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JOINT 1-D INVERSION OF TEM AND MT RESISTIVITY DATA, COMPARISON WITH MINERAL ALTERATION AND TEMPERATURE IN DRILLHOLES – CASE STUDY: KRÝSUVÍK AREA, SW-ICELAND

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ABSTRACT

In this work, the electrical resistivity of the subsurface is studied. Geophysical exploration methods to be used are discussed in some details. The joint 1-D inversion of TEM and MT data is performed. The static shift problem of MT data is solved by inverting jointly both TEM and MT data. The resistivity of the subsurface correlates with the alteration mineralogy studied in drillholes. The NW-SE profile of the Krýsuvík area appears to host a viable geothermal system. However, a discrepancy is raised between the formation and the alteration temperatures, which would mean a cooling geothermal system underneath the studied profile.

1. INTRODUCTION

Iceland is located on the Mid-Atlantic Ridge that crosses from the Southwest through the Middle and up through the North, expressed on land by the active volcanic rift zone. Associated with the active volcanic systems are high-temperature geothermal areas, one being the Krýsuvík high-temperature geothermal field located in SW-Iceland on the Reykjanes peninsula. The rock is mainly basaltic and four formations are found in the area, classified as the Upper Tertiary Plateau-basaltic formation, Upper Pliocene and Lower Pleistocene, palagonite hyaloclastite, moberg formations and postglacial formations.

The principles and applications in geophysical exploration are divided in two categories, direct and indirect methods. The direct methods deal with determining the electrical resistivity of the earth that could be related to geothermal reservoirs while the indirect method handles the structures of the earth that are not necessary linked with a geothermal resource but provide important information on dykes and intrusions in a volcanic area.

The geophysical exploration methods for geothermal resources, used in this work, deal with measurements of the physical properties of the earth. However, emphasis is put on parameters sensitive to temperature and fluid content of the rocks.

In this work, a review of geophysical exploration methods for geothermal resources will be given. Electrical conduction in the earth material will be discussed, which explains why electromagnetic measurements are selected and carried out. An overview of the geology of the Krýsuvík high-temperature field is given. This is followed by some characteristics of the Krýsuvík high-temperature geothermal field and processing and interpretation of TEM and MT data. The static shift problem, skin depth in MT, and the joint inversion of TEM and MT data are discussed in some detail. Finally, the resistivity structure of a part of the Krýsuvík high-temperature geothermal field is given

2. GEOPHYSICAL EXPLORATION METHODS FOR GEOTHERMAL RESOURCES

Geophysical exploration methods for geothermal resources are divided into two categories according to the parameters to be measured or investigated: direct methods and structural / indirect methods. The following overview is mainly based on Hersir and Björnsson (1991).

2.1 Direct methods

Thermal methods – Temperature surveys. Thermal methods directly measure temperature and heat. No other methods have such a good correspondence with the properties of the geothermal system. There are approximately four sub-categories: temperature alone (direct interpretation, mapping); geothermal gradient (vertical variation of temperature measured in soil or drillholes); heat flow (calculated from the product of gradient and thermal conductivity); heat budgets (measuring spring flow and steam output and/or integrating heat flow). Measuring the temperature alone or the geothermal gradient is of direct significance in local geothermal work, while measuring the heat flow is more of a regional or global interest.

Heat can be exchanged in different ways such as by conduction (transfer of heat through a material by atomic vibration (often in steady-state); convection (transfer of heat by motion of mass, i.e. liquid, natural circulation of hot water); and through radiation (does not play any significant role in geothermal exploration). Conduction plays a significant role in the transfer of heat in the earth's crust. Thermal convection is usually a much more effective heat transfer mechanism than thermal conduction and is most important in geothermal systems.

Electrical resistivity of rocks – Electrical methods. In geothermal exploration, measuring the electrical resistivity of the subsurface is the most powerful prospecting method. It is directly related to the properties of interest. Electrical resistivity depends on porosity and the pore structure of the rock; the salinity of the water; temperature; water-rock interaction and alteration; amount of water (saturation); pressure; and steam content in the water.

Conductivity in minerals and solutions takes place by the movement of electrons and ions, mostly through groundwater in pores and along surface rocks and solution. For moderate temperatures in the range of 0-200°C, increasing temperature and mobility of the ions implies a decrease in resistivity. Resistivity of water as a function of temperature is expressed by Dakhnov (1962) with the following equation:

$$\rho_{\omega} = \frac{\rho_{\omega_o}}{[1+\alpha(T-T_0)]} \tag{1}$$

where ρ_{ω_o} = Resistivity of the fluid at T_o (Ωm);
α = Temperature coefficient of the resistivity; $\alpha \sim 0.023°C^{-1}$ for T_o= 23°C; and $\alpha \sim 0.025°C^{-1}$ for T_o= 0°C.

At high temperatures, a decrease in the dielectric permittivity of the water results in a decrease in the number of dissociated ions in solutions. Above 300°C, this starts to increase fluid resistivity (Quist and Marshall, 1968). In most rocks, the rock matrix itself is an insulator. The conduction is mainly in aqueous solution of common salts distributed through the pores of the rock and/or at the rock-water interface.

The most important factors controlling the conductivity are porosity, temperature, salinity and water-rock interaction. In geothermal areas, the rocks are water-saturated. Ionic conduction in the saturating fluid depends on the number and mobility of ions and the connectivity of flow paths through the rock matrix. Usually, the saturating fluid is among the dominant conductors in the rock and the degree of saturation is of great importance to the bulk resistivity. The pressure dependence is negligible compared to the temperature dependence, provided that the pressure is sufficiently high so that there is no change in phase. Pores spaces must be interconnected and filled with water in order that a rock can conduct the electricity. In all types of porosity, there are larger voids, called storage pores, and finer connecting pores. Most of the resistance to electric current flow (and fluid flow) is in the connecting pores. It has been observed for many cases that the resistivity of water-saturated rocks varies approximately as the inverse square of the porosity. This empirical relationship is called *Archie's law* (Archie, 1942). It describes how resistivity depends on the porosity function of ionic conduction in the pore fluid and dominates other conduction mechanisms in the rocks. It is valid if the resistivity of the pore fluid is of the order of 2 Ωm or less, but doubts are raised if the resistivity is much higher (Flóvenz at al., 1985):

$$\rho = \rho_w a \emptyset_t^{-n} \tag{2}$$

where ρ = Bulk resistivity (Ωm);
ρ_w = Resistivity of the pore fluid (Ωm);
$\emptyset_t$ = Porosity in proportions of total volume;
a = An empirical parameter, usually around 1; and
n = Cementing factor, an empirical parameter which varies from 1.2 for unconsolidated sediments to 3.5 for crystalline rocks, usually around 2.

Since Equation 2 only applies for relatively low resistivity values of the pore fluid, several relationships have been developed where interface conduction dominates both matrix and ionic conduction. Flóvenz at al. (1985) established the following equation relating the bulk resistivity ρ, to the fracture porosity $\emptyset_f$, the temperature T and the pore fluid resistivity ρ_{w_o}at T_o=23°C. This equation has been found applicable for the uppermost 1 km of the Icelandic basaltic crust for temperatures of up to at least 100°C (Flóvenz et al., 1985):

$$\frac{1}{\rho} = \frac{0.22}{\rho_w}\left[1-\left(1-\emptyset_f\right)^{\frac{2}{3}} + \frac{\left(1-\emptyset_f\right)^{\frac{2}{3}}}{1+\left(1-\emptyset_f\right)^{\frac{1}{3}}+\left(1-\emptyset_f\right)^{\frac{1}{3}}4.9.10^{-3}}\right] + \frac{{\emptyset_f}^{1.06}}{b} \tag{3}$$

In Icelandic high-temperature geothermal areas, the interface conductivity is caused by alteration of the rock matrix. The type of alteration minerals formed depends upon the temperature and chemical composition of the fresh rocks and saturating fluid.

Electrical exploration methods. An electrical method is either a sounding method or a profiling method, depending on what kind of resistivity structure is being investigated. The sounding method is used for mapping resistivity as a function of depth. The profiling method maps resistivity at more or less constant depth and is used to map lateral resistivity changes.

Electrical methods can be divided into the following categories and subcategories. In DC-methods, a constant current ***I*** (independent of time) is introduced into the ground through a pair of electrodes at the surface of the earth. The current creates a potential field in the earth. By measuring the electrical field ***E*** (potential difference over a short interval), the subsurface resistivity can be inferred. Schlumberger soundings, head-on profiling, dipole soundings and profiling are useful DC-methods.

In EM-methods (electromagnetics, AC-methods), alternating current (AC) at various fixed frequencies or current varying with time in a controlled way is used instead of DC-current. Many different configurations of transmitters and receivers are used. Direct contact with the ground or induction coupling characterises these methods.

In magnetotellurics (MT) and audio-magnetotellurics (AMT), fluctuations in the natural magnetic field of the earth and the induced electric field are measured. Their ratio is used to determine the apparent resistivity.

The time domain or transient electromagnetic method (TEM) is a method where a magnetic field is built up by transmitting a constant current into a loop or grounded dipole; then the current is turned off and the transient decay of the magnetic field is measured. It is used to determine the resistivity.

