Natural and Controlled Source Magnetotelluric Data Processing and Modeling

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In this thesis, four studies using different geophysical electromagnetic methods are presented. In the first study dealing with airborne measurements, the noise response due to the rotation of the aircraft and the aircraft itself as a metallic conductive body on the Earth’s electromagnetic response in very low frequency and low frequency band was investigated. The magnetic fields are independent of the aircraft in the VLF band and part of the LF band. But at higher frequencies (above 100 kHz), the signals are more influenced by the aircraft. The aircraft also generates its own noise frequencies which are mixed with the radio transmitter signals. The second and third studies are applications of radio-, controlled source-magnetotellurics and electrical resistivity tomography methods at a quick-clay landslide site in southwest Sweden. The data are processed and modeled in 2D and 3D, and the models are compared with high-resolution seismic and geotechnical data. The obtained results were further validated and refined by performing synthetic tests in the second study. The third study shows that the 3D models provide larger and more continuous volume of the quick clay structure than traditional 2D models. Both studies have shown that integrated application of geophysical methods for landslides is ideal. Quick clays often overlie the coarse-grained layers showing an increase of resistivity values in the models. In the fourth study, a new audio magnetotelluric data acquisition technique is developed and is named moving magnetotellurics (MMT). In this new technique, the magnetic sensors are placed on the ground and only 15 to 20 minutes data are acquired for each station, which usually is enough to cover the frequency range 30-300 Hz. The new technique is more efficient and convenient than the traditional magnetotelluric method, and test measurements have shown that it is an applicable method in shallow depth studies.

Keywords: Geophysics, Airborne Electromagnetic Method, Radio Magnetotellurics, Controlled Source Magnetotellurics, Electrical Resistivity Tomography, Moving Magnetotellurics, 2D inversion, 3D inversion
Dedicated to my parents
单士林 魏淑华
List of Papers

This thesis is based on the following papers, which are referred to in the text by their Roman numerals.


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Additional conference proceedings written during my PhD studies, but not included in the thesis are:


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Abbreviations

2D          Two-dimensional
3D          Three-dimensional
MT          Magnetotellurics
AMT         Audio Magnetotellurics
RMT         Radio Magnetotellurics
MMT         Moving Magnetotellurics
VLF         Very Low Frequency
LF          Low Frequency
TE          Transverse Electric Mode
TM          Transverse Magnetic Mode
ERT         Electric Resistivity Tomography
CPT         Cone Penetration Test
CPTU-R      Cone Penetration Tests With Resistivity
RMS         Root Mean Square
S/N         Signal-to-noise
SGU         Geological Survey of Sweden
1 Introduction

Geophysical methods measure the electromagnetic, electrical, magnetic, gravitational or seismic source response of the subsurface and thus provide information about the physical properties of the Earth's subsurface. Geophysical measurements that are carried out on the surface or in the air can be performed quickly to gain subsurface information. Electromagnetic geophysical methods measure the variations of electrical and magnetic fields at the Earth's surface or in the air. These fields are processed to yield a subsurface resistivity image for further interpretation. Electromagnetic methods have been widely applied in mineral exploration, geological mapping, archaeology, waste disposal monitoring, hydrocarbon exploration and so on.

The airborne Very Low Frequency (VLF) technique, which covers the frequency range of 14-30 kHz has been widely used for large scale geological mapping. The successful application of ground radio magnetotellurics (RMT, 14-250 kHz) has shown that the multifrequency concept is superior to the traditional VLF technique. Thus the idea of using Low Frequency (LF) band (30-300 kHz) as supplementary frequencies in the airborne VLF technique was proposed to increase the depth resolution and the possibility of using the LF frequencies was studied in Paper I. This paper studied the noise generated due to the rotation of the aircraft and the aircraft itself as a metallic conductive body on both VLF and LF bands. The study was carried out by three different methods: 3D wave polarization, determination of transmitter direction and full tipper estimation. The results show that the magnetic field is independent on the aircraft at low frequencies in the VLF and part of the LF bands (below 100 kHz). At high frequencies (above 100 kHz), the signals are more influenced by the aircraft and the wave polarization directions are more scattered as it can be seen when the aircraft turns.

However, single geophysical dataset is often difficult to interpret because one dataset cannot give specific subsurface conditions. Furthermore, the anomalies indicated by the data could be caused by numerous sources. As a result, geophysical methods are often used in combination with other geophysical methods, geological records or other information. In Paper II and III, the electromagnetic methods, RMT, electric resistivity tomography (ERT) and controlled source tensor magnetotelluric (CSTMT) methods are interpreted jointly with reflection seismic and borehole data to study quick clays in a landslide area. In Paper II, the geometry and physical properties of geological structures at a quick-clay landslide site were studied using joint interpretation of RMT, ERT, reflection seismic and borehole data. Synthetic
tests using real field data parameters were also carried out and they confirmed the validity of the resistivity models from real data. The integration of ERT and RMT data with reflection seismic data is ideal for quick-clay landslide studies especially when the clay materials are thick. In Paper III, the RMT, CSTMT and ERT data allowed a three dimensional study and the results better delineate the geometry and physical properties of the quick-clay structures as studied in Paper II. The combined datasets CSTMT and RMT data (CSRMT) were inverted using a 3D inversion program to increase the depth of penetration by adding the CSTMT data. The RMT and ERT data were inverted individually and compared with the 3D CSRMT model. Different layers with different electrical properties can be observed from the models. The interpreted coarse-grained and bedrock horizons from the 3D migrated seismic data correspond well with structures observed in the 3D CSRMT model. The study has shown that a potential quick-clay layer at around 5-10 m depth extends in the 3D study area.

The audio magnetotelluric (AMT) method is usually applied for deep sounding. It requires long recording times and a lot of work in field setup. In Paper IV, a newly developed measurement system was proposed to make it easier, more efficient and convenient to carry out AMT field work. Hence, a new AMT data acquisition method is proposed which is named moving magnetotellurics (MMT). The three magnetometers are mounted on a frame and directed in x, y and z directions. The frame is placed on the ground during data acquisition, which makes it more convenient and efficient than the traditional MT or AMT measurements (e.g. saving time from digging and burying the sensors). Test measurements were carried out at several places in Sweden to study the noise influence on the method. The MMT method suffers from noise disturbances as the other MT methods do, but the noise level is much reduced in areas where the Earth is less resistive.

In this thesis, the fundamental concepts of the methods in my research work are introduced in Chapters 2, 3 and 4. In chapter 5, the four papers are summarized including the main research work and results in my Ph.D study.
The electromagnetic (EM) theory has been comprehensively introduced in many popular electromagnetism books (Landau and Lifshitz, 1960; Jackson, 1998; Griffiths, 1981; Greiner, Bromley and Soff, 1998). Concepts of EM wave polarization from these books have been utilized in paper I. Detailed discussions of the EM theory for applied geophysics can be found in Ward and Hohmann (1987) and Nabighian (1988, volume I and II). Applications of EM methods in common use have been described in Nabighian (1988, volume I and II). The EM methods that have been utilized in the thesis are described in the following paragraphs.

2.1 Electromagnetic waves

Electromagnetic waves are created by the vibration of electric charges. These vibrations create a wave which has both an electric and a magnetic component (Griffiths, 1981; Greiner, Bromley and Soff, 1998). The Scottish physicist James Clerk Maxwell (1831-1879) showed that electric and magnetic fields fluctuate together and form a propagating wave (Figure 2.1). The wave is a transverse wave, since the fields are perpendicular to the direction in which the wave travels. All electromagnetic waves travel through a vacuum at the same speed of light of $3\times10^8$ m/s (c). It is less than $3\times10^8$ m/s when the wave propagates through a material medium (http://www.physicsclassroom.com/mmedia/waves/em.cfm).

