Uula Autio

DEVELOPMENT AND APPLICATION OF THE MAGNETOTELLURIC METHOD TO STUDY THE CRUSTAL STRUCTURE OF CENTRAL FINNISH LAPLAND
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Abstract

In this thesis, the magnetotelluric (MT) method has been applied to study the crustal geoelectric structure of central Finnish Lapland and new advances in MT data analysis have been realized. MT data acquired in 2014 display pronounced 3D effects, reflecting the complex tectonic history of the bedrock in the area. Particularly anomalous, so-called extreme data are observed in the northern part of the study area. A definition for extreme MT data is given by the condition where the determinant of the real and/or imaginary part of the impedance tensor becomes negative. Physically, such data are associated with reversal of, for example, the electric field as compared to its direction for a one-dimensional resistivity structure. The behaviour of the phase tensor and the determinant average of the impedance tensor, for instance, have been analysed in the case of extreme data. 3D conductivity models derived using the ModEM code display high crustal conductance (thousands of siemens) in the vicinity of the Central Lapland Greenstone Belt, the Peräpohja Belt and the Kuusamo Belt. A remarkable feature is an arc-shaped conductor inside the northern part of the Central Lapland Granitoid Complex, which continues into the Central Lapland Greenstone Belt in the north. The conductive structures in the models could represent deeply buried graphite and sulphide bearing metasedimentary rocks or reactivated Archaean shear zones. The conductors in the northern and southern parts of the study area are separated by a resistor coinciding with the Central Lapland Granitoid Complex. A possible explanation for the observed pervasive E–W principal direction of the phase tensor data in the study area could be the failed rift suggested in other studies.

In addition to the magnetotelluric studies in central Lapland, as a methodological development, an express approach where time-domain electromagnetic (TEM) data are transformed and subsequently used in 2D MT determinant inversion is presented. The methodology is applied in a simple geoelectric setting located in Greece but can in principle be extended to the more complex geological environments encountered in the Fennoscandian shield.

Keywords: crust, electrical conductivity, extreme data, Fennoscandian shield, Finnish Lapland, magnetotelluric
Autio, Uula, Magnetotelluurisen menetelmän kehitys ja soveltaminen maankuoren tutkimuksessa Lapin keskiosissa.
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Tiivistelmä
Lapin keskiosien tutkimusten lisäksi tutkielmassa on kehitetty tulkintatapa, jossa aika-alueen sähkönmagneettisen menetelmän (TEM) mittausaineisto muunnetaan käytettäväksi 2D MT käänteismallinnuksessa. Lähestymistapaa kokieltiin sähkökohtuvuusksiirtoesineissä yksinkertaisessa, Kreikassa sijaitevassa geologisessa ympäristössä, mutta se on periaatteessa laajennettavissa myös haastavampiin Fennoskandian kivil geologisiin ympäristöihin.

Asiasanat: Fennoskandia, Lappi, maankuori, magnetotelluurinen, sähkökohtavaus, aärivaste
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Rovaniemi, November 2021
Uula Autio
List of abbreviations

1D One-dimensional
2D Two-dimensional
3D Three-dimensional
BMT Broadband magnetotelluric
CLGB Central Lapland Greenstrone Belt
CLGC Central Lapland Granitoid Complex
CSRMT Controlled source radiomagnetotelluric
EM Electromagnetic
HSZ Hirvaskoski shear zone
KB Kuusamo Belt
LMT Long period magnetotelluric
MaSca "Magnetotellurics in the Scandes" project
MT Magnetotelluric
Ωm Ohm metre
PSZ Pajala shear zone
PB Peräpohja Belt
PC Pudasjärvi complex
S Siemens
TEM Transient electromagnetic
VSZ Venejoki shear zone
List of original publications

This thesis is based on the following articles, which are referred to in the text by their Roman numerals (I–III):


The papers included in this thesis are the result of work by various authors. The individual contributions of the author of this thesis are listed below.

I  Leading role in data acquisition, data analysis, theoretical considerations, modelling, main author of the manuscript.

II Data analysis, inversion, interpretation, main author of the manuscript.

III Methodological development, data analysis, modelling, inversion, main author of the manuscript.
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1 Introduction

1.1 Background and structure of the thesis

The magnetotelluric (MT) method is a passive geophysical method based on the measurement of the natural time-varying electromagnetic (EM) field on the Earth's surface. MT images the electrical conductivity structure of the subsurface, and as a deep probing method, it sheds light on the structure of the Earth’s crust and upper mantle. Due to its high scalability and advanced modern 2D and 3D interpretation methods, MT has also found many other uses beyond large-scale lithospheric studies. For example, geothermal exploration has long been an important application area for MT (Munoz 2014). MT is also becoming increasingly important in mineral exploration as deeper targets are pursued and regional control for ore bodies is sought (Heinson et al. 2018, Nurmi & Rasilainen 2015).

In the Fennoscandian shield, MT and other natural source EM methods have a long history in lithospheric studies (for reviews, see Hjelt 1987, Korja & Hjelt 1993, Korja 1997, Korja et al. 2002, Korja 2007). The conductance map compiled by Korja et al. (2002) illustrates that the shield has an elaborate crustal conductivity structure. Extreme lateral variations are observed due to conductive volcano-sedimentary belts separating resistive gneiss-granitoid blocks. The conductive belts, which MT is most sensitive to, often mark fossil tectonic boundaries (Korja & Hjelt 1993). The crustal conductors in the Fennoscandian shield are typically related to buried sedimentary rock layers (Korja 1997). In Finland, such well-conducting graphitic and sulphidic metasedimentary rocks have been identified in the near-surface throughout the country (Loukola-Ruskeeniemi 1992, Loukola-Ruskeeniemi et al. 2011). Crustal conductors can also be related to shear zones, where graphite can originate either from the surface (organic) or from deeper levels (inorganic) (Korja et al. 1996). MT can additionally be used to probe the deepest part of the lithosphere, i.e., the (electric) lithosphere-asthenosphere boundary, which is anomalously conductive, possibly due to partial melts or hydrogen diffusion (e.g., Heinson 1999). In the Fennoscandian shield MT has imaged such an upper mantle conductor at a depth of approximately 200 km (Jones 1982, Engels et al. 2002, Lahti et al. 2005, Korja 2007, Cherevatova et al. 2015a).

The research encompassed in this thesis aimed at further developing the MT method in the study of the Fennoscandian lithosphere. The thesis is grounded on the three
publications (Papers I–III). At the core of the thesis are MT data acquired in the central Finnish Lapland in 2014 in the context of the Magnetotellurics in the Scandes -project (Cherevatova 2014). The locations of the MT sites on a geological map are presented in Figure 1. The new data set is much denser than existing MT data from the area, and is therefore better suited to analysing the complex geoelectric setting. Earlier MT studies have mostly been based on 1D analysis of sparse MT data from the BEAR project (Vanyan et al. 2002, Lahti et al. 2005). Indeed, as will be shown, the MT data display complex behaviour reflecting the convoluted geological-tectonic history of the area. One of the most intriguing features of the data set is that it contains multiple extreme MT data. These are defined in Paper I as MT responses that contain a negative determinant for the real and/or imaginary part of the impedance tensor. Such data have rarely been considered in the existing literature and are related to large resistivity contrasts and possibly often to complex conductor geometries. Paper I is devoted to the analysis of various aspects of extreme data. Paper II contains a more general analysis of the MT data set and provides a 3D geoelectric model of the crust of central Finnish Lapland, together with discussion of its tectonic implications. Paper III describes a methodology for the express integration of MT and transient electromagnetic (TEM) data. The methodology is applied to data measured in a sedimentary-basin environment in Greece, which is geoelectrically simple. However, although this was not pursued further in the thesis, the methodology could in principle be extended to the much more complex geoelectric environments in Fennoscandia. It is particularly useful to complement MT with TEM, since it allows the static shift problem in MT to be addressed.
1.2 Geological setting of the study area

The study area is located in the northern Fennoscandian shield, where the present-day geology is a result of Palaeoproterozoic reworking of the Archaean Karelian craton. In this section, the main aspects of the tectonic evolution are outlined.

