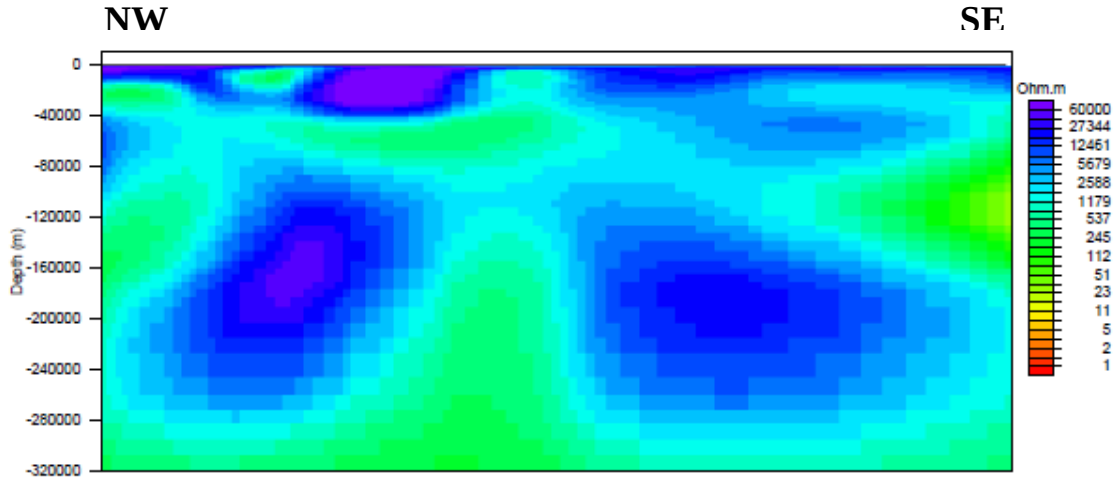


Figure B.14. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set (V.E = 0.5). The regularization parameter τ of 7 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

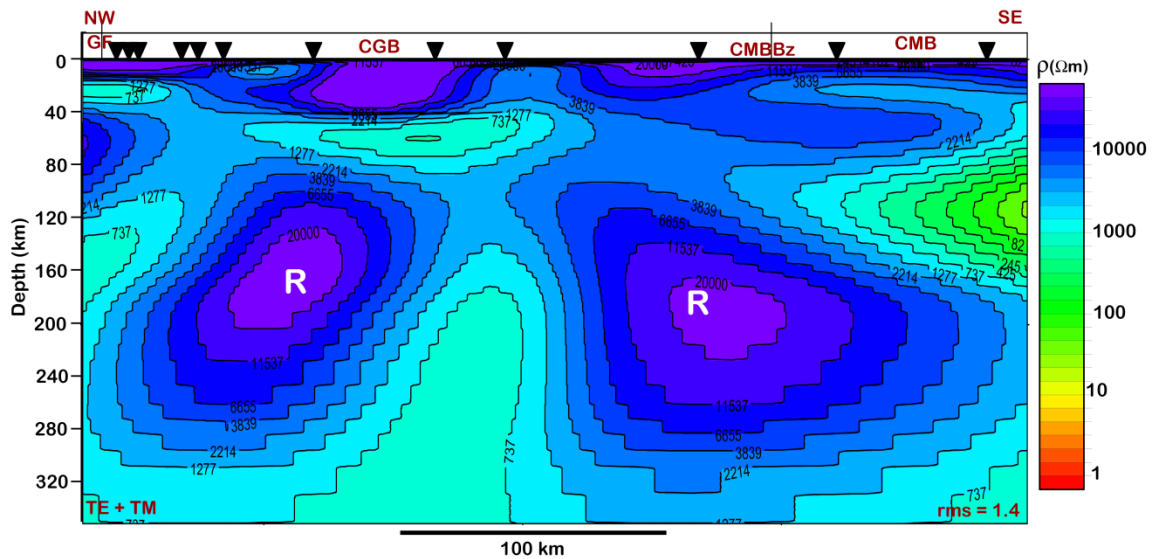


Figure B.15. Gridded resistivity model derived by joint inversion of TE and TM responses of Profile 3 for the lithospheric data set (V.E = 0.5). The model is the same as Figure 6.14 but it displays contour lines that help to better delineate resistivity variations in the model