The EM wave which is the source field for many EM methods (such as MT, AMT, RMT, etc.) is assumed to be in the form of a plane wave. The plane wave is the primary EM field, reaches the earth's surface. Part of it is reflected back and part of it penetrates into the earth. The earth acts as a good conductor, induce electric currents, known as the telluric currents. The electric currents in turn produce secondary electric and magnetic fields. A combination of the primary and secondary fields is detected by the receiver.
2.2 Maxwell’s equations

The electromagnetic fields within a material can be described by Maxwell's equations. The following formulas show the differential form of Maxwell's equation in frequency domain (Ward and Hohmann, 1987):

\[
\nabla \times \mathbf{E} = -i\mu \omega \mathbf{H} \quad (2.1)
\]

\[
\nabla \times \mathbf{H} = \mathbf{J} + i\omega \mathbf{D} \quad (2.2)
\]

\[
\nabla \cdot \mathbf{D} = q \quad (2.3)
\]

\[
\nabla \cdot \mathbf{B} = 0 \quad (2.4)
\]

where \( \mathbf{E} \) (V/m) is the electric field intensity, \( \mathbf{H} \) (A/m) is the magnetic field intensity, \( \mathbf{B} \) (T) is the magnetic induction, \( \mathbf{J} \) (A/m\(^2\)) is the electric current density, \( \mathbf{D} \) (C/m\(^2\)) is the electric displacement current, \( \omega \) (Hz) is the angular frequency and \( q \) (C/m\(^3\)) is the electric charge density. Equation (2.1) is Faraday's Law saying that a magnetic flux that changes in time generates an electric field. Equation (2.2) is Ampere-Maxwell Law states that electric current and changing electric flux produces a magnetic field. Equation (2.3) is Gauss's Law for the \( \mathbf{E} \) field. It states that electric flux through a closed surface is proportional to the charge enclosed. Equation (2.4), Guass's Law for the \( \mathbf{B} \) field says that the total magnetic flux through a closed surface is zero.

Maxwell's equation can also be related through their constitutive relationship:

\[
\mathbf{J} = \sigma \mathbf{E} \quad (2.5)
\]

\[
\mathbf{D} = \varepsilon \mathbf{E} \quad (2.6)
\]

\[
\mathbf{B} = \mu \mathbf{H} \quad (2.7)
\]

\( \sigma \), \( \varepsilon \) and \( \mu \) describe intrinsic properties of the materials through which the electromagnetic wave propagate. The quantity \( \sigma \) (S/m) is the electrical conductivity, \( \varepsilon \) (F/m) is the dielectric permittivity and \( \mu \) (H/m) is the magnetic permeability.
Under quasi-stationary conditions, the displacement current \((i\omega D\) in equation 2.2) is neglected. In the case of a homogeneous structure and for a plane wave propagating in the positive \(z\)-direction, the components of the electric and magnetic fields can be written as

\[ A = A_0 \cdot e^{i\omega t} \cdot e^{-i\alpha z} \cdot e^{-\alpha z} \]

(2.8)

with \(\alpha = \sqrt{\mu_0\sigma\omega/2}\) and \(A\) represents either the electric or magnetic field. \(A_0\) is the wave amplitude, the second and third factor is sinusoidal time and depth variations and the fourth is exponential decay. This decay can be quantified by the skin depth, \(\delta\), at which this term was decayed to 1/e (Vozoff, 1991):

\[ \delta = \sqrt{\frac{2}{\mu_0\sigma\omega}} \approx 503\sqrt{\rho T} \text{ (m)} \]

(2.9)

with \(T\) representing the period, \(\mu_0\) is the magnetic permeability of vacuum.

From the skin depth, we can characterize the investigation depth roughly. Although it has been defined for homogeneous media, its use can be extended to heterogeneous cases as well (Spies, 1989; Telford et al., 1990).

### 2.3 Plane wave transfer functions

The electric and magnetic fields have a linear relationship in Maxwell's equations. The horizontal electric fields \((E_x, E_y)\) and vertical magnetic field components \((H_z)\) are related to horizontal magnetic field components \((H_x, H_y)\) assuming plane wave conditions and a fixed frequency (Cantwell, 1948),

\[
\begin{bmatrix}
E_x \\
E_y
\end{bmatrix} = \mathbf{Z} \begin{bmatrix}
H_x \\
H_y
\end{bmatrix}
\]

(2.10)

and

\[ H_z = T^T \begin{bmatrix}
H_x \\
H_y
\end{bmatrix} \]

(2.11)

where \(\mathbf{Z}\) and \(\mathbf{T}\) are known as plane wave transfer functions. \(\mathbf{Z}\) is the complex impedance tensor given by

\[
\mathbf{Z} = \begin{bmatrix}
Z_{xx} & Z_{xy} \\
Z_{yx} & Z_{yy}
\end{bmatrix}
\]

(2.12)

whereas \(\mathbf{T}\) is the complex vertical magnetic transfer function or tipper, defined by \(\mathbf{T} = (A, B)^T\). Superscript \(T\) means transposition, \(A\) and \(B\) are the complex components of the tipper tensor.

Induction arrows which are formed by the real parts of the tipper tensor are vector representations of the complex ratios of vertical to horizontal magnetic field components and can be used to infer the presence of lateral variations in conductivity, because vertical magnetic fields are generated by lateral conductivity gradients (Simpson and Bahr, 2005).
In a two dimensional case with structures striking in the $x$ direction, the
diagonal elements are zero, $Z_{xy}$ represents the transverse electric (TE) mode
(currents flowing in a plane parallel to the strike), and $Z_{yx}$ represents the
transverse magnetic (TM) mode (current flowing in a plane perpendicular to
the strike). The determinant of impedance tensor defined as
\[ Z_{\text{det}} = \sqrt{Z_{xx}Z_{yy} - Z_{xy}Z_{yx}} \] (2.13)
is rotationally invariant and can be used for modeling the resistivity. Peder-
sen and Engels (2005) showed that 2D inversion of the determinant data is
more robust than the inversion of TE and TM modes in a 3D environment.
Most of the routine 1D and 2D inversion codes usually use apparent resistiv-
ity ($\rho_a$) and phase ($\phi$) data at different frequencies ($\omega$). They are given as:
\[ \rho_a(\omega) = \frac{1}{\omega \mu_0} |Z(\omega)|^2 \] (2.14)
\[ \phi(\omega) = \arctan \frac{\text{Im}(Z(\omega))}{\text{Re}(Z(\omega))} \] (2.15)
assuming $\mu = \mu_0$.

2.4 Wave polarization

Wave polarization generally means wave orientation. The polarization direc-
tion is along the direction of the electric field by convention. Wave polarization
usually occurs for vector fields, i.e. the electric and magnetic fields. The
electromagnetic waves are described as transverse waves, meaning that a
plane wave's electric field vector $E$ and magnetic field $H$ are in directions
perpendicular to (or "transverse" to) the direction of wave propagation (in-
tuctor.physics.lsa.umich.edu).

The EM wave which is the source field for the MT field is assumed to be
in the form of a plane wave, either linearly polarized or elliptical polarized
and propagating vertically downward to the earth (Cantwell, 1948; Torres-
Verdin and Bostick, 1992). This assumption depends on the experimental
fact that these fields are uniform over wide areas.