The Fennoscandian (Baltic) Shield is an exposed part of the East European craton (Gorbatschev & Bogdanova 1993) and covers a large part of northern Europe. The Fennoscandian Shield comprises the area bounded by the Caledonides mountains in the northwest and the East European Platform, where the bedrock is covered by...
Phanerozoic sediments, in the southeast (Figure 1b). The oldest Archaean rocks of the
shield constitute the Karelian craton (Ka) and the Kola craton (Ko) in the northeast.
The northwesternmost part of the Archaean domain, the alleged Norrbotten craton (N),
is separated from the Karelian Craton by the Pajala shear zone (PSZ, Lahtinen et al.
2015a), which coincides with the Baltic-Bothnian megashear (Berthelsen & Marker
1986). The current study area, marked with a red rectangle in Figure 1b, is mostly
covered by Palaeoproterozoic supracrustal belts and granitoids.

Studies on the tectonic evolution of northern Finland (see, e.g., Lahtinen et al.
2018, Nironen 2017, Lahtinen et al. 2015a, Tiira et al. 2014, and references therein)
suggest that multiphase intraplate rifting in the Archaean started at ca. 2.5 Ga and
led to continental break-up at ca. 2.1 Ga. A phase with subduction activity and a
subsequent accretional collisional phase commenced at ca. 1.9 Ga with the partly
toeval Lapland–Kola orogen in the northeast and the multiphase Svecofennian orogen in
southwest. Archaean basement faults played an important role in the deposition of
supracrustal rocks as well as structural reactivation during multiple stages of rifting,
basin inversion and fold-and-thrust belt formation (Skyttä et al. 2019). The area was
subject to another phase of crustal shortening around 1.8 Ga accompanied by intrusive
and hydrothermal activity and the formation of gold and copper–gold mineralizations.
Excluding some minor events, the Fennoscandian shield has essentially been intact
since then. At present central Finnish Lapland is a complex assemblage of supracrustal
volcanosedimentary belts and granitoids (Fig. 1a). The Archaean bedrock is exposed in
the south and northeast. On average, less than 10 m of Quaternary sediments cover the
bedrock. This is thin enough to impose a negligible influence on MT measurements, but
thick enough to hamper, for example, the field mapping of shear zones. Thus, to a large
extent, the faults and shear zones depicted in the area (Fig. 1a) have been indirectly
inferred from geophysical data such as aeromagnetic maps.

The supracrustal Central Lapland Greenstone Belt (CLGB), the Peräpohja Belt
(PB) and the Kuusamo Belt (KB) are composed of sedimentary and volcanic rocks that
were mainly deposited on the Archaean crust during the extensional stage (2.5–1.9
Ga). The belts have been metamorphosed under varying metamorphic conditions. The
NW–S-trending CLGB is one of the largest Palaeoproterozoic greenstone belts in the
world and can be traced from northern Norway through Sweden and Finland to Russian
Karelia (Hanski & Huhma 2005). A significant tectonic marker in the CLGB is the
Kittilä allochthon (Hanski & Huhma 2005), located just north of the current study area,
possibly thrust on top of the Karelian craton from the west in the Karelia–Norbotten
collision (Lahtinen et al. 2015a). The related Kittilä conductor (Lahti et al. 2012, Cherevatova et al. 2015a,b) is an upper crustal feature with the highest conductivities related to graphite and sulphide-bearing schists.

The Central Lapland Granitoid Complex (CLGC) is composed of various granite and gneiss assemblages with a wide range of ages from 2.7 to less than 1.8 Ga (Nironen 2005, Sorjonen-Ward & Luukkonen 2005, Ahtonen et al. 2007). Nevertheless, a large part of the rocks are late-orogenic granitoids (1.84–1.80 Ga) related to the latest stage of crustal shortening. Most of the granitoids are indicated to have been derived from Archaean crust (Ahtonen et al. 2007). Recently, the CLGC has been suggested to represent a root of an aulacogen (a failed rift arm of a triple junction) related to the break-up of the Archaean craton (Lahtinen et al. 2015a,b). This possibly explains, for instance, the intrusion of granitic magmas in the CLGC area from 2.1 Ga onwards. Later in the contractional stage, a crustal block consisting of the Archaean Pudasjärvi complex (PC) and the basement of the PB and the CLGC moved northward at ca. 1.9 Ga due to the Svecofennian collisions in the south (Nironen 2017). During this the Hirvaskoski Shear Zone (HSZ) in the southeastern part of the study area was developed (reactivated), and this could explain the large horizontal displacement along the shear zone suggested by Airo (1999). The Venejoki Shear Zone (VSZ) (Fig. 1a, after Nironen 2017) would represent the northern margin of the alleged Archaean block. The northward movement of the block would have led to thickening of the crust in the collision zone, and the area of CLGC would have become a thermal centre during the orogenic collapse.

1.3 Earlier geophysical studies

Previously acquired MT data overlapping with the current study area include the BEAR project (Korja et al. 2002), earlier MasCa data in the northwestern part and the EMMA project (Smirnov et al. 2008b) in the southeastern part. These data have been included into the analysis contained in this thesis. Older EM investigations partially overlapping with the study area include the horizontal spatial gradient and induction vector studies (Pajunpää 1988, 1989).

Various seismic experiments and research have also been conducted in central Lapland. The reflection seismic data from the FIRE project (Kukkonen & Lahtinen 2006) FIRE4 profile, which crosses the study area in a NNS–SSW direction (Fig. 1) have been thoroughly analysed by Patison et al. (2006). During the FIRE experiment, some wide-angle reflection and refraction data were also acquired, and an upper crustal
seismic velocity model was derived by Silvennoinen et al. (2010). Silvennoinen et al. (2010) additionally obtained a density model from the available regional-scale gravity data (Korhonen et al. 2002). A crustal seismic velocity model in the eastern part of the study area was defined by Tiira et al. (2014) from the wide-angle reflection and refraction HUKKA2007 profile. Data from the broadband seismic POLENET/LAPNET network have been used for teleseismic P-wave tomography (Silvennoinen et al. 2016) and seismic anisotropy studies of the lithosphere (Plomerová et al. 2011, Vinnik et al. 2014). Finally, the study area is covered by low-altitude airborne mapping by the Geological Survey of Finland (Hautaniemi et al. 2005). The nationwide airborne mapping, with a nominal flight altitude of 30 m and line spacing of 200 m, includes magnetic, shallow-penetrating fixed-wing EM and radiometric data.

1.4 The 2014 MT data set

At the core of this thesis research are the magnetotelluric data acquired in 2014 in central Finnish Lapland in the context of the MaSca project (Cherevatova et al. 2015a). MaSca MT data have previously been acquired from the northern parts of Sweden and Norway, and the current data set thus extends the data coverage to the Finnish side. Note that previously measured MT data have also been included in the analysis (Fig. 1, section 1.3), but are not described here.

During April–June 2014, total of 79 broadband (BMT) and long period (LMT) MT sites, supported with vertical magnetic field measurement at each site, were installed in central Finnish Lapland. The horizontal EM fields were measured in the geomagnetic N–S and E–W directions using the MTU2000 system developed at the University of Uppsala (Smirnov et al. 2008a). In addition, three LEMI LMT systems were borrowed from the University of Leicester, UK. The data have been rotated to align the coordinate axes x and y with the geographic N–S and E–W directions, respectively. Site spacing is ca. 10–20 km for BMT sites and 20–40 km for LMT.

The electric field was measured using Pb/PbCl₂ electrodes, with dipole lengths ranging from 70 to 100 m, depending on site conditions. The magnetic field was measured using LEMI-120 and Metronix MFS05 coils at the BMT sites and LEMI fluxgates at the LMT sites. The BMT instruments were set to record for 20–48 hours (with exceptions towards longer times). The BMT time series contain data with two sampling frequencies, at 20 Hz full-time and at 1000 Hz for two hours during the local night-time. On average, two of the total of three BMT instruments were moved to a new
location each day, and for every third site a two-day recording is thus available. Twelve LMT instruments simultaneously recorded (1-Hz sampling) for approximately for 3–4 weeks. The LMT instruments were moved once to new locations in the middle of the field campaign. A remote reference signal was (mostly) available from two and eleven remote sites, respectively for BMT and LMT.

