Determination of Near-Surface, Crustal and Lithospheric Structures in the Canadian Precambrian Shield Using Time-Domain Electromagnetic and Magnetotelluric Methods

by

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DOCTOR OF PHILOSOPHY

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Abstract

Two electromagnetic methods were used to analyse the geoelectric structure of the subsurface of regions of the Precambrian Shield in Canada: the magnetotelluric (MT) and time-domain electromagnetic (TEM) methods.

Magnetotelluric soundings were made at 60 sites in the southwestern Northwest Territories, Canada, along the LITHOPROBE SNORCLE Transect Corridor 1 and 1A, in the summer of 1996. The sites are located in southwestern Northwest Territories, Canada, between latitudes 60°-65° N and longitudes 110°-125° W, and cross the Archean Slave Province, the Proterozoic Buffalo Head, Great Bear Magmatic Arc, Hottah, Fort Simpson and Nahanni terranes, and the Great Slave Lake Shear Zone. Phanerozoic sedimentary rocks overlie the Proterozoic terranes.

Analysis of MT and well log data indicates that the conductive Phanerozoic sedimentary rocks have an approximately one-dimensional resistivity structure. They have a maximum thickness of about 1000 m at a location in the western Fort Simpson terrane (longitude 123° W) and decrease in thickness towards their margin in the western Great Bear Magmatic Arc (longitude 117° W).

The underlying Precambrian rocks have a two-dimensional structure with a strike of ~N35° E at crustal depths and of N60° E at upper mantle depths. The MT models image the resistive (>400 Ωm) Fort Simpson basin in the Fort Simpson terrane. The MT results show that the base of the basin is west-dipping (about 20°), and reaches a depth of ~20 km depth in the west. The basin extends at least 100 km along the survey line. The MT results support the interpretation of the geometry of the basin made from seismic reflection data.
A sub-vertical conductive (<100 Ωm) boundary between the Fort Simpson and Hottah terranes extends from several kilometres depth to the lower crust and mantle, and corresponds to structure in the seismic reflection response. This conductive body is interpreted as Hottah crust that was metamorphosed and deformed during collision of the Fort Simpson and Hottah terranes. There is a conductive zone (~30 Ωm) in the mid-lower crust beneath the boundary of the Hottah terrane and the Great Bear Magmatic Arc. The rocks beneath the Great Bear Magmatic Arc are interpreted to be the deformed and metamorphosed rocks of the Hottah-Slave transition (e.g. sedimentary rocks of Coronation Supergroup). The source of the enhanced conductivity could be either carbon grain-boundary films or conductive minerals.

The Great Slave Lake shear zone (GSLsz) is defined by a resistivity high and magnetic low, and corresponds to a 20 km mylonites zone. The geoelectric strike direction near the GSLsz varies with depth from approximately N30°E in the upper and middle crust to approximately N60°E in the upper mantle. Similarity of the MT and seismic SKS fast directions for the shallow structure provides evidence that the controls on these responses are related, and suggests that structural control may be an more important component of the seismic anisotropy in the crust. In the upper mantle, obliquity of the MT and seismic directions may provide a potential kinematic indicator around the GSLsz. The MT strike direction may relate to either the structures or anisotropy (or both).

The MT results reveal enhanced conductivity in the mantle along Corridor 1 and 1A. The depth to the 150 Ωm resistivity conductor is at ~100±20 km depth beneath the Nahanni, Fort Simpson, central Hottah, central Great Bear and Buffalo Head terranes, but shallows to ~40 km depth beneath the Hottah terrane. High heat flow (109 mW/m²) measured across the
Hottah, Fort Simpson and Nahanni terranes indicates that an uplift of the lithosphere-asthenosphere boundary could cause the broad-scale enhanced conductivity with the source of the high conductivity being partial melt in the asthenosphere. The more localized enhanced conductivity in the Hottah mantle could be due to either dissolved hydrogen, or carbon, or both in the mantle rocks.

The Whiteshell Research Area located near Pinawa, Manitoba, Canada has been used to investigate the concept of disposal of nuclear fuel waste in Precambrian rocks. It is located on Archean granitic rocks of the Lac du Bonnet Batholith (latitude 50°N, longitude 95°50’W), in the northwest part of the Winnipeg River Subprovince of the Superior Province of the Canadian Shield. The Lac du Bonnet TEM project is one component to evaluate the use of geophysical methods to investigate the properties of rock and groundwater.

In the Lac du Bonnet batholith, the fractures control the properties of the groundwater. The near-surface sub-vertical fractures provide pathways for fresh water to penetrate the granite, and enter groundwater flow systems in the sub-horizontal fracture zones. At greater depth, the relatively high salinity of groundwater is largely derived from dissolution of soluble salts in the host rock. The main object of this project is to map the fracture zones and fresh/saline water interface in Precambrian granitic rocks using the surface TEM method. The TEM surveys were completed at Sites B, D, URL and A. A GEONICS PROTEM47 system with a 100 m transmitter loop was used. The data were collected for receiver offsets ranging from 0-280 m on four sides of transmitter loop.

Analysis of the TEM and borehole log data indicates a basic three-layer structure: a thin conductive surface layer, a thick resistive second layer with an embedded conductive layer at some stations, and a conductive bottom layer. The conductive surface layer (~20 Ωm
at Sites B and D, <1000 Ωm at Site URL) is interpreted to be associated with the spatially variable soils, organic material in open fractures, and weathered rock. The thickness of this layer is about 0.5-2 m at Sites B and D and <10 m at Site URL. The resistive second layer (~3000 Ωm at Sites B and D, >50,000 Ωm at Site URL) is interpreted to represent a zone of sub-vertical fractures saturated with fresh groundwater percolating from the surface. There is an embedded conductive layer at most stations at 100-200 m depth at Site D (with ~0.3 S conductance), which is interpreted to correspond to a low-dipping fracture zone. At URL this layer is present at a few stations (with ~0.05 S conductance) at 120-170 m depth and is interpreted to be the sub-horizontal fracture zone FZ3. The basal conductive layer (50-400 Ωm at Site B and D, 1000-2000 Ωm at Site URL) is interpreted to correspond to sparsely fractured rocks containing brackish or saline groundwater. At Site A, the complex responses suggest the presence of significant anisotropic, 2D or 3D components in the structure.

The results of this study show the TEM method can be used to investigate the fracture zones and groundwater salinity distribution in the Precambrian granitic rocks and contribute to site investigations for nuclear waste deposit. The TEM study in the Lac du Bonnet Batholith was successful in demonstrating the potential of the TEM methods in mapping groundwater salinity in granitic batholith. The PROTEM 47 instrument, in combination with a 100 m transmitter loop, provides a suitable TEM system for mapping the resistivity structure of the Lac du Bonnet batholith down to a depth of 300-400 m. For deeper penetration and more accurate results in areas of fracture zones, other TEM systems with greater transmitter power and/or larger transmitter loops will be required.
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The data from the contracted MT V5 survey were collected by Phoenix Geophysics. The LiMS data were collected by Mr. Nick Grant, with assistance of researchers from the Geological Survey of Canada (GSC), the University of British Columbia, and the University of Manitoba. Processing of LiMS data was performed using software provide by the GSC.

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Lists of Important Symbols

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<th>Definition</th>
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<tr>
<td>$\alpha$</td>
<td>rotation angle</td>
</tr>
<tr>
<td>$\beta$</td>
<td>induction number</td>
</tr>
<tr>
<td>$\gamma_{AB}$</td>
<td>coherence between time series $A$ and $B$</td>
</tr>
<tr>
<td>$\delta_{FD}$</td>
<td>skin depth in frequency domain</td>
</tr>
<tr>
<td>$\delta_{TD}$</td>
<td>skin depth in time domain</td>
</tr>
<tr>
<td>$\epsilon$</td>
<td>electrical permittivity</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>jackknife mean</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>cylindrical wavenumber</td>
</tr>
<tr>
<td>$\Lambda$</td>
<td>Swift skew</td>
</tr>
<tr>
<td>$\mu$</td>
<td>magnetic permeability</td>
</tr>
<tr>
<td>$\xi^2$</td>
<td>jackknife variance</td>
</tr>
<tr>
<td>$\rho_a$</td>
<td>apparent resistivity</td>
</tr>
<tr>
<td>$\rho_a^e$</td>
<td>early-time apparent resistivity</td>
</tr>
<tr>
<td>$\rho_a^l$</td>
<td>late-time apparent resistivity</td>
</tr>
<tr>
<td>$\rho_f$</td>
<td>resistivity of fluid</td>
</tr>
<tr>
<td>$\rho_r$</td>
<td>resistivity of rock</td>
</tr>
<tr>
<td>$\rho_l$</td>
<td>longitudinal resistivity</td>
</tr>
<tr>
<td>$\rho_v$</td>
<td>transverse resistivity</td>
</tr>
<tr>
<td>$\rho_{GM}$</td>
<td>geometric mean resistivity</td>
</tr>
<tr>
<td>$\rho^*$</td>
<td>Schmucker resistivity</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>electrical conductivity</td>
</tr>
<tr>
<td>$\sigma^2_i$</td>
<td>variance</td>
</tr>
<tr>
<td>$T$</td>
<td>twist tensor</td>
</tr>
<tr>
<td>$\tau$</td>
<td>tolerance</td>
</tr>
<tr>
<td>$\nu$</td>
<td>regularization parameter</td>
</tr>
<tr>
<td>$\phi$</td>
<td>porosity</td>
</tr>
<tr>
<td>$\phi_{ij}$</td>
<td>phase angle (phase lead of time series $i$ over time series $j$)</td>
</tr>
<tr>
<td>$\phi_s$</td>
<td>shear angle</td>
</tr>
<tr>
<td>$\phi_t$</td>
<td>twist angle</td>
</tr>
<tr>
<td>$\chi^2$</td>
<td>chi-square misfit (least square normalized by variance)</td>
</tr>
<tr>
<td>$\psi$</td>
<td>Roughness</td>
</tr>
<tr>
<td>$\omega$</td>
<td>angular frequency</td>
</tr>
<tr>
<td>$a, m, n$</td>
<td>constant in Archie’s law</td>
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</tbody>
</table>
transmitter loop area, receiver coil area
anisotropy tensor
electric and magnetic field
magnetic induction (field) in frequency domain
inductive scale length
salinity
Groom-Bailey distortion matrix
real and imaginary induction vectors
calculated data
static shift factor
electric displacement in frequency domain
electric field in time domain
electric field in frequency domain
coefficient of anisotropy
Fréchet derivative
Fréchet derivative in terms of log(σ)
forward modelling responses
gain in Groom-Bailey decomposition
observed data
thickness of layer i
magnetic field in time domain
magnetic field in frequency domain
transmitter current
Bessel function of the first kind of order i
electric current density in frequency domain
wave number
linear Laplacian operator
absolute value misfit
least squares misfit
Least squares misfit normalized by the magnitude of data
transmitter moment in frequency domain
transmitter moment in time domain
transmitter moment magnitude
1D model
number of turns of transmitter loop, number of turns of receiver loop
\( q_o \)  
free charge density  
\( r \)  
spacing between source and receiver  
\( R \)  
remote reference magnetic field  
\( R_{\text{dd}} \)  
error covariance matrix  
\( \mathcal{R} \)  
rotation matrix  
\( S \)  
saturation  
\( S \)  
shear tensor  
\( S_{AB} \)  
cross power spectral density function between time series \( A \) and \( B \)  
\( t \)  
time  
\( t, e, s \)  
values of twist, shear and anisotropy  
\( T \)  
tipper  
\( u_i \)  
vertical wavenumber  
\( u(t) \)  
step function  
\( x, y \)  
horizontal co-ordinates (\( x \) - north direction, \( y \) - east direction)  
\( V \)  
voltage  
\( Y \)  
Tikhonov regularization function  
\( z \)  
deepth  
\( Z \)  
impedance tensor  
\( z' \)  
Schmucker depth

Note: Components of tensors or vectors are denoted by a non-bold version of the corresponding symbol. \( x, y, z \) subscripts denote vector components in the rectangular co-ordination system.
<table>
<thead>
<tr>
<th>Acronym</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>ID</td>
<td>One-dimension</td>
</tr>
<tr>
<td>2D</td>
<td>Two-dimension</td>
</tr>
<tr>
<td>3D</td>
<td>Three-dimension</td>
</tr>
<tr>
<td>AAV</td>
<td>Arithmetic Average</td>
</tr>
<tr>
<td>AECL</td>
<td>Atomic Energy of Canada Limited</td>
</tr>
<tr>
<td>AMT</td>
<td>Audio-frequency Magnetotelluric</td>
</tr>
<tr>
<td>DET</td>
<td>Determinant Average</td>
</tr>
<tr>
<td>EM</td>
<td>Electromagnetic</td>
</tr>
<tr>
<td>FZ</td>
<td>Fracture Zone</td>
</tr>
<tr>
<td>GAV</td>
<td>Geometric Average</td>
</tr>
<tr>
<td>GB</td>
<td>Groom-Bailey</td>
</tr>
<tr>
<td>GDS</td>
<td>Geomagnetic Depth Sounding</td>
</tr>
<tr>
<td>GPR</td>
<td>Ground Penetrating Radar</td>
</tr>
<tr>
<td>GSC</td>
<td>Geological Survey of Canada</td>
</tr>
<tr>
<td>GSLsz</td>
<td>Great Slave Lake shear zone</td>
</tr>
<tr>
<td>LiMS</td>
<td>Long Period Magnetotelluric System</td>
</tr>
<tr>
<td>LS</td>
<td>Least Squares</td>
</tr>
<tr>
<td>MT</td>
<td>Magnetotelluric</td>
</tr>
<tr>
<td>NLCG</td>
<td>Non-Linear Conjugate Gradient</td>
</tr>
<tr>
<td>RR</td>
<td>Remote Reference</td>
</tr>
<tr>
<td>SNORCLE</td>
<td>Slave-Northern Cordillera Lithospheric Evolution</td>
</tr>
<tr>
<td>SNR</td>
<td>Signal to Noise Ratio</td>
</tr>
<tr>
<td>TE</td>
<td>Transverse Electric</td>
</tr>
<tr>
<td>TEM</td>
<td>Time-domain Electromagnetic</td>
</tr>
<tr>
<td>TDS</td>
<td>Total dissolved Solid</td>
</tr>
<tr>
<td>TM</td>
<td>Transverse Magnetic</td>
</tr>
<tr>
<td>URL</td>
<td>Underground Research Laboratory</td>
</tr>
<tr>
<td>WRA</td>
<td>Whiteshell Research Area</td>
</tr>
</tbody>
</table>
Chapter 1  Introduction

Electromagnetic (EM) surveying methods make use of the response of the ground to the propagation of electromagnetic fields. The electrical resistivity is the most relevant physical property in EM surveys. It is externally sensitive to the composition, texture, and fluid content of the rock, and provides another facet to integrated geophysical programs of near-surface and crustal studies. The most widespread causes of enhanced electrical conductivity in the continental crust are graphite, metallic sulphides and oxides, partially melted rock, and saline water in interconnected pores and fractures (Fig. 1.1).

The exploration depth of the EM survey depends on the frequency of the EM fields and the electrical conductivity of the medium through which it is propagating. The sources of EM fields originate in either the atmosphere or ionosphere (natural source), or from electric and magnetic dipoles energized by batteries and generators (controlled source). These sources produce a very wide frequency range of EM fields, $10^1$ to $10^4$ Hz, which can provide a large range of penetration depths, from several metres to hundreds of kilometres. The sensors for receiving the response from the sub-surface are electrodes for the electric fields, and magnetometers or coils for the magnetic fields. In the presence of a body of anomalous resistivity, the EM responses will differ in both phase and amplitude from the source fields. The differences reveal the presence of anomalous bodies and provide information on their geometry and electrical properties.
Figure 1.1 Resistivity range of some earth materials (after Palacky, 1987).
Many EM survey methods can be configured using different EM sources and response sensors. For example, in the time-domain EM (TEM) method, the source provides a series of discontinuous pulses separated by inactive periods, and the sensor receives the responses in the inactive periods. The exploration depth is typically limited to between several metres and several hundred metres, depending on the transmitter loop size and resistivity of survey area. The method can be used to map major base metal deposits and the groundwater resources. In the audio-frequency magnetotelluric (AMT) method, the source is from natural EM fields in the frequency range $1 \text{ to } 10^3 \text{ Hz}$, and two horizontal electric field components, two horizontal magnetic field components and one vertical magnetic field component are recorded. The AMT exploration depth is limited to several kilometres. The AMT method can map metal deposits, groundwater resources and upper crustal structure. If the frequency range of the natural source is from $10^{-4} \text{ to } 10^{-1} \text{ Hz}$, the EM method is called the magnetotelluric (MT) method. MT surveys can investigate the crust and mantle structure at depths from several kilometres to hundreds of kilometres.

1.1 Objectives of Research

The objective of this research is to investigate the crustal and lithospheric structures, the Great Slave Lake shear zone (GSLsz) and the Phanerozoic sedimentary rocks in southwestern Northwest Territories using the MT method, and to delineate the distribution of saline groundwater in the Whiteshell Research Area in eastern Manitoba using the TEM method.

Table 1.1 contrasts the TEM and MT methods used in this research. The research
consists of three parts: the LITHOPROBE SNORCLE (Slave-Northern Cordillera Lithospheric Evolution) MT project, the Whiteshell Research Area TEM project, and synthesis studies.

Table 1.1 Overview of MT and TEM methods used in this research.

<table>
<thead>
<tr>
<th></th>
<th>Magnetotelluric (MT)</th>
<th>Time-domain EM (TEM)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EM field source</td>
<td>Natural-source (atmosphere and ionosphere)</td>
<td>Battery-powered transmitter and transmitter loop</td>
</tr>
<tr>
<td>Recording period/time</td>
<td>$10^4 - 10^6$ seconds</td>
<td>6.85 - 701 $\mu$s</td>
</tr>
<tr>
<td>Penetration depth</td>
<td>100 m to several hundred kilometres</td>
<td>10's of metres to several hundred metres</td>
</tr>
<tr>
<td>Processing domain</td>
<td>Frequency</td>
<td>Time</td>
</tr>
<tr>
<td>Common applications</td>
<td>Lithospheric structure, petroleum exploration etc.</td>
<td>Near-surface structure, mining exploration, groundwater etc.</td>
</tr>
<tr>
<td>Application in this research</td>
<td>SNORCLE Transect Corridor 1, NWT, Canada</td>
<td>Lac du Bonnet area, Manitoba, Canada</td>
</tr>
<tr>
<td>Geological setting</td>
<td>Precambrian rocks overlain by ~1000 m of Phanerozoic rocks</td>
<td>Precambrian rocks with fractures and fracture zones at less than several hundred metres depth.</td>
</tr>
<tr>
<td>Study objective</td>
<td>crustal and lithospheric resistivity structure</td>
<td>Interface between fresh and saline-water saturated granitic rock</td>
</tr>
</tbody>
</table>

LITHOPROBE is Canada's national, collaborative, multidisciplinary Earth Science research project established to develop a comprehensive understanding of the evolution of the North American continent (Clowes 1993). A series of ten transects, aimed at a number of specific geological target areas, form the basis of LITHOPROBE's principle scientific and operational components (Fig. 1.2; Clowes 1993).

The SNORCLE transect is located in the southwestern Northwest Territories, the
Figure 1.2 Location of LITHOPROBE transects on a simplified tectonic element map of North America (from Clowes, 1997). The transects are: SNORCLE - Slave-Northern Cordillera Lithospheric Evolution; SC - southern Cordillera; AB - Albert Basement; THOT - Trans-Hudson Orogen; WS - Western Superior; KSZ - Kapuskasing Structural Zone; GL - Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE); AG - Abitibi-Grenville; LE - Lithoprobe East; and ECSOOT - Eastern Canadian Shield Onshore-offshore.
Yukon territory, and northern British Columbia (Fig. 1.3). The transect consists of three corridors. In this thesis, the data from Corridor 1 and 1A will be analysed. These corridors are located in southwestern Northwest Territories, Canada, between latitudes 60°-65°N and longitudes 110°-125°W. The targets of Corridor 1 are the Archean Precambrian rocks exposed in the Slave Province in the east and the Proterozoic Precambrian rocks buried beneath Phanerozoic sedimentary rocks in the west. The Proterozoic terranes are the Great Bear Magmatic Arc, Hottah terrane, Fort Simpson terrane and Nahanni terrane. Corridor 1A crosses the GSLsz that separates the Proterozoic Wopmay Orogen to the west from the Buffalo Head magmatic arc to the east.

My research focusses on the Proterozoic part of the Corridors. The analysis of the data from Archean sites has been coordinated by Dr. Alan Jones at the Geological Survey of Canada (GSC) (Jones and Ferguson 1997, 2001; Jones et al. 2001). The main objectives of the SNORCLE analysis in this thesis include:

(1) Delineation of the resistivity structure of the Phanerozoic sedimentary rocks. Information on the variation of resistivity in the Phanerozoic sedimentary rocks can be used to assess the value of the MT method in petroleum exploration. Conductive Phanerozoic sedimentary rocks (~1000 m in SNORCLE area) can also mask the resolution of the underlying crustal structure by the MT method. Understanding the distribution of conductivity of the Phanerozoic rocks is therefore useful for modelling the deep structure.

(2) Definition of the regional conductivity structure and boundaries between terranes. Some Precambrian terranes in study area (e.g., Nahanni, Fort Simpson, Hottah) (Fig. 1.3) are covered by the Phanerozoic sedimentary rocks. The surface geology can not be used to
Figure 1.3 Location of SNORCLE Transect corridors (heavy red lines) on a map of the terranes and tectonic elements of northwestern North America (from Clowes, 1997). The Contwoyto and Anton terranes are elements of the early Archean Slave Province. The Coronation and Great Bear magmatic arc terranes are part of the early Proterozoic Wopmay orogen. Faults are heavy dark lines; teeth on upper plate for thrust faults. GSLsz: Great Slave Lake shear zone.
deduce the Precambrian crustal structure. The MT method can be used to confirm the terrane boundaries defined by the other geophysical methods, such as the gravity, magnetic and seismic reflection methods.

(3) Delineation of lithospheric depth. The Archean Slave Province forms the margin of the Proterozoic terranes in the east of Corridor 1. Examination of the spatial variations of lithospheric thickness across the margin is an important aspect of geological and geophysical research.

(4) Determination of the geometry and evolution of the GSLsz. The GSLsz is arguably the most dominant feature in the aeromagnetic map of Canada. It is a northeast-trending (N45°E) strike-slip dextral fault with 300-700 km offset in the northwestern part of the Canadian Shield (Hoffman 1987) and is a crustal-scale fault zone (Hanmer 1988). The MT results will help resolve the subsurface structure of the GSLsz and contribute to the general models of major strike-slip faults.

The Whiteshell Research Area (WRA) located near Pinawa, Manitoba, Canada has been used by the Atomic Energy of Canada Limited (AECL) to investigate the concept of disposal of nuclear fuel waste in Precambrian rocks. It is located on Archean granitic rocks of the Lac du Bonnet Batholith, in the northwest part of the Winnipeg River Subprovince of the Superior Province of the Canadian Shield (Fig. 1.4 and 1.5).

In the Lac du Bonnet Batholith, the granitic rock is homogeneous over a large area and depth and large-scale faulting is rare. Subvertical fractures are common above 200 m depth (Everitt et al. 1996; Brown et al. 1989), but below 200 m, subvertical fractures are rare and are limited to the margins of low-dipping fracture zones (Brown et al. 1989). Beneath
Figure 1.4 Subdivisions of the western Superior Province. Star indicates the location of study area (http://gaia.cg.nrcan.gc.ca/emews/ws_geology.gif).
Figure 1.5 The location of Whiteshell Research Area (WRA), Underground Research Laboratory (URL) and Lac du Bonnet Batholith (Maris et al., 1997).
the major low-dipping fracture zones, the fractures are subhorizontal. This fracture
distribution controls the groundwater movement and chemistry.

Subvertical fractures near the surface provide direct pathways for the water to
circulate with surface meteoric water. Therefore the surface water is fresh and resistive. At
greater depth the sparse subhorizontal fractures cannot provide direct pathways, the
groundwater becomes more saline and conductive because of the long residence time and
dissolution of soluble salts in the host rocks (Gascoyne et al. 1994).

Many geophysical methods have been carried out to investigate the rock and
groundwater properties in the Lac du Bonnet Batholith, including aeromagnetics, gravity,
high resolution seismic reflection, ground penetrating radar (GPR), electromagnetic, and
geophysical borehole logs. The main objectives of WRA TEM project include:

1) Evaluation of the ability of the TEM method to identify the fresh/saline-water interface,
   where the resistivity will have a significant increase.

2) Delineation of the fresh/saline-water interface at specific sites in the Lac du Bonnet area,
   Manitoba.

Both SNORCLE and WRA research projects are located in Canadian Precambrian
Shield and use EM methods to determine the near-surface structure. The results from
SNORCLE and WRA will be synthesised using several common aspects:

1) Comparison of both EM results with nearby borehole logs including detailed examination
   of the relationship between the EM models and geophysical logs.

2) Analysis of transverse anisotropy. This analysis will evaluate how the EM methods
   respond to the small-scale variability of resistivity of each site.
1.2 Introduction to Electrical Resistivity of Crustal Rocks

Most rock matrices (silicate or carbonate minerals) are electrical insulators. For example, a dry granite can have a resistivity greater than $10^6 \ \Omega m$. The bulk electrical resistivity of a rock is therefore sensitive to very small changes in the minor constituents with lower resistivity, such as aqueous fluids, partial melts and conductive minerals (magnetite, graphite and sulphide and/or oxide minerals) (Palacky 1987).

There are two main mechanisms of conduction by which significant conduction of current (ie. movement of electric charge) may occur in rocks: ionic or electrolytic conduction and metallic or electronic conduction. With ionic induction, 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. The calculation of the resistivity of the rock ($\rho_r$) follows Archie’s Law for “clean” material (with no clay):

$$\rho_r = a \phi^n S^m \rho_f$$  \hspace{1cm} \text{(1.1)}

where $\phi$ is porosity, $S$ is the proportion of pores filled with fluid (saturation), $\rho_f$ is the resistivity of the fluid, and $a$, $m$ and $n$ are empirical constants which depend on the grain-shape and degree of connectivity of the grains (Archie 1942).

Fluid resistivity depends on the fluid chemistry and varies over several orders of magnitude (0.3-100 $\Omega m$) depending on salinity (Fig. 1.1). It can be written as,
\[
\frac{1}{\rho} \text{(mS/m)} = \frac{1}{6} C \text{(ppm)}
\] (1.2)

where \( C \) is the salinity of the fluid (McNeill, 1991). Therefore, the resistivity is one indicator of the groundwater salinity.

**Resistivity in the Near-surface and Upper Crust**

From Archie's Law, a rock's resistivity is determined by its porosity, fluid resistivity and saturation. In sedimentary rocks, the clay content is also important, which can result in an increase of surface conductivity (McNeill 1980). In general, a decreasing resistivity can be observed according to the order: igneous rocks, metamorphic rocks, sedimentary rocks, sediments (Telford et al. 1990, Kearey and Brooks 1991).

The resistivity of igneous and metamorphic rocks is high because of the extremely small inter-granular porosity. The typical resistivity of Precambrian shield granite is higher than 1000 \( \Omega \text{m} \) (Palacky 1987). However, the fractures resulting from the mechanical breakage can form higher porosity. Fractured igneous rocks have lower resistivity than unfractured rocks. Therefore, the measurement of resistivity can be used to map the distribution of fractures.

The porosity of sedimentary rocks is generally inter-granular and consists of voids remaining from the compaction process, which is relatively high (20%-60%). The resistivity of most sedimentary rocks (sandstone, gravel etc.) spans several decades of \( \Omega \text{m} \) (Palacky 1987). The clay-rich sedimentary rocks, such as shales, have very low resistivity (5-30 \( \Omega \text{m} \)) (Palacky 1987; McNeill 1980). The resistivity of graphite-rich meta-sedimentary rocks, such
as graphitic schist, is also very low (0.1-10 Ωm) (Fig. 1.1).

**Resistivity in the Middle and Lower Crust**

In the middle and lower crust (at depths of greater than 10-15 km), the resistivity is significantly lower than the upper crust, particularly where the crust is of Phanerozoic age. The integrated conductance of the middle to lower crust in Phanerozoic areas has a mean value of order 300-500 S compared with values of about 20 S observed in a number of Precambrian shield locations (Hyndman and Shearer 1989; Hyndman and Klemperer 1989). The relatively low resistivity zones generally correlate with zones of high seismic reflection (Hyndman and Shearer 1989), and relatively low seismic velocities (Hyndman and Klemperer 1989).

What are the causes of the observed enhanced conductivities in the middle and lower crust? Many hypotheses have been proposed during the last two decades, and the following are the most likely candidates: (1) saline fluids, (2) carbon grain-boundary films, (3) conducting minerals, and (4) partial melts.

Bailey (1990) modelled fluid residence times and showed that the residence time of fluid in the crust is dependent on the ambient temperature. He suggested that fluids will rise rapidly to the brittle-ductile transition zone, and accumulate there in extensive reservoirs with high horizontal permeability but low vertical permeability. Jones and Gough (1995) argued that because the depth of the mid-crustal conductor in south Cordillera is correlated with the depth of the 450°C isotherm (about 10 km), it is likely to be associated with hot saline water and silicate melt in fractures and other interconnected spaces. A small percentage (less than
1% of saline-filled porosity is required to explain the observed mid to lower crustal conductivities (Jones and Gough 1995). The required porosities could be very low if the fluids are highly saline with conductivities of 500 S/m rather than 50 S/m (Jones 1992).

Many petrologists see no evidence for free water within the lower crust in their data (Jones 1992). Accordingly, other explanations, such as carbon films on grain boundaries are thought to play an important role (Mareschal et al. 1992; Wannamaker, 1997). The continental deformation and metamorphism disperse carbon from its original placement as sub-thrust packages and fluid species. To some extent, the dispersion may continue today. The increase in deep crustal conductance during the Proterozoic may be caused in part by an increase in organic-rich sediment deposition plus the advent of plate tectonic processes able to carry carbon-bearing material to great depth (Wannamaker, 1997).

The petrophysical model and MT results of the Kapuskasing uplift (Katsube & Mareschal 1993) and Trans-Hudson orogen (Jones and Craven 1990) suggest graphite can be a source to enhance the conductivity of mid-lower crust.

The conductive core in the crust of the Rio Grande rift Melt is located in 10-15 km depth. Combining the seismic results (Sanford et al. 1983 referenced in Hermance and Neumann 1991), Hermance & Neumann (1991) suggested this conductive body might be a zone of diffuse magmatic intrusions, enhanced hydrothermal circulation, or a combination of both.

Resistivity of the Upper Mantle

Mantle materials are only rarely brought to the surface, most of our information on
their properties is indirect and based on the variation of seismic velocities and resistivity with depth, combined with studies of mineral behaviour at high temperature and pressures (Heinson and Constable 1992).

The estimates of upper mantle resistivity are derived from long-period MT surveys and from analysis of the daily variation of the magnetic field (Schultz et al. 1993). Although the material of upper mantle is resistive silicate minerals, enhanced conductivity has been observed at 40-180 km depth (Jones 1992; Hirsch 1990). Partial melt and hydrogen dissolved in olivine (Karato 1990) may be two factors causing the resistivity decrease. Laboratory results from Wang et al. (1999) proved the presence of hydrogen can significantly enhance the conductivity of diopsides.

Alternatively, Boerner et al. (1999) suggested the presence of hydrous mantle minerals, specifically phlogopite, generated by metasomatising events as a possible explanation for conducting Archean mantle in southern Alberta.

Ji et al. (1996) analysed the seismic and MT anisotropies of the Grenville Front area of Canada. The results show obliquity between the polarization direction of the fast split shear wave and the most electrically conductive direction in the upper mantle in regions near a transcurrent shear zone. Ji et al. (1996) assumed the seismic and MT anisotropies are controlled by lattice-preferred orientation and shape-preferred orientation (i.e. foliation and lineation) of mantle minerals (mainly olivine) respectively. If it is true, the obliquity between seismic and MT anisotropies provides a useful indicator for shearing in the upper mantle.
1.3 Scope of Thesis Research

In this thesis, geoelectric structures in Precambrian Shield rocks are analysed in two research projects: the SNORCLE MT project and the WRA TEM project.

In the SNORCLE MT project, the MT data were collected by a contracted field assistant (Mr. Nick Grant), staff of the University of Manitoba, University of British Columbia and GSC. In this thesis research, the collected MT data are reduced, processed, analysed and interpreted. From examination of the data and one and two-dimensional modelling, it is possible to delineate the resistivity structure of the Phanerozoic sedimentary rocks, and to define the regional structure, the transitions between the terranes of Proterozoic rocks in the southwestern Northwest Territories. The structural features and anisotropy of the GSLsz are analysed using the MT method and joint examination of MT and seismic results.

In the WRA TEM project, feasibility studies of using the TEM method to delineate the interface of fresh and saline-water saturated granitic rocks were completed at AECL Underground Research Laboratory (URL) in 1996 (Wu, et al. 1996, 1997). A full TEM survey was carried out in the WRA in 1997 by V. Maris and P. Street. Maris et al. (1997) and Maris (2000) analysed the central loop soundings. The present thesis research contains a more extensive analysis of the offset data from the survey.

In Chapter 1 of this thesis, I have introduced the research objectives and study areas, and briefly reviewed the resistivity properties of the rocks. In the following chapter, I will give more detailed description of geological characteristics and existing geophysical results for both research areas. In Chapter 3, I will introduce basic EM theory, including: noise analysis and removal; correction of MT distortion; 1D, 2D and 3D modelling and inversion.
Chapter 4 and 5 describe the data processing, analysis and interpretation from the SNORCLE and WRA areas respectively. In Chapter 6, comparisons of EM results with resistivity logs and analyses of the anisotropy in the logs are described. Chapter 7 describes the conclusion of the thesis research.
Chapter 2  Background Geology and Geophysics

In this chapter, the geological setting and previous geophysical surveys completed around the SNORCLE and WRA areas will be described.

North America is an old continent. Most of the North American cratons have been coherent since 1.7 Ga (Hoffman 1989). There are a number of Archean provinces in North America: the Slave, Nain, Superior, Hearne, Rae and Wyoming provinces (Fig. 1.2). They differ from each other in the ages and arrangement of their internal constituent terranes. For example, the Superior and Slave provinces in the Canadian shield contain small areas of pre-3.5 Ga crust but are largely composed of crust formed after 3.0 Ga (Hoffman 1989). The Archean provinces are welded together by early Proterozoic collisional orogens, e.g. the Wopmay, Thelon, Trans-Hudson and Penokean orogens (Fig. 1.2). The Wopmay Orogen was generated on 2.4-2.0 Ga crust.

The Precambrian part of the continent is encircled by Phanerozoic fold belts (the Innuitian, Cordillera, Sierra Madre, Ouachitan, and Appalachian) except for the rifted northeastern margin facing Greenland (Goodwin 1996). Phanerozoic sedimentary rocks overlie most of the Proterozoic crust in southern North America. In western Canada, the rocks of the Western Canadian Sedimentary Basin also overlie Precambrian terranes (Fig. 2.1).
Figure 2.1 Selected tectonic elements and structures of the northwest Canadian Shield (after Hanmer, 1988). Na: Nahanni terrane, FS: Fort Simpson terrane, HO: Hottah terrane, GB: Great Bear magmatic arc, HR: Hay River terrane, GSLsz: Great Slave Lake shear zone, BH: Buffalo Head terrane, BFZ: Bathurst fault zone, MFZ: McDonald fault zone. The circles show the location of MT sites on Corridor 1 and 1A.
2.1 Geology and Geophysics of the SNORCLE Corridor 1 and 1A Area

2.1.1 Geological Setting

The LITHOPROBE SNORCLE Corridor 1 and 1A are located in the southwestern Northwest Territories, Canada (Fig. 1.3). Corridor 1 crosses several geological structural elements from east to west: the Archean Slave province (Anton, Sleepy Dragon and Contwoyto terranes), the Proterozoic Wopmay orogen (Hottah terrane, Great Bear Magmatic Arc and Coronation Supergroup) and the Nahanni and Fort Simpson terranes (Fig. 1.3, 2.1). Corridor 1A crosses the Proterozoic Buffalo Head terrane, GSLsz and Great Bear Magmatic Arc terrane from east to west. Table 2.1 lists the major tectonic events in the region.

Table 2.1 The main tectonic events in the Slave Province and Fort Simpson terrane areas (modified from Goodwin 1996; Cook et al. 1998; Bleeker et al. 1999a, b).

<table>
<thead>
<tr>
<th>Time (Ga)</th>
<th>Event</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;0.60</td>
<td>Phanerozoic sedimentation</td>
</tr>
<tr>
<td>1.75</td>
<td>Rifting of Fort Simpson terrane</td>
</tr>
<tr>
<td>1.84-1.87</td>
<td>Formation of Great Bear Magmatic Arc, eastward subduction of oceanic lithosphere beneath Hottah terrane</td>
</tr>
<tr>
<td>1.84</td>
<td>Fort Simpson terrane subducted into Hottah terrane eastward</td>
</tr>
<tr>
<td>1.89</td>
<td>Hottah terrane collided with western Slave craton</td>
</tr>
<tr>
<td>1.95-2.03</td>
<td>Formation of Great Slave Lake shear zone</td>
</tr>
<tr>
<td>1.95-2.40</td>
<td>Wopmay orogeny</td>
</tr>
<tr>
<td>2.0-2.40</td>
<td>Formation of Buffalo Head magmatic arc</td>
</tr>
<tr>
<td>2.60</td>
<td>Formation of Contwoyto terrane and Central Slave Basement</td>
</tr>
<tr>
<td>2.69</td>
<td>Anton terrane subducted to east, magmatism and subsequent collapse of Yellowknife Basin</td>
</tr>
<tr>
<td>2.83-4.00</td>
<td>Formation of central Slave Basement Complex</td>
</tr>
<tr>
<td>4.60</td>
<td>Oldest rock in Anton terrane</td>
</tr>
</tbody>
</table>
**Slave Province:** The oldest known rocks (about 4.6 Ga) on Earth were found in the Slave Province (Bowring et al. 1989; Isachsen et al. 1991). It is a relatively small Archean craton that exhibits distinctive rock associations and structures. The Slave Province comprises a central Meso- to Paleoarchean basement domain (Central Slave Basement Complex), flanked on either side by Neoarchean supracrustal domains (Bleeker et al. 1999a, b). The basement rocks of the south-central Slave Province are overlain by the Central Slave Cover Group (Bleeker et al. 1999a, b).

The Central Slave Basement Complex includes the Anton and Sleepy Dragon terranes (Bleeker et al. 1999a, b). Anton terrane is dominated by the granitoid orthogneisses with high-grade metamorphism. The age range is from 2.83-4.0 Ga. The Sleepy Dragon terrane is a gneissic basement of continental affinity.

The Neoarchean domain of the eastern Slave Province (Contwoyto terrane) consists of a 2.7 Ga Hackett River juvenile crust. The Central Slave Cover Group consists mainly of clastic, quartz-rich sedimentary rocks, banded iron formation and minor volcanic rocks. The age is poorly constrained (Bleeker et al. 1999a, b).

The contact between the Central Slave Basement Complex and the Central Slave Cover Group is highly deformed in most places. The results from geology (Bleeker et al. 1999a, b) and seismic reflection (van der Velden and Cook 2000) all indicate a collisional boundary between the Central Slave Basement Complex and the eastern Hackett River juvenile crust.

**Buffalo Head and Taltson Magmatic Arc Terranes:** The Buffalo Head terrane is a 2.4-2.0 Ga magmatic belt (Ross 1991) and was truncated by the GSLsz in the west at 2.03-1.95
Ga (Hoffman 1987). The Buffalo Head terrane comprises metaplutonic and subordinate felsic metavolcanic rocks (Ross 1991). The north-trending Taltson Magmatic Zone (1.99-1.90 Ga) welds the Buffalo Head terrane to the east (Thériault 1994). The Taltson Magmatic Arc is a belt of granitic to dioritic plutons (1.99-1.92 Ga) and contains magnetite- and ilmenite-series plutons. The western limit of Taltson Magmatic Arc is not exposed, but the continuity of magnetic anomalies suggests that they underlie much of northern Alberta (Hoffman 1987).

**Wopmay Orogen:** The Wopmay Orogen (Table 2.1) was generated on cryptic 2.4 to 2.0 Ga crust, according to xenocrystal and zircon dating and Pb and Nd isotopic data (Housh and Bowring 1988). It is composed of three major tectonic elements: the Hottah terrane, the Great Bear Magmatic Arc and the Coronation Supergroup (Fig. 1.3, Fig. 2.1) (Hoffman and Bowring 1984; Hildeberand, et al. 1987; Camfield et al. 1989). To the west is the older Hottah terrane (1.95-1.91 Ga), a metamorphic-plutonic complex; to the east is the Coronation Supergroup, a continental margin (back-arc) depositional prism (Hoffman and Bowring 1984; Camfield et al. 1989). Between them is the 100 km wide Great Bear Magmatic Arc, a continental “calc-alkalic” volcano-plutonic depression dated at 1.875-1.84 Ga that unconformably overlies the Hottah terrane and (deformed) Coronation margin strata. The Great Bear Magmatic Arc resembles Cenozoic continental-margin magmatic arcs and is interpreted as the product of eastward subduction of oceanic lithosphere beneath the Hottah terrane (Cook et al. 1998).

**Great Slave Lake Shear Zone (GSLsz):** The GSLsz is a northeast-trending, 25-km-wide dextral continental transform fault extending from the foothills of the Rocky Mountains
in northeast British Columbia to the south side of Great Slave Lake (Hoffman 1987). The GSLsz is linked to the Paleo_proterozoic convergence and collision between the Slave and Rae Provinces (Fig. 2.1; Gibb 1978; Hoffman 1987). The GSLsz is a crustal-scale fault zone (Hanmer 1988). Based on its magnetic expression, the GSLsz can be correlated for at least 1300 km, mostly in the subsurface. Where exposed along the southeast shore of Great Slave Lake, it comprises granulite to lower greenschist facies mylonite belts (Hoffman 1987; Hanmer et al. 1992). U-Pb zircon ages on syntectonic granites define a minimum duration for ductile shear of 2.03 to 1.95 Ga (Hoffman 1987). Along Corridor 1A, the Great Bear Magmatic Arc and the Buffalo Head terrane are located to the west and the east of the fault (Fig. 1.3, Fig. 2.1).

**Fort Simpson and Nahanni Terranes:** The Fort Simpson terrane is located in the Fort Simpson structural trend, a north-northeast-striking middle Proterozoic anticlinal culmination buried beneath the Paleozoic and Mesozoic cover (Cook and van der Velden 1993). The Fort Simpson basin overlies the Fort Simpson terrane. The Proterozoic sedimentary rocks deposited in the Fort Simpson monocline (basin) may correlate with the 1.846-0.9 Ga Mackenzie Mountains Supergroup and Wernecke Supergroup (Cook et al. 1992). The main rocks of the Mackenzie Mountains Supergroup are sandstone, mudrocks, and cherty and stromatolitic dolomite. The Wernecke Supergroup consists mainly of fine-grained clastic strata and carbonate (Aitken 1993). The Fort Simpson basin is interpreted to be a result of lithospheric extension following the collision of the Fort Simpson terrane with the western Hottah terrane at about 1.840 Ga (Cook et al. 1998).

The Nahanni terrane is located at the foot of the Cordillera orogen. The name was
introduced by Hoffman (1989) and refers to the buried crust west of the Fort Simpson trend. The age of the rocks is unknown (Ross et al. 2000). A single well showed the crystalline material to be composed of gabbroic mafic rock (Ross et al. 2000).

**Phanerozoic Sedimentary Rocks:** Phanerozoic sedimentary (< 0.6 Ga) rocks cover the western most part of the Slave craton and the Proterozoic rocks to the west (Fig. 2.1). The hundreds of wells drilled for petroleum exploration in this area show that the thickness of sedimentary cover is about 1000 m in Fort Simpson terrane and decreases to the east with the margin located in the west of the Wopmay orogen (Fig. 2.2). The sedimentary rocks consist dominantly of Ordovician and Devonian shale, siltstone and limestone (National Energy Board 1980; Aitken 1993). The geology of these rocks will be discussed in more detail during the interpretation of the MT results.

### 2.1.2 Geophysical Setting

A number of geophysical surveys have been done in and around the SNORCLE area. The results of these surveys reveal some of geological structures present.

**Potential Field Results**

Figure 2.3 shows the Bouguer gravity and magnetic anomaly maps of northwestern North America. The Slave Province is characterized by a regional gravity low with a small high near Yellowknife. The Yellowknife high correlates with the mafic rocks of the Kam Group. The regional low corresponds to the Yellowknife basin metasedimentary rocks to the east of Yellowknife and the Anton domain to the west of Yellowknife (Cook et al. 1992).
Figure 2.2 Structure contours of the basement (pre-Phanerozoic) surface in western Canada (after Aitken, 1993). Heavy line is SNORCLE transect Corridor I and IA.
Figure 2.3 Potential field maps of northwestern North America and major tectonic elements (from Clowes, 1997). (a) Bouguer gravity anomaly map, (b) Total magnetic field map. White line: SNORCLE transect Corridor 1 and 1A. NH - Nahanni; FS - Fort Simpson; HO - Hottah; GB - Great Bear; CN - Coronation; AN - Anton; YK - Yellowknife; CO - Contwoyto; KS - Ksituan; CH - Chinchaga; BH - Buffalo Head; TA - Taltson; RA - Rae, GSLsz - Great Slave Lake shear zone.
The Great Bear Magmatic Arc has a magnetic high, which can be traced from the Great Bear Magmatic Arc to the GSLsz along longitude 118°W. This magnetic high may be a buried suture between the Wopmay Orogen and Slave Province (Burwash et al. 1993). The Taltson Magnetic Arc and the Buffalo Head terrane, south of Great Slave Lake (east of longitude 118°W) also are characterized as a magnetic high. These magnetic highs are probably caused by variations in the magnetic properties of Proterozoic rocks at depth because the Phanerozoic cover rocks are thin and nonmagnetic (Cook, et al. 1992). Hoffman (1987) indicated that each high is the magnetic signature of a mid-Proterozoic magmatic arc formed between 0.2 Ga and 1.8 Ga.

The magnetic and gravity lows in the Nahanni terranes and the west of Fort Simpson terrane are associated with the Fort Simpson Basin. The east side of the Fort Simpson structure and the west of the Hottah terrane (along 120°W) are characterized by a magnetic low and a prominent gravity high. This signature can be interpreted to represent the suturing of the Fort Simpson terrane to the Hottah terrane as proposed by Hoffman and Bowring (1984), Hildebrand et al. (1987), and Hoffman (1987).

**Seismic Reflection Results**

Nearly 725 km of deep reflection data from the SNORCLE transect provides images of Proterozoic and Archean terranes and their boundaries from the Archean Slave Province on the east to the Cordillera on the west (Fig. 2.4; Cook et al. 1998).

It is difficult to delineate the transition from the Slave Province to the Wopmay orogen from the seismic reflection profile. The eastern part of the Wopmay orogen is not
Figure 2.4 Seismic reflection profile along Corridor 1 and its geological interpretation (from Cook et al., 1998) anomaly curves along the profile are shown at the top of figure.
(al., 1998). The magnetic and Bouguer gravity
exposed or is non-existent, and is covered by thin Phanerozoic sediments (Fig. 2.2). A lead isotope study of the Wopmay orogen by Housh et al. (1989) indicates that Archean rocks of the Slave Province do not extend far to the west of the medial zone in the subsurface. The main feature of the Slave-Wopmay transition is a westward tapering geometric wedge (Fig. 2.4).

From the west of Great Bear Magmatic Arc to the west, the seismic profile shows a thin sedimentary rock layer overlying a basinal feature with angular unconformity (Cook et al. 1998). In the Great Bear Magmatic Arc and Hottah terranes, the complex reflections (multiple dipping and horizontal reflections) in the crust imply the several tectonic events: uplift, fold and deformation. The complex reflections are interpreted as imaging the collision of the Hottah terrane with the western Slave Province (Bowring and Grotzinger 1988). The Moho is imaged as a set of clear events at about 40 km depth. The Moho reflection disappears further to the west at about station 4500.

Station 3000 (at about 120°W) is characterized as a structure boundary. The seismic reflection wedge is interpreted to represent the suturing of the Fort Simpson terrane with Hottah terrane (Cook et al. 1998; Snyder 2000). Moho reflections in this area dip eastward and correlate with events that project to nearly 80 km depth beneath the Hottah terrane and Great Bear Magmatic Arc.

The Fort Simpson basin is dominated by a set of west-dipping reflections. The thickness of basin increases westward from its edge near station 2600 to 24 km in the west of the profile. The Moho rises from about 37 km near stations 1000-1500 to about 30 km near station 101.
Seismic Refraction Results

During August-September 1997, LITHOPROBE conducted a seismic refraction/wide-angle reflection survey (SNORE 97) along Corridor 1 (Viejo et al. 1998; 1999). 2D velocity models (Fig 2.5) show:

(1) Crustal velocities for both Archean and Proterozoic terranes vary between 6 and 6.9 km/s. Mantle velocities vary from 8.0 to 8.3 km/s, with an anomalously low velocity (7.6 km/s) beneath the transition between the Hottah and Fort Simpson terranes. In general, mantle velocities increase to the east.

(2) In the Slave Province, the Anton terrane has higher velocities than the eastern terranes.

(3) Beneath the Fort Simpson-Hottah transition, there are relatively high velocities (7.0 km/s) above Moho. The velocity contrast with the adjacent area is 0.15 km/s. The velocity is relatively low (7.7-8.0 km/s) with minus 0.2-0.5 km/s velocity contrast from 30-75 km depth. Together with the results of the seismic reflection survey, this transition can be attributed to the subduction processes and delamination of the lower crust by the colliding Fort Simpson terrane.

(4) The Proterozoic Fort Simpson basin is well defined by a low velocity zone (6.15 km/s) with about -0.1 km/s velocity contrast in western side of the profile.

(5) The Moho appears as a flat continuous interface at 32-35 km (Viejo et al. 1999).

Electromagnetic Results from Areas Adjacent to the Present Study Area

Geomagnetic Depth Sounding (GDS) is a geomagnetic sounding method, based on recordings of the time variations of the three components of the magnetic field at the survey
Figure 2.5 2D velocity structural models along Corridor 1 (from Viejo et al., 1999). (a) Velocity of the crust; (b) velocity anomaly image (an average velocity model was obtained for the whole line and then subtracted from the final velocity models); (c) velocity of the Moho and subcrustal mantle. Velocity values are shown by the color scale (right), some values (km/s) are indicated directly by the number.
locations. A GDS survey was carried out in 1982 at eight sites along a 240 km profile across the Wopmay orogen (Fig. 2.6a), north of the SNORCLE Transect (Camfield et al. 1989). Estimations of the geomagnetic transfer functions along the profile show a clear crossover indicating the presence of a local conductor (Fig. 2.6b). The conductor is shallow, narrow, north-south trending, and is located to the east of the Wopmay fault zone (Camfield et al. 1989). Because the shortest period used in the analysis is 63 s, and the longest period used is 501 s, it is impossible to evaluate the exact location and depth of the conductor (Boerner et al. 1996). The lack of transfer function anomalies on the ends of the profile suggests the Archean basement and the Great Bear Magmatic Arc are resistive.

The SNORCLE MT data from the Slave Province along with additional MT soundings from the region were processed and analyzed by Jones and Ferguson (Jones and Ferguson 1997, 2001; Jones et al. 2001). Figure 2.7a shows a map of the phases for a period of 300 s, corresponding to the EM signal penetration 150-200 km down into the lithosphere. The averaged phases are high (80°) in the Lac de Gras region. Phases decrease from this region to the north towards the Jericho Kimberlite pipe, to the southeast towards the Kennady Kimberlite pipe, and to the southwest towards Yellowknife. A qualitative interpretation of the results is that they are explained either a higher conductivity of the mantle, or a shallower depth to the base of the lithosphere beneath Lac de Gras compared to neighboring areas (Jones et al. 2001).

The resistivity model from 2D inversion (Fig. 2.7b) shows a high conducting upper mantle region (<30 Ωm) with its top at a depth of ~80 km beneath Lac de Gras. This anomaly is consistent with the phase map. The anomaly exists between sites 205 and 217. Due to the
Figure 2.6 Results from EM survey (from Camfield et al., 1989). (a) Magnetometer measurement locations on a simplified geological map of the northern Wopmay Orogen. (b) Transfer function amplitude data for geomagnetic temporal variations at periods of 63s and 501 s. The in-phase data have been reversed to point toward conductors. The lower plots show the transfer function amplitude data (open circles) together with a 2D conductivity model response (solid lines).
Figure 2.7 MT results in Archean Slave Province (from Jones et al., 2001).
(a) average phases at 300 s; (b) resistivity model along a north-south profile.
rapid attenuation of electromagnetic fields in the conductor, it is not possible to image the resistivity structure below it. This would require much longer periods. However, the results show that lithosphere is thinnest beneath the Lac de Gras, where it has a thickness of about 80 km. The depth of lithosphere increases to south, to about 250 km in the south end of the profile (Jones et al. 2001).

The crust of the Lac de Gras region contains more conducting features than to the south. The Anton terrane in the southwest of the profile has no conducting lower crust (Jones and Ferguson 2001). The resistive lower crust in Anton terrane is in contrast to all other Archean cratons, such as the Superior, Kaapvaal, and Siberian, and younger-aged (Proterozoic and Phanerozoic) crust where the lower crust is conductive. The resistive lower crust in Anton terrane implies the absence of postulated sources of lower crustal conductivity, i.e., graphite, conductive minerals or fluids.

**Teleseismic Results**

Analyses of teleseismic data in the Slave Province reveal the mantle stratigraphy (Bostock 1997, 1998). The results reveal a layer at ~70-80 km depth and a sequence of at least two layers between 120 and 150 km depth. They suggest that the Slave craton was assembled through processes of shallow subduction resulting in a near-horizontal mantle stratigraphy. An eastward dipping layer is located at ~170 km depth in the west of the Slave Province and 230 km depth in the center of the Slave Province. It extends over a 130 km distance in the horizontal direction. Bostock (1998) suggests this layer might be due to Proterozoic subduction. The deepest interface revealed in the analysis is at about 350 km
depth, and is interpreted as the top of a layer containing a dense silicate melt fraction (Bostock 1998).

Shear-wave splitting can be used to analyze seismic structure, and has been used in the study of the Canadian Cordillera (Silver and Chan 1991; Jones et al. 1996). In general, a shear wave passing through an anisotropic region will split into fast and slow modes of propagation. The polarization directions of the split shear waves record the orientation of the elastic symmetry system (Crampin 1981). The splitting trends are commonly interpreted as indicative of either fossil anisotropy in the mantle (Silver and Chan 1991), or the direction of absolute plate motion (APM) due to present-day asthenospheric flow (Vinnik and Farra 1992).

Eaton and Asudeh (2001) presented the teleseismic results around the GSLsz. They resolve significant shear wave splitting near the GSLsz with differences in SKS arrival times from different azimuths of 0.9 to 1.4 s. Their data are generally consistent with a two-layer model containing a shallow layer with a fast-axis direction of N43°E±9° and a deeper layer with a fast-direction of N104°E±10°. The anisotropy within the shallow layer is strongest within the shear zone with the time differences for that individual layer reaching values of 1.6 s. The anisotropy of the deeper layer increases to the southeast.

2.2 Geology and Geophysics of the Lac du Bonnet Area

In Canada, the Nuclear Fuel Waste Management Program began in 1978 to assess the concept of safe disposal of nuclear fuel waste in the plutonic rock of the Precambrian Shield, and to demonstrate the technology necessary for disposal. Four main geologic research areas
in the Canadian Shield have been used for this purpose (Fig. 2.8). Two of them are located in granitic plutons: Atikokan in northwestern Ontario and Pinawa in southeastern Manitoba; one is on a gabbroic pluton (near Massey in central Ontario); and one is on gneissic terrain (near Chalk River in eastern Ontario). In 1980 AECL obtained a 21-year lease on a 3.8 km² area of the Lac du Bonnet batholith near Pinawa in Manitoba to conduct in-situ geoscientific experiments in a realistic environment within the granite batholith (Brown et al. 1989).

For safe disposal of nuclear fuel waste, the quality of the rock and the groundwater flow are two important factors. The deposit area should have low permeability and sparse fracture zones (Brown et al. 1989; Soonawala et al. 1992). The flow velocity of groundwater should be very slow, so that the groundwater will take a very long time to reach the surface. AECL has been investigating the possibility using geophysical methods to define groundwater characteristics, e.g., geophysical mapping of the interface between near-surface fresh groundwater and deeper saline groundwater in a granitic rock environment near Pinawa, Manitoba (Wu et al. 1996).

2.2.1 Geology of the Lac du Bonnet Area

The Lac du Bonnet Batholith lies in the northwest part of the Winnipeg River Subprovince of the Superior Province of the Canadian Shield (Fig. 1.4). The Archean Superior Province consists of a series of subparallel east-west trending belts of contrasting lithology, age and/or metamorphic grade (Card and Ciesielski 1986; Card 1990). Plutonic, volcanic-plutonic, high-grade gneiss, and metasedimentary rocks are the main types of rocks. The Winnipeg River Subprovince, one of the subdivisions of the Superior Province, is a
Figure 2.8 Locations of AECL geologic research areas and generalized geology (from Brown et al., 1989).
plutonic belt.

