MAGNETOTELLURIC CRUSTAL STUDIES IN KENAI, ALASKA

by

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Date__26 November 2003_

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_____Good times at CSM_____

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Magnetotelluric (MT) data are acquired, processed, inverted and integrated with aeromagnetic and gravity data for geologic interpretation. The transect is over the flank of a long-wavelength aeromagnetic anomaly high on the Kenai Peninsula in Alaska with the Border Ranges fault on the eastern edge and the Cook Inlet sedimentary basin on the western edge. The MT sounding method images subsurface electrical conductivity using time-varying electric and magnetic fields recorded at the Earth’s surface. Data collected in the field allow construction of electrical conductivity distribution in the subsurface that may be representative of geologic structure. The two-dimensional conductivity model constructed from processed apparent resistivity curves, phase curves, and tipper show a number of deep conductors. A few main results on conductivity structure are: geologic framework surrounding the Border Ranges fault is more resistive to the east, there is a deep conductive zone beneath the Cook Inlet sediments with the top at approximately 10 km depth, and the deep conductive zone comes toward the surface at the edge of the basin. The combined gravity, magnetic, and MT model shows a body that is magnetic, dense, and conductive. This suggests a mafic lithology (magnetic, dense) with possible zones of high porosity and/or serpentinization to explain the high conductivity.
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for Jonathan Goold and Green Family
CHAPTER 1

INTRODUCTION

The Kenai Peninsula is located on Alaska's southern coast and it is bordered by Cook Inlet and Gulf of Alaska, as shown in Figure 1.1. This thesis focuses on a magnetotelluric (MT) project to investigate the crustal structure in the Kenai area. The project consists of acquisition, analysis, and interpretation of MT data. The MT study area is indicated by a blue dot in Figure 1.1. Figure 1.2 shows a topographic map of the study area with dots at six MT station locations. The MT transect crosses relatively flat terrain, resulting in two benefits. The field logistics are relatively easy. More importantly, topographic effects are minimal in the acquired MT data. This chapter discusses the geologic problem, regional tectonic settings, previous geophysical work in the Kenai area, and background information on the MT sounding method.

1.1 Geologic Problem

In a study carried out by USGS (Saltus et al., 1999), a Southern Alaska magnetic high (SAMH) is identified and described. The question of origin of the SAMH requires a geological understanding of the area. The SAMH is a major crustal element of Southern Alaska and generally is associated with Jurassic arc-related rocks and their basement within the Wrangellia composite terrain of Plafker and Berg (1994). Understanding of this major crustal element is one key to unraveling the complete tectonic history of
Figure 1.1 General location of Kenai Peninsula and Cook Inlet (modified from Magoon, 1994). The MT study area is indicated in blue.
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southern Alaska (Saltus et al., 2003). Magnetic data alone are not sufficient to answer the question. Additional data, such as MT soundings, that provide complementary information and have better depth resolution are therefore required.

1.2 Major Tectonic Features and Geologic Setting

In south-central Alaska, the Pacific Plate is sliding to the NNW past southeastern Alaska on the Queen Charlotte Fairweather transform fault and then is subducting beneath southern Alaska at a rate of about 5.5 cm/yr on the Aleutian megathrust (Haeussler et al., 2000). The Yakutat Block is a microplate colliding with the southern Alaska margin, as shown in Figure 1.3.

Collision of the Yakutat terrain is inferred to be the cause of rapid and high uplift of the eastern Chugach-St. Elia Mountains, faulting in interior Alaska along the Denali and Totsschunda fault systems, and uplift of the highest part of the Alaska Range. It is also the cause for deformation in and extrusion of the forearc to the southwest (Haeussler et al., 2000). The origin of crust beneath the Cook Inlet basin is one of the questions surrounding the understanding of these plate movements. Cook Inlet evolved from a backarc basin setting during Mesozoic time to a forearc basin in the Cenozoic. Numerous high-angled reverse faults and minor normal faults indicate considerable compression throughout the Mesozoic and Cenozoic periods (Magoon, 1994). MT investigation of deeper crustal structure in the magnetic anomaly high of the Kenai area provides means for more insightful explanations of tectonic history.

A closer look at the tectonic structure around Kenai is shown in Figure 1.4. Transects labeled A-A’ and B-B’ are used in geological conclusions in Chapter 4. Figure 1.5 describes the units and symbols in Figure 1.4. The basin and geophysical study area
Figure 1.3 Regional tectonic overview of southern Alaska. The subduction zone bounding southern Alaska is created by collision of the Pacific Plate and North American Plate at a rate of about 5.4 cm/yr (from Haeussler et al., 2000).

lie between the Kenai Mountains and an arc of volcanoes that form parts of the Alaska and Aleutian Ranges.

Cook Inlet and the Kenai Mountains bound the western part of Kenai Peninsula, which is generally covered by Quaternary sediments. In the northwestern part of the peninsula, unconsolidated sediments are as thick as 230 meters. In the southern part of the peninsula, unconsolidated sediments are thinner and they are absent on many hills (Glass, 1996). Figure 1.4 shows a geologic map of the study area. Peat has developed in poorly sorted areas that have little topographic relief. The Cook Inlet contains rocks that yield oil, gas, and coal (Glass, 1996).
Figure 1.4 Major composite tectonostratigraphic terranes, terranes, and geologic units of the southern Alaska margin and adjacent areas (Plafker et al., 1994). The MT transect is indicated by a blue line. Cross-sections A-A’ and B-B’ are described in geologic conclusions in Chapter 4.
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The Kenai Mountains, located south of the study area, have a general composition of metamorphosed sedimentary and volcanic rocks of Jurassic and Cretaceous ages. Unconsolidated sediments comprise the western geologic setting. They include glacial deposits belonging to various Pleistocene glaciations, and Holocene fluvial, colluvial, and other surficial deposits (Bradley and Wilson, 1998). Intrusive rocks, shown in black, are located on the eastern edge of Figure 1.6. Though the details of surface geology are not the targets of this MT study, structures imaged at depth provide information about how basins with resources are formed. The MT investigation is aimed at inferring geologic structure beneath the Quaternary sediments.

1.3 Existing Work

Gravity and magnetic studies indicate the presence of low density by the Cenozoic sedimentary section of the Cook Inlet, and high density and magnetic susceptibility for a deeper body on the west side of the Border Ranges fault in the Kenai study area (Saltus et al., 2003). Steve Hackett reported a gravity survey in 1977 (Hackett, 1997). Richard Saltus and others reported airborne magnetic surveys in 2001 (Saltus et al., 2001).

In previous gravity work, a LaCoste-Romberg gravimeter from the University of Alaska Geophysical Institute was used for data acquisition. Gravity stations were plotted on USGS 1:62,500 and 1:250,000 scale topographic maps of the Tyonek and Kenai quadrangles. Bouguer anomalies have been computed using a reduction density of 2.67 gm/cm³ (Hackett, 1977). A broad, steep gravity gradient is associated with the East Basin-Border Ranges fault system. This wide and relatively steep gravity-gradient zone extends across the Kenai Peninsula and along the Kenai Mountain front. The zone
borders the eastern flank of the Cook Inlet basin. The cause is interpreted to be extensive basement faulting.

The aeromagnetic map of Alaska shown in Figure 1.7 is constructed from grids that combine information collected in 85 separate aeromagnetic surveys conducted between 1945 and 1982. The data from these surveys are of varying quality; large regions of Alaska are covered only by very coarse surveys with line spacing greater than 9.6 km (Saltus et al., 2001). The southern region is magnetically distinct from the rest of Alaska. It contains strong magnetic highs and lows that are continuous along strike for more than 1,000 km. These domains correlate with convergent margin tectonostratigraphic terranes including magmatic arcs, deformed flysch belts, and accreted oceanic crust (Saltus et al., 2003). The originally analog magnetic data over the Kenai MT transect is from a survey flown in 1957 and 1958. Flight line spacing varied, averaging approximately 3.2 km spacing. Flight lines were flown east-west at a barometric elevation of 762 m (Saltus et al., 2001).

The width and extent of the gravity and magnetic gradients imply considerable regional stratigraphic truncation of the Tertiary sediments along the eastern margin of the Cook Inlet basin (Hackett, 1977).