Pedersen (1982) studied the magnetic wave polarization in two dimen-
sions. He defined a quasi-monochromatic partially polarized signal as a
summation of a polarized portion and an unpolarized portion and estimated
the degree of polarization as the ratio of the intensity of the polarized part to
the total part of the signal. In paper I in this thesis, three dimensional wave
polarization is studied to distinguish radio transmitter frequencies from
background EM signal in Very Low Frequency (3 to 30 kHz) and Low Fre-
quency (30-300 kHz) bands. Generally, the EM waves produced by the radio
transmitters have higher energy than the background waves and are polar-
ized. The polarization analysis is based on the polarization theory by Mandel
and Wolf (1995) and Ellis et al. (2005).
Assume an electromagnetic field is characterized by a $3 \times 3$ correlation matrix (2.16):

$$
\mathbf{S} = \begin{bmatrix}
S_{11} & S_{12} & S_{13} \\
S_{21} & S_{22} & S_{23} \\
S_{31} & S_{32} & S_{33}
\end{bmatrix}
$$

(2.16)

where $S_{ij}$ are the cross- or power- spectra of the three magnetic field components, $H_x$, $H_y$ and $H_z$ and the subscripts 1, 2 and 3 denote the $x$, $y$ and $z$ components respectively. Since $\mathbf{S}$ is a nonnegative definite Hermitian matrix, it can be diagonalized by a unitary transformation and, moreover, its eigenvalues ($\lambda_1, \lambda_2, \lambda_3$, ordered as $\lambda_1 \geq \lambda_2 \geq \lambda_3 \geq 0$) are necessarily real and non-negative. The 3D degree of polarization is defined as the ratio of the averaged intensity of the polarized part to the total averaged intensity of the field at the point:

$$
P = \frac{\lambda_1 - \lambda_2}{\lambda_1 + \lambda_2 + \lambda_3}
$$

(2.17)

here $0 \leq P \leq 1$ with $P=0$ representing a completely unpolarized field and $P=1$ a completely polarized field at that point (Ellis et al., 2005). If the field is totally polarized, the biggest eigenvalue $\lambda_1$ is associated with the magnetic energy density and the corresponding eigenvector is associated with the equivalent magnetic field (Ellis and Dogariu, 2004).

### 2.5 Radio magnetotelluric method (RMT)

The Radio Magnetotelluric (RMT) method is an improvement of the VLF technique by using both the electric and magnetic fields from distant radio transmitters. These transmitters are vertical electric dipoles producing electromagnetic waves that can be classified as plane waves at remote observation points (Bastani, 2001; Pedersen et al., 2006). The RMT method operates in a frequency range of 14-250 kHz and can be used for near surface studies.

Turberg et al. (1994) first implemented the RMT method in hydrogeological studies. After that, this method has been widely used in environmental, geotechnical and archaeological studies (e.g. Tezkan et al., 1996; Tezkan et al., 2000; Bastani and Pedersen, 2001; Newman et al., 2003; Linde and Pedersen, 2004a; Linde and Pedersen, 2004b; Pedersen et al., 2005).

In the RMT method, five components of the electric and magnetic fields are measured simultaneously. The data processing technique is based on the natural source Magnetotelluric method. Any radio transmitter between 14-250 kHz can be clearly indentified, and this allows us to derive more information on the variations of the resistivity with depth following the famous expression (2.9). In the case of conducting sediments on top of resistive basement, the sounding depth may be reduced because electromagnetic waves lose more energy in conductive materials (Bastani, 2001).
The RMT data studied in this thesis was acquired with the EnviroMT system (Bastani, 2001). The system is frequency domain electromagnetic equipment designed for ground measurements. The apparent resistivity and phase were estimated in 9 frequency bands of 1 octave. As presented by Pedersen et al. (2006), the RMT data suffers most from the sparse or non-uniform distribution of radio transmitters and low signal to noise ratio. The signal to noise ratio is a function of the transmitter strength, distance to the transmitter and also the selected bandwidth. Signals for transmitters are selected by demanding that the signal to noise ratio is above a predefined threshold (Bastani and Pedersen, 2001).

Fundamental scalar transfer functions are estimated in the transmitter coordinate, the \( y \)-axis coincides with the horizontal magnetic field direction which is perpendicular to the wave propagation direction. The directions to radio transmitters are estimated following equation (2.18) using the two horizontal magnetic components. The reason is that the magnetic field is generally less distorted by lateral conductivity changes than the electric fields, so the estimated directions from the magnetic fields are more stable and unique. The direction or the azimuth is measured from the magnetic north towards the east.

\[
\theta_{AZ} = \frac{k\pi}{2} + 0.5\tan^{-1}\left[2\text{Re}(S_{12})/(S_{11} - S_{22})\right], \quad k=0,1. \quad (2.18)
\]

\( \theta_{AZ} \) is multi-valued corresponding to the minimum and maximum horizontal magnetic field powers. The angle of minimum power which gives the direction to the transmitter is found by checking the second derivative of \( S_{11} \) (power spectra of the magnetic field)

\[
2(S_{11} - S_{22})\cos(2\theta_{AZ}) - 4[\text{Real}(S_{12})\cos(2\theta_{AZ})], \quad (2.19)
\]

which is positive for the minimum and negative for the maximum power. The scalar transfer functions can be thus estimated after rotating the fields into the transmitter coordinate system.

\[
T_{ij} = S_{ij}/S_{ii} \quad (2.20)
\]

with \( i=1,2 \) (\( H_x, H_y \)); and \( j=3,4,5 \) (\( H_z, E_x \) and \( E_y \)).

2.6 Controlled source tensor magnetotellurics (CSTMT)

The original motivation of using controlled sources in the magnetotelluric (MT) method was to improve the signal strength problem that arises in the method (Goldstein and Strangway, 1975). The AMT signal from natural
sources in the frequency band 1Hz-10 kHz is mainly caused by distant lightning storms that propagate within the earth-ionosphere waveguide (Vozoff, 1987; Smith and Jenkins, 1998; Garcia and Jones, 2002). The properties of this waveguide cause diurnal, seasonal and solar-cycle fluctuations of the AMT fields. The temporal fluctuations cause significant signal amplitude attenuation especially at frequencies in the 1-5 kHz AMT dead band (Szarka, 1987; Qian and Pedersen, 1991; Chave and Jones, 2012). The AMT amplitude increases during summer months and nighttime. To overcome the noise problem, the Controlled Source Audio Magnetotelluric (CSAMT) method was proposed by Goldstein and Strangway (1975). The traditional practice is to set a transmitter and receiver distance at least four times larger than the skin depth (Sasaki et al., 1992; Wannamaker, 1997), so that a plane wave assumption can be made, and MT or AMT modeling can be used for interpreting the observations (Bartel and Jacobson, 1987). However, it may be difficult to define criteria for far-field conditions because the electrical property of the subsurface is unknown.

The Controlled Source Tensor Magnetotelluric (CSTMT) method was proposed by Li and Pedersen in 1991. They proved that impedance and a tipper tensor can be uniquely defined for any earth structure and any source and receiver position and that these tensors are independent of the orientation and strength of the electric or magnetic dipole sources. This allows using two sources with different orientations at the same position.

The CSTMT data used in this thesis was also collected by the EnviroMT system. Two modes can be employed for the measurements, RMT and CSTMT. In the CSTMT mode the source is composed of two perpendicular horizontal magnetic dipoles or electric dipoles remotely controlled from the receiver to operate in the frequency range of 1-100 kHz. In this data set, magnetic dipoles were used because they are much easier to install and their range is sufficient to cover distances up to about 1 km. They also have little coupling to nearby conductive structures compared to electric dipoles and they are therefore generally expected to provide better plane wave conditions and have smaller galvanic distortion effects than electric dipoles (Bastani, 2001).

A unique transfer function can be estimated for the selected transmitting frequency, because the transmitter consists of two independent horizontal dipole coils that are set-up approximately perpendicular to each other. Although the transfer functions are easily interpreted in the plane wave case, they can be still uniquely defined in the intermediate and near-field ranges as shown by Li and Pedersen (1991).