The Lac du Bonnet Batholith was emplaced at about 2.665 Ga based on a U-Pb zircon age (Brown et al. 1989). It includes pink, porphyritic granite-granodiorite; biotite-rich and gneissic granite; xenolith-bearing granite; and grey granite-quartz monzonite (Brown et al. 1989).

In the Lac du Bonnet Batholith, the granitic rock is homogeneous over a large area and depth. Large-scale faulting is rare and limited primarily to a NW-striking fault in the eastern part of the batholith (Everitt et al. 1996). There are several large-scale ENE lineaments. Their direction coincides with the long axis of the batholith and is the main orientation of many small-scale fractures throughout the batholith. Low-intermediate-dipping fracture zones or faults (10°-30°) are common at surface and can extend from the surface to at least 800 m depth, forming high-flow channels (Brown et al. 1989). Examples of these fracture zones include FZ1, FZ2 and FZ3 near the URL (Fig. 2.9). Subvertical fractures are common in outcrop. They are long and straight and divide the rock into rectangular blocks that are eroded to form a stepped topography marked by subvertical escarpments and flat to gently sloping outcrops (Fig. 2.9; Brown et al. 1989). Many subvertical fractures die out around 100 m from the surface (Agterberg 1996). Around the URL, subvertical fractures are rare and below 200 m are limited to the margins of the low-dipping fracture zones (FZ2 and FZ3) (Fig. 2.9). Under the major fracture zones, the fractures are subhorizontal. The greatest intensity of fractures is coincident with the intersection between pink, porphyritic granite and grey, more equigranular granite.

Overburden consists of glacio-lacustrine sediments and tills of variable thickness
Figure 2.9 Generalized block models in the URL showing (a) low-dipping fracture zones and location of litho-structural, (b) fracture domains. (From Everitt et al., 1996).

2.2.2 Hydrogeology of the Lac du Bonnet Area

Hydrogeologic studies of the Lac du Bonnet Batholith indicate that groundwater movement and chemistry depend strongly on the permeability and porosity, which are controlled by fractures in the pluton (Everitt et al. 1996).

The groundwater salinity in the Lac du Bonnet Batholith generally increases with depth (Gascoyne et al. 1987 referenced in Everitt et al. 1996). This is similar to the results for many other locations on the Canadian Shield (Frape et al. 1984). In the Lac du Bonnet Batholith, the groundwater is comparatively fresh in the near-surface. The total dissolved solids (TDS) is 0.3 g/L above 250 m (Fig. 2.10a, Everitt et al. 1996). Beyond this depth, the salinity increases with depth and TDS can be up to 50 g/L. The concentration of Cl⁻ shows a continuous increase with depth (Fig. 2.10b). The near-surface sub-vertical fractures provide pathways for (fresh) meteoric water to penetrate the granite, and enter groundwater flow systems in the sub-horizontal fracture zones (Everitt et al. 1996).

Gascoyne et al. (1994) analyzed the $^{36}$Cl content of groundwater in the Lac du Bonnet Batholith, and compared the results with $^{36}$Cl content of rock matrix solutions obtained from two-borehole leaching experiments. The $^{36}$Cl/Cl ratios of groundwater in the fracture zones are significantly higher (by a factor of 1.5-2.5) than those of the borehole leaching waters whose salinity is derived only from rock-matrix salts. The results suggest that the groundwater salinity is largely derived from dissolution of soluble salts in the host rock (Gascoyne et al. 1994). The fresh-saline water interface as seen on borehole resistivity logs
Figure 2.10 Variation of groundwater chemistry with depth for URL and vicinity: (a) Total dissolved solid (TDS) and (b) Chlorinity (Cl) (from Everitt et al., 1996).
varies from sharp to gradual (Stevenson et al. 1996).

2.2.3 Geophysics of the Lac du Bonnet Area

Many geophysical surveys have been carried out to analyze the rocks and groundwater properties in the Lac du Bonnet Batholith (Soonawala et al. 1992): aeromagnetics, gravity, high resolution seismic reflection, ground penetrating radar, electromagnetic, and geophysical borehole logging.

Delineation of the Batholith

The magnetic method has been used to delineate the boundaries of the Lac du Bonnet Batholith, and define the petrological structural discontinuities within the batholith. The batholith has an obvious magnetic expression. The magnetic field gradually increases from the border toward the axis of the batholith, with a change of up to 1000 nT (Soonawala et al. 1992). These results are consistent with the geologically mapped position of the batholith (Stevens et al. 1995). The vertical magnetic gradient and spatially filtered images have also been used to determine the structural direction and trends in the batholith. The gravity method was used to estimate the shape and depth of the pluton and to examine the geological setting (Soonawala et al. 1992). Boyce (1991) used the Poisson theorem to combine magnetic and gravity data and analyzed the structure of Lac du Bonnet Batholith.

Delineation of Fracture Zones

At the URL located in the lac du Bonnet Batholith, the high resolution seismic
reflection method has been used successfully to delineate and image the fracture zones in plutonic rocks up to 1000 m depth (Moon et al. 1993; Kim et al. 1994). Ground-penetrating radar has also imaged low-dipping fracture zones to 30-50 m depth (Stevens et al. 1995). The disadvantage of the radar method is that fracture zones cannot be traced to great depth below the surface of the outcrop (Soonawala et al. 1992). Crosshole seismic (Wong et al. 1985; Hayles et al. 1996) and crosshole radar (Anderson et al. 1987) surveys show clearly the location and orientation of fracture zones. The crosshole seismic tomographic surveys in URL using a high frequency seismic signal transmitter (4400 Hz) (Hayles et al. 1996) defined the geologic conditions between two boreholes. The location of fracture zones is good agreement with the fracture frequency observed in the borehole core and single-hole acoustic velocity. The surveys also imaged an anisotropy in the P-wave velocity, for which vertically travelling waves are about 4% faster than those travelling horizontally. The low velocity could be caused by the fractures.

The acoustic televiewer, caliper and nuclear logs can also define the location of fracture zones. For example, in borehole URL-12 (Fig. 2.11), the acoustic televiewer log shows major fracture zones at 145-155 m, 474-482 m and 668-674 m depths at which the acoustic velocity log shows the relatively low velocity. The caliper log shows increased hole diameter at 472-482 m depth. EM logs show low resistivity in these three depth ranges (Fig. 2.11).

Delineation of Groundwater Properties

Archie's Law (Eq. 1.1) indicates that fluid resistivity is an important factor
Figure 2.11 Geophysical logs of borehole URL-12 in the URL (from Soonawala et al., 1992).
controlling the rock resistivity. Fluid resistivity is directly related to the salinity of the groundwater (Eq. 1.2). The fluid resistivity log at borehole URL-12 exhibits a gradual decrease with depth and a very substantial drop below the fracture zone at 660 m, which indicates a variation of groundwater salinity with depth (Soonawala, et al. 1992). The results indicate that borehole EM and electrical surveys should be sensitive to the groundwater salinity (Stevenson et al. 1996).

In 1980, several EM surveys were carried out within the URL (Paterson and Watson 1981) including resistivity soundings using the Schlumberger array and complex resistivity surveys using gradient and dipole-dipole arrays. The Schlumberger soundings detected low resistivity at depth 110-220 m. The results from the dipole-dipole responses indicated significant drops of resistivity at depth 300-700 m. The variation of resistivity was interpreted in term of the variation of groundwater salinity. Other DC vertical electrical soundings, AMT and MT at the WRA (Hoover et al. 1987) indicated only a two-layer structure: a thin inhomogeneous conductive surface layer (about 10 m thick, with 10 Ωm resistivity) overlying a very resistive second layer (up to 20,000 Ωm).
Chapter 3  Methodology and Applications

This chapter will introduce electromagnetic theory and provide some details on two methods: the magnetotelluric and time-domain electromagnetic methods, including basic theory, data collection, data processing, modelling and applications related to this thesis research.

3.1 Introduction to Electromagnetic Theory

All EM phenomena are governed by Maxwell’s equations which in the nationalized MKSA system of units are written:

\[
\begin{align*}
\nabla \times E &= -\frac{\partial B}{\partial t} \\
\nabla \times H &= \frac{\partial D}{\partial t} + J \\
\n\nabla \cdot B &= 0 \\
\n\nabla \cdot D &= q_o
\end{align*}
\]  \hspace{1cm} (3.1)

where,

\[
\begin{align*}
B &= \mu H \\
D &= \varepsilon E \\
J &= \sigma E
\end{align*}
\]  \hspace{1cm} (3.2)
where $E$ is the electric field strength, $H$ is the magnetic field strength, $D$ is electric displacement, $B$ is magnetic induction (field), $J$ is electric current density, $q_o$ is free charge density, $\mu$ is permeability, $\epsilon$ is permittivity, $\sigma$ is conductivity and $t$ is time.

For most elementary EM earth problems, there are some assumptions that are usually made: all media are linear, piece-wise homogenous, and $\mu$, $\epsilon$ and $\sigma$ are independent of time, frequency, temperature, and pressure (Ward and Hohmann 1988). If the media is homogeneous, there is no charge accumulation and $\rho_f$ can be neglected (Ward and Hohmann 1988). Maxwell’s equations can be transformed to the following equations,

$$\nabla^2 E - \mu_0 \frac{\partial E}{\partial t} - \mu \epsilon \frac{\partial^2 E}{\partial t^2} = 0$$

$$\nabla^2 H - \mu_0 \frac{\partial H}{\partial t} - \mu \epsilon \frac{\partial^2 H}{\partial t^2} = 0$$

We adopt a coordinate system with $x$ to the north, $y$ to the east and $z$ vertical downwards. Assuming the electric and magnetic fields have a positive harmonic time dependence i.e. $A(x, y, z, t) = A(x, y, z)e^{i\omega t}$, equation (3.3) becomes,

$$\nabla^2 E -(i\omega \mu \sigma - \omega^2 \mu \epsilon)E = 0$$

$$\nabla^2 H -(i\omega \mu \sigma - \omega^2 \mu \epsilon)H = 0$$

Equation (3.4) is a wave equation (Helmholtz equation). The solution for most EM problems involves solving the wave equation in piece-wise uniform media using appropriate initial and boundary conditions.
3.1.1 EM Fields in Uniform Layered Media

The EM fields of a finite source in a layered earth can be represented as a superposition of plane-waves with varying, angles of incidence (Ward and Hohmann 1988). For a simple source consisting of a distant external line current oriented parallel to the x-axis, we have $E(x,y,z) = E_x(y,z)$ and $H(x,y,z) = [0, H_y(y,z), H_z(y,z)]$. The resulting current system is characterized by the absence of a vertical electric field and is defined as the transverse electric (TE) mode. For a different simple source consisting of an infinitely long ground 2D electric dipole striking parallel to the x-axis, we have $H(x,y,z) = H_y(y,z)$ and $E(x,y,z) = [0, E_x(y,z), E_z(y,z)]$. The resulting current system is characterized by the absence of a vertical magnetic field and is defined as the transverse magnetic (TM) mode (Fig. 3.1). The TE and TM modes are two basic modes of electromagnetic field propagation in a layered earth. The current systems of the two modes will interact independently with the layered earth.

From Eq. (3.4) and for the sources mentioned above, the horizontal electric field in TE mode and the horizontal magnetic field in TM mode can be written as,

$$\frac{\partial^2 E_x}{\partial y^2} + \frac{\partial^2 E_x}{\partial z^2} - (i\omega \mu \sigma - \omega^2 \mu \epsilon) E_x = 0$$

$$\frac{\partial^2 H_x}{\partial y^2} + \frac{\partial^2 H_x}{\partial z^2} - (i\omega \mu \sigma - \omega^2 \mu \epsilon) H_x = 0$$

(3.5)

The equations for the electric and magnetic fields have the same format. We can examine the solution for the TE mode by solving for the horizontal electric field. After Fourier transform, in the spatial frequency domain we have,
The fields can be separated into the TE mode which is characterized by the absence of a vertical electric field and the TM mode which is characterized by absence of a vertical magnetic field.

Figure 3.1 Electromagnetic field distribution in uniform layered media for simple source.
\[
\frac{\partial}{\partial y} = ik_y, \quad \frac{\partial^2}{\partial y^2} = -k_y^2
\]  

(3.6)

where \( k_y \) is related to the geometry of the source and the distance from the source. Thus,

\[
\frac{\partial^2 E_x}{\partial z^2} - \Theta^2 E_x = 0
\]  

(3.7)

where

\[
\Theta^2 = k_y^2 + k^2
\]  

(3.8)

and

\[
k^2 = i\omega\mu_0 - \omega^2\mu\epsilon
\]  

(3.9)

The solution of Eq. (3.7) can be written as.

\[
E_x(k_y, z, \omega) = A(k_y, \omega) e^{\Theta z} + B(k_y, \omega) e^{-\Theta z}
\]  

(3.10)

For the case the half-space, the field will decay to zero as \( z \to \infty \) and \( A = 0 \). Thus,

\[
E_x(k_y, z, \omega) = E(k_y, 0, \omega) e^{-\Theta z}
\]

\[
= E(k_y, 0, \omega) e^{-\text{Re}(\Theta) z + \text{Im}(\Theta)}
\]  

(3.11)

where \( \text{Re}(\Theta) \) parameterizes the attenuation of the field magnitude with depth and \( \text{Im}(\Theta) \) parameterizes the phase change.

### 3.1.2 Skin Depth

**Frequency Domain**

The skin depth \( \delta \) in frequency domain is defined as the depth at which uniform fields
have decayed to a magnitude of \(1/e\) (37%) of the surface value in a uniform conductor. For a uniform field, \(k_j\) is equal to zero. The skin depth can be derived,

\[
\delta_{FD} = \frac{1}{Re(\Theta)} \left[ \frac{2}{-\omega^2 \mu_\varepsilon + \sqrt{\omega^4 \mu^3 \varepsilon^2 + \omega^2 \mu^2 \varepsilon^2}} \right]^{1/2}
\]  

(3.12)

For the quasi-static approximation \((\omega \varepsilon \ll \sigma)\), which is usual at audio frequencies and below, we have,

\[
\delta_{FD} = \sqrt{\frac{2}{\omega \mu \sigma}}
\]  

(3.13)

The skin depth depends on the source frequency as well as on the conductivity structure of the earth material from the surface to that depth. For the resistive upper crust of the Canadian Shield and the conventional period range of modern acquisition \((10^4 - 3 \times 10^4\) s), the MT method can resolve electrical structures from several tens of metres to several hundred kilometres depth.

**Time Domain**

Consider a step function in an uniform external magnetic field at \(t=0\),

\[
H_f(z=0,t) = u(t)
\]  

(3.14)

where
For the quasi-static approximation, the solution of Eq. (3.7) for this type of source (Nabighian and Macnae 1991) may be written as,

\[ u(t) = \begin{cases} 0 & t < 0 \\ 1 & t \geq 0 \end{cases} \quad (3.15) \]

At a fixed time \( t \), the electric field decays with depth \( z \). At a fixed depth \( z \), the time derivative of electric field reaches a maximum value at a time,

\[ t = \frac{\mu \alpha z^2}{2} \quad (3.17) \]

This time is sometimes called the diffusion time. Rearranging Equation (3.17), we can define the time-domain skin depth,

\[ \delta_{TD} = z = \sqrt{\frac{2t}{\mu \sigma}} \quad (3.18) \]

This is the depth at which, for a source consisting of an impulse at \( t=0 \) (the time-derivative of Eq. 3.15), the magnetic field reaches its maximum value at a given time. The time-domain skin depth also depends on the conductivity of the earth. For example, in the resistive Canadian shield, if the sampling time is from 10-500 \( \mu \)s, the skin depth ranges from about 15-1500 m.
3.2 Magnetotelluric Method

3.2.1 Data Collection and Instruments

Magnetotelluric surveys use natural EM field sources: thunderstorms for the frequencies higher than 1 Hz and current systems in the ionosphere and magnetosphere driven by solar activity for frequencies below 1 Hz (Vozoff 1991). In both cases the sources are far from the surface of the earth and the EM fields behave almost like plane waves because of the very large resistivity contrast between the air and the Earth (Ward and Hohmann 1988).

MT soundings for crustal and upper mantle studies are usually carried out using simultaneous recordings of five field components varying with time: three components of the magnetic field $B_x$, $B_y$, and $B_z$; and two horizontal components of the electric field $E_x$ and $E_y$ (Fig. 3.2). Remote-reference horizontal magnetic field components are often measured, typically at a distance of several hundred metres to several tens of kilometres from the main site.

Magnetic field components are commonly recorded by fluxgate sensors or induction coils. The fluxgate sensor includes two cores of easily saturable, high-permeability material (e.g. ferromagnetic cores) wound with a common winding. When an alternating excitation current is fed into the winding, the induction magnetic field is generated. By connecting the excitation windings of two cores in series opposition, the two cores are excited in antiphase. This creates an even harmonic excitation frequency series. The sensor therefore only senses the external field, such as the Earth’s field. The core with the field parallel to the Earth’s field saturates earlier than the core with the field antiparallel to the Earth’s field. The combination
Figure 3.2 Configuration of a typical MT survey (from Phoenix Geophysics Limited).
of the excited fields forms a series of pulse functions (Kearey and Brooks, 1991). For the induction coil, when a variable magnetic field with time crosses the coil, a voltage is produced which is proportional to the area of the coil and the time derivative of magnetic field parallel to the coil axis. The induction coil produces a superior measurement of high frequency (>1 Hz) signals compared with a fluxgate sensor (Vozoff 1991).

Each horizontal electric field component is measured by recording the voltage variation versus time between a pair of grounded electrodes. An electrode in contact with the earth is an electrochemical cell. A voltage exists at the interface which depends on the chemical natures of the materials involved, on their interaction, and on the temperature. In order to minimize these voltages, the common electrodes used in MT surveys are nonpolarizing Cu- CuSO₄, Pb-PbCl₂, and Ag-AgCl cells. The typical E-line length (the distance between two electrodes) in MT surveys is 100 m.

**Quality of Data**

In MT surveys, the coherence (γ²) between two field components A and B is used to evaluate the quality of data,

\[ \gamma_{AB}^2 = \frac{|S_{AB}|^2}{S_{AA}S_{BB}} \]  
(3.19)

where

\[ S_{AB}(\omega) = \frac{\sum_{m=1}^{M} A(\omega_m)B^*(\omega_m)}{ML} \]  
(3.20)
and $S_{ab}(\omega)$ is the average power spectral density function over $M$ transform points within a given frequency window and/or different data segments within the time series which have length $L$. $*$ means complex conjugation. The commonly used unit of power spectral density function is nT$^2$/Hz for the magnetic field components and (mV/km)$^2$/Hz for the electric field.

For perfect data, the coherence $r^2$ will be 1.0 for two interrelated field components. It will be small for two un-interrelated field components and will approach zero as $M \rightarrow \infty$.

### 3.2.2 Basic Theory

Because the theory and analysis in the frequency domain are simpler than in time domain, the time series first are converted into the spectra in the frequency domain by Fourier transformation. Several frequency-domain parameters are used to determine the geoelectric structure.

**Impedance**

The MT impedance tensor $Z(\omega)$ is the frequency-domain ratio of the horizontal components of the electric and magnetic field. It is defined using

$$
E_i(\omega) = \sum_{j=1}^{2} Z_{ij}(\omega) H_j(\omega)
$$

(3.21)

where $i, j$ are $x$ or $y$ components, and the tensor impedance $Z(\omega)$ is,

$$
Z(\omega) = \begin{bmatrix}
Z_{xx}(\omega) & Z_{xy}(\omega) \\
Z_{yx}(\omega) & Z_{yy}(\omega)
\end{bmatrix}
$$

(3.22)
It is a function of frequency, direction and resistivity structure.

**Half-space**

For a half-space with uniform conductivity $\sigma$, from Eq. (3.1), (3.11) and (3.21) the impedance $Z_{xy}(\omega)$ is,

$$Z_{xy}(\omega) = \frac{i\omega \mu}{\theta}$$  \hspace{1cm} (3.23)

In a MT survey, the frequency used is relatively low, we have

$$\omega \ll \sigma$$  \hspace{1cm} (3.24)

and for a uniform source,

$$k_y = 0$$  \hspace{1cm} (3.25)

Eq. (3.8) can be written

$$\theta^2 = i\omega \mu \sigma$$  \hspace{1cm} (3.26)

Substituting Eq. (3.26) into Eq. (3.23), the impedance can be written:

$$Z_{xy}(\omega) = -Z_{yx}(\omega)$$

$$= \sqrt{\frac{i\omega \mu}{\sigma}}$$

$$= \sqrt{\frac{\omega \mu}{\sigma}} e^{i\frac{\pi}{4}}$$  \hspace{1cm} (3.27)

The measured impedance can be used to determine the conductivity of the half-space. The corresponding resistivity is defined by
\[ \rho = \frac{1}{\sigma} \frac{|Z_{xy}(\omega)|^2}{\omega \mu} \]  

(3.28)

The phase of impedance is the phase lead of the observed horizontal electric field component over the perpendicular horizontal magnetic field component, and for an uniform half-space is defined by

\[ \phi_{xy} = \frac{\pi}{4} \]  

(3.29)

Note that in the rest of this section, unless otherwise stated, all EM fields and impedance are functions of frequency.

**Layered Half-space**

For a uniform or layered half-space, the relationships between the four impedance components are

\[ Z_{xx} = Z_{yy} = 0, \quad \phi_{xx} = \phi_{yy} = 0 \]

\[ Z_{yx} = -Z_{xy}, \quad \phi_{yx} = \phi_{xy} + 180^\circ \]  

(3.30)

The theoretical impedance at the surface can be calculated by the following recursion relationship for a layered half-space,

\[ Z_i(\omega) = \frac{i \omega \mu}{\theta_i} \frac{\theta_{i+1}(\omega) + i \omega \mu \tanh(\theta_i h_i)}{\theta_i Z_{i+1}(\omega) \tanh(\theta_i h_i) + i \omega \mu} \]  

(3.31)

where \( Z_i(\omega) \) and \( Z_{i-1}(\omega) \) are the impedances at the top of layer \( i \) and \( i+1 \), \( h_i \) and \( \sigma \), are the
thickness and conductivity of layer $i$. The base of the layered sequence can be treated as a uniform half-space.

In the case of a layered half-space, the resistivity calculated from Eq. (3.28) is defined as the apparent resistivity $\rho_a$ and depends on frequency:

$$\rho_a(\omega) = \frac{|Z_{a}^\prime(\omega)|^2}{\omega \mu}$$  \hspace{1cm} (3.32)

The apparent resistivity can be viewed as a weighted-average resistivity over the penetration of the signals or an alternatively resistivity of a uniform half-space possessing the same impedance magnitude as the observed response.

Figure 3.3 shows the apparent resistivity and phase for three models. For the uniform half-space, the apparent resistivity is constant and equal to the true resistivity and the phase has a constant value of 45°. For a model containing a thin conductive layer, the apparent resistivity will decrease and the phase will increase above 45° at periods for which the skin depth exceeds the depth to the conductive layer. The apparent resistivity decreases again at periods for which the skin depth exceeds the depth to the underlying resistive layer. After the phase reaches its maximum value, it relaxes back to 45° with increasing period. For a buried resistive structure, the opposite is true: the apparent resistivity will increase, and the phase will decrease below 45°, when the skin depth exceeds the depth to the resistive layer. However, the responses are not as sensitive to the resistive layer as they are to the conductive layer.
Figure 3.3 Apparent resistivity and phase for an uniform half-space, a half-space containing a buried resistive layer and a half-space containing a buried conductive layer.
**Induction Scale Length**

A parameter that can be used to evaluate the penetration depth in a layered resistivity structure is the inductive scale length $C(\omega)$ (Schmucker and Jankowsky 1972) that is defined as,

$$C(\omega) = \frac{Z(0, \omega)}{i\omega \mu_0}$$

(3.33)

where $Z(0, \omega)$ is the impedance observed at the surface of the earth.

The real part of $C(\omega)$ provides a measure of the weighted mean depth of the inductive current distribution, and is defined as a frequency-dependent depth-scale $z(\omega)$.

$$z'(\omega) = Re \left[ C(\omega) \right]$$

(3.34)

The imaginary part of $C(\omega)$ is used to define the apparent resistivity $\rho'$ at depth $z'(\omega)$.

$$\rho'(z') = 2\omega \mu_0 \left( Im[C(\omega)] \right)^2$$

(3.35)

**2D Structure**

In 2D structures, the EM fields are separated into two independent modes: TE mode and TM mode. If the axes of the data acquisition coordinate system are parallel or perpendicular to the geoelectric strike, then $Z_{xx} = Z_{yy} = 0$; if the x-axis is parallel to the geoelectric strike $\sigma = \sigma(y, z)$, then $Z_{xy} = Z_{TE}$ and $Z_{yx} = -Z_{TM}$.

For a resistivity contrast boundary in a simple 2D model (Fig. 3.4a), the TE and TM apparent resistivity and phase show different sensitivity to the structure. Both responses are frequency dependent (Fig. 3.4b, c). At short periods <0.1 s, the TE and TM apparent resistivity and phase all show the discontinuity across the boundary. With the increasing
Figure 3.4 (a) 2D model, (b) apparent resistivity and phase pseudosections along the profile, (c) TE and TM apparent resistivity and phase at periods 0.001 s and 10 s. The irregularities in the contours in the pseudosections arise because of the limited number of sites used to define the data.
period, TM apparent resistivity still shows discontinuous. The TE apparent resistivity and TE and TM phases show the smooth variation across the boundary (Fig. 3.4). These results indicate that TM apparent resistivity is more sensitive to lateral structural boundary than TE responses and can provides the location of the boundary.

If neither axis is along strike then $Z_{xy} = -Z_{yx} = 0$. It is possible to rotate the coordinate system of the data, from the acquisition orientation to axes parallel and perpendicular to the geoelectric strike. These two directions are called the principal axes. After a clockwise rotation $\alpha$,

$$Z = \mathbf{R}Z'\mathbf{R}^T$$  \hspace{1cm} (3.36)

where $Z$ is the observed impedance, $Z'$ is the rotated impedance, $\mathbf{R}$ is the rotation matrix, $\mathbf{R}^T$ is the transpose of $\mathbf{R}$, and

$$\mathbf{R} = \begin{pmatrix} \cos \alpha & -\sin \alpha \\ \sin \alpha & \cos \alpha \end{pmatrix}$$  \hspace{1cm} (3.37)

Several different methods have been used to find the rotation angle $\alpha$ between measurement direction and geoelectric strike. One method is to use the impedance polar diagram, which rotates the impedance in steps and plots the results on a polar diagram. The picked angle will be the one that maximizes or minimizes some combination of the impedance magnitude (Vozoff 1991). Swift's method is another way to find the angle $\alpha$ (Vozoff 1991). This method maximizes the quantity

$$|Z_{xy}|^2 + |Z_{yx}|^2$$  \hspace{1cm} (3.38)

at each frequency. The solution can be written,
4\alpha = \tan^{-1} \frac{(Z_{xx} - Z_{yy})(Z_{xy} + Z_{yx})^* + (Z_{xx} - Z_{yy})^*(Z_{xy} + Z_{yx})}{|Z_{xx} - Z_{yy}|^2 - |Z_{xy} + Z_{yx}|^2} \tag{3.39}

Note that in this solution there is a 90° ambiguity in the definition of the geoelectric strike. As well as the impedance, apparent resistivity, and phase, additional parameters can be used to analyse the 2D geoelectric structure.

**Tipper and Induction Vector**

In a 2D geoelectric structure and for uniform source fields, the vertical magnetic field is related to the horizontal magnetic field perpendicular to the strike direction. The relationship between \(H_z\), \(H_x\) and \(H_y\) can be written,

\[ H_z = T_x H_x + T_y H_y \tag{3.40} \]

\(T = [T_x, T_y]\) is called the tipper and is period dependent. For a 1D case, the vertical magnetic field is zero, therefore the tipper is zero, \(T = 0\). In a 2D case, if the \(x\) direction is along the strike, \(H_z\) is only related to \(E_x\) and \(H_x\) and not \(H_y\). Therefore, \(T_x\) is zero. \(T_y\) is rarely as great as 1, with 0.1 to 0.6 being the common range (Vozoff 1991). Figure 3.5 shows the tipper sections for a simple 2D model (Fig. 3.4a). The real part and imaginary part of tipper \(T_y\) all show an anomalous zone across the structure boundary, which is period dependent. With increasing period, the width of this anomalous zone increases and the magnitude of tipper increases (Fig. 3.5a, b). The magnitude of \(T_y\) is asymmetric across the boundary, decaying at a slower rate above the relatively resistive structure (Fig. 3.5c).
Figure 3.5 Tipper component $T_y$ along profile in Figure 3.4a. (a) Pseudosection of real part, (b) pseudosection of imaginary part, (c) profiles at periods 0.001 s and 10 s.
The induction vector is the geographical representation of the transfer function between the vertical and inducing horizontal magnetic field components. It is formed from the real and imaginary (quadrature) parts by plotting,

\[
C_r = -Re (T_x)i - Re (T_y)j \\
C_q = Im (T_x)i + Im (T_y)j
\]  

(3.41)

where \(i\) and \(j\) are unit vectors to the geographical north and east respectively (Schmucker 1970, Lilley and Arora 1982). The real induction vector, when reversed, usually points towards the regions of enhanced conductivity (Jones 1986). The imaginary component of the induction vector should be either parallel or anti-parallel to the real component, and departures from such a situation suggest the presence of 3D structures (Ferguson et al. 2000). The length of the induction vector is proportional to the conductivity contrast, but decreases with distance from the conductor.

**Maximum phase split**

The phase split refers to the phase difference between off-diagonal impedance components. The impedance is rotated until there is a maximum phase split. There is a \(\pm 90^\circ\) ambiguity in determining whether the direction of maximum phase split is the strike direction. The larger-phase direction will usually (but not always) correspond to the geoelectric strike when there are isolated conductors in resistive background rocks (Ferguson et al. 2000).
3D Structure

The EM fields are more complex for 3D structures because of the effects of boundary charges. The swift impedance skew $\Lambda$ is a parameter used to measure three-dimensionality (Vozoff, 1991). It is defined as,

$$\Lambda = \frac{|Z_{xx} + Z_{yy}|}{|Z_{xy} - Z_{yx}|} \quad (3.42)$$

For 1D and 2D structures for which the data have been rotated to the principle coordinate system, $Z_{xx}$ and $Z_{yy}$ will be zero (for noise free data), and the skew will be zero.

An other parameter, phase sensitive skew (PSS), is also used to measure the local 3D distortion. This parameter is based on impedance phase (Bahr 1991). If the value of PSS is greater than 0.3 the data can be considered 3D. However, the dimensionality may be ambiguous if the PPS is below 0.3.

Distortion from local inhomogeneities

Small near-surface inhomogeneities can result in the distortion of MT responses. This response is termed a “local” response to distinguish it from the “regional” response associated with the larger-scale structures. The definition of “local” is frequency dependent. At lower frequencies, the sounding penetrates over larger scales such that the structure which is classified as regional of high frequencies becomes the local structure (Wannamaker et al. 1984b). It is difficult to define the boundary of local and regional structures. The distortion from 2D and/or 3D inhomogeneities in the near-surface includes current channelling
distortion and galvanic distortion (Vozoff 1991; Zhang et al. 1987; Jones 1988).

Current channelling distortion is associated with EM induction in a conductive structure which is approximately 2D on a regional scale but is locally 3D. Even though the inductive behaviour is 2D, the resulting impedance tensor can be shown to have a 3D behaviour (Jones 1983; Zhang et al. 1993; Groom and Bailey 1989). This effect is more significant in sedimentary basins and coastal environments because of the high conductivity of the ocean and sediments. The process is frequency dependent.

Galvanic distortion is caused by the electric charges that accumulate along conductivity boundaries (or gradients) associated with the local-scale surface structure with dimensions that which are small compared with the penetration depth (Bahr 1988). These charges influence the electric field (Bahr 1988; Groom & Bailey 1989; Groom et al. 1993; Smith 1995) and may also alter the magnetic field when the charges deflect regional electric currents (Zhang et al. 1993; Chave and Smith 1994). The estimated impedance and strike direction will be changed. This distortion is a non-inductive distortion.

Electric field galvanic distortion is frequency independent. For a 2D case, if the orientation of the measurements is parallel to the strike of the distorting structure, the galvanic distortion will simply be a static shift. Because the conductor is small, its effect on phase disappears at frequencies higher than the highest available frequency (Vozoff 1991). There is a shift of the apparent resistivity curve by the same multiplicative factor at all frequencies (Jones 1988) and no change to the phase response.

The static shift in sedimentary areas is much smaller than in crystalline areas (Zhang et al. 1993), because there are more near-surface inhomogeneities with higher conductivity
contrasts in crystalline areas (Zhang et al. 1987).

Magnetic-field galvanic distortion is frequency dependant because it contains a component proportional to the impedance. Chave and Smith (1994) discuss magnetic field galvanic distortion. This distortion maybe important at periods under a few thousand seconds, but at longer periods it is not as important.

**Average Impedance Responses**

For 2D and 3D responses, it is common to use an averaged response to represent the impedance in order to simplify responses. There are a number of ways, for example, the arithmetic average $Z_{\text{avr}}$,

$$Z_{\text{avr}} = \frac{Z_{ij} + Z_{ji}}{2} \quad (3.43)$$

the geometric average $Z_{\text{gav}}$

$$Z_{\text{gav}} = \sqrt{Z_{ij}Z_{ji}} \quad (3.44)$$

and the determinant average $Z_{\text{det}}$,

$$Z_{\text{det}} = \sqrt{Z_{ij}Z_{ji} - Z_{ii}Z_{jj}} \quad (3.45)$$

The arithmetic average impedance ($Z_{\text{avr}}$) and geometric average impedance ($Z_{\text{gav}}$) are computed from the off-diagonal impedance components (Berdichevsky & Dmitriev 1976, Beamish 1986), whereas the determinant average impedance ($Z_{\text{det}}$) is computed from four
impedance components (Berdichevsky et al. 1980, Ranganayaki 1984). The apparent resistivity and phase based on these averages can be calculated by Eq. (3.28). These estimates provide the responses which can be interpreted using 1D methods. The resulting conductivity structure will be an approximation of the true structure.

The determinant apparent resistivity $\rho_{\text{det}}$ (Ωm) is computed from the determinant impedance $Z_{\text{det}}$, and can be written as,

$$\rho_{\text{det}} = \frac{T}{5} |Z_{\text{det}}|^2$$  \hspace{1cm} (3.46)

where $T$ is period.

### 3.2.3 Data Processing

The objective of MT data processing is to extract from the stochastic electric and magnetic field time series, a set of smooth, repeatable functions representing the earth’s response, which can be used to interpret the conductivity structure (Vozoff 1991). Conventional least square and remote reference processes are efficient for stationary Gaussian (normal) distribution of the noise (Egbert and Booker 1986) and signals which are stationary (statistical properties do not change with time). However, the natural sources generally do not exhibit simple, wide-sense stationary, statistical distributions. They include some nonstationary phenomena, for example, geomagnetic storms and outliers (spikes). The Robust estimation is successful for these noise distributions.
3.2.3.1 Impedance Estimation

*Least Squares Method*

Impedance estimation is the process of determination of the tensor $Z(\omega)$ from observed data. One common method to estimate the impedance is the least-squares analysis, which minimizes the mean value of $|E - Z^{LS}H|^2$. Here $E$ and $H$ are matrices containing the vector components of horizontal magnetic and electric fields, and $Z^{LS}$ is a matrix containing the tensor components of impedance. The solution for $Z^{LS}$ can be written as (Gamble et al. 1979a; 1979b),

$$Z^{LS} = [EH] [HH]^{-1}$$  \hspace{1cm} (3.47)

where $[AB]$ indicates a crosspower matrix, defined by

$$[AB] = \begin{vmatrix} S_{A,B_x} & S_{A,B_y} \\ S_{A,B_x} & S_{A,B_y} \end{vmatrix}$$ \hspace{1cm} (3.48)

The four impedance components defined by equation (3.47) can be written,

$$Z^{LS}_{xx} = S_{E,H_x} S_{H,H_x} - S_{E,H_x} S_{H,H_y} / D$$

$$Z^{LS}_{xy} = S_{E,H_x} S_{H,H_y} - S_{E,H_y} S_{H,H_x} / D$$

$$Z^{LS}_{yx} = S_{E,H_y} S_{H,H_x} - S_{E,H_x} S_{H,H_y} / D$$

$$Z^{LS}_{yy} = S_{E,H_y} S_{H,H_y} - S_{E,H_y} S_{H,H_x} / D$$

$$D = S_{H,H_x} S_{H,H_x} - S_{H,H_y} S_{H,H_y}$$  \hspace{1cm} (3.49)

This solution for the impedance will produce an unbiased impedance only if there is
no noise present on the magnetic components and no noise is correlated with the signal. The least squares impedance estimations will be biased by noise on the magnetic field because of the inclusion of autopowers of magnetic field components in the equations. Noise on the electric field components will not introduce any bias into the impedance estimation.

For most MT measurements, the electric and magnetic fields are contaminated by some level of noise. There are a number of sources of noise, for example, instrument noise, cultural noise, and thunderstorms.

Remote Reference Method

In order to avoid the use of the autopowers in the estimation of the impedance, two remote reference fields $R_x$ and $R_y$, usually horizontal magnetic fields, are recorded (Fig. 3.2). The magnetic fields in equation (3.49) can be replaced by the remote reference field $R$ (Gamble et al. 1979a; 1979b), i.e.

$$Z_{RR} = \begin{bmatrix} E \end{bmatrix} \begin{bmatrix} H \end{bmatrix}^{-1}$$

(3.50)

The corresponding impedance components are,

$$Z_{xx}^{RR} = S_{E,R_x}S_{H,R_x} - S_{E,R_x}S_{H,R_x} / D^{RR}$$

$$Z_{xy}^{RR} = S_{E,R_y}S_{H,R_x} - S_{E,R_y}S_{H,R_x} / D^{RR}$$

$$Z_{yx}^{RR} = S_{E,R_x}S_{H,R_y} - S_{E,R_x}S_{H,R_y} / D^{RR}$$

$$Z_{yy}^{RR} = S_{E,R_y}S_{H,R_y} - S_{E,R_y}S_{H,R_y} / D^{RR}$$

$$D^{RR} = S_{H,R_x}S_{H,R_y} - S_{H,R_y}S_{H,R_x}$$

(3.51)
The impedance elements calculated by the crosspower between reference and measurement fields will be unbiased if the noise on remote reference fields is incoherent with that at the main MT site and their signals are coherent (Gamble et al. 1979a; 1979b). In general, remote reference sites are located a few hundred metres to several tens kilometres away from the main MT site (Fig. 3.2).

At high geomagnetic latitude, the electrojet current systems produce non-uniform ($k_i 
eq 0$ in Eq. 3.6) and non-stationary signals, and the influence on recorded MT data is particularly high at night. The bias of MT impedance is largest at long periods (Jones 1980; Cassels and Jones 1998).

**Robust Estimation**

The aim of robust estimation is to reduce the effect of nonstationary field sources and noise (Egbert & Booker 1986; Chave et al. 1987; Larsen 1989; Larsen et al. 1996). It can take many forms,

1. The median filter. From repeated determinations of the response at each frequency, a "best" estimate is determined.
2. Absolute values ($L_1$ norm) method. This method weights each estimate according to its absolute difference from the best estimate, minimizing the sum of the differences. It is used for exponential noise distributions (Vozoff 1991; Chave et al. 1987).
3. Robust-$M$ estimation. This method is a weighted least square method. The weights are computed based on the residuals and scale estimate from the previous iteration (Chave and Thompson 1989). It will downweight the response from the outliers. In the case of
MT analysis, the MT impedance derived from segments of the time-series segments containing outliers differs from that derived from the rest of the time series (Egbert and Booker 1986; Chave et al. 1987; Chave and Thompson 1989). These estimates are therefore downweighted. The main goal is to determine the appropriate weighting (Vozoff 1991).

(4) Smooth robust estimation. This method uses an iterative re-weighting method on the time series data (Larsen et al. 1996). The time series are first weighted by whitening weights and section weights. The data are then corrected for outliers and gaps (Larsen 1989). Then a correction function is computed by the least squares or remote reference method. This process is repeated until the number of new outliers is less than 0.5 percent of the total number of data and the correction function is approximately unity (Larsen et al. 1996).

Cascade Decimation Method

The cascade decimation method uses a low pass digital filter and decimation to compute power spectra (Wight and Bostick 1980; Jones and Jodicke 1984). Sine and cosine transforms are applied. The calculated power spectra represent constant percentage bandwidths. This method is suited for the rejection of the bad data, and is a commonly used method in MT analyses.

The cascade decimation method can compute the power spectra for single or multi-channel time series data. Two programs have been developed by Alan Jones at the GSC for processing MT time series: the single-station-weighted cascade decimation and all-station-weighted cascade decimation. The single-station-weighted method uses one remote reference
station to calculate the impedance. The all-station-weighted method uses multiple remote reference stations to calculate the impedance.

Jones et al. (1989) compared the different techniques for MT response function estimation. The remote reference processing is necessary to minimize bias errors. The robust estimation is useful to remove the affection from strong geomagnetic activities (e.g. outliers, electrojet). The cascade decimation method is suited to reject intervals of bad data. The robust estimation has been successfully used to process Trans-Hudson Orogen MT data (Fig. 1.2: Garcia et al. 1997). For the SNORCLE MT data, the non-uniform and non-stationary sources from electrojet current systems at high geomagnetic latitude cause bias of the impedance estimate (Cassels and Jones 1998).

3.2.3.2 Error Estimation

Jackknife error estimation is a common method of error estimation in robust schemes for which there are no standard statistical distribution based uncertainty estimates. For a data set with \( N \) observations, the jackknife error estimate is obtained by dividing the data into \( N \) groups of size \( N-1 \) each by deleting an entry in turn from the whole set. Let \( \kappa \) be a statistical parameter of whole set and \( \kappa_i \) be the statistical parameter of the \( i \)th subset for which the \( i \)th datum has been removed. The jackknife mean is defined as,

\[
\bar{\kappa} = N\kappa - \frac{N-1}{N} \sum_{i=1}^{N} \kappa_i
\]  

(3.52)

The \( \kappa \) value of a group containing bad data points (e.g. the signal from outliers) will show a difference from the statistical parameter. The jackknife estimate defines the confidence in
terms of the variance as,

\[ \xi^2 = \frac{N-1}{N} \sum_{i=1}^{N} (\kappa_{i,i} - C)^2 \]  

(3.53)

where

\[ C = \frac{1}{N} \sum_{i=1}^{N} \kappa_{i,i} \]  

(3.54)

Data points lying farther from the statistical estimate than \( \xi \) will be considered as the bad data points and removed.

### 3.2.3.3 Tensor Decomposition

As discussed in section 3.2.2, 2D or 3D near-surface structures can cause distortion of MT responses. This distortion is defined by

\[ Z^{obs} = \mathcal{R} C Z \mathcal{R}^T \]  

(3.55)

where \( Z^{obs} \) is the observed impedance, \( Z \) is the impedance tensor of the regional structure, \( C \) is the distortion tensor, and \( \mathcal{R} \) is the rotation matrix defined in equation (3.37).

Bahr (1988), Groom and Bailey (1989) and Smith (1995) have developed different parameterizations of the distortion tensor. The Groom-Bailey (GB) decomposition is based on the approximation that the regional geoelectric structure is either 1D or 2D, and the impedance tensor consists of two parts: regional 1D/2D induction and local frequency-independent telluric distortion. The tensor operator \( C \) in GB decomposition is factored into the product of a scale factor \( g \), and matrices \( T, S \) and \( A \).
The effect of twist is simply to rotate the electric field vectors through a clockwise angle $\phi = \tan^{-1}t$. The shear develops anisotropy on axes which bisect the regional inductive principle axes. If $e$ is a positive value, an electric field vector on the $x$-axis will be deflected clockwise by an angle $\phi_x = \tan^{-1}e$, and an electric field vector on the $y$-axis will be deflected counter-clockwise by the same angle. The anisotropy tensor $A$ simply stretches the two field components by different factors $(1+s$ and $1-s)$ (Groom and Bailey 1989). The gain $g$ is a scaling of electric fields.

If the distortion is frequency independent, $g$ and $A$ will not change the principal apparent resistivity curve shapes and phases. Therefore, they can be treated as the components of static shift and be absorbed into $Z$. The GB decomposition cannot remove static shift.

If the physical model of distortion is appropriate, the GB decomposition can correctly recover the two principal impedances and provide a regional strike direction $\alpha$. The strike has a $\pm 90^\circ$ ambiguity. Once the regional strike angle $\alpha$ is known correctly, rotation of the data to the principal axis coordinate system can give impedance elements with the correct phase for the regional impedance. However, the apparent resistivity needs to be corrected for static shift.

### 3.2.3.4 Removal of Static Shift

Static shift can be removed either by using independent information acquired by
additional equipment or by using the MT responses from multiple sites to determine a static shift factor. The alternative equipment methods include conducting a controlled-source survey in which magnetic fields alone are measured and arraying many electrode dipole pairs simultaneously (the Electromagnetic Array Profiling technique) (Bostick 1987).

There are various methods which can be used to find the static shift factor from the MT response. For example, one can derive a regional or global resistivity curve from the profile and shift the data to match this curve in some manner (Vanyan et al. 1983). Alternatively, one could choose a representative frequency and determine an area-averaged apparent resistivity by weighting surrounding sites using a chosen smoothing window (Sternberg et al. 1985; Moroz 1986).

Jones (1988) determined the correction factors by comparing borehole logs data with a sufficient number of MT observations of the shallow resistivity. He assumed that the earth consists of two layers. The first layer is a 3D thin and inhomogeneous layer. For the second layer, the lateral variation of parameters (resistivity and depth) is slow along profile and can be described in a simple parametric fashion. The first step is to calculate the resistivity and depth of two layers along the profile by 1D inversion. The second step is to find a simple function to fit the resistivity of second layer. Then the correction factor is obtained using the ratio of the fitted function and the 1D inversion resistivity at each site.

Berdichevsky et al. (1989) assumed that for a 2D structure the site-to-site apparent resistivity curves for the TE mode have the same low frequency responses if the earth below a given depth is 1D. This depth is deeper than the region of interest (Jones and Dumas 1993), for example, an upper mantle depth of 50-100 km depth corresponds to a period range of 50-
100 s. The average TE apparent resistivity at the chosen low period can be determined. All TE apparent resistivity curves are then shifted to this value. At the scale corresponding to the highest frequencies, the structure is again considered to be 1D, therefore, the TM apparent resistivity is matched to the TE apparent resistivity.

The relationship between the apparent resistivity $\rho$ and the static shift factor $D$ is defined in this thesis as,

$$D = \sqrt[4]{\rho_{\text{correct}} / \rho_{\text{observe}}}$$  \hspace{1cm} (3.57)

For 1D and 2D structure where the axes of the data acquisition coordinate system are parallel or perpendicular to the 2D geoelectric strike, the impedance $Z_{xx}$ and $Z_{yy}$ are close to zero and the determinant impedance (Eq. 3.43) can be approximately by,

$$Z_{\text{det}} = \sqrt{Z_{xy} Z_{yx}}$$  \hspace{1cm} (3.58)

Combining Eq. (3.46), (3.57) and (3.58), the static shift factor of the determinant apparent resistivity can be written as,

$$D_{\text{det}} = \sqrt{D_{xy} D_{yx}}$$  \hspace{1cm} (3.59)

where $D_{xy}$ and $D_{yx}$ are the distortion factors for $\rho_{xy}$ and $\rho_{yx}$ respectively. Ferguson (1988) analysed the effect of the static shift on inversion model parameters. The parameters should be corrected by the multiplicative factors: $D$ for depth, $D^2$ for resistivity, and $D^1$ for conductance.
3.2.4 Modern Forward Modelling and Inversion Methods

3.2.4.1 Forward modelling

Section 3.1 and 3.2 contain introductions to the basic MT theory. For 1D layered structure, the MT response (apparent resistivity) at the surface can be obtained using equation 3.31. For 2D and 3D structures, the aim of forward modelling is to solve Maxwell's equations in piece-wise uniform media. Forward modelling methods are based mainly on the integral equation (Hohmann 1983; Wannamaker *et al.* 1984a), finite element (Wannamaker *et al.* 1987; Best *et al.* 1985), finite difference (Mackie *et al.* 1993; Mackie and Madden 1993a) and thin sheet (Dawson and Weaver 1979; Dawson *et al.* 1982; Dawson 1983; McKirdy and Weaver 1984, 1985) approaches.

The integral equation method is used for 2D or 3D structural models (Chave and Booker 1987; Wannamaker *et al.* 1984a). In the method, 2D or 3D inhomogeneous bodies can be replaced by an equivalent current distribution which is approximated by pulse basis function (Wannamaker 1991). It is well-suited for treating isolated bodies embedded in a simple substrate, because the numerical complexity is limited to the body itself.

The finite element method discretizes the 2D or 3D structure into the finite elements, and calculates the EM fields in each element using the discrete integral form (Salazar-Palma *et al.* 1998). It requires special attention at low frequencies, when body dimensions are small compared with the skin depth. The finite element algorithm developed by Wannamaker *et al.* (1987) overcomes this difficulty. The finite element methods are usually the most accurate of the forward modelling methods, and the finite difference methods are the quickest and simplest (Madden and Mackie 1989).
The finite difference method was the first numerical method applied to solve an EM problem (Salazar-Palma et al. 1998). It discretizes 2D or 3D structure into piecewise rectangular or cubic grids, and calculates the unknown EM fields at the points of space discretization using finite differences, and the discrete approximations to partial derivatives (Salazar-Palma et al. 1998). This method is better suited than integral equation methods for modelling arbitrarily complex geometries. The method leads to large sparse systems of equations to be solved for the unknown field values (Mackie et al. 1993). Two methods developed for solving the finite difference equations are the conjugate gradient method (Mackie and Madden 1993a, 1993b) and direct solution (Mackie et al. 1993). Some problems of the finite difference method are the number of iterations required and the ill-conditioned nature of the matrix systems.

The thin sheet method is used to model surface inhomogeneities. The earth is represented by a uniformly conducting or layered half-space overlain by a surface layer of variable conductance (Dawson et al. 1979, 1982; Dawson 1983). The thin sheet approximation simplifies the 3D induction problem. The EM fields can be solved analytically above and below the non-uniform sheet, leaving only the sheet itself as the region where numerical methods must be used. Schmucker (1995) developed a 2D thin-sheet modelling method using integral equations.

3.2.4.2 Inversion

Inversion is the procedure for finding a physical earth model from a set of observed data. All inversion methods seek to minimize some functional which penalizes data misfit
and model structure. The followings are several data misfit measures used in inversions:

(a) Least squares

\[
L_2 = \frac{1}{M} \left[ \sum_{i=1}^{M} (d_i - g_i)^2 \right]^{1/2}
\]

(b) Non-normalized absolute value

\[
L_1 = \frac{1}{M} \sum_{i=1}^{M} |d_i - g_i|
\]

(c) Least squares normalized by the magnitude of data

\[
L'_2 = \frac{1}{M} \left[ \sum_{i=1}^{M} \frac{(d_i - g_i)^2}{d_i^2} \right]^{1/2}
\]

(d) Least squares normalized by variance (\(\chi^2\) misfit)

\[
\chi^2 = \sum_{i=1}^{M} \frac{(d_i - g_i)^2}{\sigma_i^2}
\]

where \(d_i\) is calculated data, \(g_i\) is observed data, \(M\) is the number of samples, and \(\sigma_i^2\) is the variance. The \(L_2\) norms are optimal for errors with a Gaussian distribution. The magnitude normalized least-squares misfit weights down errors associated with larger data. The \(L_1\) norm is optimal for errors with exponential distribution. As mentioned earlier this estimate is more robust than least-squares (Menke 1984).

Sensitivity

Most nonlinear EM inversion problems are treated by linearization. If a small perturbation of conductivity \(\delta \sigma\) produces a small response change (apparent resistivity or
impedance) $\delta g$, then

$$
\delta g = \int_M F(\sigma, z) \delta \sigma(z) \, dz 
$$

(3.64)

$F(\sigma, z)$ is called the Fréchet derivative and is also called the sensitivity as it describes the variation of response produced by the change of the model parameters. For 1D layered resistivity model, Eq. (3.64) can be approximated by,

$$
F(z) = \frac{\Delta g}{\Delta \sigma \Delta z}
$$

(3.65)

where $\Delta g$ is change of response at the surface when a conductivity change $\Delta \sigma$ occurs over a depth range $\Delta z$ at some depth. $F(z)$ is the sensitivity of surface response $g$ and is an approximation of the Fréchet derivative. This parameter can be used to indicate penetration depth of EM signals and to gain a qualitative feel for resolvability.

The Fréchet derivative can be computed numerically directly from its definition by Eq. (3.65) (Gómez-Trevino and Edwards 1983). It can also be estimated from analytical solution. For a uniform layered half-space, the changes of impedance (\(\delta Z\)) with the variation of conductivity and depth can be written as (Parker 1977; Chave 1984; Gómez-Trevino 1987),

$$
\delta Z = \int_0^z Z_{(0, \omega)} \left( \frac{E(z, \omega)}{E(0, \omega)} \right)^2 \delta \sigma(z) \, dz
$$

(3.66)

Therefore,

$$
F(z) = Z_{(0, \omega)} \left( \frac{E(z, \omega)}{E(0, \omega)} \right)^2
$$

(3.67)

The numerical solution is an approximation of this analytical solution.
In the earth the conductivity can vary by many orders of magnitude (Chapter 2). It is therefore sometimes appropriate to parameterize the sensitivity in term of the logarithm of conductivity. We can have,

\[ \delta(\log(\sigma)) = \frac{d(\log(\sigma))}{d\sigma} \delta\sigma \]

\[ = \frac{\delta\sigma}{\sigma} \]

Substituting Eq. (3.68) into (3.65), we have,

\[ R(z) = \frac{\Delta g}{\sigma} \frac{\Delta(\log(\sigma))}{\Delta z} \]

Defining a new Fréchet derivative in term of \( \log(\sigma) \),

\[ F'(z) = \frac{\Delta g}{\Delta(\log(\sigma))} \Delta z \]

we have

\[ F'(z) = \sigma F(z) \]

For the MT problem,

\[ F'(z) = \sigma Z_{0,\omega} \left( \frac{E(z,\omega)}{E(0,\omega)} \right)^2 \]

Note that the Fréchet derivative is a complex quantity. In order to examine its depth dependance, the magnitude of the quantity normalized by either its maximum or surface value often is often considered.

Figure 3.6 shows the analytical Fréchet derivatives for the \( \log \) variation of conductivity (Eq. 3.72) for three conductivity models. The magnitude of Fréchet derivative
Figure 3.6 Frechet derivative magnitude from analytical solution at period 1 s which is normalized by the surface value. Here, the variation of conductivity used is log(σ).
decreases smoothly with depth for an uniform half space, increases in the conductive layer and decreases in the resistive layer. The large change of the Fréchet derivative magnitude in the conductive layer indicates the EM responses are very sensitive to the buried conductive layer, and is less sensitive for the buried resistive layer.

1D Inversion

Parker (1980) and Parker and Whaler (1981) analysed the existence of inversion solutions and methods for the construction of inversion solutions for the 1D MT problem. They found that if no 1D model fits a data set exactly, the conductivity models with the smallest least squares or $\chi^2$ misfit will always be composed of a finite layers. Parker called this case $D^-$ model. There are an infinite number of possible models between the best-fitting $D^-$ type and one with any larger value of the misfit. This situation is referred to as the equivalence of the inversion solution.

The determination of the simplest or smoothest model which fits the data within the prescribed $\chi^2$ is one of a variety of criteria used to stabilize the inversion (Constable et al. 1987; Smith and Booker 1988). Constable et al. (1987) applied this criterion to the 1D EM inversion method calling the resulting algorithm “Occam” inversion. The Occam MT inversion produces 1D resistivity models which make the roughness (the integral of the square of the first or second derivative of the resistivity with respect to depth) as small as possible, while $\chi^2$ achieves an acceptable value. The roughness ($\psi$) is defined as,

$$\psi = \int (\frac{dm}{dz})^2 \, dz$$ (3.73)
\[ \psi = \int (d^2 m / dz^2)^2 \, dz \]  \hspace{1cm} (3.74)

where \( m \) is the \( \log(p) \). The misfit used to evaluate a model is the tolerance (\( \tau \)):

\[ \tau = \sqrt{\frac{1}{M} \sum_{i=1}^{M} \frac{(g_i - F_i[m])^2}{\sigma_i^2}} \]  \hspace{1cm} (3.75)

where \( g_i \) is the observed data, \( F_i[m] \) is the forward modelling responses from model \( m \), \( n \) is the number of data (at various periods), \( \sigma_i \) is the uncertainty in the \( i \)th datum (assuming statistical independence in the error), and \( z \) is depth. The tolerance is a normalized \( \chi^2 \) misfit (and should be close to unity for a statistically appropriate fit to the data).

Smith and Booker (1988) developed a similar method, plus a technique for testing whether systematic regions of underfit or overfit exist in the model structure. Alternative methods for developing simple models also exist. Fischer et al. (1981) and Fischer and LeQuang (1982) developed a 1D MT inversion method which gives a best-fitting response to the observed data based on a minimum number of layers in the model.