1.4 Magnetotelluric Method

MT provides resistivity as a new parameter and yields information at depths previously unresolved by potential field geophysics. Interpretation of MT gives resistivities and depths, not just anomalous highs and lows in the data. Therefore, depth interpretation based on MT data is much more definitive than that based on gravity or magnetic data (Vozoff, 1972).
Figure 1.7 Merged aeromagnetic data of Alaska in a shaded relief image (illumination from the northwest). Individual datasets are adjusted by upward and downward continuation to produce a uniform specification of 305 m draped above terrain. This grid provides a consistent basis for digital filtering of these data. The black lines are major faults (Saltus et al., 2001), with the exception of the arrow that points to the aeromagnetic anomaly high that the MT transect crosses. The southern Alaska magnetic high parallels the subduction zone of the Pacific and North American plates.
The first papers to discuss MT theory for one-dimensional (1D) structures were written by Tikhonov (1950) and Cagniard (1953). Tikhonov showed that at low frequencies the derivative of the horizontal magnetic field \( H \) is proportional to the orthogonal component of the electric field \( E \). Cagniard, often credited to be the "father" of MT, developed formulas relating \( E_x \) and \( H_y \) on the surface of a layered medium with a plane wave source. Vozoff (1972, 1991) wrote good summary papers on the applications and background of MT.

The amplitude, phase, and directional relationships between electric \( E \) and magnetic \( H \) fields on the surface depend on the distribution of electrical resistivity in the subsurface (Vozoff, 1991). Depending on signal frequency and resistivity of material being studied, the MT method can resolve geoelectric structure from depths of tens of meters to depths of tens of kilometers. Lower frequency signals with longer wavelengths, obtainable by longer recording time, have greater penetration depth.

MT relies on two sources, lightning and solar wind. Lightning produces electromagnetic (EM) waves in a frequency range from approximately 1 Hz to 10 kHz, with which we can probe the shallower depth of the subsurface. Lightning discharges generate Schumann resonances (approximately 8 Hz) within the earth-ionosphere cavity. The ionosphere, present above 80 to 160 km above the Earth’s surface, is an electrically conducting set of layers of the Earth's atmosphere consisting of ionized rarefied atmospheric gases caused by incident solar radiation (Kaufman and Keller, 1981).

Induction of current in the ionosphere leads to displacement of mass and interaction of magnetic and inertial (resistance to acceleration) forces that gives rise to hydrodynamic waves and trapped charged particles. The flow of ionized particulate material thrown off by the Sun, solar wind, produces EM waves in a frequency range from approximately 0.001 Hz to less than 1 Hz (Vozoff, 1991). Solar wind streams past the Earth at speeds of more than 500 km per second causing complex interactions with the Earth’s magnetic field (NASA, 2003). A fundamental property of magnetic fields is that they exert forces on moving electrical charges. When solar wind encounters the
Earth's magnetosphere, electrons and protons are deflected in opposite directions giving rise to an electrical current in the plasma and a magnetic-field effect. This interface moves back and forth erratically as energy of solar wind is arriving. Resulting magnetic effects that arise at the magnetopause are strongly modified by the time they penetrate the Earth's surface and are observed (Kaufman and Keller, 1981).

The MT method depends on the penetration of the EM energy into the Earth. The MT method operates under two main assumptions: First, we assume a quasistatic approximation by neglecting displacement currents. Mathematically, the wave equation of EM propagation becomes the diffusion equation. Secondly, we assume a plane wave source. After impinging upon the Earth, the natural EM fields propagate essentially vertically into Earth because of the large resistivity contrast at the air-Earth interface, which causes a vertical refraction of both fields transmitted into the earth (Vozoff, 1972). The changing horizontal magnetic field induces a changing horizontal electric field at right angles through Faraday’s law. The electric field in the conducting earth drives the telluric currents (Vozoff, 1991).

An MT station is used to measure the $E$ and $H$ fields at any particular time and location to obtain sample time records. MT is a frequency-domain method, therefore we extract frequency information from time series segments. The MT stations measure two orthogonal electrical components and three orthogonal magnetic components. The station spacing, the resistivity structure, and the frequencies recorded determine lateral resolution of the interpreted resistivity model (Vozoff, 1991).

Processing of MT data begins with conversion of time series segments into frequency domain fields by Fourier Transforms. Amplitude and phase characteristics as a function of frequency are expressed in real and imaginary parts. We measure a vector $E$ field and a vector $H$ field and relate them with an impedance tensor. This two by two tensor has two principal components, which in a 1D or 2D situation constitute two independent EM polarizations. Because the components are complex, we examine magnitude and phase separately (Park, 2003, written communication). Resulting
apparent resistivity and phase curves are analyzed and modeled to derive an interpreted
g eolectric structure of the MT transect. If the Earth’s subsurface is homogeneous and
isotropic, interpretation of the EM field yields true resistivity of the subsurface. For a
heterogeneous subsurface, apparent resistivity can be calculated from the amplitude,
phase, and directional relationships between electric and magnetic fields on the surface

The resistivity of geologic units is largely dependent upon their fluid content,
porosity, degree of fracturing, temperature, and conductive mineral content. Fault zones
will show low resistivity (less then 100 Ωm) when they are comprised of rocks fractured
enough to have hosted fluid transport and consequent mineralogical alteration.
Unaltered, metamorphic rocks (non-graphitic) have moderate to high resistivity
(hundreds to thousands of Ωm) (Keller, 1987).

The sensitivity of MT to conductive regions is the reason why the method is often
used to examine anomalous crust such as the Kenai MT study area. The Kenai MT data
is expected to provide a depth constraint on sources of a magnetic high and help to
answer the question whether it could be trapped and altered oceanic crust or underplated
 mafic magmas.
The Kenai MT transect is approximately perpendicular to the north-south trending Border Ranges Fault, with one MT station on the east side and five MT stations on the west side. The six stations have an average spacing of 7.4 km forming a survey line approximately 38 km long. Table 1 includes site locations and dipole lengths for every station. Two stations were collected during a previous survey carried out on 10 July 2001 by the USGS (Sampson, 2003, USGS, written communication). The two stations, labeled k001 and k002 in Table 1, were recorded using an MT-1 system made by ElectroMagnetic Instruments Incorporated (EMI, 1996). The MT-1 is a broad band, lightweight system with three magnetic and two electric dipole sensors. Louise Pellerin, Jonathan Goold and I acquired four additional stations that complement stations k001 and k002. The focus of this chapter is on the acquisition of the four additional stations, labeled k201 through k204 in Table 1.

2.1 Equipment

Stations k201 through k204 were recorded using the MT-24 system. The MT-24 configuration recorded orthogonal electric and magnetic fields, which were processed to provide tensor impedances for interpreting two-dimensional (2D) geoelectric structure.
Table 1 MT station location, dipole length, azimuth, and electrode contact resistance.

Sites k201, k202, k203, and k204 have azimuth measured clockwise from geographic north. Sites k001 and k002 have azimuth measured clockwise from magnetic north.

<table>
<thead>
<tr>
<th>Site ID</th>
<th>k201</th>
<th>k202</th>
<th>k203</th>
<th>k204</th>
<th>k001</th>
<th>k002</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude, NAD27</td>
<td>60.4395</td>
<td>60.49127</td>
<td>60.48458</td>
<td>60.43477</td>
<td>60.44956</td>
<td>60.472</td>
</tr>
<tr>
<td>Longitude, NAD27</td>
<td>-150.317</td>
<td>-150.614</td>
<td>-150.804</td>
<td>-151.116</td>
<td>-150.943</td>
<td>-150.405</td>
</tr>
<tr>
<td>Elevation (m)</td>
<td>71</td>
<td>59</td>
<td>82</td>
<td>47</td>
<td>85</td>
<td>95</td>
</tr>
<tr>
<td>Dipole Length (m) of Ey</td>
<td>64.5</td>
<td>90</td>
<td>90</td>
<td>90</td>
<td>30</td>
<td>30</td>
</tr>
<tr>
<td>Dipole Length (m) of Ex</td>
<td>90</td>
<td>90</td>
<td>90</td>
<td>67</td>
<td>30</td>
<td>30</td>
</tr>
<tr>
<td>Azimuth of Ey and Hy</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>47</td>
<td>35</td>
<td>0</td>
</tr>
<tr>
<td>Azimuth of Ex and Hx</td>
<td>90</td>
<td>90</td>
<td>90</td>
<td>137</td>
<td>125</td>
<td>270</td>
</tr>
<tr>
<td>Contact Resistance (kΩ) of Ex</td>
<td>9.8</td>
<td>12</td>
<td>2.3</td>
<td>5.5</td>
<td>4.7</td>
<td>9.75</td>
</tr>
<tr>
<td>Contact Resistance (kΩ) of Ey</td>
<td>6.6</td>
<td>1</td>
<td>1.2</td>
<td>7.8</td>
<td>3.2</td>
<td>10.75</td>
</tr>
</tbody>
</table>

The coordinate system for electric and magnetic field directions are defined with y parallel to strike, x perpendicular to strike, and z pointing vertically downward. The acquired user coordinate system generally defines strike in a north-south direction.

Three permalloy-cored induction coils measured orthogonal x, y, and z components of the magnetic field. The two horizontal coils were buried in the ground, and the vertical coil was only partially buried. To help provide stability and maintain constant temperatures, dirt mounds were made around exposed portions of the vertical magnetic coil and an upside-down plastic container was secured over the coil.