If the measurement is carried out in the near-field or transition zone, near-field effects are present in the estimated transfer functions. According to Li and Pedersen (1991), the near field effect from a magnetic dipole will cause the logarithmic apparent resistivities to decay with a slope of 45 degrees as a function of log frequency, and the phases approaching 90 degrees. The real part of the tipper tensor points toward the transmitter.
2.7 Electrical resistivity tomography (ERT)

The Electrical Resistivity Tomography (ERT) method is one of the geoelectric methods. In this method, the spatial variation of electrical resistivity $\rho$ is estimated using four electrodes. One pair of transmitting electrodes (electrodes A and B in figure 2.2) injects current ($I$) into the subsurface and the other pair (electrode M and N) measures the potential difference (voltage $V$), which then together permits the determination of apparent resistivity as:

$$\rho_a = K \frac{V}{I}$$  \hspace{1cm} (2.21)

where $K$ is a geometric factor that depends on the positions of the electrodes (Daily et al., 2005). The measured apparent resistivities, for example along a profile, are used to estimate the true resistivity of structures below the profile.

![Battery](Image)

*Figure 2.2. Four-electrode resistivity array. Solid lines show the injected current and dashed lines show the equipotentials. Figure from Carpenter et al. (2012).*

The application of ERT is especially recommended for exploring horizontal structures, however, its use brings quality results also with regard to vertical structures (Drahor et al., 2006). The maximal depth at which changes of the electric resistivity can be detected is given by approximately as one fifth to one eighth of the maximum spacing of current electrodes (Panek et al., 2008).

In this thesis, the commercial software RES2DINV and RES3DINV are used for ERT data processing. Information about the software can be found at http://www.geotomosoft.com/.
2.8 Moving magnetotelluric method (MMT)

The Moving Magnetotelluric (MMT) method is a newly developed EM data acquisition method proposed in this thesis. This method is based on traditional AMT or MT method. Five field components are measured (two electric and three magnetic). The two electric fields are collected by four electrodes forming two telluric dipoles. The three magnetic field components are measured by three induction coils. But instead of burying the electrodes and coils in the ground, in MMT measurements, standard lead-lead chloride electrodes which are optimal for measuring electric components at low frequencies are replaced by steel electrodes which can easily be inserted into the ground and the magnetometers are put into three orange cylinders (see Figure 2.3) which are firmly fixed onto a wooden frame to allow the three magnetometers have the same vibrations. The frame is placed on the ground for data acquisition. The mounted coils are oriented following the right hand rule with the z-axis positive downwards.

To cover the frequency range from 30-300 Hz, the data is only acquired for a maximum of 15-20 minutes. Normal MT processing techniques are used for the MMT data to yield frequency domain transfer functions. Man-made noise such as power lines or railways reduces the data quality to a high degree, but its influence can be reduced during data analysis. Due to the less human activity at night, MMT data quality can be improved by night time recordings. In areas where the earth is resistive and strongly influenced by man-made noise, strong near-field effects are seen in both apparent resistivity and phase data. The data quality is much higher when the measurement is carried out in a less resistive area because electromagnetic noise propagates to smaller distance within a conductive earth.

The new MMT method has several advantages compared to the standard AMT or MT method, including efficiency, reduced space and time demanding and lower weight to carry. It is an applicable method in shallow depth (2 or 3 km) studies, especially in areas where normal MT measurements are inconvenient and/or too expensive to carry out.
Figure 2.3. Moving MT setup. Three magnetic field sensors are mounted on the wooden frame and are orientated following right-hand rule.
3 Other techniques combined with electromagnetic methods.

In this thesis the RMT and ERT data are jointly interpreted with seismic reflection data and cone penetration tests (CPT). The RMT and ERT methods produce apparent resistivities as a function of frequency or electrode spacing. These apparent resistivities, and for the RMT method additionally the impedance phase are later inverted to yield a resistivity model that varies laterally as well as with depth. During the inversion process, smoothness constrains are usually imposed to constrain those parts of the model for which the data constraints are insufficient. For certain lithological interfaces the seismic reflection method may be suited to image the geometry of the subsurface and it is complementary to electromagnetic and resistivity methods to better resolve the subsurface structures. The CPT method can provide conductivity/resistivity information of the subsurface as a function of depth. It is a direct method that can help to improve the uncertainties in the interpretation of geophysical methods.

3.1 Seismic reflection method

The seismic reflection method is a method that uses the principles of seismology to estimate the velocity and geometry properties of the Earth's subsurface from reflected seismic waves. Dynamite, a specialized air gun or a seismic vibrator can be used as controlled seismic energy sources.

When a seismic wave travels between two materials with different acoustic impedances, some of the wave energy will be reflected from the interface and some will be refracted and travel through the interface. In the seismic reflection technique, first the seismic waves are generated and then the time taken for the waves to travel from the source are recorded by an array of detectors (or receivers, or geophones) at the surface (Sheriff and Geldart, 1995). Knowing the travel times from the source to receivers and the velocity of the seismic waves, the depth and trend of the reflecting surface(s) are calculated. Reflection horizons from a number of underlying reflecting layers can be detected.

Combinations of seismic reflection and electromagnetic methods have been widely applied in different near surface investigations, such as hydrogeologic and environmental investigation, landslide studies and gas hydrate
stability studies (Meyer and Fine, 1997; Bichler et al., 2004; Gehrmann, 2011). In this thesis, integrated interpretation of 2D and 3D modeling of ERT, RMT data and seismic reflection data is presented.

Figure 3.1. Seismic wave paths showing reflections from the top of bedrock to detectors (or geophones). (http://www.geologicresources.com/seismic_reflection_method.html.)

3.2 Cone penetration test

The cone penetration test (CPT) is a method used to determine the subsurface conditions (i.e. resistivity/conductivity) up to certain depths. It is a valuable method of assessing subsurface stratigraphy associated with soft materials or potentially liquefiable materials (silt, sand) and landslides.

CPTU-R data (cone penetration tests with resistivity measurements) are shown in this thesis since they can be directly connected to the results from RMT and ERT data. The CPTU-R data identifies layers of quick clay and coarse-grained materials. Geophysical models always contain uncertainties even when several methods are combined to make a joint interpretation (Donohue et al., 2012). Since the CPT data is more direct in showing the property and depth of different materials, it is helpful for the confirmation of geophysical interpretations.
Inversion is the process of estimating a model that represents the subsurface structures from measurements carried out on the Earth's surface (or in boreholes). Inversion relies heavily on forward modeling which is the heart of the inversion process. It is used for computing the sensitivity matrix and responses for calculating the misfit from the measured data. The finite difference approach is applied in the forward modeling of the inversion schemes used in this thesis. It is one of the main three approaches to solve Maxwell's equations numerically (see (2.1) and (2.2)). A discrete system $Ax=b$ is formed where $b$ contains the boundary values and $x$ represents the unknown electric or magnetic fields.

For a more complete description of the following inversion scheme, the reader is referred to Siripunvaraporn and Egbert (2000) and Siripunvaraporn et al. (2005).

Consider the earth as discretized into a series of $M$ constant resistivity blocks, $m=[m_1, m_2, ..., m_M]$, $N$ observed data $d=[d_1, d_2, ..., d_N]$. The fit of the theoretical model responses $F[m]$ to the observational data can be expressed as

$$X_d^2 = (d - F[m])^T C_d^{-1} (d - F[m])$$

where the $T$ represents matrix transpose, and $C_d^{-1}$ is the inverse of the data covariance matrix.