Measuring the electrical resistivity of the subsurface is the most powerful prospecting method in geothermal exploration. Resistivity is directly related to the properties of interest, like salinity, temperature, porosity (permeability) and alteration mineralogy. To a great extent, these parameters characterise a reservoir (Hersir and Björnsson, 1991). The specific resistivity, ρ, is defined through Ohm's law, which states that the electrical field strength ***E*** (V/m) at a point in a material is proportional to the current density ***j*** (A/m^2):

$$\boldsymbol{E} = \rho \boldsymbol{j} \tag{4}$$

The proportional constant, ρ, depends on the material and is called the (specific) resistivity and measured in Ωm. The reciprocal of resistivity is conductivity ($1/\rho = \sigma$). Resistivity can also be defined as the ratio of the potential difference ΔV (V/m) to the current I (A); across material which has a cross-sectional area of 1 m^2 and is 1 m long.

$$\rho = \frac{\Delta V}{I} \tag{5}$$

The common principle of all resistivity sounding methods is to induce an electrical current into the earth and monitor signals, normally at the surface, generated by the current distribution. In conventional direct current soundings such as Schlumberger soundings, this is done by injecting current into the ground through electrodes at the surface; the signal measured is the electric field (the potential difference over a short distance) generated at the surface. In magnetotellurics (MT), the current in the ground is induced by time variations in the earth's magnetic field and the signal measured is the electric field at the surface. In transient electromagnetics (TEM), the current is also induced by a time varying magnetic field but, in this case, the source is of a controlled magnitude, generated by the current in a loop or grounded dipole and the monitored signal is the decaying magnetic field at the surface.

The processing of magnetotelluric data involves concepts from electromagnetic theory, time series analysis and linear systems for studying natural electric and magnetic field variations recorded at the earth's surface. The electromagnetic field relationships lead to either scalar transfer impedance which couples an electric component to an orthogonal magnetic component at the surface of a plane-layered earth, or tensor transfer impedance which couples each electrical component to both magnetic components in the vicinity of a lateral inhomogeneity.

A number of time series spectral analysis methods can be used for estimating the complex spectral coefficients of the various field quantities. These, in turn, are used for estimating the nature of the transfer function or tensor impedance. For two dimensional situations, the tensor impedance can be rotated to determine the principle direction of the electrical structure. In general for real data, estimates of the apparent resistivity are more stable when calculated from the tensor elements rather than simple orthogonal field ratios (Cagniard estimates), even when the fields are measured in the principal coordinates (Cagniard, 1953).

The processing and inversion of the Transient Electromagnetic data, used in this work, involves a number of software products through which different operations are performed such as the TemX and TEMTD programs. The TemX program is written in ANSI-C and uses the XForms library (a free software available online) for interactive graphics. The program comes in two versions: TemX which reads and processes central-loop TEM data recorded by a PROTEM receiver from GEONICS Ltd, and TemXZ which reads and processes central-loop TEM data recorded by a GDP-32 receiver from Zonge Engineering & Research Organisation, Inc.

In this work, the TEMTD program performs 1-D inversion with horizontally layered earth models of Central-loop Transient Electromagnetic (TEM) and Magnetotelluric data. It can be used to invert only TEM or MT data, or to do a joint inversion, in which case it determines the best static shift parameter for the MT data. For TEM data, the program assumes that the source loop is a square loop and that the receiver coil/loop is at the centre of the source loop. The current wave form is assumed to be a half-duty bipolar semi-square wave (equal current-on and current-off) segments, with exponential current turn-on and linear current turn-off. For MT data, the program assumes standard EDI files for impedance and/or apparent resistivity and phase data. The program is written in ANSI-C and runs under the UNIX/LINUX operating systems.

2.2 Structural/indirect methods

Magnetic surveys. Together with gravity and seismic refraction measurements, magnetic methods in geothermal exploration are applied in mapping geological structures (structural method). The most important applications are location and depth of concealed intrusives, tracing dykes and faults, locating buried lava, depth to basement, locating hydrothermally-altered areas, and paleomagnetism.

Gravity surveys. The gravity force between two masses, m_1 and m_2, at a distance r apart, is given by Newton's law of gravitation (m_1 is the apple and m_2 is the earth):

$$F = G\frac{m_1 m_2}{r^2} \tag{6}$$

where G is the universal gravitation constant, $G = 6.670 \times 10^{-11}$ Nm^2/kg^2.

Different types of rocks in the crust and mantle have different densities, hence different gravitational forces. In geothermal exploration, the gravity method is used to detect geological formations with different densities. Therefore, it is a typical structural method.

Gravity variations are measured with a gravimeter. These are very sensitive mechanical instruments which measure the change in acceleration (g) at one place relative to another reference place (relative measurement). The unity for acceleration is m/s^2. A gravity unit (g.u.) is equal to 10^{-6} m/s^2. On the other hand, geophysicists usually use the unit, Gal. One g.u. is equal to 0.1 mgal. The sensitivity of gravimeters is about 0.005 mgal. Absolute gravimeters are now developed. They are already used in the field to measure gravity at selected base stations. Their accuracy is similar to the accuracy of conventional gravimeters.

In order to obtain information about the subsurface density from gravity measurements, it is necessary to make several corrections to the measured gravity values before they can be represented in terms of geological structures. The final corrected value for the gravity anomaly is called the Bouguer anomaly, Δg_B. It can be expressed as:

$$\Delta g_B = g_M + C_{FA} - C_B + C_T - g_N \qquad (7)$$

Where g_M = Measured gravity corrected for tidal effects (attraction of the moon and the sun) and drift in the gravimeter.

C_{FA} = Elevation correction or free-air correction; it can be written as: $C_{FA} = 0.3086H$, where H is the height of the station in m above sea-level.

C_B, = Correction for excess mass material between the station and sea-level, Bouguer correction. It can be written as $C_B = 0.04191\rho H$, where ρ is the density (g/cm^3). The density must be well known in order to be able to calculate the Bouguer correction.

C_T = Topographical correction, correction for local terrain variations near the station. It includes distortion by hills and valleys. It can be written as $C_T = \rho GL$, where GL is a parameter characterising the topography around the station. Topographic corrections always decrease measured gravity.

g_N = Normal reference gravity, according to an international formula. It takes into account the latitude of the station, i.e. the earth's shape (ellipsoid), and centrifugal acceleration.

3. SELECTED EXPLORATION METHODS FOR THE KRÝSUVÍK AREA

3.1 Magnetotellurics (MT)

The magnetotelluric method was proposed by Tikhonov, Kato and Kikuchi, and Rikitake in the 1950s and later by Cagniard (1953), although the principle was recognized much earlier. In its forms (MT, AMT and CSAMT), the method is now used in a broad variety of applications, from very deep and large scale studies of the crust and upper mantle, through a wide range of exploration applications, geothermal and petroleum exploration, to the shallow problems of epithermal gold, 'deep' base metals and groundwater. In many of these applications, MT provides a kind of information which can be obtained in no other way.

The magnetotelluric (MT) method uses time variations of the Earth's natural electromagnetic field to determine the electrical conductivity of the Earth. The solar system ejects charged particles called *solar wind*. These particles are mainly protons and electrons interacting with the Earth's magnetosphere and producing electromagnetic fields. The charges in the ionosphere cause displacement currents and conduction currents to produce lightning and thunderstorms. The flow of charged particles in ionospheric zones or magnetospheric layers, as hydromagnetic waves or plasma, generates electrical and magnetic fields that propagate towards the Earth (Vozoff, 1991). The strength of the electromagnetic fields depends on time and position (in latitude) due to the relationship between the Sun and the Earth in rotation and the ejection of the Earth's magnetic field. The magnetic field of the Earth varies from the equator to the poles. The interactions of solar particles and the Earth's magnetic fields result in electromagnetic waves of various frequencies.

The geomagnetic fluctuations range between periods of 10^{-3} s and 10^{5} s (or between frequencies of 10^{3} Hz and 10^{-5} Hz) depending on their origin (Vozoff, 1991). When reaching the ground, the electromagnetic waves penetrate to great depths and interact with subsurface layers that produce secondary electromagnetic fields that are measured by MT instruments.

Between 0.5-5 Hz lies the dead band at which natural EM fluctuation has a low intensity. MT measurements in this frequency range usually suffer from poor data quality. Magnetotelluric measurements presented throughout this work have a band width of 10^{-2}-10^{3} s. The period ranges between 1 and 10^{3} s is due to interactions between solar particles, wind, the Earth's magnetosphere and the ionosphere.

The propagation of electromagnetic fields can be described by a set of four relations, called the Maxwell's equations. Given a polarisable and magnetisable medium containing no electric and magnetic sources, the following relations hold at all times for all frequencies:

$$\nabla \cdot \boldsymbol{B} = 0 \tag{8}$$

$$\nabla \cdot \boldsymbol{D} = \varrho \tag{9}$$

$$\nabla \times \boldsymbol{E} = -\frac{\partial \boldsymbol{B}}{\partial t} \tag{10}$$

$$\nabla \times \boldsymbol{H} = \boldsymbol{j} + \frac{\partial \boldsymbol{D}}{\partial t} \tag{11}$$

Equation 8 states that magnetic fields, defined by the magnetic induction **B** (in T) are always source-free, and no free magnetic poles exist. The electric displacement $\boldsymbol{D}$ (C/m^2) is solely due to an electric charge density ϱ (C/m^3). Faraday's law (Figure 1) in Equation 10 shows the coupling of an induced electric field $\boldsymbol{E}$ (V/m) in a closed loop due to a time varying magnetic field $\boldsymbol{B}$ along the axis of the induced electric field.