The two horizontal electric-field measurements paralleled the horizontal magnetic-field measurements. Nonpolarizable lead-lead chloride electrodes were used for potential-difference measurements. They are made of lead wire placed in a porous container filled with saturated lead-chloride and sodium-chloride solutions. These electrodes stabilize within 30 minutes to one hour. Electrode contact resistance for Kenai stations is given in Table 1.
Once the layout of an MT site is determined, electrodes are placed in the ground first. The bottom of holes dug for electrodes were lined with clay surrounded by dirt saturated with salt water for optimal contact resistance when necessary. By the time the rest of the station is set-up and ready to record electromagnetic signals, the electrodes have stabilized (Petiau and Dupis, 1980).

Field acquisition modules (FAM) for the MT-24 system receive power from 12 volt batteries. The computer-storing unit (CSU) is powered by a separate 12 volt battery to prevent spiking of data being recorded when voltage of the battery drops due to transmission of data through a cable. The typical station set up is shown in Figure 2.1.

![Figure 2.1 General layout of the MT-24 stations.](image)
2.2 Site Location and Field Conditions

In an MT survey, we need to know the surrounding of each site since there are often sources of noise and interference with equipment. Site selection depends on scientific objectives, accessibility, and sources of cultural noise. Natural and cultural noise that can affect data quality include power lines, pipelines and fences, active byways, curious animals, severe close-by thunderstorms, high winds, extreme temperature fluctuations, and topographic features. Burying equipment in the ground and placing heavy objects along the length of wire connecting electrodes help to minimize noise effects of weather and animals. Temperature fluctuation is less below the ground surface and weighting down the wire minimizes wind effects. Careful site selection and application of appropriate signal processing techniques also contributes to data quality. Reasonable site locations at appropriate spacing intervals with adequate access were found in Kenai.

Site location and field conditions for collection of MT-24 data were generally fair. Each station was set on flat terrain and the weather varied from sunny to light rain. Station k201, at Upper Skilak Lake Campground, was in a relatively undisturbed area. Station k202, in a wet field, was near a dirt road with about two cars passing each hour during the day. Power lines and phone lines were approximately 300 meters from the site, but prominent spikes were not apparent in the data. Station k203, in an open field with dry soil, was approximately one kilometer from power lines in a quiet area of suburban Soldatna. Station k204, in a field near a gravel pit, was 7.5 km from the Soldatna Homer Highway (to the west) and about one kilometer north of stationary heavy machinery. Sensor orientation of k204 had to be rotated 47 degrees to keep dipoles between a gravel pit and dirt road (Pellerin, 2002, written communication). Data recorded at this site were especially noisy. In general, field conditions allowed reasonable lengths of time series to be recorded.
2.3 Recording Data

MT equipment allows simultaneous measurements of the EM field. The phase and amplitude of the $E/H$ ratios, as a function of frequency, are dependent on the electrical resistivity within the Earth.

MT data were recorded using both continuous recording and stacking mode for collection of raw time series. Global Positioning System (GPS) synchronized continuous recording of two stations for 13.5 to 14 hours at 50 Hz. Three to four 10 minute recordings at 100 Hz for each station were stacked. A low-pass filter applied prior to digitization of the data avoided aliasing.

Stations k201 and k202 were installed in the afternoon on 24 July 2002. The next day, we moved equipment to install stations k203 and k204. Once we set each station, approximately ten minutes of recordings allowed least-squares analysis to be completed for high-band frequencies. The first station of each day had a timer set to begin recording when we predicted the second station would be ready to record. Each pair of stations were synchronized by means of the Global Positioning System (GPS) to within +/- 1 microseconds (EMI, 2002), which allowed remote-reference processing for middle and low-band frequencies. Sites k201 and k202 were synchronized in recording for 13.5 hours and sites k203 and k204 were synchronized in recording for 14 hours. The ACQ24 Version 6.4 program was used to read and combine time series segments for stations that recorded simultaneously, a segment of which is shown in Figure 2.2 (EMI, 2002).
Figure 2.2 Time series segment of approximately 2.5 minutes for stations k201 and k202. Electric fields are shown in red and magnetic-fields are shown in blue.
CHAPTER 3

PROCESSING

The objective of data processing is to extract sets of smooth, repeatable earth response functions from time series signals. This chapter begins with background information on apparent resistivity then discusses processing steps used for the Kenai data. A flow chart of processing steps is shown in Figure 3.1. The data processing shown in Figure 3.1 resulted in reliable values of earth response functions for the Kenai data.

**Figure 3.1** Processing steps used for the Kenai data.
3.1 Skin Depth and Impedance Tensor

Conductivity removes (attenuates) energy from the EM wave through the work of moving charge. Higher frequency EM waves lose energy more quickly than low-frequency waves because they move more charge in a given time. The depth at which the amplitude of a plane EM wave will be attenuated to \( \frac{1}{e} \approx 0.37 \) of its surface amplitude is called the skin depth, \( \delta \). Skin depth gives a relative measure of how deep the EM wave can penetrate into the Earth. Skin depth is related to resistivity as:

\[
\delta = (\pi \mu \sigma)^{-\frac{1}{2}} \approx 0.503 \sqrt{\frac{\rho}{f}} \text{ (km)},
\]

where

\[
f = \text{frequency in Hz},
\]

\[
\sigma = \text{conductivity in } \frac{S}{m},
\]

\[
\rho = \text{resistivity in } \Omega \cdot m,
\]

\[
\mu = \text{magnetic permeability in } \frac{H}{m}.
\]

The answer with \((\pi \mu)^{-\frac{1}{2}} = 0.503\) assumes a free space permeability of \(4\pi \cdot 10^{-7} \text{ H/m}\) (Vozoff, 1991). The approximate depth of penetration for the MT-24 stations is 25 km using an averaged reliable lowest frequency recorded of 0.01 Hz and assuming an average resistivity of 25 \( \Omega \cdot m \). The earlier MT-1 stations were recorded for only a couple of hours, resulting in smaller penetration depths.

The impedance tensor relates electric-field measurements to magnetic-field measurements. Assuming a 1D earth, the impedance \( Z \) is given by:

\[
Z_{xy} = \frac{E_x}{H_y} = \frac{\omega \mu}{k} = (1 + i \left( \frac{\omega \mu}{2 \sigma} \right)^{\frac{1}{2}}),
\]

where
\[ E_x = \text{electric field in the } x \text{ direction in } \frac{V}{m}, \]
\[ H_y = \text{magnetic field in the } y \text{ direction in } \frac{A}{m}, \]
\[ \omega = \text{angular frequency}, \]
\[ \mu = \text{magnetic permeability in } \frac{H}{m}, \]
and \( k = \text{propagation constant, or wave number (Vozoff, 1991).} \)

Apparent resistivity, \( \rho_a \), for a layered earth is given by:

\[
\rho_a = \frac{1}{\omega \mu} \left| \frac{E_x}{H_y} \right|^2, \tag{3}
\]

(Cagniard, 1953). Phase is proportional to the slope of the apparent resistivity curve on a log-log scale from a baseline at 45 degrees. In a uniform earth, apparent resistivity has to be the same at every frequency, and the electric field leads the magnetic field in-phase by 45 degrees at all frequencies (Vozoff, 1972). Apparent resistivity and phase are plotted on log-log scale versus frequency or period (inverse of frequency).

Two independent modes of the impedance are analyzed for 2D earth analysis in a right hand Cartesian coordinate system with \( y \) parallel to strike and \( x \) perpendicular to strike. Transverse electric (TE) mode is when the electric field is parallel to strike. Transverse magnetic (TM) mode is when the magnetic field is parallel to strike. Diagonal terms of the impedance tensor for a perfectly 2D earth are zero:

\[
Z = \begin{bmatrix}
0 & Z_{xy} \\
Z_{yx} & 0
\end{bmatrix}, \tag{4}
\]

where

\[
Z_{yx} = Z_{TE} = \frac{E_y}{H_x},
\]
\[
Z_{sy} = Z_{TM} = \frac{E_x}{H_y}.
\]
Assuming that the data are acquired in the user coordinate system \((x', y')\) and the structural coordinate system is \((x, y)\) where \(y\) is parallel to strike, MT data is rotated from \((x', y')\) to \((x, y)\). If the rotation angle from \((x', y')\) to \((x, y)\) is \(\theta\), then we apply the rotation matrix:

\[
R = \begin{pmatrix}
\cos \theta & \sin \theta \\
-\sin \theta & \cos \theta
\end{pmatrix}
\]

such that

\[
E = RE' \quad \text{and} \quad H = RH'.
\]

Under such a rotation, we obtain an impedance tensor in equation (4). In practice, we do not know \(\theta\) exactly, and the geology is not strictly 2D. Therefore, we estimate \(\theta\) by minimizing the diagonal terms \(Z_{xx}\) and \(Z_{yy}\) of the rotated impedance tensor \(Z\) (Swift, 1967).