The transfer functions (impedance tensor and tipper) were processed from the recorded time series using the robust remote-reference code described in Smirnov (2003) and Smirnov & Pedersen (2009). As a result, good quality transfer functions are available within the period ranges of approximately 0.003–1000 s (BMT) and 10–10000 s (LMT). Approximately 10% of the sites were discarded due to poor quality.

The study area is located at a relatively high latitude (67°N), where the proximity of the source currents (the polar electrojet, for instance) may violate the planewave condition assumed in MT (e.g., Osipova et al. 1989, Engels et al. 2002, Viljanen 2012). The source-field effects were investigated in the BEAR project and it was concluded that using robust processing methods, the plane-wave assumption was valid up to a period of 10000 s (Varentsov et al. 2003b,a, Sokolova et al. 2007). However, the tipper transfer function is already likely to be affected by the source field at shorter periods. This is illustrated by an averaged representation of the tippers in Figure 2. As can be seen, the length of the real induction vectors increases in an exponential fashion towards the longer end of the period range. At the same time, the vectors rotate towards the north, i.e., towards the source. The increase in the length is similar to that predicted by the simple source field model by Rokityansky (1982), represented by the hashed line in the figure. However, it would be difficult to estimate at which point the source field effect becomes non-negligible (e.g. > 5%) in the case of real data. To mitigate the possible source field effects, periods greater than ca. 2000 s have been excluded from further tipper data analysis.

Cherevatova et al. (2015a) additionally recognised that their tipper data were affected by the Atlantic Ocean in the west. The current study area is located further east and no effect from the ocean (the coast effect) is expected. In fact, the decay of the tipper as a function of distance from the northern Norwegian coast towards Sweden and Finland was already demonstrated by Jones (1981).
Fig. 2. Average behavior of induction vectors. The circles represent the arithmetic mean length and the arrows on top the mean direction (Wiese) of the observed real induction vectors. In addition, the predicted response from tipper inversion from Paper II is presented. For reference, the hashed line represents the simple theoretical estimate for the ratio of the vertical to horizontal normal magnetic field amplitude ($|H^0_z / H^0_h|$), equal to 0 in case of no source effect), based on formula 6.42 of Rokityansky (1982). In the formula, a distance of 800 km to the “polar electrojet” source was used and the apparent resistivities were taken as an average computed from all the 2014 survey determinant impedance data. Reprinted, with permission, from Paper I © Elsevier.
2 Electromagnetic methods

2.1 General

Since this thesis is mainly concerned with the MT method, it will be emphasized in this chapter. Only a brief description of the TEM method, relevant in Paper III, will be provided. Additionally, the transfer functions from the controlled source MT method CSRMT, also used in Paper III, are equivalent to MT data. Details of the practical aspects of the CSRMT are given by Bastani (2001).

The behaviour of electromagnetic fields considered in the thesis is governed by Maxwell’s equations. The relevant methods (MT, TEM) concern relatively slowly varying EM fields, which renders the field behaviour diffusive rather than wave-like. This so-called quasistatic assumption holds under the condition

\[ \sigma \gg \omega \varepsilon, \]

where \( \sigma \) is the electrical conductivity, \( \omega = 2\pi/T \) is the angular frequency of the EM field, \( T \) being the period, and \( \varepsilon \) is the permittivity. Instead of conductivity, one may refer to resistivity \( \rho = 1/\sigma \). In this thesis, both resistivity and conductivity are used interchangeably. In brief, the quasistatic behaviour of the EM field can be described in the frequency domain and in terms of the electric field by the following system of partial differential equations (e.g. Egbert & Kelbert 2012)

\[ \nabla \times \nabla \times \mathbf{E} + i\omega \mu_0 \sigma \mathbf{E} = 0. \]

Note that in Eq. 2 and throughout the thesis, \( e^{i\omega t} \) sign convention is assumed. The electric field is a function of position \( \mathbf{r} \) and frequency, i.e., \( \mathbf{E} = \mathbf{E}(\mathbf{r}, \omega) \). In Eq. (2), \( \mu_0 \) is the magnetic permeability of free space. Thus, as is common, possible magnetic effects are ignored. Also note that permittivity is not present in the equation, a consequence of the quasistatic assumption (displacement currents are ignored). Additionally, it is assumed that the medium is linear, isotropic and frequency independent. For a given conductivity \( \sigma(\mathbf{r}) \) defined within some domain, Eq. (2) together with the applied boundary conditions completely describes the behavior of the electric field. Note that Eq. (2) does not contain a source term, which means that the source must be enforced through the boundary conditions. This is customary for MT, but for controlled source methods a source term can be explicitly included in (2). Of course, Eq. (2) is only one
way of representing the physics of the EM methods, and in practice, the formulation of
the Maxwell’s equations depends on the specific problem at hand. However, Eq. (2)
is used as the forward modelling starting point in ModEM (Egbert & Kelbert 2012),
where an approximate solution is sought numerically by discretizing the equation using
a staggered grid finite difference method. After the electric field solution has been
obtained, the magnetic field can be obtained from Faraday’s Law.

As a result of the slowly varying fields (quasistatic assumption), methods such as MT
and TEM are sensitive to spatially averaged volumes of the geoelectric structure. This
contrasts with high-frequency wave propagation methods such as ground penetrating
radar, which sense reflections from layer interfaces of different media. The low frequency
(long period) allows a greater depth of investigation at the expense of resolution (Ward &
Hohmann 1988). The investigation depth of electromagnetic methods is characterized
by the skin-depth ($\delta$), defined as the depth by which the amplitude of a harmonically
varying planar EM wave travelling in a homogeneous resistivity medium has decayed to
1/e of its initial value.

$$\delta = 500\sqrt{\rho T},$$  (3)

where $\rho$ is the resistivity of the medium and $T$ is the period of the EM field. Although
the above skin-depth formula is best used as a rough guideline for plane-wave methods
such as MT, the principle is general and affects all EM fields. Therefore, the depth of
investigation is controlled by varying the period of the EM field (slower varying fields
penetrate deeper). For controlled source methods, geometric attenuation and the limited
strength of the artificial source field additionally affect the achievable maximum depth of
investigation.

### 2.2 Electrical conductivity

The EM methods considered in the thesis image the electrical conductivity structure of
the subsurface. Conductivity is the ability of the material to carry electrical current,
physically arising from the movement of net electrical charge within the material. This
is described by Ohm’s, law which states that an applied electric field gives rise to a
current density $j$, which is proportional to the conductivity,

$$j = \sigma E.$$  (4)
Knowledge of the electrical conductivity distribution is useful, because it varies greatly between different geological materials. Figure 3 presents resistivity ranges for various geological materials.

Most rock-forming minerals are highly resistive and the bulk resistivity is often controlled by volumetrically minor phases. However, to have an appreciable effect, the minor good conducting phase needs to be well interconnected. In Precambrian shields, candidates for strong crustal-scale conductors are graphite, sulphides and iron oxides. In Fennoscandia, particularly rocks bearing graphite of organic origin generally play a major role in crustal scale conductivity anomalies (Korja 1997, Hjelt et al. 2006).

It has been also demonstrated by MT studies worldwide that the continental lower crust is often relatively conductive. Attempted explanations include saline pore fluids and graphite (Haak & Hutton 1986, Hyndman et al. 1993). In the eastern part of our study area, which is void of strong crustal conductors (Paper II), a relatively conductive crustal layer starting from a depth of ca. 20 km is resolved by MT (Vanyan et al. 2002, Lahti et al. 2005). Korja et al. (2002) contrasted this with many other Archaean parts of Fennoscandia which show highly resistive crust throughout. In our study area, however, the very strong upper/mid crustal conductors generally prevent imaging the possible enhanced conductivity in the lower crust (Paper II).