Consider a general model norm to be

$$X_m^2 = (m - m_0)^T C_m^{-1} (m - m_0)$$

where $m_0$ is a initial (or a priori) model, $C_m^{-1}$ is an inverse of a model covariance matrix which characterizes the expected magnitude and smoothness of resistivity variations relative to $m_0$. Usually a Lagrange multiplier $\lambda$ is introduced to solve minimization problem, resulting this objective functional $U(m, \lambda)$:

$$U(m, \lambda) = (m - m_0)^T C_m^{-1} (m - m_0) + \lambda^{-1} \{(d - F[m])^T C_d^{-1} (d - F[m]) - X_e^2\}$$

where $\lambda$ acts to 'trade off' between minimizing the norm of data misfit and the norm of the model (Tikhonov and Arsenin, 1977; Parker, 1994).

The minimum structure solution is found by minimizing $X_m^2$ subject to $X_d^2 = X_e^2$, where $X_e^2$ is the desired level of misfit (Constable et al., 1987). The inversion process divides the subsurface into a number of small rectangles (2D) or rectangular prisms (3D), and attempts to determine the resistivity values of the rectangles or prisms so as to minimize the difference be-
tween the calculated and observed data. The difference can be measured by the misfit $X^2_d$, a critical aspect in inversion, or the root-mean-squared (RMS) error, with $\text{RMS} = \sqrt{\frac{\sum X^2_d}{N}}$.

A priori information which is independent of the observations, such as information from geology or other studies can be used as input together with observed data. They both can help to limit the non-uniqueness of the recovered model. The prior knowledge acts as constraints upon the optimization problem. A priori information is used in Paper IV.

In the joint inversion of RMT and ERT, individual weights need to be assigned to different data sets to avoid domination by one data set which has the highest number of data points. In Paper II, the weights for different data sets are manually tested to make sure that the data of the different methods fitted equally well.

### 4.1 2D and 3D inversion of ERT data

The inversion routines used for 2D and 3D inversions in this thesis are Res2DINV and RES3DINV (http://www.geotomosoft.com/downloads.php). The programs are based on the smoothness-constrained least-squares method (deGroot-Hedlin and Constable, 1990; Sasaki, 1992; Loke et al., 2003). The observed data contains the coordinates for each electrode array and the corresponding apparent resistivity. The coordinates include the positions in $x$ and $z$ directions for 2D inversion, and in $x$, $y$ and $z$ directions for 3D inversion. Topography information can be included into the input data file for inversions with topography. Boundary information can also be included for inversion models with sharp boundaries. Usually the model with sharp boundaries shows a more distinct boundary between two layers. These two programs use the Gauss-Newton method that recalculates the Jacobian matrix of partial derivatives after each iteration (Loke and Dahlin, 2002).

### 4.2 2D and 3D inversion of EM data

The inversion code EMILIA (ElectroMagnetic Inversion with Least Intricate Algorithms, see Kalscheuer et al., 2010), which is an extended version of the REBOCC program (Siripunvaraporn and Egbert, 2000) is used for 2D single and joint inversions of the RMT and ERT data. The forward and sensitivity calculation for RMT data is based on REBOCC (Reduced data space OCCAM approaches). The 3D inversion code WSINV3DMT (Siripunvaraporn et al., 2005) is used for studying the 3D RMT and CSRMT data. It is based on a 3D data-space Occam inversion scheme and closely follows the 2D inversion by Siripunvaraporn and Egbert (2000).
In 2D MT inversions, usually the apparent resistivity and phase data calculated from the off diagonal elements of the impedance tensor $Z$ are inverted. However, for the 3D case, the diagonal terms ($Z_{xx}$ and $Z_{yy}$) can also become significant, and are also included in the inversion. In this program, the full impedance tensors with both real and imaginary parts are inverted.

Because of the non-linearity of the MT inverse problem, an iterative approach is required:

$$ F[m_{k+1}] = F[m_k + \Delta m] = F[m_k] + J_k (m_{k+1} - m_k) $$  \hspace{1cm} (4.4)

Here the subscript $k$ denotes iteration number, and $J_k$ is the $N \times M$ sensitivity matrix.

Siripunvaraporn and Egbert (2000) transformed the inverse problem from the model space into the data space, by expressing the solution as a linear combination of rows of the sensitivity matrix smoothed by the model covariance:

$$ m_{k+1} - m_0 = C_m^T \left[ \lambda C_d + J_k C_m \right] \cdot \left[ d - F[m_k] + J_k (m_{k+1} - m_k) \right] $$  \hspace{1cm} (4.5)

This transformation reduces the size of the system of equations to be solved from $M \times M$ to $N \times N$. Since the number of model parameters $M$ usually is much larger than the number of data points $N$, a significant decrease in both CPU time and memory can be achieved with this approach (Siripunvaraporn and Egbert, 2000).

### 4.3 Model assessment

In geophysical inverse problems, usually the number of model parameters ($M$) is much bigger than the number of observation data points ($N$) and usually there will be more than one best-fitting model. There are almost always many possible answers to an inverse problem which cannot be distinguished by the available observations (Scales et al., 2001).

A way to find the best answer (bested fitting model) is the measure of RMS. The inversion models chosen to show in the thesis are all with minimum RMS among several iterations. However, in some situations the minimum RMS is reached but not ideal compared to the desired RMS defined in the inversion parameters. In this case, usually the data-fit is checked. The data-fit is the difference between original and predicted data. If there are poor fits at specific frequencies and stations, the data points are excluded from the inversion process. The RMS is usually reduced after removing poorly fitting data points.

The earth is three dimensional with irregular subsurface structures. In geophysics, it is common to use a two dimensional model to explain the real three dimensional earth. Sometimes problems arise because the data are contaminated by 3D effects. One of the solutions is to acquire data in a 3D pattern and carry out 3D analysis. This is done in Paper III. Another solution is
to invert the determinant data (mentioned in Chapter 2.3), which is often more robust against 3D effects than the TE-mode or TM-mode impedances.

Due to the smoothness and regularization effects in the inversion, it is difficult to determine how well the structures are recovered by the inversion models. Synthetic studies are a way to test the resolution of the model from real data. In paper II, a synthetic study was carried out. The synthetic data are inverted and compared with the model from real data for verification.
5 Summary of papers

5.1 Interference effects of aircraft on the Earth's electromagnetic response at VLF and LF frequencies

The airborne very low frequency (VLF) technique has been widely and successfully applied in mineral and groundwater explorations in the last few decades. Noise effects due to the rotation of the aircraft and the aircraft itself as a metallic conductive body on the Earth's electromagnetic response were studied in the Very Low Frequency (VLF) and Low Frequency (LF) bands. A large number of radio transmitters with frequencies in these bands can be used for geophysical prospecting. The electromagnetic fields produced by these transmitters have much higher amplitudes than the background fields. Some potential radio transmitter frequencies in the VLF and LF bands were chosen from the power spectra plots of the measured magnetic fields. The effects from the aircraft on the Earth's electromagnetic fields were studied by three different methods:

1. The 3D wave polarization method, which is based on the polarization theory by Mandel and Wolf (1995), using the eigenvalues and eigenvectors to study the polarization status of the electromagnetic fields. All the chosen signal have rather high degrees of polarization (estimated from the eigenvalues). From the eigenvectors the wave polarization direction can be estimated and some 'dummy' (signal produced by the aircraft) radio transmitter frequencies are detected (e.g. 101.7 kHz and 203.4 kHz) (Figure 5.1). At some frequencies the estimated polarization directions are more scattered, which indicates that the aircraft itself adds a secondary field to the measured total magnetic fields (171 kHz, 177 kHz and 225 kHz).