Using off-diagonal elements, apparent resistivity, \(\rho\), is defined as a function of frequency by:

\[
\rho_{xy} = \frac{1}{\omega \mu} \left| Z_{xy} \right|^2,
\]

and

\[
\rho_{yx} = \frac{1}{\omega \mu} \left| Z_{yx} \right|^2,
\]

where

\[
\omega = \text{angular frequency}, \quad \text{and} \quad \mu = \text{magnetic permeability}.
\]

When geology approximates the 2D case, rotated MT data has diagonal elements very close to zero. Since MT is sensitive to vertical and horizontal variations in resistivity, the method is capable of establishing whether EM fields are responding to 1D, 2D, or 3D subsurface structures (Vozoff, 1991). Groom Bailey analysis, impedance polar diagrams, tipper, and skew can show how strike direction may be changing with depth and dimensionality of the subsurface – helping to find an optimal \(\theta\) for calculation of a more
accurate impedance tensor. The impedance tensor is also referred to as the MT Transfer Function, and I will use the two terms interchangeably.

3.2 Robust Processing for MT Transfer Functions

Smooth robust estimates made by the least-squares and remote reference methods yield MT transfer functions. The objective of smooth robust estimates is to find MT transfer functions with the least amount of curvature, meaning the least amount of interpreted layers, consistent with most of the data. The transfer function is found iteratively and the process converges within six to eight iterations. The MT transfer function is made robust by frequency- and time-domain weights that remove the effects of outliers. The set of weights should increase the coherence between $E$ and $H$ data while down-weighting small ($<20\%$) percentages of data (Larsen, 1989). Outliers may be caused by instrumental and source problems. If the weighted residuals for remote reference data have uncorrelated noise, then the contribution to the noise by the electric and magnetic data can be estimated and used to evaluate the least-squares and remote reference estimates. Smooth continuous representation is necessary for estimating geomagnetic variations at all frequencies because the MT transfer function is expected to be smooth and continuous for most geologic sites that are not close to surface discontinuities (Larsen, 1989).

Least-squares frequency band estimates are found using electric and magnetic data from a single site. This method minimizes variance of residuals. Residuals, $D$, are found by:

$$D = E_{obs} - E_{pred},$$  \hspace{1cm} (7)

where
\[ E_{\text{obs}} = \text{observed electrical data and} \]
\[ E_{\text{pred}} = \text{electrical data predicted from magnetic data.} \]

Residuals found to scatter around zero are considered good. Least-squares method estimates are biased by noise in magnetic data (Larsen, 1989). The remote reference method attempts to alleviate this bias by obtaining reference magnetic fields at a separate station where noise is expected to be uncorrelated with fields at the local MT station (Gamble et al., 1979). This method uses simultaneous local and remote magnetic data to minimize the modulus of the covariance between locally and remotely derived residuals (Larsen, 1989). Estimates are unbiased if noise is uncorrelated between the remote and local stations, therefore improving upon estimates made from the least-squares method.

Residuals are related to the MT impedance tensor by the following equations:

\[ E = H \cdot Z + D, \]
\[ E = \underline{H} \cdot Z + \underline{D}, \]

where
\[ E = \text{frequency domain electric data}, \]
\[ H = \text{local magnetic data}, \underline{H} = \text{remote magnetic data}, \]
\[ Z = \text{MT transfer function}, \]
\[ D = \text{residuals derived by the standard method}, \]
\[ \underline{D} = \text{residuals derived by the remote reference method}. \]

Noise variance estimates for calculated MT transfer functions are made with the assumption that the noise resides mainly in the weighted electric field. Four weights are applied to the observed data in the following order: (1) frequency weights that prewhiten the continuum part of the residuals making them nearly independent, (2) frequency weights that eliminate any spectral peaks in the residuals, (3) time weights that eliminate the outliers in the time domain, and (4) frequency weights that postwhiten the weighted residuals. In general, confidence limits on results can be readily constructed because the weighted residuals are found to be approximately independent, identically, normally
distributed (Larsen, 1989). Both least-squares and remote reference methods are used in robust processing to calculate the MT transfer function for the Kenai MT data.

The least-squares method is used for robust processing of MT transfer functions for all frequencies of the two earlier stations k001 and k002, and for high-frequency band transfer functions of stations k201, k202, k203, and k204. The high-frequency band data, recorded in stacking mode at 500 Hz, acquired 300,000 data points from each recorded time segment of approximately 10 minutes. The number of high-frequency band runs recorded for each station included two for k201, three for k202, three for k203, and four for k204. More time segments are recorded for stations with noisy signal.

Remote reference method is used for robust processing of MT transfer functions for middle- and low-frequency band data of stations k201, k202, k203, and k204. We used cascade decimation, which down-sampled data that were recorded continuously at 50 Hz, for middle- and low-frequency band calculations. In cascade decimation, sine and cosine transforms are applied to sequences of data produced by successively applying a low-pass digital filter to original data. Resulting spectra are the average of estimates that are independent in time and represent constant percentage bandwidths (Wight and Bostick, 1980). Sites k201 and k202 had a total of 2,424,750 synchronized samples, while sites k203 and k204 had a total of 2,520,000 synchronized samples. The program used 8192 data points per time segment. For middle-band calculations, a decimation of two was applied to the data (25 Hz). For low-band calculations, a decimation of 32 was applied to the data (1.5625 Hz). Results for the processed and overlapped high-, middle-, and low-band calculations for the transfer functions and their associated errors are shown in Figure 3.2, Figure 3.4, Figure 3.5, and Figure 3.7. Results for the two earlier stations k001 and k002 are shown in Figure 3.3 and Figure 3.6.
Figure 3.2 Apparent resistivity and phase versus period for station k201.
Figure 3.3 Apparent resistivity and phase versus frequency for station k002.
Figure 3.4 Apparent resistivity and phase versus period for station k202.
Figure 3.5 Apparent resistivity and phase versus period for station k203.
Figure 3.6 Apparent resistivity and phase versus frequency for station k001.
**Figure 3.7** Apparent resistivity and phase versus period for station k204.
3.3 Rotation of Impedance Tensor and Earth Dimensionality

Apparent resistivity sounding curves depend on orientation of the axis. Due to resistivity distribution below the observation point, curves may be distorted by induction effect caused by excessive currents and galvanic effects caused by excessive charges (Berdichevsky, 1976). Optimal rotation angles with respect to principal directions of $Z$ were found by various methods including geologic strike estimated from a previous gravity survey, analysis of Groom-Bailey decomposition of the impedance tensor, tipper magnitude, and skew. Pages 23 and 24 discuss how data are rotated from an acquired user coordinate system to a structural coordinate system.

Based on gravity data, the regional geologic strike for Kenai appears to be between 17 and 23 degrees east from true north for all stations. Figure 3.8 shows MT station locations plotted on an isostatic residual gravity map. Angle estimates for regional geologic strike are measured with respect to the regional trend of the gravity data.

Groom Bailey Analysis

Groom Bailey analysis is used to separate local and regional parameters under the assumption that regional structure is at most 2D and local structure causes only galvanic (frequency independent) scattering of electric fields. Decomposition of the impedance tensor separates the effects of 3D channeling of electric currents from those of 2D induction. This method recovers principal axes of induction and two principal impedances in the form of a product factorization (Groom and Bailey, 1989).
Figure 3.8 Isostatic residual/gravity anomaly (Saltus et al., 2001). The MT station locations are plotted as dots. The regional geologic strike is measured between approximately 17 and 23 degrees from true north for all MT stations.
Decomposition of the impedance tensor yields:

\[ Z_m = R(\theta)TSZ_2R^T(\theta), \]  

where

\[ Z_m = \text{measured impedance tensor}, \]
\[ R(\theta) = \text{rotation matrix}, \]
\[ R^T(\theta) = \text{transpose of rotation matrix}, \]
\[ T = \text{twist tensor}, \]
\[ S = \text{shear tensor}, \]
\[ Z_2 = \text{regional induction impedance tensor scaled by unknown, but frequency-independent factors}. \]

The shear tensor, characterized by a shear angle equal to \(\tan^{-1} e\) where \(e\) is shear, detects anisotropy on axes that bisect the regional inductive principal axes. In foliated rocks the resistivity parallel to foliation is less than that perpendicular to foliation. The twist tensor rotates the electric-field vectors through a clockwise angle \(\tan^{-1} t\), where \(t\) is twist (Groom and Bailey, 1989). If the telluric distortion is truly frequency-independent, this method simultaneously extracts estimates of the distortion parameters \(e\) and \(t\), the regional strike \((\theta)\), and the modified 2D regional impedances by least-squares fitting of the model to the data. Any of the model parameters can be constrained (Groom and Bailey, 1991).