At upper mantle depths, the electrical conductivity is to a large extend controlled by temperature due to the semiconductor nature of the minerals, and electrical resistivity therefore generally decreases as a function of depth. However, the decrease is not enough to explain the observed enhanced conductivity at the electrical lithosphere–asthenosphere boundary (ELAB), the top of which in the Fennoscandian shield has been interpreted at depths of ca. 200 km (Jones 1982, Engels et al. 2002, Lahti et al. 2005, Korja 2007, Cherevatova et al. 2015a). The ca. 200 km depth is also in agreement with the data analysed in this research, as demonstrated in Paper II. The inferred resistivity of the ELAB is in the order of 10 Ωm, while the electrical conductivity of dry olivine at such a depth is in the order of $10^3$ Ωm (e.g., Constable 2006), assuming a temperature of around 1200 °C from the geotherm estimated by Kukkonen & Peltonen (1999). Partial melts or hydrogen diffusion in olivine are usual candidates for explaining the enhanced conductivity of the ELAB (see, e.g., Heinson 1999).
2.3 The magnetotelluric method

In the magnetotelluric (MT) method, temporal variations of the Earth's natural electromagnetic (EM) field are measured on its surface (Tikhonov 1950, Cagniard 1953). Generators of the natural EM fields include global thunderstorm activity ($T < 1$ s) and electric currents flowing in the ionosphere and magnetosphere ($T > 1$ s) (e.g., Viljanen 2012). Although the measured EM field variations can be considered essentially random, relationships exist between field components (e.g., Berdichevsky & Zhdanov 1984, Weidelt & Chave 2012), which allow the resistivity information to be extracted. The relevant linear relationships (transfer functions) are introduced in the following.

We use a coordinate system where the $x$-axis points towards geographic north, $y$-axis towards geographic east and $z$-axis vertically downwards. At a given location on the surface of the Earth, the horizontal electric ($E_x, E_y$) and magnetic ($H_x, H_y$) fields in the frequency domain are related through the impedance tensor

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

(5)
where, $Z_{xx}$, $Z_{xy}$, $Z_{yx}$ and $Z_{yy}$ are the impedance tensor components. In shorter matrix notation, we write

$$E = ZH \quad \text{or} \quad E = (X + iY)H,$$

(6)

where $i$ denotes the imaginary unit and, for the purposes of this thesis, the impedance tensor has been split into its real ($X$) and imaginary ($Y$) parts. Expanded, the matrices read

$$X = \begin{pmatrix} X_{xx} & X_{xy} \\ X_{yx} & X_{yy} \end{pmatrix} \quad \text{and} \quad Y = \begin{pmatrix} Y_{xx} & Y_{xy} \\ Y_{yx} & Y_{yy} \end{pmatrix}$$

(7)

If, in addition to the horizontal magnetic field, the vertical magnetic field ($H_z$) is also measured, we can additionally define the vertical magnetic transfer function (tipper), which relates the fields by

$$H_z = T_{zx} H_x + T_{zy} H_y,$$

(8)

where $T_{zx}$ and $T_{zy}$ are the tipper elements.

It is assumed that both the impedance tensor and the tipper are complex functions of only the location, the period (excitation frequency of the source field) and the subsurface resistivity structure (Berdichevsky & Zhdanov 1984). In particular, it is not dependent on the source field morphology, which is assumed to be a vertically incident plane wave. This assumption can be violated at long periods due to the proximity of the source currents (e.g., Viljanen 2012). As was discussed in the section 1.4, violation of the plane-wave condition was also observed in the case of the tipper data considered in this thesis.

Due to the analytical content of this thesis (in particular, Paper I), it is necessary to summarize various ways of representing the information contained in the impedance tensor and tipper. Commonly, the impedances are represented as apparent resistivity ($\rho \ [\Omega m]$) and the impedance phase ($\phi \ [\text{deg}]$), which are defined, for example, for $Z_{xy}$ (analogously for the rest of the elements) as

$$\rho_{xy} = \frac{T}{2\pi \mu_0} |Z_{xy}|^2,$$

(9)

and

$$\phi_{xy} = \text{arg}(Z_{xy}).$$

(10)

For a uniform resistivity earth (homogeneous half-space), $\phi_{xy} = 45^\circ$. Strictly, then $\phi_{yx} = -135^\circ$, but for convenience, it is customary in graphical representation to set
\( \phi_{xy} + 180^\circ \). The determinant (average) of the impedance tensor \( Z_{\text{det}} \) is defined as (Berdichevsky & Dmitriev 1976)

\[
Z_{\text{det}} = \sqrt{Z_{xx}Z_{yy} - Z_{xy}Z_{yx}} = \sqrt{|\det(Z)|} = |Z_{\text{det}}|e^{i\phi_{\text{det}}},
\]

where the absolute value is given by

\[
|Z_{\text{det}}| = \sqrt{|\det(Z)|}
\]

and \( \phi_{\text{det}} \) is the determinant phase given by

\[
\phi_{\text{det}} = \frac{1}{2} \arg(\det(Z)).
\]

Apparent resistivity for the determinant impedance is defined analogously as for individual impedance tensor elements (Eq. 9). \( Z_{\text{det}} \) is rotationally invariant, i.e., unlike the individual impedance tensor elements in general, it is independent of the orientation of the coordinate system it is expressed in.

A direct graphical representation of the impedance tensor called telluric vectors (e.g., Bahr 1988) are constructed by (here, only the real part is considered, similarly for the imaginary part)

\[
e_x = X_{xx} \hat{x} + X_{yx} \hat{y}
\]

\[
e_y = X_{yx} \hat{x} + X_{yy} \hat{y},
\]

where \( \hat{x} \) and \( \hat{y} \) are the north- and east-directed unit vectors, respectively. \( e_x \) is the electric field related to the unit magnetic field varying in the x-direction and \( e_y \) is the electric field related to the unit magnetic field varying in the y-direction. As an example, for a 1D structure, the telluric vectors are orthogonal to the corresponding magnetic field direction and thus also to each other.

The phase tensor \( \Phi \) introduced by Caldwell et al. (2004) is defined as

\[
\Phi = \frac{1}{\det(X)} \begin{bmatrix} X_{yy}Y_{xx} - X_{xy}Y_{yx} & X_{xy}Y_{xx} - X_{xx}Y_{yy} \\ X_{xx}Y_{xy} - X_{yx}Y_{xy} & X_{xx}Y_{yy} - X_{xy}Y_{xx} \end{bmatrix} = \begin{bmatrix} \Phi_{xx} & \Phi_{xy} \\ \Phi_{yx} & \Phi_{yy} \end{bmatrix},
\]

where \( \Phi_{xx}, \Phi_{xy}, \Phi_{yx} \) and \( \Phi_{yy} \) are the phase tensor elements which are real. In the 2D case, when the coordinate system axes are aligned in parallel and perpendicular to the strike direction,

\[
\Phi = \begin{bmatrix} \Phi_{xx} & 0 \\ 0 & \Phi_{yy} \end{bmatrix} = \begin{bmatrix} \tan(\phi_{xx}) & 0 \\ 0 & \tan(\phi_{xy}) \end{bmatrix}.
\]
The phase tensor can be visualised as an ellipse via multiplying it by the unit circle
\( c = (\cos(\alpha) \sin(\alpha))^T, 0^\circ < \alpha < 360^\circ \). In practice, however, the ellipse parameters are
easily obtained from the parameterization described below. In the 2D case, the ellipse
axes can only be oriented parallel or perpendicular to the geoelectric strike. Note that the
ellipse axis in the \( x \)-direction (\( \Phi_{xx} \)) is equivalent to the impedance phase related to
the electric field in the \( y \)-direction (\( \phi_{yx} = \tan^{-1}(\Phi_{xx}) \)). In the
same way, \( \phi_{xy} = \tan^{-1}(\Phi_{yx}) \). Thus, when interpreting either of the ellipse axes in a
2D scenario, it should be kept in mind that it corresponds to the current flow in the
orthogonal direction. In practice, when the phase tensor is parameterized as described
next, the strike-aligned phase tangents are immediately recovered as the phase tensor
principal values.