2D Inversion

Everett (1994) proved that there exists at least one optimal solution to the 2D EM inverse problem. A number of 2D inversion methods have been developed over the past two decades. Rodi et al. (1984) first applied the Backus-Gilbert joint minimization of statistical variance and model roughness (Chave and Booker 1987). They used a smoothness constraint on the model to stabilize the solution (Rodi et al. 1984; Sasaki 1989).

deGroot-Hedlin and Constable (1990) developed 2D Occam inversion which determines the smoothest model that fits the observed data to within a prescribed error
(Constable et al. 1987). The forward modelling responses is calculated using Wannamaker’s 2D finite element forward solution (Wannamaker et al. 1987). The program will find \( \log(\rho) \) that has roughness \( R \) as small as possible, while tolerance \( T \) achieves as acceptable value.

The 2D Occam inversion algorithm requires substantial memory and computational speed (Siripunvaraporn and Egbert 2000). Siripunvaraporn and Egbert improved this inversion scheme creating the REBOCC (the reduced basis OCCAM’s inversion) algorithm, which is faster and more stable.

Some algorithms have been developed that are faster than Occam method. Smith and Booker (1991) developed an approximate 2D MT rapid relaxation inversion (RRJ). It uses 1D inversion and a full 2D forward problem with an iterative matrix solution to compute field gradients and to assess convergence.

Mackie and Rodi (1996) and Rodi and Mackie (2001) used non-linear conjugate gradients (NLCG) for direct iterative minimization of the Tikhonov regularization function \( Y(m) \),

\[
Y(m) = (g - F(m))^T R_{dd}^{-1} (g - F(m)) + \nu |L(m - m_0)|^2
\]

where \( L \) is a linear Laplacian operator, \( \nu \) is regularization parameter, and \( R_{dd} \) is error covariance matrix,

\[
R_{dd} = \begin{bmatrix}
\sigma_1^2 & 0 & 0 & 0 \\
0 & \sigma_2^2 & 0 & 0 \\
0 & 0 & \ldots & 0 \\
0 & 0 & 0 & \sigma_n^2
\end{bmatrix}
\]

The data misfit criterion used is again a normalized \( \chi^2 \) misfit. The implementation of the
inversion is accomplished by calculating $F(m)$ using a finite difference algorithm. The mesh defining the model used in forward modelling is also used as the regularization grid in the inversion (Mackie et al. 1997b).

3D Inversion

Madden and Mackie (1989) have developed a full 3D MT inversion method using a minimum structure constraint to regularize the solution. Mackie and Madden (1993b) reported an inversion procedure using the conjugate gradient relaxation method, which solves the maximum likelihood equations instead of solving the system directly using matrix inversion routines. Wang and Lilley (1999) developed a non-linear inversion method for thin-sheet modelling based on least-squares criteria (Eq. 3.63). The inversion tries to find a smooth model as proposed by Constable et al. (1987) and uses the conjugate gradient relaxation method (Mackie and Madden 1993a, b) to update the model perturbation. The data for the inversion are the induction vector responses.

3.2.5 Examples of MT Applications Related to Thesis Research

The MT method has been used successfully to investigate anomalous bodies, structural boundaries and fault systems from local to lithospheric scales. In this section, several applications of MT study pertinent to the thesis research are examined in order to illustrate the method.

MT results from the Alberta Basin (Boerner et al. 1995, 1996) revealed a surface conductive layer with 2-3 km thickness. In the upper-middle crust, the conductivity model
is essentially resistive with numerous conductive bodies interspersed at a depth of 10-20 km. These isolated conductors were interpreted as euxinic graphitic-sulphidic shale sequences formed in Paleooproterozoic foredeep sequences (1.9-2.5 Ga). From west to east, the depth to the conductive subcrustal mantle decreases from ~200 km in Paleooproterozoic Lacombe domain to ~100 km under the Archean Hearne Province (Boerner et al. 1996).

Results synthesized in Jones and Gough (1995) show the southern Cordillera has a variably conductive upper crust, a generally conductive middle crust and a quite conductive lower crust (forming the Canadian Cordillera Regional conductor) (Fig. 3.7). In the upper crust, the enhanced conductivity is likely to be caused by a mixture of contributions from fluid connected films or isolated conducting phases. For example, the regional high conductivity at a few kilometres depth in the Purcell Anticlinorium is explained by mineralization (sulphides etc.). In the middle crust, the depth of the conductor is correlated with the depth of the 450° C isotherm (about 10 km) (Jones and Gough 1995), and it is far more likely to be associated with hot saline water and silicate melt in fractures and other interconnected spaces. The interpretation of MT data can highlight the current physical state of the crust by mapping the brittle-ductile transition, as manifested by the rapid increase in conductivity within the middle crust (Jones and Gough 1995). In the southern Omineca and Intermontane belts, the determination of a thin crust with ~32 km Moho depth and the observed crustal extension and basaltic extrusions during Tertiary time have led to the proposal that an elongated active upflow in the mantle now lies beneath the Intermontane Belt and Omineca Belt (Jones and Gough 1995).

MT has also been used to image major fault systems. Jones (1992) and Jones and
Figure 3.7 Apparent resistivity log$_{10}(\rho_s)$ maps for depth (a) 5 km, (b) 10 km, (c) 20 km, and (d) 30 km from MT surveys in the southern Cordillera (Jones and Gough 1995).
Gough (1995) imaged the Fraser Fault and the Slocan Lake Fault in British Columbia using the MT method. MT models have yielded a 3D image of the Tintina Fault in the northern Rocky Mountain (Fig. 3.8, Ledo et al. 2001). The models image a vertical contact extending from the surface. Around five kilometres depth there is a sub-horizontal detachment zone separating structures of different conductivity. MT and seismic reflection surveys at Parkfield, California, show that the San Andreas Fault zone is characterized by a vertical zone of low electrical resistivity (Unsworth et al. 1997; 1999). This zone is approximately 500 m wide and extends to a depth of ~4000 m. The low electrical resistivity is attributed to high porosity of saline fluids present in the highly fractured fault zone (Unsworth et al. 1997; 1999).

3.3 Time-domain Electromagnetic Method

The description of the TEM method which is provided in this section will focus on methods and the background theory used in my research.

3.3.1 Data Collection and Instruments

The traditional controlled-source EM methods use an alternating current in an ungrounded loop on or above the surface of the earth as a source of EM fields. The primary field of the loop will induce eddy currents in conductors in the earth. The secondary electromagnetic fields due to these induced currents, together with the primary EM field, are recorded by a receiver (Fig. 3.9a). The receiver can be either inside the loop or outside the loop (Fig. 3.9b). The secondary EM field contains information regarding the underground
Figure 3.8. 2D MT inversion models for the profiles crossing the Trench-Tintina Fault (TTF) (from Ledo et. al., 2001). Distances in km.
Figure 3.9 (a) General principle of EM survey (from Kearey and Brooks, 1991). (b) Configuration of TEM survey (after GEONICS 1992a).
conductors, but is much smaller than the primary field.

The TEM method overcomes this problem and records the secondary field in the absence of the primary field using a transient primary field. A strong direct current is passed through an ungrounded loop. After a finite time, this current is abruptly interrupted (creating a step function excitation). According to Faraday's law of induction, the rapid change in transmitter primary field will induce eddy currents in conductors in the earth. The eddy currents produce secondary magnetic fields, which can be measured in the absence of the primary field. In the absence of a source, the induced eddy currents will decay with time. The decay rate of the currents and of the accompanying magnetic field depends on the conductivity, size, and shape of the underlying conductor (Nabighian and Macnae 1991).

The vertical component of the magnetic field \((H_z(t))\) or its time derivative \((\partial H_z(t)/\partial t)\) resulting from the induced currents is the most common form of data recorded in a TEM survey. The latter can be measured in terms of the voltage induced in a horizontal loop or coil.

3.3.2 Basic Theory

This section will first introduce the concept of “induction number”, then describe the basic calculation of EM responses in a uniform half-space and a layered half-space.

**Induction Number \((\beta)\)**

The induction number \(\beta\) is an important parameter for describing the EM results. It is defined,
\[ \beta = \frac{r}{\delta} \] (3.78)

where \( r \) is spacing which is the distance between source and receiver, and the skin depth \( \delta \) can be either \( \delta_{PD} \) or \( \delta_{RD} \).

When the induction number is large, \( i.e. \) the skin depth is very small compared to the spacing, the responses depend on the transmitter-receiver spacing. The terms “wave zone”, “far zone”, or “early time” are used for this case (Spies and Frischknecht 1991). In contrast, when the induction number is small, the skin depth is large, the shape of the measured sounding curves is independent of the transmitter-receiver spacing. The terms “inductive zone”, “near zone”, or “late time” are used.

The initial \( (r=0) \) surface current distribution on an underlying conductive body is independent of the conductivity of the body and is a function only of the size and shape of the conductor. As a consequence, the early-time stage of the currents in the transient process are only weakly dependent on the conductivity of the body. The late-time stage of the currents is strongly dependent on the conductivity of the body.

**TEM Fields for uniform Half-space**

For a homogeneous half-space of conductivity \( \sigma \), if the source is a horizontal transmitter loop with transmitter moment \( m(\omega) \) at the surface and the receiver is located on the surface at offset \( r \), the frequency-domain solution in the cylindrical coordination system (Fig. 3.10) for the EM field components can be written as (Kaufman and Keller 1983, Spies and Frischknecht 1991),
Figure 3.10 Cylindrical coordination system for the horizontally layered conductivity model. $\sigma_i$ and $h_i$ are the conductivity and thickness of $i$th layer.
where the transmitter moment is defined as \( m(\omega) = f(\omega)AN \), \( f(\omega) \) is current, \( A \) is the area and \( N \) is the number of turns. \( J_i \) is the Bessel function of the first kind of order \( i \), \( k \) is defined in Eq. (3.9). \( \lambda \) is the cylindrical wavenumber, and the vertical wavenumber \( \mu \) is given by \( \mu = (\lambda^2 + k^2)^{1/2} = (\lambda^2 - \omega^2 \mu \varepsilon + i\omega \mu \sigma)^{1/2} \) (3.80)

The ideal source for TEM surveys is the impulse function because of its simple Laplace transformation and higher resolution (Spies and Frischknecht 1991). However, the corresponding depth of penetration is low. Triangular and half-sine waveform are better sources for the deeper exploration, but have more complex transformation equations (Spies and Frischknecht 1991). A source commonly used in TEM surveys is step function (Eq. 3.14 and 3.15).

To obtain Laplace-domain solutions for the response of a uniform half-space, Eq. (3.79) and (3.80) are generalized to include the Laplace frequency \( s \) and the source function is specified in the Laplace domain \( m(s) \) (Ferguson and Edwards 1994). The time-domain solution is then obtained with an inverse Laplace transform. For a step function transmitter moment:

\[
m(t) = m_0 u(t)
\] (3.81)
Kaufman and Keller (1983) and Spies and Frischknecht (1991) derived and showed the time-domain solution:

\[
e_e(t) = -\frac{m_o}{2\pi r^3} \left[ 3\text{erf}(\eta r) - \frac{2}{\sqrt{\pi}} \eta r \left( 3 + 2\eta^2 r^2 \right) e^{-\eta^2 r^2} \right]
\]

\[
h_e(t) = -\frac{m_o}{2\pi r^3} e^{-\eta^2 r^2} \left[ \left( 1 + \frac{4}{\eta^2 r^2} \right) I_1(\eta^2 r^2) - I_0(\eta^2 r^2) \right]
\]

\[
h_m(t) = \frac{m_o}{4\pi r^3} \left[ \frac{9}{2\eta^2 r^2} \text{erf}(\eta r) - \frac{1}{\sqrt{\pi}} \left( \frac{9}{\eta r} + 4\eta r \right) e^{-\eta^2 r^2} \right]
\]

where \( e_e(t), h_e(t) \) and \( h_m(t) \) are electric and magnetic field components in time domain and

\[
I_1(x) = -\frac{r}{\sqrt{2\pi a}} e^{a} \int_{0}^{\frac{r^2}{2a}} e^{-\frac{r^2}{2a}} J_1(\lambda r) d\lambda
\]

and

\[
\eta = \sqrt{\frac{\mu_0 \sigma}{4\pi}}
\]

If we consider the early-time response (\( t\to 0, \eta r >> 1 \)) and the late-time response (\( t\to \infty, \eta r < 1 \)), Eq. (3.82) can be simplified. The variation of vertical magnetic field with time \( (\partial h_m/\partial t) \) for early-time and late-time responses can be written as,

\[
\frac{\partial h_m^e(t)}{\partial t} = \frac{9m_o}{2\pi \mu_0 r^5 \sigma}
\]

\[
\frac{\partial h_m^l(t)}{\partial t} = -\frac{m_o h_0^{3/2}}{20\pi^{3/2} \sigma^{3/2} t^{3/2}}
\]

Note that the early-time response is independent of time. If these expressions are applied to
the fields measured over a layered half-space, the early and late time apparent resistivity at the surface can be obtained,

\[
\rho'_{0}(t) = \frac{2\pi\mu_{0}r^{5}}{9m_{0}} \frac{\partial h_{z}}{\partial t}
\]

(3.87)

\[
\rho'_{0}(t) = \frac{m_{0}^{23\mu_{0}}}{20^{21/3\pi^{3/3}} (\frac{\partial h_{z}}{\partial t})^{-2/3}}
\]

(3.88)

For a uniform half-space these equations yield the true resistivity.

The relationship between the voltage measured in a horizontal receiver loop and the variation of magnetic field with time is,

\[
\frac{dh_{z}}{dt} = \frac{aV}{\mu_{0}A_{R}N_{R}2^n}
\]

(3.89)

where \(A_{R}\) and \(N_{R}\) are the area and the number of turns of the receiver, \(n\) is the gain, and \(a\) is a constant which depends on the instrument. Figure 3.10 shows the voltage, normalized by the transmitter moment, at different offset for an uniform half space. The early-time responses are independent of the time (but are not shown in the figure). The late-time responses decay with time and the slope of curves is \(t^{-1.1}\). A sign change in the voltage response occurring at times intermediate between early and late time is observed for offset receiver locations (Fig. 3.11).

**TEM Fields for a Layered Half-space**

Spies and Frischknecht (1991) list the EM responses (electric and magnetic field
Figure 3.11 The TEM voltage responses at different offset for a uniform half space (100 ohm.m). The source is a 100 m transmitter loop. The voltage is normalized by the transmitter current. dash line: negative value.
components and apparent resistivity) for a homogeneous half-space in both the frequency domain and time domain for commonly used transmitter-receiver configurations.

The calculation of EM responses for the different transmitter-receiver configurations for a horizontal layered half-space has been described by Mallick and Verma (1979), Verma and Mallick (1984), Kaufman and Keller (1983) and Fullagar and Oldenburg (1984). For example, Kaufman and Keller (1983) derived the expressions of the EM field components for a vertical magnetic dipole, which has horizontal transmitter and receiver loop, at the surface \( (z = 0) \). In the frequency domain,

\[
E_\phi(r,0,\omega) = \frac{i\omega\mu m(\omega)}{2\pi} \int_0^\lambda \frac{\lambda^2}{\lambda + u_i/R_n} J_i(\lambda r) d\lambda
\]

\[
H_x(r,0,\omega) = \frac{m(\omega)}{2\pi} \int_0^\lambda \frac{u_1}{R_n \lambda + u_i/R_n} \lambda^2 J_i(\lambda r) d\lambda
\]

\[
H_z(r,0,\omega) = \frac{m(\omega)}{2\pi} \int_0^\lambda \frac{\lambda^3}{\lambda + u_i/R_n} J_0(\lambda r) d\lambda
\]

where

\[
R_n = \coth \left( u_1 \frac{h_1}{r} + \frac{u_1}{u_2} \coth \left( u_2 \frac{h_2}{r} + \text{coth}^{-1} \left( u_2 \frac{h_2}{r} \right) \right) \right) \frac{u_2}{u_3} \coth \left( u_3 \frac{h_3}{r} + \text{coth}^{-1} \left( u_3 \frac{h_3}{r} \right) \right)
\]

(3.91)

and

\[
u_i = (\lambda^2 + k^2)^{1/2} = (\lambda^2 - \omega^2 \mu \epsilon + i\omega\mu \sigma)^{1/2}
\]

(3.92)

and where \( h_i \) is the thickness of layer \( i \). Equation (3.91) corresponds to a recursion formula,
as described for the MT work (Eq. 3.31).

In order to obtain the time-domain solution, the first step is that Eq. (3.90) and (3.91) and source function are generalized to the Laplace domain. The time-domain solution is then obtained with numerical inverse Laplace transform. Equation (3.87) and (3.88) can be applied to calculate the apparent resistivity of the layered half-space.

Figure 3.12 shows the time-domain responses of a two-layer model with varying conductivity contrast. The early-time responses (<6×10⁻² s) are independent of time. The late-time responses decay with time and show multiple sign reversals. The sign reversals observed at a constant offset are related to the conductivity contrast of the model. Therefore, the sign reversals of responses are one of the aspects that can be used to understand the variation of conductivity.

3.3.3 Data Processing

This section will describe the noise levels in TEM survey and the reduction of noise.

Environmental Noise Level

The EM noise recorded during the TEM surveys contains contributions from ionospheric and atmospheric sources (McCracken et al. 1984; 1986), from local EM sources such as powerlines and pipelines, and from the instrument itself. The noise can distort the solution and decrease the exploration depth.

In order to understand the environmental noise level at a survey site and compare it with the signal level, the noise level can be determined by taking a number of recordings
Figure 3.12 TEM responses (normalized voltage) of two-layer models with varying conductivity contrasts (Spies and Frischknecht, 1991). $d/r = 0.25$, $r$ is the source-receiver separation.
with the transmitter turned off and averaging the results to reduce the statistical scatter in the data. This method records only the environmental noise and omits any noise from the transmitter electronics. The typical environmental noise levels of TEM instruments after data stacking range from $10^{-9}$ to $10^{-10}$ V/m$^2$ (Fitterman and Stewart 1986; Fitterman 1989), but under quiet conditions can drop as low as $10^{-11}$ V/m$^2$ (Fitterman and Stewart 1986).

The recordings of TEM system include both the signal and noise. In order to increase the signal to noise ratio (SNR), TEM instruments stack the individual decays. Stacking of $N$ responses at each receiver position will reduce the noise amplitude by $\sqrt{N}$ i.e., increase the SNR (defined in terms of power spectral density) by a factor of $N$. Additional recording and averaging of multiple responses will further increase the SNR. The SNR also can be increased by increasing the transmitter moment (i.e. either the current or the loop area) (Fitterman 1989, Massie 1993).

### 3.3.4 TEM Forward Modelling and Inversion

**Forward Modelling**

Most TEM modelling uses a numerical Hankel transform (e.g. Chave and Booker 1987) and a numerical inverse Laplace transform (e.g. Ferguson and Edwards 1994) to calculate the EM responses (equations 3.90-3.92). Various methods have been developed to accurately evaluate Hankel transforms, such as the linear digital filter method (Anderson 1979, Guptasarma and Singh 1997) and the direct numerical integration scheme with Padé convergence acceleration (Chave 1983).

For 1D structure, the response at the surface can be calculated in the Laplace
wavenumber domain using the recursion equation (e.g., Eq. 3.85 and 3.86). In forward modelling of the response of 2D or 3D structures, the kernel functions are complicated since the impedance is no longer a simple recursion relationship. Numerical solutions, e.g. finite-difference, finite-element and integral equation, have been developed to calculate the EM fields. The methods are similar to those used for MT modelling.

Stoyer and Greenfield (1976) and Oristaglio and Hohmann (1984) developed 2D finite-difference methods for computing the EM fields. Wang and Hohmann (1993) developed a finite-difference solution for 3D TEM problems. Coggon (1971) used the finite-element method to compute the EM anomalies of 2D structures. Pridmore et al. (1981) demonstrated that the finite-element method may be used to model the complex 3D Earth. Gupta et al. (1989) analysed the finite element solution for 3D TEM responses. The TEM integral equation solution has been discussed by SanFilipo and Hohmann (1985) and Newman et al. (1986). Eaton and Hohmann (1989) developed an integral-equation based 3D inversion algorithm for TEM soundings that uses imaging results as a starting model.

Inversion

1D Inversion

1D inversion methods include linearized least-squares and Backus-Gilbert methods. The least squares method minimizes an objective function of the model to find a special model. For example, the inversion methods by Sandberg (1990) and Farquharson and Oldenburg (1993) minimize the misfit between the model and observed responses. Fullagar and Oldenburg (1984) developed finite-element method based on the Backus-Gilbert method
to compute the responses of the horizontal loop.

An interpretation method providing an alternative to full inversion is the imaging technique. It is based on tracking a descending image of the transmitting loop that replicates the observed magnetic field. The descending image consists of currents diffusing in the earth. The variation in the depth and speed of these image currents with time can be converted into a conductivity-depth section. This is a smoothed approximation of the true conductivity structure (Macnae and Lamontagne 1987; Smith et al. 1994).

2D and 3D Inversion

Inversion codes have been developed for various controlled-source EM problems. Li and Oldenburg (1994) developed a 3D finite difference algorithm to invert DC resistivity data. Ellis and Oldenburg (1994) presented a conjugate gradient method, which minimizes a nonlinear objective function and applies the adjoint equation to compute the gradient of the objective function. Zhang et al. (1995) applied conjugate gradient relaxation to solve 3D DC resistivity forward and inversion problems. There are few commercially available codes for full 3D inversion of TEM data. These methods will be developed and applied to TEM forward modelling and inversion in the future.

3.3.5 Examples of TEM Applications Related to Thesis Research

The TEM method has been used successfully to investigate fractures, fault systems, and shallow (<1000 m) groundwater resources. In this section, several TEM applications which are pertinent to the thesis research are examined.
The TEM method was used in the exploration for, and characterization of, underground repositories for high-level nuclear waste in the Yucca Mountain, Nevada Test Site (Frischknecht and Raab 1984). The structure of this area is complex, with geologic boundaries including faults and lithologic contacts. Because of lateral changes in the resistivity, the soundings from large-scale EM methods (e.g. geoelectrical depth sounding and horizontal profiling method) represent 2D or 3D structures and are difficult to interpret. The TEM results provided a more localized response. They define a three-layer model which fits all of the data very well, and delineated the major faults or fault zones along the survey traverse (Fig. 3.13). Schlumberger resistivity soundings detect only the first interface mapped by the TEM method, and appear to be more distorted by the lateral variations than the TEM sounding curves (Frischknecht and Raab 1984).

TEM surveys in Fraser Valley, British Columbia in 1994 showed that EM methods can map surficial geological features for ground water to depths between 150 m and 200 m (Best et al. 1995). Figure 3.14 shows the TEM models at three sites. At site VED-100, the TEM model delineates 140 Ωm lacustrine silts and clays with 6 m thickness overlying conductive (3.5-21 Ωm) marine sediments. The resistive bottom layer (4500 Ωm) is crystalline bedrock. At site DHHOS, the TEM model provided a four-layer model: a very shallow resistive surface layer, salt water saturated layer with 1.26 Ωm resistivity and 35 m thickness, a brackish water saturated layer with 6.14 Ωm resistivity and 24 m thickness, and a resistive bottom layer with >400 Ωm resistivity. This model successfully reveals the variation of salinity of groundwater. The TEM model at site AGA-120 produced a three-layer model, which is different with the models at site VED-100 and DHHOS. The resistive top
Figure 3.13 Calculated resistivity versus depth cross-section for one TEM survey line at Nevada Test Site (Frischknecht and Raab, 1984).
Figure 3.14 TEM soundings and inversion models at three sites in Fraser Valley, British Columbia in 1994 (from Best et al. 1995).
layer with 840 Ωm resistivity and 41.4 m thickness is interpreted to be sands and gravels containing little or no clay. The second layer with 100 Ωm resistivity and 97 m thickness may correspond to drift with increased clay content. The very conductive bottom layer (12 Ωm resistivity) is probably marine sediments.

Fitterman and Stewart (1986) also found TEM method could map the fresh/salt water interface in aquifer. TEM sounding is sensitive to the present of saltwater. However, the sensitivity depends on the resistivity, depth and thickness of saltwater layer.
Chapter 4 Magnetotelluric Study of SNORCLE Transect Corridor 1 and 1A

In this chapter, I will first introduce the SNORCLE transect Corridor 1 and 1A survey location (Fig. 4.1), the MT instruments used in the SNORCLE survey and the data collection. This description will be followed by a discussion of the determination of the impedance and tipper from the stochastic electric and magnetic field time series (Vozoff 1991). Next I will describe the use of the Groom Bailey regional strike angle, direction of maximum phase difference and induction vectors to examine the lateral and depth variation of the geoelectric strike along the survey line. To reduce the distortion from near-surface inhomogeneities Groom Bailey decomposition and static shift removal were applied to the data. These procedures and the results will be described in the chapter. The chapter will include a detailed description of the one-dimensional forward modelling and inversion used to determine the structure of shallow Phanerozoic sedimentary rocks, and the 2D forward modelling and inversions used to examine the crustal and lithospheric structure. In the last section of this chapter, MT results are compared with seismic reflection, seismic refraction, teleseismic, gravity and magnetic data in order to perform the final geological interpretation.

4.1 MT Survey

This section will introduce the SNORCLE MT survey location, survey configuration
Figure 4.1 Selected tectonic elements and structures of the northwest Canadian Shield (after Hanmer, 1988). Na: Nahanni terrane, FS: Fort Simpson terrane, HO: Hottah terrane, GB: Great Bear magmatic arc, HR: Hay River terrane, GSLsz: Great Slave Lake shear zone, BH: Buffalo Head terrane, BFZ: Bathurst fault zone, MFZ: McDonald fault zone. The circles show the location of MT sites on Corridor 1 and 1A.
and data collection.

4.1.1 Site Locations

Magnetotelluric soundings were completed at 60 sites along LITHOPROBE SNORCLE transect Corridor 1 and 1A in the southwestern part of the Northwest Territories, Canada during the summer of 1996 (Fig. 4.1). On Corridor 1, there were 25 sites on the Nahanni, Fort Simpson and Hottah terranes, 8 sites on the Great Bear Magmatic Arc, and 12 sites on the exposed Slave craton. Phanerozoic sedimentary rocks overlie most of the terranes except the Slave Province (Fig. 4.1). On Corridor 1A, there were 15 sites on a transect crossing the Great Bear Magmatic Arc, GSLsz and Buffalo Head terrane. As shown Fig. 4.1, the position of the GSLsz on geological maps (e.g., Hanmer et al. 1992) is between sites 153 and 154, but the centre of the magnetic low with which the shear zone is correlated (Eaton and Asudeh 2001) lies between sites 155 and 156. To the northwest of the GSLsz the Corridor 1A crosses a 10 km wide Middle to Upper Devonian barrier complex (sites 151 and 153), the Pine Point barrier which forms part of the larger scale Presqu'ile Barrier.

Table 4.1 lists the geographic and geomagnetic location of the sites. Because the sites were located at high geomagnetic latitudes (66°- 69°N), auroral geomagnetic activity may have produced non-uniform and non-stationary source fields, causing appreciable bias to MT responses (Jones 1980; Garcia et al. 1997).

4.1.2 Data Collection and Survey Configuration

Shorter period (10^4-100 s) data for each site were acquired using the Phoenix V5 MT
Table 4.1 Geographic and geomagnetic location of MT observation sites along SNORCLE Transect Corridor 1 and 1A.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Recording Time (1996)</th>
<th>Geomagnetic Latitude (N)</th>
<th>Geomagnetic Longitude (W)</th>
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<td>62° 39' 47&quot;</td>
<td>116° 12' 32&quot;</td>
<td>08/08-09/19</td>
<td>68° 59'</td>
<td>65° 48'</td>
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<td>62° 30' 45&quot;</td>
<td>116° 27' 49&quot;</td>
<td>08/06-09/03</td>
<td>68° 49'</td>
<td>65° 42'</td>
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<tr>
<td>103</td>
<td>62° 45' 55&quot;</td>
<td>115° 51' 14&quot;</td>
<td>08/08-08/26</td>
<td>69° 09'</td>
<td>64° 06'</td>
</tr>
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<td>62° 43' 52&quot;</td>
<td>115° 39' 49&quot;</td>
<td>08/07</td>
<td>69° 09'</td>
<td>64° 24'</td>
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<td>62° 28' 13&quot;</td>
<td>114° 36' 22&quot;</td>
<td>08/10-08/20</td>
<td>69° 05'</td>
<td>63° 54'</td>
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<td>64° 42'</td>
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<td>115° 14' 22&quot;</td>
<td>08/12-08/26</td>
<td>69° 08'</td>
<td>63° 00'</td>
</tr>
<tr>
<td>108</td>
<td>62° 33' 31&quot;</td>
<td>115° 05' 16&quot;</td>
<td>08/14-08/22</td>
<td>69° 05'</td>
<td>63° 12'</td>
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<td>08/15-08/22</td>
<td>69° 05'</td>
<td>63° 36'</td>
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<td>08/16-06/22</td>
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<td>62° 12'</td>
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<tr>
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<td>114° 04' 47&quot;</td>
<td>08/17-08/27</td>
<td>69° 14'</td>
<td>62° 30'</td>
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<td>08/19-09/08</td>
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<td>113° 34' 51&quot;</td>
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<td>10/12-10/18</td>
<td>67°02'</td>
<td>59°12'</td>
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* The sites with one-day recording listed have only V5 data.
system. A commercial contractor, Phoenix Geophysics, completed these measurements using three nights of recording at each site. Additionally, one to four week recordings of long period data were completed at each location (except sites 159 and 160) using LiMS units of the GSC. These soundings provided a longer period range response (20 - 3 × 10^4 s). They were done by a contracted field assistant (Mr. Nick Grant) and researchers from the University of Manitoba, the University of British Columbia and the GSC.

Both V5 and LiMS systems measure time series of the natural time-varying fluctuations in the earth's electric and magnetic fields (Chapter 2). Table 4.2 lists the configurations of the Phoenix V5 and LiMS systems. The horizontal electric field variations are measured in two perpendicular directions by 100 m electric dipoles coupled to the ground by electrodes. Wherever possible, the two perpendicular directions for recording orientation were the geomagnetic north and east. In the V5 survey, horizontal magnetic variations were measured in this recording orientation by induction coil sensors consisting of iron-cored coils. The vertical magnetic field component was measured by a multi-turn square loop (laid on the ground horizontally to measure the vertical magnetic flux). The remote-reference was a second site recording simultaneously.

For the LiMS system also, the horizontal electric variations are measured in the geomagnetic north and east recording orientation using by 100 m dipoles coupled to the ground by electrodes. The horizontal magnetic variations are measured along the geomagnetic north-south and east-west directions using fluxgate sensors, and the vertical magnetic variation is measured using a vertical fluxgate sensor. The magnetic field sensors are 3-component NAROD ring-core fluxgate sensors mounted in a waterproof cylinder.
Recordings taken during the same time period at adjacent sites (typically 20-40 km apart) provided remote-reference time-series.

**Table 4.2** Configurations of V5 and LiMS systems used in SNORCLE.

<table>
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<tr>
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<th>LiMS</th>
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<td>$10^4$ s - 100 s</td>
<td>20 s - 30,000 s</td>
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<tr>
<td>Record durations</td>
<td>three night</td>
<td>one to four weeks</td>
</tr>
<tr>
<td>Horizontal electric field sensors</td>
<td>porous Pb-PbCl$_2$ electrodes.</td>
<td></td>
</tr>
<tr>
<td>E-lines</td>
<td>the spacing of a pair of electrodes is 100 m</td>
<td></td>
</tr>
<tr>
<td>Horizontal magnetic field sensors</td>
<td>iron cored coils with an integral low noise preamplifier.</td>
<td>3-component NAROD ring-core fluxgate sensor mounted in a waterproof cylinder</td>
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<tr>
<td>Vertical magnetic Field</td>
<td>10 m square air loop</td>
<td></td>
</tr>
<tr>
<td>Power</td>
<td>12V battery</td>
<td>two 12 V batteries</td>
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<tr>
<td>Environmental conditions</td>
<td>operating temperature range $-10^0$ C to $+50^0$ C.</td>
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</table>

Electrodes used for both surveys were porous Pb-PbCl$_2$ electrodes, which have optimal noise characteristics (Vozoff 1991). The LiMS installation at each site always followed the removal of the Phoenix equipment and used the same electrode holes, but a different set of electrodes. Therefore, the LiMS and V5 data should have the same distortion associated with near-surface inhomogeneities.

### 4.2 Data Processing

#### 4.2.1 V5 Data

Phoenix Geophysics provided the V5 MT responses in the form of auto- and cross-
power spectra of five electric and magnetic field components and two remote reference field components in SEG EDI (Society of Exploration Geophysicists Electronic Data Interchange) format. The spectra were computed using robust statistical techniques based on the Jones-Jödicke approach (Jones and Jödicke 1984; Jones et al. 1989). All spectra were rotated to geographic north. The data were imported into a GEOTOOLS MT Workstation for processing, analysis and display.

The spectra and the SNR of the electric and magnetic fields at a typical site (site 127) are shown in Fig. 4.2. The SNR is defined as the ratio of the signal to the signal plus noise in these plots. Site 127 is located in the Hottah terrane, where Paleozoic sedimentary rocks overlie the Precambrian shield.

The magnitude of horizontal electric and magnetic field components and the SNR is lower around periods of 1 s and $5 \times 10^4$ s (Fig. 4.2a, b, c, d). These two periods are defined as the "dead bands" in MT and AMT soundings. For the AMT dead band (around $5 \times 10^4$ s), the source EM fields from the atmosphere and ionosphere are relatively weak. For the MT dead band, there is relative low signal level in the transition between different sources of the MT field. Thunderstorm activity is the primary source above 1 s, and micropulsations in the magnetic field of the earth are main sources below 1 s. Wind noise may also contribute to the lower SNR around 1 s. Except for these two bands, the spectra of the horizontal magnetic field components increases with period in a stable manner, with negligible noise effects. The ratio, signal/(signal and noise), is high and close to 1.0. These characteristics of the horizontal field components can be observed at most sites.

For site 127, the amplitude spectrum of the vertical magnetic field is lower than those
Figure 4.2 Spectra and SNR (ratio of signal to signal+noise) of electric and magnetic components of V5 data at site 127.
horizontal magnetic fields (Fig. 4.2b). The very low amplitude spectrum of the vertical magnetic field at periods less than 1 s suggests the surface sedimentary rocks are very uniform with no lateral near-surface conductors. The jump in the spectrum at 0.01 s may be due to the use of different amplifiers for the different period bands. The low signal level of the vertical magnetic field at short periods means the SNR is relatively low.

Examination of the coherence at site 127 shows that each horizontal magnetic field component is coherent with the orthogonal electric field component \((H_x, H_y)\) with the coherence close to 1.0 (Fig. 4.3b, 4.3c). The horizontal magnetic field component is less coherent with the parallel electric field component \((H_x, E_y)\) at periods less than 10 s, with values of 0.1-0.2. The coherence ranges between 0.4-0.9 at periods longer than 10 s. The coherence results imply an approximate 1D structure or a 2D structure aligned with the data acquisition orientation at period less than 10 s (upper-middle crust) and a more complex structure at periods longer than 10 s (middle-lower crust and mantle). For most of the other sites on the corridors, the coherence values are also close to 1.0 for perpendicular horizontal magnetic and electric field components, and coherence values are less than 0.5 for parallel horizontal magnetic and electric field components.

Figure 4.4 shows apparent resistivities and phases for off-diagonal component of the unrotated impedance tensor. The \(xy\) mode corresponds to north-south electric fields and east-west magnetic fields. The \(yx\) mode corresponds to east-west electric fields and north-south magnetic fields. Small error bars indicate the error is <2%. Larger error bars at periods around \(10^3\) s and at longer periods (>100 s) are due to lower signal level and correspond to error of 3%.
Figure 4.3 Coherence between the electric and magnetic field components at site 127.
Figure 4.4 Apparent resistivity (a) and phase (b) of V5 data at site 127.
4.2.2 LiMS Data

LiMS instruments record the time series of the electric and magnetic fields. Figure 4.5 shows an example of the recordings at site 130. During the recording there was strong geomagnetic activity associated with the auroral electrojet, such as around midnight each day for the period September 10 to 15.

In order to remove the effects of non-uniform source fields produced by the auroral electrojet, three methods of spectral analysis (Chapter 3) were used to calculate MT impedances:

(1) single-station-weighted cascade decimation;
(2) all-station-weighted cascade decimation (Jones and Jödicke 1984; Jones et al. 1989); (3) robust remote reference estimation (Chave et al. 1987; Chave and Thompson 1989). The spectral analysis was performed with the time series rotated to the true north relative to a site declination.

Figure 4.6 shows the apparent resistivity and phase derived from the impedances calculated from the different spectral estimate methods at site 130. The results of the all-station-weighted cascade decimation show less variance between adjacent estimates and smaller error bars than the single-station-weighted cascade decimation (Fig. 4.6a). The former method is more effective at removing the incoherent noise, because it averages more signals. The results from the robust remote reference estimation are close to the all-station-weighted results at periods less than 2000 s (Fig. 4.6b). However, the robust results are more erratic with larger error bars at periods longer than 2000s. The bias of the MT responses at long periods, caused by non-uniform field sources, may not be corrected properly by the
Figure 4.5 The raw time series for the magnetic components \((B_x, B_y, B_z)\) and the electric components \((E_x, E_y)\) at site 130 for the period from September 9 to 16, 1996 (Universal Time). The numbers on the bottom are the scales of the field components in the order plotted.
Figure 4.6 The apparent resistivities and phases at site 130 determined from the different processing methods. (a) The results of the single-station-weighted cascade decimation (black symbols) and the all-station-weighted cascade decimation (open symbols). (b) Robust remote reference (black symbols) which used site 122 as remote reference and all-station-weighted cascade decimation (open symbols). Circle: xy mode, square: yx mode.
robust methods used here. The results from all-station-weighted cascade decimation will be used in the following analysis. This method is based on the largest number of sites, so it should provide the best estimate of the uniform-filed response.

4.2.3 Data Merge and Response Analysis

In order to obtain a MT response extending over the full period range (10^{-4} - 10^{4} s), the V5 and LiMS responses, need to be merged. The V5 data at longer periods (>100 s) is relatively poor with larger error bars (Fig. 4.4), therefore, we merged V5 data for periods less than 20 s and LiMS data with periods larger than 20 s.

Figure 4.7a shows V5 and LiMS data at site 130 before merged. In the period range 20-1000 s, there is a significant mismatch of apparent resistivity between the LiMS and V5 responses, observed at most sites. The difference is larger for apparent resistivity than for phase and strongest for the xy mode. The LiMS results from the all-station cascade method are a little closer to the V5 estimate than those from the robust estimate. The effect is not consistent between the sites recorded by the same LiMS instruments indicating it is not due to the LiMS equipment. Detailed comparison of data collected with V5 and LiMS systems revealed the problem arose from a small, systematic error in the calibration of the V5 system.

The data were corrected by GSC researchers following provision of new calibrations by Phoenix Limited in 1998. The corrected data shows a better match in merged period range of V5 and LiMS data (Fig. 4.7b). The merged V5 and LiMS data were rotated to the geographic north coordinate system, converted into EDI format, and imported into the GEOTOOLS MT Workstation for analysis. Appendix A contains plots of the merged
Figure 4.7 (a) V5 (opened symbols) and LiMS (black symbols) data for site 130 before merging. LiMS time series are processed by all-station-weighted cascade decimation method. Circle: $xy$ mode; square: $yx$ mode. The offset between the V5 and LiMS data is caused by incorrect calibrations of the V5 system. The data was recorded along geomagnetic north. (b) Merged data after the correction of the calibrations. The data was rotated to geographic north.
apparent resistivity and phase responses for all 60 sites on Corridor 1 and 1A.

4.3 Relationship of Period and Depth

Two parameters used to evaluate the exploration depth of the MT response were introduced in last chapter: skin depth and Schmucker depth $z^*$ (Eq. 3.34). The skin depth is an effective "maximum" exploration depth for a uniform half-space model, the depth at which the fields decay to a magnitude of $1/e$ (37%) of the surface value (Eq. 3.13). In contrast the Schmucker depth represents a weighted mean depth of the inductive current distribution (Eq. 3.34). Both measures are period-dependent. The Schmucker depth is shallower than the skin depth.

Figure 4.8 shows the Schmucker depth for a representative site in the Nahanni terrane (site 143), the Fort Simpson terrane (site 135), the Hottah terrane (site 127), the Great Bear Magmatic Arc (site 115) and the Anton terrane (site 106) along Corridor 1. Table 4.3 lists the Schmucker depth at several periods.

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Figure 4.8 Schmucker depth versus period at some sites along Corridor 1. (a) $xy$ mode, (b) $yx$ mode.
The Schmucker depth for the $xy$ and $yx$ modes are quite similar. The sites in the Great Bear Magmatic Arc and Slave Province (sites 115 and 106) have larger Schmucker depth than sites in the Nahanni, Fort Simpson and Hottah terranes (sites 143, 135 and 127). This difference is caused by the absorption of signal by the conductive Phanerozoic sedimentary rocks overlying the Nahanni, Fort Simpson and Hottah terranes, which decreases the exploration depth.

Figure 4.9 shows the Schmucker depth for the representative sites on Corridor 1A, and Table 4.4 lists the values at some periods. Sites in the Buffalo Head terrane (sites 153 and 156) have larger Schmucker depth than sites in the Great Bear Magmatic Arc (sites 148 and 150). The exploration depth for sites near the GSLsz (sites 152 and 151) are close to those for sites in the Great Bear Magmatic Arc at periods less than 0.1 s; and are close to those for sites in the Buffalo Head terrane at periods longer than 1 s. These results indicate a structural transformation across the GSLsz.

<table>
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<td>227 229</td>
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<td>212 255</td>
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</table>
Figure 4.9 Schmucker depth versus period at some sites along Corridor 1A. (a) xy mode, (b) yx mode.
4.4 Geoelectric Strike Determination

As discussed in Chapter 3, a variety of methods can be used to define the geoelectric strike, including (real and imaginary) induction vectors, maximum phase split directions and Groom-Bailey (GB) regional strike. Figure 4.10 shows the three strike parameters. The period ranges used here in analysing the GB strike are 0.003-0.16 s, 0.2-14.2 s, 20-1700 s and >200 s. The periods of the induction vectors and maximum phase split estimate are 0.01 s, 10 s, 100 s and 2500 s. The exploration depth for these periods may be determined from Tables 4.3 and 4.4. The resolution scale on the GB plot, represented by the length of the strike symbol, is inversely proportional to misfit $\chi^2$ error between the GB decomposition impedance and the observed impedance. The GB plots do not include the unresolved strike angles. The programs used to calculate the GB regional strike, induction vector and maximum phase difference were provided by the GSC, and were written by Dr. Ross Groom and Dr. Alan Jones respectively. The results indicate the significant differences between the near-surface and deeper resistivity structures, and between the east and west ends of the profile.

4.4.1 Geoelectric Strike Directions for Corridor 1

<10 s Period

At short periods the real and imaginary induction vectors have magnitudes less than 0.05 (Fig. 4.10a). The maximum phase differences are less than 10° at most sites (except the sites in the Slave Province). Because of the ±90° ambiguity when GB regional strike angle is used to determine the strike direction, the strike could also be N30°W. The strike angles
Figure 4.10. Groom-Bailey regional strike (symbol length is inversely proportional to the $\chi^2$ error between the GB decomposition impedance and observed impedance), maximum phase split, real induction vector and imaginary induction vector along the SNORCLE Transect Corridor 1 and 1A at periods (a) 0.01 s, (b) 10 s, (c) 100 s, (d) 2500 s. There is +/-90 degree ambiguity when using the GB strike or maximum phase split to determine the strike direction.
Strike Angle

Maximum Phase Difference

Real Induction Vector

Imaginary Induction Vector

(b) strike directions at 10 s
Strike Angle

Maximum Phase Difference

Real Induction Vector

Imaginary Induction Vector

( c ) strike directions at 100 s
(d) strike directions at 2500 s
are irregular in Slave Craton.

The period range of Fig. 4.10a (<0.01 s) corresponds to penetration depth of <100 m at sites with the Phanerozoic cover and <1.5 km at sites on the Precambrian shield. The small induction vectors and maximum phase differences indicate low conductivity contrast and almost 1D resistivity structure in the Phanerozoic sedimentary rocks. The skew values at this period range are less than 0.05 (very close to zero) in the Phanerozoic areas (Fig. 4.11), which supports this interpretation. The irregular strike angles in the Slave Craton indicate a more complex near-surface conductivity structure in the exposed Proterozoic and Archean areas.

10 s Period

This period corresponds to signal penetration of several kilometres for sites lying on the Phanerozoic cover to several tens of kilometres for sites on Corridor 1 in the Great Bear Magmatic Arc and Slave Province (Table 4.3). The observed orientations at 10 indicate:

- The real induction vectors at 10 s period have a magnitude of about 0.2 in the Nahanni and Great Bear Magmatic Arc terranes and point to the Fort Simpson terrane where the magnitude of real induction vectors is <0.2. There is a reversal of the east-west component of vector direction near the margin of the Nahanni and Fort Simpson terranes. The real induction vectors point a southward in the Slave Province and have a southward component elsewhere. The imaginary induction vectors have a magnitude of less than 0.1 but have a southwards component at many sites.

- The maximum phase differences are >25° at most sites except for the sites in the Fort
Simpson terrane and eastern Nahanni terrane. The direction of maximum phase is close to N20°E in the western Nahanni and Hottah terranes and in the Great Bear Magmatic Arc, east-west in the Fort Simpson and eastern Nahanni terranes, and variable in the Slave Province.

- The GB regional strike angles are close to N20°E along the whole profile, and are parallel to the maximum phase difference direction in western Nahanni and Hottah terranes and the Great Bear Magmatic Arc. However, the strike direction in the Fort Simpson terrane is slightly different from the maximum phase difference direction.

The reversal of the east-west component of the induction vector components across the Fort Simpson terrane must be explained by a north-south striking structure in the deep Phanerozoic cover or else in the upper 2 km of the Precambrian crust because the penetration depth is several kilometres around the Fort Simpson terrane. The overall southward component may be caused by the increasing thickness of the Phanerozoic sedimentary rocks to the south. Pervasive orientations of induction vectors apparently unrelated to local structures have been interpreted as being related to nearby sedimentary rocks in other locations where thick sequences of sedimentary rocks occur adjacent to more resistive crystalline rocks (e.g. Wang and Lilley 1999). The long period response on the LITHOPROBE Tans Hudson Orogen line includes induction vectors that point towards thick sediments in the Williston Basin (Stevens 1998).

100 s Period

This period corresponds to signal penetration depth 25-30 km for sites lying on the
Phanerozoic cover and about 100-200 km for sites in the Great Bear Magmatic Arc and Slave Province (Table 4.3). The observed orientations at 100 s reveal:

- The real induction vectors have a magnitude of about 0.1 and point to west in the eastern Nahanni and Fort Simpson terranes, a magnitude of about 0.25 and a southwest direction (S30°-60°W) in the Hottah terrane and western Great Bear Magmatic Arc, a magnitude of about 0.25 and a west and northwest direction in the eastern Great Bear Magmatic Arc, and a variable magnitude (0-0.6) and direction in the Slave Province. The imaginary induction vectors have a magnitude of about 0.25 and point to southwest in the Nahanni, Fort Simpson and Hottah terranes and western Great Bear Magmatic Arc. Their magnitude and direction are variable in the western Great Bear Magmatic Arc and Slave Province.

- In the western part of the profile the magnitude of maximum phase difference is 20°-50° and the direction of phase split is approximately N45°E, whereas in the Great Bear Magmatic Arc the magnitude is about 15° orientated N60°E-N90°E. The maximum phase difference has a northwestward orientation in the Slave Province.

- The GB regional strike angles are near N45°E in the west of profile, near N60°E in the Great Bear Magmatic Arc, and near N30°E in Slave Province.

The variable real and imaginary induction vector direction and larger swift's skew values (Fig. 4.11) observed in the Slave Province indicate the presence of 3D structures or galvanic distortion. The similarity of the maximum phase difference direction and the regional strike angle from the Nahanni terrane to the west Great Bear Magmatic Arc indicate a strike direction of approximate N45°E for the structure.
Figure 4.11: Skewness contour maps at period 0.01 s, 10 s, 100 s and 2500 s around SNORICLIE transect Corridor 1 and IA.
>10^d s Period

This period corresponds to signal penetration through the deepest crust and into lithospheric mantle (Table 4.3) at all sites on Corridor 1.

- The real induction vectors point southwest (~S45°W) and have a magnitude of >0.2 in the western part of the Great Bear Magmatic Arc. The direction and magnitude of the real induction vectors are variable in the Great Bear Magmatic Arc and the Slave Province. The imaginary induction vectors are neither parallel nor perpendicular to the real part, and their directions are variable. The magnitude of the imaginary induction vectors in the Fort Simpson area, ~0.3, is larger than that in the Hottah terrane and Great Bear Magmatic Arc.

- In the western part of the profile the direction of maximum phase difference is approximately north-south (with a ±90° ambiguity), and in the Great Bear Magmatic Arc the direction is N30°W- N45°W. The maximum phase direction is variable in the Slave Craton.

- The GB regional strike trends north-south in the Nahanni, Fort Simpson and Hottah terranes, but is variable in the Great Bear Magmatic Arc. The orientation changes to ~N30°E in the Slave Craton.

The variable real and imaginary induction vectors, maximum phase direction and GB regional strike indicate significant spacial variations in the deep resistivity structure across Corridor 1.

*Rose Diagram of Strike Angles*

Rose diagrams can be used to synthesize the strike information. Figure 4.12 shows
True North

0.003-0.16 s

Total 29 sites

0.2-14.2 s

Total 27 sites

20-1700 s

Total 29 sites

>2000 s

Total 28 sites

**Figure 4.12** Rose diagram of the Groom-Bailey regional strike angle for four period ranges. A total of 29 sites located in Nahanni terrane to Great Bear magnetic arc along Corridor 1 are included. The angle bin is 10 degree.
rose diagrams of the GB regional strike angle at different period ranges for sites located in terranes between the Nahanni terrane and the Great Bear Magmatic Arc. At short periods (0.003-0.16 s) there is a uniform distribution from 20°-70° with a number of sites having strikes between 50°-70°E; 20°-30° is the most common strike angle for periods between 0.2 s and 14.2 s. The strike angle swings further east-west, to about 40°-60° at periods 20-1700 s. The strike angle is north-south at periods longer than 2000 s.

For 2D modelling and inversion, the plane of the resistivity models should be perpendicular to the strike direction. The average strike angle for the period range of 0.1 s to 1000 s, N34°E, will be used for initial modelling and inversion. This period range is appropriate for examination of upper crustal to lithospheric depths (Table 4.3). The impedance of each site was rotated to this direction. The TE mode corresponds to the electric fields aligned the strike direction and the TM mode corresponds to the electric fields perpendicular to the strike direction.

4.4.2 Geoelectric Strike Directions for Corridor 1A

<10 s Period

The period range at <10 s period corresponds to penetration depth of <80 m for sites on Corridor 1A. The magnitudes of the real and imaginary induction vectors are <0.05 (Fig. 4.10a). The maximum phase differences are <10°. The GB regional strike direction is ~N60°E (or N30°W) around the GSLsz, and is irregular in the Great Bear Magmatic Arc. The real and imaginary induction vectors and maximum phase difference results are similar to results from the adjacent Proterozoic terranes on Corridor 1, indicating a response dominated
by the approximately 1D resistivity structure of the Phanerozoic sedimentary rocks.

10 s Period

This period corresponds to signal penetration of 5 km to the west of the GSLsz and about 17 km for sites to the east of the GSLsz (Table 4.4). At 10 s period, the real induction vectors have a direction of southwest in the western part of the GSLsz, with a magnitude of about 0.2-0.3. The direction of real induction vectors at sites east of the GSLsz is variable and the magnitude in the range is 0.1-0.2. Imaginary induction vectors have a magnitude less than 0.1. The maximum phase difference has a magnitude of 45° and a direction of about N30°E at most sites of Corridor 1A, except two sites in the middle of the Corridor 1A (site 152 and 151). At site 152 and 151, maximum phase difference has a magnitude of 15° and a direction of N45°E. The GB regional strike direction is similar to the direction of maximum phase difference.

100 s Period

At 100 s period, corresponding the penetration depths of 30 km in the west of the GSLsz and >55 km in the east of the GSLsz (Table 4.4), the real induction vectors in the Great Bear Magmatic Arc point southwest. They change in direction to northwest and west in the Buffalo Head terrane. The magnitude of imaginary induction vectors is small at most sites. The maximum phase difference direction is approximately N60°E in the Great Bear Magmatic Arc, and is close to north-south in the Buffalo Head. The regional strike direction is approximately N60°E at sites crossing the GSLsz. The variation of the induction vectors
and maximum phase difference direction along Corridor 1A at periods of 10s - 100 s implies a change in the resistivity structure across the GSLsz in middle to lower crust.

>10¹ s Period

This period corresponds to signal penetration through the deepest crust and into lithospheric mantle (Table 4.4). At this period, the real induction vectors have a magnitude of 0.2-0.3 and point to the southwest in the western part of the Great Bear Magmatic Arc, and to the west in the Buffalo Head terrane. Imaginary induction vectors have a magnitude of >0.3, and a direction of east to southeast in the Buffalo Head terrane and variable direction in the Great Bear Magmatic Arc. The strike and maximum phase difference directions trend close to north-south in the Great Bear Magmatic Arc, but the strike direction changes to about N60°E and the maximum phase difference direction to about N30°W in the Buffalo Head terrane. The long period induction vectors, maximum phase difference and regional strike angle all show a difference across the GSLsz, indicating an azimuthal dependency around the GSLsz in the lower crust and upper mantle.

Rose Diagram of Strike Angle

Rose diagrams of the GB regional strike angle along the Corridor 1A are shown in Fig. 4.13. At short periods (0.003-0.16 s) the strike angle of most sites trends N30°-35°E. The same angle N30°-35°E is the most common strike angle for periods between 0.2-14.2 s. This direction is sub-parallel to the orientation of the GSLsz. The strike angle increases to N55°-65°E at periods 20-1700 s. There are two main strike directions at periods >2000 s, they are
Figure 4.13 Rose diagram of the Groom-Bailey regional strike angle for four period ranges. A total of 15 sites located in Corridor 1A are included. The angle bin is 10 degree.
N85°-90°E observed in the Great Bear Magmatic Arc and N55°-60°E observed in the Buffalo Head terrane. This difference indicates a lateral variation of the structure across the GSLsz.

4.5 Distortion

Groom-Bailey tensor decomposition (Groom and Bailey 1989) was used to examine distortion from near-surface inhomogeneities. Figures 4.14 and 4.15 show contour plots of the shear and twist determined using unconstrained GB decomposition along Corridor 1 and 1A. The results are somewhat erratic because of the presence of errors in the data and the fact that no smoothing has been applied.

In the Nahanni, Fort Simpson, Hottah, western Great Bear Magmatic Arc and Buffalo Head terranes (sites 145-117 along Corridor 1, 146-150 along Corridor 1A) where sites are located on Phanerozoic sedimentary rocks, the shear and twist angles are very small (< 5°) at periods less than 10 s. This observation indicates an absence of strong near-surface inhomogeneities, a result which is to be expected for sites lying on the shallow-dipping Phanerozoic sedimentary rocks. Larger shear and twist angles (~20°) at periods longer than 10 s indicate stronger distortion associated with deeper structures. However, distortion is weaker than in the eastern Great Bear Magmatic Arc and Archean Slave Province. The skew value also is lower at sites covered by the Phanerozoic sedimentary rocks than at sites in the eastern Great Bear Magmatic Arc and Slave Province (Fig. 4.11).

Between site 131 and site 124, larger shear (20°) and twist (10°) angles around period 1000 s show considerable spatial uniformity (Fig. 4.14). The Swift’s skew value is relatively high (Fig. 4.11). These results suggest the distortion may be due to larger scale 3D structures
Shear Angle

Twist Angle

Figure 4.14 Contour map of shear and twist from GB decomposition of sites along Corridor 1. Note the number “1” in front of site number is taken away for some sites in the west of Corridor 1 because of the small space, for example 145 is changed to 45. This is same for the following profiles.
Figure 4.15 Contour map of shear and twist from GB decomposition of sites along Corridor 1A.
rather than local galvanic distortion.

In the eastern Great Bear Magmatic Arc and the Slave Province, the shear and twist angles are larger at all periods (typically around 25° in the Slave Province) than that at sites in the Nahanni terrane and western Great Bear Magmatic Arc. However, the distortion is smaller than noted in many Precambrian shield settings (Jones and Ferguson 2001). For example, for most sites in the Flin Flon Belt, Trans-Hudson Orogen, the shear angles are >30° at period >5 s (Grant 1997; Stevens 1998; Ferguson et al. 2000). The stronger distortion in Trans-Hudson Orogen is interpreted to be due to more inhomogeneous near-surface structure in the exposed Precambrian rocks.

In the Slave Province, analysis of electric-field distortions shows the response is only weakly dependent on geoelectric strike (Jones et al. 2001). For example, MT phases are almost independent of the rotational coordinate. This result requires the conductivity structure of the crust and mantle to vary slowly with lateral distance in both crust and mantle. Therefore, near-surface inhomogeneities are the primary distortion source. This interpretation is supported by Fig 4.14 which shows the distortion varies from site to site and is independent of period.