In viewing results of an unconstrained galvanic decomposition of the Kenai data, values for regional strike, shear, and twist tensors with the least amount of root-mean-square (RMS) relative error of fit are used for bounds in a constrained galvanic decomposition. Bounds for strike are also determined by regional strike estimated from gravity data. Figure 3.9 through Figure 3.11 shows resulting plots for the constrained galvanic decomposition. Regional strike is calculated from an initial value of 20 degrees with bounds of plus or minus three degrees for all stations. Initial values for twist and shear and respective bounds for each station are shown in Table 2.
Figure 3.9 Groom Bailey strike for MT stations.
Figure 3.10 Groom Bailey twist for MT stations.
Figure 3.11 Groom Bailey shear for MT stations.
Table 2 Twist and Shear bounds for MT stations.

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Twist</th>
<th>Shear</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>initial value</td>
<td>bound (±)</td>
</tr>
<tr>
<td>k001</td>
<td>5</td>
<td>5</td>
</tr>
<tr>
<td>k002</td>
<td>5</td>
<td>5</td>
</tr>
<tr>
<td>k201</td>
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<td>5</td>
</tr>
<tr>
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<td>10</td>
<td>5</td>
</tr>
<tr>
<td>k204</td>
<td>-5</td>
<td>2</td>
</tr>
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</table>

**Tipper Magnitude**

The tipper is a measure of the vertical magnetic field normalized by the horizontal magnetic field. In a 1D earth, the vertical magnetic component is zero. In general, for a multi-dimensional earth the vertical magnetic component is equal to $T_x H_x + T_y H_y$, where $T$ represents Tipper. Tipper magnitude is defined by:

$$|T| = \sqrt{|T_x|^2 + |T_y|^2}.$$  \hspace{1cm} (10)

The tipper helps to resolve ambiguity in strike, and helps to show which side of a contact is more conductive (Vozoff, 1991). The modulus of the tipper is rarely as great as 1, with 0.1 to 0.5 being the common range. Since the vertical component of the magnetic field is weak, the lower part of the range is often noisy. Tipper can show which side of a contact is more conductive because near a conductor-resistor boundary, the near-surface current density parallel to strike is larger on the conductive side (Vozoff, 1991). Plots of Tipper magnitude for stations k201 through k204 are shown in Figure 3.12. Tipper magnitude of stations k201 and k202 clearly define vertical structure on either side of a contact. The more conductive side is underneath station k202, where the Tipper magnitude is greatest.
Figure 3.12 Tipper magnitude for MT stations.
**Polar Plots and Skew**

Impedance polar plots provide a measure for the MT data dimensionality. Polar plots show the modulus of a component of the impedance tensor as a function of the rotation angle $\theta$ ($0 < \theta < 2\pi$) at different frequencies:

\[
\begin{align*}
Z_{xy}(\theta) &= Z_{xy} \cos^2 \theta + (Z_{yy} - Z_{xx}) \sin \theta \cos \theta - Z_{yx} \sin^2 \theta, \\
Z_{xx}(\theta) &= Z_{xx} \cos^2 \theta + (Z_{yy} - Z_{yx}) \sin \theta \cos \theta + Z_{yy} \sin^2 \theta.
\end{align*}
\] (11)

Principal impedances are $Z_{xy}$ and $Z_{yx}$, and diagonal impedances are $Z_{xx}$ and $Z_{yy}$ (Reddy et al., 1977). Analysis of the shape of polar plots provides information about the level of 3D distortion and/or noise that may occur within data. The diagonal impedance elements, that we try to minimize, are normalized with respect to principal impedance. For 1D geoelectric structure, the principal impedance polar diagrams are circles. For 2D or 3D structure, they elongate in a direction either parallel or perpendicular to strike, depending on the position of the node with respect to the discontinuity (Reddy et al., 1977).

Figure 3.13 shows how impedance polar diagrams at the surface of the earth appear surrounding a contact. The principal impedance, $Z_{xy}$, is shown by a thicker line than the diagonal impedance tensor, $Z_{xx}$. Over resistors the principal-impedance polar diagram elongates perpendicular to strike direction, and over conductors the principal impedance polar diagram elongates parallel to strike direction. For 2D resistivity structures, the diagonal impedance polar diagram, $Z_{xx}$, attains the shape of a symmetric cloverleaf. For 3D resistivity structures, the diagonal impedance polar diagram, $Z_{xx}$, elongates in one direction and its amplitude becomes comparable to that of the principal impedance polar diagram (Reddy et al., 1977). Figure 3.14 through Figure 3.19 for sites k201 through k204 show impedance polar diagrams for $Z_{xy}$ in blue and $Z_{xx}$ in red.

Interpretation of polar impedance plots is incorporated with interpretation and comparison of the impedance skew values for the six MT stations.
Figure 3.13 Tensor impedance polar diagrams at the surface of the earth. The inner diagram (thin line) is for a diagonal element normalized with respect to the outer diagram (thick line), which is for an off-diagonal element of the tensor impedance matrix (Reddy et al., 1977).
Figure 3.14 Station k201 polar impedance plots at frequencies indicated below each plot.

The blue indicates the principal impedance, $Z_{xy}$. The red indicates the diagonal impedance, $Z_{xx}$. 

Zxy  Zxx  Map View - North Up (Frequency in Hz)
Figure 3.15 Station k002 polar impedance plots at frequencies indicated below each plot.
The blue indicates the principal impedance, $Z_{xy}$. The red indicates the diagonal impedance, $Z_{xx}$.
Figure 3.16 Station k202 polar impedance plots at frequencies indicated below each plot. The blue indicates the principal impedance, $Z_{xy}$. The red indicates the diagonal impedance, $Z_{xx}$. 

Map View - North Up (Frequency in Hz)
Figure 3.17 Station k203 polar impedance plots at frequencies indicated below each plot. The blue indicates the principal impedance, $Z_{xy}$, The red indicates the diagonal impedance, $Z_{xx}$. Map View - North Up (Frequency in Hz)
Figure 3.18 Station k001 polar impedance plots at frequencies indicated below each plot.

The blue indicates the principal impedance, $Z_{xy}$. The red indicates the diagonal impedance, $Z_{xx}$.

Map View - North Up (Frequency in Hz)
<table>
<thead>
<tr>
<th>Z_{xy}</th>
<th>Z_{xx}</th>
<th>Map View - North Up (Frequency in Hz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>48.94</td>
<td>60.06</td>
<td></td>
</tr>
<tr>
<td>48.83</td>
<td>41.50</td>
<td></td>
</tr>
<tr>
<td>34.38</td>
<td></td>
<td></td>
</tr>
<tr>
<td>28.32</td>
<td>24.02</td>
<td></td>
</tr>
<tr>
<td>19.04</td>
<td>14.60</td>
<td></td>
</tr>
<tr>
<td>12.21</td>
<td></td>
<td></td>
</tr>
<tr>
<td>11.33</td>
<td>7.32</td>
<td></td>
</tr>
<tr>
<td>4.88</td>
<td>4.39</td>
<td></td>
</tr>
<tr>
<td>3.45</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2.93</td>
<td>2.44</td>
<td></td>
</tr>
<tr>
<td>1.76</td>
<td>1.20</td>
<td></td>
</tr>
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<td>0.83</td>
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<td></td>
</tr>
<tr>
<td>0.57</td>
<td>0.38</td>
<td></td>
</tr>
<tr>
<td>0.34</td>
<td>0.24</td>
<td></td>
</tr>
<tr>
<td>0.17</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Figure 3.19** Station k204 polar impedance plots at frequencies indicated below each plot. The blue indicates the principal impedance, $Z_{xy}$. The red indicates the diagonal impedance, $Z_{xx}$. 
Another measure of the dimensionality or noise for MT data is provided by the impedance skew of the impedance tensor (Vozoff, 1972). Skew is given by:

\[
S = \frac{|Z_{xx} + Z_{xy}|}{|Z_{xy} - Z_{yx}|}.
\]  

(12)

If the geology beneath an MT station in a noiseless environment is truly 1D or 2D, then the skew will be zero. Skew values for the MT-24 stations are plotted in Figure 3.20. Field skew below 0.2 are indicative of low noise level recordings. Skews values above 0.2 are an indication of 3D geology and/or higher levels of noise (Rodriguez, 2002, oral communication). Analyzing polar plots and skew values for each MT station in Kenai shows there is geologic structure ranging from 1D to 2D to either 3D geologic structure or noisy data.