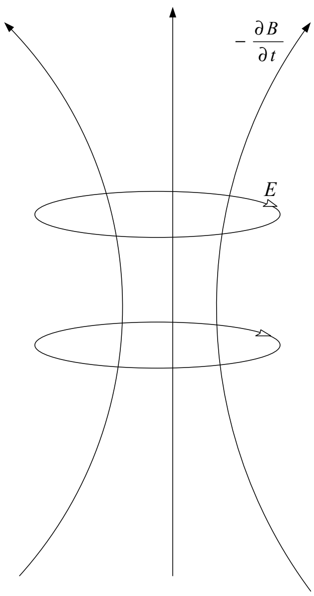

FIGURE 1: Faraday's law

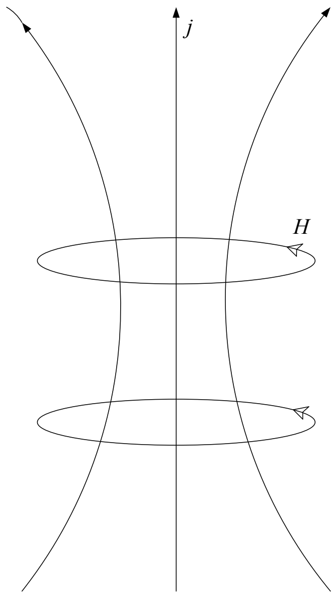

FIGURE 2: Ampere's law

The magnetic intensity $\boldsymbol{H}$ (A/m) is caused by electric current densities $\boldsymbol{j}$ (A/m^2) and time varying electric displacements (Figure 2). This is expressed mathematically in Equation 11 which is known as Ampere's law. Furthermore, currents in the presence of an electric field can only flow if the media has conductivity σ (S/m), then

$$\boldsymbol{j} = \sigma \boldsymbol{E} \tag{12}$$

In MT, magnetic and electrical fields are measured in the frequency band 10^{-4} to 10 kHz, with high frequencies (>1 Hz) coming from thunderstorm activities in the equatorial belt while low frequencies (<1 Hz) occur due to the interaction between the solar wind and the earth´s magnetic field (magnetosphere) and ionosphere (Figure 3).

Natural electric and magnetic field strength are simultaneously recorded in two orthogonal, horizontal directions as a function of time. For geothermal exploration, the MT method targets deep brine reservoirs and hot (or partially molten) rocks that act as the heat source for a geothermal system (Ward and Wannamaker, 1983).

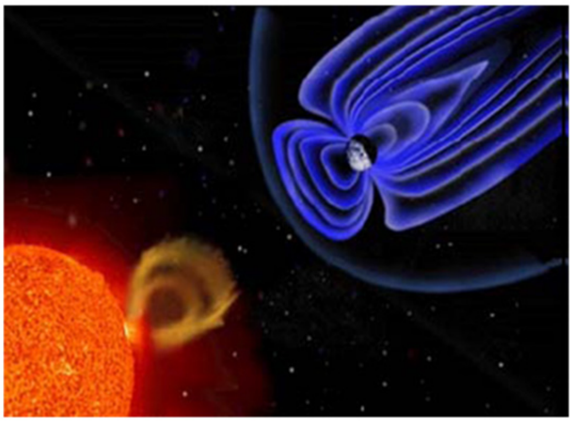

FIGURE 3: Interaction of solar wind with the magnetosphere. Source region for low frequency (< 1Hz) natural EM fields (taken from Encyclopaedia Britannica, 2010)

Static shift in magnetotellurics. The static shift is due to inhomogeneities in the near-surface geological structures of the area. Electric currents shift towards highly conductive bodies/structure while the magnetic field tends to be stable. Sternberg et al. (1988) described a method for correcting static shift in the magnetotelluric by incorporating the electromagnetic sounding from the same vicinity.

The TEM sounding should be very close to or at least within 100-300 m distance from the MT site. TEM has no static shift effects due to near-surface inhomogeneities because it does not monitor the electrical field. Here, the static shift is corrected through a joint inversion of TEM and MT soundings, using the inversion program TEMTD developed by Árnason (2006a).

Electromagnetic induction in a homogeneous earth. For an electromagnetic wave of frequency ω the ratio of electric to magnetic field intensity is a characteristic measurement of the electromagnetic properties, often called the characteristic impedance:

$$Z_{xy} = \frac{E_x}{H_y} = \frac{i\omega\mu_0}{k} \tag{13}$$

$$Z_{yx} = \frac{E_y}{H_x} = -\frac{i\omega\mu_0}{k} \tag{14}$$

where Z_{xy} and Z_{yx} = The characteristic impedance in x and y directions;
μ_0 = The magnetic permeability in vacuum (H/m);
E_x and E_y = Electric field intensity (V/m) in x, y directions;
H_x and H_y = The magnetic field intensity (A/m) in x, y directions; and
k = The wave propagation constant.

If the earth is homogeneous and isotropic then the true resistivity of the earth is related to the characteristic impedance through the following relationship (Hermance, 1973):

$$\rho = \frac{1}{\mu\omega}\left|Z_{xy}\right|^2 = \frac{1}{\mu\omega}\left|Z_{yx}\right|^2 \tag{15}$$

For a non-homogeneous earth the apparent resistivity (ρ_a) can be defined as if the earth were homogeneous using the same formula. In practical units for homogeneous earth, the apparent resistivity, ρ_a can be written as:

$$\rho_a = 0.2T|Z|^2 = 0.2T\left|\frac{E_x}{B_y}\right|^2 \tag{16}$$

where E = The electrical field (mV/km); and
B = The magnetic field (nT).

For a non-homogeneous earth, the apparent resistivity (ρ_a) and the phase (θ_a) are given by:

$$\rho_a = 0.2T|Z_0|^2 \tag{17}$$

and

$$\theta_a = \arg(Z_0) \neq \frac{\pi}{4} \tag{18}$$

Skin depth. The skin depth, δ, is the depth where the electromagnetic field has been reduced to e^{-1} of its original value at the surface. For oscillating electromagnetic fields, that is $\boldsymbol{H}$, $\boldsymbol{E} \sim e^{i\omega t}$, a vertically incidental plane wave, the skin depth is a scale length for the time varying field, or an estimate of how deep such a wave penetrates into the earth, and is given by:

$$\delta = \frac{1}{Real(k)} = \frac{1}{Real\left(\sqrt{i\omega\mu\sigma}\right)} = \sqrt{\frac{2}{\omega\mu\sigma}} = \sqrt{\frac{2T\rho}{2\pi\ 4\pi\ 10^{-7}}} = \frac{10^3}{\pi}\sqrt{\frac{20}{8}}\sqrt{T\rho} \tag{19}$$

or

$$\delta = 500\sqrt{T\rho} \tag{20}$$

where δ = Skin depth (m);
T = Period; and
ρ = Resistivity (Ωm).

Formula 20 shows that in the magnetotelluric method for constant resistivity, the depth of probing increases when the period T increases and decreases when the period decreases. This was done before the survey by playing on the frequency at which one would expect the soundings to be recorded.

3.2 Transient-electromagnetic method

In the transient electromagnetic method (TEM), an electrical current is induced in the ground and a magnetic field created is measured at the surface, from which the resistance of the underground rocks is determined. The current in the ground is generated by a time varying magnetic field. Yet, unlike MT-soundings, the magnetic field is not the randomly varying natural field but a field of controlled magnitude generated by a source loop. A loop of wire is placed on the ground and a constant magnetic field of known strength is built up by transmitting a constant current into the loop. The current is then abruptly turned off (Figure 4). The decaying magnetic field induces electrical current in the ground. The current distribution in the ground induces a secondary magnetic field decaying with time. The decay rate of the secondary magnetic field is monitored by measuring the voltage induced in a receiver coil (or a small loop) at the centre of the transmitter loop.

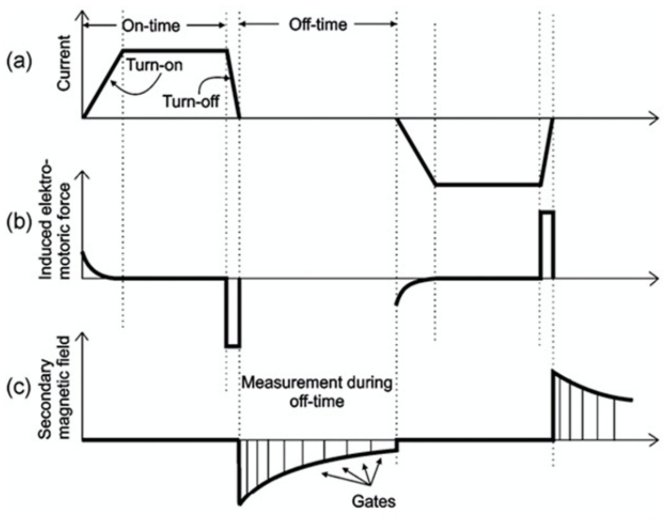

FIGURE 4: Basic nomenclature and principles of the TEM method; (a) shows the current in the transmitter loop; (b) is the induced electromotive force in the ground; and (c) is the secondary magnetic field measured in the receiver coil. For the graphs of the induced electromotive force and the secondary magnetic field, it is assumed that the receiver coil is located in the centre of the transmitter loop (Christensen et al., 2006)

The current distribution and the decay rate of the secondary magnetic field depend on the resistivity structure of the earth. The decay rate, recorded as a function of time after the current in the transmitter loop is turned off can, therefore, be interpreted in terms of the subsurface resistivity structure. The depth of penetration in the central loop TEM-sounding is dependent on how long the induction in the receiver coil can be traced in time before it is drowned in noise.