From the examination of exploration depths (Fig. 4.8 and 4.9), the period required for EM signals to penetrate about 1 km of conductive sedimentary rock is less than 10 s. Therefore, at sites in the Phanerozoic sedimentary rocks the very low shear and twist angle at periods less than 10 s indicate an absence of inhomogeneities in the Phanerozoic sedimentary rocks. On the basis of this results it is unlikely that inhomogeneities in the Phanerozoic sedimentary rocks contribute to the distortion of long period data. The distortion
observed at long periods may be from the inhomogeneities near the top of the Precambrian rocks, or from large-scale variations in the thickness of the Phanerozoic cover.

Along Corridor 1A, shear and twist angles vary spatially across the GSLsz at periods longer than 30 s (Fig. 4.15). Sites to the east of the GSLsz (153 to 157) have higher shear angles (10°-30°) and twist angles (<10°) than sites to the west of the GSLsz (10°-25° shear and <5° twist). A clear, sub-vertical "boundary" is shown beneath sites 152 and 151, where the shear and twist are close to zero. The observations indicate lateral variation of the resistivity structure, with larger distortion to the east of the GSLsz.

The MT responses used in subsequent modelling are based on the GB analysis using average shear and twist for the period range 0.01 s - 1000 s at each site, and a constant strike azimuth for the whole profile. This period range corresponds to the signal penetration from the near-surface into the mantle. The responses at period <0.01 s were not used to calculate the average shear and twist in order to disregard the effect from the near-surface.

Figure 4.16 compares the raw MT responses with GB responses at two sites. For site 136, the raw responses differ from the GB responses at periods less than 0.1 s, but are similar at periods longer than 0.1 s (Fig. 4.16a). For site 129 (Fig. 4.16b), GB responses are very close to the raw responses at periods less than 20 s, indicating low levels of distortion. However, the GB responses are different from the raw responses at periods longer than 20 s. The GB TE apparent resistivity is smaller than the non GB apparent resistivity, whereas the GB TM apparent resistivity is higher than the non GB apparent resistivity. Because of the relatively small shear angle (-1.7°) and twist angle (8.2°), distortion cannot be responsible for the large difference between the raw and GB responses (Fig. 4.16b). The main reason for this
Figure 4.16 Comparison of raw responses (no rotation) and GB responses at (a) site 136, twist and shear are -5.6° and -0.9° respectively, and (b) site 129, twist and shear are -1.7° and 8.2° respectively. In GB decomposition, the specify strike angle is 34°.
difference is the GB strike angle of 34°. This result indicates the TE and TM responses at periods >20 s are strongly dependent on strike direction.

**Static Shift**

The GB tensor decomposition cannot correct the static shift of the MT data, therefore the static shift has to be examined using other methods. In the GB method the static shift is corrected by calculating the geometric mean of off-diagonal impedances at high frequencies to fit the data (Groom *et al.*, 1993). In this thesis, the Sternberg and Jones methods are used to complete the static shift correction for the MT data.

For the Sternberg method (described in Chapter 3), the apparent resistivity at period 100 s is used to determine the static shift factor, because the apparent resistivity is relatively stable and varies smoothly (100-200 Ωm). The regionally-averaged apparent resistivity is obtained by weighting surrounding sites using a chosen smoothing window (for example, by averaging the apparent resistivity of the adjacent three or five sites). The profile was separated into three segments to calculate the regionally-average apparent resistivity: Nahanni terrane, Fort Simpson terrane and Hottah terrane. Figure 4.17 shows the raw and regionally-average TE and TM apparent resistivity at 100 s period. The static shift factor is the ratio between them as defined in Eq. (3.57).

As described in Chapter 3, the “Jones method” of static shift correction involves the two-layer 1D inversion and comparison with borehole logs. In this study we use only the first part of the procedure. The first step is to calculate the resistivity and depth of two layers along the profile by 1D inversion. The second step is to find a simple function to fit the
Figure 4.17. Examination of static shift using the Sternberg method.
resistivity of second layer. Then the correction factor is obtained using the ratio of the fitted function and the 1D inversion resistivity at each site. The detail comparison with borehole logs will be discussed in Chapter 6. Figure 4.18 and Table 4.5 show the static shift factors at sites 145 - 122 in Corridor 1 from Sternberg and Jones methods.

Table 4.5 The static shift factors for $\rho_\nu$ and $\rho_\sigma$ at some stations from the Sternberg and Jones methods.

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<tr>
<th>Station</th>
<th>Sternberg method</th>
<th>Jones method</th>
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</thead>
<tbody>
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<td>Factor of $\rho_\sigma$</td>
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The ratio of Sternberg factors and Jones factors is close to unity except sites 143-141,
Figure 4.18 (a) Comparison of static shift factors between the Sternberg and Jones methods for site 145 to site 122, (b) ratio of Sternberg factors and Jones factors.
indicating the static shift factors from both methods are similar at most of sites (Fig. 4.18b). Correction factors close to unity (0.75-1.37) indicate low static shift in Fort Simpson and Hottah terranes and an absence of near-surface small-scale heterogeneities. The same methods of static shift correction were applied in other parts of the study areas. The static shift is also small (0.75-1.3) along Corridor 1A. In the Great Bear magmatic arc and Slave Province, the static shift range is from 0.5 to 2.0, which is larger than at sites on the Phanerozoic sedimentary rocks.

Note the static shift factors based on the short period responses can be used to correct both the V5 and LiMS data. V5 and LiMS data will have the same static shift environment, because they used the same electrode holes. The static shift correction was not made prior to the final modelling and inversion because of the small values, but its affect is examined in Chapter 6.

4.6 Apparent Resistivity and Phase Response

Figure 4.19 and 4.20 show the GB TE and TM mode apparent resistivity and phase pseudosections along Corridor 1 and 1A. In order to display data using a pseudosection the sites are projected onto a corresponding straight line. The relative orientation of the different segments of the corridors and the geoelectric strike (N34°E) made it necessary to project the sites onto a series of different lines. The western segment of the profile is perpendicular to the geoelectric strike (N34°E). The eastern and middle segments of the profile are approximately parallel to the survey line, because the sites will be very close, or even switch their order, after they are projected to the profile perpendicular to the geoelectric strike (Fig.
Figure 4.19. Groom-Bailey apparent resistivity and phase pseudosections for TE and TM modes along Corridor 1 (no static shift correction applied). The results are divided into three profiles. The western one is perpendicular to regional geoelectrical strike. The other two profiles are approximately parallel to the survey line. Na: Nahanni, FS: Fort Simpson, Ho: Hottah, GB: Great Bear Magmatic Arc, An: Anton, Co: Contwoyto.
4.19. The profile along Corridor 1A is perpendicular to the strike N41°E (Fig. 4.20).

4.6.1 Pseudosections for Corridor 1

In this section the pseudosection responses for the different period ranges will be examined.

*Short period (<1 s)*

In the Slave Province, the TE and TM apparent resistivities are resistive (>1000 Ωm) and erratic at period <1 s, and the TE and TM phases are relatively stable. The response is characterized by strong static shifts. In the Nahanni terrane, Fort Simpson terrane, Hottah terrane and the western part of the Great Bear Magmatic Arc, low TE and TM apparent resistivities (<50 Ωm) extend from the shortest periods to periods of about 1 s. The TE and TM phases are also relatively high at this period range. This conductive response is observed at progressively shorter periods with distance east across the Great Bear Magmatic Arc, and can be attributed to the Phanerozoic sedimentary rocks thinning eastwards.

At sites in the Nahanni to Hottah terranes the TE and TM apparent resistivities are relatively high (~60 Ωm) compared with at the Great Bear Magmatic Arc (<50 Ωm) at periods less than 0.01 s, and low (<20 Ωm) between 0.01-1 s. The TE and TM phases are higher than 45° at periods less than 0.1 s. The responses show a spatial variation at short periods 0.1-1 s near site 142 (the extreme western end of Corridor 1). Both TE and TM phases are relatively low and both TE and TM resistivities are relatively high compared with surrounding areas. There is an enhanced conductivity response (<10 Ωm) near sites 135 and
136 at periods less than 0.1 s, where the phases are relatively high.

**Intermediate Periods (1-10³ s)**

This period range corresponds to penetration from upper crust to upper mantle (Table 4.3 and Figure 4.8). The TE and TM apparent resistivity values in this period range increase from about 200 Ωm in the west, to about 2000 Ωm beneath the Great Bear Magmatic Arc, to >5000 Ωm in the Slave Province (Fig. 4.19). The TE and TM phases show the relatively low values which are observed at progressively shorter periods with distance east across the Great Bear Magmatic Arc. A discontinuity of TE responses is observed at sites 131 and 130, in which the TE resistivity is relatively low (about 100 Ωm) and TE phase is relatively high comparing with adjacent sites. This anomaly extends from about 1 s to 10⁴ s. However, there is no clear variation of apparent resistivity and phase for the TM mode.

**Long periods (>10⁴ s)**

This long period range corresponds to signal penetration to deep into the lithosphere and into the asthenosphere. At the longest periods a TE and TM apparent resistivity decrease can be observed beneath the Fort Simpson and Hottah terranes, but cannot be observed beneath the east of the Great Bear Magmatic Arc and the Slave Province. A TE and TM phase increase is observed in the Hottah terrane to the Slave Province. The penetration depth at 1000 s period is about 130 km in the Nahanni to Hottah terranes, which is shallower than 450 km in the Slave Province (Table 4.3). Therefore, the decrease of resistivity and increase of phase may be related to the asthenosphere becoming more shallow in the western part of
4.6.2 Pseudosections for Corridor 1A

Short Periods (<1 s)

This period range corresponds to signal penetration to several hundred metres (Table 4.4). Both TE and TM apparent resistivities show an abrupt change at site 153: the resistivity is >60 Ωm to the east and <20 Ωm to the west (Fig. 4.20). The TE and TM phases are relatively low at sites 152 and 151 compared with at adjacent sites. This anomalous phase zone corresponds to a boundary in the apparent resistivity. The apparent resistivity and phase to the west of site 152 are similar to those in the Hottah terrane.

Intermediate Periods (1-10¹ s)

This period range corresponds to signal penetration to the upper crust and mantle. The TE and TM apparent resistivities are relatively high in this period range compared with those for periods <1 s. The resistivity in the Buffalo Head terrane (>400 Ωm) is higher than that in the Great Bear Magmatic Arc (<400 Ωm). A transition in apparent resistivity occurs around sites 150 and 152. This transition shifts west with increasing periods from 0.5 s to 5 s, then changes to a vertical transition at periods >5 s (Fig. 4.20). The TE and TM phases all show a deflection to shorter periods (10-100 s) beneath sites 152 and 151. These responses suggest the presence of a structural anomaly beneath the Phanerzoic cover around sites 152-153.
**Long Period (>10^3 s)**

At this period range, there is a decrease of TE and TM apparent resistivity in the Hottah terrane to the Great Bear Magmatic Arc. This feature is not visible in the Buffalo Head terrane. An increase of TE and TM phases can be observed beneath the Hottah terrane and Great Bear Magmatic Arc at very long periods.

**4.7 Conductivity Structure of Surface Sedimentary Rocks**

MT apparent resistivity and phase responses indicate the Phanerozoic sedimentary rocks form a relatively thick and conductive surface layer (Fig. 4.19 and 4.20). The responses suggest the thickness of this layer decreases in an eastward direction from the Fort Simpson terrane to the Great Bear Magmatic Arc.

At short periods (<1 s), the similarities of the TE and TM apparent resistivity responses, small induction vectors, and small maximum phase difference (Fig. 4.10), imply an approximately 1D near-surface structure. Therefore, 1D modelling and inversion were used to determine the geoelectric structure of the surface sedimentary rocks. In this part of the study the data were truncated at 10 s period corresponding approximately to a signal penetration of 3 km (Table 4.3), which is through the sedimentary rocks.

**4.7.1 Geoelectric Model of Surface Sedimentary Rocks**

Inversion of the TE, TM and determinant average responses was completed for each site using Occam (Constable *et al.* 1987), Marquardt (Marquardt 1963) and Fischer (Fischer *et al.* 1981) methods, as implemented in GEOTOOLS. For the Occam inversion, the starting
model is a half space, the maximum number of iterations is 15, and the stopping tolerance (Eq. 3.75) is set as 0.1. The Fischer model was used as the initial model in the Marquardt least squares inversion.

Models from the three inversion methods are similar and reveal a three-layered shallow structure: a thin relatively resistive first layer, a conductive second layer, and a very resistive basement. The responses of three inversion models fit the observed data very well at short periods (<10 s). Figure 4.21 shows the 1D Occam and Fischer inversion models and data responses for the TE, TM and determinant modes at a typical site, site 127.

The static shift factor is close to 1 at this site (Table 4.5), and does not affect the inversion models. The model responses fit the observed data very well at periods less than 10 s. The inversion models have the same basic elements, containing a relatively resistive layer with 50 Ωm resistivity and 50 m thickness, a conductive middle layer with 9 Ωm resistivity and 600 m thickness, and a resistive basement with 600 Ωm resistivity. The depth to the top of the resistive basement in the Fisher model is about 650 m. The poorer data fit at long periods implies a more complex 2D or 3D deep geoelectric structure.

The Occam algorithm provides the smoothest model fitting the data at a specified tolerance (Chapter 3). Therefore, the smooth Occam model does not resolve the basement depth clearly (Fig. 4.21), although the model responses fit observed data very well. The Fischer algorithm which allows abrupt resistivity changes, such as expected on the basis of the geological structure, can provide a better determination of the depth of interface between the very conductive and resistive layers.

TE and TM responses and models at short periods are similar to the determinant
Responses from 1D Occam model

Responses from 1D Fischer model

Figure 4.21. Model responses from 1D Occam and Fischer inversions at site 127. (a) TE and TM models and responses, (b) determinant models and responses. The depth is plotted on a horizontal axis to permit comparison between the resistivity model and the apparent resistivity response.
Responses from 1D Occam model

Responses from 1D Fischer model

(b) determinant models and responses
average response and model (Fig. 4.21). Therefore, the determinant average response was used in subsequent modelling to delineate the shallow structure.

Inversions were done at each site. The 1D resistivity models were aligned together along the profile with the model at each site is represented by a column in which the resistivity value which varies with depth. This presentation is called stitched 1D model in MT processing. Figure 4.22 shows the stitched 1D determinant model of Fischer inversion for the profiles from the Nahanni terrane to the Great Bear Magmatic Arc along Corridor 1 and from the Great Bear Magmatic Arc to the Buffalo Head along Corridor 1A. The inversion result reveals significant lateral and vertical resistivity variations within the sedimentary rocks.

Along Corridor 1 (Fig. 4.22a), there is a moderately resistive upper layer (>50 Ωm) which is about 100 m thickness in the eastern Fort Simpson and western Hottah terranes. The layer becomes thinner to the west in the western Fort Simpson and Nahanni terranes (20 m). It also becomes thinner to the east with its margin in the Great Bear Magmatic Arc. More conductive rocks (~10 Ωm) underlie this surface layer. The depth to the base of this conductive layer decreases from 1050 m at site 139 in the Fort Simpson terrane to 200 m at site 117 in the Great Bear Magmatic Arc. Results for site 142 are very different from surrounding sites. The sedimentary rocks at this site are much thinner than at adjacent sites with a total thickness of about 120 m.

An equivalence analysis was done in order to examine the resolution of the structures in the inversion models. The program was provided by A. Jones of GSC (Jones 1982). The equivalence was examined using a 1D model and an inversion code based on singular value
Figure 4.22 Stitched one-dimensional models from Fischer inversions of the determinant response along Corridor 1 and 1A.
decomposition. The program provides a resolution matrix showing the covariance between the parameters and also provides estimates of the parameter uncertainty using the 95% confidence limit. If the resolution matrix is the identity matrix (1.0 along the diagonal and 0.0 elsewhere), then each model parameter is uniquely determined. Otherwise, the resolved parameters are logarithmic combinations of the model parameters (Menke 1984).

The equivalence was examined using a four-layer inversion model at a typical site, 124, where the very conductive layer consists of two parts (Table 4.6b). The data error used for the resolution analysis was 1% for the apparent resistivity and 0.5° for the phase. The periods used in inversion were from 0.0001 s to 7.1 s.

The resolution matrix element corresponding to the thickness of each layer has relatively high value (0.959 for \( t_1 \), 0.92 for \( t_2 \), and 0.974 for \( t_3 \)). The 95% confidence limits, which provide a relatively conservative error estimate, are quite large, e.g. the bound of third layer thickness is from 360 m to 770 m with a best-fit value 528 m (Table 4.6b).

The results indicate the MT method provides reasonable resolution for the conductive structure above the resistive basement. The exact resolution will be examined in Chapter 6, where the MT results will be compared with the borehole log data.

Along Corridor 1A (Fig. 4.22b), the conductive surface layer is 450-700 m thick. The resistivity of the sedimentary rocks shows differences across the profile. A resistivity boundary at site 153 can be seen clearly. To the west, the sedimentary rocks overlying the Great Bear Magmatic Arc have a similar conductivity structure as the Fort Simpson and Hottah terranes: a thin, moderately resistive, upper layer (≈50 Ωm) with a depth to the base of 20-40 m and a conductive second layer (≈10 Ωm) extending to 450-700 m depth. To the
east, the sedimentary rocks overlying the Buffalo Head terrane consist of two layers, both of which are more resistive than to the west. The structure includes a relatively resistive upper layer (~500 $\Omega$m) with a thickness of 350 m, and a conductive second layer (~50 $\Omega$m) about 300 m thick. The shallow resistivity results indicate a change in the Phanerozoic sedimentary rocks.

**Table 4.6 (a) Resolution matrix, (b) estimates of the parameter uncertainty for a three-layer model at site 124.**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>1$^{st}$ layer resistivity ($\rho_1$)</td>
<td>0.999</td>
</tr>
<tr>
<td>2$^{nd}$ layer resistivity ($\rho_2$)</td>
<td>0.001</td>
</tr>
<tr>
<td>3$^{rd}$ layer resistivity ($\rho_3$)</td>
<td>0</td>
</tr>
<tr>
<td>4$^{th}$ layer resistivity ($\rho_4$)</td>
<td>0</td>
</tr>
<tr>
<td>1$^{st}$ layer thickness ($t_1$)</td>
<td>0.063</td>
</tr>
<tr>
<td>2$^{nd}$ layer thickness ($t_2$)</td>
<td>0.07</td>
</tr>
<tr>
<td>3$^{rd}$ layer thickness ($t_3$)</td>
<td>0.001</td>
</tr>
</tbody>
</table>

(a)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Minimum Value</th>
<th>Best-fit Value</th>
<th>Maximum Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho_1$</td>
<td>52.7</td>
<td>55.7</td>
<td>58.7</td>
</tr>
<tr>
<td>$\rho_2$</td>
<td>8.4</td>
<td>17.7</td>
<td>37.1</td>
</tr>
<tr>
<td>$\rho_3$</td>
<td>7</td>
<td>8.6</td>
<td>10.6</td>
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<tr>
<td>$\rho_4$</td>
<td>304.6</td>
<td>437.9</td>
<td>629.5</td>
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<td>$t_1$</td>
<td>60.8</td>
<td>96.8</td>
<td>153.9</td>
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<tr>
<td>$t_2$</td>
<td>64</td>
<td>122.9</td>
<td>236.1</td>
</tr>
<tr>
<td>$t_3$</td>
<td>363.9</td>
<td>528.3</td>
<td>767</td>
</tr>
</tbody>
</table>

(b)
4.7.2 Interpretation of Shallow Conductivity Structure

The 1D MT inversion results were compared with information from nearby borehole logs (Table 4.7). The total thickness of the Phanerozoic sedimentary rocks, comprising Devonian dolomite, limestone, sandstone, siltstone and shale, thins to the northeast.

**Corridor 1**

The resistivity structure of the Phanerozoic sedimentary rocks overlying the Nahanni, Fort Simpson, Hottah, Great Bear Magmatic Arc and Buffalo Head terranes may be interpreted as:

- A relatively resistive 20-100 m thick surface (50-100 Ωm) corresponding to the upper Devonian bioclastic limestone and sandy limestone (Redknife Formation). A detailed comparison with borehole log and interpretation will be discussed in Chapter 6.

- The conductive layer with about 10 Ωm resistivity has about 750-850 m thickness in the Fort Simpson terrane and decreases towards the east. Comparing MT results with the borehole logs (National Energy Board 1980), this layer can be associated with the upper Devonian shale and silty sandstone deposits of Fort Simpson formation and 1st Black Shale unit (Hills et al. 1981). These sedimentary rocks are found in the Phanerozoic rocks overlying the Fort Simpson terrane, Hottah terrane and the eastern part of the Great Bear Magmatic Arc. The thickness of this layer is agreement with the borehole data. More details on the comparison of MT results and boreholes will be provided in Chapter 6.

- The conductive surface layer at site 142 (with 200 m thickness) is thinner than in surrounding areas. This feature corresponds to the Liard High, which was formed during
the early Paleozoic (Cambrian and early Ordovician) (Law 1971; Merjer-Drees 1975).

Sedimentary rocks deposited after the middle Ordovician overlie the Liard High.

<table>
<thead>
<tr>
<th>Borehole</th>
<th>Latitude</th>
<th>Longitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cominco Pine Point Test G-1</td>
<td>60°51'05&quot;</td>
<td>114°24'30&quot;</td>
</tr>
<tr>
<td>Cominco Sulphur Pt. Test G-4</td>
<td>60°54'00&quot;</td>
<td>114°46'00&quot;</td>
</tr>
<tr>
<td>Horn River et al Hay River B-52</td>
<td>60°51'04&quot;</td>
<td>115°40'17&quot;</td>
</tr>
<tr>
<td>Alex Falls B-07</td>
<td>60°36'01&quot;</td>
<td>115°45'39&quot;</td>
</tr>
<tr>
<td>Shell Alexandra No. 5</td>
<td>60°34'15&quot;</td>
<td>115°44'39&quot;</td>
</tr>
<tr>
<td>N.W.T. Heart Lake No. 1</td>
<td>60°48'00&quot;</td>
<td>116°48'00&quot;</td>
</tr>
<tr>
<td>Mills Lake B-41</td>
<td>61°40'06&quot;</td>
<td>116°53'17&quot;</td>
</tr>
<tr>
<td>Placid Chevron N.E.Tathlina D-50</td>
<td>60°49'07&quot;</td>
<td>117°09'08&quot;</td>
</tr>
<tr>
<td>Mills Lake C-12</td>
<td>61°01'02&quot;</td>
<td>117°17'55&quot;</td>
</tr>
<tr>
<td>Pan Am-Shell Kakisa H-36</td>
<td>60°55'18&quot;</td>
<td>117°21'05&quot;</td>
</tr>
<tr>
<td>Providence A-47</td>
<td>61°26'14&quot;</td>
<td>117°22'33&quot;</td>
</tr>
<tr>
<td>Mills Lake B-75</td>
<td>61°14'05&quot;</td>
<td>117°28'42&quot;</td>
</tr>
<tr>
<td>Mills Lake L-10</td>
<td>61°09'41&quot;</td>
<td>117°31'39&quot;</td>
</tr>
<tr>
<td>Mills Lake I-57</td>
<td>61°16'37&quot;</td>
<td>117°39'36&quot;</td>
</tr>
<tr>
<td>Mills Lake A-70</td>
<td>61°09'07&quot;</td>
<td>117°56'29&quot;</td>
</tr>
<tr>
<td>Mills Lake C-03</td>
<td>61°12'12&quot;</td>
<td>118°01'00&quot;</td>
</tr>
<tr>
<td>Mills Lake C-60</td>
<td>61°19'13&quot;</td>
<td>118°10'45&quot;</td>
</tr>
<tr>
<td>Mills Lake J-74</td>
<td>61°13'38&quot;</td>
<td>118°13'48&quot;</td>
</tr>
<tr>
<td>Kakisa –73</td>
<td>61°02'58&quot;</td>
<td>118°14'25&quot;</td>
</tr>
<tr>
<td>Foetus Lake C-49</td>
<td>61°08'12&quot;</td>
<td>118°38'28&quot;</td>
</tr>
<tr>
<td>Foetus Lake P-56</td>
<td>61°05'49&quot;</td>
<td>118°39'37&quot;</td>
</tr>
<tr>
<td>N.W.T. No. 2 (K-59)</td>
<td>61°18'32&quot;</td>
<td>118°40'30&quot;</td>
</tr>
<tr>
<td>Redknife I-24</td>
<td>61°03'40&quot;</td>
<td>119°18'49&quot;</td>
</tr>
<tr>
<td>Redknife J-21</td>
<td>61°00'41&quot;</td>
<td>119°19'14&quot;</td>
</tr>
<tr>
<td>Redknife H-28</td>
<td>61°27'16&quot;</td>
<td>119°31'51&quot;</td>
</tr>
<tr>
<td>Trout D-66</td>
<td>61°35'04&quot;</td>
<td>119°58'02&quot;</td>
</tr>
<tr>
<td>Briggs Trout River No.4</td>
<td>60°59'11&quot;</td>
<td>120°15'02&quot;</td>
</tr>
<tr>
<td>Jean Marie E-07</td>
<td>61°16'20&quot;</td>
<td>120°46'45&quot;</td>
</tr>
<tr>
<td>Blackstone F-72</td>
<td>61°11'27&quot;</td>
<td>122°54'54&quot;</td>
</tr>
<tr>
<td>Liard River No. 2 (D-75)</td>
<td>61°14'06&quot;</td>
<td>122°44'59&quot;</td>
</tr>
</tbody>
</table>
Corridor 1A

Along most of the length of the profile the total thickness of the conductive surface layer in the resistivity models for Corridor 1A (Fig. 4.22b) is in quite good agreement (±100 m) with the thickness of Phanerozoic sedimentary rocks determined from the borehole logs (Fig. 4.23). It is difficult to make an accurate comparison east of site 156 because there are fewer well logs near the profile. The moderately resistive (50-100 Ωm), 20 to 40 m thick, surface-layer present at sites northwest of 153 is interpreted to correspond to overburden. The thickness of this layer is in good agreement with the observed thickness of the surficial sediments in the area (e.g. Rhodes et al. 1984). The layer is not discerned to the east, probably because the resistivity of the overburden is more similar to that of the underlying rocks than is the case to the northwest.

To the east of site 153, the relatively resistive surface layer (~500 Ωm resistivity, ~350 m thickness) is interpreted to correspond to the middle Devonian carbonate and evaporate dominated units including the Nyarling, Little Buffalo, and Chinchaga Formations. The underlying relatively conductive layer (~50 Ωm resistivity, ~300 m thickness) corresponds to the Middle Ordovician (?)-Middle Devonian Mirage Point Formation which includes red and green shale beds, carbonate, sandy carbonate and gypsum (Glass, 1990; Rhodes et al. 1984; Douglas and Norris 1973) and the underlying Ordovician Old Fort Island Formation which consists of quartz sandstone, siltstone and shale. The thickness of the lower unit in the model is significantly greater than the thickness of the Mirage Point Formation in the cross-section but this difference may be due in part to the divergence between the eastern end of the Corridor 1a and the location of the cross-section. Rhodes et al. (1984) note
that the regional variation in the thickness of the Mirage Point Formation is up to 180 m.

The conductive (<10 Ωm) zone subcropping between sites 153 and 148 corresponds to the thick dark grey and black shale sequences of the Upper Devonian Horn River Formation (Fig. 4.23, Douglas and Norris 1973). The slightly more resistive (10-25 Ωm) zone subcropping northwest of site 147 correspond to units including the Upper Devonian Twin Falls, Tathlina, and Redknife Formations which include bioclastic and reef carbonates as well as sandy limestone, shale and siltstone. The geological cross-section shows the Mirage Point Formation and the overlying resistive evaporative dolostone formations extending to the northwest beneath the conductive Horn River Formation.

It is not possible to resolve these units in the MT models because the MT response is relatively insensitive to resistive layers located beneath conductive layers. The presence of these units may have resulted in a small overestimate in the thickness of the overlying Horn River Formation in the MT models. The MT models for sites 150 to 153 indicate the presence of the Presqu’ile Barrier, along strike from the Pine Point deposit. Between sites 150 and 152 there is a rapid decrease in the thickness of the conductive layer corresponding to the basinal shale units. This change is sharper than shown on the geological cross-section and is interpreted to represent a lateral transition from the Upper Devonian Horn Point Formation to younger, topographically high, dolomitic units of the Presqu’ile Barrier. The eastern margin of the barrier is not discerned on the MT models because it involves a contact between dolomitic units in the barrier and carbonates of the Nyarling and Little Buffalo Formations. These units will have similar resistivity.
Figure 4.23 Geological map of south seashore of the Great Slave Lake and structure section along A-D (from Douglas 1973; Douglas and Norris 1973). Qal: Deltaic and alluvial sands and silts of Slave River, uDHR: Greenish grey shale, limestone, sandstone, mDSP: Aphanitic to fine-grained, laminated, thinly bedded, brown limestone, mDSUP: Aphanitic to fine-grained, brown limestone, green shale and dolomitic limestone, mDr: Massive, coarse-grained, vuggy dolomite, mDPP: Fine-grained limestone; shale; finely-grained, aphanitic, granular and highly porous brown dolomite, mDr: Grey to green calcareous shale with iron sulphide concretions, mDNY: White gypsum; minor aphanitic brown limestone, mDLIB: Aphanitic argillaceous, brown limestone and dolomite, mDC: White laminated gypsum; salt; fine-grained brown limestone and dolomite, DMP: Red and purplish red dolomite, sandy dolomite; red and green shale; gypsum.
4.8 Conductivity Structure of Precambrian Rocks

4.8.1 Effect of Phanerozoic Sedimentary Rocks on Long Period MT Responses

In this section I will examine whether the large-scale spatial variation in the conductive sedimentary rocks can distort the long period MT responses. In SNORCLE area, the Phanerozoic sedimentary rocks are conductive (10 Ωm) and thick vary from zero to 1 km in thickness.

Two-dimensional forward modelling was used to calculate the MT responses of some simple models and investigate the sensitivity of responses to various near-surface structure. The modelling used is Dr. Philip Wannamaker's 2D finite element algorithm (Wannamaker et al. 1987) included with GEOTOOLS (GEOTOOLS 1997). This method uses a finite element mesh based on rectangular cells of uniform resistivity with the margins of the cells defined by columns and rows of arbitrary widths. Each rectangular cell is further subdivided into four triangle elements.

Table 4.8 lists three models which are used to examine the sensitivity of responses for the various surface structure.

<table>
<thead>
<tr>
<th>Model</th>
<th>Shallow</th>
<th>Deep</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model A</td>
<td>Two uniform layers. 1st: 50 Ωm, 100 m; 2nd: 9 Ωm, 900 m</td>
<td>400 Ωm half space</td>
</tr>
<tr>
<td>Model B</td>
<td>Two wedge shape layers. The resistivities are same as for Model A.</td>
<td>400 Ωm half space</td>
</tr>
<tr>
<td>Model C</td>
<td>Conductive layers from 1D inversion</td>
<td>400 Ωm half space</td>
</tr>
</tbody>
</table>
Model A consists of two conductive horizontal layers overlying a resistive half space (Fig. 4.24a). Model B consists of two west-dipping conductive layers overlying the resistive basement (Fig. 4.24b). For Model C, the structure above 1000 m depth is defined by the 1D inversion results (Fig. 4.22a) and overlies the resistive basement.

The length of the profile between the first site and the last site is about 250 km, but the full width of the forward model is about 5000 km. The broad margins outside the actual profile are designed to satisfy the boundary conditions of modelling algorithm (Wannamaker et al. 1987). At the edges of the model, the width of finite element mesh columns increases logarithmically away from the sites, and the thickness of the rows typically increases geometrically with depth. The number of horizontal nodes in the model is 122. The number of vertical nodes is 77.

The first test completed was a comparison of responses from Model A and B. Subsequently the responses from Model A and C are compared. For Model A, the TE mode apparent resistivity and phase are exactly same as the TM mode response for each site (Fig. 4.24c) as required for a 1D structure. For Model B, the first relatively resistive layer thins from a 100 m thickness at site 145 to zero at site 126. The eastward-thinning second layer causes more resistive responses in the eastern part of the profile than in the western part at long periods. However, the TE and TM responses are very similar. The wedge shape produces differences between the TE and TM responses in the short period responses only at site 126 located near the margin of the upper layer. The TE apparent resistivity at this site shows a small difference from the TM apparent resistivity at periods less than 0.01 s. Although there is a small separation of apparent resistivity at several sites at long periods,
Figure 4.24 Comparison of the apparent resistivity and phase responses for two models. (a) Model A: uniform layered half-space, (b) Model B: west-dipping sedimentary rocks overlying the uniform half space, (c) apparent resistivity and phase from Model A (circle symbols) and Model B (solid line: TM response, dash line: TE response).
the separation is still small with a maximum of 17% difference (linear scale). The TE and TM phases are very close. Therefore, the results suggest the thickness change of conductive Phanerozoic sedimentary rocks will not produce significant 2D distortion of the long period responses observed in SNORCLE area.

Figure 4.25 compares the responses from Model C with Model A. At periods <1 s, the corresponding responses for Model C are different from Model A because Model C has more complex surface structure than Model A. At these periods the apparent resistivities of TE and TM modes for Model C are very close at most sites, except for sites 134 and 133. The slight split of TE and TM phases at periods <0.01 s indicates the inhomogeneities at near surface (<100 m, Fig. 4.8). At periods >1 s, the TE responses from Model C approaches the responses from Model A. Separation of the Model C TE and TM apparent resistivity occurs at sites 143 and 142 where a local 2D structure is present, the thinning of sedimentary rocks over the Precambrian basement (Liard high). At these sites the Model C TE and TM phase responses remain close and similar to the response from Model A.

The gradual thinning to the east of the conductive sedimentary rocks does not cause major distortion of long period responses in the western part of Corridor 1. Phanerozoic structures produce some static shift but do not affect the long period phase responses. The separation of apparent resistivity and phase at some sites is caused by the local 2D structures. Modelling and inversion studied by Cox et al. (1986) showed that a surface conductive layer of sediment and fractured basalt at the surface did not affect the determination of deep structure.
Figure 4.25 Comparison of the apparent resistivity and phase for Model A (circle symbols) and Model C (lines). Model C consists of conductive Phanerozoic sedimentary rocks from 1D inversion model overlying the uniform half space with 400 ohm.m resistivity. Solid line: TM response, dash line: TE response.
<table>
<thead>
<tr>
<th>Site 133</th>
<th>Site 132</th>
<th>Site 131</th>
<th>Site 130</th>
</tr>
</thead>
<tbody>
<tr>
<td><img src="b" alt="Graph" /></td>
<td><img src="b" alt="Graph" /></td>
<td><img src="b" alt="Graph" /></td>
<td><img src="b" alt="Graph" /></td>
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<tr>
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<td>Site 128</td>
<td>Site 127</td>
<td>Site 126</td>
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<td><img src="b" alt="Graph" /></td>
<td><img src="b" alt="Graph" /></td>
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<tr>
<td>Site 125</td>
<td>Site 124</td>
<td>Site 123</td>
<td>Site 122</td>
</tr>
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<td><img src="b" alt="Graph" /></td>
<td><img src="b" alt="Graph" /></td>
<td><img src="b" alt="Graph" /></td>
</tr>
</tbody>
</table>
4.8.2 Modelling Strategy for the Crust and Upper Mantle

The TE and TM apparent resistivity and phase pseudosections (Fig. 4.19 and 4.20) showed the variation of MT responses with both spatial position and period, indicating the presence of significant 2D and/or 3D crustal and lithospheric resistivity structures. The 1D inversion models shown in Fig. 4.22 can produce a good data fit to only the short period data (<10 s) for most sites, also indicating a more complex deep geoelectric structure. As discussed in section 4.4, the maximum phase difference, induction vectors and GB regional strike all indicate an approximately 2D deep structure. Therefore 2D forward modelling and inversion were applied to more accurately resolve the crustal and lithospheric structure.

The 2D forward modelling method used is the 2D finite element method (Wannamaker et al. 1987) discussed in last section. The 2D inversion methods used in this thesis include 2D Occam (deGroot-Hedlin and Constable 1990) and 2D Non-Linear Conjugate Gradient (NLCG) (Mackie and Rodi 1996; Rodi and Mackie 2001) inversions.

In 2D MT inversions, there are typically three complex responses to fit: the TE and TM impedances and the tipper. As shown in Chapter 3 the TE mode has lower resolution for the lateral position of structure because its current flow is parallel to the structural strike. However it is more sensitive to the conductance of conductive regions than the TM mode. The TM response is more sensitive to lateral boundaries than the TE mode because the TM mode involves a current flow perpendicular to the structural strike. As shown in Fig. 3.5, the tipper which is a TE mode response is sensitive to 2D conductivity boundaries. Therefore, the inversion of TE mode can resolve the depth and the conductance of the conducting regions and the inversion of TM mode can resolve the horizontal variation of conductivity.
Inversion of the TM model can also resolve the presence of thin resistive layers, not resolved by the TE mode (Agarwal et al. 1993; Ferguson and Edwards 1994). The inversion of both modes (including the tipper) can provide a more accurate resistivity structure.

Because of the profile orientation of Corridor 1 and 1A, inversions were performed on several groups of sites:

- A 2D inversion profile was created for sites located in the Nahanni, Fort Simpson and Hattah terranes along Corridor 1. The sites were projected onto a N124°E line which is perpendicular to the average strike direction.

- The geoelectric strike direction is sub-parallel to the survey line in the Great Bear Magmatic Arc (Fig. 4.10), and sites in the arc will be very close together after being projected to a line perpendicular to the strike. Only 1D inversions will be done.

- A separate inversion profile was completed for sites along Corridor 1A. For modelling of the whole crustal structure the sites were projected onto a N131°E line which is perpendicular to the average strike angle of Corridor 1A.

- As discussed in section 4.4.2, the geoelectric strike direction varies with period around the GSLsz. The strike direction is approximately N30°E at ~10 s, whereas at 100 s the strike direction is approximate N60°E. Therefore, for modelling of the upper crustal structure and the lower crust and mantle structure, additional inversions were run using the responses rotated to N30°E (periods <10 s) and to N60°E (periods 30-3000 s) respectively.

The inversions were performed on a Sun Microsystems Enterprise 450 (4 X UltraSPARC-II 400MHz) at GSC. The system clock frequency is 100 MHz, the memory size is 2048 MB and each of the four CPU’s has 4 MB of cache. The run time is dependent on the
number of data (the number of responses inverted and the number of frequencies) and the model size (the number of rows and columns of the regularization grid and finite-element meshes). One iteration of an Occam inversion needs about 1-2 hours for 1500 inversion data and a typical model size (about 150 columns x 80 rows), so, it requires about one day to finish one Occam inversion (10-15 iterations). The NLCG method is much faster than the Occam method, and takes only 2-3 hours to finish one inversion (40-50 iterations).

Limitations of the GEOTOOLS (version 7.2) implementation of the 2D inversion algorithms mean the program does not allow the inclusion of the static information into the inversion of the GB corrected data. Consequently, the static shift correction will not be included in the following inversion and modelling.

4.8.3 Two-dimensional Occam Inversion Methodology

Design of Mesh and Grid

Regularization grids define rectangular cells with each grid cell containing an integral number of finite element rectangular cells. The design of the finite element mesh and regularization grid follows a series of rules described in GEOTOOLS manual (GEOTOOLS 1997).

The design of meshes (columns and rows) for 2D forward models was discussed in section 4.8.1. From experimentation it was determined that the ratio of the column width and mesh row width should be less than 6 in order to avoid the program crashing due to overflow. The finite element mesh consists of 121 columns and 71 rows for the model along Corridor 1, and 114 columns and 71 rows for the model along Corridor 1A.
The thickness of the regularization grid elements in the vertical direction (rows) typically increases geometrically with depth. This geometry is consistent with the MT resolution which decreases exponentially with depth. In order to accommodate the near-surface resistivity variation associated with the Phanerozoic rocks, the row thickness at shallow depth should be small. Table 4.9 lists the row depth used for modelling the SNORCLE data.

There are at least two finite element mesh rows in each grid row and there is one vertical regularization grid node between each two sites. Adjacent columns in the regularization grid can be combined at some depths in order to decrease the total number of inversion parameters. The number of grid cells (the total number of inversion parameters) is 807 for the models along Corridor 1, and 774 for the models along Corridor 1A.

Table 4.9 Grid row depth in Occam inversion model for the SNORCLE area.

<table>
<thead>
<tr>
<th>No. of Row</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
<th>13</th>
<th>14</th>
<th>15</th>
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</thead>
<tbody>
<tr>
<td>Depth (km)</td>
<td>0.1</td>
<td>0.3</td>
<td>0.6</td>
<td>0.9</td>
<td>1.2</td>
<td>1.8</td>
<td>2.5</td>
<td>3.5</td>
<td>5</td>
<td>7</td>
<td>10</td>
<td>14</td>
<td>18</td>
<td>23</td>
<td>29</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>No. of Row</th>
<th>16</th>
<th>17</th>
<th>18</th>
<th>19</th>
<th>20</th>
<th>21</th>
<th>22</th>
<th>23</th>
<th>24</th>
<th>25</th>
<th>26</th>
<th>27</th>
<th>28</th>
<th>29</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth (km)</td>
<td>36</td>
<td>44</td>
<td>53</td>
<td>63</td>
<td>75</td>
<td>90</td>
<td>110</td>
<td>140</td>
<td>180</td>
<td>230</td>
<td>300</td>
<td>400</td>
<td>580</td>
<td>800</td>
</tr>
</tbody>
</table>

Inversion Data and Error Floors

In the various inversions of the Corridor 1 and 1A data, the GB TE and TM apparent resistivity, phase and tipper were inverted. The tipper responses inverted were the real and imaginary components of the tipper perpendicular to the strike of the model. The number of inversion periods used was 14. Table 4.10 lists the periods chosen in inverting the
SNORCLE data. These periods are sufficient to represent the variation of the observed responses and are approximately evenly spaced in log space. The penetration depth of periods chosen is up to 200 km (Fig. 4.8 and 4.9).

**Table 4.10 Inversion periods.**

<table>
<thead>
<tr>
<th>No.</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
</tr>
</thead>
<tbody>
<tr>
<td>Period (s)</td>
<td>0</td>
<td>0</td>
<td>0.01</td>
<td>0.031</td>
<td>0.083</td>
<td>0.312</td>
<td>0.833</td>
</tr>
<tr>
<td>No.</td>
<td>8</td>
<td>9</td>
<td>10</td>
<td>11</td>
<td>12</td>
<td>13</td>
<td>14</td>
</tr>
<tr>
<td>Period (s)</td>
<td>3.846</td>
<td>10</td>
<td>38.46</td>
<td>100</td>
<td>312</td>
<td>833</td>
<td>3125</td>
</tr>
</tbody>
</table>

Error floors (user-defined minimum error limits) were applied to the data because of uncertainties in the spectral error estimates. A 5% impedance error floor corresponds to 2.86° phase error and a 10% apparent resistivity error. GEOTOOLS uses a log unit to specify the error for apparent resistivity. A ±10% linear unit error \((\rho/1.1 \sim \rho_0 \times 1.1)\) corresponds to ±0.0434 log unit error \((\log \rho_0 \pm 0.0434)\), because \(10^{0.0434} \approx 1.1\). In the case of the tipper GEOTOOLS uses the variance to specify the error, for example, a 0.0225 error specified in GEOTOOLS corresponds to 0.15 tipper error.

For cases in which a specific response (e.g. \(\rho_{TM}\) and \(\phi_{TM}\)) or data at a specific period needs to be fitted as closely as possible, a smaller error can be set. It is also possible to use a larger error for the data that is to be downweighted in an inversion (e.g. ±1 log unit for the apparent resistivity; ±50° for phase). For most sites on the Phanerozoic sedimentary rocks in SNORCLE area, the static shift estimates is small, about 0.8-1.2 (Table. 4.5). In order to account for minor static shift in the apparent resistivity responses it is possible to increase the error floors on the apparent resistivity responses while forcing the inversion to fit the phase and/or tipper more closely.
Summary of Occam Inversions Done

A range of inversions need to be performed in order to obtain a satisfactory geoelectric model and to examine its uniqueness and resolution, e.g., inversions with different initial models, different inversion error floors, and different data responses (TE mode only, TM mode only, both modes, both modes and tipper etc.). Table 4.11 and 4.12 summarise some inversions completed for the profiles along Corridor 1 and 1A.

Table 4.13 lists the tolerance and roughness of each iteration from a typical inversion for each profile. Typically, after each iteration, the misfit (tolerance) decreased and the roughness increased. However, the roughness showed some oscillations. The model picked as the final model or the initial model for the next inversion was not always the final iteration result. It was chosen such that its tolerance was close to minimum value and its roughness was not extreme.

The iteration result shown in the black font in Table 4.13 is a model chosen as the initial model for the final inversion of the data from each profile. The specified RMS misfit used in the final inversion was set to the RMS of the best fit plus 10% (Table 4.11 and 4.12).

4.8.4 Two-dimensional Occam Inversion Results

Figure 4.26 compares the 2D Occam inversion models from inversion 8, 10 and 13 in Table 4.11 for the profile from Nahanni terrane to Hottah terrane along Corridor 1. These models are based on inverting TE data (TE model), TM data (TM model), and both TE and TM data and tipper (joint model) respectively.
Table 4.11 Occam inversions completed for the profile along Corridor 1.

<table>
<thead>
<tr>
<th>No.</th>
<th>GB $\rho_{1E}$ (log unit)</th>
<th>GB $\phi_{1E}$ (degree)</th>
<th>GB $\rho_{TM}$ (log unit)</th>
<th>GB $\phi_{TM}$ (degree)</th>
<th>Tipper</th>
<th>Specified RMS Misfit</th>
<th>Initial Model</th>
<th>No. of Invert Data</th>
<th>No. of Iteration$^1$</th>
<th>Inversion Results Roughness Tolerance</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.1</td>
<td>2.86</td>
<td>0.1</td>
<td>2.86</td>
<td>NI$^2$</td>
<td>1.0</td>
<td>200 $\Omega$m half-space</td>
<td>1312</td>
<td>15</td>
<td>59.5</td>
</tr>
<tr>
<td>2</td>
<td>0.1</td>
<td>2.86</td>
<td>0.1</td>
<td>2.86</td>
<td>0.15</td>
<td>1.0</td>
<td>200 $\Omega$m half-space</td>
<td>1914</td>
<td>15</td>
<td>40.9</td>
</tr>
<tr>
<td>3</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>1.0</td>
<td>horizontal fault at 1 km depth</td>
<td>1312</td>
<td>program stop after 1 iteration</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.15</td>
<td>1.0</td>
<td>horizontal fault at 1 km depth</td>
<td>1914</td>
<td>program stop after 6 iterations</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>1.0</td>
<td>fix surface layer by 1D inversion</td>
<td>1312</td>
<td>12</td>
<td>42.6</td>
</tr>
<tr>
<td>6</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.15</td>
<td>1.0</td>
<td>fix surface layer by 1D inversion</td>
<td>1914</td>
<td>15</td>
<td>37.6</td>
</tr>
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<td>7</td>
<td>NI</td>
<td>NI</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
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<td>200 $\Omega$m half-space</td>
<td>638</td>
<td>15</td>
<td>16.8</td>
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<tr>
<td>8</td>
<td>NI</td>
<td>NI</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>1.0</td>
<td>one of iterations in last inversion</td>
<td>638</td>
<td>11</td>
<td>24.3</td>
</tr>
<tr>
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<td>NI</td>
<td>NI</td>
<td>NI</td>
<td>1.0</td>
<td>200 $\Omega$m half-space</td>
<td>638</td>
<td>15</td>
<td>16.5</td>
</tr>
<tr>
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<td>2.86</td>
<td>NI</td>
<td>NI</td>
<td>NI</td>
<td>1.0</td>
<td>one of iterations in last inversion</td>
<td>638</td>
<td>8</td>
<td>21.1</td>
</tr>
<tr>
<td>11</td>
<td>0.0868</td>
<td>5.73</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>1.0</td>
<td>200 $\Omega$m half space</td>
<td>1312</td>
<td>15</td>
<td>35.3</td>
</tr>
<tr>
<td>12</td>
<td>0.0868</td>
<td>5.73</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.15</td>
<td>1.0</td>
<td>one of iterations in last inversion</td>
<td>1914</td>
<td>15</td>
<td>42.6</td>
</tr>
<tr>
<td>13</td>
<td>0.0868</td>
<td>5.73</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.15</td>
<td>1.85</td>
<td>one of iterations in last inversion</td>
<td>1914</td>
<td>6</td>
<td>29.6</td>
</tr>
</tbody>
</table>

$^1$ For each inversion a total of 15 iterations was specified. Entries less than 15 indicate the inversions stop at a smaller number.

$^2$ NI: parameter not included in the inversion.
Table 4.12 Occam inversions completed for the profile along Corridor 1A.

<table>
<thead>
<tr>
<th>No.</th>
<th>Inversion Parameters and Error Floor</th>
<th>Specified RMS Misfit</th>
<th>Initial Model</th>
<th>No. of Invert Data</th>
<th>No. of Iteration</th>
<th>Roughness Tolerance</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>GB $\rho_E$ (log unit)</td>
<td>GB $\phi_E$ (degree)</td>
<td>GB $\rho_M$ (log unit)</td>
<td>GB $\phi_M$ (degree)</td>
<td>Tipper</td>
<td></td>
</tr>
<tr>
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<td>2.86</td>
<td>0.1</td>
<td>2.86</td>
<td>NI²</td>
<td>1.0</td>
</tr>
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<td>0.1</td>
<td>2.86</td>
<td>0.15</td>
<td>1.0</td>
</tr>
<tr>
<td>3</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>1.0</td>
</tr>
<tr>
<td>4</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>1.0</td>
</tr>
<tr>
<td>5</td>
<td>NI</td>
<td>NI</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>1.0</td>
</tr>
<tr>
<td>6</td>
<td>NI</td>
<td>NI</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>2.0</td>
</tr>
<tr>
<td>7</td>
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<td>2.86</td>
<td>NI</td>
<td>NI</td>
<td>NI</td>
<td>1.0</td>
</tr>
<tr>
<td>8</td>
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<td>2.86</td>
<td>NI</td>
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<td>2.0</td>
</tr>
<tr>
<td>9</td>
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<td>5.73</td>
<td>0.0434</td>
<td>2.86</td>
<td>NI</td>
<td>1.0</td>
</tr>
<tr>
<td>10</td>
<td>0.0868</td>
<td>5.73</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.15</td>
<td>1.0</td>
</tr>
<tr>
<td>11</td>
<td>0.0868</td>
<td>5.73</td>
<td>0.0434</td>
<td>2.86</td>
<td>0.15</td>
<td>3.15</td>
</tr>
</tbody>
</table>

¹For each inversion a total of 15 iterations was specified. Entries less than 15 indicate the inversions stop at a smaller number.
²NI: parameter not included in the inversion
The GB responses of the TE and TM models are shown in Appendix B. The overall fit to the data is moderately good with normalized $\chi^2$ misfits (tolerance) of 2.44 for the TE model and 2.07 for the TM model (Table 4.11). Examination of the results reveals a number of similarities and differences between the TE and TM models (Fig. 4.26a, b):

- The TE model is more conductive than the TM model.
- Both models show the conductive Phanerozoic sedimentary rocks at the near-surface.
- Both TE and TM models show a relative resistive structure beneath several kilometres depth at sites 145-134 and above 30 km depth at sites 129-122.
Figure 4.26 Comparison of Occam inversion models, (a) inversion of only TE apparent resistivity and phase, (b) inversion of only TM apparent resistivity and phase, (c) inversion of TE and TM data and tipper.
A stronger resistivity contrast is observed between sites 134 and 133 in the TE model but the contrast occurs between sites 130 and 129 in the TM model. However, in both models the resistivity is lower in the east than in the west.

The conductive body at 15-50 km depth beneath sites 131 and 130 in the TE model is not observed in the TM model. This observation is consistent with the observed apparent resistivity and phase (Fig. 4.19). The observed TE responses showed a conductive zone at periods >10 s beneath sites 131 and 130. Because of the lower TE apparent resistivity and higher TE phase at periods >100 s beneath sites 130-122 (Fig. 4.19) compared with the corresponding TM responses, the TE inversion model includes a more conductive lower-crust and upper mantle than the TM model (Fig. 4.26a, b).

The differences between the TE and TM models imply the existence of significant 2D or 3D structures. Therefore, the inversion of both TE and TM mode responses is necessary in order to provide a more accurate resistivity structure (Fig. 4.26c).

The GB responses of the joint model are shown in Fig. 4.27-4.29. The normalized χ² misfit is 1.66 for this model. The model apparent resistivity and phase responses fit the observed data well. However, because of edge effects, the misfit between the model responses and the observed data is larger at sites which are close to the edge of the model, for example sites 145, 144 and 122. The magnitude of the tipper is approximately zero at periods <1 s and <0.2 at periods >10 s. The fit to the tipper is good at periods <10 s but poor at long periods. The observed tipper is poorly resolved at period >100 s.

Figure 4.30 compares the 2D Occam inversion models from inversion 6 (TE data only), 8 (TM data only) and 11 (TE and TM data and tipper) for the profile along Corridor
Figure 4.27 Comparison of the apparent resistivity for the Occam inversion model responses and observed data at the sites on the profile from Nahanni terrane to Hottah terrane. Solid line: TM response, dash line: TE response, +: TM observed data, o: TE observed data.
Figure 4.28 Comparison of the phase for the Occam inversion model responses (line) and observed data (symbol) at the sites on the profile from Nahanni terrane to Hottah terrane. (a) TE Phase, (b) TM phase.
(b) TM phase
Figure 4.29 Data fit for the tipper at sites along profile from the Nahanni terrane to Hotah terrane Square: real part of observed tipper Ty, triangle: the imaginary part of observed tipper Ty, solid line: calculated real part of tipper, dash line: calculated imaginary part of tipper.
Figure 4.30 Comparison of Occam inversion models along Corridor IA, (a) only invert the TE apparent resistivity and phase, (b) only invert the TM apparent resistivity, (c) invert the TE and TM data and tipper.
The GB responses of the TE and TM models are shown in Appendix C. The normalized $\chi^2$ misfit is 4.18 for the TE model and 4.03 for the TM model (Table 4.11). The data fit of TE model responses is poorer than TM model responses. Both models reveal a number of similarities and differences (Fig. 4.30a, b):

- The TE model is more conductive than the TM model.
- Both models show the conductive Phanerozoic sedimentary rocks at the near-surface.
- Both models show a resistive middle and lower crust beneath sites 154-160.
- Both TE and TM models include a relatively conductive zone beneath sites 150 and 152 (Fig. 4.30a, b). This conductive zone has a $\sim 300 \Omega m$ resistivity and extends from the near-surface to the upper mantle in the TM model. In the TE model, this zone is more conductive ($<100 \Omega m$) and is present only below 50 km.
- The TE model includes a conductive middle crust and mantle beneath sites 129-149 ($<100 \Omega m$), which differs with the TM model.

Inversion of both TE and TM responses is necessary to provide a more accurate resistivity structure (Fig. 4.30c), which combines the structural features from TE and TM models. The GB responses of this model (apparent resistivity, phase and tipper) are shown in Fig. 4.31-4.33. The normalized $\chi^2$ misfit is 2.84 (Table 4.12). The apparent resistivity and phase show a poor fit at periods longer than 10 s at some sites, for example sites 147 to 154. The model tipper responses have the same form as the observed responses at most sites, except for the long period responses at sites 153-155. These results suggest a more complex crustal structure at depth, possibly including 3D structure, around sites 147-155.
Figure 4.31 Comparison of the apparent resistivity for the Occam inversion model responses and observed data at the sites on Corridor IA. Solid line: TM response, dash line: TE response, +: TM observed data, o: TE observed data.
Figure 4.32 Comparison of the phase for the Occam inversion model responses (line) and observed data (symbol) at the sites on Corridor 1A. (a) TE Phase, (b) TM phase.
(b) TM phase
Figure 4.33 Data fit of tipper at sites along Corridor 1A. Square: real part of observed tipper Ty, triangle: the imaginary part of observed tipper Ty, solid line: calculated real part of tipper, dash line: calculated imaginary part of tipper.
4.8.5 Two-dimensional NLCG Inversion Methodology

Design of Meshes and Grids

The design of the rows in the mesh for the NLCG inversions is the same as the Occam model. The design of columns in the mesh is dependent on the spatial variation of resistivity. There should be only one site in each column. In the NLCG inversion model, the regularization grid is identical to the finite element mesh.

Input Data and Inversion Parameters

The following information is required for each NLCG inversion.

1) A starting model (see Table 4.14a). The shallow part of the Occam inversion models (<1 km) are also used to defined the near-surface structure of the stating model.

2) RMS error for termination of inversion. This parameter is the same as the tolerance specified in the Occam inversions. It was set to 1.0 for most inversions (Table 4.14b). The program will terminate after RMS reaches this value if the specific maximum number of inversion iterations has not been reached.

3) A parameter Tau, which controls the tradeoff between fitting the data and adhering to the model constraint. Larger values of tau produce a smoother model at the expense of a worse data fit. Tau values between 3 and 300 seem to be suitable for most inversions (Mackie et al. 1997b).