Appendix C

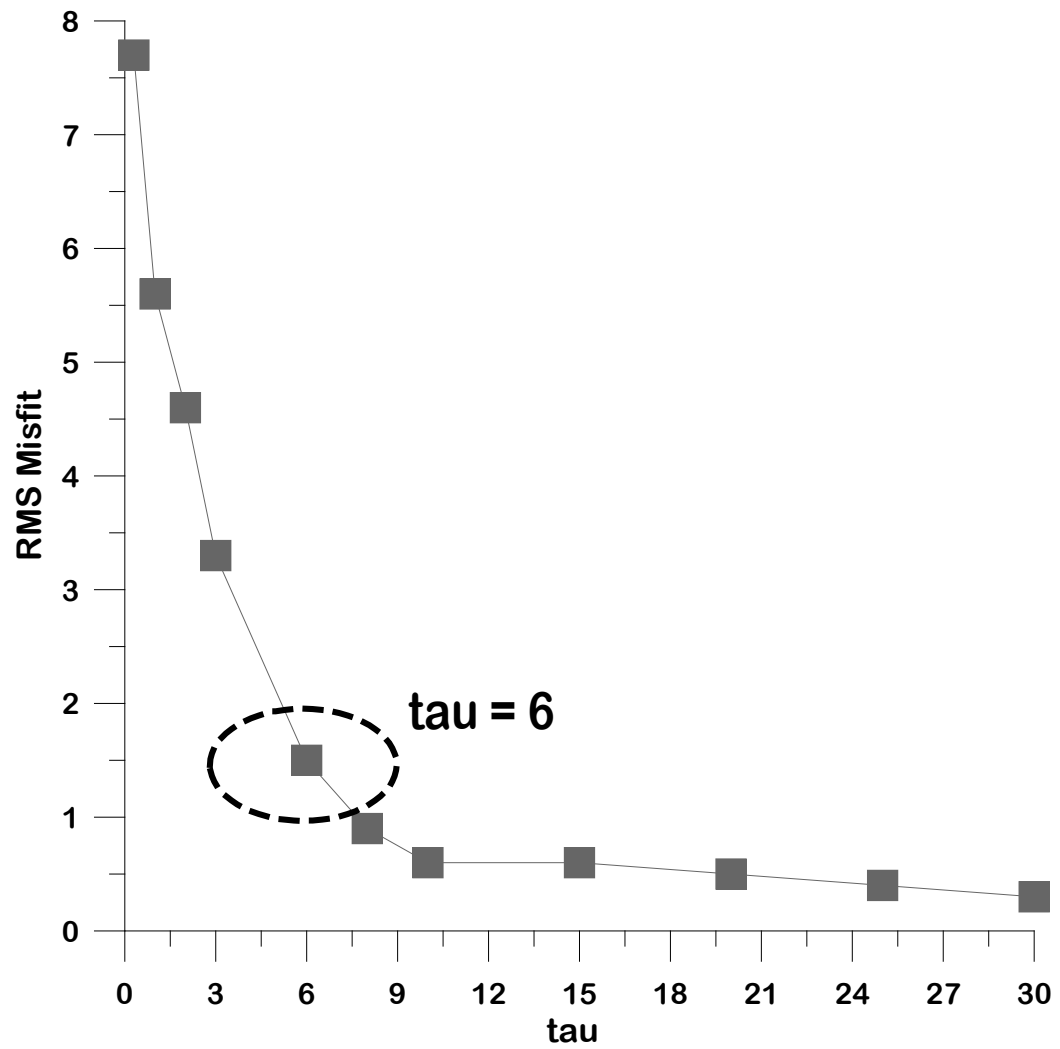


Figure C.1. L-curve for 2-D inversion inversion models of Profile 4 showing RMS misfit versus a range of tau (τ) values.