The apparent resistivity, $\rho_a(t)$, (for a homogeneous half-space) in terms of induced voltage at late times after the source current is turned off, is given by (Árnason, 1989):

$$\rho_a(t) = \frac{\mu_0}{4\pi}\left\{\frac{2\mu_0 I_0 A_r n_r A_s n_s}{5t^{\frac{5}{2}}V(t,r)}\right\}^{\frac{2}{3}} \tag{21}$$

where A_r = Cross-sectional area of the receiver coil (m^2);
n_r = Number of windings in the receiver coil;
μ_0 = Magnetic permeability in vacuum (H/m);
A_s = Cross-sectional area of the transmitter loop (m^2);
n_s = Number of windings in the transmitter loop;
$V(t)$ = Voltage as a function of time;
$\rho_a(t)$ = Apparent resistivity as a function of time; and
t = Time elapsed, after the transmitter current was turned off.

4. GEOPHYSICAL PROSPECTING OF THE KRÝSUVÍK GEOTHERMAL FIELD

4.1 Methodology

The rocks within the earth´s subsurface have many physical properties that vary from place to place. The common ones include electrical conductivity, sound velocity, magnetism and density. As mentioned earlier in this work, geophysics is the science that studies the variations of any of these properties across the surface of the earth. If the measured property over an area is different from that of the surrounding regions, it is referred to as anomalous. In the present work, transient electromagnetic (TEM) and magnetotelluric (MT) resistivity soundings were carried out at the same locations in the Krýsuvík geothermal field.

4.2 Acquisition of data and instrumentation

A total of 96 MT and more than 200 TEM sites have been acquired from the Krýsuvík high-temperature geothermal field and the surrounding area since 1989 and 1-D inverted (Eysteinsson, 1999; 2001; Hersir et al., 2010). These data were used in this report with the permission of HS Orka to study the subsurface resistivity distribution of the Krýsuvík high-temperature geothermal field.

In this work, the data were acquired using different instruments according to the specific methods. For TEM, a PROTEM receiver from GEONICS Ltd. was used. TemX software program developed at ISOR (Árnason, 2006b), reads and processes central-loop TEM raw data recorded from that instrument. The transient signal (Figure 5) was recorded at logarithmically spaced sampling gates after the current was turned off. For each frequency, several repeated transients were stacked and stored in a memory cache inside the GEONICS data logger and then transferred to a personal computer for processing.

For MT: the data was acquired using a 5-channel MT data acquisition system (MTU-5) from Phoenix Geophysics . The instrumentation consisted of a data recorder, induction coils, non-polarizing electrodes, Global Positioning System (GPS), 12 V battery, flash memory for data recording, and telluric and magnetic cables (Figure 6).

MT data were acquired for about 20 hrs, giving the frequency range from 320 Hz and often up to few thousand seconds, which generally ensures investigation depths down to several tens of km. During MT data acquisition, one 5-component MT station was installed far away from the field of focus and maintained as a remote reference station which was later used during data processing to reduce the effects of local cultural noise.

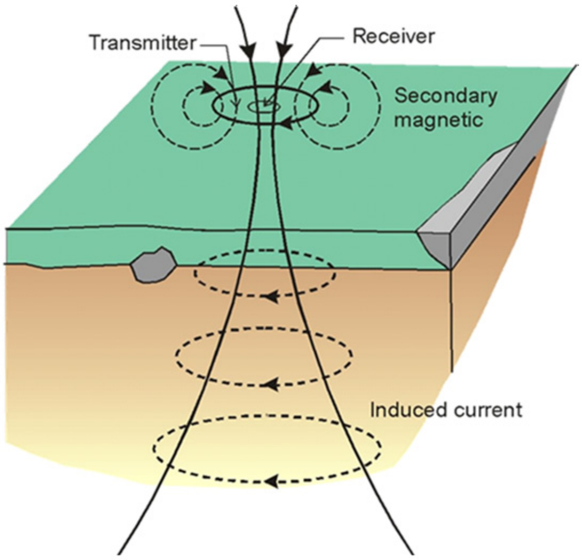

FIGURE 5: Transient current flow in the ground (from Hersir and Björnsson, 1991)

4.3 Characteristics of the Krýsuvík high-temperature geothermal field

Geological setting. The geology of Iceland is governed by the tectonic and volcanic evolution through the last 16-18 million years. The major geothermal activities occur within and/or along the Mid-Atlantic Ridge that crosses the island from the southwest through the centre and up to the northeast of Iceland. The Krýsuvík high-temperature field is one of the areas geologically affected by intense volcanic activity. It is located in SW-Iceland on the Reykjanes Peninsula.

Two majors hyaloclastite ridges, trending NE-SW dominate the Krýsuvík area. Faulting in the area is also mainly NE-SW. The rocks are of basaltic origin and from different periods, some having been erupted under glacier. Different geological formations are hyaloclastite, erupted during the last few interglacial periods. The rocks are basaltic and occur as: olivine tholeiites, usually fine-grained to medium-grained; dolerites made of coarse-grained basalt; glassy basalts, referred to as pillow lava. There are craters in several places. Geothermal manifestations are located near the main structures and often have a NE-SW trending direction.

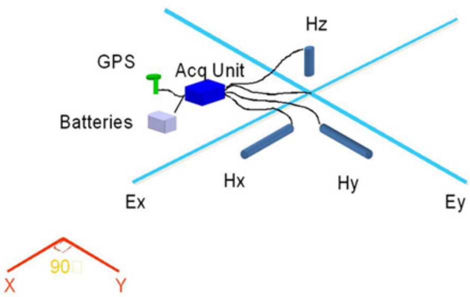

FIGURE 6: MT field set up with the five channels (from Phoenix Geophysics, Ltd., 2005)

Hydrothermal alteration at the surface was probably caused mostly by shallow intrusive bodies such as dykes, and major faults. In each area, the alteration grade varies slightly from the peripherals to the centre of the geothermal manifestations. The alteration minerals are mainly clay and mixed-clay minerals, such as: kaolin, smectite, silica, calcite, iron oxides, zeolite and chlorite (Mortensen et al., 2006).

Resistivity structure of high-temperature geothermal fields in Iceland. All high-temperature geothermal fields in Iceland, and in general where the host rocks are volcanic not sedimentary, have similar resistivity structures. They are characterised by some kind of a convex structure. At a certain depth, a low-resistivity (conductive) cap or zone domes up (the outer margin of the reservoir) and is underlain by higher resistivity, a resistive core (see Figure 7). Not so many years ago, most scientists found it hard to believe that resistivity would increase at this depth since temperature obviously

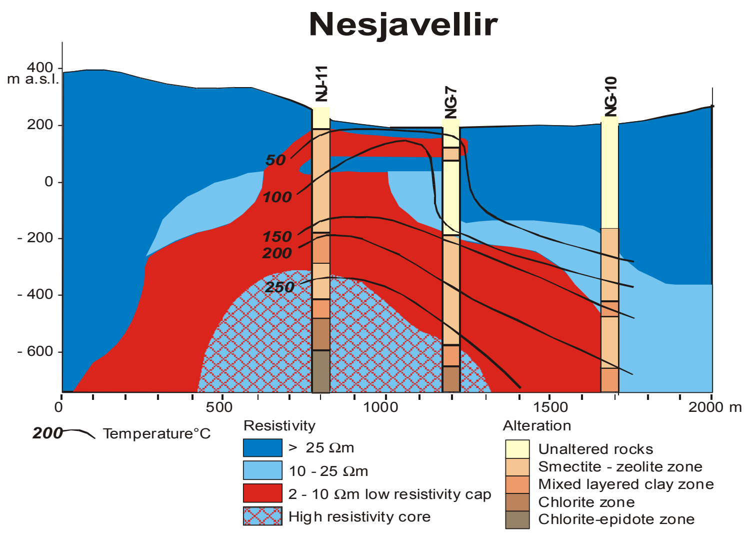

FIGURE 7: Resistivity cross-section from Nesjavellir geothermal field, SW-Iceland, alteration zoning and temperature (Árnason et al., 1987)

increased with depth. This was a kind of puzzle and different theories were put forward. This structure can be seen in a joint 2-D interpretation of different datasets at Nesjavellir, SW-Iceland showing a detailed picture of the resistivity structure in the uppermost 1 km which could be compared to data from nearby boreholes (Figure 7).

As Figure 7 shows, the subsurface resistivity structure in high-temperature geothermal fields reflects the hydrothermal alteration. The primary minerals in the host rock matrix are transformed into different minerals because of water-rock interaction and chemical transport by geothermal fluids. Alteration minerals depend on the type of primary minerals and the chemical composition of the geothermal fluid but mainly on temperature, although porosity and permeability also control the intensity of the alteration. If the alteration and temperature are in equilibrium, the subsurface structure reflects not only the alteration but also shows which temperatures to expect.

5. PROCESSING AND INTERPRETATION OF DATA FROM THE KRÝSUVÍK GEOTHERMAL FIELD

The present work was designed to find the connection between the resistivity structure of the subsurface underneath the studied area and the main alteration minerals in the drillholes. Figure 8 features a map locating the MT and TEM soundings with the associated drillholes on a studied profile.

5.1 MT data processing

The processing of magnetotelluric data involves concepts from electromagnetic theory, time series analysis and linear system theory for reducing natural electric and magnetic field variations recorded at the earth's surface to form suitable forms for studying the electrical properties of the earth's interior.

Spectral analysis of time series. Interpretation of magnetotelluric data is usually done in the frequency domain. Spectral analysis of the raw data is an important aspect of data processing. This involves creating a time series x_j; by sampling a signal $x(t)$; at equal intervals of time Δt from $j = 1$ to N, where $T_0 = (N-1)\Delta t$ is the duration of the signal.