4) The TE and TM mode apparent resistivities, phases and error floors for each period at each station. The NLCG program used does not calculate a tipper response.
Table 4.14 (a) Starting model in NLCG inversion and (b) NLCG inversions completed for profile along Corridor 1 and 1A.

<table>
<thead>
<tr>
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</thead>
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<td>100</td>
</tr>
<tr>
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<tr>
<td>3&lt;sup&gt;rd&lt;/sup&gt; layer</td>
<td>300</td>
<td>Half-space</td>
</tr>
</tbody>
</table>

(a)

<table>
<thead>
<tr>
<th>No</th>
<th>Inversion Parameter and Error Floor</th>
<th>Tau</th>
<th>Specified RMS Error</th>
<th>No. of Iterations Corridor 1</th>
<th>Corridor 1A</th>
</tr>
</thead>
<tbody>
<tr>
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<td>GB ρ&lt;sub&gt;TE&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TE&lt;/sub&gt; 2.5% GB ρ&lt;sub&gt;TM&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TM&lt;/sub&gt; 2.5%</td>
<td>3</td>
<td>1</td>
<td>53</td>
<td>60</td>
</tr>
<tr>
<td>2</td>
<td>GB ρ&lt;sub&gt;TE&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TE&lt;/sub&gt; 2.5% GB ρ&lt;sub&gt;TM&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TM&lt;/sub&gt; 2.5%</td>
<td>30</td>
<td>1</td>
<td>43</td>
<td>60</td>
</tr>
<tr>
<td>3</td>
<td>GB ρ&lt;sub&gt;TE&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TE&lt;/sub&gt; 2.5% GB ρ&lt;sub&gt;TM&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TM&lt;/sub&gt; 2.5%</td>
<td>100</td>
<td>1</td>
<td>32</td>
<td>53</td>
</tr>
<tr>
<td>4</td>
<td>GB ρ&lt;sub&gt;TE&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TE&lt;/sub&gt; 2.5% GB ρ&lt;sub&gt;TM&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TM&lt;/sub&gt; 2.5%</td>
<td>300</td>
<td>1</td>
<td>45</td>
<td>36</td>
</tr>
<tr>
<td>5</td>
<td>GB ρ&lt;sub&gt;TE&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TE&lt;/sub&gt; 2.5% GB ρ&lt;sub&gt;TM&lt;/sub&gt; 5% GB ϕ&lt;sub&gt;TM&lt;/sub&gt; 2.5%</td>
<td>NI</td>
<td>NI</td>
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<td>GB ϕ&lt;sub&gt;TM&lt;/sub&gt; 2.5%</td>
<td>30</td>
</tr>
</tbody>
</table>

NI: parameter not included in the inversion

(b)

As for the 2D Occam inversions, multiple inversions using different inversion parameters (τ<sub>au</sub>, error floors), starting models and data responses (TE and/or TM responses, apparent resistivity and/or phase) were completed. Table 4.14 lists some inversions for profiles along Corridor 1 and 1A. The inversion periods chosen in NLCG inversions are the same as those in the Occam inversions (Table 4.10).

4.8.6 Two-dimensional NLCG Inversion Results

Figure 4.34 compares the 2D NLCG models for different τ<sub>au</sub> values (3, 30 and 300) for the profile from the Nahanni to Hottah terranes along Corridor 1. Figure 4.35 shows the comparison of the model responses with τ<sub>au</sub>=30 and the observed data as well as the
Figure 4.34 Comparison of NLCG inversion models along Corridor 1. The inversions used the various $\tau$ values: (a) 3, (b) 30, (c) 300.
Figure 4.35 Comparison of the apparent resistivity and phase from NLCG inversion model responses and the observed data at the sites on the profile from Nahanni terrane to Hottah terranes. Solid line: TM response, dash line: TE response, ●: TM observed data, ○: TE observed data. RMS is misfit. (a) sites 145-134, (b) sites 133-122.
normalized \( \chi^2 \) values for each site. The responses for the other two models are shown in Appendix D. The models provide the reasonably good fit to the data. The overall root mean square misfit, based on error floors of 5\% for the apparent resistivity data and 2.5\% for the phase data, is 3.4, 3.8 and 4.4 for \( \text{Tau} = 3, 30 \) and 300 respectively.

The three models show the same basic structure. Model B (\( \text{Tau} = 30 \)) is similar to model A (\( \text{Tau} = 3 \)), but as expected is smoother. Model C is the smoothest model (\( \text{Tau} = 300 \)) and the boundaries of different resistivity bodies are very gradual (Fig. 4.34c). If the \( \text{Tau} \) value is too low, the model will be relatively rough. Therefore, the inversion model with 30 \( \text{Tau} \) was taken as the final model in this study (Fig. 4.34b) because it provided both reasonable fit to data and reasonably smooth model.

Figure 4.36 compares the 2D NLCG models for the different \( \text{Tau} \) value (3, 30 and 300) for the profile along Corridor 1A. The model responses and data fitting RMS value for model with \( \text{Tau} = 30 \) are shown in Fig. 4.37 and the responses for the other two models are shown in Appendix E. The models provide the reasonably good fit to the data. The overall root mean square misfit, based on error floors of 5\% for the apparent resistivity data and 2.5\% for the phase data, is 4.3, 4.5 and 5.4 for \( \text{Tau} = 3, 30 \) and 300 respectively.

The NLCG model responses provide the much better fit to the data (Fig. 4.37) than Occam model responses (Fig. 4.31 and 4.32). For example, at site 146 the NLCG model responses fit the data very well (Fig. 4.37), but the Occam model responses do not fit the data at periods >10 s (Fig. 4.31). It is difficult to get a two-dimensional NLCG model to fit both the TE and TM data well at several sites near the centre of the profile suggesting a more complex crustal structure (with a 3D component). In the inversion model obtained from the
Figure 4.36 Comparison of NLCG inversion models along Corridor IA. The inversions used the different Tau values: (a) 3, (b) 30, (c) 300.
Figure 4.37 Comparison of the apparent resistivity and phase from NLCG inversion model responses and the observed data at the sites along Corridor 1A. Solid line: TM response, dash line: TE response, ●: TM observed data, ○: TE observed data. RMS is the misfit. (a) sites 129-149, (b) sites 150-160.
combined data set the TE data shows the poorest fits, particularly at periods longer than 10 s at sites 150, 151, 154, 155, and 156. Modelling studies have shown that the TM response is more robust than the TE response in the presence of three-dimensional structures (Jones 1983, Wannamaker et al. 1984b) so the good fit to the TM response provides increased confidence in the two-dimensional models.

4.8.7 Examination of 2D Inversion Models

The shallow (< 2km) structures observed from the 2D Occam inversion are shown in Fig. 4.38, which provide more detailed resolution of the Phanerozoic sedimentary rock than the 2D NLCG inversion. Figure 4.39 and 4.40 compare the deeper structure in the final 2D Occam and NL CG inversion models along Corridor 1 and 1A. The structures in the different geographical regions of the models will now be discussed.

4.8.7.1 Nahanni Terrane, Fort Simpson Terrane and Western Hottah Terrane

The 2D Occam inversion model (Fig. 4.38a) also provided a conductive shallow structure (<1 km) as the 1D inversion results (Fig. 4.22a). Both 1D and 2D models show the thickness of the conductive (~10 Ωm) Phanerozoic sedimentary rocks is about 1000 m beneath the Fort Simpson terrane and thin to the east. Both models include a resistive zone beneath sites 143 and 142 at the boundary of the Nahanni and Fort Simpson terranes. This similarity between 2D and 1D inversions combined with a good fit to the data at short periods, provides increased confidence in the 1D inversion results.

The crustal resistivity structure shows significant lateral variations. Key features are
Figure 4.38 Shallow structures from 2D Occam inversion. (a) Profile from Nahanni terrane to Hottah terrane along the Corridor 1, (b) profile along Corridor 1A.
Figure 4.39 Final 2D inversion models from (a) Occam inversion and (b) NLCG inversion along the profile from Nahanni terrane to Hottah terrane. Letters refer to features discussed in text.
Figure 4.40 Final 2D inversion models from (a) Occam inversion and (b) NLCG inversion along Corridor 1A.
listed below (from west to east):

- There is a resistivity boundary beneath sites 143 and 142 at the boundary of Nahanni and Fort Simpson terranes. This structure is labelled A in Fig. 4.39. A relatively conductive body (about 230 $\Omega$m) extending to 20 km depth separates more resistive zone (>500 $\Omega$m) to the east and west. Although this structure is defined mainly by the response at site 143, the feature can be observed in both TE and TM apparent resistivity and phase pseudosections (Fig. 4.19). The TE and TM phases are relatively high at periods of 0.1 s to 100 s (20-40°) compared with phases in surrounding areas (10-30°). The real induction vector is relatively large (0.3) and points to southwest at sites 143 and 142 (Fig. 4.10b,c).

- A relatively resistive body is present in the upper crust in the Fort Simpson terrane, labelled B in Fig. 4.39. This body is resolved more clearly in NLCG model (Fig. 4.39b) than in Occam model (Fig. 4.39a). The resistivity of the body is 400-800 $\Omega$m in the Occam model and >1000 $\Omega$m in the NLCG model. The depth to the base of the body increases from about one kilometre under site 133 to about 20 km under site 141. The west dipping base (15°-20°) flattens at the west of the profile. This body is underlain by a relatively conductive zone (~130-250 $\Omega$m), labelled C in Fig. 4.39.

The maximum phase difference (Fig. 4.10) shows an east-west direction at sites 140-134 at period 10 s (3-7 km depth, Table 4.3), which is different with N20°E in the surrounding area (Fig. 4.10b). The real induction vectors have a reversal of direction at sites in the Fort Simpson terrane at period 10 s (Fig. 4.10b) and point to the west in the eastern Nahanni and Fort Simpson terranes at period 100 s (Fig. 4.10c). These results indicate the lateral and vertical variations of the structures in the Nahanni and Fort
Simpson terranes.

One-dimensional methods can be used to examine the resolution of the resistive body, however, the 1D methods used will provide only a guide to the true resolution of the 2D structures. The program used in this analysis calculates 1D Fréchet derivative and is provided by Dr. Ian Ferguson. Figure 4.41 plots the MT model and the corresponding 1D Fréchet derivative at site 139. The model clearly shows a resistive layer at above 25 km depth. Within the depth of the resistive zone the normalized Fréchet derivative is approximately one-half of the surface values. The magnitude of the Fréchet derivative increases at about 20-25 km depth. The results suggest the MT response can resolve the depth to the base of the resistive layer and provide some resolution of the thickness and resistivity of the resistive layer.

Two-dimensional forward modelling was done to check the validity of the 1D Fréchet derivative results. The results show a change in the depth to the top of the conductor (either shallower or deeper) causes a variation of the TE and TM apparent resistivity and phase. This indicates the MT responses can resolve the depth to the top of the conductive layer (or the base of the resistive layer). The responses show only slight variation when the structure beneath the conductive layer is changed in the 2D model, indicating the low resolution of the features beneath the conductors. These results agree with the analysis using 1D Fréchet derivatives and support the use of 1D Fréchet derivative for examining other parts of the model.

An equivalence analysis was completed for the site 139 using a simplified three-layer model and the SVD method. The inversion frequencies used in 1D inversion are
Figure 4.41 (a) MT Model at site 139 derived from the 2D model, (b) normalized Frechet derivative magnitude at site 139.
from 0.001 s to 1000 s. The model, resolution matrix and estimates of parameter uncertainty are shown in Table 4.15.

Table 4.15 (a) Simplified three-layer model at site 139, (b) its resolution matrix, and (c) estimates of parameter uncertainty.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Resistivity (Ωm)</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>20</td>
<td>1</td>
</tr>
<tr>
<td>2</td>
<td>2000</td>
<td>24</td>
</tr>
<tr>
<td>3</td>
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(a)

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</thead>
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<td>2nd layer resistivity ($\rho_2$)</td>
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<tr>
<td>3rd layer resistivity ($\rho_3$)</td>
<td>0 -0.001 0.502</td>
</tr>
<tr>
<td>1st layer thickness ($t_1$)</td>
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</tr>
<tr>
<td>2nd layer thickness ($t_2$)</td>
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(b)

<table>
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<td>24</td>
</tr>
<tr>
<td>$\rho_2$</td>
<td>208</td>
<td>2000</td>
<td>19234</td>
</tr>
<tr>
<td>$\rho_3$</td>
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<td>508</td>
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<td>$t_2$</td>
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<td>117</td>
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</tbody>
</table>

(c)

The resistivity and thickness of the first layer can be resolved well. The thickness of the second layer has relatively low resolution because it depends partly on its resistivity.
and the underlying layer (Table 4.15). The resolution of the thickness increases as the ratio of the resistivity of the second layer to the resistivity of the third layer increases.

- A clear sub-vertical conductive zone is present at crustal depths beneath sites 131 and 130 and extends to site 129 in Occam models (labelled D in Fig. 4.39). The zone has a resistivity of less than 100 $\Omega$m and extends from ~10 km depth to the lower crust and mantle. This conductive zone includes 3D structure at depth, because the models provide a poor data fit (including to the tipper) at long periods (>100 s) for these two sites.

The presence of this structure is supported by the observed responses. The TE responses show relatively low apparent resistivity and relatively high phase beneath sites 132-130 (Fig. 4.19). The real induction vectors are larger in the east of site 129 at period 100 s and point to the southwest, indicating the present of conductor (Fig. 4.10c).

A Fréchet derivative analysis for the model at site 130 (Fig. 4.42) clearly shows the high normalized Fréchet derivative at depth 10-30 km and very low value (close to zero) beneath 40 km. The results suggest the MT response can provide some resolution of the thickness and resistivity of the conductive layer. The very low normalized Fréchet derivative indicates the change of response is very small with the variation of resistivity beneath 40 km. That means there is no resolution for the structure beneath the conductor, at depths exceeding ~40 km.

An equivalence analysis was completed for the site 130 using a simplified four-layer model. The model, resolution matrix and estimates of parameter uncertainty are shown in Table 4.16.
Figure 4.42 (a) MT Model at site 130, (b) normalized Frechet derivative magnitude at site 130. The sensitivity is relatively high for the conductive layer.
Table 4.16 (a) Simplified four-layer model at site 130, (b) its resolution matrix, and (c) estimates of parameter uncertainty.

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<td>3rd layer thickness ($t_3$)</td>
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(b)

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<th>Parameter</th>
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<th>Best-fit Value</th>
<th>Maximum Value</th>
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<td>9948</td>
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<td>$\rho_3$ (Ωm)</td>
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<td>8</td>
<td>18</td>
</tr>
<tr>
<td>$\rho_4$ (Ωm)</td>
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<td>428</td>
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<td>$t_1$ (km)</td>
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<td>$t_3$ (km)</td>
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<td>127</td>
</tr>
</tbody>
</table>

(c)
The resolution matrix elements for the thickness of the first and second layers are 0.968 and 0.906 respectively. The 95% confidence bounds on the thickness of the first layer are from 660 m to 1500 m with a best-fit value of 1 km and from 4.4 km to 18 km with a 9 km best fit value for the second layer (Table 4.16c). The Fréchet derivative results suggest the depth to the top of the conductive third layer (~10 km) can be resolved by MT data. The thickness of conductive third layer cannot be resolved as accurately. Therefore, the base to the conductor cannot be resolved.

- There is a broad conductive zone (less than 150 Ωm) in the mantle beneath the Hottah terrane which is labelled E in Fig. 4.39. The zone starts in the Fort Simpson terrane at about 110 km depth and rises beneath the Hottah terrane. It is best resolved in the NLCG model. This conductive region can be observed in TE Occam model and the region is more conductive than crust to the west in the TM Occam model (Fig. 4.26).

This conductive body is not immediately evident in the apparent resistivity and phase pseudosections (Fig. 4.19 and 4.20), but is required by the separation of TE and TM apparent resistivity and phase at long periods (sites 129-122 in Fig. 4.35). As discussed in section 4.8.1, the spatial variations of the conductive Phanerozoic sedimentary rocks cannot cause large-scale separation of TE and TM responses at long periods. Therefore, the separation of the TE and TM data implies a 2D or 3D crustal and/or mantle structure.

To the east, this conductive zone appears to connect with a more conductive zone (~20 Ωm) in the lower crust (Fig. 4.39a) with its top at ~40 km depth. The model responses fit the observed data (including the tipper) well in this part of the model (Fig. 4.27-4.29), which adds supports for the inversion model.
4.8.7.2 Hottah Terrane, Great Bear Magmatic Arc and Buffalo Head Terrane

The Occam and NLCG inversion models from Corridor 1 show a conductive zone beneath the Hottah terrane which is located at the right edge of the model (Fig. 4.39a and Fig. 4.39b). Corridor 1A shows a corresponding conductive zone at the western end of the profile. In order to determine whether this is an edge effect of 2D Occam inversion or a real structure, Corridor 1A was extended further to the west. The extended profile includes all sites on Corridor 1A and sites 129-122 from Corridor 1. The deeper crustal structures in the Occam and NLCG inversion models of this data set include (Fig. 4.40):

- The broad conductive zone (less than 130 Ωm) at >60 km depth beneath the Hottah terrane, labelled F1 in Fig. 4.40. All inversion models from both the Occam and NLCG methods (including those using the tipper) show this conductive body. These results indicate that it is a real geoelectric structure, and not an edge effect. This zone may connect with a more conductive zone (~30 Ωm) in the lower crust beneath the boundary of the Hottah terrane and Great Bear Magmatic Arc, labelled F2 in Fig. 4.40. There is sharp eastern edge to this zone at site 147. Seismic results indicate the depth to the Moho on Corridor 1 at its intersection with Corridor 1A is ~40 km (Cook et al. 1999). The MT images therefore indicate that the conductive zone extends into the mantle.

Figure 4.43 shows the MT model and magnitude of Fréchet derivative at site 122. The normalized Fréchet derivative increases at depths greater than the about 20 km, indicating the conductive body should be resolved. There is no resolution of the structure beneath this conductor, at depths exceeding ~80 km.

The equivalence analysis for the simplified four-layer model at site 122 indicates
Figure 4.43 (a) MT Model at site 122, (b) normalized Frechet derivative magnitude at site 122.
that it is not possible to resolve exactly the depth to the conductive body below the resistive layer (Table 4.17).

**Table 4.17** (a) Simplified four-layer model at site 122, (b) its resolution matrix, and (c) estimates of parameter uncertainty.

<table>
<thead>
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<th>Resistivity (Ωm)</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
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(a)

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(b)

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<td>$t_3$</td>
<td>14</td>
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<td>186</td>
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The resolution matrix element of the second-layer thickness is 0.497, and the bound of thickness is from 3.7 km to 97 km for a best-fit value 19 km. Therefore, the thickness of second layer cannot be resolved well and affects the resolution of underlying layers (e.g. the correlation resolution matrix element is 0.274 between the second-layer thickness and the third-layer resistivity).

- In the NLCG model, there is a conductive zone (< 50 Ωm, labelled G) in the centre of the Great Bear terrane with its top at ~15 km depth below sites 149 and 150. Its presence is supported by the observations in the apparent resistivity pseudosections (Fig. 4.20) which show moderately low resistivity at periods of 10 to 1000 s beneath sites 149 and 150.

- There is an anomalous conductive zone beneath sites 150 to 154 (labelled H in Fig. 4.40). The resistivity of the zone reaches values of <100 Ωm in all of models. The upper surface of the zone appears to have an eastward dip. In some models this conductor connects with the previous conductor but this aspect of the model is not well resolved. The eastern margin of this conductor at sites 154 and 155 consists of a sharp transition to more resistive zone to the east.

For a 1D model based on the 2D resistivity structure beneath site 153, the Fréchet derivative begins to increase at about 45 km depth at periods longer than 100 s (Fig. 4.44). The Schmucker depth is deeper than 57 km at period longer than 100 s (Table. 4.4). These results indicate that the long period (>100 s) responses are sensitive to deep conductor.

The equivalence analysis for a simplified four-layer model at site 153 is shown in Table 4.18.
Figure 4.44 (a) MT Model at site 153, (b) normalized Frechet derivative magnitude at site 153.
Table 4.18 (a) Simplified four-layer model at site 153, (b) its resolution matrix, and (c) estimates of parameter uncertainty.

<table>
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</tr>
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(a)

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<td>4&lt;sup&gt;th&lt;/sup&gt; layer resistivity ((\rho_4))</td>
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<tr>
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</tr>
</tbody>
</table>

(b)

<table>
<thead>
<tr>
<th>Parameter</th>
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<th>Best-fit Value</th>
<th>Maximum Value</th>
</tr>
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<td>(t_3) (km)</td>
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</table>

(c)

The MT data can resolve the thickness and resistivity of the first layer, and resistivity of the second layer. The resolution of the thickness of the second layer is
moderately low (0.882 resolution matrix element) because of the partial dependency on the resistivity of the underlying layer (\(R_{t_j\rho_j}=0.1\)). The thickness bound of the second layer is relatively high, from 5.9 km to 28 km for a best-fit value 13 km. The thickness of third layer can be resolved (\(R_{t_j\rho_j}=0.916\)). Therefore, the long period MT responses (>100 s) can resolve the resistivity variations below 40 km depth.

- There is a sub-vertical resistive zone (~10,000 \(\Omega m\)) beneath sites 155 and 156, which is included in both Occam and NLCG models (labelled I in Fig. 4.40). The observed TE and TM apparent resistivities at sites 154-156 are also high at periods 10-300 s (Fig. 4.20).

  The tipper has a higher value at periods >10 s at sites 154 and 155 (Fig. 4.33), which suggests a strong resistivity contrast. The induction vectors show a slight change of direction between locations to the west of site 155 and to the east of site 155 at a period of 10 s (Fig. 4.10c). Maximum phase difference direction is approximately N60°E in the Great Bear Magmatic Arc, and is close to north-south in the Buffalo Head at period 100 s. These results all support a 2D or 3D structure in the boundary around sites 154 and 155.

- The upper crust in the Buffalo Head terrane is more conductive than the lower crust (in the zone labelled J in Fig. 4.40). There is an indication of eastward dipping crustal structure but because the period range of the data from sites 159 and 160 is less than at other sites this feature is not well resolved.

4.8.8 Conductivity Structure of Lithosphere

The TE and TM apparent resistivity and phase pseudosections (Fig. 4.19 and 4.20) showed the low resistivity (<100 \(\Omega m\)) and high phase beneath the Fort Simpson terrane,
Hottah terrane and Great Bear Magmatic Arc at periods $>1000$ s. These features are not observed beneath the Slave Province and the Buffalo Head terrane. These observations indicate a spatial variation in the deep lithosphere along Corridor 1 and 1A.

In order to obtain a preliminary indication of the deep lithospheric structure the 1D and 2D inversions were completed for the long period MT responses ($>10$ s). 1D Fischer inversion models for TE and TM responses are shown in Fig. 4.45. The models include a somewhat similar conductive deep structure. The depth to the top of the conductive layer is shallower beneath the Hottah terrane and the western Great Bear Magmatic Arc and becomes deeper towards the Fort Simpson terrane in the west and the Buffalo Head terrane in the east. The TE model shows a generally more shallow depth than the TM model and the resistivity in the TE model ($\sim10$ $\Omega$m) is lower than in the TM model ($50$ $\Omega$m). These results indicate the influence of 2D structures on the resolution of the deep lithospheric structure.

In order to examine and improve the deep structures, 2D Occam inversions for long periods MT responses ($>10$ s) were completed along Corridor 1 and 1A. The TE and TM apparent resistivities and phases and tipper were inverted. The final tolerance (normalized $\chi^2$ misfit) and roughness is 2.2 and 45.3 for Corridor 1, and 3.3 and 21.3 for Corridor 1A. Both models (Fig. 4.46) show a conductive deep structure ($<100$ $\Omega$m). There are four relative conductive bodies in the middle and lower crust: beneath sites 137-136, 131-130, 122-146, and 151-154. Because of the effect of these conductors, the depth to the top of the conductive deep structure can be resolved only at some sites (e.g., east of site 128, Fig. 4.46). The resolution is very low below the black dashed lines (Fig. 4.46). These is sufficient resolution to show that the depth to the top of the conductive body decreases from the Fort
Figure 4.45 Stitched 1D Fischer inversion models from the Nahanni terrane to Buffalo Head terrane. Blanking area: region of no penetration.
Fig. 4.46 2D Occam inversion models. Only the data at periods >10 s were inverted in order to determine the deep structures. Resolution is low below the black dashed line.
Simpson terrane (~150 km) to Hottah terrane (~40 km), and is about 140 km beneath Buffalo Head terrane.

There are some differences between the Corridor 1 and Corridor 1A models in the region of overlap because these sites are near the edge of the inversion models. The differences do not affect interpretation.

4.9 Geological Interpretation

The analysis of MT data revealed a series of geoelectric structures along the SNORCLE Corridor 1 and 1A (Fig. 4.38, 4.39 and 4.40). In order to understand the geological structures presented by the geoelectric features, other geophysical results will be used in the geological interpretation, for example, seismic reflection, seismic refraction, teleseismic, gravity and magnetic results.

4.9.1 Liard High and Bulmer Lake Arc

The 1D and 2D MT inversions show a resistivity boundary at sites 143 and 142 (Fig. 4.22 and 4.38). The conductive surface layer at site 142 (<40 Ωm) is thinner (with 300 m thickness) than in surrounding areas.

Examination of geological literature reveals that this resistivity feature coincides with the Liard High (Fig. 4.47). Results from the well logs show the thickness of the Phanerozoic sedimentary rocks is ~600 m at site 142, which is thinner than that in surrounding area (~1000 m; Fig. 2.1). The Liard High lies above the Bulmer Lake Arc, a Precambrian paleo-topographic high, formed by folded granitic and metamorphic basement (Merjer-Drees 1975).
Figure 4.47 Structural and stratigraphic cross-section around the Liard High (from Merjer-Drees, 1975)
The basement high is overlain by the Ordovician sandstone.

The thickness of the sedimentary rocks on the Liard High indicated by MT results is smaller than that from the well logs. The Liard High is a relatively narrow structure, so the distance between MT site and well log (4.6 km) could be one reason for the difference. An other reason could be the presence of local 3D structure.

The regional gravity and magnetic data do not define the local structure of the Bulmer Lake Arc because of their limited spatial resolution. However, the gravity data show a spatial gradient in the east-west direction around sites 143 and 142; to the east the Bouguer anomaly is higher than to the west (Fig. 4.48). This gravity feature extends to the north of the survey profile.

Comparison of the MT and geological results suggests the MT data provide resolution of the shallow geological structure. The MT models show a resistive body beneath ~400 m depth at sites 143, 142. The MT response will provide poor resolution of the true resistivity because the body is beneath the surface conductor. However, the MT models constrain the resistivity of the body is to be above130 Ωm, which is consistent with the expected resistivity of granitic and metamorphic basement (Chapter 1). The relatively conductive body (~230 Ωm) extending to 20 km depth beneath sites 143 and 142 (Fig. 4.39, labelled A) may relate to the Bulmer Lake Arc. The site density of the MT survey is insufficient to resolve the structure in more detail.

The MT data resolved a conductive surface layer (<40 Ωm resistivity and <300 m thickness) at sites 140-122. On the basis of the geological results, it is interpreted to be the Ordovician sandstone forming a "crest" overlying the granitic and metamorphic basement.
Figure 4.48 (a) Bouguer gravity anomaly and (b) total magnetic field anomaly maps. Black dots: MT survey sites. The gravity and magnetic data were provided by GSC.
4.9.2 Upper Crustal Resistive Body in the Fort Simpson Terrane

MT results from Occam and NLCG inversions show a relatively resistive body with a west-dipping (~20°) base in the Fort Simpson terrane, labelled B in Fig. 4.39. The MT method is not able to resolve the resistivity of the body exactly, because it lies beneath the conductive sedimentary rocks. However, the results do constrain the resistivity to be greater than 400 Ωm. The body extends at least 100 km along the survey line. The depth to the top of the body is about 1 km. The depth to the base is about 20 km at site 141 and flattens to about 25 km beneath sites 141 to 139. The thickness of the body decreases towards the east. It is underlain by a relatively conductive body (130-250 Ωm).

The seismic reflection data shows a package of west dipping reflections beneath stations 0-2500 corresponding to the location of MT sites 142-132 (Fig. 4.49; Cook et al. 1998). These reflectors dip at about 20°-30° near station 2100 (MT site 134), but become horizontal toward the west end of the profile. The depth to the base of the package of reflectors is at least 20 km in the west. The seismic data have been interpreted to represent the Fort Simpson basin which overlies the Fort Simpson terrane (Fig. 4.49). The basin thins to the east and disappears near the boundary between the Fort Simpson and Hottah terranes (Cook et al. 1998). The thickness of the basin is about 20 km in the west. The Moho depth in this area is about 30 km in the west of the Fort Simpson terrane and decreases to the east. (Cook et al. 1998).

The seismic refraction data reveals a low velocity crust (6.15 km/s) at depth less than 20 km on the western side of the Fort Simpson terrane, where the velocity anomaly relative to the average velocity in the crust is about -0.1 km/s (Fig. 2.5, Viejo et al. 1999). Allowing
Figure 4.49 (a) Seismic reflection profile and 2D MT NLCG inversion model at Fort Simpson terrane. The figure shows seismic data (Cook et al., 1998) overlying the resistivity model. The magnetic and Bouguer gravity anomaly curves along the profile are shown in the top of figure. (b) Geological interpretation of seismic reflection data.
for a regional trend the Bouguer gravity anomaly at MT sites 142-130 is relatively high (Fig. 4.49a).

Comparison of the MT results and the seismic results shows the relatively resistive body (>400 $\Omega$m, above ~20 km depth) in the Fort Simpson terrane corresponds to the low velocity, reflective zone, which is interpreted as the Fort Simpson basin (Cook et al. 1998). The basin is believed to have formed as a result of lithospheric extension following collision of the Fort Simpson with the western Hottah terranes at ~1.84 Ga (Table 2.1; Chapter 2).

MT results support the seismic interpretation. The relatively high resistivity of the body in the MT model (>400 $\Omega$m) is consistent with the presence of the Proterozoic sedimentary rocks, consisting of sandstone, siltstone and argillite deposits on the western side of the Fort Simpson basin (Cook et al. 1992; 1998; Hills et al. 1981). A relatively high resistivity of about 1500 $\Omega$m was determined for the Proterozoic sedimentary rocks in the Lewis thrust sheet of southeastern British Columbia, where the main rocks are alluvial fan and megabreccia-landslide deposits (Gupta and Jones 1990). Interpretation of the MT model indicates the base of basin is west-dipping (about 20°). It is at about 20 km depth beneath sites 141-139 and rises to the east. This geometry is consistent with the interpretation of the seismic reflection results (Fig. 4.49). Therefore, MT results confirm the seismic interpretation and resolve the shape and depth of Fort Simpson basin. MT results show the western margin of the basin is the Bulmer Lake Arc (MT site 142), labelled A in Fig. 4.39.

The relatively conductive body (130-250 $\Omega$m resistivity) beneath the Fort Simpson basin at 30 km (labelled C in Fig. 4.39) may be interpreted to be Proterozoic mantle. As discussed in Chapter 1, the reasons for the enhanced conductivity in the mantle could be
partial melt, hydrogen (Karato 1990), and hydrous minerals (Boerner et al. 1999).

4.9.3 Crustal Fort Simpson-Hottah Conductor

MT results for both the Occam and NLCG inversion models, show a clear sub-vertical conductive zone (less than 100 $\Omega$m) beneath sites 131 and 130 (labelled D in Fig. 4.39), at the boundary between the Fort Simpson and Hottah terranes. This zone extends from several kilometres depth to $\sim 40$ km. Seismic data indicate a crustal thickness of $\sim 30$ km in this area, indicating the base of the anomaly resistivity zone could be as deep as the upper mantle. As discussed in last section, the MT data can resolve the depth to the top of the conductor, but the resolution of the depth to the base is low.

The seismic reflection data define a wedge-shaped structure at several kilometres to $\sim 25$ km depth between stations 1800 and 3200 (reflections TW in Fig. 4.50). Seismic reflection station 3000 is near MT site 131 (Fig. 4.50). The sub-vertical conductive zone is spatially correlated with reflection zone TW. Because of the limited resolution of the MT method, the precise geometry of the conductive zone is not defined by the MT results and it is represented in the model by a block shape which is different from the wedge shape in seismic reflection.

Wide-angle reflection results show an anomalous velocity zone beneath the Fort Simpson-Hottah transition, consisting of a relatively high velocity (7.0 km/s, 0.15 km/s velocity anomaly) above the Moho ($\sim 30$ km) depth (Fig. 2.5, Viejo et al. 1999). The velocity anomaly corresponds to the conductive zone and seismic reflection zone TW. Beneath the Moho, there is a relatively low velocity zone (7.7-8.0 km/s) at 30-75 km depth with a
Figure 4.50 (a) Seismic reflection profile and 2D MT Occam inversion model from Fort Simpson to Hottah terranes along Corridor 1. The figure shows seismic data (Cook et al., 1998) overlying the resistivity model. The magnetic and Bouguer gravity anomaly curves along the profile are shown on the top of figure. (b) Geological interpretation of seismic reflection data.
negative 0.2-0.5 km/s velocity anomaly (Fig. 2.5, Viejo et al. 1999).

These geophysical data all indicate the presence of an anomalous zone at the boundary between the Fort Simpson and Hottah terranes, and provide information on the geometry of the Fort Simpson-Hottah transition. This transition zone was interpreted by Cook et al (1998) as a structural boundary associated with the collision between the Fort Simpson and Hottah terranes. According to Snyder (2000), the older Hottah terrane (1.95-1.91 Ga) collided into the eastern flank of the younger Fort Simpson terrane (1.8 Ga) creating a wedged shape geometry. The upper crust of the Fort Simpson terrane was detached and thrust over the Hottah terrane. The ocean lithosphere that separated the Fort Simpson terrane from the Hottah terrane was subducted beneath the crust of the Hottah terrane. Later transpression caused the metamorphism and deformation of the Proterozoic sedimentary rocks of the Fort Simpson and Hottah terranes (and intrusion of intrusive calc-alkaline plutons; see Chapter 2).

There are four main hypotheses adopted to explain the enhanced conductivities in the middle and lower crust: saline fluids, carbon grain-boundary film, conducting metallic minerals, and partial melt (Chapter 1). Either carbon grain-boundary film or conducting minerals could be viable mechanisms for the enhanced conductivity of the Fort Simpson-Hottah transition. Deformation and metamorphism of ocean lithosphere containing a fluid phase can disperse carbon (Mareschal et al. 1992; Wannamaker, 1997), which can contribute to the higher conductivity. This process may have occurred in the subducting oceanic lithosphere beneath the Hottah terrane. High temperature and pressure could have produced graphitic schists and gneisses formed carbonate rocks, mudstone and siltstone, resulting in
the observed high conductivity (Boerner et al. 1996; Wannamaker 1997). Conductors within deformed and metamorphosed fold and thrust belts have also been observed in other areas, for example, the Penokean Orogen and Cheyenne belt (Boerner et al. 1996). Saline fluids are unlikely to be the reason for the enhanced conductivity in the SNORCLE area because the age of structure is very old (1.91 Ga). Bailey (1990) modelled fluid residence times and showed that the residence time of free aqueous fluids in deep (ductile) crust is very short.

4.9.4 Hottah Mantle Conductor

All MT models include a broad conductive zone beneath the Hottah terrane (labelled E in Fig. 4.39; labelled F1 in Fig. 4.40). The resistivity of this zone is less than 150 Ωm in the west of Hottah terrane. The zone may connect with a more conductive (~30 Ωm) zone in the crust beneath the boundary of the Hottah terrane and Great Bear Magmatic Arc (labelled F2 in Fig. 4.40). The depth to the top of this conductor decreases eastwards (from about 70 km in the west to 20 km in the east) and it has a sharp western edge at site 147.

This conductive anomaly beneath sites 126-146 does not appear to coincide with the gravity responses (Fig. 4.48). Because the depth of the top of this conductor is deeper than 25 km, a direct correlation with the magnetic response would not be expected.

The conductive zone beneath the Hottah terrane correlates to a zone of the weak and discontinuous seismic reflections below 30 km that is interpreted as Hottah mantle by Cook et al., (1998; 1999) (Fig. 4.50). Seismic refraction results show the velocity is about 7.7 km/s in the upper mantle beneath the Hottah terrane, which is lower than the adjacent areas (8.2 km/s) (Fig. 2.5; Viejo et al. 1998; 1999). Comparison of the results from seismic reflection,
refraction and MT, the data suggest a conductive upper mantle beneath the Hottah terrane.

Several hypotheses are used to explain the enhanced conductivity in the upper mantle conductors (Chapter 1). An uplift lithosphere-asthenosphere boundary can produce the high conductivity due to partial melt in the asthenosphere. Such a conductor will be associated with high heat flow. Alternatively, hydrogen dissolved in olivine has been proposed as a factor causing decreased resistivity in the mantle (Karato 1990). Carbon-bearing material in the mantle may also create enhanced conductivity (Wannamaker, 1997). In the laboratory, carbon has been observed at crustal conditions to be deposited in graphite form on mineral surfaces during fracturing (Roberts, et al. 1999). Alternatively, Boerner et al. (1999) suggested the presence of hydrous mantle minerals, specifically phlogopite, generated by metasomatising events as a possible explanation for conducting Archean mantle in southern Alberta. It is speculated that beneath the Hottah terrane, either carbon or dissolved hydrogen (or both) are the mechanisms for the enhanced conductivity (Jones et al. 2001). Measurement of heat flow by Lewis and Hyndman (2001) shows a high heat flow (109 mW/m²) across the Hottah, Fort Simpson and Nahanni terranes. Therefore, the thin lithosphere could be an alternative hypothesis to interpret the enhanced conductivity in Hottah terrane.

4.9.5 Hottah-Great Bear Crustal Conductor

The conductive upper mantle beneath the Hottah terrane connects with a more conductive zone (~30 Ωm) beneath the boundary of the Hottah terrane and Great Bear Magmatic Arc in which the depth of the top of the conductive zone is about 20 km (labelled F2 in Fig. 4.40). The conductive middle and lower crust beneath the boundary between the
Hottah terrane and Great Bear Magmatic Arc contrasts with the resistive middle and lower crust further east in Great Bear Magmatic Arc.

The Great Bear Magmatic Arc has been interpreted by Cook et al. (1999) as the product of eastward subduction of oceanic lithosphere beneath the Hottah terrane at 1.84-1.87 Ga (Table 2.1). Cook et al. (1999) suggest the Great Bear Magmatic Arc is relatively thin (~3-4.5 km) and lies above either Hottah crust or imbricated rocks of the Coronation margin. Although the origin of the rocks beneath the Great Bear Magmatic Arc is not known clearly, it is certain that they are deformed rocks of the Hottah-Slave transition (Cook et al. 1999), for example, deformed and metamorphosed Coronation Supergroup sediment layers.

Therefore, the source of the enhanced conductivity in the middle and lower crust beneath the boundary between the Hottah terrane and Great Bear Magmatic Arc could be either carbon grain-boundary film or conductive minerals formed during the deformation and metamorphism. In studies of rocks from the Kapuskasing uplift (Katsube and Mareschal 1993) and Trans-Hudson orogen (Jones and Craven 1990), petrophysical models and MT results have suggested graphite can be a source to enhance the conductivity of mid-lower crust.

Alternatively, Gupta and Jones (1990) and Jones and Ferguson (1997) discuss extensive conductivity anomalies in the crust caused by interconnected sulphide mineralization. In either case, the enhanced conductivity is caused by electronic conduction in interconnected metasediments.

There is also conductive crust beneath the central Great Bear Magmatic Arc (labelled G in Fig. 4.40) with its upper surface at around 10 km depth. Because this structure is further
from margin of the Great Bear Magmatic Arc its source is less clear than that of the conductor to the northwest. However, it can again be attributed to electronic conduction in interconnected metasediments of either Hottah terrane or Coronation Supergroup.

4.9.6 Buffalo Head Terrane

There is a clear resistivity boundary at sites 156 and 157 near the west margin of the Buffalo Head terrane (Fig. 4.40). The relatively high resistivity in the Buffalo Head terrane is associated with the metaplutonic and subordinate felsic metavolcanic rocks, forming the terrane. The lower resistivity at mid to lower crustal depths in the east of the Buffalo Head terrane may be interpreted as the westwards extension of the Taltson Magmatic Arc beneath the Buffalo Head terrane (Hoffman, 1987). Additional MT soundings in the east of the Buffalo Head terrane and Taltson Magmatic Arc are needed to examine the extension of the Taltson magmatic zone in more detail.

4.9.7 Great Slave Lake Shear Zone

MT Results

As discussed in section 4.4.2, the MT responses define clear azimuthal dependence in the area of the GSLsz. The geoelectric strike changes from ~N32°E at periods corresponding to the upper crust to N65°E at periods corresponding to lithospheric penetration with the transition occurring over the period range from 10 to 100 s (Fig. 4.10). The real induction vectors have a direction of southwest to the west of the GSLsz and a variable direction to the east of the GSLsz at 10 s period, and point to southwest in the Great
Bear Magmatic Arc and change to northwest and west in the Buffalo Head terrane at 100 s period. The variation of the induction vectors and strike direction at periods of 10-100 s imply a change in the resistivity structure across the GSLsz in the middle and lower crust.

The apparent resistivity and phase pseudosections along Corridor 1A show a transition zone between sites 154 and 157 at periods of 0.3-300 s (Fig. 4.51):

- The TE phase shows a boundary between the relatively high values at site 154 at periods 0.3-5 s (30°) and the relatively small values at adjacent sites (24°). The TE phase beneath site 155 is 18° which is lower than the adjacent sites. The TM phase is lower at site 154-156 (12°) than at sites 152-153 (18°).

- The TE apparent resistivity at site 155 is higher in the period range 10-50 s (>1000 Ωm) than at the adjacent sites (<800 Ωm). The TM apparent resistivity shows a resistive zone between sites 154 and 157 at periods 4-400 s (>1500 Ωm). The lateral extent of the resistive zone is dependent on the rotation angle. The resistive zone moves west slightly when the data examined at a rotation angle of N41°E.

The relatively low phases and high resistivity beneath sites 155 and 156, particularly in the TM response, imply a resistive zone. In order to examine the strike dependence of MT models from the upper crust to the upper mantle across the GSLsz, inversions of MT data from the different period ranges and with the different strike angle were completed (Table 4.19). The NLCG inversion models are shown in Fig. 4.52. The responses of Model C have been shown in Fig. 4.35. The responses for Models A and B are shown in Appendix F. The data fit for Model A are good for most of sites, except the TE responses at site 156 at periods >0.4 s. The TM responses of Model B fit the observed data well at every site. The TE
Figure 4.51 Apparent resistivity and phase pseudosections around GSLsz along Corridor 1A. Rotation angle used is N60°E.
apparent resistivity of Model B is higher than the observed data at sites 148, 150, 151, 154, and 156, and lower at site 152. However, the data fit to the TE phase is good at these sites.

Table 4.19 2D inversions for Corridor 1A data using different strike angles.

<table>
<thead>
<tr>
<th>No.</th>
<th>Inversion Period Range (s)</th>
<th>Rotation Angle</th>
<th>Inversion Method</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$10^{-1}$ - 10</td>
<td>N30°E</td>
<td>NLCG</td>
</tr>
<tr>
<td>2</td>
<td>30-3000</td>
<td>N60°E</td>
<td>NLCG</td>
</tr>
<tr>
<td>3</td>
<td>$10^{-1}$ - 3000</td>
<td>N41°E</td>
<td>NLCG and OCCAM</td>
</tr>
</tbody>
</table>

The inversion model obtained using short period data (<10 s) and N30°E rotation angle is designed to resolve upper to middle crustal structure (Model A, Fig. 4.52a). The resolution depth for a period of 10 s along Corridor 1A is less than 20 km (Table 4.4). The model clearly shows a west-dipping, relatively conductive zone (50-1000 Ωm) beneath sites 148 to 152, and a very resistive body (>10,000 Ωm) beneath sites 151 to 157. The inversion obtained using long period data (30-3000 s) and N60°E rotation angle is designed to resolve middle to lower crust and upper mantle structure (Model B, Fig. 4.52b). The model includes a conductive zone extending from several kilometres to 40 km depth beneath sites 149 and 150, and a very resistive body (>10,000 Ωm) beneath sites 155-158 at depths of 10-70 km. Model C in Fig. 4.52 has been discussed in last section, and also includes a conductive zone beneath sites 149 to 153 and a resistive zone beneath sites 155 and 156.

All of the models include a conductive zone extending through the crust to the upper mantle beneath sites 149 to 153, and a resistive zone extending through the crust beneath
Figure 4.52 Three NLCG inversion models along Corridor 1A. (a) inversion of data at periods <10 s with 30° strike angle, (b) inversion of data at periods >30 s with 60° strike angle, (c) inversion of data at periods 0.001-3000 s with 41° strike angle.
sites 155 and 156. The locations of the margins of the resistive body are dependent on the rotation angle. However the region beneath site 155 and 156 is resistive for all rotations. For rotations close to north-south the body extends further to the west. The presence of the resistive zone in the 2D models is supported by the apparent resistivity and phase (Fig. 4.51).

**Seismic Results**

Teleseismic data has been analysed to determine the seismic fast polarization direction of shear-waves (SKS) and anisotropy. Eaton and Asudeh (2001) give results for stations across the GSLsz located near the MT profile. The average fast polarization directions of SKS splitting at some seismic stations are shown in Fig. 4.53 and listed in Table 4.20.

Eaton and Asudeh (2001) resolve significant shear wave splitting near the GSLsz with differences in SKS arrival times from different azimuths of 0.9 to 1.4 s. The SKS data revealed a two-layer anisotropy: the upper layer has a variable fast polarization direction with an overall axis close to 43°±9°; the lower layer has a fast polarization direction of 104°±10°. The inferred splitting times of shallow and deep layer are 1.2±0.3 s and 0.8±0.5 s respectively. The splitting times for the upper layer increases as the GSLsz is approached suggesting a possible increase in the thickness of this layer (Eaton and Asudeh 2001). The interference between these two layers explains the rapid variations in splitting observed near the GSLsz. The anisotropy within the shallow layer is strongest within the shear zone with the time differences for that individual layer reaching values of 1.6 s. The anisotropy of the deeper layer increases to the southeast.
Figure 4.53 Comparison of magnetic field data, MT azimuths, and SKS fast directions. Upper panel shows MT strikes for the period band 1-20 s and the SKS direction for the shallow seismic layer. Lower panel shows the MT strikes for the period band 20-500 s and the SKS direction for the deep seismic layer.
Table 4.20 Comparison of the average fast polarization direction (φ) of shear-wave (SKS) splitting cross the GSLsz and the maximum phase split (MPS) direction of some MT sites (modified from Eaton and Asudeh 2001).

<table>
<thead>
<tr>
<th>Seismic Station</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>φ (°) Shallow</th>
<th>Deep</th>
<th>MT Site Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>MPS (°) at 1-20 s</th>
<th>MPS (°) at 20-500 s</th>
</tr>
</thead>
<tbody>
<tr>
<td>GS01</td>
<td>60.7342</td>
<td>115.5775</td>
<td>51</td>
<td>101</td>
<td>152</td>
<td>60.8139</td>
<td>115.5722</td>
<td>38</td>
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<tr>
<td>GS02</td>
<td>60.73</td>
<td>115.313</td>
<td>50</td>
<td>105</td>
<td>151</td>
<td>60.75</td>
<td>115.4917</td>
<td>10</td>
</tr>
<tr>
<td>GS03</td>
<td>60.7142</td>
<td>115.0911</td>
<td>46</td>
<td>116</td>
<td>153</td>
<td>60.7778</td>
<td>115.0639</td>
<td>36</td>
</tr>
<tr>
<td>GS05</td>
<td>60.742</td>
<td>114.812</td>
<td>32</td>
<td>86</td>
<td>154</td>
<td>60.7694</td>
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<td>42</td>
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<tr>
<td>GS09</td>
<td>60.6703</td>
<td>114.594</td>
<td>38</td>
<td>111</td>
<td>155</td>
<td>60.7111</td>
<td>114.7111</td>
<td>34</td>
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<tr>
<td>GS10</td>
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<td>38</td>
<td>98</td>
<td>156</td>
<td>60.6222</td>
<td>114.5222</td>
<td>39</td>
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<tr>
<td>GS13</td>
<td>60.278</td>
<td>114.0263</td>
<td>58</td>
<td>96</td>
<td>157</td>
<td>60.4889</td>
<td>114.3583</td>
<td>34</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>158</td>
<td>60.3667</td>
<td>114.3806</td>
<td>16</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>159</td>
<td>60.4333</td>
<td>114.0806</td>
<td>14</td>
</tr>
</tbody>
</table>

The fast polarization direction of the deep layer is oblique to the direction of absolute plate motion (225°), implying that the inferred mantle anisotropy is more likely to be a manifestation of fossil anisotropy than shear-induced flow at the base of the lithosphere. Eaton and Asudeh (2001) interpret this two-layer model as a simplified representation of strain-dependent rotation of olivine crystalline axes. The upper layer is a localized zone of higher strain beneath the fault zone, which overlies an extensive but mildly deformed layer in the lower lithosphere.

**Comparison of MT, Magnetic Results and Seismic**

The MT responses define clear azimuthal dependence in the area of the GSLsz. The geoelectric strike changes from ~N32°E at periods corresponding to the upper crust to N65°E
at periods corresponding to lithospheric penetration. The geoelectric strike in the crust is more north-south than the regional (>40 km scale) strike of the GSLsz south of Slave Province which is N60°E (Hanmer et al. 1992).

Figure 4.54 compares the magnetic anomaly with 2D MT inversion models. The anomalous low magnetic zone corresponds is spatially correlated with the resistive zone (>3000 Ωm) beneath MT sites 155 and 156. The location of magnetic high to the west of the low magnetic zone corresponds to the wide conductive zone beneath sites 154 and 153. However, the magnetic low also suggests an azimuth similar to the MT results (Fig. 4.53, 4.54).

The results suggest that the strike of the GSLsz where it crosses Corridor 1A is locally close to N30°E. Deflections of the shear zone away from the regional N60°E azimuth are also observed on the exposed portion of the GSLsz to the northwest. Hanmer (1988) attributes such deflections to late stage deformation wrapping around constrictions. The geoelectric strike at periods corresponding to signal penetration to lithospheric depths is N60°E. This orientation is parallel to the larger scale strike of the shear zone to the south of the Slave Province. The result suggests the larger scale tectonic motion associated with the fault has been recorded by the resistivity structure of the mantle lithosphere.

Table 4.20 and Figure 4.53 compare the seismic results with the maximum phase split direction for period ranges of 1-20 s and 20-500 s at several MT sites. The period range 1-20 s and 20-500 s correspond the exploration depth into the upper-middle crust (<25 km), and, lower crust and upper mantle (25-100 km) respectively (Fig. 4.9).

As shown in Fig. 4.53 there is excellent agreement between the azimuth of the MT
Figure 4.54 Comparison of magnetic anomaly and 2D MT inversion models around Great Slave Lake shear zone. (a) total magnetic field anomaly, (b) Occam inversion model and (c) NLCG inversion model. The orientation of the model is N49°W. The location shown in figure is from the geological map (Fig. 4.1). Other studies (e.g. Eaton and Asudeh 2001) equate the location of the GSLsz with the magnetic low.
strike at periods 1-20 s and the fast SKS direction in the shallow layer within the GSLsz. Outside the shear zone the MT strike tends to have a more north-south orientation than the SKS azimuth. Both the MT and seismic inversions suggest an increase in the degree of heterogeneity within the shear zone. The results suggest a common source for the MT and SKS directions within the shear zone and, because 20 s period corresponds to MT signal penetration to within the crust, they allow the seismic anisotropy to be attributed to the crustal layer.

The MT strike for the period range 20 to 500 s is ~N60°E and is oblique to the SKS fast direction of N104°E. As the absolute plate-motion direction is 225° Eaton & Asudeh (2001) attribute the response of the deeper layer to represent frozen anisotropy within the mantle lithosphere rather than to modern anisotropy in the asthenosphere. In the present analysis the MT response was modelled using a 2D structural model rather than a model containing anisotropic conductivity and the MT responses provided support for an increase in lithospheric mantle conductivity to the northwest of the GSLsz. However, it is possible that the MT model may include a component of inherent mantle anisotropy and this aspect will be examined in future studies.

At sites near the Grenville Front in eastern Canada, an average 23° obliquity between SKS and MT responses has been interpreted as a kinematic indicator of dextral shearing. The seismic and MT anisotropy are interpreted to be caused by lattice-preferred and shape-preferred orientation of mantle minerals respectively (Ji et al. 1996). The relative orientation of the deeper MT and seismic responses near the GSLsz has the same sense as at the Grenville Front, although the obliquity near the GSLsz is larger, and is consistent with
dextral shearing. Eaton and Asudeh (2001) interpret the fast SKS direction as representing lattice-preferred orientation of mantle minerals in the direction of the maximum compressive palaeostress associated with the continental collision that produced the GSLsz. The strike of the MT responses near the GSLsz is interpreted to indicate the direction of large-scale shear. In contrast to the interpretation of Ji et al. (1996), in this study the MT responses are modelled in terms of macroscopic structures rather than shape preferred orientation of mantle minerals.

Interpretations

Two-dimensional MT modelling results reveal a sub-vertical resistive zone (>3000 Ωm) in the upper-lower crust beneath sites 155-156, which is coincident with the magnetic anomaly low (Fig. 4.54). The resistive zone is interpreted to represent the electrical signature of the GSLsz. Interpretation of the resistive response of the GSLsz requires more detailed consideration of the geology of the shear zone.

As shown by Hanmer et al. (1992), the GSLsz exposed to the northwest is a bundles of upright belts of mylonites. The oldest mylonites are of granulite facies and formed a belt with a width of >10 km, possibly up to 25 km. The metamorphic grade of the mylonites decreases with decreasing age with the decreasing grade reflecting both a decrease in temperature and pressure (Hanmer 1988). The subsequent strain produced progressively narrowing belts of upper and lower amphibolite grade mylonite and lower greenschist-facies chlorite-bearing mylonites. The late stage of deformation evolved through the ductile-brittle transition and the latest stages involved dilational faulting and development of quartz
stockworks (Hanmer 1988; Hanmer et al. 1992). The mylonites are interpreted to have formed from a mixed protolith of hornblende-biotite, magnetite-bearing, granite and granodiorite that was intruded both pre- and syntectonically.

The relatively high resistivity within the GSLsz suggests that all of the mylonite units are have high electrical resistivity. Previous electromagnetic studies have suggested that hydrous chlorite-bearing rocks will have moderately enhanced conductivity (e.g. Lee et al. 1983) but in a later detailed study Olhoeft (1981) found that structural water provided little contribution to the electrical conductivity. Within the GSLsz on the exposed shield greenschist facies mylonite occurs in a network of belts up to 2 km wide and 70 km long. Given the probability that greenschist mylonites occur in the study area, the high resistivity that is observed supports Olhoeft’s (1981) conclusions. However, it is possible that the effective electrical conductivity of the lower grade mylonite belts is reduced by the quartz “stockwork” which consists of vertical veins up to 25 m wide and up to 40 km long that cut the mylonite structures. Further electromagnetic soundings on both the exposed shield and with greater site density across the GSLsz are needed to fully characterize the electrical character of the different mylonites.

The Tintina Fault in northwestern Canada also is defined as a resistive fault by the MT data (Ledo et al. 2001). The MT study also showed the San Andreas at Carrizo Plain is resistive at mid-crustal depth (Mackie et al. 1997a). These results show major strike slip faults, like the Tintina and San Andreas faults, may have resistive characterise. However, the MT surveys also defined some conductive faults, for example, the Alpine fault in New Zealand for which the conductive zone is related to the hot deep fluids associated to the
tectonic uplift of the southern Alps (Ingham and Brown, 1998), the Denali fault system in Alaska for which the conductive rocks are caused by carbon and/or metallic mineral films associated to largely tectonically emplaced (Stanley et al. 1990).

The average geoelectric strike near the GSLsz, determined at periods corresponding to signal penetration to crustal depths, is N32°E. This orientation is more north-south than the regional (>40 km scale) strike of the part of the GSLsz south of Slave Province which is N60°E (Hanmer et al. 1992). The western boundary of the magnetic low where it crosses the profile and the eastern boundary to the south of the profile also have strike angles that are closer to north south. The MT and magnetic results suggest that the strike of the GSLsz where it crosses Corridor 1a is locally close to N30°E. Deflections of the shear zone away from the N60°E azimuth are also observed on the exposed portion to the northwest and Hanmer (1988) attributes such deflections to later stage deformation wrapping around constrictions.

The similarity of the MT and seismic response directions for the shallow layer provides evidence that the controls on these responses are related. The present study suggests a large component of the MT response can be explained by structural features rather than small-scale material anisotropy. These results do not exclude the possibility of material anisotropy near the GSLsz, but it does suggest that structural control may be a more important component of the MT response in the upper-mid crust. The geoelectric strike direction of GSLsz depends on the geological structure strike (Fig. 4.1). The similarity of the MT and seismic azimuths suggests that large scale structure may also contribute to the seismic anisotropy.
4.9.8 Lithosphere

The MT responses (Fig. 4.19 and 4.20), 1D inversion models (Fig. 4.45) and 2D inversion models (Fig. 4.46) all show a region at enhanced conductivity at a maximum depth of 100-200 km across the mantle in the Nahanni, Fort Simpson, Hottah, Great Bear Magmatic Arc and Buffalo Head terranes. Because of the effect of the conductive Phanerozoic sedimentary rocks and crustal conductors, the depth to the top of the conductive deep structure and its resistivity can not be resolved at all points along the profile (Fig. 4.46).