The phase tensor can be parameterized after Booker (2014) as
\[
\Phi = R^T(\theta) \begin{pmatrix} \Phi_{max} & 0 \\ 0 & \Phi_{min} \end{pmatrix} R(2\beta) R(\theta). \tag{18}
\]
To conform to the original notation of Caldwell et al. (2004), \( \beta \) is used as the skew
(indicator of 3D effects) instead of \( \psi = 2\beta \) by Booker (2014). Additionally, the principal
values, the magnitudes of which represent the lengths of the phase tensor ellipse
semi-major and semi-minor axes, are ordered so that \(|\Phi_{max}| \geq |\Phi_{min}| \). The angle \( \theta \) is an
azimuth of the phase tensor ellipse major axis (measured positive clockwise from the
"north"-pointing \( x \)-axis, the \( y \)-axis points towards "east"), \( R(\theta) = (\cos \theta \ \sin \theta \ \sin \theta \ \cos \theta) \) and \( T \) denotes the transpose operation. Note that \( \theta \) is equal to \( \alpha - \beta \) of Caldwell et al. (2004),
and by substitution, it is shown that (18) is equivalent to their original parameterization.
All the parameters in (18) may be obtained using the formulae provided by Bibby et al.
(2005), summarized in Appendix 1. Note that the four-quadrant inverse tangent function
\( \text{atan2} \) must be used in the computation of \( \beta \) and \( \theta \), otherwise, incorrect parameterization
may result.

The tipper (Eq. 8) can be visualised via the real (\( T^{Re} \)) and imaginary (\( T^{Im} \)) induction
vectors, defined as
\[
T^{Re} = \text{Re}(T_{zx})\hat{x} + \text{Re}(T_{zy})\hat{y}, \tag{19}
\]
\[
T^{Im} = \text{Im}(T_{zx})\hat{x} + \text{Im}(T_{zy})\hat{y}. \tag{20}
\]
Formulae (19) and (20) conform to the Wiese convention (Wiese 1962) in which the real
vectors typically point away from conductors. An alternative is to use the Parkinson
convention (Parkinson 1959), where the direction of the real vector is reversed compared
to the Wiese convention. In addition, changing the sign convention would lead to reversing the direction of $T_{im}$ (Lilley & Arora 1982).

### 2.4 Static shift in MT

In the presence of an electric field, excess charge accumulates on resistivity gradients (e.g., interface between two different resistivity domains), which can lead to the so-called galvanic distortion of the measured electric field (Jiracek 1990, Zhang et al. 1987). Here, we do not consider the possible effects on the magnetic field (e.g., Smith 1997). The basic scenario is that a "small" near-surface resistivity anomaly is the source of the distortion and if the period at which the response is considered is sufficiently long, the distortion of the measured electric field is frequency independent. In a simple scenario of a 1D regional structure affected by galvanic distortion this leads to a frequency independent shift (static shift) in the apparent resistivity curve, while the impedance phase would be unaffected.

Generally, the effect of galvanic distortion of the electric field in MT can be described by a galvanic distortion tensor ($C$) applied to the undistorted impedance tensor ($Z$). The distorted (superscript $m$) impedance becomes

$$Z^m = CZ = \begin{pmatrix} C_{xx} & C_{xy} \\ C_{yx} & C_{yy} \end{pmatrix} \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix}$$

(21)

Elements of $C$ are real and frequency independent. Considering a 1D structure, only the off-diagonal elements are non-zero in the undistorted impedance tensor and it is easy to show that the measured apparent resistivity, e.g., for the $xy$-impedance element, becomes

$$\rho^m_{xy} = s_{xy} \rho_{xy},$$

(22)

where $s_{xy} = C^2_{xy}$. For the determinant average the effect of galvanic distortion of apparent resistivity is

$$\rho^m_{det} = |\det(C)| \rho_{det}$$

(23)

In practice, the effect of galvanic distortion for the determinant impedance is the same form as depicted for an individual impedance element by Eq. 22. The difference is that for the determinant impedance, such an equation is true for any dimensionality of the regional structure and, moreover, the distortion is rotationally invariant. The unknown static shift factor can be corrected for, for example, by coincident TEM measurement,
which is assumed to be free from the effect of the near surface inhomogeneities (Pellerin & Hohmann 1990, Meju 1996, Paper III). One such procedure is discussed in Paper III. The determinant phase in the case of galvanic distortion is

\[ \phi_{det}^m = \frac{1}{2} \text{arg} (\det(C) \det(Z)). \]  

(24)

From this it is evident that galvanic distortion will only affect the determinant phase if the determinant of the galvanic distortion matrix is negative, leading to a quadrant change of the phase. Note that such a condition manifests extreme data (see next section). Synthetic modelling presented in Paper I suggests that such a negative determinant for the galvanic distortion tensor may exist.

The phase tensor is not affected by galvanic distortion since (following the above notation for \( Z \))

\[ \Phi^m = (C X^{-1}) C Y = X^{-1} C Y = X^{-1} Y = \Phi. \]  

(25)

However, if the phase tensor is represented in the parameterize form described in Paper I and summarized in the next section, galvanic distortion may lead to out-of-quadrant principal value phases if the galvanic distortion tensor is negative. For this to be recognized requires that the data are available through a sufficient period range to track the possible quadrant changes.

2.5 Extreme MT data

In Paper I, extreme MT data were defined as an impedance tensor whose real and/or imaginary part has a negative determinant, i.e., if one or both of the following holds

\[ \text{det}(X) < 0 \]  

(26)

\[ \text{det}(Y) < 0. \]  

(27)

Extreme data can result from 2D and 3D conductivity structures. In the 2D case, extreme data can be directly related to off-diagonal impedance phases in the strike coordinates (Eq. 17). In particular, extreme data are related to cases where the TE mode phase goes out of its nominal quadrant, i.e., the quadrant where the phase generated by 1D structures must be located. The TM mode phase cannot leave the nominal quadrant (Weidelt & Kaikkonen 1994). However, in the general 3D case, an out-of-quadrant off-diagonal phase is not a sufficient condition for extreme data. On the other hand, if data are extreme, out-of-quadrant off-diagonal phases must exist at least for some orientation of the measurement axis (Lilley 1998).
The implications of extreme data for various data types are now summarised (Paper I). For the phase tensor, its relation to extreme data is indicated by (Bibby et al. 2005)

\[
det(\Phi) = \frac{det(Y)}{det(X)} = \Phi_{\text{max}}\Phi_{\text{min}}
\]  

(28)

showing that extreme data have a connection to negative phase tensor principal values. In particular, if one of the determinants is negative, then one of the principal values must also be negative. In general, it is desirable to allow both principal values to obtain negative values (Booker 2014, Paper I). However, it turns out that when the parameterization is constructed using the formulae of Bibby et al. (2005), summarized in Appendix 1, the maximum principal value is positive by definition. Although the formulae always lead to mathematically valid parameterization, the parameters may endure inconvenient quadrant jumps in the case of extreme data. This is illustrated in Figure 4 by BMT data from central Lapland. At this site, extreme data are encountered in the period range of approximately 30 s to 100 s, and \(det(X) < 0\). In a), the parameters are as given by the formulae of Bibby et al. (2005). It is observed that approaching the extreme data section e.g. from the shorter period end, the maximum principal value phase approaches 90° (maximum principal value approaches infinity). The phase exceeding 90° means that the principal value changes sign. However, what instead happens is that the parameterization changes due to the positivity requirement of \(\Phi_{\text{max}}\). In particular, the following changes take place

\[
\Phi_{\text{max}} \rightarrow -\Phi_{\text{max}}
\]

(29)

\[
\Phi_{\text{min}} \rightarrow -\Phi_{\text{min}}
\]

(30)

\[
\beta \rightarrow \beta \pm 90°.
\]

(31)

By appreciating that \(R(2\beta) = -R(2\beta \pm 180°)\), it is easy to see that the above modification does not affect the validity of the parameterization. To bring the parameterization to a more convenient form in the period range of extreme data, the modification (29)-(31) is applied for the parameters presented in a), rendering the parameters continuous and more easily interpreted, as shown in b). The final point to consider is the fact that the phase tensor does not inherently contain information about the principal phase quadrants. That is, if one considers the parameters at any isolated period, the phase quadrant cannot be deduced without additional information. The proper quadrant can be recovered by fixing the quadrant at some period and then tracking the quadrant changes as a function of the period. For the above example and for most MT data in general, this should be
possible, since most data can be assumed to be non-extreme for the major part of their period range.