One intuitively feels that for the sampled data to represent the original signal adequately, the signal should be smooth over the sampling interval Δt in a more rigorous sense. An absolute requirement is that the signal to be sampled should have negligible spectral energy at periods shorter than twice the sampling interval.

There are a number of techniques for transforming the time series into spectral information in the frequency domain. Although, in principle, all of these methods are equivalent, since they result in the

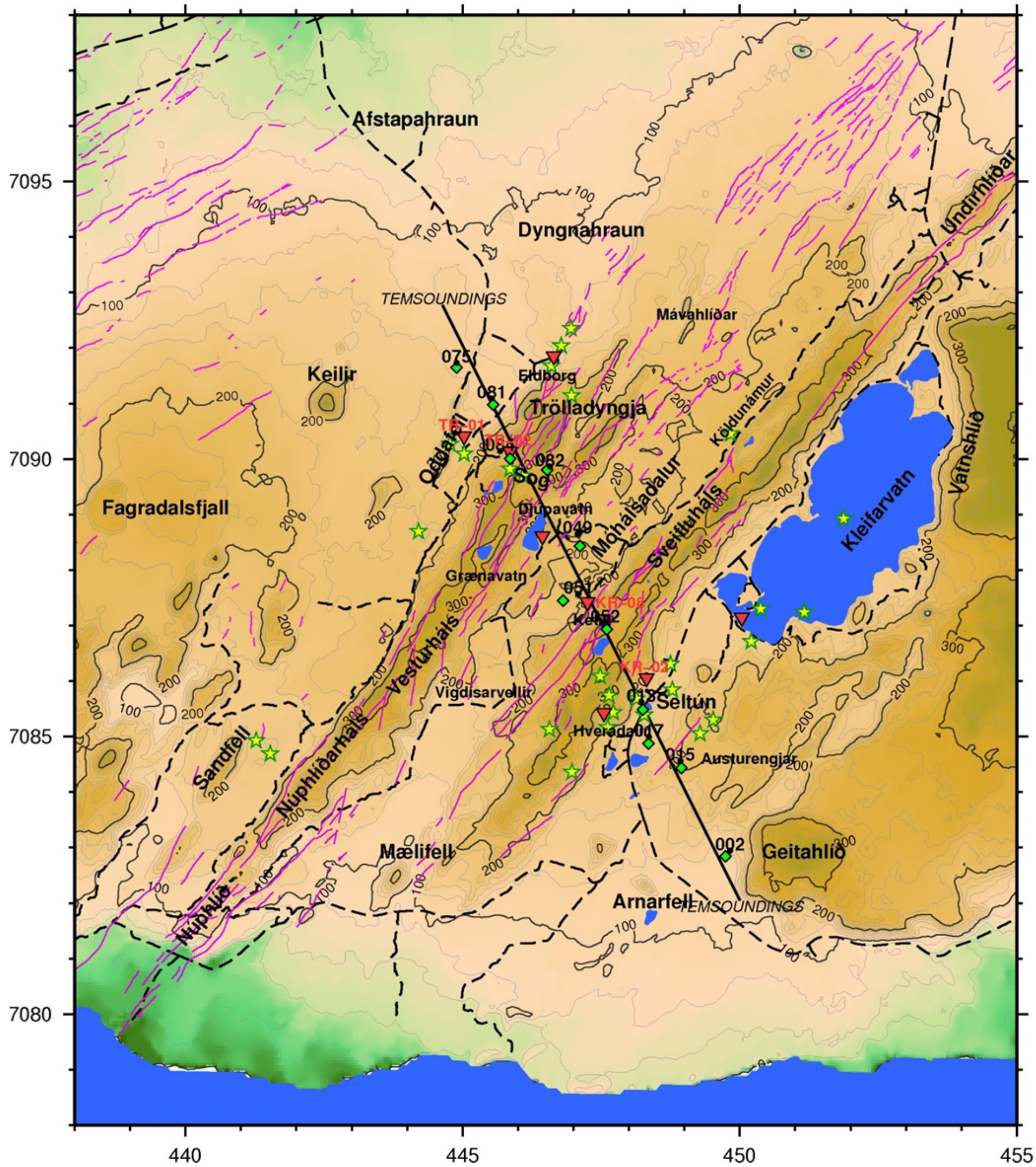

FIGURE 8: Location of the resistivity profile showing: TEM stations (diamonds), MT stations (dots), geothermal wells (inverted triangles), surface manifestations (stars), faults and fissures

Fourier components of the record, in practice there are important differences that depend on the character of the signal and the underlying assumptions regarding the nature of the data.

Fourier harmonic analysis. The fundamental assumption in Fourier harmonic analysis is that the times series of length T_o is cyclic or periodic with fundamental period T_o, such that $x(t) = x(t+ T_o)$. This assumption places strict constraints on the spacing between harmonics as well as implying that the record section can be optimally approximated over its length T_o by a finite number of harmonics. This can be seen in the following way. Over an interval of time $\left(-\frac{T_0}{2},\frac{T_0}{2}\right)$ a function $x(t)$ can be expressed as the sum of a Fourier series:

$$x(t) = \sum_{k=-\infty}^{k=+\infty} X_k \, e^{ik\omega_0 t}, \; \omega_0 = \frac{2\pi}{T_0} \tag{21}$$

where X_k is given by:

$$X_k = \frac{1}{T_0} \int_{-\frac{T_0}{2}}^{\frac{T_0}{2}} x(t) e^{-ik\omega_0 t} \, dt \tag{22}$$

The amplitude spectral lines X_k are spaced at intervals k/T_o along the frequency axis, which are harmonics of the record length T_o. Since we have required that the signal *x(t)* be band-limited to frequencies less than the folding-frequency ($1/2\Delta t$), and since T_o = (N-1) Δt, then for the maximum harmonics, as well as a term for k = 0, the total number of harmonics needed is n, the same as our number of data points (Simpson and Bahr, 2005).

Time-series data downloaded from a flash memory of the MT equipment was processed by the SSMT2000 program, provided by the equipment manufacturer, Phoenix Geophysics of Canada. First the parameter file was edited to reflect the data acquisition setup and then the resulting time-series data were Fourier transformed to the frequency domain. The time series data were converted into the frequency domain by using a cascade decimation method. From the Fourier transform band, averaged cross-powers and auto-powers were calculated using the robust processing method. The cross-powers were then graphically edited by the MT-Editor program to remove the noisy data points and evaluate ”smooth” curves for both phase and apparent resistivity. The final cross- and auto-powers, as well as all relevant MT parameters calculated from them (impedances, apparent resistivity, phase coherences, strike directions, etc.) were stored in industry-standard file formats (EDI files) ready for joint inversion.

5.2 TEM data processing

TemX software is the program used to process TEM raw data (Árnason, 2006b). The program comes in two versions, TEMX and TEMXZ which read and process central loop TEM raw data recorded by a PROTEM receiver form GEONICS Ltd., and by a GDP-32 receiver from Zonge Engineering, respectively. The program reads files with data sets from a sounding. Then it performs a normalization of voltages with respect to transmitted current. Afterwards, the data are displayed graphically to allow users to omit outliers and calculate averages over datasets and late apparent resistivity. Appendix I shows the TEM data and its 1-D interpretation.

Correction for the static shift in magnetotellurics using transient electromagnetic soundings. The static shift is mainly caused by the accumulation of charges at resistivity boundaries causing the electrical field not to be continuous across this boundary. The static shift is expressed by scaling of the apparent resistivity by an unknown factor (shifted on log scale). This shift is independent of frequency. The static shift can be a big problem in volcanic areas where resistivity variations close to the surface are often extreme. These parallel shifts in apparent resistivity curves lead to large errors in inverted data. For instance, a shift downwards by S=0.1 will, in interpretation, result in ten times too low resistivity values and about three times too small depths to the resistivity boundary (Árnason, 2008).

The TEMTD program was used in this work to perform 1-D inversion with horizontally layered models of central-loop transient electromagnetic (TEM) and magnetotelluric data. It can be used to invert only TEM or MT data and also for joint inversion of both TEM and MT data, in which case it determines the best static shift parameter for MT data; then the static shift problem can be solved (Árnason et al., 2010).

5.3 Description of the processing methods for both MT and TEM data and joint interpretation

TEM and MT data were processed to study the subsurface resistivity of the studied area. The TEM data were acquired using Geonics equipment at a frequency of 2.5 and 25 Hz. The TEM raw data were first processed using TemX software that runs on a computer with a Linux operating system. The program was used to remove bad data. The data from TemX were plotted as a smooth curve on a bilogarithmic scale and modelled using the TEMTD program (Occam inversion). The TEMTD program generated and estimated automatically the characteristics of the misfit curve compared to measured data. The fits were assumed reasonable at χ^2 less than 1.

The magnetotelluric (MT) data were acquired using MTU-5 magnetotelluric instruments made by Phoenix Canada. Four sets of instruments (1857, 1926, 1925 and 1924) were deployed in the field. One instrument was used as a remote reference station, placed far away from the field of focus (10-20 km). The three others ran simultaneously with the remote reference station. The five components H_x, H_y, H_z, E_x and E_y were measured by two sets of instruments and one measured only the two electrical components E_x and E_y. The magnetic field tended to be stable in the same area. Using H_x and H_y from the nearest station for the two component site, the impedances could be computed.