For station k201, skew values above 0.2 only show above 30 Hz and below 0.003 Hz. This indicates there are low noise levels, meaning geologic structure can be accurately interpreted for this survey for most of the frequency range recorded. The polar plot at 48.94 Hz, with \(Z_{xy}\) in a peanut shape and \(Z_{xx}\) elongating in one direction, correlates with high skew for this frequency; however, adjacent polar plots indicate 2D geology. Thus, these high skew values most likely indicate noise. The polar plots below 2 Hz have skew values close to or below 0.1, meaning the plots of \(Z_{xy}\) and \(Z_{xx}\) in Figure 3.14 can be compared to the theoretical plots in Figure 3.13. Comparison of these plots with Figure 3.13 shows an agreement with Tipper magnitude that station k201 is located on the resistive side of a contact.

Station k202 has skew above 0.2 around 0.1 Hz, which correlates with larger calculated error bars on the apparent resistivity and phase curves. Frequencies above 10 Hz and below 0.001 Hz also have higher skew. Except for noise in part of the mid-band, station k202 recorded a similar frequency range of data with minimal error like the remote reference station k201. The peak skew values for k202 are in the frequency range of power line noise. Polar impedance plots are close to 1D for frequencies below 20 Hz.
Figure 3.20 MT stations skew plots. Skew values above 0.5 were eliminated when trying different starting models described in Chapter 4.
Station k002 has skew values above 0.2 for some frequencies above 90 Hz, between 0.3 and 1 Hz, and for the lowest frequency recorded at 0.009 Hz. Data were not acquired at lower frequencies because stations k002 and k001 only recorded for a few hours. Data is more accurate for frequencies between 30 Hz and 86 Hz at k001 and k002 compared to the other four stations. The polar plots between 24 Hz and 86 Hz for station k002 plot $Z_{xy}$ in almost circular shapes, which indicates 1D behavior and agrees with the skew values being below 0.1 for these frequencies. The polar plot at 0.34 Hz, with $Z_{xy}$ and $Z_{xx}$ comparable in size, correlates with the highest skew value around 0.5 for station k002 indicating noise rather than 3D behavior because adjacent polar plots indicate 2D behavior. Station k001 shows a similar polar plot at 0.34 Hz. Skew values are below 0.1 for frequencies between 0.1 and 0.03 Hz, indicating that noise is minimal, so the two lowest frequency polar plots in Figure 3.15 are comparable to Figure 3.13. These plots, with $Z_{xx}$ attaining the shape of a symmetric cloverleaf, show 2D resistivity structures. The two lowest frequency polar plots for station k001 in Figure 3.18, with skew values close to 0.1, similarly show 2D resistivity structures.

Skew is below 0.2 from 0.02 Hz to 0.4 Hz and again from 3 Hz and higher for station k203. Skew is above 0.2 for frequency ranges 0.002 Hz to 0.01 Hz and 0.5 Hz to 1 Hz, and apparent resistivity and phase curves have large error bars. Polar impedance plots generally show 2D behavior. This station has minimal help from its remote site, k204. More than half of the skew for station k204 plotted above 0.2. Skew as high as 0.8 appears around 0.9 Hz. Polar impedance plots for $Z_{xx}$ have comparable amplitude to $Z_{xy}$ around 0.9 Hz. Error bars are also very large for the apparent resistivity and phase curves. Sources of noise documented in Chapter 2 are most likely to blame for the results of station k204. 3D geology may also contribute to the scattered data of this station.
Inversion constructs a resistivity model from the MT data. It is necessary since MT data are integrated responses that are sensitive to a volume of resistivity in the subsurface. Therefore, interpretation is based on the inverted resistivity model. Discussion of inversion in this chapter covers background information, 1D inversion, and 2D inversion. Geophysical interpretation is based on the MT model, gravity data, and magnetic data. The MT model shows subsurface geometry. Density and susceptibility are assigned to different regions of subsurface that are identified from the inverted resistivity model. An integrated interpretation represented by distributions of resistivity, density, and susceptibility is thus formed and it is consistent with all three geophysical data sets. Geological interpretation is obtained from the integrated model.

4.1 Inversion

Inversion allows us to derive from geophysical data a model of physical properties describing the subsurface. A good inverted model is one that reproduces the observed data and is consistent with other available information. In MT, we try to find resistivity structure that is consistent with the processed apparent resistivity, phase, and tipper for a range of frequencies at each station along a survey line. We invert the six Kenai MT stations to construct a geo-electric section using a 2D algorithm. 2D inversion
required a reference model that represents the background resistivity structure. I choose
to establish such a reference model by using the inverted 1D model at station k201. I
used Occam’s 1D inversion for station k201 and an algorithm for 2D inversion of the
Kenai survey line described by Rodi and Mackie (2001).

### Background

1D inversion is done in the same way as 2D inversion in principle. The algorithm
is often referred to as Occam’s inversion since it looks for the simplest model that
explains the data. The term was coined by Constable et al. (1987), although the same
approach was used previously by others (e.g., Oldenburg, 1983).

The 2D inversion code described by Rodi and Mackie (2001) is based on
Tikhonov regularization, introduced by Tikhonov and Arsenin (1977). The inverse
problem is solved by minimizing an objective function, $S$. The objective function
consists of the weighted sum of the model objective function and data misfit:

$$S = \phi_d + \tau \phi_m,$$  \hspace{0.5cm} (13)

where $\phi_d$ is the data misfit and $\phi_m$ is the model objective function. $\tau$ is the
regularization parameter that controls the tradeoff between $\phi_d$ and $\phi_m$. $\tau$ is chosen so
$\phi_d$ is equal to a value consistent with the errors in the data. Larger values of $\tau$ lead to
smoother models at the expense of data fit.

The inversion code by Rodi and Mackie (2001) sets the model objective function
and data misfit equal to:

$$\phi_m = \|L(m - m_0)\|^2 \text{ and}$$

$$\phi_d = (d - F(m))^T R^{-1}_{dd} (d - F(m))$$
where
\[ L = \text{a linear operator}, \]
\[ m = \text{unknown model vector}, \]
\[ m_0 = \text{a prior model}, \]
\[ d = \text{observed data vector}, \]
\[ F = \text{forward modeling operator}, \]
\[ R_{dd} = \text{error covariance matrix}. \]

The linear operator is chosen as Laplacian, i.e. \[ L = \Delta : \]
\[
\|L(m - m_0)\|^2 = \int (\Delta(m(x, z) - m_0(x, z)))^2 \, dxdz. \tag{14}
\]

(Rodi and Mackie, 2001). The model vector contains resistivity values in cells of the
discretized model:
\[ m(x, z) = \log \rho(x, z), \tag{15} \]
where \( m = (m_1, m_2, \ldots, m_m) \). Each datum \( d_i \) is either the log amplitude or phase of TE or
TM complex apparent resistivity or the tipper. The forward modeling operator calculates
the theoretical response due to an assumed set of subsurface conductivities.

Solution of the inverse problem is done by an iterative, non-linear conjugate
gradients (NLCG) method. NLCG applies directly to the minimization of \( S \). The model
sequence is given by:
\[ m^{j+1} = m^j + \alpha^{j+1} h^{j+1} \tag{16} \]
where \( h^{j+1} \) is a search direction obtained from NLCG and \( \alpha^{j+1} \) is computed to minimize
\( S \) by a line search (Rodi and Mackie, 2001).

The average misfit between the predicted data and the observed data is
represented by the RMS error, which is different from how RMS is usually defined. The
RMS error for the 2D inversion program is defined by:
\[
\sqrt{\frac{(d - F(m))^T R_{dd}^{-1} (d - F(m))}{N}}, \tag{17}
\]
where \( N \) is the number of data points (Mackie et al., 1997). \( R_{dd} \) is the error covariance matrix whose diagonal terms are the variance of each datum. This matrix allows the user to define a noise floor value to keep the inversion from trying to reproduce noise. An optimally chosen value of \( \tau \) should yield an RMS error of 1.0 assuming \( R_{dd} \) is correct. Noise unaccounted for in the data and 3D effects may prohibit the inversion code from reaching the target RMS misfit values, which is especially true for TE mode data. Typical values for most MT inversions have an RMS error between 3 and 300 (Mackie et al., 1997) when the error covariance derived from robust impedance estimation is directly used in the inversion.

**1D Inversion**

A 1D Occam inversion model from station k201 provides good estimates on what resistivity values to use below 23 km depth for the starting model of the 2D inversion. This station is good for 1D analysis because it is located off of the high magnetic anomaly and because the data contains small amounts of noise. Figure 4.1 shows the result from 10 iterations of 1D inversion on a geometric mean average of the TM and TE mode data. MT processing software called Geotools® (1997) is used for this inversion.