It is also interesting to note that at site r68, in the measurement coordinate system, the off-diagonal impedance elements do not show anything out of the ordinary, such as out-of-quadrant phases (see Paper II, Fig. 8). On the other hand, site r14 shows an out-of-quadrant yx-phase although the data is not extreme (see also Paper II, Fig. 8).

\[
\tan^{-1}\left(\Phi_{\text{max}}\right) - \tan^{-1}\left(\Phi_{\text{min}}\right) = \beta - \theta
\]

Fig. 4. Behaviour of the phase tensor parameters from BMT site r68. a) Unmodified parameterization. b) Modified parameterization obtained from the parameters depicted in a). The shaded area shows the period range for extreme data \((\det(X) < 0)\). Reprinted, with permission, from Paper I © Elsevier.

In terms of the (real, analogous for imaginary part) telluric vectors, the condition for extreme data can be stated by writing the determinant as

\[
\det(X) = |e_x||e_y|\sin\gamma,
\]

where \(\gamma\) is the clockwise angle from \(e_x\) to \(e_y\). Thus, the determinant is negative for \(-180^\circ < \gamma < 0^\circ\). It is finally noted that in case either \(\det(X) = 0\) or \(\det(Y) = 0\), the telluric vectors are parallel or antiparallel and the phase tensor ellipse reduces to a line (\(\Phi_{\text{min}}/\Phi_{\text{max}} = 0\)).

To provide an explicit example of how extreme data may originate, we consider a simple example, that is, the 2D model originally considered by Selway et al. (2012) and further analysed in Paper I. The model is depicted in the Figure 5 and consists of a semi-infinite conductive thin sheet embedded on top of a resistive background. The response is considered on top of the conductor at some distance from the sheet edge. Extreme data behaviour is observed in the TE mode (current flows "into the page").
Fig. 5. The 2D model of Selway et al. (2012) and its MT response computed by REBOCC. a) The resistivity model. The MT site location is depicted by the black triangle. b) Phase tensor principal value phases, equivalent to the TE and TM impedance phases. c) The phase tensor ellipses from the MT site. The model strike is aligned in the x-direction. Reprinted, with permission, from Paper I © Elsevier.

Generally, the 2D impedance tensor in the strike coordinates is

$$Z_{2D} = \begin{pmatrix} 0 & X_{xy} \\ X_{yx} & 0 \end{pmatrix} + i \begin{pmatrix} 0 & Y_{xy} \\ Y_{yx} & 0 \end{pmatrix}$$

(33)

The determinants of $X$ and $Y$ follow as

$$\det(X_{2D}) = -X_{xy}X_{yx} \quad \text{and} \quad \det(Y_{2D}) = -Y_{xy}Y_{yx}$$

(34)

$$\phi_{xy} = \arg(Z_{xy}) = \tan^{-1}\left(\frac{Y_{xy}}{X_{xy}}\right).$$

(35)

In the considered example, the phase becomes negative, meaning that $\det(Y)$ becomes negative as well. Generally, in the 2D case, the signs of $\det(X)$ and $\det(Y)$ are determined by the quadrant that the TE mode phase is located in. This is also evident by comparing the complex plane representation in Figure 6 to Eq. 34.
2.6 Inversion of MT data

To obtain a quantitative estimate of the conductivity distribution in the subsurface, the data are inverted using an inverse modelling algorithm. Here, we describe the principle of MT data inversion for the 3D case used in the inversion code ModEM (Egbert & Kelbert 2012, Kelbert et al. 2014, Meqbel 2009). For the 2D data inversion in Paper III, we used REBOCC (Siripunvaraporn & Egbert 2000) with the modification of Pedersen & Engels (2005) to allow inversion of the determinant impedance. Although there are differences in the implementation, the general formulation of the inverse problem (Eq. 36) is the same for REBOCC and ModEM. One notable difference is that REBOCC uses the apparent resistivity and impedance phase as input data, while ModEM uses directly the complex impedances.

In the forward computation needed for inversion, the algorithm of ModEM is based on Eq. 2. The total electric field solution is obtained using a staggered-grid finite-difference formulation. As described in section 2.1, the equations are solved using the quasistatic approximation (the displacement currents are neglected) and the magnetic permeability is equal to vacuum permeability. The boundary conditions are derived from 2D solutions at the side edges of the model, the perfect conductor condition ($E_x=E_y=0$) is used at the bottom edge and the constant field value (the plane-wave condition), polarized in the $x$- or $y$-direction, at the top edge of the model, which includes the virtually perfectly resistive air layer. The magnetic field can be obtained from the electric
field solution through Faraday’s law. Using the solutions for the two polarizations, the impedance tensor (Eq. 5) and the tipper (Eq. 8) predicted by the resistivity model can be defined.

For inversion, ModEM uses a nonlinear conjugate gradient scheme to minimize an objective functional formulated in a least squares sense. The algorithm attempts to find a spatially smoothly varying resistivity model, preferably similar to the prior model, which also predicts the measured data within the estimated data uncertainty. Like any other gradient-based algorithm, a local rather than global minimum of the objective function will be generally achieved. The objective functional to be minimized consists of data misfit and model regularisation terms and is given by (Egbert & Kelbert 2012)

\[ \Phi = (d - f(m))^T C_d^{-1} (d - f(m)) + \lambda (m - m_0)^T C_m^{-1} (m - m_0). \] (36)

In (36), \( d \) is the data, \( m \) is the model parameter, \( m_0 \) is the \textit{a priori} model parameter, \( f(m) \) is the forward functional, \( C_d \) is the data covariance matrix and \( C_m \) is the model covariance matrix. Note that the model covariance matrix is often defined by \( C_m^{-1} = D^T D \), \( D \) being, for example, the derivative operator, but ModEM constructs the model covariance matrix through a recursive autoregressive scheme (Kelbert \textit{et al.} 2014). In practice, the effect is very similar and the amount of smoothing applied can be controlled. The trade-off between data misfit and model regularisation is controlled by \( \lambda \). ModEM uses a cooling scheme, whereby during early iterations, a relatively high \( \lambda \) is used, but it is made smaller once the change in misfit becomes smaller than a preset threshold value between subsequent iterations.

The misfit between the measured data (\( d \)) and the data predicted (\( f \)) from a given conductivity model is given by a (normalised) rms,

\[ \text{rms} = \sqrt{\frac{1}{N} \sum_{i=1}^{N} \left( \frac{d_i - f_i}{\Delta d_i} \right)^2}. \] (37)

where \( N \) is the number of data points and \( \Delta d_i \) is the uncertainty prescribed to the \( i \)th data. Thus, rms equal to unit means that the model predicts the measured data “just within” the prescribed uncertainties, on average. \( N \) amounts, for example in the case of full impedance tensor data from multiple sites, to the number of sites \( \times \) number of periods \( \times 8 \) (the elements of \( X \) and \( Y \)). \( \Delta d_i \) is the larger one from the uncertainty derived from the measurements (95% confidence intervals) and a selected error floor. The absolute error floor is defined as \( |Z_{xy}| \times 10^p \) for real and imaginary parts of \( Z_{xy} \) and \( Z_{xx} \) and as
error floor = \(|Z_{yx}| \times \frac{p}{100}\) for real and imaginary parts of \(Z_{yx}\) and \(Z_{yy}\), where \(p\) is the prescribed percent error floor. For tipper, the error floor is given in absolute units.

2.7 The TEM method

Within the scope of this thesis, the transient electromagnetic method (TEM) is considered as an inductively coupled method where both the transmitter and receiver are of the ungrounded type (closed-loop transmitter, receiver sensitive to the magnetic field). Essentially, a steady-state current fed into the transmitter loop is abruptly terminated and the induced, decaying secondary EM field is recorded by the receiver.