The raw MT data were processed using the SSMT2000 program which transforms data from the time domain to the frequency domain. The inputs were reviewed to meet the time consistency between the remote station and the field instruments. Computing the Fourier coefficients and processing the MT data from the time domain to the frequency domain were performed, after which the MT signals were edited based on the remote reference station to reduce or minimise, respectively, the upward or downward bias effects caused by the electric and magnetic fields. The impedance was computed from the electrical and magnetic fields spectra whereas the noise was minimized by auto and cross power spectral analysis. The computed frequency data were edited using the MT-Editor program from Phoenix to remove bad data points and produce a smooth curve. Then, the processed output in the form of MPK files were used to make standard industrial data, EDI files, shown in Appendix II. The processed MT data are EDI files. They were inverted jointly with the TEM data using the TEMTD program. Appendix III shows the results from the joint MT and TEM inversion.

Figure 9 shows the NW-SE resistivity cross-section probing down to 800 m, using TEM data only. Close to the surface it reveals high resistivities of some hundreds of Ωm, followed by lower resistivities of 10-32 Ωm. Then a conductive cap with a resistivity below 10 Ωm, and even close to 1 Ωm, is seen. Beneath the conductive cap there is a resistive core.

5.4 Joint 1-D inversion of TEM and MT data

The joint 1-D inversion of TEM and MT soundings was designed to solve the static shift problem in MT data in the volcanic environment of the Krýsuvík area. Care was taken in measuring TEM soundings at the same places as the MT soundings. The TEMTD software used here offers several options of inversion of TEM and MT data, separately or jointly. For our case, we inverted them jointly.

In joint 1-D inversion of TEM and MT data, one more parameter is inverted for in addition to the layered model resistivity and thickness parameters, namely a static shift multiplier by which the apparent resistivity has to be divided so that both the TEM and MT data can be fitted with the same model. The

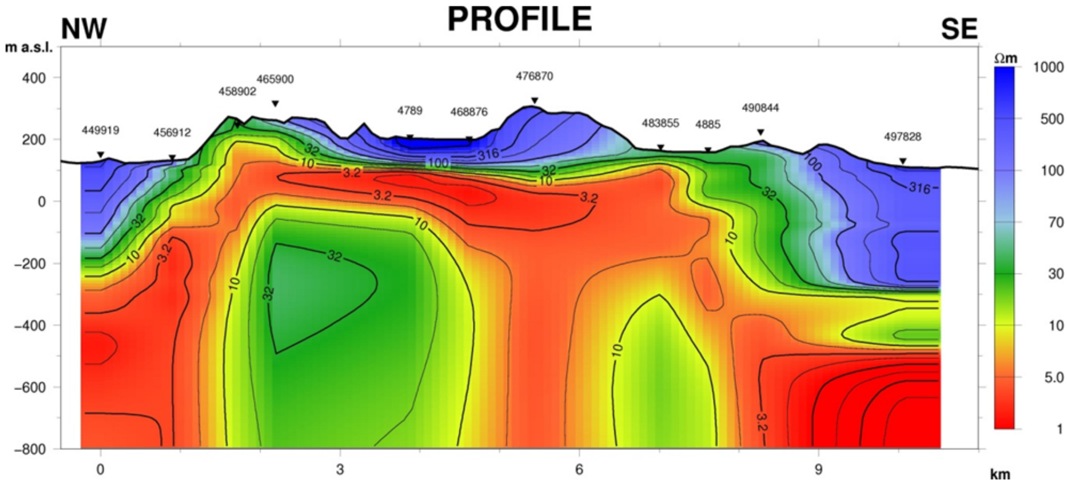

FIGURE 9: The NW-SE resistivity cross-section in the Krýsuvík area probing down to 800 m, using TEM data only

program can do both standard layered inversion (inverting resistivity values and layered thicknesses) and Occam inversion with exponentially increasing layer thicknesses with depth. A joint 1-D Occam inversion was performed for the rotationally invariant determinant apparent resistivity and phase of the Krýsuvík soundings and the associated TEM soundings.

Figure 10 shows the NW-SE resistivity cross-section obtained from joint 1-D inversion of TEM and MT data down to 2500 m depth. It reveals high resistivities in the upper most part; underneath which relatively low resistivity is found and then a very low resistive cap follows. From a depth of 1000 m and down to 2500 m, the highly resistive core is found. Towards southeast the resistivity within the core decreases considerably.

Figure 11 also shows the NW-SE resistivity cross-section obtained from joint 1-D inversion of TEM and MT data, but now it reaches down to 5000 m. The profile shows the low-resistivity cap at 0-1200 m depth b.s.l. and a highly resistive core below it.

Finally, Figure 12 shows once again the NW-SE resistivity cross-section obtained from joint 1-D inversion of TEM and MT data, and this time probing down to 20,000 m. The cross-section shows that the resistivity in the profile is quite high at great depths, as seen in the interpreted resistivities between 8,000 and 20,000 m.

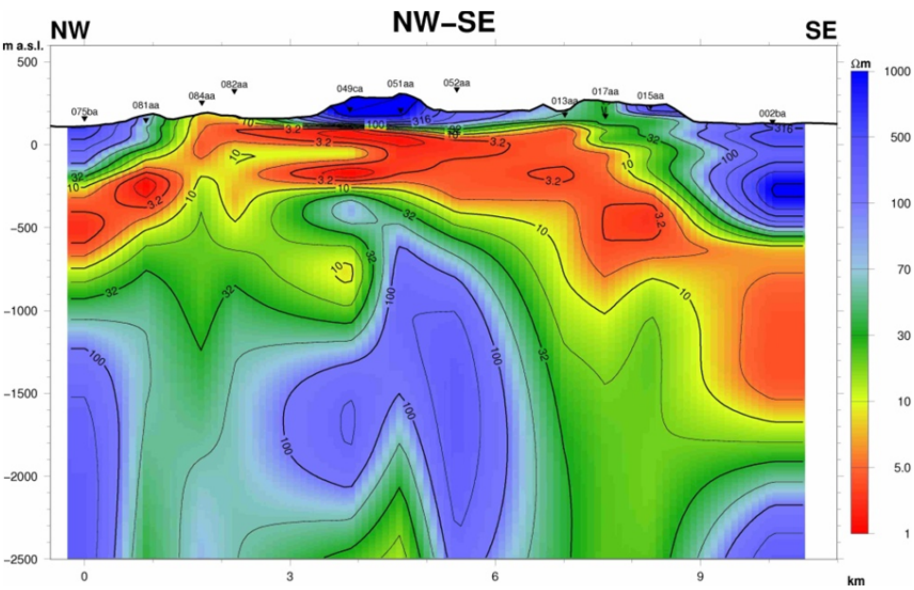

FIGURE 10: The NW-SE resistivity cross-section in the Krýsuvík area probing down to 2500 m depth, based on 1-D joint inversion of the TEM and MT data

6. COMPARISON OF RESISTIVITY RESULTS WITH MINERAL ALTERATION AND TEMPERATURE IN DRILLHOLES IN THE KRÝSUVÍK AREA

The interpretation of the results from the MT and TEM geophysical methods is expected to suggest the presence (or absence) of a heat source, reservoir characteristics, up-flow and discharge zones of a geothermal prospect (Árnason, 1990). Data from the MT and TEM soundings is often displayed as resistivity versus time, frequency or depth curve.

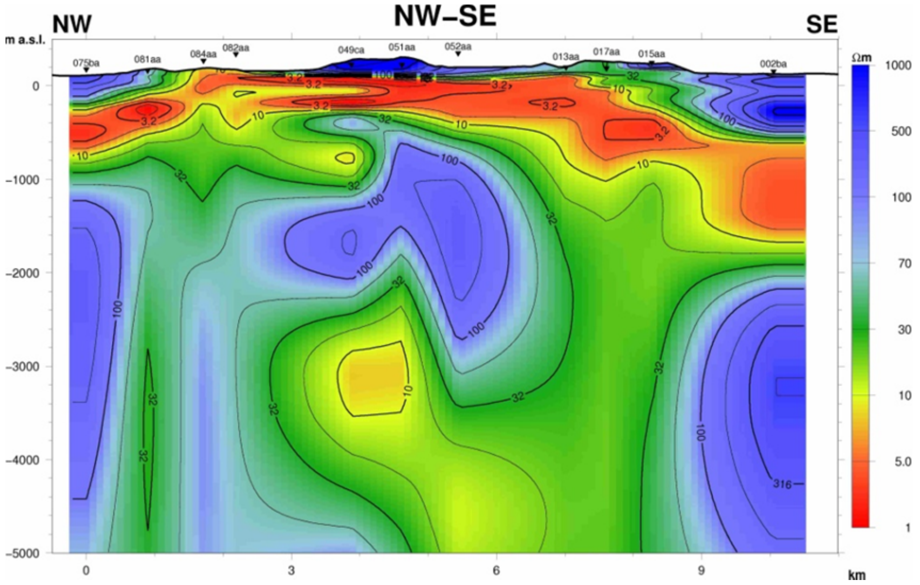

FIGURE 11: The NW-SE resistivity cross-section in Krýsuvík, probing down to 5000 m depth, based on 1-D joint inversion of TEM and MT data

The low-resistivity section of the curve suggests a conductive region. Resistivity anomalies can be related to temperature and alteration in a geothermal field (Figure 13). Most of the time a good correlation is found between resistivity and alteration mineralogy. The resistivity is relatively high in cold unaltered rocks (dry lava fields) outside the reservoir and slightly lower at temperatures of 50-100°C; alteration intensity is normally low at this temperature range.