The depth to a 150 Ωm resistivity occurs at ~100±20 km depth beneath the Nahanni, Fort Simpson, central Hottah, central Great Bear and Buffalo Head terranes, the depth to this value decreases to ~40 km in the Hottah terrane and the eastern Great Bear Magmatic Arc.

This depth is interpreted to be the base of the lithosphere. High heat flow (109 mW/m²) have been measured across the Hottah, Fort Simpson and Nahanni terranes by Lewis and Hyndman (2001) and along the Presqu’ile Barrier to the west of the study area (130-200 mW/m²) by Majorowicz et al. (1989). Although there is probably an influence from local topographic features and region fluid flow on these results, the values can be plausibly explained by a relatively shallow lithosphere-asthenosphere boundary. The source of the higher conductivity would be partial melt in the asthenosphere.

The enhanced conductivity at the mantle depths beneath the boundary of the Hottah Terrane and Great Bear Magmatic Arc and beneath the eastern Great Bear Magmatic Arc is more localized and is interpreted to represent increased conductivity within the lithosphere rather than a decrease in the depth to the asthenosphere. Seismic reflection results show delamination structures extending to 100 km depth in the mantle beneath the Great Bear
Magmatic Arc providing evidence that lower crustal rocks were emplaced in the mantle during subduction (Cook et al. 1999). This subduction was associated with collision and accretion of the Hottah terrane to the western margin of the Slave craton at 1.9-1.88 Ga. The source of the enhanced conductivity in the mantle beneath the western Great Bear Magmatic Arc may be either hydrogen or carbon introduced into the mantle through the subduction process.

At sites near the GSLsz the geoelectric strike at periods corresponding to signal penetration to lithospheric depths is N60°E, which is parallel to the larger scale strike of the shear zone in the area to the south of the Slave Province. The result suggests the larger scale tectonic motion associated with the fault has been recorded by the resistivity structure of the mantle lithosphere. In this study the response was modelled using a 2D structural model rather than a model containing anisotropic conductivity. However, it is possible that the azimuthally dependent MT responses could be due to inherent mantle anisotropy with an azimuth of N60°E.

There is a significant boundary in the mantle conductivity at the GSLsz with the conductivity being higher to the northwest of the boundary. The exact geometry of this boundary is fairly poorly constrained by the MT data because it is occurring at the longest periods available in the response and is obscured in part by the overlying crustal structures. Nevertheless the existence of the conductive mantle to the northwest of the GSLsz is supported by the observed high phases and low apparent resistivity at periods exceeding 10^3 s at sites northwest of 154 (Fig. 4.53). The truncation of the mantle conductor beneath the eastern Great Bear Magmatic Arc at the GSLsz suggests the fault involved significant strike-slip movement of the mantle lithosphere as well as the crust.
Chapter 5 Time-domain Electromagnetic Study of Lac du Bonnet Batholith

Many geophysical surveys have been carried out to examine the rock and groundwater properties in the Lac du Bonnet Batholith in order to investigate the concept of safe disposal of nuclear fuel waste, for example aeromagnetics, gravity, high resolution seismic reflection, GPR, EM and geophysical borehole logging.

In the granitic rock of Lac du Bonnet Batholith, small subvertical fractures are common above 200 m depth and rare below 200 m where they are limited to the margins of low-dipping fracture zones (Brown et al. 1989). The fractures are subhorizontal beneath the major low-dipping fracture zones. This fracture distribution controls groundwater movement and chemistry.

Groundwater salinity in the Lac du Bonnet Batholith generally increases with depth. Groundwater in the Lac du Bonnet Batholith is comparatively fresh above 250 m because the near-surface sub-vertical fractures provide pathways for (fresh) meteoric water to penetrate the granite (Everitt et al. 1996). Beyond 250 m depth, groundwater salinity increases with depth because of the dissolution of soluble salts in the host rock (Gascoyne et al. 1994). However, the depth of the fresh-saline water interface is spatially variable (Stevenson et al. 1996).

Three TEM studies have been performed to investigate the fresh/saline-water
interface in the Lac du Bonnet area in a joint project with AECL. Two feasibility studies were completed in the summer and fall of 1996, using 10 m, 20 m and 40 m transmitter loops. They were used to examine the ability of TEM method to identify the fresh/saline water interface (Wu et al. 1996; 1997). A full survey was completed at four selected locations around Lac du Bonnet in the summer of 1997, using a 100 m transmitter loop (Maris et al. 1997; Maris 2000). The data were collected with 0-280 m offset range and 20 m spacing on four sides of the transmitter loop. The central loop soundings at the four sites had been analysed by Maris et al. (1997). This chapter will focus on the analysis and interpretation of offset data from this survey.

5.1 Survey Locations

The feasibility studies included data collection at three sites: Site 1, 2 and 3 (Fig. 5.1b). For the full survey, data were collected at four sites (Site A, B, D and URL, Fig. 5.1a). Table 5.1 lists the transmitter loop size, receiver offset and survey time at each survey site.

The feasibility studies used 10-40 m transmitter loops and a horizontal coil to measure the vertical component of magnetic field changes. At each site, the responses were measured at stations on lines oriented north, south, east and west of centre of the loop. The offset range is 0-45 m. The station interval on the lines was 5 m. Site 2 is located beneath power lines and was used only to analyse noise. The data were collected with help from I.J. Ferguson, V. Maris and I. Shiozaki in August and October, 1996 (Wu et al. 1996; 1997).
Figure 5.1 (a) Location of AECL permit areas in the Lac du Bonnet area in which TEM surveys were carried out (Sites A, B, C and URL) (Maris et al., 1997), (b) TEM survey sites in the URL lease area.
Table 5.1 PROTEM47 surveys in the Lac du Bonnet area, Manitoba.

<table>
<thead>
<tr>
<th>Study</th>
<th>Site</th>
<th>Transmitter Loop Size (m)</th>
<th>Receiver Offset (m)</th>
<th>Survey Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>1st Feasibility Test</td>
<td>Site 1</td>
<td>10</td>
<td>0 - 45</td>
<td>Aug. 5, 1996</td>
</tr>
<tr>
<td></td>
<td>Site 2</td>
<td>20, 40</td>
<td>0 - 45</td>
<td>Aug. 15, 1996</td>
</tr>
<tr>
<td></td>
<td>Noise test only</td>
<td></td>
<td>–</td>
<td>Aug. 5, 1996</td>
</tr>
<tr>
<td>2nd Feasibility Test</td>
<td>Site 3</td>
<td>40</td>
<td>0 - 46</td>
<td>Oct. 24, 1996</td>
</tr>
<tr>
<td>1997 Full Survey</td>
<td>Site URL</td>
<td>100</td>
<td>0 - 280</td>
<td>June 9-July 4, 1997</td>
</tr>
<tr>
<td></td>
<td>Site D</td>
<td>100</td>
<td>0 - 280</td>
<td>July 7-Aug. 7, 1997</td>
</tr>
<tr>
<td></td>
<td>Site A</td>
<td>100</td>
<td>0 - 280</td>
<td>Aug. 11-13, 1997</td>
</tr>
<tr>
<td></td>
<td>Site B</td>
<td>100</td>
<td>0 - 280</td>
<td>Aug. 19-20, 1997</td>
</tr>
</tbody>
</table>

The full survey used a 100 m transmitter loop and a horizontal receiver coil (vertical dipole). Data were collected by V. Maris and P. Street in the summer of 1997 (Maris et al. 1997; Maris 2000). The offset range is 0-280 m, which is limited by the length of the available reference cable connecting the transmitter and receiver. The station interval on the lines was 20 m.

The criteria employed in choosing the full survey sites were: (1) the presence of nearby boreholes with available geological and geophysical logs that can provide additional information about the interface between fresh and saline pore fluid; (2) the presence of large outcrops and thin conductive overburden to permit set up of the transmitter loop and receiver coil and increase exploration depth; (3) the distance from sources of cultural noise such as power lines.

Figure 5.2 shows the layout of the survey and the ground surface on which the
Figure 5.2 Stations and surface features at (a) Site A, (b) Site B, (c) Site D, (d) Site URL. This figure shows the layout of the survey and the ground surface on which the receiver coil was placed (Maris et. al., 1997).
receiver coil was placed at each site of the full survey (Maris, et al. 1997). The transmitter loops were mostly placed on exposed granitic rock; some of the receiver stations outside the loop extended into thin soil covered areas. V. Maris documented field records of survey conditions at each site (Maris, et al. 1997).

Site A was one of the more difficult surveys to position the loop (Fig. 5.2a). The transmitter loop crosses the access road in the final survey position. The south line terminates at a gravel provincial road and power line 170 m from the transmitter loop centre. A long and linear valley, which is probably fracture-related, parallels the east lines. A barbed-wire fence intersected the north line near the 80 m station. The west line ended at the 240 m offset at a pasture and a second barbed-wire fence. The ground surface was wet because of the intermittent rainfall during most days of the survey and heavy rainfall at night. The outcrop is highly fractured at this site and is covered with thick vegetation cover.

Site B has a large, open outcrop with small, scattered patches of thin overburden. It is bordered on the east and north side by 7 m cliffs (Fig. 5.2b). The transmitter loop was laid on the outcrop, with the east side at the edge of the cliff. The east survey line extended over the cliff and into an adjacent, densely forested valley. The north line stopped at a gravel roadway after crossing a cliff with a rubble base. An additional diagonal survey line was added to take advantage of a long, clear outcrop.

The transmitter loop at Site D was laid on a large outcrop, covered with moss and thin vegetation (Fig. 5.2c). This site was the most completely surveyed, with only one of the regularly-spaced stations missed. The north line progressed on uneven outcrop, with occasional bordering valleys. The west line continued over a sharp 2 m cliff near the 200 m
station. The east line descended off the outcrop near 100 m and into an area of thick forest.

At Site URL, the transmitter loop was laid on an open, hilly outcrop with patches of vegetation (Fig. 5.2d). The final station on the east line and the 140 m station on the south line were omitted because the ground was too swampy to set up the receiver. Two data transmission lines to a groundwater pressure transducer in borehole M-6 crossed the west line, near the 220 m and 240 m stations.

**5.2 Instrument and Configuration**

The instrument used in the Lac du Bonnet TEM study was a GEONICS PROTEM47 system belonging to the University of Manitoba. The recordings are voltages (in mV). Table 5.2 lists the time-gates of data collection, which is the UH frequency set of time-gates. The PROTEM47 instrument provides both short ($2^8$) and long ($2^{10}$) stacking times in order to reduce the effect of environmental noise and improve the SNR.

The survey configurations used in the feasibility and full survey are listed in Table 5.3. During the surveys, both short ($2^8$) and long ($2^{10}$) stacking times were used. In order to provide maximum signal strength while avoiding saturation, a variable instrument gain was used at each station (GEONICS 1992a). There are 8 gains for PROTEM 47. A gain of 6 or 7 was used in the feasibility studies. Gains of 4 and 6 were usually used for small offsets in full survey, and values increased to a maximum of 7 as the offset distance increased. In order to maximize signal strength, GEONICS LIMITED recommends that recordings not be made using the maximum instrument gain setting of 8 (Gil Levy, GEONICS LIMITED, pers. comm. 1997). At each station, the measurements were taken three or four times. The average
of these measurements can reduce the effect of environmental noise and increase the SNR.

### Table 5.2 PROTEM 47 UH gate times

<table>
<thead>
<tr>
<th>Gate</th>
<th>Time (μs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>6.85</td>
</tr>
<tr>
<td>2</td>
<td>8.95</td>
</tr>
<tr>
<td>3</td>
<td>12.08</td>
</tr>
<tr>
<td>4</td>
<td>15.72</td>
</tr>
<tr>
<td>5</td>
<td>20.05</td>
</tr>
<tr>
<td>6</td>
<td>26.17</td>
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<tr>
<td>7</td>
<td>33.45</td>
</tr>
<tr>
<td>8</td>
<td>42.1</td>
</tr>
<tr>
<td>9</td>
<td>54.1</td>
</tr>
<tr>
<td>10</td>
<td>68.2</td>
</tr>
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<td>11</td>
<td>83.8</td>
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<tr>
<td>12</td>
<td>104.6</td>
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<tr>
<td>13</td>
<td>135.6</td>
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<td>14</td>
<td>172.3</td>
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<td>214.9</td>
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<td>16</td>
<td>275</td>
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<td>17</td>
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<td>436</td>
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<td>19</td>
<td>555</td>
</tr>
<tr>
<td>20</td>
<td>701</td>
</tr>
</tbody>
</table>

### 5.3 Data Processing

The aim of TEM data processing is to increase the signal noise ratio (SNR) and convert the recorded data to a form that can be used for modelling and inversion, such as $dB/dt$ or late time apparent resistivity responses.
Table 5.3 Survey configurations of PROTEM 47 used in Lac du Bonnet study (GEONICS 1992a).  

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Feasibility Study</th>
<th>Full Survey</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transmitter (single-turn rectangular loop) (m)</td>
<td>10×10</td>
<td>20×20</td>
</tr>
<tr>
<td>Transmitter Current (A)</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>Turn-off Time (µs)¹</td>
<td>0.8</td>
<td>1.4</td>
</tr>
<tr>
<td>Diameter of Receiver Coil (m)</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Receiver Coil Orientation</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Frequency/Time Gates (µs)</td>
<td>6.85-701</td>
<td></td>
</tr>
<tr>
<td>Averaging</td>
<td>2^6 and 2^10 stacking times</td>
<td></td>
</tr>
<tr>
<td>Instrument Gain</td>
<td>6-7</td>
<td>6-7</td>
</tr>
</tbody>
</table>

¹The turn-off times used were the values recommended by GEONICS (GEONICS 1992a).

Data Averaging and Noise Analysis

The field recordings were first downloaded from the PROTEM47 using GEONICS software (GEONICS 1992b), then the data from repeated measurements at each station were averaged. During this process, bad data points (e.g., caused by spherics) were removed and the recordings were normalized to the same gain.

The environmental noise was recorded by taking a number of recordings with the transmitter turned off which omits any signal and noise from the transmitter. The average magnitude of the noise for each site are shown in Fig. 5.3. For a random noise source, the noise level should decrease in proportion to the increase in the width of the time windows; therefore the decrease in noise level reflects the increasing length of the time gates.

The noise levels measured in the 1996 surveys were compared with the noise level near the URL from a 1993 survey (I. Ferguson, pers. comm. 1997) (Fig. 5.3a). The 1993 survey was located approximately 400 m north of the main buildings. The noise levels from Site 2
Figure 5.3 Average background noise levels at (a) Sites 1, 2, 3 and 1993 survey; (b) Site A, B, D and URL. The average noise level was calculated by averaging the absolute value of the noise at each time-gate.
of 1996 survey are similar to those near the URL. These two sites are beneath or close to small powerlines, contributing to the high noise level. The low noise levels at Site 1 and Site 3 suggest an absence of such local noise source. There was no local thunderstorm activity during the feasibility surveys.

For the full survey, the background noise levels were measured at the start of the day and at the end of each line. Figure 5.3b shows the average background noise levels at Site A, B, D and URL. The profiles display a steady decrease from a maximum of 0.8 mV for the first gate to a minimum of 0.01 mV for the last time gate.

The noise levels at the four sites were similar. The small spread between the profiles is less than half a decade (or a factor of 3) wide. The spread between the curves increases with increasing time. Site A has the noisiest first ten gates, and Site D has the noisiest last ten gates. Site URL has the lowest average background noise levels. These differences indicate the different environmental noise level at each site.

**Response Normalization**

The recorded voltage \( V \) (mV) from the PROTEM47 can be converted to \( dB/dt \) (nT-s\(^{-1}\)) using the following formula

\[
\frac{dB}{dt} = \frac{19200V}{A_R N_R 2^n}
\]

(5.1)

where \( A_R \) (m\(^2\)) is the receiver coil area, \( N_R \) is the number of turn of the receiver coil, and \( 2^n \) is the instrument gain used (GEONICS 1992a, b).

Equations (3.87) and (3.88) can be used to calculate the early-time and late-time
apparent resistivity, that commonly is used to describe the geoelectric characteristics of the earth. Prior to calculation of the apparent resistivity, the voltage is normalized by the moment of transmitter loop and the effective area of the receiver coil (Chapter 3).

Investigations using boreholes and hydrological methods (Stevens et al. 1995; Soonawala et al. 1992; Frape et al. 1984) have shown that the granite containing resistive fresh water is at least 150 m thick in the Lac du Bonnet Batholith. When a typical resistivity value of granite saturated with fresh water (10,000 Ωm) is used to calculate the induction number β, it is evident that the recordings in our survey are all located at late-time (Fig. 5.4). Therefore, in the following data processing and analysis, late-time responses will be used to examine the TEM responses (Eq. 3.88).

5.4 Forward Modelling and Inversion

Computer programs used for forward modelling and inversion of TEM responses include RECTAN (GEONICS 1992b), EINVERT4 (Sandberg 1990) and TEMIX (Interpex Limited 1996). RECTAN is a 1D forward modelling program. It uses the digital filter method to calculate the vertical magnetic field derivative and the late-time apparent resistivity. RECTAN can calculate the central loop and offset responses of layered models.

EINVRT4 is a 1D inversion program in which the forward modelling is based on the same algorithm as RECTAN. The inversion minimizes the squared misfit (Eq. 3.62; least squares normalized by the magnitude of data). EINVRT4 can process only central loop responses.

TEMIX is used to process both central loop and offset responses. It calculates TEM
Figure 5.4 The variation of induction number (β) with offset and time. The resistivity used to calculate the skin depth is 10,000 ohm.m. The shaded area defines 0-280 m offsets (x-coordinate) and 6.85-701 ms times (y-coordinate) used in the survey.
magnetic fields using evaluations of Hankel transforms by the linear digital filter method (Anderson 1979). The program provides two forward modelling methods to calculate the Hankel transforms. The first one uses a pre-splined kernel function that gives reasonable accuracy. The second one calculates the original kernel functions and provides more accurate results (Interpex Limited 1996). The calculation time for the original kernel function is 3 times that of the pre-spline kernel function. The inversion uses a ridge regression method (Inman 1975). The fitting criterion is a least square misfit between the observed and model responses (eq. 3.60).

The choice of a good initial model is critical to the success of TEMIX inversions. If an unreasonable initial model is set up, the inversion will not converge, or will converge to an unacceptable model. Other researchers have noted this limitation when using TEMIX to invert data from resistive Precambrian terranes (Gil Levy, pers. comm. 1998).

TEMIX includes an Occam inversion which can produce a smooth model (Constable et al. 1987). The Occam model should not necessarily be regarded as the true resistivity structure because it is a smoothed model containing only those structures actually required by the data (e.g., Constable et al. 1987). The true geological structure may have greater roughness. However, the Occam model can be used as a starting model for other inversions.

TEMIX also can produce a parameter resolution matrix (Interpex Limited 1996), which can indicate whether the parameters of the inversion model can be independently predicted, or resolved (Menke 1984). If the resolution matrix is the identity matrix (1.0 along the diagonal and 0.0 elsewhere), then each model parameter is uniquely determined. Otherwise, the resolved parameters are logarithmic combinations of the model parameters (Interpex
For thin conductive layers at depth, the individual resistivity and thickness cannot be resolved. In this case the parameter resolution matrix elements for the individual thickness and the individual resistivity are close to ±0.5. The resolution matrix element corresponding to the cross terms between thickness and resistivity should be close to ±0.5 (Interpex Limited 1996).

Because the solution to an inversion problem is seldom unique, inversion modelling allows the existence of equivalent models that would fit the data equally well. TEMIX uses a specified error limit in determining the equivalent models. This parameter should indicate how much noise is acceptable in the model. If the best-fit error is $E_{\text{best}}$ and the error limit is 10 percent, then inversion models with error less than $E_{\text{best}} \times 1.1$ will be considered as equivalent models. Within the error limit, TEMIX tests the parameter changes indicated by the resolution matrix, and find the approximate bounds of layer thicknesses, resistivities and depths (Interpex Limited 1996).

5.5 Data Analysis and Interpretation: Feasibility Studies

5.5.1 Signal Responses

Before forward modelling and inversion, it is necessary to analyse the signal responses and compare them with the noise level so that the data containing the real geological information can be separated and inverted. Figure 5.5 shows the signal and noise levels at Site 1 within the URL. The signal levels for responses measured with the 10 m transmitter loop approach the noise level at times exceeding $3 \times 10^{-5}$ s (Fig. 5.5a). The SNR is very low.
Figure 5.5. TEM sounding at Site 1. (a) 10 m transmitter loop; (b) 40 m transmitter loop. Offset N14 means the receiver is located 14 m north of the transmitter center. The average noise level at Site 1 is included for comparison.
The signal responses vary erratically between nearby receiver positions, but are consistent for repeated measurements at each receiver position. The results for the 20 m transmitter loop are more consistent than for the 10 m loop but still show the erratic variation and low SNR.

For the 40 m transmitter loop, signal levels at offsets less than 25 m are consistent and higher than the noise level at times earlier than about $2 \times 10^{-4}$ s (Fig. 5.5b). The signal levels at 45 m offset approaches the noise level at times exceeding $3 \times 10^{-5}$ s. The offset recordings deviate from the central-loop result at late times, and as the offset increases, the deviation from the central loop result begins at earlier times. The results suggest that the actual SNR is low even though the signal is well above the measured environmental noise level. It is possible that there is additional noise associated with the transmitter.

At Site 3, the responses at all offsets are relatively consistent at times earlier than about $4 \times 10^{-4}$ s (Fig. 5.6a). At offsets greater than 15 m, a sign reversal occurs at $-3 \times 10^{-5}$ s. The signal levels are higher than the noise level, indicating that the recorded signals are not significantly affected by the environmental noise. However, the responses are relatively erratic at late time suggesting an additional source of noise. The signal levels at Site 3 are higher than Site 1 at early time (Fig. 5.6b), indicating a more conductive response.

Figure 5.7 shows the dependance of theoretical TEM responses on the shallow resistivity. If the shallow structure is more conductive, the voltage response is higher and the apparent resistivity response is lower. This characteristic is most obvious for early-time responses and decreases with increasing time: the effect of the shallow conductive layer is smaller for late-time responses. Although the TEM data has low SNR, it suggests a corresponding response is observed at Site 1 and 3 (Fig. 5.6b). The higher early-time
Figure 5.6. (a) TEM sounding at Site 3. Results for the 40 m transmitter loop showing the normalized voltage responses at some offsets to the north of the loop. The average noise level is included for comparison. (b) Comparison of TEM sounding at Site 1 and Site 3. For both sites the receiver is located in the centre of the 40 m transmitter loop. Note that the response at Site 3 is larger at early time.
Figure 5.7 Dependent of responses on the variation of shallow resistivity ($\rho_1$).
(a) Normalized voltage. (b) Apparent resistivity for central loop sounding for 40 m transmitter loop. The program RECTAN was used to generate the responses.
response (voltage) at Site 3 may be a result of the effect of both the conductive overburden and the near-surface fractures.

5.5.2 Inversion Results

As indicated by the erratic nature and low SNR of the responses from the 10×10 m and 20×20 m transmitter loop at Site 1 (Fig. 5.5a), the true earth responses have been masked by noise. Determination of the conductivity structure at Site 1 was therefore based on the 40 m transmitter loop responses.

Forward modelling was used to provide the basic geoelectric model for inversion. The initial results from RECTAN indicated a three-layer structure, consisting of a relatively conductive surface layer, a resistive middle layer, and a more conductive lower layer. This model was used as the starting model for the inversions.

The inversion was only based on time gates 2 to 18 (i.e. 8.95 μs to 436 μs) of the central loop sounding. The omission of the late time gates was due to the lower SNR (Fig. 5.5b). The first time gate was omitted because the response in this gate varied erratically between closely spaced receiver positions. Although the recorded voltage for this time gate was lower than the saturation limit for the instrument, it was much higher than all the other time-gates. One reason for this response may be effects associated with ramp-off time. However, the ramp off time set on the transmitter loop in the survey was the value specified by GEONICS (GEONICS, 1992a,b).

The ramp-off time is set in order to accomplish the synchronization between the transmitter and receiver with the reference cable (GEONICS, 1992a, b). The runon correction
is included in the modelling and inversion of some programs in order to correct influence of previous pulses in the wave train on the decay curve (TEMIX, 1996). EINVERT does not include these parameters during inverting the data. However, the effect of ramp-off time and runon correction for the resistivity models were examined using TEMIX. The results shows these parameters can be neglected in the inversion of TEM data.

An initial unconstrained inversion for the central loop sounding using the EINVERT4 inversion program was completed (Fig. 5.8a). The $L_1$ and $L_r$ misfits are 0.02558 and 0.09354 respectively. The fit of the response to the observation is very poor at early-time but better at late time. The response provides a very poor fit at times less than 20 μs, where the TEM response appears to be significantly affected by noise.

Because the main aim of this study is to map the depth to the conductive saline-water interface, an additional investigation of the resolution of the depth to the conductor was done. Based on the unconstrained inversions and consideration of the geological structure, the parameters of the top layer were fixed: 8 m thickness and 200 Ωm resistivity. In subsequent inversions, the depth to the conductive layer was constrained to progressively higher values and the data inverted for the conductivity of layers 2 and 3. The minimum misfit for each inversion model was examined in order to determine the range of allowable models (i.e., those models which could provided a reasonable fit to the data). The results of the inversions are listed in Table 5.4 and the variation of misfit ($L_2$, $L_r$) with the depth are shown in Fig. 5.9.

The $L_r$ misfit shows significant curvature with a clear minimum around 185 m depth (Fig. 5.9b). This misfit increases significantly at depths less than 170 m and greater than 200 m. This indicates the misfit is sensitive to variation of depth of the interface and the variance
Figure 5.8. (a) Result of unconstrained inversion, with 0.02558 L2 misfit and 0.09354 L1 misfit. (b) Result of constrained inversion, with 0.0234 L2 misfit and 0.085 L1 misfit for central loop response of 40 m transmitter at Site 1. In the inversion, the first and last two gates were omitted.
Figure 5.9 Minimum misfit of constrained inversions versus the depth to the conductive layer. (a) L2 misfit as defined by eq. (3.44), (b) L1 misfit as defined by eq. (3.46)
of this model parameter is relatively low (Menke 1984). The $L_2$ misfit shows a broad misfit curve with a minimum value around 193 m depth (Fig. 5.9a). The misfit increases significantly at depths less than 180 m and greater than 260 m. The resolution of this inversion is, therefore, limited to about ±40 m. Figure 5.8b shows the inversion results for the case for which the $L_1$ and $L_2$ misfit are 0.089 and 0.0234 respectively.

<table>
<thead>
<tr>
<th>Depth D (m)</th>
<th>Thickness $h_2$ (m)</th>
<th>Conductivity $\sigma_2$ (S.m$^{-1}$)</th>
<th>Conductivity $\sigma_3$ (S.m$^{-1}$)</th>
<th>Misfit $L_1$</th>
<th>Misfit $L_2$</th>
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<tbody>
<tr>
<td>108</td>
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<tr>
<td>133</td>
<td>125</td>
<td>$1.3 \times 10^5$</td>
<td>$1.7 \times 10^3$</td>
<td>0.0455</td>
<td>0.135</td>
</tr>
<tr>
<td>158</td>
<td>150</td>
<td>$5.2 \times 10^6$</td>
<td>$2.5 \times 10^3$</td>
<td>0.0328</td>
<td>0.102</td>
</tr>
<tr>
<td>173</td>
<td>165</td>
<td>$2.1 \times 10^5$</td>
<td>$2.9 \times 10^3$</td>
<td>0.0287</td>
<td>0.089</td>
</tr>
<tr>
<td>183</td>
<td>175</td>
<td>$1.1 \times 10^5$</td>
<td>$3.4 \times 10^3$</td>
<td>0.0255</td>
<td>0.087</td>
</tr>
<tr>
<td>188</td>
<td>180</td>
<td>$1.2 \times 10^3$</td>
<td>$3.5 \times 10^3$</td>
<td>0.0240</td>
<td>0.085</td>
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<tr>
<td>193</td>
<td>185</td>
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<td>0.0234</td>
<td>0.089</td>
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<td>$5.0 \times 10^3$</td>
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<td>0.097</td>
</tr>
<tr>
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<td>$5.0 \times 10^3$</td>
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<td>0.110</td>
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<tr>
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<td>$7.9 \times 10^3$</td>
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<td>0.133</td>
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<tr>
<td>283</td>
<td>275</td>
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<td>$5.0 \times 10^3$</td>
<td>0.0546</td>
<td>0.180</td>
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<tr>
<td>308</td>
<td>300</td>
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<td>$1.4 \times 10^2$</td>
<td>0.0858</td>
<td>0.246</td>
</tr>
</tbody>
</table>

The inversion results along with additional forward modelling suggest that the first conductive layer must be less than 10 m thickness and have a resistivity of more than 200 Ωm, otherwise the misfit increases. The resolution of the second layer resistivity is very low, but the resistivity is very high, greater than $10^4$ Ωm. The resistivity of the bottom layer is constrained to be between 150-300 Ωm. The constrained inversions (Table 5.4) suggest the
depth to the conductive layer is 200±40 m.

The central loop response was also inverted using TEMIX (which was not available at the time of the initial analysis). Figure 5.10 shows the inversion model, data fit and equivalence analysis. The TEMIX inversions confirmed a three-layer geoelectrical structure: thin and conductive first layer (7 m thickness and 200 Ωm resistivity), resistive second layer (210 m thickness and 2×10^5 Ωm resistivity), and conductive bottom layer (300 Ωm resistivity).

Table 5.5 lists the resolution matrix. The order of parameter resolution is: thickness of the resistive second layer (the resolution matrix element is 0.96), conductance of the conductive first layer and resistivity of the conductive third layer. From the equivalence analysis, the depth to the conductive third layer is about 200-250 m.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>1st layer resistivity (ρ₁)</td>
<td>0.79</td>
</tr>
<tr>
<td>2nd layer resistivity (ρ₂)</td>
<td>0.01</td>
</tr>
<tr>
<td>3rd layer resistivity (ρ₃)</td>
<td>-0.05</td>
</tr>
<tr>
<td>1st layer thickness (t₁)</td>
<td>-0.19</td>
</tr>
<tr>
<td>2nd layer thickness (t₂)</td>
<td>0.01</td>
</tr>
</tbody>
</table>

At Site 3, the EINVERT4 and TEMIX inversions (Fig. 5.11) for the central loop sounding yield a four-layer model: a thicker and relatively resistive first layer, a thin conductive second layer with about 0.2 S conductance, and a relatively resistive third layer,
Figure 5.10 (a) Layered inversion model (solid line) and equivalent models (dash lines), (b) observed data and model response for central loop station at Site 1. The bad data were excluded from the inversion.
Figure 5.11 (a) Layered inversion model, (b) data response (apparent resistivity for central loop at Site 3.)
and a relatively conductive bottom layer. The data fit is poor at 10th-13th time gates.

Table 5.6 lists the resolution matrix for the TEMIX model. The resolution matrix indicates that the thickness of the first layer, conductance of the second layer and resistivity of the third layer can be resolved. The resistivity of the first and fourth layer and the thickness of the third layer can not be resolved. The thin conductive second layer masked the response of the deep structure. That means the soundings with a 40 m transmitter loop can not delineate deeper conductivity variation, such as the interface between fresh and saline water at this site.

Table 5.6 Resolution matrix for central loop model at Site 3.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>1st layer resistivity ($\rho_1$)</td>
<td>0.3</td>
</tr>
<tr>
<td>2nd layer resistivity ($\rho_2$)</td>
<td>0.13</td>
</tr>
<tr>
<td>3rd layer resistivity ($\rho_3$)</td>
<td>0.02</td>
</tr>
<tr>
<td>4th layer resistivity ($\rho_4$)</td>
<td>-0.02</td>
</tr>
<tr>
<td>1st layer thickness ($t_1$)</td>
<td>0.03</td>
</tr>
<tr>
<td>2nd layer thickness ($t_2$)</td>
<td>0.07</td>
</tr>
<tr>
<td>3rd layer thickness ($t_3$)</td>
<td>-0.08</td>
</tr>
</tbody>
</table>

5.5.3 Interpretation

At Site 1, the inversion models produce a three-layer geoelectric structure: a thin and conductive first layer (<10 m thickness, <220 $\Omega$m resistivity), a relatively thick and resistive second layer (>180 m thickness, >10$^4$ $\Omega$m resistivity), and a relatively conductive bottom
layer (150-350 Ωm).

Comparison of the results with nearby borehole logs (URL12 and URL13, Fig. 5.12) and near-surface geological information indicates that shallow conductivity probably corresponds to the overburden and near-surface weathered granite in which soils and organic material are deposited in open fractures. This result is same as that determined by VLF-EM resistivity survey and Schlumberger resistivity soundings, which determined a conductive overburden (10-27 Ωm) with 5-10 m thickness (Paterson and Watson 1981 referenced in Brown et al. 1989). The presence of clay in the soil contributes to an enhanced conductivity due to cation exchange on the surface of clay grains.

The resistive middle layer would correspond to granite with sub-vertical fractures containing the fresh water. The lower conductive layer can be attributed to the saline water zones. The depth of the fresh/saline water interface from TEM results, 195±40 m.

At Site 3, the inversions produce a four-layer model: a thicker and relatively resistive first layer, a thin conductive second layer at 100 m depth with 0.2 S conductance, a relatively resistive third layer with unresolved thickness, and a relatively conductive bottom layer. The first resistive layer is interpreted to correspond to the granite with sub-vertical fractures containing the fresh water. Comparison of the resistivity model with logs from nearby borehole URL-3 (Fig. 5.13) and high resolution seismic reflection results (Moon et al. 1993) suggests the conductive second layer is the shallow fracture zones FZ2 (Fig. 2.9). The shallow fractures may be a major reason of the enhanced conductivity at Site 3.

The conductive shallow fractures decreased the resolution depth of the TEM survey, and make it difficult to resolve the thickness of the resistive third layer and the resistivity of
Figure 5.12 Geophysical logs of borehole URL-13 (provided by Whiteshell Research Laboratory, 1996).
Figure 5.13 Borehole logs of borehole URL-3, which is located approximately 90 m north of Site 3 (provided by Whiteshell Research Laboratory, 1996).
the conductive bottom layer. Therefore, it is impossible to delineate the interface between fresh and saline water at Site 3 using a 40 m transmitter loop configuration.

5.5.4 Recommendation Based on Feasibility Studies

The results of the feasibility survey were used to recommend the configuration of a full TEM survey to be used to delineate the fresh/saline water interface in Lac du Bonnet area.

Factors influencing the effective exploration depth of a TEM system include the time range of the recording gates and the conductivity of the media (eq. 3.16). The skin depth provides a limiting factor for the exploration depth. For a $10^4 \, \Omega \, m$ half space (a typical resistivity value for granite rocks), the theoretical skin depth can be up to 3300 m for the UH recording time used by PROTEM 47 (Table 5.3). However, sedimentary rocks near the surface usually have been weathered and contain clay and fractures, which results in higher conductivity, decreasing the skin depth. The actual skin depth is less than 3300 m.

A second important factor affecting resolution depth is SNR (Fitterman 1989). The noise limits the time to which the signal can be used to observe the underground geoelectric structure (Fitterman 1989). For example, the signal level at Site 1 is close to the noise level at time greater than $2 \times 10^{-4} \, s$ (Fig. 5.5b). The recorded data at times greater than $2 \times 10^{-4} \, s$ are not useful for resolving the geoelectric structure. One way to improve the SNR at all recording times is to increase the transmitter moment (Fitterman 1989) which is a function of transmitter loop size, turn number and transmitter current. Equation (5.2) shows an approximate relationship between transmitter loop size ($l$) and depth of exploration ($d$) for the PROTEM system (GEONICS 1992a, b). It indicates that a larger transmitter loop can
increase the exploration depth.

\[ d = 8.94l^{0.4} \rho^{0.25} \]  

(5.2)

For a 40 m transmitter loop and a conductive structure with 200 Ωm resistivity, the exploration depth predicted by Eq. 5.2 is about 150 m. The responses of 40 m transmitter loop at Site 1 are close to the noise level at times >10^-4 s (Fig. 5.5b) and the inversion results show that the TEM survey using a 40 m transmitter loop configuration can only successfully map the interface between the resistive and conductive layers at depths less than 200 m. The resolution depth at this site is thus limited to about 200 m depth because of near-surface structure and noise.

In order to increase the SNR, improve the resolution depth and map deeper resistivity structure, it was determined that it would be necessary to employ a larger transmitter loop, e.g., 100 m, in the full survey at Lac du Bonnet Batholith. In order to limit budget costs it was recommended that the PROTEM47 belonging to University of Manitoba be used in the survey. The transmitter current 3 A will be used for 100 m transmitter loop, which is a maximum value recommended by GEONICS (GEONICS 1992a). The results of the feasibility study indicated that the full survey should include soundings up to the maximum offset allowed by the reference cable between transmitter loop and receiver coil (280 m). The offset soundings are expected to have higher SNR because the instrument noise from the transmitter loop decreases with the increasing offset.
5.6 Data Analysis: Full Survey

5.6.1 Preliminary Analysis

The details of the data collection, background noise, recorded voltages, vertical magnetic fields and apparent resistivity responses for all receiver stations at the four sites of the full survey have been presented in Maris et al. (1997) and Maris (2000). Maris et al. (1997) also compared the responses for different offsets for each site and the responses for the same offset at different sites, and analysed the central loop soundings by modelling and inversion, using RECTAN and EINVRT4.

The results of Maris et al. (1997) suggested a three-layer subsurface resistivity model at four sites: a thin (less than 3 m thickness) and relatively conductive surface layer that is interpreted to be overburden, a very resistive middle layer and a more conductive bottom layer (Fig. 5.14). Constrained inversions of the central loop data were completed by fixing the top layer parameters and varying the interface depth between the second and third layers. The minimum misfit for each inversion model was examined to determine the range of allowable models. The best estimate of the depth to the top of the lower conductive layer is $370 \pm 30$ m for Site A, $160 \pm 20$ m for Site B, $145 \pm 20$ m for Site D, and $540 \pm 80$ m for Site URL (Table 5.7, Maris et al. 1997). The interface depths for Sites B and D were well determined, whereas the depth at URL was less well resolved by the data. The results at Site A are less reliable because the responses contain multiple sign reversals not fitted by the data.

In this thesis, the analysis of the TEM data is extended and expanded to include the offset loop data. This approach increases the volume of data used in the analysis from 4 to over 200 TEM receiver stations. The expanded analysis is, therefore, expected to provide
Figure 5.14 The inversion model and data fit for central loop station at Site B (Maris et al., 1997).
more accurate and more detailed information on the resistivity structure at each site. The analysis is based on the use of the program TEMIX (Interpex Limited 1996) which allows both 1D forward modelling and inversion of central loop and offset TEM data.

**Table 5.7** Best-fitting resistivity models for each site determined from central loop data. (Maris et al. 1997)

<table>
<thead>
<tr>
<th>Site</th>
<th>Layer 1</th>
<th>Layer 2</th>
<th>Layer 3</th>
<th>Layer 1/Layer 2</th>
<th>Layer 2/Layer 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Resistivity (Ωm)</td>
<td>45</td>
<td>50</td>
<td>45</td>
<td>50</td>
<td>50</td>
</tr>
<tr>
<td>Interface depth (m)</td>
<td>1.8</td>
<td>2.7</td>
<td>1.2</td>
<td>1.5</td>
<td>1.5</td>
</tr>
</tbody>
</table>

**5.6.2 Signal Responses**

Figure 5.15 shows TEM soundings with the same offset on different azimuths at Site A, B, D and URL, and compares them with the average noise level at each site. In general, the signal responses consist of a decay from a maximum of $\sim 10^4$ mV for the first gate to a value of $\sim 0.05$-1 mV for the last gate. The decay varies from smooth to irregular, or a combination of both. The most common response consists of an initial, early-time irregular decay followed by a smooth decay at later times, however, the signals for Site A show irregular decay at all recording times. Additional data are shown in Maris et al. (1997).

At all four sites, the signal responses are higher than the environmental noise levels, in particular for the first ten gates. The signal to noise ratio decreases for the last ten time-gates. Site B has the highest SNR. The signal levels for Site A and URL decay faster than
Figure 5.15. TEM sounding with the same offset at Site A, B, D and URL. The average noise level of each site is included for comparison.
those for Site B and D. The signal levels of the last six time-gates at Site A and URL are lower and close to the noise level, although Site A and URL have lower late-time noise levels than Site B and D (Fig. 5.3). The observations show differences in the signal response at four sites.

Figures 5.16-5.19 show the contour plots of the apparent resistivity for Site B, D, URL and A, which were generated for west to east and south to north profiles. The apparent resistivity is contoured as a function of position and time. The plots use a logarithmic time axis because the TEM responses are recorded using logarithmically spaced time gates (Table 5.1). The apparent resistivity may vary over several orders of magnitude, so the log of the apparent resistivity is contoured. A kriging algorithm was used to interpolate the data and the program SURFER (Keckler 1994) was used for contouring and plotting. In order to allow closer evaluation of features on the plots, profiles of the response for individual time gates are shown in Appendix G. In subsequent discussions of results, the notation INN refers to the results for offset NN on line I (e.g. N20 for the receiver location 20 m north of loop centre).

All responses show erratic short-time responses at time less than $6 \times 10^{-5}$ s. For later time gates, the apparent resistivity is high and decreases with increasing time at Site B, D and URL. The response of Site A is different from the other three sites and shows multiple sign reversals. Some anomalous zones can be observed in these contour plots. A detailed examination of the plots will be undertaken in following paragraphs.

**Responses at Time $<6 \times 10^{-5}$ s**

The responses at time $<6 \times 10^{-5}$ s are relatively erratic with large response changes
Figure 5.16 Late-time apparent resistivity versus time profiles at Site B. (a) east-west profile, (b) north-south profile. The vertical axis shows the logarithm of time and the horizontal axis shows the receiver offset. The contour interval equals 0.1 decades of the logarithm of the apparent resistivity.
Figure 5.17 Late-time apparent resistivity versus time profiles at Site D. (a) east-west profile, (b) north-south profile. The vertical axis shows the logarithm of time and the horizontal axis shows the receiver offset. The contour interval equals 0.1 decades of the logarithm of the apparent resistivity.
Figure 5.18 Late-time apparent resistivity versus time profiles at Site URL. (a) east-west profile, (b) north-south profile. The vertical axis shows the logarithm of time and the horizontal axis shows the receiver offset. The contour interval equals 0.1 decades of the logarithm of the apparent resistivity.
Figure 5.19 Late-time apparent resistivity versus time profiles at Site A. (a) east-west profile, (b) north-south profile. The vertical axis shows the logarithm of time and the horizontal axis shows the receiver offset. The contour interval equals 0.1 decades of the logarithm of the apparent resistivity.
between adjacent stations (Fig. 5.16-5.19). The responses are further complicated by sign reversals and jumps between adjacent time-gates. The erratic values and sign reversals are particularly evident at small offsets (<100 m). These effects decrease with increasing time and increasing offset.

The erratic responses with sign reversals at short-time are interpreted as being due to various sources of noise. Noise from geological structure (inhomogeneous near-surface resistivity structures) affects nearby stations. Environmental noise may affect the responses at all stations. System noise from instruments affects the stations close to the transmitter loop.

The early-time response at Site URL and A are more irregular than at Site B and D. This is interpreted to be due to the stronger lateral variations in near-surface resistivity at Site URL and A, and lower signal levels associated with higher resistivity.

At Site URL (Fig. 5.18), the early-time effects typically extend over the first 10 time gates (to $7 \times 10^{-5}$ s), i.e., later than at Site B and D where they typically extend over 6 time gates. The apparent resistivity is higher than at the other sites and reaches values as high as 10,000 $\Omega$m.

**Responses at Time >6$\times 10^{-5}$ s**

At Sites B, D and URL, the responses are more regular and vary smoothly as a function of position and time at times greater than $2 \times 10^{-5}$ s and offsets greater than 100 m. The apparent resistivity decreases with increasing time although there is a minor increase at intermediate time at Site D (Fig. 5.16-5.18). The later-time responses at Site URL (Fig. 5.18)
are more resistive and erratic than at Site B and D.

At Site B, the apparent resistivities along the west-east and south-north profiles gradually decrease with increasing time, from about 2500 Ωm at $4.2 \times 10^3$ s to about 400 Ωm at $7 \times 10^4$ s. Examination of the response for individual time gates shows that resistivity decreases towards both ends of the east-west profile, and also towards the north end of the north-south profile (Appendix G). These effects may be explained by either decreasing resistivity of rocks or by a decrease in depth to an underlying conductive layer.

At Site D, apparent resistivities along west-east and south-north profiles decrease with increasing time, from approximately 2000 Ωm at $2.6 \times 10^3$ s to 800 Ωm at $1.72 \times 10^4$ s. There is an increase of apparent resistivity at $2.15 \times 10^4$ s, to about 1000 Ωm, followed by a smooth decrease at later times at most stations. The apparent resistivity decreases significantly at the north end of the profile at stations beyond N200. For several locations at Site D (i.e. at the ends of the east and south lines) the response at the late time-gates is more erratic than at Site B. Examination of the noise response (Fig. 5.3) shows that Site D has relatively high noise levels at the latest time gates.

At Site URL, apparent resistivities along both west-east and south-north profiles gradually decrease with increasing time from approximately 10,000 Ωm at $7 \times 10^5$ s to approximately 2200 Ωm at $2.7 \times 10^4$ s. The lower late-time signal levels at URL relative to at Site B and D indicate a more resistive structure.

*Local Conductive Responses*

At Site B, D and URL, there are several anomalous resistivity zones (Fig. 5.16-5.18),
which only are presented at two or three stations. At W10-W40 and S40-S60 of Site B (Fig. 5.16) conductive responses (<800 Ωm) extend from the earliest time recorded. At W200 and E200 of Site D (Fig. 5.17) conductive responses (<600 Ωm) extend to late time. The conductive zone around W200 extends to W260 and the conductive zone around E200 extends from E150 to E240. At Site URL, there are two relatively conductive zones (<2000 Ωm) around W220 and E50. The width of these zones is narrow, about 20-30 m.

Responses at Site A

The responses at Site A (Fig. 5.19) are different than those from the other three sites. Figure 5.19 shows that multiple zones of higher and lower resistivity occur in the late time responses. These zones are continuous from west to east and from south to north. The anomalous zones are characterized by two sign reversals:

1) a sign change at 54-68 µs, which can be observed along both south-north and west-east profiles.

2) a sign change at 275-436 µs, which is observed at most stations.

The times of the sign reversals are consistent between adjacent stations indicating that the reversals are true geophysical responses and not spurious results due to noise. Because of the sign changes, it is difficult to characterize the overall trends in apparent resistivity at this site. Note the TEM responses at stations beyond N180 on the north line: a prominent resistive feature is observed to extend from the earliest time, $6.9 \times 10^5$ s to relatively late time $1.35 \times 10^4$ s, and appears to "dip" north.
5.6.3 1D Layered Inversion Results

The inversions were completed using TEMIX. The first step in the inversion involved examination of the smooth models from the Occam inversion for the different stations. The smooth models were used to assess the resistivity structure. For some stations, the Occam models provide a good fit to the data (Figure 5.20a), and can be used as initial models for layered inversion in the subsequent step. However, at other stations the program can not provide perfectly smooth models to fit the data (Figure 5.20b). This problem occurs most often for the complex responses: the smooth model can not represent the sharp variation in resistivity. If the input data provide only weak constraints on depth parameters, the inversion solution may determine a pathological solution with unrealistically high or low depth estimates (Interpex Limited 1996).

*Site B*

For Site B, TEMIX provided reasonably fitting smooth models at a total of 17 stations. The program did not provide good models for stations with offsets between about 20-80 m on both west-east and south-north profiles due to noisy early time responses (Fig. 5.16).

Figure 5.21 shows a contoured figure of the good fitting smooth models for Site B. The crosses on the figure show the actual data points used to define the contours. There are two main resistivity transitions. A shallow conductive layer can be seen at a number of sites (e.g., S140, 0, N100). This near-surface conductive zone overlies a very resistive zone (>4000 Qm). The resistive zone is absent or less resistive at the ends of the east-west profile, and is thinner at large offsets to the north. The resistive zone overlies a more conductive zone.
Figure 5.20 1D Occam inversion model and response for (a) station S100 at Site D for which the occam model response provides a good fit to observed data, (b) station N240 at Site D for which the occam model needs to be improved to fit the data better.
Figure 5.21 Contour plot of one-dimensional Occam inversion models for the Site B. (a) east-west, (b) south-north profiles. The crosses show the location of data points used in the contouring. Only those Occam models which provide a good fit to the data were used.
(resistivity approximately 250-700 $\Omega$m). The transition to the more conductive zone occurs at a depth of approximately 200 m below the centre of the transmitter loop.

If the true resistivity structure contains abrupt resistivity variations, several features will not be clearly resolved in the smooth TEM inversion models. Firstly, the Occam inversion results will tend to show a conductive surface layer as being less conductive and thicker than the true structure. Secondly at zero offset, there is a relatively thick conductive zone beneath the very resistive layer at 200-300 m depth. This feature probably arises as an artifact of the regularization constraint requiring the transition from the very resistive (or conductive) to conductive (or resistive) zone to be as smooth as possible. Inversions based on a minimum number of layer, described below, show sharp resistivity changes more clearly.

Figure 5.22 shows an example of the layered inversion model, equivalence analysis, and model responses (voltage and apparent resistivity) for station E220 at Site B. The model responses fit the data very well (Fig. 5.22c and 5.22d). The inversion model indicates a four-layer structure. A thin (<2 m) and very conductive (15 $\Omega$m) surface layer (not easily visible in the figure) overlies a very resistive layer (~10,000 $\Omega$m). The resistive layer extends to a depth of 50 m. It overlies a relatively conductive layer (~300 $\Omega$m) extending to 320 m depth. The bottom layer in the model is more conductive, with a resistivity of approximately 80 $\Omega$m.

Table 5.8 lists the parameter resolution matrix of this inversion model. The resolution matrix elements for the thickness and resistivity of the thin surface layer ($R_p$ and $R_t$) are 0.95. The cross value between the surface layer thickness and resistivity ($R_p$ and $R_t$) is very low, -0.04. These results indicate that the individual thickness and resistivity of the first layer
Figure 5.22 Layered inversion model and response for station E220 at Site B. (a) Four-layer inversion model, the surface layer is not visible on figure. (b) equivalence analysis (final model—solid line, equivalence model - dashed line), (c) voltage response, and (d) apparent resistivity response.
are quite well resolved. The result is supported by the good fit of the model response to the data. The thickness of the resistive second layer \((R_t t_2)\) is well resolved \((1.0)\), but its resistivity \((R_\rho_2)\) is poorly resolved. The resistivity is not dependent on the thickness (there is a small cross resolution value \(R_\rho_t t_2 = 0.01\)). The thickness of the third layer is resolved quite well \((R_t t_3 = 0.95)\). The value for \(R_\rho_3 t_3\) is 0.8 and for \(R_\rho_3 t_3\) is -0.16, indicating that the resistivity of this layer is somewhat dependent on the thickness. The resistivity of the underlying layer is dependant on the third layer resistivity \((R_\rho_4 t_3 = -0.25)\) and thickness \((R_\rho_4 t_3 = -0.2)\), and is poorly resolved \((R_\rho_4 t_4 = 0.47)\).

**Table 5.8 Resolution matrix for station E220 at Site B.**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>1\textsuperscript{st} layer resistivity ((\rho_1))</td>
<td>0.95</td>
</tr>
<tr>
<td>2\textsuperscript{nd} layer resistivity ((\rho_2))</td>
<td>0.05</td>
</tr>
<tr>
<td>3\textsuperscript{rd} layer resistivity ((\rho_3))</td>
<td>0.03</td>
</tr>
<tr>
<td>4\textsuperscript{th} layer resistivity ((\rho_4))</td>
<td>0.8</td>
</tr>
<tr>
<td>1\textsuperscript{st} layer thickness ((t_1))</td>
<td>0.06</td>
</tr>
<tr>
<td>2\textsuperscript{nd} layer thickness ((t_2))</td>
<td>0.01</td>
</tr>
<tr>
<td>3\textsuperscript{rd} layer thickness ((t_3))</td>
<td>0.01</td>
</tr>
</tbody>
</table>

Further tests were done to examine the resolution of the depth to base of the resistive layer. In particular, the sensitivity of this depth to change in the overburden layer was examined. For stations near the centre of the loop, for which the early time data were well resolved, it is possible to correctly determine the surface layer and delineate the correct depth.
to the base of the resistive layer. For large offsets, the early time data contains sign changes that dependent on the resistivity and thickness of the surface layer and could not be correctly fitted in the inversions. For these sites it is more difficult to separate the effects from the surface layer in the resolution of the deeper structure. However, it was found that a large change in the surface layer (e.g. factor of 2 change in conductance) corresponded to a relative small change in the depth to the base of the resistive layer (~10%). The results indicate that the depth to the base of the resistive layer at large offsets could be slightly deeper than determined if the surface layer is more conductive than that shown in my models. Correction of this effect would require use of other EM surveys to determine the resistivity of the near surface layer.

One-dimensional inversions were done for all stations at Site B and a good fit between the model and observed TEM data was obtained at 16 stations. For small to intermediate offsets (<100 m), large misfits occurred because of noise. There are some stations at large offsets (e.g., beyond W160 and S220) where the response is smooth but could not be fitted using a layered model.

Appendix H shows layered inversion models, equivalence analysis, model responses and resolution matrices for typical stations from the different lines (west, south and north) at this site. The inversion models from 16 stations are shown in Fig. 5.23.

Figure 5.23 is an image of the large-scale resistivity structure at Site B. The basic four layer model is noted at all stations. The thickness of individual layers varies from station to station and the correlation of each layer between stations is shown on the figure.

- Conductive Surface Layer. The thin, conductive surface layer has resistivity 10-30 Ωm and
Figure 5.23 Profiles of layered inversion models at B Site. The profiles include only those inversion models which provide a good fit to the TEM data. The dashed lines show the correlation of major resistivity layers between stations. At this plotting scale, the thin conductive surface layer cannot be seen clearly.
thickness 0.5-2.0 m. The higher resolution matrix elements indicate this layer can be resolved by the TEM models (Table 5.8 and Appendix H).

- **Resistive Layer.** The resistivity of the second layer exceeds 2000 Ωm at all sites. The thickness varies from 180 m near the centre of the transmitter loop to 100 m at the end of the south line and 50 m at the ends of the east and north lines. On the west line the thickness decreases to 70 m (at W140) before increasing at greater offset. The Occam model results show a corresponding change on the west line (Fig. 5.21).

However, the layered inversion models are consistent with the central loop inversion results from Maris et al. (1997), in which the base depth of this resistive layer was determined to be 160 ± 20 m.

- **Moderately Conductive Layer.** The resistivity of the third layer varies between 50 Ωm and 300 Ωm. The base of the layer lies at a depth of 300-350 m over most of the two profiles, but rises near the west, east, and north ends. These results reflect the trends seen in the TEM responses (Fig. 5.16) in which the apparent resistivity at late times decreased towards the ends of the lines. The resistivity and thickness of this layer can be resolved, however they have the relatively large range. The resistivity varies from 50 Ωm to 400 Ωm and the thickness varies from 300 m to 370 m.

- **Very Conductive Bottom Layer.** The resolution of the resistivity of the base layer is low, however, the resistivity model and equivalence analysis show that this layer is conductive (<85 Ωm) (Fig. 5.23 and Appendix H).
Site D

At Site D, TEMIx provided reasonable Occam smooth models for a total of 18 stations. Figure 5.24 shows the contoured smooth-model profiles based on the acceptable models. Note that the data points on which the contours are based are sparsely distributed at smaller offsets.

The Occam models for Site D are similar to those for Site B, but resolve an additional resistivity transition. The conductive surface layer is present at some stations (E140, W140, S160, and S200), but is absent at others (W100, W120, N140, N160). The near-surface conductive zone overlies a resistive zone (>2000 $\Omega$m), which is absent or less resistive at the ends of the east-west profile, and is thinner at large offsets on the north line. The depth to the bottom of this resistive zone is less than 200 m. Its resistivity is less than that of the corresponding resistive zone at Site B. The resistive zone overlies a more conductive zone (approximately 200 $\Omega$m) which has almost uniform thickness along the profiles. Beneath this conductive zone, the resistivity increases to 1000 $\Omega$m. This increase in resistivity is much sharper than the corresponding resistivity increase observed at Site B. For Site B, the resistivity increase was not required by the data (as shown by the non-regularized inversions) and was interpreted as an artifact of the Occam inversion. At Site D the increase is definitely required by the data and is needed to fit a deflection in the apparent resistivity curves at approximately $10^4$ s (Fig. 5.20). Beneath the resistive layer, the models become more conductive with depth.