The main differences compared to the MT method are that the TEM method uses a controlled source, the source field is more localized (non-plane-wave) and the data analysis and interpretation are usually carried out in the time-domain. Note that for TEM or for other controlled source methods for which the temporal and spatial behaviour of the source field is known, there is no necessity to define transfer functions. In terms of forward modelling TEM is considerably more challenging compared to MT. 1D solutions already require numerical approaches in contrast to simple analytical expressions for MT.

In this thesis, data measured by the TEM-FAST48 instrument have been analysed in Paper III. This is a low-power TEM-instrument meant for relatively near-surface studies and uses a single-loop configuration. 1D forward modelling and inversion capability is provided by the included TEM-Researcher software. The used modelling algorithms are described in Barsukov et al. (2015).
3 Summary of results

3.1 Paper I: Extreme MT data

Paper I concentrates on the analysis of the extreme MT data observed in central Finnish Lapland. The high data quality made the data set particularly useful to provide further insights into such rarely reported behaviour. The paper provides the definition for extreme data and discusses their properties for various data representations. Both measured data from central Lapland and data from synthetic modelling were used to shed light on extreme data. Based on the synthetic modelling study, a resistivity contrast of at least 1:1000 in the subsurface is required to observe extreme data (see also Selway et al. 2012).

Figure 7 presents phase tensor ellipses from the study area at the period of 128 s. A cluster of extreme data is encircled by the hashed ellipse and individual extreme responses are marked by the minus signs (-) next to the phase tensor ellipses. The surface projection of the conductor revealed by 3D inversion (Paper II) is also shown on the map. It is worth noting that extreme data are observed at both BMT and LMT sites and additionally at a site from the BEAR project.
In Paper I, various concepts related to extreme data were analysed. This included the phase tensor, the telluric vectors and also the determinant impedance. The behaviour of the phase tensor parameterization especially needed clarification. Figure 8 presents the phase tensor principal value phases from central Lapland. Phases greater than 90° or less than 0° signify extreme data. Such a representation requires modifications to the original phase tensor parameterization, which is discussed in section 2.5 together with an illustrative example from central Lapland. For central Lapland data, most extreme data show phases greater than 90°. Similarly as for high impedance phases in general, the reason for this can be that conductive material is being sensed below more resistive
material. However, the exact source for the extreme data remains unclear. The predicted data can explain the main trends in the observed data (see Fig. 11c in the next section and Paper II), but the extreme parts of the responses are not reproduced. However, the extreme data are most likely related to the strong arc-shaped conductor revealed by 3D inversion (Paper II). It was also speculated that possible near-surface graphitic conductors, also identified in the vicinity of the extreme responses, may have an effect.

Fig. 8. Phase tensor principal value phases from the sites within the area encircled by the hashed ellipse in Fig. 7. At many sites, principal value phases exceeding 90° are observed. Note that such a condition is different from and more anomalous than an out-of-quadrant off-diagonal impedance phase. Reprinted, with permission, from Paper I © Elsevier.

Synthetic modelling was undertaken to provide at least some specific scenarios for how extreme MT data maybe generated. The simplest case is possibly the 2D model after Selway et al. (2012) discussed in section 2.5. A more complex 3D model, presented in Figure 9, was also devised. The model mimics a folded conductor squeezed in between otherwise resistive rock. The electric current is forced to follow the path of the conductor, also leading to the reversal of direction of the electric field, which leads to the highly complex MT data behaviour including extreme responses. The model also predicts a negative determinant for the galvanic distortion matrix C. This leads to out-of-quadrant phase tensor principal value phases persisting for long periods and also to an out-of-quadrant determinant phase (Fig. 9c).
Fig. 9. The fold model. a) Model plan view and phase tensor ellipses. A minus sign (-) indicates an extreme response. The surface projection of the conductor is represented by a grey colour. Note that the response is symmetric through the origin of the model. b) Additional model information. c) Phase tensor parameters and determinant phase from the site located at the model origin (within the black circle). Note that due to crossing of the principal phase curves, the principal values are denoted as $\Phi_a$ and $\Phi_b$ (see Paper I). Reprinted, with permission, from Paper I © Elsevier.

3.2 Paper II: Tectonic implications for the Fennoscandian shield

In contrast to Paper I, Paper II was devoted to more general interpretation of the central Lapland MT data (Fig. 1). In particular, a geoelectric model for the bedrock under the study area was obtained via 3D inversion. The used inversion methodology is described in section 2.6. The results were addressed together with existing geological
and geophysical data available from the area, and the implications for the tectonic evolution of central Lapland were discussed. The interpretation was mainly confined to crustal depths. An indication of enhanced conductivity at the assumed base of the lithosphere at ca. 200 km also exists, but the resolution is degraded by the very strong crustal conductors.

The main data characteristics are evident from the induction vector and phase tensor data represented in Figure 10. At the period of 1 s, the induction vectors clearly correlate with the near-surface conductors evident from the airborne apparent resistivity map (a). At longer periods, a clear trend in the data directionality is observed. This trend is shown at the period of 256 s in the figure (b,c), but the trend already starts to develop at much shorter periods (10 s). The induction vectors point towards east (NEE), which would indicate a N–S oriented geoelectric structure. However, this inference does not hold directly within the study area. Although the MT data contain pronounced 3D effects, seen, for example, from the high phase tensor skews ($|\beta| > 3$), the structures in the preferred inversion model from impedance and tipper data (Fig. 11b) appear to have more E–W rather than N–S trends. In particular, highly conductive zones in the north and in the south are separated by the prominent resistor CR (CLGC resistor). An explanation for the circa east-pointing induction vectors could be enhanced conductivity outside the study area in the west. However, confirmation of this would require further investigations.

Two better-resolved higher conductive features are the conductors CC in the north and KB in the southeast. High conductivity areas also exist in the PB area as, but large variations between different inversion runs (not shown) were observed. Denser MT data coverage is required in the PB area to better image its conductivity structure. The conductor KC (Kuusamo conductor) runs along the NE-striking extension of the HSZ (Fig 1) in the northwestern margin of KB. Since the supracrustal KB has been suggested to be rather thin (up to 3 km, Nironen 2017), the large depth of the conductor (top at ca. 15 km) supports the idea of a major reactivated Archaean shear zone (Nironen 2017, Piippo et al. 2019). The main part of the HSZ is located further south outside the current study area, however, and its conductivity is thus presently unknown.

The conductor CC (CLGC Conductor) appears to be a particularly interesting feature. It has conductance in excess of 10000 S, while its top is predicted at the depth of ca. 15 km. This conductor is the source of the local induction vector anomaly in the northern part of the study area and coincides with the area of the observed extreme responses. The conductor CC correlates with results from the FIRE4 seismic reflection profile,
which crosses the study area in a NNW–SSE direction (Fig. 1 Patison et al. 2006). Major south-dipping shear zones in the CLGC have been inferred from the FIRE4 reflection sections. In addition, distinct deformed units are also evident in the parts of the FIRE4 profile spatially correlating with the conductor CC. The conductor CC might represent deeply buried graphite-rich metasediments, and due to its bent geometry, it was possibly affected by both E–W (Norbotten–Karelia) and N–S (Svecofennian orogenies in the south) directed collisional events.

Fig. 10. Phase tensor and tipper data at 1 s on top of an aeroelectromagnetic apparent resistivity map (Airo 2005, © Geological Survey of Finland). Red colours represent low apparent resistivity values, which are interpreted as near-surface (max depth ca. 100 m) conductors. Abbreviations of geological units as in Fig. 1. b) Phase tensor and tipper data at 256 s on top of a smoothed grayscale map of the arithmetic mean of the principal value phases. For comparison, old induction vector data from Pajunpää (1989) are also presented at the period of 300 s. c) Rose diagrams of regional strike indicators. The red dot in the diagrams indicates the direction N80°E. Reprinted, with permission, from Paper II © Elsevier.