2. An algorithm for automatic determination of directions to radio transmitters is given by Bastani (2001). Two directions of minimum and maximum horizontal magnetic field powers are calculated corresponding to the radio transmitter direction and wave polarization direction. Scalar transfer functions are calculated in the transmitter coordinate system. The estimated scalar transfer functions (both real and imaginary parts) have strong sign dependence on the flight directions for 171, 177 and 225 kHz. This is caused by the aircraft producing an additional vertical (and horizontal) field to the primary field and the secondary field from the earth, whereby the behavior of the magnetic transfer functions is changed.
3. Full tipper estimation. Using at least two transmitter frequencies, the
full tippers are estimated by assuming that the anomalies do not vary much
with frequency as long as the frequency band is limited to one octave. Usu-
ally the magnetic transfer functions distribute around zero, which is the case
for the transfer functions estimated from low frequencies pairs. But for the
higher frequency pair (e.g. 177&225 kHz) there is a bias of around 10%
which appears in both the real and imaginary parts of the transfer functions
(Figure 5.2), again proving that the aircraft itself generates secondary fields
which are added to the earth's transfer functions.

The results from the three different methods have confirmed that the elec-
tromagnetic responses are strongly perturbed by the aircraft for frequencies
above 100 kHz.

Figure 5.1. Wave polarization direction. The equivalent magnetic field is
separated into real and imaginary parts and plotted as arrows. Red arrows
represent data collected from southwest to northeast and blue arrows repre-
sent the opposite direction. Green arrows correspond to turning of the air-
craft.
Figure 5.2. Example magnetic transfer functions estimated from two frequency pairs. Real and imaginary part of $T_{31}$ and $T_{32}$ from the frequency pair 177 & 225 kHz. $T_{31}$ and $T_{32}$ are the magnetic transfer functions estimated following equation (2.20).

5.2 Integrated 2D modeling and interpretation of geophysical and geotechnical data to image quick-clays at a landslide site in southwest Sweden

RMT, ERT and high-resolution reflection seismic data were collected along four lines at a landslide site in southwest Sweden. The aim of this study is to image the geometry and physical properties of geological structures (Figure 5.3). Landslides are harmful natural disasters occurring frequently all over the world. They cause significant damages to infrastructures, natural resources and take thousands of human lives every year. Geotechnical data collected in this area suggests presence of quick clay above coarse-grained layers.

Geophysical methods have been widely applied for landslide studies. From the RMT and ERT data we can get depth-resistivity models so that the resistivity information from the models can be related to the electrical properties of quick clays or coarse-grained layers. Berger (1980) and Solberg (2007) classified the resistivities for different materials in landslides sites (Table 1).

The RMT and ERT data along four lines in this study were inverted singly and jointly with the inversion code EMILIA (Kalscheuer et al., 2010). The ERT data was also inverted using the commercial code RES2DINV.
From the models, we can interpret that the area generally has five different layers, the first layer is the dry crust with a resistivity higher than the underlying layer (see Figure 5.4). The second layer has a resistivity of 10 to 80 Ωm and overlays a layer of coarse-grained sediments with a resistivity of 100-200 Ωm. The most resistive feature in all of the models that deepens towards the river (see Figure 5.3) represents the crystalline bedrock (granitic type). The bedrock is overlain by more conductive overburden with internal layering (mainly four layers) with varying thicknesses and properties.

Due to the regularization and smoothing in the inversion scheme, the geological units having sharp boundaries with rapid changes in the physical properties cannot be clearly modeled. The interfaces of different structures shown in the models mostly do not reveal the true depth of the interface. In this study, another geophysical data see from reflection seismics, is used to better image the sharp boundaries and also be able to image deeper structures.

Model validation by synthetic simulations was also carried out. This enables us to study the limitations of the collected ERT and RMT data to resolve the coarse-grained layer at the depths imaged by the reflection seismics and detected by the geotechnical data. It also helps to show the effects on the resolution of the geophysical models (mainly 2D resistivity models). A synthetic 2D model was built from the 2D inversion of real data along one line. Both synthetic RMT and ERT data are generated and inverted singly and jointly. Sharp boundaries are also included in the inversion of ERT data using RES2DINV. The synthetic results show that the joint ERT and RMT inversion are superior in resolving different layers, both in depth and resistivity. They also confirm the validity of the resistivity models from the real field data.

This study has shown that the integrated application of geophysical methods for landslides is ideal. Together with geotechnical data, a potential quick clay deposit area has been identified. The quick clays often overlay the coarse-grained layers showing an increase of resistivity values with depth in the models.
Figure 5.3. Surface geology of the study area. The map was provided by the Geological Survey of Sweden. The black lines show the geophysical profiles presented in this study. The black dots show the location of geotechnical boreholes. The black rectangle shows the area of 3D survey. Results from the 3D data are presented in paper III and will be summarized in Chapter 5.3.

Table 1. Resistivity classifications of different soils in an quick-clay area from Solberg (2007) and Berger(1980).

<table>
<thead>
<tr>
<th>Classification of soil material from resistivity (Berger,1980)</th>
<th>Resistivity interval</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil material</td>
<td></td>
</tr>
<tr>
<td>Clay, salt</td>
<td>1-20 Ωm</td>
</tr>
<tr>
<td>Clay, leached</td>
<td>20-90 Ωm</td>
</tr>
<tr>
<td>Clay, dry crust</td>
<td>70-300 Ωm</td>
</tr>
<tr>
<td>Silt, wet</td>
<td>50-200 Ωm</td>
</tr>
<tr>
<td>Sand, saturated</td>
<td>200-1000 Ωm</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Classification of soil material from resistivity (Solberg, 2007)</th>
<th>Resistivity interval</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil material</td>
<td></td>
</tr>
<tr>
<td>Salt/intact marine clay</td>
<td>1-10 Ωm</td>
</tr>
<tr>
<td>Leached, possible quick clay</td>
<td>10-80 Ωm</td>
</tr>
<tr>
<td>Dry crust clay, slide deposits, coarser</td>
<td>&gt;80 Ωm</td>
</tr>
</tbody>
</table>
Figure 5.4. Results from 2D inversion of RMT and ERT data along line 5. The interpreted reflectors from reflection seismic data are marked with red lines; (a) inversion of determinant RMT data with a final RMS 1.33, (b) joint inversion of RMT and ERT data with final RMSs 1.00 and 1.46 for RMT and ERT data and (c) seismic section. The black arrows at the ground surface indicate the locations of geotechnical boreholes along this line. Reflection B1 is interpreted to be from the bedrock, S1 and S2 from two sets of coarse-grained materials (Malehmir et al., 2013).
5.3 Integration of controlled source and radio-magnetotellurics, electrical resistivity tomography and reflection seismics to delineate 3D structures of a quick-clay landslide site in southwest Sweden

The RMT, CSRMT (combined method of RMT and CSTMT) and ERT data were acquired for three dimensional analysis over a quick-clay landslide site in south-west Sweden (the rectangular area shown in Figure 5.3). Before this study, two dimensional RMT and ERT data have been studied along four lines in the same study area (paper II). In some parts of the study area, the 2D assumption might not be valid due to the discontinued structures at the earth surface, for example, the landslide scar distributed in the northeast of the studying area (see Figure 5.3) is circular shaped and thus implies 3D geometry. This is part of the reason of acquiring data in a 3D pattern and carrying out 3D investigation. The results will provide more detailed and clearer structures with different electrical properties and geometry.

The RMT and ERT data were measured along 11 lines running from south to north. The line spacing is 20 m. The RMT data has a station spacing of 10 m and the ERT data has a minimum dipole length of 5 m. The CSRMT data was only acquired along every second profile with a station spacing of 40 m.