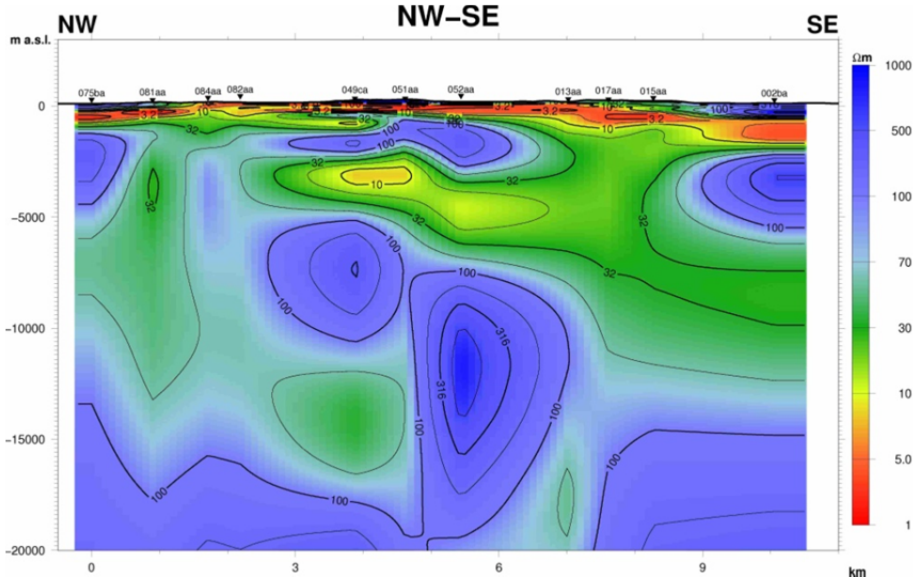

FIGURE 12: The NW-SE resistivity cross-section in Krýsuvík probing down to 20,000 m depth, based on 1-D joint inversion of TEM and MT data

At temperatures between 100 and 220°C, low-temperature zeolites and the clay mineral smectite are formed. Zeolites and smectite have hydrated and loosely bound cations between the silica plates, making the minerals conductive and with a high cation exchange capacity. The low resistivity reflects the smectite-zeolite zone, owing to the high conductivity of the zeolites and smectite. In the temperature range from 220 to about 240°C the low-temperature zeolites disappear and the smectite is transformed into chlorite in a transition zone; the so-called mixed-layered clay zone is where smectite and chlorite coexist. This zone forms the transition from the low-resistivity cap to the more resistive core.

At about 240°C, smectite disappears and chlorite becomes the dominant mineral. This marks the beginning of the chlorite zone, hence high resistivity since chlorite minerals have cations that are fixed in a crystal lattice making the mineral resistive. This zone corresponds to the resistive core (Árnason et al., 2000; Hersir and Árnason, 2009).

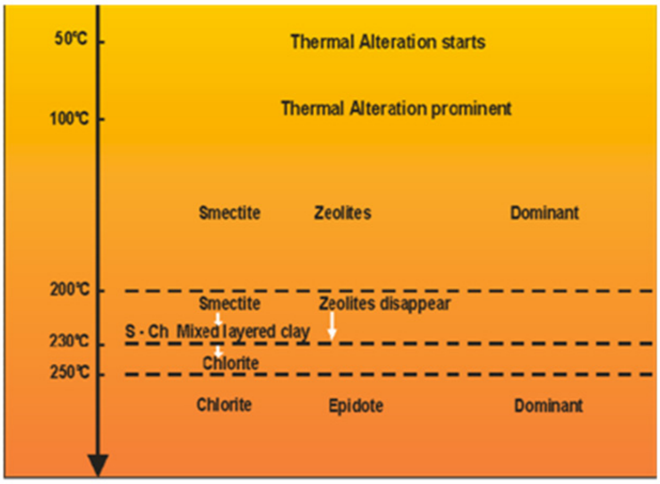

FIGURE 13: Alteration mineralogy and temperature (from Hersir and Árnason, 2009)

The following wells are located close to the NW-SE resistivity cross-section:

TR-01 is 2307 m deep and the temperature at the bottom was measured at 320°C Alteration has been found to be fairly significant and increases with depth. There is a smectite-zeolite zone close to the surface at a few metres depth. Mixed-layer clay minerals are the dominant alteration minerals at 150-500 m depth. There is a chlorite zone at 500 m depth, a chlorite-epidote zone at 700 m depth and an epidote-amphibole zone at 2000 m depth. Temperature measurements show formation temperature of 100°C near the surface, 190°C at 150 m, and 250°C at 500 m depth. The estimated formation temperature in the middle section of the well, from 1100 to 1600 m, is somewhat lower than the alteration temperature, which is probably caused by some cooling. The measured well bottom temperature was 320°C (Fridleifsson et al., 2002; Hersir et al., 2010).

TR-02 is 2280 m deep. There is a smectite-zeolite zone close to the surface at 25-50 m depth, and a mixed-layer clay zone takes over at 150 m depth. Circulation losses occurred at 600-800 m depth and the analysis of alteration minerals was therefore not possible for that section of the well. Epidote is first seen at 520 m depth and becomes the main alteration mineral at 800 m depth, possibly higher. Chlorite is first found at 500-550 m depth. The formation temperature in the well was estimated, after more than 3 months of warming up and an attempt to let it blow. The well temperature has not been measured since 2006. It has been found that at 120 m depth the temperature was 100°C, at 200 m 150°C, at 1200 m 200°C and 300°C at 1550 m depth. The highest measured temperature was 330°C at just over 2100 m depth. A comparison of formation temperature and alteration temperature showed that cooling seems to have taken place down to around 1200 m depth. The alteration temperature is somewhat lower than the formation temperature below that depth, which indicates that formation alteration has not yet progressed in line with temperature (this is possibly still a rather young system). Note that there is a gap in the analysis of alteration minerals in well TR-02, a result of circulation losses between 600-800 m deep while drilling; the rather deep reaching low resistivity under Sveifluháls is only based on one measurement (Mortensen et al., 2006; Hersir et al., 2010).

KR-02 was drilled in 1960 and is 1200 m deep. The highest measured temperature in the well was 197°C at 350 m depth. The temperature was lower at deeper levels, measured as 167°C at 1000 m depth. The smectite-zeolite zone extends from the surface down to 140 m depth, where a zone of mixed-layer clay minerals layer takes over. The chlorite zone starts at 380 m and the chlorite-epidote zone at 737 m depth (Kamah, 1996; Hersir et al., 2010).

KR-08: Alteration in well KR-08, which is 933 m deep and located at Ketill beneath Sveifluháls ridge, is considerably less pronounced than in wells KR-05 and KR-06. Smectite is found in the well from just above 200 down to 400 m, mixed-layer clay from below 300 down to below 700 m, chlorite from just below 400 m and all the way down, and epidote near the well bottom, at just below 800 m depth. The highest temperature in the well was recorded at 190°C at around 400 m depth, but from there it decreases down to the bottom, where it is 170°C (Arnórsson et al., 1975; Hersir et al., 2010).

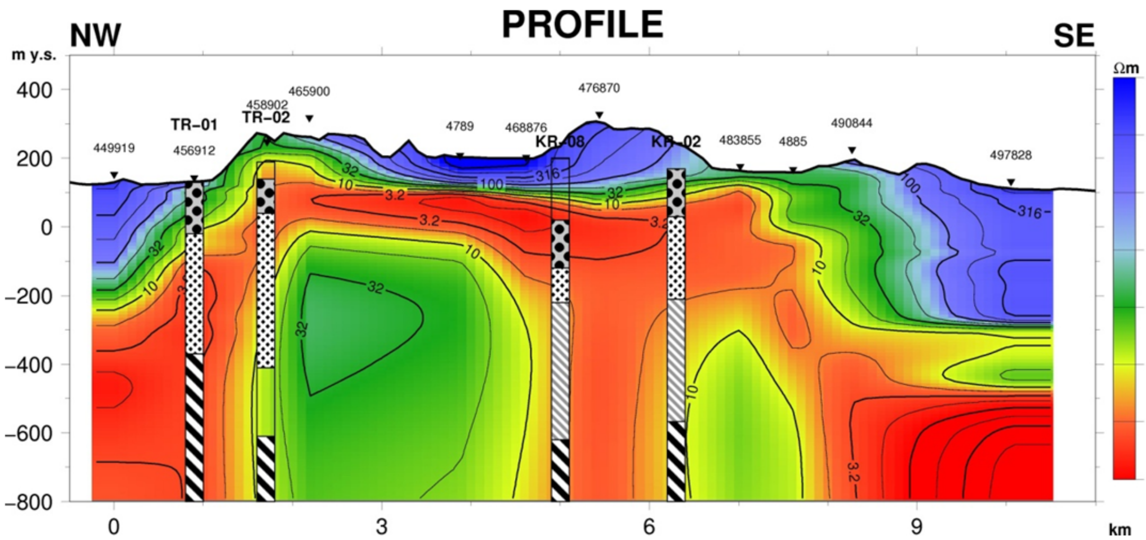

FIGURE 14: The TEM cross-section showing the subsurface resistivity structure and the main alteration mineral zones in drillholes; the order for the alteration minerals in the wells from the wellhead downwards is usually: smectite-zeolites; mixed-layer clay; chlorite; chlorite-epidote

Information on alteration minerals in wells and resistivity values from nearby soundings is summarized in Tables 1 and 2 in Appendix IV. In Figure 14, information on alteration minerals from the wells has been included in the NW-SE cross-section.

7. FINDINGS AND DISCUSSION

Looking at different models from joint 1-D inversion, it has been found that the low-resistivity cap of the subsurface below the NE-SW profile is related to alteration mineralogy (Figure 14). The temperature measured in the drillholes partly reflects the same anomaly. There is a decrease in temperatures at some depths and the geothermal system underneath the profile has probably started cooling down.