**2D Inversion**

Accuracy of a 2D MT inversion is affected by data quality and mesh design. Dimensions of the grid cells must be kept small compared with the skin depth in order that the numerical solution may be generally valid (Uluggergerli and Candansayer, 2002). The mesh is made with rectangular cells 42 columns wide and 28 rows deep. Cell width gradually increases the further away its location is from an MT station. Cell thickness gradually increases with depth.
To speed up the 2D inversion, an initial model based on the 1D inversion of station k201 is used below 23 km depth. The initial model is a layer-cake subsurface with resistivity assigned to each layer according to the approximate resistivity value derived from the 1D inversion discussed in the preceding subsection. The first layer of the model is approximately 23 km thick and consists of 13 rows of cells. This initial model, homogeneous above 23 km depth, allows the 2D inversion to try to minimize lateral and vertical resistivity gradients in the region of the 2D model.

Figure 4.1 1D Occam inversion of station k201 using averaged TE and TM mode data.
The NLGC algorithm described by Rodi and Mackie (2001) is used to invert the data set for a 2D model of resistivity beneath the Kenai MT transect. Results from three models are analyzed. All three models used the starting model described above, with resistivity structure below 23 km depth based on the 1D inversion of station k201. Model-I inverted TE mode data, TM mode data, and tipper. Results of model-I show it would be best to initially exclude the noisier TE mode data. Model-II inverts TM mode data and tipper. Model-III inverts TM mode data, TE mode data, and tipper. The TE mode data is assigned higher standard deviations in Model-III.

Model-I

Input data for model-I includes tipper, apparent resistivity for TE and TM modes, and phase for TE and TM modes covering frequencies in the range from 0.003 Hz to 100 Hz. Static shift factors are assigned to stations k203 and k001 for TE and TM mode in the model input files. Apparent resistivities are multiplied by the designated static shift factor before inversion begins. Static shift factors are determined from analysis of apparent resistivity curves. Table 3 shows input static shift factors for the two stations that use them. Station k204 is too noisy to determine an appropriate static shift factor. The static shift factors are used for all 2D inversion models. The first run of 100 iterations for model-I contains all data points with noise standard deviations set to a floor value of 5%. Input errors based on robust estimation that are below the floor value are reset to the floor value. RMS error for model-I is given in Table 4. Model-I inserts a relatively large and geologically unreasonable conductor west of the sites. Comparison of the predicted data with the observed data shows large misfits, especially in the tipper. The large RMS values for the entire model, especially the TE mode, indicate it would be helpful to increase the noise floor of the TE mode data. To estimate a more reliable noise level in the TE mode data and try to assess possible 3D geologic effects, model-II is inverted without TE mode data.
Table 3 Input values for the static shift factors of stations k001 and k203. Apparent resistivities are multiplied by the designated static shift factor before inversion begins.

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Static Shift Factor</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>TE mode</td>
</tr>
<tr>
<td>k001</td>
<td>0.85</td>
</tr>
<tr>
<td>k203</td>
<td>0.69</td>
</tr>
</tbody>
</table>

Table 4 RMS error for the first model with error floors set to 5% for TM mode data, TE mode data, and tipper.

<table>
<thead>
<tr>
<th>Site ID</th>
<th>RMS error</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>TM mode</td>
</tr>
<tr>
<td>k201</td>
<td>2.4</td>
</tr>
<tr>
<td>k002</td>
<td>6.17</td>
</tr>
<tr>
<td>k202</td>
<td>5.92</td>
</tr>
<tr>
<td>k203</td>
<td>11.53</td>
</tr>
<tr>
<td>k001</td>
<td>5.23</td>
</tr>
<tr>
<td>k204</td>
<td>12.92</td>
</tr>
<tr>
<td>average</td>
<td>7.36</td>
</tr>
<tr>
<td>total average</td>
<td>8.93</td>
</tr>
</tbody>
</table>

Model-II

To better define resistivity values, input parameters are redefined to eliminate TE mode data and a few erroneous TM mode resistivity and phase data points from sites k203 and k204 for model-II. Error floors are kept at 5%. After 400 iterations, model-II obtains an average RMS value of 3.08. This model is shown in Figure 4.2. The significant decrease in RMS for model-II shows that the TE mode data are definitely in need of increased error floors.
Model-III

The final inversion model includes apparent resistivity for TE and TM mode data, phase for TE and TM mode data, tipper, and errors for all data covering frequencies in the range from 0.003 Hz to 100 Hz. The standard deviation for TM mode data and tipper is set to 5%. The error for TE mode data is set to 20% to reduce the effects of noise in TE mode data. A summary table of RMS error for specific modes at specific stations is given in Table 5. The model is shown in Figure 4.3.

Table 5 Summary table of RMS error calculated for the final model shown in Figure 4.3.

<table>
<thead>
<tr>
<th>Site ID</th>
<th>RMS error</th>
<th>RMS error</th>
<th>RMS error</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>TM mode</td>
<td>TE mode</td>
<td>Tipper</td>
</tr>
<tr>
<td>k201</td>
<td>1.59</td>
<td>3.14</td>
<td>2.79</td>
</tr>
<tr>
<td>k002</td>
<td>5.26</td>
<td>4.12</td>
<td>3.48</td>
</tr>
<tr>
<td>k202</td>
<td>1.9</td>
<td>2.53</td>
<td>6.12</td>
</tr>
<tr>
<td>k203</td>
<td>1.25</td>
<td>1.65</td>
<td>1.55</td>
</tr>
<tr>
<td>k001</td>
<td>4.22</td>
<td>6.13</td>
<td>5.1</td>
</tr>
<tr>
<td>k204</td>
<td>3.18</td>
<td>3.32</td>
<td>2.7</td>
</tr>
<tr>
<td>average</td>
<td>2.90</td>
<td>3.48</td>
<td>3.62</td>
</tr>
<tr>
<td>total average</td>
<td>3.83</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Verification of 2D Inversion

Additional inversions are performed using starting models made from the final model in Figure 4.3. Prominent resistors or conductors are eliminated from model-III, then run through 25 iterations to see if the inversion requires each feature. Data with high skew, above an impedance ratio of approximately 0.5, are eliminated prior to running each test model. The conductor, in the upper 20 km beneath and west of sites k001 and k204, is required by the inversion. The resistor in the upper 30 km beneath and
Figure 4.2 Results from 400 iterations of 2D inversion for TM mode data and tipper from all MT stations.

Figure 4.3 Results from 380 iterations of 2D inversion for TM mode data, TE mode data, and tipper from all MT stations.
east of k002 and k201 also prove to be required by the inversion. The small conductor in
the upper 7 km beneath station k201 also reappears. The reliability of the shallow
resistive bump beneath station k002 is uncertain. Further study could employ hypothesis
testing to verify the requirement of this feature by the data. Figure 4.4 shows comparison
plots of how well predicted data from model-III shown in Figure 4.3 fit observed data.
TM results are plotted in orange squares for predicted data, and blue triangles for
observed data. TE results are plotted in green circles for predicted data, and small
magenta circles for observed data.

4.2 Integration of Results

The MT model allows us to define the geometry of different regions in the
subsurface to approximately 25 km depth. Once the geometry of subsurface bodies is
defined, assignment and adjustment of susceptibility and density values is done.
Incorporating resistivity from MT, density from gravity, and susceptibility from magnetic
data produces a geophysical model with which we can interpret geological structure.

An airborne magnetic map (Saltus et al., 2001) for the Kenai study area is shown
in Figure 4.5. Site locations of the MT-24 stations are indicated with blue dots, and site
locations of the two stations acquired earlier are indicated with red dots. Gravity and
magnetic profile data are shown with the final MT model calculated with TE mode data,
TM mode data, and tipper in Figure 4.6.

Integrating the results from gravity, magnetics and MT show a few main results
on geoelectric structure. The Border Ranges fault is more resistive to the east. There is a
deep conductive zone beneath the Cook Inlet approximating 12 km height by 22 km
width with the top at approximately 10 km depth. The conductive zone ramps toward the
surface at the edge of the basin.
**Figure 4.4** Comparison plot of model from Figure 4.3 versus MT data. TM results are in orange squares for the model, and blue triangles for data points. TE results are in green circles for the model, and small magenta circles for data points.
Figure 4.5 Aeromagnetic map of Kenai study area.
Figure 4.6 Integrating magnetic, gravity, and MT results. The Border Ranges fault is on the eastern edge and the Cook Inlet sediments are on the western edge of the MT model.
Inversion of the Kenai MT data provides a resistivity model that allows improvement on calculated structure for magnetic and gravity data. Calculated data are much closer to observed data for estimations of density contrast and magnetic susceptibility. The improved results for density contrast and magnetic susceptibility, combined with the modeled resistivity of subsurface structures from MT, aids in understanding geologic structure.

4.3 Geologic Interpretation

First, geologic conclusions that can be drawn from the final geophysical model in Figure 4.6 are examined. Next, I show how the MT model helps to answer earlier questions posed about the magnetic anomaly high in the Kenai area. In attempting to envision how geologic conclusions within the MT model fit into the regional geological framework of southern Alaska, comparison of the potential field data to cross-sections derived from seismic data and well information will conclude this chapter.