In the CSRMT data, a remotely controlled source (a double magnetic loop) was used so that it was possible to measure smaller frequencies (compared to RMT frequencies) from 2 kHz up to 10 kHz and furthermore increase the penetration depth of the electromagnetic waves. Near-field effects can be seen in the controlled source data when the source is located closer to the receivers, usually less than five times the penetration depth (Li and Pedersen, 1991). The apparent resistivity drops below plane wave response and the apparent phase rises above plane wave response for a magnetic source. The observed apparent resistivity and phase from the CSRMT data luckily was not visibly disturbed by the magnetic source and it was possible to include the controlled source frequencies into inversion assuming plane wave conditions.

The 3D RMT and CSRMT data were inverted using the WSINV3DMT program (Siripunvaraporn et al., 2005), which is a full 3D inversion program for MT data. The WSINV3DMT program uses all the elements of the impedance tensor $Z$, including both real and imaginary parts. The 3D ERT data was inverted by RES3DINV which automatically determines a 3D resistivity model for the subsurface (Loke et al., 2003).

In paper II, the limited penetration depth of all EM methods due to the conductive overburden was discussed. The same behavior can also be seen in the 3D RMT and ERT models (Figure 5.5). However, the CSRMT model shows a clearer image of the resistive bedrock both in geometry and resistivity value (Figure 5.5 (c)).
The 3D inversion resistivity model of CSRMT data is shown in Figure 5.6. The model generally shows the same structures as the 2D resistivity models interpreted by Shan et al. (2014) where we give a detailed interpretation of the resistivity structures based on the resistivity classification of soils by Solberg et al. (2008) and Berger (1980). The inversion (Figure 5.6) very well recovered the dry crust and landslide scars in the northeast corner, which correlates well with the geological observations on the earth surface. It also shows a thick marine clay layer with resistivities of 10 $\Omega$m. The bottom layer with rising resistivity values has an advantage of giving general information about the shape of the bedrock. The top surface of the bedrock is much shallower in the south and deepens a lot towards the river in the north. The bedrock also uplifts in the northeastern corner close to the boundary of the 3D area where there is small river to the east and the surface elevation is higher.

Comparisons between the CSRMT data and 3D migrated seismic data show that the CSRMT data recover the bedrock very well and the top surface of the bedrock correlates well with the seismic horizons. However, the layer on top of the coarse-grained layer is not well pictured in the CSRMT data because of lack of data points. It is better resolved by the RMT and ERT models. The three resistivity models generally picture the same second layer above the coarse-grained layer and this layer has been proved to be a possible quick clay deposit by the CPT-R data.

The 3D data allow a three dimensional investigation for landslide risk assessment. They indicate a larger and more continuous volume of the quick clay structure than the traditional 2D data which may be due to smoothing in 2D inversion process.
Figure 5.5. 2D resistivity sections from different 3D models. (a) ERT, RMS 2.00; (b) RMT, RMS 1.45 and (c) CSRMT, RMS 1.263.
5.4 Moving magnetotellurics - experimental setup and case studies from Sweden

In traditional MT or AMT field measurements five components of electric and magnetic fields are measured. The three magnetic field components are measured by three magnetic sensors that are buried into the ground. In this study we developed a new way to measure the three components of the magnetic fields by placing the three magnetic sensors above the earth surface. These sensors are fixed in a rigid wooden frame and are directed according to the right hand rule. The electric fields are measured by steel electrodes with short dipole lengths of 20 m. See Figure 2.3 for the field set up. The frame and electrodes are moved from station to station so that a profile can be covered. We name the new data acquisition technique as Moving Magnetotellurics (MMT).

The idea of MMT was not the initial idea in this study. At the beginning the idea was to measure three component magnetic fields on board an aircraft or on a moving device in a frequency band from 30 Hz to 1000 Hz. A rigid frame was built for this purpose (the yellow frame in Figure 2.3) and
was put on a trailer behind a car. It was also placed on board of an airplane operated by SGU. Several test measurements were carried out while moving the car or airplane at a high speed. All the data has confirmed that strong electronic noise from the engine was generated in the measured magnetic fields when the data acquisition is in a moving mode.

The stability of the measured signal is also tested by putting the three magnetic sensors in parallel. The frame was placed on a platform and data were acquired while dragging the platform on ice at a walking speed. Since the three magnetometers are parallel they should give the same response. If they do not, the explanation must be sought in their small, but significant rotations with respect to each other, which generate uncorrelated signals (Pedersen, 1982). The data were measured both when the frame was moving and being still. We estimated the coherences of the fields between each two channels and concluded that small rotational movement can cause strong and independent vibrations in each sensor so that the measured fields are contaminated with strong noise.

Finally, the MMT idea came up and test measurements were carried out at several places in Sweden. At sites where the artificial electromagnetic noise was strong and the subsurface had high resistivity, strong near-field effects were observed in the phase data as they were distributed around zero. A comparison of night time recording and day time recording data at one of these sites showed that the manmade noise in the night time recording data is reduced (Figure 5.7).

Another test site located rather remotely was chosen in Gotland, Sweden with conductive mudstones overlying crystalline basement (Pedersen et al., 2005; Linde and Pedersen, 2004a). The estimated apparent resistivities and phases have high quality in the frequency range of 10 to 300 Hz. These data were inverted in 2D. The data fit was checked between the response and observed data, examples from several stations are shown in Figure 5.8. The resistivity model is compared with the RMT model from the same profile, the two models show similar resistive structures. Because of different frequencies used in the two methods, the RMT model shows a rather good resolution at shallow depth (above 150 m), while the MMT model has a good resolution of the conductive mudstones and the crystalline basement (Figure 5.9).

The MMT method suffers from noise disturbances as the other MT methods do. But it has a few superiorities over the MT or AMT methods, such as efficiency and convenience. The successful test measurement in Gotland implies that the MMT method is an applicable method in shallow depth studies, especially in areas where normal MT measurements are inconvenient and/or too expensive to carry out.
Figure 5.7. Comparison of apparent resistivity (top) and phase (bottom) from day time and night time recordings. Red curves are from the midnight recording and black curves are from the day time recording.
Figure 5.8. Data-fit of four example stations from the inversion of TM mode data. Under station numbers are the resistivity and phase, respectively. The phase has better fits than the resistivity.

Figure 5.9. 2D inversion model of Gotland data.
6 Discussion and conclusions

In this thesis, the results from geophysical field measurements with different electromagnetic methods are shown and discussed. The conclusions can be drawn for each paper.

In paper I, the interference effects of aircraft on the Earth's electromagnetic response at VLF and LF frequency bands are studied. Three different methods were used to study the noise generated by the rotation of the aircraft and the aircraft itself as a metallic body. In the VLF band and part of the LF band, the magnetic field is independent of the aircraft. At high frequencies (above 100 kHz), the signals are more influenced by the aircraft. Some aircraft generated noise which is mixed with the radio transmitter signals are detected as 'dummy' signals. The estimated transfer functions have dependence on the aircraft flight direction at high frequencies.

In paper II, RMT, ERT and high-resolution reflection seismic data were processed and modeled in 2D to delineate the geometry and physical properties of geological structures at a quick-clay landslide site in southwest Sweden. The geotechnical data collected in this area suggest the presence of quick clay above coarse-grained layers and these layers play a key role in the formation of quick clays and landslide triggering. The geometry and location of shallower structures resolved in the 2D resistivity models were compared with the geotechnical data. We observed that quick clays overlying the coarse-grained layer have higher electrical resistivity than the marine clays. The RMT and ERT inversion models correlated well with images from the reflection seismic data and observed geotechnical borehole data. The models are poor in resolving deeper resistive bedrock where the thickness of the conductive overburden exceeds a certain limit. However, information from the reflection seismic data could be used to estimate the depth to the top of the bedrock. The obtained results were further validated and refined by performing synthetic tests. This study shows that integration of ERT and RMT data with reflection seismic data is ideal for quick-clay landslide studies.