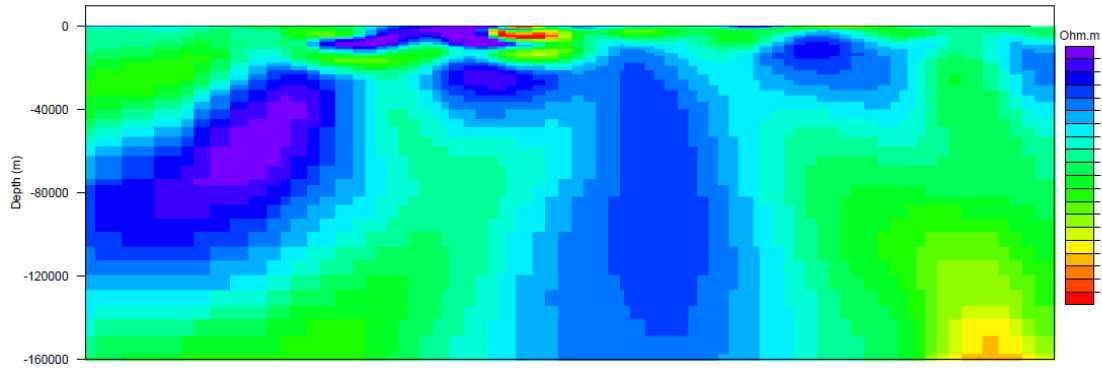


Figure C.2. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 1 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

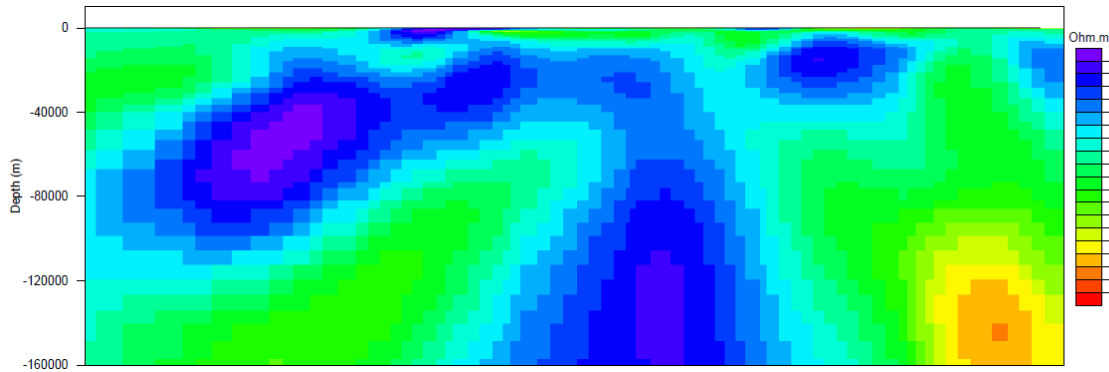


Figure C.3. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 3 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

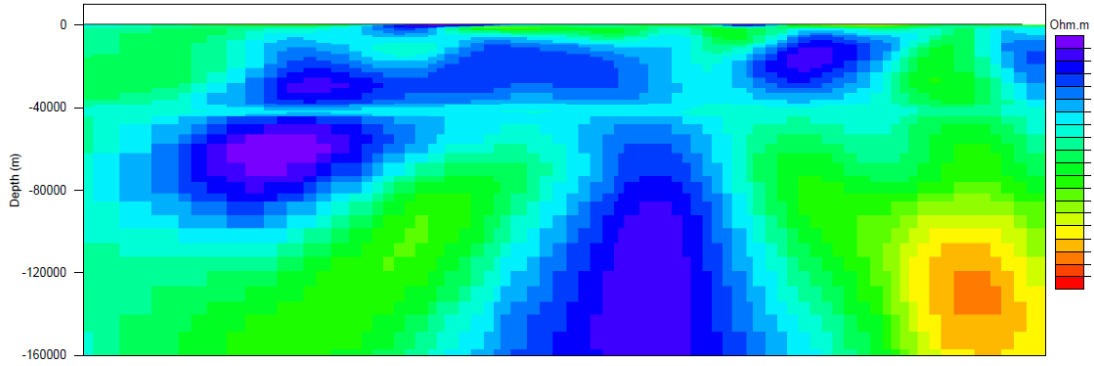


Figure C.4. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 5 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