MT model-III provides two general structural conclusions in Kenai; one being the presence of a distinctive trend dipping approximately 45 degrees northwest from the surface location of the fault, another being the offset of geo-electrically contrasting material on either side of the Border Ranges fault. On the west side of the fault, there is a very shallow layer that is low density, weakly magnetic, and conductive that probably corresponds to water and saturated sediments. Then there is a deeper, but still shallow, layer that is low density, non-magnetic and resistive and presumably corresponds to the deeper Cenozoic deposits and possibly a portion of the Talkeetna formation. Finally, there is the deep (> 10 km) zone that we model as high density, magnetic, and conductive. Similarly, in the east there are several layers to the model. The gravity high, magnetic high, and resistivity high beneath station 002 on the eastern side of the fault suggests that intrusive rocks are pervasive near this station. Low-density rocks calculated
from gravity data correlate with low-resistivity rocks in the subsurface at the center of model-III. High-density rocks correlate with high-resistivity rocks in the subsurface of the eastern half of model-III. A dipolar anomaly in magnetic data correlates with where the Border Ranges fault nears the surface in model-III with higher-resistivity rocks in the east and lower-resistivity rocks in the center. The magnetic data show a gentle increasing trend to the west suggesting subsurface rocks of high magnetic susceptibility.

The question of origin for the high magnetic anomaly in the Kenai area requires a geological understanding of the area. Earlier hypotheses for the cause of the magnetic anomaly high include correlation with Jurassic arc-related rocks and their basement, Wrangellia composite terrain (Saltus et al., 1999) and speculation about possible serpentinite zones (Saltus et al., 2001). The MT data shows a deep-seated (> 10 km) conductive body that coincides with the position of the magnetic anomaly high. Extending northeast from this conductive body towards the surface is a conductive region indicative of the dip of the Border Ranges Fault. Therefore, the MT results suggest that the causative body for the magnetic high, which correlates spatially with the mapped MT feature, has a similar geometry. In particular, the MT results strongly suggest that an anomalously conductive feature occupies a large portion of the mid to upper crust with its top at about 10 km beneath the Cook Inlet. This conductive feature appears to extend toward the surface west of the Border Ranges fault, therefore it is important to examine the geology along the BRF for candidate lithologies.

In viewing geologic cross-sections reported by Plafker and others (1994), it is obvious that information below 10 km is minimal in southern Alaska. From accreted rocks locally overlain by basinal deposits followed by extensive intrusion by tertiary granitoid plutons, it is hard to surmise what is below 10 km depth. Table 6 is helpful in understanding how possible geologic explanations are linked to the resistivity model by providing a layout of typical ranges of resistivities of earth materials (Palacky, 1987). The resistive zone in the upper 10 km near station k204 (Figure 4.6) is modeled with low density and moderate to low susceptibility, which could be indicative of porous
Table 6 Typical ranges of resistivities of earth materials (Palacky, 1987).
limestone. The material west of the Border Ranges Fault contains conductive regions, which could be caused by dissolved minerals in the electrolytically interconnected pore spaces where nearby minerals may have formed. Further information is needed to know geological composition. Figure 1.4 shows a map of the locations of cross-sections in Figure 4.7, which we use in making hypothesis of subsurface material causing the high magnetic anomaly. The cross-section labeled A-A’ is in an area with similar regional geologic trend as the MT model, though it is in the Gulf of Alaska. The cross-section labeled B-B’ is east of the Kenai MT model and transects a portion of the Southern Alaska magnetic high.

Cross-section A-A’ in Figure 4.7 provides an image of the accretionary prism based on seismic-reflection data. This cross-section is located approximately 175 km south of the MT model. Deep-marine rocks south of the Border Ranges fault system constitute one of the largest subduction-related accretionary complexes in the world with intermittent offscraping and underplating occurring from roughly late Triassic to the present (Plafker et al., 1994). It is postulated that Lower Jurassic to upper Triassic volcanic and sedimentary rocks make up the crust to the northwest of the BRF in the location similar to that of the Kenai transect.

The more speculative cross-section B-B’ in Figure 4.7, the location of which is shown in Figure 1.4, is approximately 200 km east of the Kenai transect. In the region of the conductivity high of the MT model, cross-section B-B’ shows that Plafker is postulating a combination of Jurassic mafic and ultramafic rocks plus an underlying flysch and/or mélangé and blueschist assembly. The gabbroic belt exposed along strike in the Valdez Quadrangle is described as consisting, in part, of a tectonic mélangé that includes a matrix of serpentine or sheared mylonitic gabbro (Burns, 1982). Measurements on these rocks indicate they are generally dense and magnetic, although there is a range in density possibly related to serpentinization. In the geophysical model shown in Figure 4.6, two possible explanations can link the geophysical data with the geology this cross-section proposes. A common geometry for the density, susceptibility,
Figure 4.7 Interpretation of cross-sections shown in Figure 1.4. Cross-section A-A’ is based on seismic-reflection data. Cross-section B-B’ shows interpreted structural relations between terranes, rocks that intrude them, and overlap assemblages. (Plafker et al., 1994)
and conductivity body favors the idea of mafic/ultramafic rocks within a serpentinized matrix of some sort. Another possibility is a composite model in which a somewhat thinner layer of dense and magnetic rocks is underlain, or intermixed with a flysch layer to give the high conductivities. Based on Occam’s principles the geometry proposed first, with a simpler single source, should be the preferred model absent other constraints. Further study is warranted including a well linked and coincident MT and seismic program (Saltus, 2003, written communication).
CHAPTER 5

CONCLUSIONS

The magnetotelluric sounding method relates time varying electric and magnetic fields and how they interact with the subsurface of the Earth. MT measurements along a transect in Kenai, Alaska are processed and inverted for a geo-electrical section. The resulting geometry obtained from the resistivity model is used to constrain the density and magnetic susceptibility model to better fit observed gravity and aeromagnetic data along the transect. Consistency in gravity, magnetic susceptibility, and resistivity results allows structural geologic interpretation.

Three magnetic and two electric components for six stations along the transect are measured. The ratio of orthogonal electric and magnetic fields gives the impedance tensor. Robust estimation of the impedance tensor for every station is made using two methods. Stations k001 and k002 are processed using the least-squares method. Stations k201, k202, k203, and k204 are processed using the remote reference method for middle and low band frequencies, and the least-squares method for high band frequencies. Data are rotated from the measuring coordinate system to the structural coordinate system whose x-axis is perpendicular and y-axis is parallel to geologic strike. The strike direction is determined based on gravity and magnetic data and analysis of MT impedance tensor using polar impedance plots, tipper, skew, and Groom Bailey decomposition. The rotated apparent resistivity, phase, and tipper from each station are used in inversion.
Apparent resistivity and phase for TE mode data, apparent resistivity and phase for TM mode data, and tipper at all six MT stations are used in 2D inversions to construct a geo-electric model. The inverted resistivity model clearly images the Border Ranges Fault and Cook Inlet sedimentary basin. In addition, the model provides structural information for deeper regions of the section. Density and magnetic susceptibility models based on the geometry from the MT inversion are able to reproduce respectively the observed gravity and magnetic data. Therefore, the geophysical model is consistent with three different data sets. Geologic inference is based on the geophysical model with guidance from previous geologic interpretations.

The Border Ranges Fault is imaged well by all three geophysical methods in the Kenai area. The geophysical model images the fault with a dip to the northwest. On the west side of the fault, there is a resistive region beneath the conductive Cook Inlet sedimentary basin. Beneath the resistive region is a highly conductive body approximately 12 km in height and 22 km in width, with the top of the body at 10 km depth. This conductive body has a high susceptibility based on magnetic modeling and it is the source of the magnetic anomaly high. MT study has therefore imaged the causes of one of the most prominent magnetic anomalies within the Southern Alaska magnetic high. High conductivity values follow the dip of the fault from this body to the shallow subsurface. Given that we assume the imaged conductive body is the source of aeromagnetic and gravity anomalies in the region, then this common source body is dense, magnetic, and conductive. One candidate lithology for such a body would be mafic/ultramafic rocks within a serpentinized matrix as observed in the appropriate geologic position along strike of the Border Ranges fault (Burns, 1982). This possibility is consistent with some regional seismic/geologic models for the area (Plafker, 1994).

Extending the MT survey line further west could define the western border of the conductive body with a top at 10 km. Additional MT survey lines north and south of the Kenai transect can show how far the conductive body extends and if it has depth variation. Are there conductive bodies throughout the entire Southern Alaska magnetic
high? Do similar structural features by faults near this extensive magnetic high provide a
guide to the tectonic setting of other potentially resourceful basins, including the Copper
River basin? Further study, including seismic data in the same area, is needed to know
what materials are causing the magnetic high.
REFERNCES CITED