A number of inversions using a three-layer model as the initial model failed to converge to a model providing a good fit to the data. The data from Site D could, however,
Figure 5.24 Contour plot of one-dimensional Occam inversion models for the Site D. (a) east-west, (b) south-north profiles. The crosses show the location of data points used in the contouring. Only those Occam models which provide a good fit to the data were used.
be fitted using a four-layer model in which there is a surficial layer (at some sites), an upper resistive layer 300-400 m thick containing an embedded conductive layer at 100-200 m depth, and a more conductive layer at depth.

Figure 5.25 shows an example of the layered inversion model, equivalence analysis, and model responses (voltage and apparent resistivity) for station N140 at Site D. Because of the irregular data at earlier times, the data fit is poor. But the model responses fit the data very well at later times (Fig. 5.25c and 5.25d). As noted in the Occam models, the conductive surface layer is absent at this station.

The inversion model for N140 at Site D indicates a four-layer structure: a resistive first layer (350 m thickness, >3000 Ωm resistivity) with an embedded thin conductive layer (approximately 0.3 S conductance), and a conductive bottom layer (approximately 70 Ωm resistivity).

The resistivity and thickness of the resistive layer are very well resolved for the part of the layer above the embedded conductive zone (R\rho1\rho1=0.96 and Rt1t1=0.99, Table 5.9). The equivalence models also provided consistent parameter values (Fig. 5.25b). The resistivity is poorly resolved beneath the embedded conductive layer (R\rho3\rho3 =0.01), but the best fitting models suggest the resistivity is similar above and below the conductive layer. The resistivity of bottom layer is only moderately well resolved (R\rho4\rho4=0.25).

For the embedded conductive layer, the resolution element R\rho2\rho2 is 0.59 and Rt2t2 is 0.52. The resistivity and thickness of layer are strongly dependent on each other (R\rho2t2 =-0.44). These values indicate a very high covariance between thickness and resistivity of this layer. The conductance of this layer is well resolved with a value of 0.3 S (for example, the
Figure 5.25 Layered inversion model and response for station N140 at Site D. (a) Layered inversion model. (b) equivalence analysis (final model - solid line, equivalence model - dash line). (c) voltage response, and (d) apparent resistivity response.
resistivity is ~100 $\Omega$m for a thickness of 30 m). The equivalence analysis shows the depth to the embedded conductive layer is also quite well resolved, at 130±10 m (Fig. 5.25b).

Table 5.9 Resolution matrix for station N140 at Site D.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>$1^\text{st}$ layer resistivity ($\rho_1$)</td>
<td>0.96</td>
</tr>
<tr>
<td>$2^\text{nd}$ layer resistivity ($\rho_2$)</td>
<td>-0.03 0.59</td>
</tr>
<tr>
<td>$3^\text{rd}$ layer resistivity ($\rho_3$)</td>
<td>0.01 0.01 0.01</td>
</tr>
<tr>
<td>$4^\text{th}$ layer resistivity ($\rho_4$)</td>
<td>-0.03 0.07 -0.03 0.25</td>
</tr>
<tr>
<td>$1^\text{st}$ layer thickness ($t_1$)</td>
<td>0.01 0.04 0 -0.01 0.99</td>
</tr>
<tr>
<td>$2^\text{nd}$ layer thickness ($t_2$)</td>
<td>-0.03 -0.44 -0.04 0.14 0.04 0.52</td>
</tr>
<tr>
<td>$3^\text{rd}$ layer thickness ($t_3$)</td>
<td>0 -0.01 0.03 -0.02 0 -0.01 0.98</td>
</tr>
</tbody>
</table>

Appendix H shows layered inversion models, equivalence analysis, model responses and resolution matrices for typical stations from different lines (east, west and south) at Site D. Figure 5.26 provides an image of the resistivity structure at Site D based on satisfactory layered model inversions obtained at 18 stations. A similar structure is observed at most stations although the thickness of individual layers varies from station to station.

- **Conductive Surface Layer.** A conductive surface layer with resistivity of 10-30 $\Omega$m and thickness 0.5-2.0 m is present at some stations. The resistivity and thickness of this layer show significant lateral variation (Fig. 5.24), however, the resolution matrix shows high values for these parameters (Table 5.9), indicating they can be resolved.

- **Resistive Layer.** The resistivity of the second layer is >1500 $\Omega$m. Equivalence analysis
Figure 5.26 Profiles of layered inversion models at D Site. The profiles include only those inversion models which provide a good fit to the TEM data. The dashed lines show the correlation of major resistivity layers between stations. At this plotting scale, the thin conductive surface layer cannot be seen clearly.
shows that this parameter has large variation, in the range of 1500-100,000 $\Omega$m, and the resolution of this parameter is very low. The thickness varies from 400 m near the centre of the transmitter loop to approximately 300 m near the end of the lines. The resolution of thickness is high (Fig. 5.25 and Table 5.9). The structure is somewhat different at stations W180, S240, E200 (Fig. 5.17).

- **Embedded Conductive Layer.** The conductance of the embedded conductive layer is approximately 0.3 S, and is relatively uniform along the profiles (about $\sim$30 m thickness for 100 $\Omega$m resistivity). The individual resistivity and thickness of this layer can not be resolved, only conductance can be resolved. The top of this embedded conductive layer is at 100-210 m depth along the profiles.

  The analysis of the central loop data from Site D by Maris et al. (1997) showed the resistive/conductive interface is at about 145$\pm$20 m depth. This depth correlates most closely with the depth to the embedded conductive layer in the present inversions (210 m).

- **Bottom Conductive Layer.** The resistivity of the bottom layer in the inversion models is less than 200 $\Omega$m and reaches values as low as 10-20 $\Omega$m. However, the resolution of this parameter is low, and equivalence results show there is a large variation range, 5-200 $\Omega$m. The depth to the top of this layer is about 400 at near the centre of the transmitter loop, and decreases to about 330 m towards the northwest and to about 200 m towards the southeast. The equivalence analysis shows this depth is resolved (Fig. 5.25, Table 5.9 and Appendix H).
At Site URL, TEMIX provided reasonable-fitting smooth models at a total of 22 stations. The inversions did not provide good models for stations with offsets between 20 and 80 m because of noisy early time responses. Figure 5.27 shows a contour figure of the satisfactory smooth models.

The contour plots show a three-layer structure. There is a shallow conductive layer at most stations, which overlies a very resistive zone (with resistivity $>10^5 \Omega$ m). This resistive zone is less resistive at W140 and near N180. The overall resistivity of the zone is much greater than at Sites B and D. The resistive zone overlies a more conductive zone (approximately 1000 $\Omega$ m). The transition to more conductive zone occurs at a depth of approximately 450-500 m at most stations and at slightly greater depth at S100 and W100. The resistivity of the lower zone is also much greater than at Sites B and D.

Figure 5.28 shows an example of the layered inversion model, equivalence analysis, and model responses (voltage and apparent resistivity) for station S120 at Site URL. Because the data has lower signal to noise ratio at this site (Fig. 5.15), it is not possible to fit the observed data as well as at Sites B and D (Fig. 5.20 and 5.25) and the resolution of the interpreted model parameters is accordingly reduced (Table 5.10).

The inversion model indicates a three-layer structure: a thin conductive surface layer (<10 m thickness, <1000 $\Omega$ m resistivity), a thick very resistive layer (400 m thickness, $>100,000 \Omega$ m resistivity) and a conductive bottom layer ($>300 \Omega$ m resistivity). The thickness and resistivity of the thin surface layer can not be resolved very well ($R_{\rho_1}$,$\rho_1$=0.85, $R_{t_1}$,$t_1$=0.83) and the data fit at earlier times is poor (Fig. 5.28). The two parameters of this
Figure 5.27 Contour plot of one-dimensional Occam inversion models for the Site URL. (a) east-west, (b) south-north profiles. The crosses show the location of data points used in the contouring. Only those Occam models which provide a good fit to the data were used.
Figure 5.28 Layered inversion model and response for station S120 at Site URL. (a) Layered inversion model, (b) equivalence analysis (final model - solid line, equivalence model - dashed line), (c) voltage response, and (d) apparent resistivity response.
layer are dependent \((R_p, t_1 = -0.16)\). The thickness of the resistive second layer is well resolved \((R_t, t_2 = 0.99)\), but its resistivity is poorly resolved \((R_p, \rho_2 = 0.40)\). The resistivity of the bottom layer can not be resolved very well \((R_p, \rho_3 = 0.80)\).

**Table 5.10** Resolution matrix for station S120 at Site URL.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>1st layer resistivity ((\rho_1))</td>
<td>0.85</td>
</tr>
<tr>
<td>2nd layer resistivity ((\rho_2))</td>
<td>0.21 0.4</td>
</tr>
<tr>
<td>3rd layer resistivity ((\rho_3))</td>
<td>0.1 0 0.8</td>
</tr>
<tr>
<td>1st layer thickness ((t_1))</td>
<td>-0.2 0.22 0.1 0.83</td>
</tr>
<tr>
<td>2nd layer thickness ((t_2))</td>
<td>0 0 0 0 0.99</td>
</tr>
</tbody>
</table>

Appendix H shows some layered inversion models, equivalence analysis, model responses and resolution matrices for stations from the different lines (east, west and north) at Site URL. Figure 5.29 provides an image of the resistivity structure at Site URL based on the satisfactory layered model inversions obtained at 22 stations. The majority of stations were fitted with three layer models (Fig. 5.28, Fig. G.7 and G.9). At two stations (N180 and N200), it was necessary to include an embedded conductive layer within the resistive layer, a structure similar to the models at Site D (Fig. G.9). At stations north of N200 the resistivity of the resistive layer is lower than at other stations (Fig. 5.29). Apart from these structures the resistivity structure at this site is relatively uniform along much of the profiles.

- **Conductive Surface Layer.** The thin, conductive, surface layer is present at all stations. The equivalence results show the resolution of its resistivity and thickness is relatively low.
Figure 5.29 Profiles of layered inversion models at URL Site. The profiles include only those inversion models which provide a good fit to the TEM data. The dashed lines show the correlation of major resistivity layers between stations. At this plotting scale, the thin conductive surface layer cannot be clearly seen.
(Fig. 5.28, Table 5.10 and Appendix H). However, the results suggest the resistivity of surface layer is less than 1000 $\Omega$m and that the thickness is less than 10 m.

- **Resistive Layer.** The resistivity of the second layer is constrained to be $>50,000$ $\Omega$m, but the exact resistivity value is poorly resolved (very low resolution matrix elements, Table G.3). The best-fitting models suggest the resistivity is close to 400,000 $\Omega$m at most stations. This value is significantly higher than at Sites B and D. At Site URL, the thickness of this layer along the north-south profile is approximately 400 m at all stations except for S100 where it is 450 m. The thickness along the east-west profile increases from 350 m at W280, to approximately 460 m at E160 with a locally thicker value, 540 m, at W100. The thickness decreases again at the eastern end of the profile. The locally thicker values at W100 and S100 were also observed in the Occam models (Fig. 5.28).

An analysis of central loop data from this site by Maris et al. (1997) found the depth of the very resistive layer to be about 540±80 m. Because of the relatively low resolution of all model parameters at Site URL (cf Sites B and D), the confidence limits on the base depth of resistive layer are large (±80 m). Considering this factor, the result from TEMIX (~400 m) is close to that from Maris et al. (1997). The difference may be explained in part by the relatively noisy TEM data at this site.

- **Embedded Conductive Layer.** Only the conductance and depth of this conductive layer can be resolved. The conductance of the embedded conductive layer at N180 and N200 is 0.05 S, considerably less than the value for the embedded layer at Site D. This layer is located at 125-155 m depth.

- **Bottom Conductive Layer.** The resistivity of the bottom layer in the inversion models is
300-1900 Ωm and is relatively uniform across the profiles. The exact resistivity value is poorly resolved. However, the bottom layer is constrained to be significantly more resistive than at Sites B and D.

In 1980, several ground geophysical surveys were carried out within the URL lease area (Paterson and Watson 1981). These included complex resistivity surveys using gradient and dipole-dipole arrays, as well as DC-resistivity soundings using the Schlumberger array. Several two-layer interpretations of dipole-dipole responses indicated significant drops in resistivity at depths ranging from 300 m to 700 m. The Schlumberger soundings also detected low resistivities at depths ranging from 110 m to 220 m.

**Site A**

It was not possible to obtain satisfactory Occam or layered inversion models for any of the stations at Site A. Although TEMIX can produce forward model responses containing multiple sign reversals, it is not able to satisfactorily invert data containing such sign reversals (Gil Levy, GEONICS Limited, pers. comm. 1998).

A TEMIX inversion of the central loop data, ignoring sign changes and based on a three-layer model, shows a depth of about 300 m to a bottom conductive layer. The response of the resulting model lies relatively close to the observed apparent resistivity and voltage data but does not contain the sign changes noted in the observed voltage data.

Maris et al. (1997) also constructed a three-layer model for the central-loop data by fitting a late-time apparent resistivity response that neglected sign changes in the data. The inversions used EINVERT4 obtained a depth to a basal conducting layer of 370±30 m, which
is close to the result of the present study.

5.7 Interpretation for Full Survey

Approximately 170 TEM offset soundings from Sites B, D and URL were inverted in this study. The resistivity structure at Sites B, D and URL based on the satisfactory layered model inversions are shown in Fig. 5.23, 5.26 and 5.29 and summarised in Table 5.11.

Table 5.11 Geoelectric structure from 1D inversions at Site B, D and URL.

<table>
<thead>
<tr>
<th></th>
<th>Site B</th>
<th>Site D</th>
<th>Site URL</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>resistivity (Ωm)</td>
<td>depth to base (m)</td>
<td>resistivity (Ωm)</td>
</tr>
<tr>
<td>thin conductive surface layer</td>
<td>10-30</td>
<td>&lt;2</td>
<td>10-30</td>
</tr>
<tr>
<td>thick resistive layer</td>
<td>&gt;2000</td>
<td>50-180</td>
<td>1500-4000 300-400</td>
</tr>
<tr>
<td>conductive lower layer</td>
<td>top part 50-400 300-350</td>
<td>&lt;200</td>
<td></td>
</tr>
<tr>
<td></td>
<td>lower part &lt;85</td>
<td></td>
<td></td>
</tr>
<tr>
<td>embedded conductive layer</td>
<td>-</td>
<td>Present at most stations ~ 0.3 S, the depth of top is 100-200 m.</td>
<td></td>
</tr>
</tbody>
</table>

Ranges represented in depth to base mean the lateral variation range along the profiles.

The inversion results show that the large-scale resistivity structures at Sites B, D and URL have a similar form, with a thin conductive surface layer overlying a thick resistive layer which, in turn, overlies a lower conductive layer (Table. 5.11). The thickness of individual layers varies from station to station, and the resistivity and thickness of individual
layers varies from site to site. The results from Site A are different from the other sites and will be discussed separately in terms of the interpretation.

5.7.1 Comparison of TEM Models with Borehole Logs

The resistivity structure determined from the TEM data was compared with geophysical logs and fracture logs from nearby boreholes. Geophysical logs include 40 cm (16 inch) normal resistivity, fluid resistivity, caliper and fluid temperature. Generalized lithology, where available, has been incorporated as colour fill in the plotted logs.

- **Site B**: Two boreholes, WB-1 and WB-2, are collared close to the south line of the TEM survey (Fig. 5.30). Figure 5.31 compares these geophysical logs WB-1 and WB-2 with TEM models. Logs of WB-1 and WB-2 indicate a decrease in rock resistivity at 120-150 m depth. A fractured, more electrically conductive interval extends from this depth to about 270 m in WB-1 and to about 300 m in WB-2, and is interpreted to be related to a zone of mixed fresh-saline water saturated granite. Below 270 m, the borehole logs show resistivity decreases which are interpreted to be related to a transition to saline water in sparsely fractured rock.

The TEM resistivity models agree very well with the observed logs and show two conductive layers beneath 100 m. On the south line, the depth to the top of the first conductive layer is 100-150 m, and the models include a more conductive layer at ~300 m depth (Fig. 5.23, Fig. 5.31). Quantitative comparison between the borehole log and TEM model will be discussed in Chapter 6.

- **Site D**: The closest boreholes to the Site D TEM survey are WD-1, WD-2, and WD-3, all
Figure 5.30 TEM sounding location at Site B. The red square shows the transmitter loop location and the red lines show the locations of the receiver profiles. One mapped lineament is shown as a brown line (modified from Sikorsky, 1996).
Figure 5.31 Comparison of geophysical logs from nearby boreholes with 1D TEM inversion model at Site B. Shown are caliper, temperature, fluid resistivity, 40 cm normal resistivity, TEM layered-inversion model and fracture logs with generalized lithology represented as colour fill. (a) Borehole WB-1 with TEM model for station S220, (b) Borehole WB-2 with TEM model for station S220. (From Wu et al. 1999).
Borehole WB-2

Caliper Temperature Fluid Resistivity 40cm Normal Resistivity Layered-Model S220 Fracture Frequency

Depth (m)

mm Celsius ohm-m ohm-m ohm-m ohm-m Fractures/m
of which are at least 500 m from the nearest TEM station (Fig. 5.32). Figure 5.33 compares the borehole logs WD-1 and WD-2 with the TEM response from a station on the north line (N200). The logs show the presence of fractures at 90-130 m depth at WD-1 and 70 m depth at WD-2, which could relate to the conductive embedded layer at depth 100-200 m in the TEM model. The logs suggest the fracture zone dips to the south. The TEM results from the north line also show a conductive zone deeping from 150 m at N200 to 200 m at the centre of the transmitter loop. It is more difficult to correlate the depth of the lower conducting layer in the TEM models with a change in resistivity in the borehole logs. The decrease of resistivity in the borehole logs consists of a gradual decrease below 130 m with superimposed anomalous variations over small depth intervals. In contrast, the TEM results suggest a significant decrease in resistivity at 300 m to 400 m depth, and the logs do not extend to this depth. However, these differences may be due to the 500 m distance between the location of the TEM soundings and available boreholes (Fig. 5.32).

**Site URL:** The closest borehole with lithology and fracture logs to the TEM location at Site URL is URL-2, located approximately 600 m north of the centre of the transmitter loop (Fig. 5.34). Fig. 5.35 compares the geophysical logs from borehole URL-2 with TEM models. The borehole log M12 was not used because of the absence of the lithology and fracture logs. The logs show a fractured zone with prominent fractures at 90 and 170 m depth. TEM models for the north line include a embedded conductive layer at 120-170 m depth, which could correspond to the fracture zone in URL-2. The depth of the top of this embedded conductive layer in the TEM model can be resolved (Appendix H) and agrees quite well with the borehole results.
Figure 5.32 TEM sounding location at Site D. The red square shows the transmitter loop location and the red lines show the locations of the receiver profiles. Mapped lineaments are shown as brown lines (modified from Ejekam et al., 1990).
Figure 5.33 Comparison of geophysical logs from nearby boreholes with 1D TEM inversion model at Site D. Shown are caliper, temperature, fluid resistivity, 40 cm normal resistivity, TEM layered-inversion model and fracture logs with generalized lithology represented as colour fill. (a) Borehole WD-1 with TEM model for station N200, (b) Borehole WD-2 with TEM model for station N200. (From Wu et al. 1999).
Figure 5.34 TEM sounding location at Site URL. The red square shows the transmitter loop location and the red lines show the locations of the receiver profiles. One mapped thrust fault is shown as a brown line (modified from Sikorsky, 1996).
Figure 5.35 Comparison of geophysical logs from nearby boreholes URL-2 with 1D TEM inversion model of station W280 at Site URL. Showed are caliper, temperature, fluid resistivity, 40 cm normal resistivity, TEM layered-inversion model (and fracture logs with generalized lithology represented as color fill). (From Wu et al. 1999).
The borehole URL-2 log shows that the resistivity gradually decreases at 180-250 m depth, is relatively constant over the depth range of 250-400 m, and decreases again beginning at the depth of 400 m. The depth to the lower conductive layer of TEM models is about 400 m, which can be resolved (Fig.5.28). The TEM models place the lower conductive layer at the bottom of the gradual change observed in the log.

More quantitative analysis of logs and TEM responses will be discussed in Chapter 6. Combined with geological and hydrological information, the TEM results can be used to interpret the geological structure.

5.7.2 Geological Interpretation of TEM Results

As discussed in Chapter 2, the Lac du Bonnet Batholith consists of various granitic rocks (Brown et al. 1989), which are overlain by thin glacio-lacustrine sediments and tills (Thorne 1990). The granitic rocks are relatively homogeneous over a large area and depth (Brown et al. 1989; Fig. 1.5). Large-scale faulting is rare and limited to the eastern part of the batholith (Everitt et al. 1996). Low to intermediate-dipping fracture zones or faults with dips of 10°-30° are common at surface and can extend from the surface to at least 800 m depth (Brown et al. 1989), for example, the fracture zones FZ1, FZ2 and FZ3 in the URL lease area (Fig. 2.9). Subvertical fractures are common in outcrop but below 200 m are limited to the margins of the low-dipping fracture zones (Fig. 2.9; Brown et al. 1989). Beneath the major fracture zones, the fractures are subhorizontal.

The groundwater is comparatively fresh above 250 m (TDS <0.3 g/L) (Fig. 2.10a). The near-surface sub-vertical fractures provide pathways for fresh meteoric water to penetrate the
granite, and enter groundwater flow systems in the sub-horizontal fracture zones (Everitt et al. 1996). Beneath 250 m depth, the salinity increases with depth (up to 50 g/l TDS) because of the dissolution of soluble salts in the host rock (Gascoyne et al. 1994).

**Conductive Surface Layer**

The thin, conductive surface layer with resistivity 10-30 Ωm and thickness 0.5-2.0 m is present at all receiver stations at Site B and some receiver stations at Site D (Fig. 5.23 and 5.26). At Site URL, the resistivity of this layer is greater than 1000 Ωm, and its thickness is less than 10 m at all stations. The upper layer of the resistivity model determined from the TEM survey at URL is consistent with EM31 results (>500 Ωm) determined at URL during the University of Manitoba geophysical field school in 1995 (I. Ferguson, pers. comm. 2001). A TEM survey completed in 1987 (Hoover et al. 1987) also showed a discontinuous, conductive overburden in the G lease area (Fig. 5.1).

The conductive surface layer is associated with the various near-surface lithologies (glacio-lacustrine sediments, tills and granitoids), weathered to different extents depending on near-surface fracturing and mineralogy. This layer is interpreted as being thin, spatially-variable soils and organic material in open fractures associated weathered rock at the sites (Thorne 1990; Thorne and Gascoyne 1995).

**Resistive Second Layer**

The resistivity of the second layer is approximately 3000 Ωm at Sites B and D, and greater than 50,000 Ωm at Site URL (Table 5.11). The depth to its base is 50-180 m at Site
B, 300-400 m at Site D, and 350-460 m at Site URL, with some spatial variation at each site. On most profiles, there is a smooth 50 to 100 m change in the thickness of the layer over a 400 m length of the profile. Localized features of 40 to 80 m lateral extent exist in which there is a 50 to 200 m change in thickness (Fig. 5.23, 5.26, 5.29).

This resistive layer is interpreted as corresponding to a zone of sub-vertical fractures saturated with fresh groundwater percolating from the surface. The variations in layer depth along profiles at single station indicate the spatial variation in the depth to the base of the sub-vertical fractures. The results of Moon et al. (1993) also suggest some variability in the depth to the base of fracturing.

The resistivity values of 1500-50,000 Ωm determined from TEM survey are full within the reported range for Precambrian granite (300-1000,000 Ωm in Fig. 1.1). The results are also comparable to the resistivity of near-surface Precambrian rocks of the Superior Province in the north-central Abitibi Subprovince of >10,000 Ωm (Zhang et al. 1995); near East Bull Lake of >5000 Ωm (Kurtz et al. 1986); and in the English River Sub-province of several thousand to 100,000 Ωm (Koziar and Strangway 1978).

The difference in resistivity of this layer between Sites B and D (>3000 Ωm) and Site URL (~50,000 Ωm) may be due to the different density of fracturing of the rocks (and thus the overall porosity and permeability) or to the different groundwater salinity. The difference in porosity required to explain the observed results can be examined in terms of Archie's law (Eq. 1.1).

Table 5.12 lists the parameters of fresh and saline water layers used to calculate porosity. The TDS used are the average values for the layers from Fig. 2.10, and Eq (1.2) is
used to calculate the fluid resistivity. The values of the constant parameters in Archie’s law were taken to be: \( a = 1 \) and \( m = 1.5 \) which are typical values for the low porosity rocks (<4%), for example dense igneous and carbonate rocks (Ward and Hohmann 1990; Keller 1988). The saturation is set as 100% \( (S=1) \).

Table 5.12 Possible parameters of fresh-water layer and saline-water layer in URL area.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Fresh-water Layer</th>
<th>Saline-water Layer</th>
</tr>
</thead>
<tbody>
<tr>
<td>TDS (g/L)</td>
<td>0.3</td>
<td>50</td>
</tr>
<tr>
<td>Salinity C (ppm)</td>
<td>300</td>
<td>50000</td>
</tr>
<tr>
<td>Fluid Resistivity ( \rho_f (\Omega m) = 6/C (ppm) )</td>
<td>20</td>
<td>0.12</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Fresh-water Layer</th>
<th>Saline-water Layer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Resistivity of Rock ( \rho_r (\Omega m) ) (from TEM model)</td>
<td>1500-4000</td>
<td>&gt;50000</td>
</tr>
<tr>
<td>Site B, D Site URL</td>
<td>Site B, D Site URL</td>
<td>Site B, D Site URL</td>
</tr>
<tr>
<td>Apparent Porosity ( \phi (%) ) ( \rho_r = a \phi^n S^n \rho_f )</td>
<td>2.93-5.62</td>
<td>&lt;0.54</td>
</tr>
<tr>
<td></td>
<td>0.45-1.8</td>
<td>~0.15</td>
</tr>
</tbody>
</table>

From Archie’s Law, the apparent porosity is 2.93%–5.92% at Site B and D and <0.54% at Site URL. The apparent porosity at Site B and D is 5.5–10.4 times of that at Site URL. The apparent porosity values at Site URL (from 0.54% to 0.15%) are close to the porosity values measured by Frost et al. (1995) in URL: from 0.3% in the upper 1000 m of moderately fracture rock to 0.1% at 3200 m depth and below (sparsely fracture rock).

From the laboratory experiments, the typical porosity of the unfractured granite is 0.26%–0.43% (Katsube and Mareschal 1993) with these measurements determined from
granites formed at mid-crustal depths. The porosity of the unfractured granite is clearly lower than the apparent porosity of the fractured granite at Site B and D in Lac du Bonnet Batholith.

The TEM results are comparable with the results of other geophysical surveys in the URL lease area. A dipole-dipole array survey performed in 1980 detected significant decreases in resistivity at depths ranging from 300 m to 700 m (Paterson and Watson 1981). Borehole logs and GPR surveys have shown a reduced resistivity associated with saline water at depths greater than 300-400 m (Stevens, et al. 1995). The present TEM study shows an interface between the resistive and conductive layers at about 400 m depth at Site URL.

The fracture zone FZ2 is exposed at the surface in the northwest of the URL lease area (Fig. 5.1b) and has a southeasterly dip of 20°-25° (Stevens et al. 1995). The depth of FZ2 is about 40 m at borehole M-10 and about 200 m at URL shaft (Stevens et al. 1995; Fig. 2.9). The resistivity interface from the feasibility study Site 1 is at about 200 m depth, which is similar to the depth of the FZ2 in URL shaft. The FZ2 splays to form multiple fracture zones with increasing depth and the shallow fracture zones vanish toward southeast (Fig. 2.9). The dipping angle of the FZ2 is larger in southeast of the URL lease area than in northwest. Therefore the resistivity interface at about 400 m depth at Site URL (Fig. 5.1b) determined from TEM survey likely corresponds to the location of fracture zone FZ2.

**Bottom Conductive Layer**

The resistivity of the basal conductive layer is 50-400 Ωm at Sites B and D, and 1000-2000 Ωm at Site URL. At Site B, this layer is subdivided into an upper layer with resistivity
50-400 Ωm and a lower layer with resistivity less than 85 Ωm. The base of the upper layer lies at a depth of 300-350 m over most of the two profiles at that site, but is shallower near the west, east, and north ends. The resistivity of the bottom conductive layer at Site URL is significantly larger than that at Sites B and D. This conductive layer is interpreted as corresponding to sparsely fractured rock (containing a few sub-horizontal fractures) saturated with saline groundwater or brine.

The difference in the resistivity of this layer between Sites B and D (50-400 Ωm) and Site URL (1000-2000 Ωm) may also be explained in terms of the different density of fracturing of the rocks. The apparent porosity of saturated bottom conductive layer calculated by the Archie’s law is shown in Table 5.12. The apparent porosity is 0.45%-1.8% at Sites B and D, which is 3-12 times the apparent porosity at Site URL (~0.15%).

**Embedded Conductive Layer**

For most stations at Site D and two stations at Site URL (N180 and N200), there is an embedded conductive layer within the resistive layer. The conductance of this layer is approximately 0.23-0.35 S at Site D and approximately 0.04-0.06 S at Site URL. The range of depths to this layer is 100-200 m at Site D, and 120-170 m at Site URL. The depth and conductance of this layer are well resolved by the TEM data, but the individual thickness and resistivity are not resolved. The thin conductive layer is interpreted as representing an approximation of a more complex resistivity structure containing more finely laminated conductive zones.

At Site D, bedrock mapping to the north of the survey area has revealed surface traces
of several low-dipping fracture zones (Fig. 5.32). Borehole WD-1 and WD-2 show the fracture zones at 100 m and 70 m depth respectively. These observations clearly indicate there are some low-dipping fracture zones at Site D. Based on the geometry of these fracture zones, the low-dipping fracture zones could extend into the area investigated by the TEM survey. Quantitative comparison of the conductance in the fracture zones between the borehole log and TEM model will be discussed in Chapter 6.

At Site URL, the embedded conductive layer is interpreted as being a sub-horizontal fracture zone FZ3 (Fig. 2.9), which is a localized fracture zone that is observed at about 100 m depth at URL shaft. The GPR results also showed a reflection event at about 15-20 m depth to the southwest of borehole M-10 (Stevens et al. 1995). However, there is no direct surface evidence for this interpretation. A conductive layer was also noted in Schlumberger soundings (110-220m depth) recorded in 1980 at URL (Paterson and Watson 1981). The Schlumberger soundings did not provide information about deeper geoelectric structure. However, the penetration depth of a DC resistivity survey using Schlumberger soundings is limited by electrical power and electrode spacing. The depth to the top of the conductive layer as determined by DC resistivity survey is consistent with the depth to the top of the embedded conductive layer in the present TEM study and of the conductive layer in the feasibility study.

As described by McCrank (1985) and Sikorsky (1996), the filling materials of the fracture zone could be calcite, chlorite, hematite, limonite and sulfide. These materials along with the fluids in the fractures can decrease the resistivity.
Local Complex Resistivity Structure

The TEM observations for Sites B and D included a number of stations that have different responses from nearby stations (Fig. 5.16 and 5.17). These anomalous responses sometimes occurred at single stations (e.g., W140 at Site B) and sometimes extended over several stations (on the western and eastern lines beyond W160 and E240 at Site B, on the western, eastern and southern lines beyond W180, S240 and E200 at Site D). The responses from these stations could not be fitted well by the layered models. This observation supports the interpretation that the anomalous responses are associated with more complex local resistivity structures that need 2D and/or 3D modelling.

The presence of the local resistivity structures is consistent with features of the surface geology. The site map (Fig. 5.32) shows the surface contact of a fracture zone to the area of the west line beyond W180 of Site D. The survey notes for this site (Fig. 5.2c) also recorded the presence of a cliff at W200 which is likely associated with the same feature. These observations permit the firm conclusion that the fracture zone at this location creates a local 2D or 3D resistivity structure. The conductive zone near E200 may also be associated with a second fracture zone which is mapped to the north of the east line.

5.7.3 Interpretation of Results at Site A

The TEM response at Site A including the central loop sounding is characterized by multiple sign reversals. Theoretical studies (e.g., Spies & Frischknecht 1991) show that for a 1D resistivity model, the central loop response does not have any sign changes, and that offset responses have multiple sign changes only when there are strong resistivity contrasts
at depths.

It was impossible to produce satisfactory Occam or layer inversion models to fit observed data for any of the stations at Site A. If the sign changes are neglected, the inversion of the central loop data for Site A using TEMIX produces a three-layer resistivity model similar to those for the other sites, with a depth to the lower conducting layer of approximately 300 m. The magnitude of the responses of the resulting model are relatively close to the magnitude of the observed apparent resistivity and voltage data. Maris et al. (1997) determined a three-layer model with 370±30 m depth to the base conducting layer, while also neglecting sign changes in the data. However, the simple three-layer model represents only an approximation of the true structure.

Boreholes WA-1 and WA-2 are located about 150 m east of the centre of the transmitter loop (Fig. 5.36). Figure 5.37 compares the geophysical logs with the central loop TEM models. The geophysical logs show a gradual decrease in resistivity between 300 and 400 depth. These results agree quite well with the TEM results which placed the depth to the bottom conducting layer at 300 m. The logs indicate fracturing at 250 to 300 m depth, but this is not resolved in the TEM models.

TEMIX can produce forward model responses for multi-layer structure containing the multiple sign reversals. The exact time of each sign reversal depends on a number of the model parameters, for example, the thickness of each layer and resistivity contrast between layers. Figure 5.38 shows the sign reversals from a three-layer model for a 280 m offset. The time of the sign reversals around 0.1 ms depends on the resistivity contrast between the second and third layers and the thickness of the second layer.
Figure 5.36 TEM sounding location at Site A. The red square shows the transmitter loop location and the red lines show the locations of the receiver profiles. Mapped lineaments are shown as brown lines (modified from Sikorsky, 1996).
Figure 5.37 Comparison of geophysical logs from boreholes near Site A with central loop inversion model. Shown are caliper, temperature, fluid resistivity, 40 cm normal resistivity, TEM central loop inversion model and fracture logs with generalized lithology represented as colour fill. (a) Borehole WA-1, (b) Borehole WB-2. (From Wu et al. 1999).
Resistivity Model

Resistivity (ohm.m)

Time (ms)

Normalized Voltage (nV/A.m^2)

Figure 5.38 (a) Three-layer resistivity model, (b) the observed data (square and diamond) and model response (solid line) at station offset N280 at Site A. Transmitter loop size is 100 m. There are two sign reversals in the forward modelling response. The time of second sign reversal is close to the sign reversal time of the observed data.
Therefore, the multiple sign reversals at Site A, which can not be fully modelled by a layered conductivity structure, imply a more complex geoelectric structure or more complex EM induction process than occurring at the other sites. The sign reversals may be due to significant anisotropy, two or three-dimensional components in the structure, and/or induced polarization (IP) effects.

In TEM sounding, polarizable conductive structures in the underground can produce multiple sign reversals (i.e. IP effect) (Smith and West 1988; 1989; Zadorozhnaya and Lepeshkin 1998). In other studies strong polarization has been observed in thin layers speculated to contain graphite, carbonaceous shales, iron oxides, clays, or clayey-sands (Tyne and Dagger 1990). In the Lac du Bonnet batholith, a possible source of polarization is iron oxides and clays within the fractures. At Site A, the outcrop is noted to be highly fractured. McCrank (1985) and Sikorsky (1996) noted the presence of chlorite and sulfide in the fractures in the batholith.

Site A is close to the northern boundary of the Lac du Bonnet Batholith. It is possible that compositional or structural features associated with this boundary may have influenced the resistivity structure. There is also a long and linear valley which is probably fracture-related that parallels the east line.

5.8 Recommendations for Future Work

The TEM results at Lac du Bonnet indicate that the GEONICS PROTEM 47 instrument, in combination with a 100 m transmitter loop, provides a suitable TEM system for mapping the resistivity structure from the surface to depths of 300-500 m. The
quantitative relationship between the TEM models and borehole resistivity logs will be discussed in Chapter 6.

The availability of early-time results (\( <20 \mu s \)) assisted in the resolution of the very resistive layer and the 100 m transmitter loop provided satisfactory signal levels and penetration of signals through heterogeneous near-surface layers in the resistive environment of the Lac du Bonnet Batholith. For deeper penetration, other available TEM systems with greater transmitter power, larger transmitter loop and longer time gates may be useful (EM37).

Hoover et al. (1987) mentioned that the discontinuous and conductive overburden could distort the TEM data. The results of the present study also show the effects of distortion by near-surface features. MT distortion was discussed in detail in Chapter 4. In a future analysis the TEM distortion could be examined in more detail.

Further analysis is required to fully understand the TEM responses at Site A. As discussed in last section, the multiple sign reversals at Site A imply a more complex geoelectric structure or more complex EM induction process than at the other sites. This analysis may require consideration of the effects of anisotropy, 2D/3D components of the resistivity structure, and IP effects. Three-component \((x, y, z)\) receiver coils can be used to discriminate between IP and 2D/3D effects.

Figures 5.23 and 5.26 show the depth to the base of resistive layer decreases to the northeast at Site B and to the southeast at Site D. As discussed in section 5.6.3, the decrease of this depth at offsets could be from the effect of the overburden which is difficult to separate it from the response. There is a possibility that this result could be due to the large-
scale effects of fractures. Greater confidence can be placed on the TEM results at stations close to the centre of the transmitter loop. Therefore, in future studies 2D and 3D forward modelling and inversion should be used to examine the effects of the fractures and structures in the Lac du Bonnet area. Application of these techniques was beyond the scope of the current work.

The cross-hole seismic study by Hayles et al (1996) in the URL shaft indicates an anisotropy of the P-wave velocity. The velocity in the vertical direction is about 4% faster than those in the horizontal direction. This response is more or less constant throughout the URL lease area. The anisotropy is probably caused by vertical/steeply dipping fractures and microfractures in the rock (Hayles et al 1996). Does this seismic anisotropy from the vertical fracture zones produce an electric anisotropy and affect the observation of resistivity? This possibility needs to be considered in future research.
Chapter 6  Comparison of MT and TEM Results with Borehole Logs

The results presented in the last two chapters have shown that the MT method can delineate the variation of resistivity from the near-surface to the upper mantle (>100 km) and that the TEM method (using a PROTEM47 and a 100 m transmitter loop) can delineate the variation of resistivity at depths less than 600 m. The MT method has also defined the thickness variation of conductive Phanerozoic sedimentary rocks from several tens metres to 1000 m depth along SNORCLE corridor 1 and 1A. In order to understand the relationship between the MT and TEM responses and the fine-scale resistivity variations in each environment, several electrical geophysical logs will be examined in detail. The MT or TEM models will be compared quantitatively with the well log data, e.g. the observed MT or TEM responses will be compared with the synthetic responses computed from the resistivity logs. The effect of transverse anisotropy associated with fine-scale layering in each environment will also be examined in detail.

6.1 Comparison of Large-scale Structures from Well Logs and MT Models

Based on 125 well logs archived by the National Energy Board (1980), a sedimentary isopach map of the surface of Precambrian basement along SNORCLE corridor 1 and 1A has been constructed (Fig. 6.1). The depth to the Precambrian basement was picked from each
Figure 5.1 Sedimentary isopach map for the surface of Precambrian basement. Data acquired from borehole logs (National Energy Board, 1980).
log. These values are contoured as a function of latitude and longitude. A kriging algorithm (variogram model) was used to interpolate the data and the program SURFER (Keckler 1994) was used for contouring and plotting. The edge of Phanerozoic sedimentary rocks in the east is defined from Fig. 2.2.

The thickness of the Phanerozoic sedimentary rocks is over 1000 m in the southwest of Corridor 1 and decreases towards northeast. The sedimentary rock is relatively thin around MT sites 142 and 141, where it has a thickness of <700 m.

Figure 6.2 compares the depth to the base of the Phanerozoic sedimentary rocks determined from 1D MT inversion models along Corridor 1 and 1A (Fig. 4.22) and from some well logs which are relatively close to the MT sites (Table 6.1).

**Table 6.1** Location of well logs and MT sites (National Energy Board 1980).

<table>
<thead>
<tr>
<th>Well</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Depth to Base (m)</th>
<th>MT site Latitude (N)</th>
<th>Longitude (W)</th>
<th>Depth to Base (m)</th>
<th>Distance to MT site (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Liard River D-75</td>
<td>61°14'06&quot;</td>
<td>122°44'59&quot;</td>
<td>575</td>
<td>142 61°12'11&quot;</td>
<td>122°41'37&quot;</td>
<td>300</td>
<td>4.6</td>
</tr>
<tr>
<td>Jean Marie E-7</td>
<td>61°16'20&quot;</td>
<td>120°46'45&quot;</td>
<td>950</td>
<td>134 61°19'04&quot;</td>
<td>120°40'22&quot;</td>
<td>1165</td>
<td>7.6</td>
</tr>
<tr>
<td>Foetus Lake C-49</td>
<td>61°05'49&quot;</td>
<td>118°39'37&quot;</td>
<td>793</td>
<td>127 61°05'32&quot;</td>
<td>118°44'12&quot;</td>
<td>720</td>
<td>4.1</td>
</tr>
<tr>
<td>Mill Lake J-74</td>
<td>61°13'38&quot;</td>
<td>118°13'48&quot;</td>
<td>694</td>
<td>124 61°06'52&quot;</td>
<td>118°00'28&quot;</td>
<td>750</td>
<td>17</td>
</tr>
<tr>
<td>Calstan Mills Lake C-3</td>
<td>61°12'15&quot;</td>
<td>118°01'00&quot;</td>
<td>740</td>
<td>124 61°06'52&quot;</td>
<td>118°00'28&quot;</td>
<td>750</td>
<td>10</td>
</tr>
<tr>
<td>Mill Lake A-70</td>
<td>61°09'07&quot;</td>
<td>117°56'29&quot;</td>
<td>730</td>
<td>123 61°08'27&quot;</td>
<td>117°44'56&quot;</td>
<td>620</td>
<td>10.4</td>
</tr>
<tr>
<td>Mill Lake L-10</td>
<td>61°09'41&quot;</td>
<td>117°31'39&quot;</td>
<td>625</td>
<td>122 61°04'27&quot;</td>
<td>117°32'36&quot;</td>
<td>580</td>
<td>9.7</td>
</tr>
<tr>
<td>Mill Lake I-57</td>
<td>61°16'37&quot;</td>
<td>117°39'36&quot;</td>
<td>580</td>
<td>121 61°20'11&quot;</td>
<td>117°33'08&quot;</td>
<td>520</td>
<td>8.7</td>
</tr>
<tr>
<td>IOE Province A-47</td>
<td>61°30'00&quot;</td>
<td>117°15'00&quot;</td>
<td>490</td>
<td>120 61°28'28&quot;</td>
<td>117°18'41&quot;</td>
<td>330</td>
<td>4.3</td>
</tr>
<tr>
<td>Mill Lake B-41</td>
<td>61°40'06&quot;</td>
<td>116°53'17&quot;</td>
<td>410</td>
<td>118 61°45'06&quot;</td>
<td>116°49'23&quot;</td>
<td>290</td>
<td>9.9</td>
</tr>
<tr>
<td>Mill Lake C-12</td>
<td>61°09'41&quot;</td>
<td>117°31'39&quot;</td>
<td>660</td>
<td>146 60°56'22&quot;</td>
<td>117°05'46&quot;</td>
<td>700</td>
<td>33.9</td>
</tr>
<tr>
<td>Horn River B-52</td>
<td>60°51'04&quot;</td>
<td>115°40'17&quot;</td>
<td>610</td>
<td>152 60°43'35&quot;</td>
<td>115°33'08&quot;</td>
<td>710</td>
<td>15.3</td>
</tr>
</tbody>
</table>
Figure 6.2 Comparison of the depth to the base of the Phanerozoic sedimentary rocks from well log data (symbol square) and that from 1D MT inversion models along Corridor 1 and 1A (line).
The data from the well logs are close to the 1D MT inversion results. However, there is some misfit between well log and 1D MT results. The misfit is typically ±100 m and varies randomly along the corridors. There could be several reasons for the misfit:

1. The distance between the MT sites and the associated wells (Table 6.1).
2. The limited resolution of MT data (discussed later).
3. The static shift of MT responses (discussed later).
4. The depth limit of the well logs. For example, well log E-7 did not reach the Precambrian rocks.

However, the similarity of well log and MT data indicates MT data can be used to determine the approximate thickness of the conductive Phanerozoic sedimentary rocks overlying the resistive Precambrian rocks. More detailed analysis of several logs along Corridor 1 will be done to examine the fine-scale structures.

6.2 Detailed Comparison of Electrical Logs and MT Responses

6.2.1 MT Sites and Well Logs Used in Comparison

Three MT sites along Corridor 1 were used to examine the fine-scale structures: site 134 in the Fort Simpson terrane, site 124 in the Hottah terrane, and site 120 in the Great Bear Magmatic Arc. The three well logs used in comparison are wells E-7, C-3 and A-47 (Table 6.1). They are the closest wells to each MT site (the distance between MT site and well log is less than 10 km, Table 6.1). Table 6.2 lists the stratigraphic units and the depth to top of each unit recorded in each well. Sandstone, shale and limestone are main sedimentary rocks above the Precambrian basement (granite).
Table 6.2 Depth to the top of stratigraphic units for three logs (National Energy Board 1980; Hills et al. 1981).

<table>
<thead>
<tr>
<th>Stratigraphic Unit</th>
<th>Lithology</th>
<th>Depth to top from well log (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>E-7</td>
</tr>
<tr>
<td>Fort Simpson Formation</td>
<td>grey shale, mudstone,</td>
<td>215</td>
</tr>
<tr>
<td></td>
<td>silty or sandy</td>
<td></td>
</tr>
<tr>
<td>1st Black Shale</td>
<td>shale</td>
<td>772</td>
</tr>
<tr>
<td>Horn River Formation</td>
<td>sandy limestone</td>
<td>791</td>
</tr>
<tr>
<td>Lonely Bay Formation</td>
<td>limestone, dolostone</td>
<td>814</td>
</tr>
<tr>
<td>Willow Lake Formation</td>
<td>finely crystalline dolostone</td>
<td>864</td>
</tr>
<tr>
<td>Ebbutt Member</td>
<td>shale, siltstone, sandstone</td>
<td>877</td>
</tr>
<tr>
<td>Bear Rock Formation</td>
<td>dolomitic limestone</td>
<td>907</td>
</tr>
<tr>
<td>Chinchaga Formation</td>
<td>dolostone, anhydrite</td>
<td>-</td>
</tr>
<tr>
<td>Red Beds</td>
<td>silty sandstone</td>
<td>-</td>
</tr>
<tr>
<td>Basalt Sandstone</td>
<td>sandstone</td>
<td>940</td>
</tr>
<tr>
<td>Precambrian</td>
<td>granite</td>
<td>uncertain</td>
</tr>
</tbody>
</table>

Figure 6.3 shows 40 cm induction electrical logs and sonic logs from three wells. The induction logs clearly show three zones: a relatively resistive surface zone (~300 Ωm, not visible in A-47), a very conductive second zone (~8 Ωm), and a very resistive bottom zone (>1000 Ωm). The depth to the top of the bottom resistive layer decreases from west to east. The sonic logs also show a three-layer structure: a surface layer with variable velocity, a middle layer with ~3000 m/s velocity, and a bottom layer with high velocity (~5800 m/s). The depth of the high-velocity bottom layer decreases from west to east.

6.2.2 Examination of Dimensionality of High-frequency MT Responses

As discussed in section 4.7, the near-surface structure along Corridor 1 and 1A is approximately one-dimensional. In order to make it possible to quantitatively compare the
Figure 6.3 Three well logs from the southwest Northwest Territories: Jean Marie E-7, Mills Lake C-3 and Province A-47. (a) Resistivity curves from induction logs, (b) velocity curves from sonic logs. Dash line: values are out of the recording range (provided by International Datasahre).
(b) Sonic logs
well log data with MT responses, the dimensionality of high-frequency MT responses needs to be examined. The influence from 2D structures will make the comparison between MT models and well logs more difficult.

Figure 6.4 shows the impedance polar diagrams of impedance $Z_{xy}$ and $Z_{zx}$ at a range of periods for these three sites. The impedance polar diagrams were obtained by rotating the impedance in steps and plotting the results on a polar diagram. The marked angle is the one that maximizes or minimizes some combination of the impedance magnitude (Vozoff 1991). In the absence of galvanic distortion the impedance polar diagram can be used to find the rotation angle $\theta$ between measurement direction and geoelectrical strike. In Fig. 6.4, the polar diagrams are approximately circular for impedance $Z_{xy}$ and are very small for impedance $Z_{zx}$. The impedance response is therefore only very weakly dependent on the rotation angle. This result indicates an approximately 1D shallow structure.

As discussed in Chapter 4, for Corridor 1 and 1A sites the TE mode was defined to correspond to the electric fields aligned parallel N34°E and the TM mode to correspond to the electric fields perpendicular this direction. This strike direction is defined by the longer period responses related to the underlying Precambrian rocks. Figures 6.5-6.7 show the TE and TM mode apparent resistivities and phases at sites 134, 124 and 120. At these three sites, TE and TM apparent resistivities and phases are close to each other at periods $<1$ s. This response also indicates an approximately 1D shallow structure. As the short period data are close to 1D, the determinant response (Eq. 3.45) will be used to define the shallow structure in the following study.

The observed TE and TM phases separate slightly at periods $<3 \times 10^{-3}$ s for site 134,
Figure 6.4 Polar diagrams of impedance $Z_{xy}$ and $Z_{xx}$ versus periods at sites 134, 124 and 120.
and at periods $<6 \times 10^{-3}$ s for sites 120. This response indicates a locally 2D structure at $<40$ m depth for site 134 and at $<20$ m depth for site 120 (Table 4.3). For site 124, the TE and TM phases are close at periods of $10^{-1}$-1 s, but the TE apparent resistivity is higher than the TM apparent resistivity. This response implies a static shift at this site. Table 6.3 lists the static shift factors (Eq. 3.45) for these three sites.

Table 6.3 The static shift factors for $\rho_{TE}$ and $\rho_{TM}$ at three sites from the Sternberg and Jones methods.

<table>
<thead>
<tr>
<th>Station</th>
<th>Sternberg method</th>
<th>Jones method</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Factor of $\rho_{TE}$</td>
<td>Factor of $\rho_{TM}$</td>
</tr>
<tr>
<td>134</td>
<td>0.92</td>
<td>0.93</td>
</tr>
<tr>
<td>124</td>
<td>0.85</td>
<td>0.95</td>
</tr>
<tr>
<td>120</td>
<td>1.06</td>
<td>0.99</td>
</tr>
</tbody>
</table>

6.2.3 Comparison of Synthetic MT Responses from Induction Logs with Observed MT Responses

Synthetic MT responses from the induction logs will be compared with the observed MT responses in order to examine the MT results (Fig. 6.5-6.7). The resistivity curves from the induction logs (Fig. 6.2a) were digitized at 0.2 m spacing, so that layered models (consisting of $>3000$ layers) could be created. The resistivity value at the base of the induction log was used as the resistivity of bottom layer (half-space) for the calculations, and the average resistivity of the near surface layers in the log was used as the resistivity of the first layer for the calculations.

At site 134 (Fig. 6.5), the observed MT responses are similar to the synthetic responses of log E-7 at periods $>0.1$ s. In the period range 0.004-0.1 s, the observed apparent
Figure 6.5 Comparison of the observed TE and TM apparent resistivity and phase at site 134 with MT response of well log Jean Marie E-7.
Figure 6.6 Comparison of the observed TE and TM apparent resistivity and phase at site 124 with MT response of well log Mill Lake C-3 where the resistivity of half-space beneath the well log depth used include the last data of well log (solid line) and 3500 ohm.m (dash line).
Figure 6.7 Comparison of the observed TE and TM apparent resistivity and phase at site 120 with MT response of well log Province A-47 where the resistivity of half-space beneath the well log depth used include the last data of well log (solid line) and 3500 ohm.m (dash line).
resistivity is smaller than the synthetic MT responses (by a factor of 0.6) and the observed phase is lower (by 5°). At periods <0.004 s, the observed phase is higher than the synthetic response (by 5°). These differences indicate that the well log data correspond to a location with a thicker and more resistive surface layer than the MT data (Table 4.3). The reason for this difference could be the 7.6 km distance between the MT site and the well (Table 6.1).

At site 124, the observed TE apparent resistivity is higher than the TM apparent resistivity (Fig. 6.6). This is caused by the static shift of 0.85 in the TE apparent resistivity (Table 6.1). The synthetic apparent resistivity for log C-3 is closer to the observed TM apparent resistivity. This result implies the static shift of MT responses can be corrected using the well log information. The observed phase is lower than the synthetic response (by 3°) at periods <0.05 s, indicating the conductive surface layer in the well log is thicker than in the MT model. This difference may again be due to the distance between the MT site and the well C-3 (10 km, Table 6.1).

At site 120, the observed MT responses are close to the synthetic responses of log A-47 at periods <0.4 s (Fig. 6.7). At periods >0.4 s, the observed apparent resistivity is higher than the synthetic response and the observed phase is lower. The relatively low apparent resistivity and high phase of the synthetic response indicates the layered model from the induction log has a more conductive bottom layer. The resistivity of bottom layer used in forward modelling is 308 Ωm. This value is the deepest resistivity value in the induction log and is lower than the typical resistivity of the Precambrian rocks (Fig. 1.1). As induction logs are not sensitive to resistivities larger than a few hundreds Ωm, it is possible that the induction log underestimates the true resistivity at this depth. If a higher bottom resistivity
value is chosen, for example 3500 Ωm (a typical resistivity of granite), the synthetic responses of long periods fit the observed MT data better (Fig. 6.7).

The similarity of observed MT responses and the synthetic response from nearby induction logs indicates the MT responses have defined the correct form of responses corresponding to the underlying resistivity structure. Because of equivalence the results considered above do not prove the MT response can resolve the fine details of resistivity structure.

6.2.4 Comparison of Induction Logs and MT Inversion Models

Quantitative comparison of the induction logs and the MT inversion models will be discussed in this section. As discussed in section 6.2.2, the near-surface structure is approximately 1D. Therefore, the determinant response was inverted to determine the shallow structure.

The 1D inversion methods used include the Occam, Marquardt and Fisher methods. For each site, the inversion models from the three inversion methods are very similar. Figure 6.8 compares the Fischer inversion models at three sites with the induction logs.

Induction Log for Well E-7 and MT Mode for Site 134

The induction log for well E-7 (Fig. 6.8a) shows a relatively high resistivity (~50 Ωm) from the surface to 215 m depth. The lithology of this layer was not recorded with the log. Based on comparisons with nearby logs (IOE Sun Blackstone E-72), the lithology is expected to be limestone and sandy limestone (the Jean Marie member and Redknife formation, Hills
Figure 6.8 Comparison of the 40 cm induction log resistivity and the MT model (heavy solid line). (a) well Jean Marie E-7 with MT site 134, (b) well Mill Lake C-3 with MT site 124. (c) well Province A-47 with MT site 120. Legend is same as Figure 6.3a. Dash line: the MT model after static shift correction. The geological legend is same with Fig. 6.3.
et al. 1981). From 215 to 810 m depth, the sedimentary rocks are very uniform and conductive (~8 Ωm), and consist of shale, mudstone, silty or sandy limestone of the Fort Simpson Formation, 1" Black Shale and Horn River Formation. From 814 to 864 m depth, the resistivity is about 300 Ωm in the reef carbonates of the Lonely Bay Formation (Table 6.2). From 864 to 907 m depth, the resistivity is about 25 Ωm in the Willow Lake Formation and Ebbutt Member. Between 907 m and 940 m depth the resistivity is >500 Ωm in dominantly limestone and dolostone units. The variation in resistivity between 810-940 m depth may be due in part to variation of the fluid in the pores since the sedimentary rocks have similar velocity (Fig. 6.3b). Beneath 940 m, in Cambrian quartzite, the resistivity increases significantly. This well did not reach Precambrian rocks.

At site 134, the inversion model provides a three-layered structure: a relatively resistive surface layer (60 Ωm), a conductive second layer (10 Ωm) and a resistive bottom layer (>1000 Ωm). The relatively resistive first layer with 110 m thickness is thinner than the silty and limestone layer (and corresponding relatively high resistivity) in well log E-7 (210 m). This may be due to the 7.6 km distance between the well log and MT site. Comparison of the observed and synthetic MT responses also suggested a thicker surface layer at the well site. Another reason for the difference may be the instrument resolution and environmental noise. Most MT recordings have relatively larger error bars at periods less than 10⁻³ s in the AMT dead band. These points were removed prior to processing (Fig. 6.5). Therefore, the resolution is relatively low at depths above 100 m in 60 Ωm sedimentary rocks.

The base of the conductive second layer in the MT model is at ~1165 m depth, which is deeper than the maximum log depth of 950 m. Because the well log E-7 only reached the
Cambrian stratigraphy and did not reach the Precambrian rocks, it is not possible to complete a definitive comparison of the thickness of the conductive sedimentary rocks and the second layer in the MT model. However, the results show MT data can not resolve the resistive and thin (50 m thick) Lonely Bay reef between 810-860 m because of the limited sensitivity of the MT response to thin resistive layer (Chapter 3). The MT method also did not resolve the resistive Cambrian quartzite as a separate unit.

Table 6.4 shows the resolution matrix and parameter uncertainty for the three-layer model at site 134. The method of determining these results was discussed in Chapter 4. The results show that the thickness and resistivity of the first and second layers can be resolved very well.