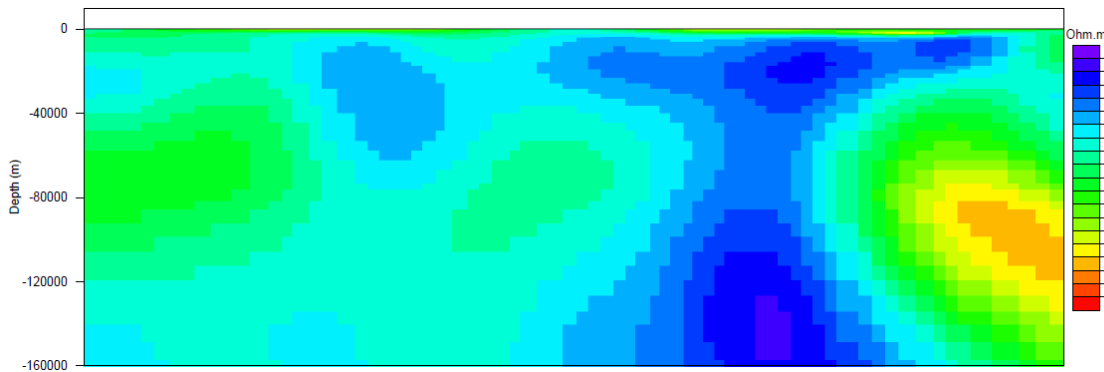


Figure C.5. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 7 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

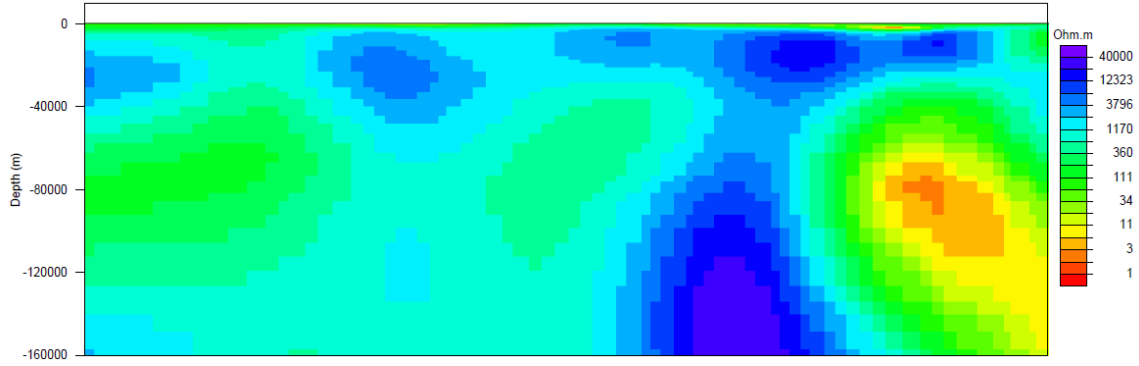


Figure C.6. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 8 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.

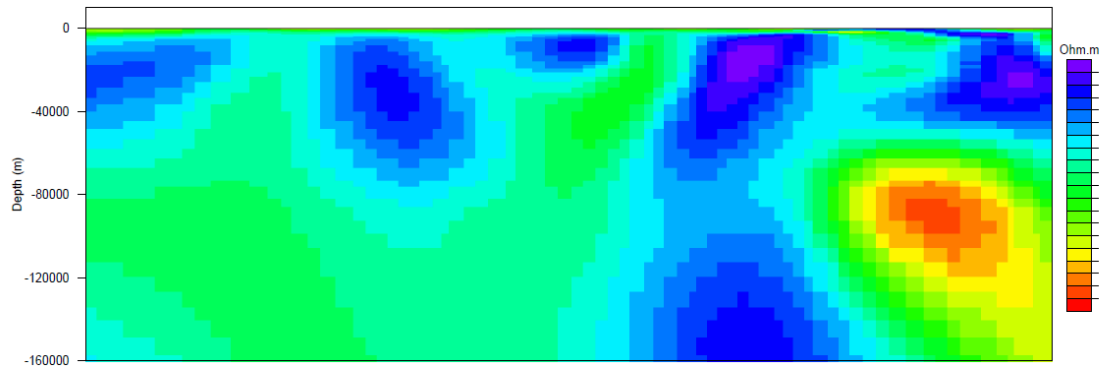


Figure C.7. 2-D resistivity model derived by joint inversion of TE and TM responses of Profile 4 for the lithospheric data set ($V.E = 0.5$). The regularization parameter τ of 10 was used to derive this model. Other weighting function parameters were set at $\alpha = 1$, $\beta = 0.3$, $H = 500$ and $V = 500$ for the inversion. The crustal section (upper 48 km) of the model is deemed incorrect as the projection of data for the 2-D inversion used a modelling azimuth that was not optimal for the crustal part of the response.