In paper III, we present 3D inversion results of RMT, ERT and CSRMT data collected at a quick clay site. These results are jointly interpreted with 3D seismic data and geotechnical data to image the potential quick clay deposit. The CSRMT data has a better penetration depth than the RMT and ERT data. It better images the bedrock underlying the marine conductive clay, but it is less reliable at resolving the top potential quick-clay and coarse-grained layers because of the lack of data coverage. The coarse-grained layer is rather flat with an average depth of 20 m. Above the coarse-
grained layer, there is a potential quick clay deposit with thickness around 5 m imaged in the resistivity models. Two boreholes have proven the presence of quick clay with thickness around 1-2 m on top of the coarse-grained layer. Due to the smoothing scheme used during the inversion process, the thickness determined in the resistivity models is uncertain and the value recovered in the inversion might be bigger than the real quick clay thickness. The potential quick clay layer almost extends over the entire studying area beneath the top dry crust. The 3D RMT, CSRMT and ERT models allow a three-dimensional investigation for landslide risk assessment. They provide a larger and more continuous volume of the quick clay structure than the traditional 2D data.

In paper IV, an idea of measuring three-component magnetic fields on board an aircraft or on a moving device in a frequency band from 30 Hz to 1000 Hz was tested. The three magnetometers were mounted on a frame. The frame was put on a trailer behind a car and on board an aircraft. In both cases, we observed strong engine noise. Moreover, uncorrelated signals were generated during movements even when moving the frame at a small walking speed. It was finally decided to measure the three magnetic field component when the frame is at rest on the ground. The new data acquisition technique is based on traditional Magnetotellurics (MT) but more efficient in several ways. We name it Moving Magnetotellurics (MMT). The MMT method suffers from noise disturbances as the other MT methods do. Man-made noise such as power lines or railways reduces the data quality to a high degree, but its influence can be reduced during data analysis. Generally man-made noise is reduced during night time and MMT data quality can be improved by night time recordings. In areas where the earth is resistive and strongly influenced by man-made noise, a strong near-field effect can be seen in both apparent resistivity and phase data. Data quality is much higher when measurements are carried out in areas with low resistivity because man-made electromagnetic noise propagates to greater distance within a resistive earth. The MMT method is an applicable method in shallow depth studies, especially in areas where normal MT measurements are inconvenient and/or too expensive to carry out.
Sammanfattning på svenska


Till skillnad från Artikel II, III och IV, undersöktes i artikel I bruset som genererades på grund av flygplanets rotation och flygplanet själv i form av en metallisk konduktiv kropp vid både VLF och LF (Låg Frekvens). Det finns ett stort antal radiosändare med frekvenser inom dessa band som kan användas vid geofysik prospektering. De elektromagnetiska fälten som dessa sändare avger har mycket högre amplitud än bakgrundsfälten och därför valdes några potentiella frekvenser inom VLF och LF från effektspektrumet av de uppmätta magnetiska fälten. Studien genomfördes med tre olika metoder; ”3D vågpolarisation”, ”determinering av transmitterriktning” och ”full tipper estimering”. Resultatet visar att magnefältet är oberoende av flygplanet vid låga frekvenser i VLF och delar av LF bandet (under 100 kHz). Vid höga frekvenser (över 100 kHz), påverkas signalen mer av flyg-
planet och vågpolarisationsriktningen är mer spridd vilket syns när flygplan-
et svänger.

I Artikel II studerades geometrin och fysiska egenskaper av geologiska
strukturer vid ett jordskredsområde med kvicklera genom gemensam tolk-
ning av radiomagnetotellurik (RMT) , elektrisk resistivitetstomografi (ERT),
reflektionssseismik och borrhålsdata. Syntetiska tester med verklig fältdata
genomfördes också och de bekräftade validiteten av resistivitetsmodellerna
från verklig data. Utifrån modellerna kan vi tolka att området generellt sett
har fem olika lager. Det översta lagret består av torrt sediment med en resis-
tivetit högre än det nästföljande lagret. Det andra lagret har en resistivitet på
10 till 80 $\Omega\text{m}$ och ligger ovanpå ett grovkornigt sedimentlager med en resis-
tivitet på 100-200 $\Omega\text{m}$. Det mest resistiva lagret i alla modeller går djupare
mot älven och representeras av kristallin berggrund (granit). Berggrunden är
täckt av mer konduktiva sedimentlager (oftast fyra lager) med varierande
tjocklek och egenskaper. Att kombinera flera geofysiska metoderna vid jord-
skred är mycket fördelaktigt. Tillsammans med geoteknisk data har potenti-
ella lager av kvicklera identifierats. Kvickleran återfinns ofta ovanpå de
grovkorniga lagren vilket visar sig med ökad resistivitet i modellerna.

I Artikel III studerar tredimensionell RMT, CSTMT (controlled source
tensor magnetotelluric) och ERT data för att bättre kunna avgränsa kvickle-
rans geometri och fysiska egenskaper som studerats i Artikel II. Datasetet
från CSRMT (controlled source radio magnetotelluric) (kombination av da-
taseten från CSTMT och RMT inverterades med ett 3D-inversionsschema
för att öka penetrationsdjupet. RMT och ERT data med fler mätpunkter in-
verterades separat och jämfördes med 3D CSRMT modellen. Olika lager
med olika elektriska egenskaper kunde observeras i modellen. CSRMT data
har bättre penetrationsdjup än RMT och ERT data. Det illustrerar bättre
berggrunden under den marina konduktiva laran, men avslöjar mindre om de
övre lagren av kvicklera och de grovkorniga lagren på grund av avsaknad av
mätpunkter. De grovkorniga lagren är relativt platta och håller ett genom-
snittligt djup of 20 meter. Ovanpå det grovkorniga lagret finns en potentiell
förekomst av kvicklera och en tjocklek på 5 meter vilket går att tolka från
resistivitetsmodellerna. 3D RMT, CSRMT och ERT data ger möjlighet till
e en tredimensionell undersökning för att bedöma risken för jordskred. Det ger
en större och mer kontinuerlig volym av kvicklerans strukturer än tradition-
nell 2D data.

I Artikel IV föreslås en ny AMT datasamling som kallas ”Moving
Magnetotellurics” (MMT). De tre magnetometrarna monteras på en ram och
riktas i x, y och z-riktning. Ramen placeras på marken när mätningen sker
vilket är mer bekvämt och effektivt än de traditionella MT- eller AMT-
mätningarna. För varje station räcker 15 till 20 minuter med datasamling
för att täcka frekvensintervallet 30 till 300 Hz. Testmätningar genomfördes
på flera ställen för att studera brusets inverkan på metoden. MMT-metoden
påverkas av brus precis som de andra MT-metoderna gör. Artificiellt brus så
som kraftledningar och järnvägar minskar kraftigt kvalitén på insamlad data,
men denna inverkan kan reduceras vid dataanalysen. Generellt reduceras artificiellt brus under natten och kvalitén på data från MMT kan ökas genom att mäta nattetid. I områden där jorden är resistiv och starkt påverkad av artificiellt brus kan stark närflätseffekt ses i både skenbar resistivitet och fasdata. Datakvalitén är mycket högre när mätningar genomförs i områden med låg resistivitet eftersom artificiellt elektromagnetiskt brus överförs längre i resistiv jord. MMT-metoden är en användbar metod vid studier av ytnära strukturer, speciellt i områden där normal MT-metod är svår och/eller för dyr att genomföra.
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