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3.3 Paper III: Complementing MT with TEM

In Paper III, a methodology for combining MT and TEM measurements was presented. Combining the two methods is particularly useful, because TEM can be used to address and correct for the static shift afflicting the MT method. The process was
demonstrated for data from the Mygdonian sedimentary basin located in Greece, where the resistivity structure is essentially layered. In principle, combining the methods in more complex geoelectric settings, such as those encountered in Fennoscandia, could also be advantageous.

The method combining the MT and TEM measurements follows the approach described in Pellerin & Hohmann (1990). Essentially, the TEM data is subjected to 1D inversion after which the synthetic MT response is calculated from the inversion model. The novelty in the methodology presented here is that

1. The TEM–MT-transformed data are considered as determinant average MT data;
2. The data from the TEM–MT-transform are subjected to 2D MT inversion together with coinciding MT data

If the prior assumption of 1D structure with galvanic distortion is perfectly satisfied, then there is no difference whether determinant data or one of the off-diagonal impedances is used. However, in the case of lateral conductivity variations, the determinant is more robust as it is rotationally invariant. In the 2D inversion, the assumption is used that the top part of the model, to which the TEM–MT-transformed data are sensitive, is 1D. The possible explicit static shift correction is optional, as it can also be addressed automatically by the inversion. In this case, one would apply higher error floors for the MT apparent resistivity data.

Figure 12 illustrates one of the main results from Paper III, which is the 2D MT inversion model from combined MT and TEM data from the Mygdonian basin. Figure 12b displays the inversion result for MT data only and Figure 12a presents the combined inversion, in which the TEM–MT-transformed data have also been used. In the data set from the Mygdonian basin, the TEM–MT-transformed data extend the MT data range towards shorter periods, resulting in better resolution in the near-surface part of the model. For the combined data result, a static shift correction has been also applied for the MT data. As shown in Figure 13, the MT apparent resistivity curve has been shifted upwards by a factor of 1.5 at site 1409. Comparing the models in Figure 12, it is evident how large an effect the static shift can introduce in the inversion. In this case, the large conductor in Figure 12a has been greatly reduced in size after the static shift correction and inclusion of the TEM–MT-transformed data.
Fig. 12. REBOCC 2D determinant inversion result of TEM–MT-transformed and MT data (a) and MT only data (b). In the MT only result, no static shift correction was applied to site 1409 (located approximately at 1500 m on the profile). Site locations are shown with black triangles for TEM and with red triangles for MT. Vertical and horizontal distances are scaled one to one. Reprinted, with permission, from Paper III © Elsevier.
Fig. 13. Combined response from TEM site 097 (black dots) and MT site 1409 (red dots). Apparent resistivity (a) and impedance phase (b) calculated from the determinant average of the impedance tensor are shown. The MT apparent resistivity curve has been shifted upwards by a factor of 1.5 to correct for static shift. The unshifted apparent resistivity curve is also displayed (grey dots). The site map (c) shows the TEM (squares) and MT sites (triangles) on the inversion profile. Reprinted, with permission, from Paper III © Elsevier.
4 Concluding remarks and recommendations for future research

In this thesis, new methodologies for analysing MT data have been developed and the MT method has been applied to image the geoelectric structure of central Finnish Lapland. The thesis is grounded on the research published in three Papers, I–III. In Paper I, extreme MT data were defined as data for which the determinant of the real or imaginary part of the impedance tensor is negative. Multiple examples of such data were observed in central Finnish Lapland, clustered particularly in the northern part of the study area. Various aspects of extreme data were analysed and clarified, as such data have rarely been considered in the existing literature. In Paper II, a geoelectric model for central Lapland area was obtained via 3D MT inversion and its implications for the tectonic evolution of the area were discussed. In Paper III, a simple methodology for complementing 2D MT inversion with TEM data was introduced.

The exact geoelectric origin of the observed extreme MT data in central Lapland remains unknown, although the revealed CC conductor (Fig. 11b) is most likely a related feature. A further MT study with denser site spacing in the vicinity of the extreme data cluster (Fig. 7) is recommended in the future. The measurements should be supported by extensive synthetic model tests. Such study could bring further insights into extreme data behaviour and would also provide more detailed information on the geoelectric structure of the area, which appears crucial in understanding the tectonic evolution of central Lapland. Major south-dipping shear zones and deformed units can be identified in a FIRE4 reflection seismic profile in the vicinity of the CC (Patison et al. 2006). In addition, the Kittilä allochthon (Hanski & Huhma 2005) is not located far to the north of the CC. To the west exists the suggested boundary, marked by the PSZ, between the Karelia and Norrbotten craton. Possibly due to the Karelia–Norrbotten collision, the Kittilä allochthon was thrust into its current position from the west (Lahtinen et al. 2015a).

The nature of the PSZ itself is also obscure. Recently, MT studies have been conducted to the west of the PSZ (Vadoodi et al. 2018), and combining data from both sides could thus provide insights into its geoelectric nature. The preliminary qualitative impression from the two data sets is, however, that there does not appear to be abrupt
change across the PSZ. A combined study from a larger area would also be warranted to resolve the circa eastward-pointing induction vectors in the current study area.

Denser MT data are needed to resolve the crustal conductivity structure in the PB area. Similarly to the northern part of the study area, near-surface graphite schists cover large parts of the belt. They are often thin and localized features with complex geometries and possess a very high conductivity contrast to neighboring rocks. However, possible elongated, deep-reaching extensions of such conductors are often tectonically significant, holding clues about past evolution. Finally, modelling MT data from such complex geoelectric structures is challenging, which drives the need for new advancements in forward and inverse modelling algorithms. Exploring the TEM–MT methodology of Paper III in a 3D scenario could also prove useful. Instead of or in addition to TEM, other EM methods could also be used to connect smaller scale structures to larger scale structures resolved by MT. For example, the low-altitude airborne electromagnetic data of the Geological Survey of Finland (Hautaniemi et al. 2005) cover the whole of Finland.


References


Special Paper 43.


Appendix 1 Calculation of the phase tensor parameters

For completeness, the formulae for the phase tensor parameters are summarized (Caldwell et al. 2004, Bibby et al. 2005). Note that the formulae in Caldwell et al. (2004) may lead to incorrect results when applied to extreme data (Moorkamp 2007), and thus the formulae from Bibby et al. (2005) are used instead. The original phase tensor parameterization is written as

\[
\Phi = R^T (\alpha - \beta) \begin{pmatrix} \Phi_{\text{max}} & 0 \\ 0 & \Phi_{\text{min}} \end{pmatrix} R (\alpha + \beta). \quad (A.1)
\]

The parameters can be calculated by

\[
\Phi_{\text{max}} = \frac{1}{2} \left[ (\Phi_{xx} - \Phi_{yy})^2 + (\Phi_{xy} + \Phi_{yx})^2 \right]^{\frac{1}{2}} + \frac{1}{2} \left[ (\Phi_{xx} + \Phi_{yy})^2 + (\Phi_{xy} - \Phi_{yx})^2 \right]^{\frac{1}{2}},
\]

\[
\Phi_{\text{min}} = -\frac{1}{2} \left[ (\Phi_{xx} - \Phi_{yy})^2 + (\Phi_{xy} + \Phi_{yx})^2 \right]^{\frac{1}{2}} + \frac{1}{2} \left[ (\Phi_{xx} + \Phi_{yy})^2 + (\Phi_{xy} - \Phi_{yx})^2 \right]^{\frac{1}{2}},
\]

\[
\alpha = \frac{1}{2} \tan^{-1} \left( \frac{\Phi_{xy} + \Phi_{yx}}{\Phi_{xx} - \Phi_{yy}} \right), \quad (A.4)
\]

\[
\beta = \frac{1}{2} \tan^{-1} \left( \frac{\Phi_{xy} - \Phi_{yx}}{\Phi_{xx} + \Phi_{yy}} \right). \quad (A.5)
\]

The \( \theta \) in equation Eq. 18 can be calculated by

\[
\theta = \alpha - \beta. \quad (A.6)
\]

Using (A.6), the parameterization (A.1) can be reduced to (18). It is noted that the four-quadrant (two-argument) inverse tangent function must be used in the calculation of \( \alpha \) and \( \beta \) to ensure correct parameterization in all cases.
Original publications


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