**Table 6.4** (a) Resolution matrix for a three-layer model at site 134, (b) estimates of parameter uncertainty 95% confident limits.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Resolution</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>1st layer resistivity ($\rho_1$)</td>
<td>1.0</td>
<td>0.0</td>
<td>0.0</td>
<td></td>
</tr>
<tr>
<td>2nd layer resistivity ($\rho_2$)</td>
<td>0.0</td>
<td>1.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>3rd layer resistivity ($\rho_3$)</td>
<td>0.0</td>
<td>0.0</td>
<td>0.957</td>
<td>1.0</td>
</tr>
<tr>
<td>1st layer thickness ($t_1$)</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>2nd layer thickness ($t_2$)</td>
<td>0.0</td>
<td>0.0</td>
<td>-0.002</td>
<td>0.0</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Minimum Value</th>
<th>Best-fit Value</th>
<th>Maximum Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho_1$ (Ωm)</td>
<td>58</td>
<td>59.7</td>
<td>61.3</td>
</tr>
<tr>
<td>$\rho_2$ (Ωm)</td>
<td>10.7</td>
<td>10.9</td>
<td>11.1</td>
</tr>
<tr>
<td>$\rho_3$ (Ωm)</td>
<td>1315</td>
<td>2121</td>
<td>3422</td>
</tr>
<tr>
<td>$t_1$ (m)</td>
<td>105.3</td>
<td>108.4</td>
<td>111.6</td>
</tr>
<tr>
<td>$t_2$ (m)</td>
<td>926.3</td>
<td>953.6</td>
<td>981.7</td>
</tr>
</tbody>
</table>

(b)
It is necessary to examine whether the difference in the depth to the resistive unit (>1000 Ωm) in the MT models and resistivity logs can be explained by static shift. As shown in the polar diagrams (Fig. 6.4), the impedance $Z_{st}$ is close to zero, therefore, the static shift factor of the determinant apparent resistivity ($D_{det}$) can be calculated by Eq. (3.56).

For site 134, the static shift factor of the determinant apparent resistivity for the Jones's method is 0.957. Figure (6.8a) shows the MT model after the static shift correction. The resistivity of each layer decreases and the depth to the top of the resistive bottom layer decreases, resulting in an interface that is closer to that in the well log data.

The difference between the model and the log responses can also be examined in terms of the integrated conductance of near-surface layer. The MT response provides an accurate estimate of this quantity even if it cannot resolve the exact structure. The integrated conductance above 940 m (maximum depth of well log) from the induction log is 80.5 S. For the same depth (940 m), the integrated conductance of the Phanerozoic sedimentary rocks in the MT model is 78.1 S. Considering the static shift in the MT model, the corrected conductance is 81.6, which is closer to the conductance of the induction log (80.5 S). The conductance difference of 1.1 S may rise from the thickness variation of the sedimentary rocks because of 7.6 km distance between MT site and well log, for example, an 11 m difference in the thickness of the conductive second layer (10 Ωm) can cause 1.1 S difference of conductance.

**Induction Log for Well C-3 and MT Model for Site 124**

The geological log for well C-3 shows the thickness of the Fort Simpson Formation
and underlying 1st Black shale unit is 480 m (Table 6.3). The induction log provides more details and suggests the existence of sub units in the Fort Simpson formation. The rocks are relatively resistive (~35 $\Omega$m) above 120 m, and are very conductive (5-12 $\Omega$m) between 120 m and 480 m depth. From 480 m to 630 m depth, the resistivity is 5-30 $\Omega$m corresponding to the sandy limestone in the Horn River Formation. From 630 m to 683 m depth, the high resistivity (>500 $\Omega$m) is related to the reef carbonates of the Lonely Bay Formation. From 683 to 740 m, the resistivity is about 10 $\Omega$m and corresponds to dolostone and silty sandstone in the geological log. Beneath 740 m, very high resistivity is associated with Precambrian rocks. The overall variation of resistivity defined by the induction log in well C-3 is similar to that in well E-7 (Fig. 6.8a, b).

The MT responses for site 124 provide a four-layered model (Fig. 6.8b): a relatively resistive surface layer (55 $\Omega$m), a conductive second layer (17 $\Omega$m), a more conductive third layer (8 $\Omega$m), and a resistive bottom layer (>400 $\Omega$m). The relatively resistive surface layer which has a thickness of ~100 m is related to the sandy limestone. The conductive second layer at about 100-200 m depth with 18 $\Omega$m resistivity corresponds to the upper Fort Simpson formation. The more conductive third layer with 9 $\Omega$m resistivity below 220 m depth is related to the lower Fort Simpson formation and underlying 1st Black Shale. However, the MT method does not resolve the resistive and thin (50 m thickness) Lonely Bay reef. The depth to the base of the conductive Phanerozoic sedimentary rocks in MT model is about 750 m before static shift correction and 680 m after static shift correction ($D_{det}^2=0.83$).

Table 6.5 provides the resolution matrix and parameter uncertainty for this four-layer
MT model.

Table 6.5 (a) Resolution matrix for a four-layer model at site 124, (b) estimates of parameter uncertainty.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>1\textsuperscript{st} layer resistivity ($\rho_1$)</td>
<td>0.999</td>
</tr>
<tr>
<td>2\textsuperscript{nd} layer resistivity ($\rho_2$)</td>
<td>0.0  0.896</td>
</tr>
<tr>
<td>3\textsuperscript{rd} layer resistivity ($\rho_3$)</td>
<td>0.0  0.0  0.992</td>
</tr>
<tr>
<td>4\textsuperscript{th} layer resistivity ($\rho_4$)</td>
<td>0.0  0.0  0.0  0.975</td>
</tr>
<tr>
<td>1\textsuperscript{st} layer thickness ($t_1$)</td>
<td>0.0  0.06  0.01  0.0  0.959</td>
</tr>
<tr>
<td>2\textsuperscript{nd} layer thickness ($t_2$)</td>
<td>0.0  0.07  0.02  0.0  0.0  0.92</td>
</tr>
<tr>
<td>3\textsuperscript{rd} layer thickness ($t_3$)</td>
<td>0.0  0.0  0.0  0.0  0.02  0.04  0.974</td>
</tr>
</tbody>
</table>

\[
\begin{pmatrix}
\rho_1 & \rho_2 & \rho_3 & \rho_4 & t_1 & t_2 & t_3 \\
\end{pmatrix}
\]

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Minimum Value</th>
<th>Best-fit Value</th>
<th>Maximum Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho_1$ (\Omega m)</td>
<td>52.8</td>
<td>55.7</td>
<td>58.7</td>
</tr>
<tr>
<td>$\rho_2$ (\Omega m)</td>
<td>8.4</td>
<td>17.7</td>
<td>37.1</td>
</tr>
<tr>
<td>$\rho_3$ (\Omega m)</td>
<td>7.0</td>
<td>8.6</td>
<td>10.6</td>
</tr>
<tr>
<td>$\rho_4$ (\Omega m)</td>
<td>304.6</td>
<td>437.9</td>
<td>629.5</td>
</tr>
<tr>
<td>$t_1$ (m)</td>
<td>60.8</td>
<td>96.8</td>
<td>153.9</td>
</tr>
<tr>
<td>$t_2$ (m)</td>
<td>64.0</td>
<td>122.9</td>
<td>236.1</td>
</tr>
<tr>
<td>$t_3$ (m)</td>
<td>363.9</td>
<td>528.3</td>
<td>767.0</td>
</tr>
</tbody>
</table>

(b)

The results show the thickness of each layer is accurately resolved, however, the resistivity of second layer is relatively poorly resolved ($R_{\rho_2\rho_2}=0.896$). The resolution matrix element corresponding to the thickness of each layer has relatively high value (0.959 for $t_1$, 0.92 for $t_2$ and 0.974 for $t_3$). The 95% confidence limits, which provide a relatively
conservative error estimate, are quite large, *e.g.* the limits of the third layer thickness are from 360 m to 770 m compared with the best-fit value 528 m (Table 6.5b). The result suggest that the difference between the depth to the resistive basement in the well log and in the MT model is not statistically resolved.

Above 740 m (maximum well log depth), the integrated conductance from the well C-3 is 85.4 S, which is higher than the integrated conductance of Phanerozoic sedimentary rocks in the MT model, 70.2 S. Considering the static shift, the conductance of MT model will increase to 77.1 S, which is still lower than the integrated conductance of well C-3. Comparison of the observed and synthetic MT responses also suggested a more conductive surface layer at the well site. The conductance difference of 8 S is very large and would correspond to 400 m thickness of the surface layer (50 Ωm) or ~80 m thickness of the conductive third layer (8.6 Ωm). Therefore, the variation of the sedimentary rocks between MT site 124 and well C-3 can not explain this large difference. One reason for the difference could be the limited resolution of MT response.

*Induction Log for Well A-47 and MT Model for Site 120*

The induction log of well A-47 shows the conductive Fort Simpson unit with 12 Ωm resistivity and 110 m thickness (Fig. 6.8c, Table 6.2). From 110 m to 335 m depth, the resistivity is ~8 Ωm in the sandy limestone of the Horn River Formation. Beneath 335 m depth, the reef carbonates of the Lonely Bay Formation (40 m thickness) are thicker than in the other wells. Although the resistivity shows variation within the Lonely Bay Formation, it is still high (Fig. 6.8c). It is hard to distinguish the thin dolostone of Chinchaga Formation
(15 m thickness) on the induction log as it has similar resistivity with the underlying Precambrian rocks. The depth to the top of the Precambrian basement is not clear in this log.

The MT model for site 120 provides a model with two main layers (Fig. 6.8c): a conductive shallow layer (~10 Ωm) and a resistive bottom layer (~880 Ωm). The interface between two layers is at about 330 m depth, which is close to the boundary between sandy limestone of the Horn River Formation and the reef carbonates of the Lonely Bay Formation. It is not possible to delineate the depth to the top of the resistive Precambrian rocks because of the effect of resistive and thick Lonely Bay reef.

Above 330 m depth (thickness of conductive Phanerozoic sedimentary rocks), the integrated conductance of induction log A-47 is 31.2 S, and the integrated conductance of MT model is 32.7 S. Considering the static shift factor, \( D_{st}^2 = 1.05 \) at this site, the conductance of MT model will decrease to 31.9 S. The consistency between induction log and MT result indicates that the MT method provides high resolution for the depth to the base of the conductive layer when the Lonely Bay reef or other resistive formations is the basal unit. The conductance difference of 0.8 S may rise from the thickness variation of the sedimentary rocks over the 4.3 km distance between the MT site and the well. For example, a 40 m variation in the thickness of the overburden (with a typical 50 Ωm resistivity, McNeill 1980), or a 7 m difference in the thickness of the conductive second layer (~9 Ωm) would produce the observed difference in conductance of MT site 120 and well A-47.

### 6.2.5 Summary

The quantitative comparison of the three induction logs and the MT responses at
three sites shows:

- The observed MT responses are similar to the synthetic MT responses derived from induction logs. The small differences are attributed to the thickness variation of the thickness of the surface sediments caused by the distance between MT site and well log (e.g. site 134 and well E-7), and/or the limited resolution of MT response.

- The MT models provide a similar geoelectric structure as the induction logs: a relatively conductive surface layer, a conductive second layer, and a resistive basement. Comparison of the MT models and well logs indicates that the depth to the top of resistive layer is resolved to <100 m by MT sites separated from the wells by ~10 km.

- After correction of static shift, the integrated conductance of the MT model is close to that of the induction log (<10% misfit at site 124, and <2% misfit at sites 134 and 120).

  Therefore, the high frequency MT responses are concluded to be reliable and can be used to investigate the shallow structure.

**6.3 Detailed Comparison of Lac du Bonnet Well Logs and TEM Responses**

In order to examine the TEM results, the observed TEM responses will be compared with synthetic responses from nearby 40 cm (short normal) resistivity logs, and the TEM model will be quantitatively compared with the resistivity log data.

**6.3.1 Comparison of Synthetic TEM Responses from Resistivity Logs and Observed TEM Responses**

The synthetic TEM responses from the resistivity logs will be compared with the
observed TEM responses in this section. The resistivity curve from the resistivity log (Fig. 6.9) was simplified to a ten-layer 1D model and RECTAN was used to calculate the synthetic TEM response of the resistivity log.

Figure 6.10 compares the synthetic TEM responses of four resistivity logs and the observed TEM responses at four stations (one station at each site). Overall, the synthetic responses of well logs are significantly more resistive than the observed TEM data. For the log from well WB-1, the synthetic TEM response is higher than the observed TEM response at station S220 at Site B by 0.6-0.8 decade. For the log of well WD-1, the synthetic TEM response is higher than the observed TEM response at station N220 at Site D by up to one decade. The difference of TEM response between log from well URL-2 and TEM station N200 at Site URL is less than 0.3 decade. The observed TEM response at the central loop station of Site A is erratic and makes the comparison difficult, but the observed values are still lower than the synthetic response based on the log of well WA-1.

There are several factors that may contribute to the higher apparent resistivity in borehole logs compared with the TEM models:

- Resolution of resistivity log. As long as there is no invasion, the resistivity log will correctly measure the resistivity of thick resistive beds (Schlumberger Canada Limited 2000). Figure 6.11 shows the response of the normal resistivity log in thick and thin beds which are less resistive than the surrounding formations. The resistivity log will estimate the true resistivity of a thick conductive layer. However, the resistivity log will overestimate the resistivity of the thin conductive layer, for example, a 40 cm resistivity log will not resolve the true resistivity of 20 cm thick conductive layer (Fig. 6.11). In the
Figure 6.9 Comparison of TEM inversion model (heavy line) and nearby 40 cm resistivity log. (a) TEM station S220 at Site B and log WB-1, (b) TEM station N200 at Site D and log WD-1, (c) TEM station N200 at Site URL and log URL-2, (d) TEM central loop station at Site A and log WA-1.
Figure 6.10. Comparison of observed TEM apparent resistivity (symbol circle) and TEM response of resistivity log (solid line).
Figure 6.11 Curves of normal resistivity log in thick and thin beds that are more conductive than adjacent formations. $R_a$: observed resistivity, $R_s$: resistivity of adjacent formation, $R_t$: resistivity of bed layer, $R_m$: resistivity of mud, $h$: thickness of bed layer, AM: spacing of resistivity log (Schlumberger Canada Limited, 2000).
Lac du Bonnet batholith the resistivity of the relatively thin and conductive fractures will strongly affect the bulk resistivity of the rock, so the rock resistivity will be overestimated by the 40 cm resistivity log.

- Fresh water invasion. The drilling of the well at WRA employed fresh surface water and the well logs were acquired several weeks or longer after the drilling (G. Lodha, pers. comm. 2000). The invasion of fresh water into country rocks may have caused the resistivity measured from the borehole logs to be greater than the true resistivity of rocks. This effect will be most significant around the fracture zones.

- Scale dependence of resistivity measurements. Laboratory experiments on unfractured granite have shown a fractal distribution of microfractures (e.g. Turcotte and Huang 1995). The fractal analysis of rocks near Yucca Mountain, Nevada has also indicated the fracturing of the rocks is scale dependent (Turcotte and Huang 1995). In situations in which the resistivity of fractures contributes to the bulk resistivity of rock, this spatial variation of fractures will affect the observations of resistivity. Therefore, the resistivity measured in EM or DC resistivity surveys may depend on the measurement scale.

Jones (1995) found that the observed resistivity from normal resistivity logs decreases with an increase of electrode spacing. For example, in boring H2 within Oracle granite near Oracle, Arizona the resistivity measured by a 40 cm resistivity log is larger than that measured by 160 cm resistivity log (by a factor of 1.5). Using a time domain induced polarization system (Scintrex IPR-10), the resistivity measured by 1.2 m spacing is larger than that measured by 1.6 m spacing by half a decade. Using cross-hole measurements, the results also show the resistivity measured by 6 m spacing is larger than
that measured by 55 m spacing by a factor of 2.0. The larger electrode spacing are most likely affected by lower resistivity material outside the test zone of the smaller scale measurements.

The TEM measurement made in the present study will measure the resistivity over a scale of 10's to 100's of metres. So, if the same scale dependence occurs in the Lac du Bonnet Batholith as in the Oracle granite, the resistivity measured by a 64 cm normal resistivity log will be significant higher than measured in a TEM study.

6.3.2 Comparison of Resistivity Logs and TEM Models

Figure 6.9 compares the resistivity logs with the inversion models for nearby TEM stations. As determined in the last section using synthetic TEM responses, the layered TEM models correspond to more conductive geoelectric structure than the resistivity log at Sites B and D (Fig. 6.9a, b). Comparison of the TEM models with the resistivity logs also shows an overall higher resistivity in well logs.

At Site B, the resistivity of well WB-1 is larger than TEM sounding by half decade above 100 m depth, by one decade at 100-320 m depth, and by 1.5 decade below 320 m depth. At Site D, the resistivity of well WD-1 is also larger than TEM sounding by 1.5 decade above the thin conductive fracture zone, by 0.2 decade below the thin conductive layer, and by one decade for the conductive layer. However, the borehole log WD-1 does not reach the saline water layer shown in the TEM model. As discussed in Chapter 5, the resolution of the resistivity in the resistive layer is low. The equivalence analysis indicated a relatively large variation range of resistivity, 0.5-1.5 decade, for resistive layer (Fig. 5.23,
25 and Appendix H). It could contribute to the difference between the TEM model and borehole log. However, for the conductive layer the equivalence analysis indicated the resistivity could be better resolved, but its value in TEM model was still lower than that in the borehole log. At Site URL, although the resistivity of well URL-2 is lower than that in the TEM model in the resistive fresh water layer (<430 m depth), it is still larger in the conductive saline water layer (>430 m) by half decade.

Due to complex TEM responses at Site A, the TEM model from central loop station only represents an approximation of the real structure (Fig. 6.9d). A quantitative comparison of the two structures will not be completed in this study.

In general, the TEM resistivity models at sites B, D, URL and A agree with the resistivity expected from the distribution of fracture zones and the tendency of nearby resistivity logs. TEM data can resolve the variation of resistivity in the Lac du Bonnet area, but not the exact resistivity values measured in logs. The sources of the difference could be:

- **Accuracy of resistivity log.** It is possible to calculate the expected rock resistivity based on measured salinity (TDS) values and estimates of porosity in the Lac du Bonnet batholith. For depths >350 m, the TDS is 50 g/L (Everitt et al. 1996) and from Eq. (1.2) the fluid resistivity is 0.12 Ωm. Assuming a porosity 0.3% (Frost et al. 1995) and constant parameters $a=1$, $m=1.5$, the bulk resistivity of rock would be 730 Ωm. Assuming a porosity of 2%, the resistivity of fracture zones would be 42 Ωm. However, the borehole logs (Fig. 6.9) do not show values as conductive as these values, suggesting the logs overestimate the true resistivity.

At Site D, the conductance of fracture zones (0.006 S) in well WD-1 is much lower
than that in TEM model (~0.3 S). This again suggests that the 64 cm resistivity log overestimates the resistivity of conductive layers.

- Scale dependence. The subvertical fractures and subhorizontal fracture zones in the Lac du Bonnet batholith show spatial variability (Brown et al. 1989; Everitt et al. 1996; Hayles et al. 1996), and the multifractal modelling of fractures (Agterberg 1996) also showed a scale dependence of the fractures. It is therefore possible that 40 cm normal resistivity measurements in the wells do not fully characterize the bulk resistivity of the rocks. Based on the scale dependence of resistivity noted by Jones (1995) in the Oracle granite it is expected that the large-scale TEM soundings should be more conductive than the 40 cm resistivity log. The difference between the TEM models and well logs results is greatest in the lower layer where fractures are sparse. This results supports the interpretation that the difference are due to the spatial and scale variations of the fractures.

6.4 Effect of Transverse Anisotropy on MT and TEM Responses

Some small-scale structures cannot be resolved in geophysical EM surveys because of the limited resolution of the EM methods, however, they will affect the EM responses. In this section, the effect of transverse anisotropy will be analysed and a detailed transverse anisotropy analysis of three SNORCLE induction logs and four AECL resistivity logs will be presented.

6.4.1 Transverse Anisotropy

In general, for an arbitrary medium, the electrical resistivity is a tensor quantity
relating the electrical field in each direction to current flow in every direction (Edwards et al. 1988). In geologically oriented studies, the tensor is approximated by either a single scalar quantity or by a vector quantity defining the resistivity in three orthogonal directions. For example, the resistivity of graphite depends on the cleavage direction. The resistivity parallel and perpendicular to cleavage differs by orders of magnitude. In the case of ionic conduction in the fractured rocks saturated with the groundwater, the current paths that are perpendicular to the fractures are more resistive than current paths that run parallel to fractures.

Directionally-dependent resistivity at a scale for which the causative structures cannot be resolved by measurements is called anisotropy (Edwards et al. 1988). The classification of anisotropy is thus related to the resolution of the particular EM sounding method. For example, if the resolution of an EM method is 20 metres and alternating layers of differing resistivity occur at 2 metres intervals, then this layering would be blurred in the EM responses and would be considered as part of the anisotropy of the material. If the individual layers are thicker and can be resolved by EM method, the structure is treated as a 1D stack of layers with different, but isotropic conductivity.

The treatment of electrical anisotropy in rocks has often been restricted to transverse isotropy, in which the resistivity is characterized by two values: one value for all directions within some plane and one value at right angles to the plane (Edwards et al. 1988). Two resistivity parameters can be defined: longitudinal resistivity and transverse resistivity.

Consider a structure consisting of a $N$ horizontal layers of total thickness $H$. Let the thickness and isotropic resistivity of the $i$th layer be $h_i$ and $\rho_i$ respectively. In the horizontal direction, the admittances $h_i/\rho_i$ (thickness divided by resistivity) of horizontal layers add in
parallel to produce $H/\rho$, the total admittance (Fig. 6.12a). The corresponding resistivity $\rho_h$ may also be referred to as longitudinal resistivity (McNeill 1980). Longitudinal resistivity would be sensed by EM responses involving to current flow parallel to the horizontal layers. *i.e.* the TE responses of MT and the TEM measurements. However, in the vertical direction, the impedances $h_i\rho_i$ (thickness times resistivity) of all horizontal layers add to produce the total impedance $H\rho$, (Fig. 6.12b). The corresponding resistivity $\rho$, may be referred to as transverse resistivity. Consequently, the geophysically-measured longitudinal and transverse resistivities for the block of material are not the same and are described mathematically using,

$$\frac{H}{\rho_h} = \sum_{i=1}^{N} \frac{h_i}{\rho}$$  \hspace{1cm} (6.1)

and

$$H\rho = \sum_{i=1}^{N} h_i \rho$$  \hspace{1cm} (6.2)

A coefficient of anisotropy, $f$, based on the ratio of the transverse resistivity to the longitudinal resistivity is used to parameterize the anisotropy,

$$f = \sqrt{\frac{\rho}{\rho_h}}$$  \hspace{1cm} (6.3)

A second parameter, the geometric mean resistivity $\rho_{GM}$, is defined as,

$$\rho_{GM} = \sqrt{\rho \rho_h}$$  \hspace{1cm} (6.4)
Figure 6.12 Physical equivalent of electrical anisotropy in rocks. (a) Equivalent electrical model of geological structure, (b) admittance ($h_i / \rho_i$), (c) impedance ($h_i \rho_i$). The definition of admittance and impedance is shown in text. $H$: total thickness of $N$ layers, $\rho$ and $h$: resistivity and thickness of $i$th layer, $\rho$: longitudinal resistivity, $\rho_t$: transverse resistivity, $R$: $i$th resistance, $R$: the equivalent resistance.
Transverse Anisotropy of Typical Rocks

For most rocks, the values of $f$ are less than about 2.0 (McNeill 1980). In sediments, anisotropy is caused by the gravity-induced alignment of elongated and flattened grains parallel to the bedding planes. Edwards et al. (1988) analysed electrical geophysical logs from Arctic sediments containing permafrost and found coefficients of anisotropy ranging from 1.1 to 3.0. The average range of anisotropy coefficients in sediments is high, ranging from 1.01 to 7.5 (McNeill 1980, Edwards et al. 1988). In crystalline rocks, anisotropy arises from the preferred orientation of mineral crystals, cracks, fractures, and cleavages as a result of pressure and dynamic processes. Al Hagrey (1994) measured anisotropy coefficients in granite of 1.15-1.34, ranging up to 1.75 within a fracture zone.

A number of studies have also examined the theory of how anisotropy affects the measured EM responses. For example, Edward et al. (1984) examined the effect of anisotropy on marine EM measurements in offshore sedimentary basins. Al Hagrey (1994) studied the fracture anisotropy using DC resistivity data from Falkenberg, Germany.

6.4.2 Modelling of the Effect of Transverse Anisotropy

Figure 6.13 shows two layered resistivity models and their transverse resistivity, longitudinal resistivity and geometric mean resistivity. The model consists of multiple layers in which each layer has same thickness (100 m). For Model A, the resistivity value of successive layers beneath the surface alternates between 100 $\Omega$m and 1000 $\Omega$m with the surface layer being 100 $\Omega$m. For Model B, the resistivity value starts from 1000 $\Omega$m in the surface layer. The different resistivity parameters (transverse, longitudinal and geometric
Figure 6.13 Two layered resistivity models and their transverse resistivity, longitudinal resistivity and geometric mean resistivity.
mean) are the same for both models. The longitudinal resistivity is lower than transverse resistivity, indicating longitudinal resistivity is more sensitive to conductive layers than transverse resistivity.

The MT apparent resistivity and phase for Model A and B are shown in Fig. 6.14. Although MT apparent resistivity is affected by the surface layer at short periods (<0.001 s), it is close to longitudinal resistivity at periods 0.001-0.1 s. This result is expected as the MT response in a 1D structure is a TE response. The result confirms that the MT responses are more sensitive to conductive layers than resistive layers. The TEM and DC resistivity responses for Model A and B are show in Fig. 6.15. The late-time TEM responses (>0.001 s) are close to longitudinal resistivity, as for the MT response. DC responses are sensitive to the resistivity of surface layer for the small electrode spacing and close to the geometric mean resistivity for large electrode spacing. The comparison of these different EM methods shows that MT and TEM responses will be biased towards the conductive responses in a finely-layered structure. A DC-resistivity sounding would indicate a more resistive deep structure than the MT and TEM soundings.

6.4.3 Transverse Anisotropy Studies of SNORCLE and Lac du Bonnet Well Log Data

A method assuming a continuous variation of resistivity with depth was used to analyse transverse anisotropy (Edwards et al. 1988). At depth \(d\), the resistivity values within \(d \pm 10\%d\) depth are used to calculate the longitudinal and transverse resistivities (Eq. 6.1 and 6.2). The increment of the depth \(d\) is 0.2 m. This method smooths the anisotropy response of the downhole log.
Figure 6.14 Comparison of MT responses and transverse, longitudinal and geometric mean resistivities for two layered models in Figure 6.12.
Figure 6.15 Comparison of TEM and DC resistivity responses with transverse, longitudinal and geometric mean resistivities for two layered models in Figure 6.12. (a) Central loop TEM responses, the transmitter loop is 1000 m. (b) DC resistivity responses.
The results of this study are based on the assumption that the log data provide a true estimate of the internal resistivity values $\rho_i$. Because the geophysical logs used are 40 cm induction logs and 40 cm resistivity logs which provide an integrated measure of the resistivity over a $\sim$0.5 m scale, the layers that are thinner than this scale will not contribute to the anisotropy estimates. For small scale structures the induction log used in the SNORCLE boreholes will provide a measure of the longitudinal resistivity and the resistivity log used in the Lac du Bonnet boreholes will provide a measure of the transverse resistivity.

Because the fracture zones at URL contain thinly conductive layers that are less than 0.5 m thickness (small fractures), the transverse anisotropy determined in this study may underestimate the true value.

**SNORCLE Study**

The results of anisotropy analysis for the well logs Jean Marie E-7, Mills Lake C-3 and Province A-47 are shown in Fig. 6.16. For all three wells, the transverse resistivity is higher and more sensitive to the resistive layers than the longitudinal resistivity. The longitudinal resistivity is relatively sensitive to thin conductive layers but not thin resistive layers. For example, the thin resistive layers at depth range of 815-865 m in well E-7 (Fig. 6.16a) and depth range of 630-680 m in well C-3 (Fig. 6.16b) are only weakly represented in the longitudinal resistivity. The transverse resistivity shows stronger variation than the longitudinal resistivity. The contribution of the thin resistive layers will be poorly resolved by the MT responses (Fig. 6.8). One reason that the depth to the top of the resistive bottom layer in the MT model for site 134 is deeper than transition to resistive unit in well E-7 (Fig.
Figure 6.16 Comparison of resistivities (induction log, transverse, longitudinal, MT inversion model) and anisotropy coefficient. (a) well E-7 and MT site 134, (b) well C-3 and MT site 124, (c) well A-47 and MT site 120.
Resistivity (ohm.m)

深度 (m)

Anisotropy

(b) Well C-3 and MT site 124
Resistivity (ohm.m)

Anisotropy

(c) Well A-47 and MT site 120
is because the MT response is biased towards the resistivity of the conductive layers interbedded with the resistive layers near the base of the Phanerozoic sedimentary rocks.

For well E-7, the anisotropy coefficient shows great variation with depth (Fig. 6.16a). Above 215 m, the anisotropy coefficient is about 1.02-1.17, which is in the range of values reported for interbedded shale and sandstone (McNeill 1980). Between 215-780 m, the coefficient is very small (1-1.05) which is the typical value of shale (McNeill 1980). Below 780 m, the coefficient is higher (2.0-2.3) which is in the value range of the interbedded limestone and limey shale (McNeill 1980). The lower value (of about 1.75) at 850-900 m is associated with thin and relatively conductive layers located with the shale of Ebbutt Member.

For well C-3, the anisotropy coefficient is small (1-1.25) above 600 m in the silty sandstone, shale, and sandy limestone of Fort Simpson Formation, 1st Black Shale and Horn River Formation, and higher (2.0-2.5) below 600 m in the reef carbonates of Lonely Bay Formation and dolomite of Chinchaga Formation (Fig. 6.16b). For well A-47, the anisotropy coefficient is small above 300 m. Below 300 m the anisotropy coefficient is high and shows significant variation (1.2-2.2) with depth (Fig. 6.16c).

Lac du Bonnet Study

A detailed anisotropy analysis for four borehole logs in the Lac du Bonnet area, (WB-1, WD-1, URL-2 and WA-1) was done and the results were compared with the TEM model from nearby TEM station in Figures 6.17 and 6.18. For all four logs, the longitudinal resistivity is lower and more sensitive to the conductive fracture zones than the transverse
Figure 6.17 Comparison of resistivities (40 cm resistivity log, transverse, longitudinal, TEM inversion model) and anisotropy coefficient. (a) Log WB-1 and TEM station S220 of Site B, (b) log WD-1 and TEM station N200 of Site D, (c) log URL-2 and TEM station N200 of Site URL.
Resistivity (ohm.m)

Anisotropy

(b) Log WD-1
Resistivity (ohm.m)

Depth (m.)

Anisotropy

(c) Log URL-2
Figure 6.18 (a) Comparison of resistivities (40 cm resistivity log, transverse, longitudinal, TEM inversion model) at log WA-1 and TEM central loop station of Site A. (b) Anisotropy coefficient from resistivity log WA-1.
resistivity, for example, at depths of 90-130 m in the log from well WD-1 (Fig. 6.17b) and depths of 40-170 m in the log from well URL-2 (Fig. 6.18c).

The anisotropy coefficients for the borehole logs show higher values than the background at some depths, for example, at borehole WB-1, the anisotropy coefficient values are 1.05-1.3 above 25 m, 1.02 around 55 m, and 1.07 between 125 m and 190 m (Fig. 6.17a). These depths correspond to the fracture zones (Fig. 5.31). Although the TEM models do not resolve all fracture zones shown in the anisotropy analysis, they resolve a resistivity boundary at 105 m depth, which possibly corresponds to the fracture zone at 125 m depth. This result suggests the bulk resistivity (and thus the salinity of groundwater and the distribution of subvertical fractures) is related to the fracture zones at this depth.

For borehole WD-1, the relatively high anisotropy coefficients are observed at 10 m depth with 1.15 value, 60 m depth with 1.17 value, and 90-120 m depth with 1.4-1.65 value (Fig. 6.17b). These depths also correspond to the fracture zones (Fig. 5.33). The fracture zone at 90-120 m depth may correspond to the conductive zone at 150-170 m depth in TEM model. As discussed in Chapter 5, the different depth between the borehole and the TEM model may be due to the 500 m distance between the location of the TEM soundings and the boreholes (Fig. 5.32).

For borehole URL-2, the fracture zones at 40-170 m depth (Fig. 5.35) correspond to high anisotropy coefficients (1.3-1.7 value, Fig. 6.17c), and also to the conductive layer at ~100 m depth in the TEM models at several stations. The resistivity interface at 415 m depth in the TEM model corresponds to a slight change of anisotropy coefficient.

For borehole WA-1, the anisotropy coefficients are high at 280-350 m depth (Fig.
6.18). The TEM model also includes a conductive layer at 210-270 m depth. However, a quantitative comparison of the features can not be made for this site because the TEM model represents only an approximation of the true structure (as explained in Chapter 5).

The anisotropy analysis also indicates the DC-resistivity sounding would provide a relatively resistive structure than the TEM sounding. This aspect should be considered in any future quantitative comparison of DC-resistivity sounding results (e.g. Paterson and Watson 1981) and TEM results in the study area.

6.5 Comparison of Results from SNORCLE and Lac du Bonnet Studies

Structures

Comparison of EM results and well logs show that three-layer structures derived from the MT results in SNORCLE area are in good quantitative agreement with the well log data. For example the synthetic MT response of induction log is close to the observed MT response, and the integrated conductance of induction log is close to that of the MT model. The three-layer structure from the TEM results in Lac du Bonnet area have the same trends as the well log data, but the resistivity of TEM sounding is lower than that of well log. The differences are largest in the zone of sparse fracturing at depth. The reason for this difference could be the accuracy with which the resistivity log measures the resistivity of thin conductive layers and the scale dependence of resistivity in the granite of the Lac du Bonnet Batholith.
Resolution

The inversion of EM responses shows that the MT method delineates the depth to the base of the conductive Phanerozoic rocks, and the TEM method delineates the depth to the base of the resistive fresh water saturated granite. Both methods are sensitive to the conductive structures. However, the MT method can not resolve the thin resistive layers, e.g. 50 m thick Lonely Bay reef in SNORCLE area, because of the limited resolution of MT sounding. The TEM method can resolve only the conductance of the fracture zone, not the exact resistivity value, because of the problem of equivalence.

Transverse Anisotropy

The anisotropy analysis shows the MT and TEM responses which provide a measurement of the longitudinal resistivity are more sensitive to the conductive layers than resistive layers in the SNORCLE and Lac du Bonnet areas. Anisotropy analysis revealed relatively high values of the coefficient of transverse anisotropy in the thin sedimentary layers in SNORCLE area (e.g. >2.0 values in resistive reef carbonates of Lonely Bay Formation), and in the fracture zones in the Lac du Bonnet area (e.g. >1.4 values at Site D).

Methodology

Based on the results shown in last several sections, high frequency MT data could be used to map the fresh/saline water interface in Lac du Bonnet area. A TEM survey with a larger transmitter loop size (e.g., 200 m or 300 m) could be used to investigate resistivity variations in sedimentary rocks of the SNORCLE area.
Chapter 7  Conclusions

This thesis study used two EM methods to investigate the structures in crystalline rocks of the Precambrian Shield: MT and TEM. The MT method provided information on the Proterozoic tectonics from the crust to mantle depths in the southwestern Northwest Territories and on the overlying Phanerozoic sedimentary rocks. The TEM method delineated the interface of fresh/saline water saturated granitic rocks at depths above 600 m in the Lac du Bonnet batholith, Manitoba, using a PROTEM47 instrument with a 100 m transmitter loop.

MT Results from the SNORCLE Study

The SNORCLE magnetotelluric project is part of LITHOPROBE, Canada’s national, collaborative, multidisciplinary Earth Science research project. Magnetotelluric soundings were made at 60 sites in the southwestern Northwest Territories, Canada, along the LITHOPROBE SNORCLE Transect Corridor 1 and 1A, in the summer of 1996. The sites are located in southwestern Northwest Territories, Canada, between latitudes 60°-65°N and longitudes 110°-125°W. The main objective of SNORCLE study in this thesis is to delineate regional conductivity structures and use this information to investigate the structures of the Proterozoic terranes, the structures within the Phanerozoic sedimentary rocks, the structures of the Great Slave Lake shear zone, and the variation in lithospheric thickness.
The MT data collected (electric and magnetic field time series) were converted to impedances by spectral analysis, then used to calculate apparent resistivity and phase. The Groom Bailey regional strike angle, direction of maximum phase difference and induction vectors were used to examine the geoelectric strike. In order to reduce the distortion from near-surface inhomogeneities Groom Bailey decomposition and static shift removal were applied to the data. After that, one-dimensional forward modelling and inversion was used to determine the structure of shallow Phanerozoic sedimentary rocks, and 2D forward modelling and 2D inversion (Occam and Non-Linear Conjugation Gradient methods) were used to examine the crustal and lithospheric structure. Results of the seismic reflection, seismic refraction, seismology, gravity and magnetic field were used to interpret the MT results.

The MT study along SNORCLE transect Corridor 1 and 1A reveals the geoelectric structure of the Phanerozoic sedimentary rocks and the underlying Proterozoic crust and lithosphere. The MT results confirm the following structures and tectonics which already have been revealed by the other geophysical methods:

- The conductive Phanerozoic sedimentary rocks have an approximately 1D resistivity structure. They have a maximum thickness along the corridor of about 1000 m at a location in the western Fort Simpson terrane (longitude 123°W) and decrease in thickness towards their eastern margin in the eastern Great Bear Magmatic Arc (longitude 117°W). The resistivity structure consists of two layers: a relatively resistive surface layer (50-100 Ωm, 20-100 m) which is interpreted using well log information to be upper Devonian sandy limestone; and a conductive lower layer (10 Ωm, 0-850 m) which is interpreted to
correspond the upper Devonian shale and silty sandstone. The MT results identified the Liard High, a structure formed during the early Paleozoic. Well logs show the thickness of the Phanerozoic sedimentary rocks is thinner over the Liard high than surrounding area. The MT results also show a relatively thin conductive layer at this location.

The sedimentary rocks overlying the Buffalo Head terrane consist of relatively resistive middle Devonian limestone and salt deposits (~500 $\Omega$m, ~350 m) overlying the relatively conductive sandy-silty limestone and shale layer (~50 $\Omega$m, ~300 m). The discontinuity of resistivity near the boundary between the Buffalo Head terrane and the Great Bear Magmatic Arc noted in the MT results is consistent with the presence of a unconformity in the Phanerozoic sedimentary rocks (Fig. 4.23).

Detailed comparison of MT results and borehole logs shows the observed MT responses are similar to synthetic MT responses derived from induction logs. The small differences are attributed to the thickness variation of the thickness of the geological units caused by the distance between MT site and well log (e.g. site 134 and well E-7), and/or the limited resolution of MT response (e.g. site 124 and well C-3). The MT models provide a similar geoelectric structure as the induction logs: a relatively conductive surface layer, a conductive second layer, and a resistive basement. Comparison of the MT models and well logs indicates that the difference between the depth to the top of resistive layer in MT models and well logs (separated by ~10 km) is <100 m. After correction of static shift, the integrated conductance of the MT model is close to that of the induction log (1-10% misfit). Therefore, the consistence between the MT results and well logs proves that the high frequency MT responses can be used to investigate the structure of the
sedimentary rocks where the borehole logs are scarce.

- The MT results imaged the resistive (>400 Ωm) Fort Simpson basin. The resistivity structure shows the base of the basin is west-dipping (about 20°), and reaches a depth of ~20 km depth in the west, and that the basin extends at least 100 km along the survey line. The MT results support the interpretation of the geometry of the basin made from seismic reflection data. The basin is believed to have formed as a result of lithospheric extension following collision of the Fort Simpson terrane with the western Hottah terrane at ~1.84 Ga.

  The MT results show the western margin of the basin is the Bulmer Lake Arc. The basin overlies the relatively conductive (<250 Ωm) Proterozoic mantle.

- The MT results reveal that a sub-vertical and conductive (<100 Ωm) boundary between the Fort Simpson and Hottah terranes extends from several kilometres depth to the lower crust. This conductive zone corresponds to a wedge-shaped structure occurring at depth between several kilometres and ~25 km in the seismic reflection response and a relatively high velocity zone in the seismic refraction response. The precise geometry of the conductive zone is not defined by the MT results because of the limited resolution of the MT method.

  This conductive body is interpreted as a structural boundary associated with the collision between the Fort Simpson and Hottah terranes. The older Hottah terrane (1.95-1.91 Ga) collided into the eastern flank of the younger Fort Simpson terrane (1.8 Ga) creating a wedged shape geometry. The upper crust of the Fort Simpson terrane was detached and thrust over the Hottah terrane, and the ocean lithosphere that separated the
Fort Simpson terrane from the Hottah terrane was subducted beneath the crust of the Hottah terrane. Later transpression caused the metamorphism and deformation of the Proterozoic sedimentary rocks of the Fort Simpson and Hottah terranes. The conductor is interpreted to be associated with deformed oceanic lithosphere. Either carbon grain-boundary films or conducting minerals (graphite or sulphide or oxides minerals) could be a viable mechanisms for the enhanced conductivity.

The MT results reveals the following new structures:

- The MT results reveal a conductive (~30 Ωm) mid-lower crust beneath the boundary of the Hottah terrane and the Great Bear Magmatic Arc for which the depth to the top of the conductive zone is ~20 km. The Great Bear Magmatic Arc has been interpreted as the product of eastward subduction of oceanic lithosphere beneath the Hottah terrane at 1.84-1.87 Ga. On the basis of the seismic results it is interpreted to be a relatively thin layer (3-4.5 km), overlying either Hottah crust or imbricated rocks of the Coronation margin. The origin of the rocks beneath the Great Bear Magmatic Arc could be deformed and metamorphosed rocks of the Hottah-Slave transition, for example, sedimentary rocks of the Coronation Supergroup. The sources of the enhanced conductivity could be either carbon grain-boundary film or conductive minerals (graphite or metallic oxides or sulphide) formed during the deformation and metamorphism of the rocks in the Hottah-Slave transition.

- The MT results show that the upper crust of the Buffalo Head terrane is characterized as an east-dipping resistive structure (> 1000 Ωm), corresponding to metaplutonic and subordinate felsic metavolcanic rocks. The relatively conductive middle and lower crust
of the eastern Buffalo Head terrane (<500 Ωm) may be interpreted as the westwards extension of the Taltson magmatic arc beneath the Buffalo Head terrane.

- Within the crust the GSLsz forms a resistive zone at least 20 km wide and at the surface is correlated with a magnetic low. The rocks within the GSLsz consist of greenschist to granulite facies mylonites. The high resistivity (>5000 Ωm) is interpreted to be due to the resistive nature of the granitic protolith of the mylonites and the fact that the GSLsz are dominated volumetrically by rocks deformed within the ductile regime.

The geoelectric strike direction near the GSLsz varies with depth from ~N33°E at periods less than 20 s corresponding to the upper and middle crust to ~N62°E at period range of 20-1500 s corresponding to the lower crust and lithospheric mantle. These results and two-dimensional modelling show that the GSLsz is a lithospheric-scale feature.

The geoelectric strike at crustal depths in the vicinity of the GSLsz is more north-south than the overall azimuth of the shear zone to the south of the Slave Province (N60°E) but is consistent with the strike of the magnetic anomaly interpreted to correspond to the shear zone and with the fast direction of SKS seismic phases within the upper layer of a two-layer model. This strike is interpreted to reflect a local scale (<50 km) deflection of the mylonite belts, an interpretation supported by the geometry of the magnetic anomalies within the study area. The geoelectric strike at periods corresponding to lower crust and mantle lithosphere depths is N62°E. This orientation is close to the large-scale (>50 km) azimuth of the shear zone south of the Slave Province. In the present study the long period MT responses were fitted quite well by a resistivity model consisting of two-dimensional isotropic structures but is possible the data could also be fitted by a
model incorporating both anisotropic conductors and structural elements. The deeper MT strike direction is oblique to the fast SKS direction within the lower layer of a two-layer model. The relative orientation of the two sets of data is the same as in the area of the Grenville Front in eastern Canada and is consistent with dextral shear recorded in the mantle lithosphere.

- The MT method reveals a conductive structure in the upper mantle beneath the Hottah terrane (<150 Ωm). The depth to the top of this conductor is about 70 km in the west, decreases towards the east, and the conductor may connects with the conductive mid-lower crust at 20 km depth in the east. The conductive zone correlates to a zone of the weak and discontinuous seismic reflections below 30 km and a low velocity zone (7.7 km/s). The mechanisms for the enhanced conductivity could be carbon, dissolved hydrogen, or an uplift lithosphere-asthenosphere boundary.

- The MT results reveal an enhanced conductivity in the mantle along Corridor 1 and 1A. The depth to the 150 Ωm resistivity conductor occurs at ~100±20 km depth beneath the Nahanni, Fort Simpson, central Hottah, central Great Bear and Buffalo Head terranes. The conductive mantle is the shallowest in the Hottah terrane (~40 km depth). High heat flow (109 mW/m²) across the Hottah, Fort Simpson and Nahanni terranes measured by Lewis and Hyndman (2001) indicates that an uplift of the lithosphere-asthenosphere boundary could contribute to the enhanced conductivity. The source of the high conductivity would be partial melt in the asthenosphere. The enhanced conductivity in the Hottah mantle is more localized and therefore could be from either dissolved hydrogen, or carbon, or both.
TEM Results from the Lac du Bonnet Study

The WRA located near Pinawa, Manitoba, Canada has been used by the AECL to investigate the concept of disposal of nuclear fuel waste in Precambrian rocks. The Lac du Bonnet TEM project is one component to evaluate the use of geophysical methods to investigate the properties of rock and groundwater. The main objectives of the TEM project are to evaluate the ability of the TEM method to identify the fresh/saline-water interface and delineate the fresh/saline-water interface at specific sites in the Lac du Bonnet area.

Two feasibility studies were completed in the summer and fall of 1996, using 10 m, 20 m and 40 m transmitter loops. A full survey was completed at four selected locations around Lac du Bonnet (Site B, D, URL, and A) in the summer of 1997, using a 100 m transmitter loop. The data were collected with 0-280 m offset range and 20 m spacing on four sides of transmitter loop.

The TEM surveys successfully mapped the fresh/saline water interface in the Lac du Bonnet area. The results imaged a basic three-layer structure at Sites B, D and URL: a thin conductive surface layer, a thick resistive layer with an embedded conductive layer at some stations, and a conductive lower layer.

The conductive surface layer (~20 Ωm) is resolved at all stations of Site B and some stations at Site D with 0.5-2.0 m thickness, and all stations of Site URL with <10 m thickness. It is interpreted to be associated with the spatially-variable soils, organic material in open fractures, and weathered rock.

The resistivity of the second layer is approximately 3000 Ωm at Sites B and D, and greater than 50,000 Ωm at Site URL. This layer is interpreted to represent the zone of sub-
vertical fractures saturated with fresh groundwater percolating from the surface. The depth to the base of the resistive layer is deeper at Site URL (~460 m) and decreases to the northwest (180 m at Site B). It shows small scale spatial variation along the profiles at each site. The apparent porosity of this layer is 2.93-5.62% at sites B and D, which is higher than 0.54% at site URL.

There is an embedded conductive layer at most stations at 100-200 m depth at Site D (~0.3 S conductance), which is interpreted to correspond a low-dipping fracture zone. At URL the embedded conductive layer is present at 120-170 m depth at a few stations (~0.05 S conductance) and is interpreted to be the part of the sub-horizontal fracture zone FZ3.

The resistivity of the basal conductive layer is 50-400 $\Omega$m at Sites B and D, and 1000-2000 $\Omega$m at Site URL. This conductive layer is interpreted to correspond to sparsely fractured rocks (containing a few sub-horizontal fractures) saturated with saline groundwater or brine. The apparent porosity is 0.45-1.8% at sites B and D, which is higher than 0.15% at site URL.

At Site A, although it was impossible to obtain satisfactory layered inversion models because of the multiple sign reversals in the observed data, the inversion results suggest a three-layer "background" resistivity structure that is similar to the results at Sites B, D and URL. The responses at Site A indicate the presence of significant anisotropic, 2D or 3D components in the structure.

Comparisons of TEM results and borehole logs show TEM method provides some resolution of the fracture and groundwater salinity distributions in the Lac du Bonnet Batholith. The method resolved the depth to the fresh/saline water interface at Site B and
resolved the fracture zones at Sites D and URL. These results show the potential of the TEM method for site investigation in granitic batholith.

The TEM method did not provide complete resolution of the groundwater salinity distribution at Sites D and URL. The depth to the base of a resistive layer at Sites B and URL is greater than the depth to the fresh/saline water interface from borehole logs. The conductive fracture zones within the overlying resistive layer and the limited effective penetration depth of the TEM system could contribute to this difference. Also quantitative comparison of the TEM models and borehole logs showed the resistivity determined by the TEM method was less than that determined by borehole logs. The reasons for this difference are interpreted to be the accuracy with which the resistivity log measures the resistivity of thin conductive layers and the scale dependence of resistivity measurements in the borehole log.

The TEM study in the Lac du Bonnet Batholith was successful in demonstrating the potential of the TEM methods in mapping groundwater salinity in granitic batholith. The PROTEM 47 instrument, in combination with a 100 m transmitter loop, provides a suitable TEM system for mapping the resistivity structure of the Lac du Bonnet batholith down to a depth of 300-400 m. For deeper penetration and more accurate results in areas of fracture zones, other TEM systems with greater transmitter power and/or larger transmitter loops will be required. Further analysis is required to fully understand the TEM responses at Site A and the local structures at all sites, which may require consideration of the effects of anisotropy and 2D/3D components of the resistivity structure. The work can be started by applying 2D and 3D forward modelling and inversion.
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Appendix A  Apparent Resistivity and Phase Curves for MT Sites in SNORCLE Transect Corridor 1 And 1A

This appendix shows the merged apparent resistivity ($\rho_\alpha$ and $\rho_{yz}$) and phase($\phi_y$ and $\phi_{yx}$) at 60 sites along the SNORCLE transect Corridor 1 and 1A. The circles present the $xy$ mode responses and the black dots present the $yx$ mode responses. The data are rotated to a geographic north coordinate system (section 4.2.3).
Appendix B  Comparison of 2D Occam Inversion Model Responses and Observed Data at Sites along Corridor 1

This appendix compares the responses of two 2D occam inversion models (Fig. 4.26a and 4.26b) with the observed data at sites along Corridor 1. For the model Fig. 4.26a, only TE responses are compared because only the TE apparent resistivity and phase data were inverted. For the model Fig. 4.26b, only the TM apparent resistivity and phase data were inverted, so the comparison of only TM responses are shown.
TE Model Responses

Site 145

Site 144

Site 143

Site 142

Site 141

Site 140

Site 139

Site 138

Site 137

Site 136

Site 135

Site 134

Phase (Degree)

Appendant Resistivity (ohm.m)

Period (s)
Appendix C  Comparison of 2D Occam Inversion Model Responses and Observed Data at Sites along Corridor 1A

This appendix compares the responses of two 2D occam inversion models (Fig. 4.30a and 4.30b) with the observed data at sites along Corridor 1A. For the model Fig. 4.30a, only TE responses are compared because only TE apparent resistivity and phase data were inverted. For the model Fig. 4.30b, only the TM apparent resistivity and phase data were inverted, so the comparison of only TM responses are shown.
TM Model Responses

Site 129

Site 128

Site 127

Site 126

Site 125

Site 124

Site 123

Site 122

Site 146

Site 147

Site 148

Site 149

Apparent Reactivity (abs/m)

Phase (degree)

Period (s)
Appendix D  Comparison of 2D NLCG Inversion Model Responses and Observed Data at Sites along Corridor 1

This appendix compares the responses of two 2D NLCG inversion models (Fig. 4.34a and 4.34c) with the observed data at sites along Corridor 1. The Tau value used in the inversion is 3 for model Fig. 4.34a and 300 for model Fig. 4.34c.
Appendix E  Comparison of 2D NLCG Inversion Model Responses and Observed Data at Sites along Corridor 1A

This appendix compares the responses of two 2D NLCG inversion models (Fig. 4.36a and 4.36c) with the observed data at sites along Corridor 1A. The $\tau_u$ value used in the inversion is 3 for model Fig. 4.36a and 300 for model Fig. 4.36c.
\textbf{Tau = 300}

\begin{itemize}
  \item Site 129
  \item Site 128
  \item Site 127
  \item Site 126
  \item Site 125
  \item Site 124
  \item Site 123
  \item Site 122
  \item Site 146
  \item Site 147
  \item Site 148
  \item Site 149
\end{itemize}

\begin{itemize}
  \item Phase (Degree)
  \item Apparent Resistivity (ohm m)
\end{itemize}

Period (s)
Appendix F  Comparison of 2D NLCG Inversion Model Responses and Observed Data at Sites around GSLsz

This appendix compares the responses of two 2D NLCG inversion models (Fig. 4.52a and 4.52b) with the observed data at sites around GSLsz. The model Fig. 4.52a inverted the short period MT data (10^{-1}-10 s) with a N30°E rotation angle. The model Fig. 4.52b inverted the long period MT data (30-3\times10^3 s) with a N60°E rotation angle.
Long Period Responses

Site 129

Site 128

Site 127

Apparent Resistivity (ohm.m)

Phase (Degree)

10^{-1}

10^{-2}

10^{-3}

Site 126

Site 125

Site 124

Apparent Resistivity (ohm.m)

Phase (Degree)

10^{-1}

10^{-2}

10^{-3}

Site 123

Site 122

Site 121

Apparent Resistivity (ohm.m)

Phase (Degree)

10^{-1}

10^{-2}

10^{-3}

Site 141

Site 140

Site 149

Apparent Resistivity (ohm.m)

Phase (Degree)

10^{-1}

10^{-2}

10^{-3}

10^{-4}

Period (s)
Appendix G  Profiles of TEM Late-time Apparent Resistivity

The following figures show the late time resistivity versus offset for each time-gate at Sites B, D, URL and A.
Figure G.1 Profiles of late time resistivity for individual time-gate at Site B. The data for the 19 time-gates have been divided into two groups: gates 1-5 and gates 6-19. See Table 5.3 for the time of each time-gate.
Figure G.2 Profiles of late time resistivity for individual time-gate at Site D. The data for the 19 time-gates have been divided into two groups: gates 6-14 and gates 14-18. The first five gates and last gate are very erratic and are not shown. See Table 5.3 for the time of each time-gate.
Figure G.3 Profiles of late time resistivity for individual time-gate at Site URL. The data for the 19 time-gates have been divided into two groups: gates 6-14 and gates 14-18. The first five gates and last gate are very erratic and are not shown. See Table 5.3 for the time of each time-gate.
Figure G.4 Profiles of late time resistivity for individual time-gate at Site A. The data for the 19 time-gates have been divided into two groups: gates 6-14 and gates 15-19. The first five gates are very erratic and are not shown. See Table 5.3 for the time of each time-gate.
Appendix H  Layered Inversion Models, Responses and Resolution Matrices for Selected Stations from Sites B, D and URL

The following figures show the layered inversion models, equivalence analysis and responses for some stations from different lines (east, west, south, north) at Site B, D and URL. The tables list the resolution matrix of the contrastive station. In the tables, \( P \) presents the resistivity, \( T \) presents the thickness, the number \( N \) following \( P \) and \( T \) is \( N \)th layer of model. For example, \( PI \) and \( TI \) mean the resistivity and thickness of the first layer individually.
Figure 3.1: Layered inversion model and response for station W100 at Site B. (1) Layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Figure H.2 Layered inversion model and response for station S100 at Site B. (1) layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Figure H.3 Layered inversion model and response for station N121 at Site B. (1) layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Table H.1 Resolution Matrix for stations (a) W100, (b) S100, (c) N121 at Site B.

(a) W100

<table>
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<tr>
<th></th>
<th>P1</th>
<th>P2</th>
<th>P3</th>
<th>P4</th>
<th>T1</th>
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</tr>
<tr>
<td>P3</td>
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<td>-0.02</td>
<td>0.91</td>
<td></td>
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<tr>
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<td>-0.2</td>
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<td></td>
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<td></td>
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</tr>
<tr>
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<td>-0.06</td>
<td>-0.15</td>
<td>0</td>
<td>0.01</td>
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(b) S100

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(c) N121

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Figure H.4 Layered inversion model and response for station W100 at Site D. (1) layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Figure H.5 Layered inversion model and response for station E100 at Site D. (1) layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Figure H.6 Layered inversion model and response for station S100 at Site D. (1) layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Table H.2 Resolution Matrix for stations (a) W100, (b) E100, (c) S100 at Site D.

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Figure H.7 Layered inversion model and response for station W280 at Site URL. (1) layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Figure H.8 Layered inversion model and response for station E240 at Site URL. (1) layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Figure H.9 Layered inversion model and response for station N200 at Site URL. (1) layered inversion model, (2) equivalence analysis (final model - solid line, equivalence model - dashed line), (3) voltage response, and (4) apparent resistivity response.
Table H.3 Resolution Matrix for stations (a) W280, (b) E240, (c) N200 at Site URL.

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