Geologically, the NW-SE profile of the studied area reflects highly resistive lava flows in the subsurface, followed by a smectite-zeolite zone, reflecting quite low resistivity as the smectite-zeolite zone is quite conductive. It has been found that the low-resistivity cap is underlain by relatively resistive layers made of mixed-clay minerals. However, a highly resistive core is found at still deeper levels, underneath the mixed-clay zone, which relates to chlorite-epidote; as chlorite is much more resistive because of its fixed ions within a crystal lattice.

Schematically; the NW-SE profile reflects the four main substructures in terms of low resistivity and alteration mineralogy. The uppermost is the highly resistive surface expressed in the subsurface by unaltered rocks; the second shows relatively high resistivity that is related to saturation of the rocks with groundwater, filling pores and aquifers and found at some tens of metres depth. Then a low-resistivity layer can be seen, which is predominantly controlled by the smectite-zeolite minerals. Below the low-resistivity cap, at a few hundred metres depth, a highly resistive core is found, which is probably related to hot and resistive rock with alteration of chlorite-epidote. This is probably the geothermal reservoir. At still deeper levels between 3000 and 5000 m, a probable heat source, characterized by fairly low resistivities ranging between 10 and 32 Ωm, can be inferred.

8. CONCLUSIONS AND RECOMMENDATIONS

The electrical resistivity of the subsurface for the NW-SE profile of the Krýsuvík area was calculated and interpreted. The resistivity of the subsurface correlates fairly well with the alteration mineralogy studied in drillholes. The area underneath the profile appears to host a viable geothermal system. However, the formation temperature seems to be somewhat lower than the alteration temperature which indicates that the system has been cooling down. A number of additional TEM and MT soundings are needed to cover the area better. In future, further studies should be carried out to get a better understanding of the Krýsuvík geothermal system.

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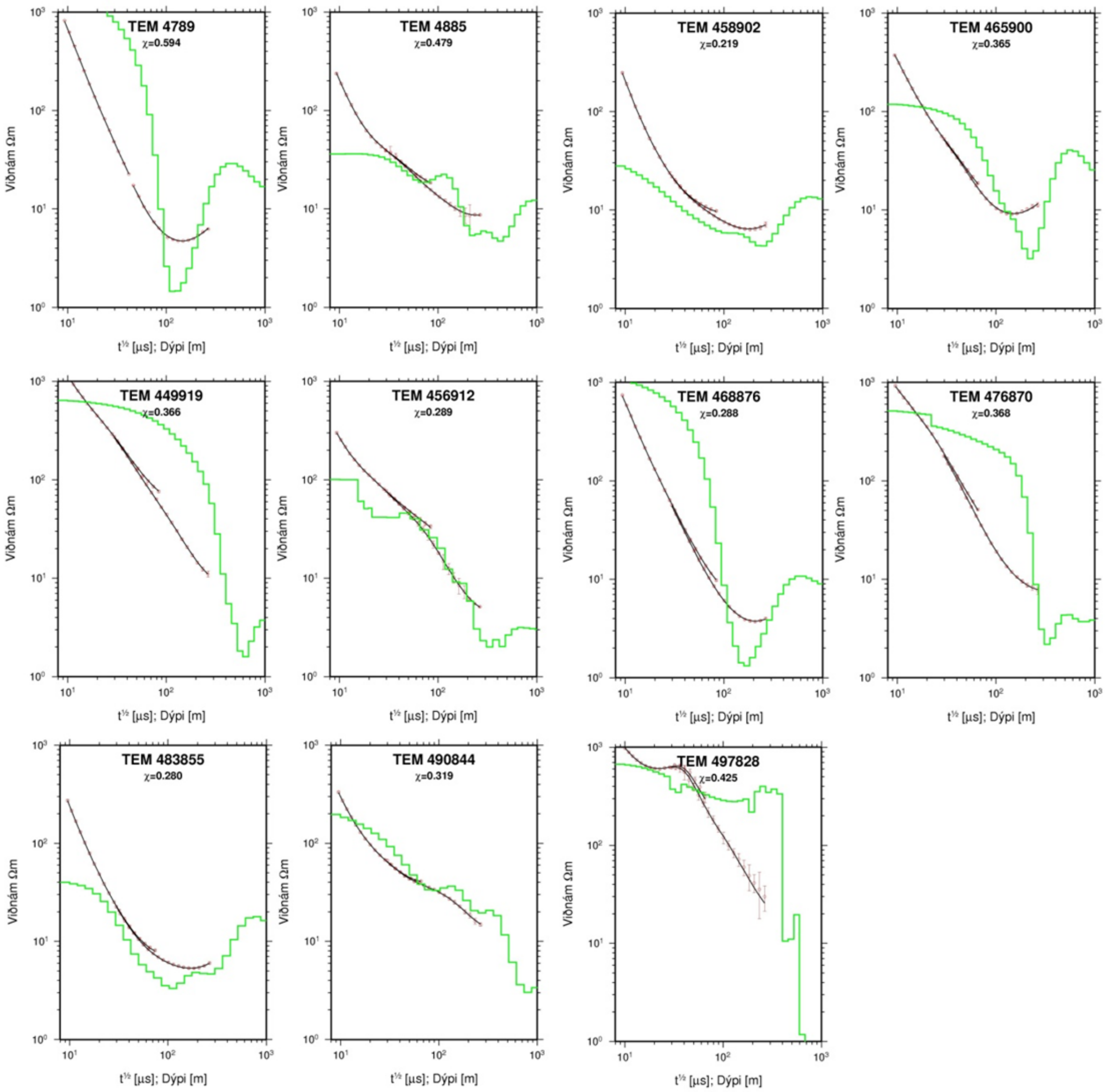
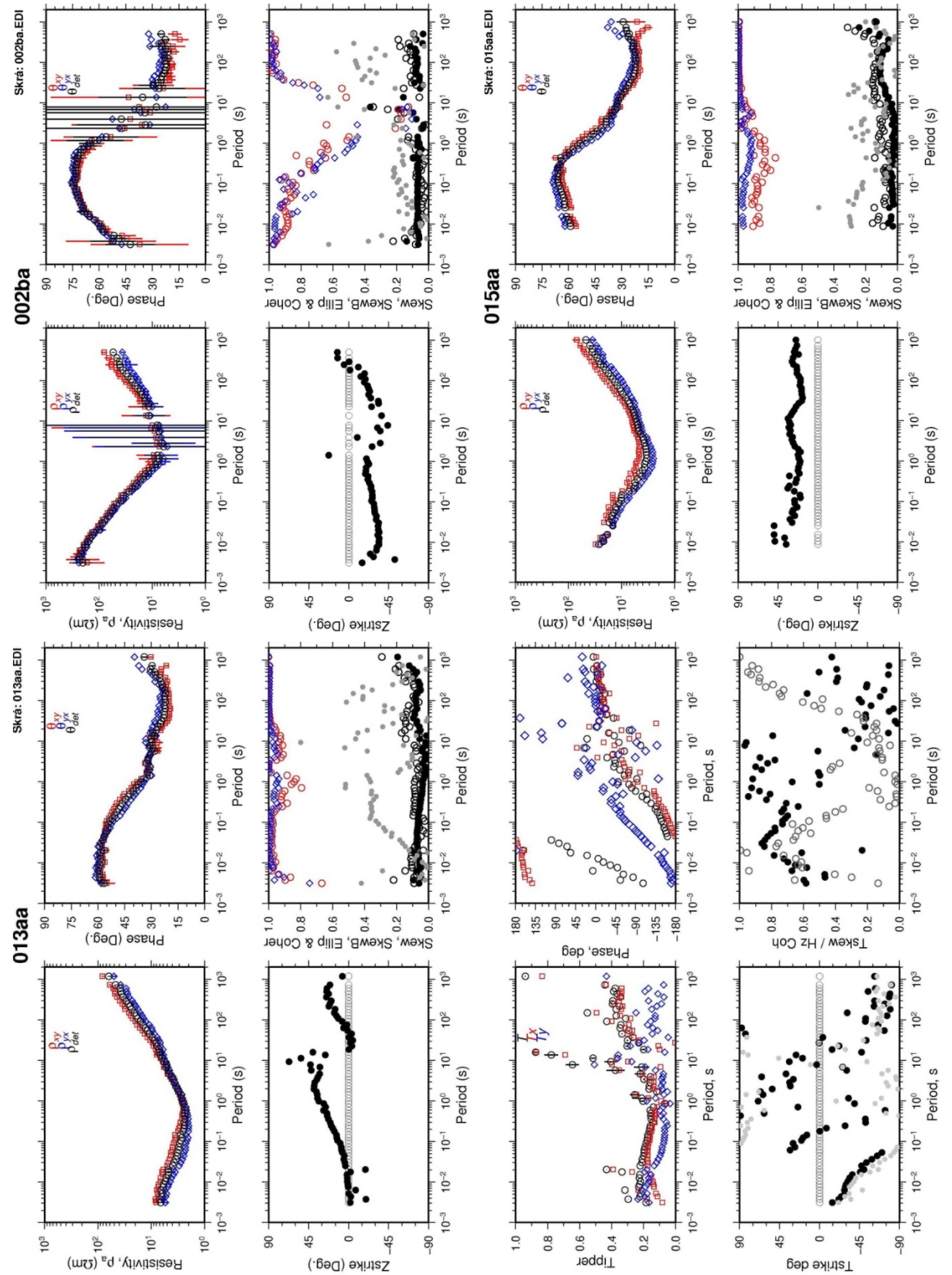
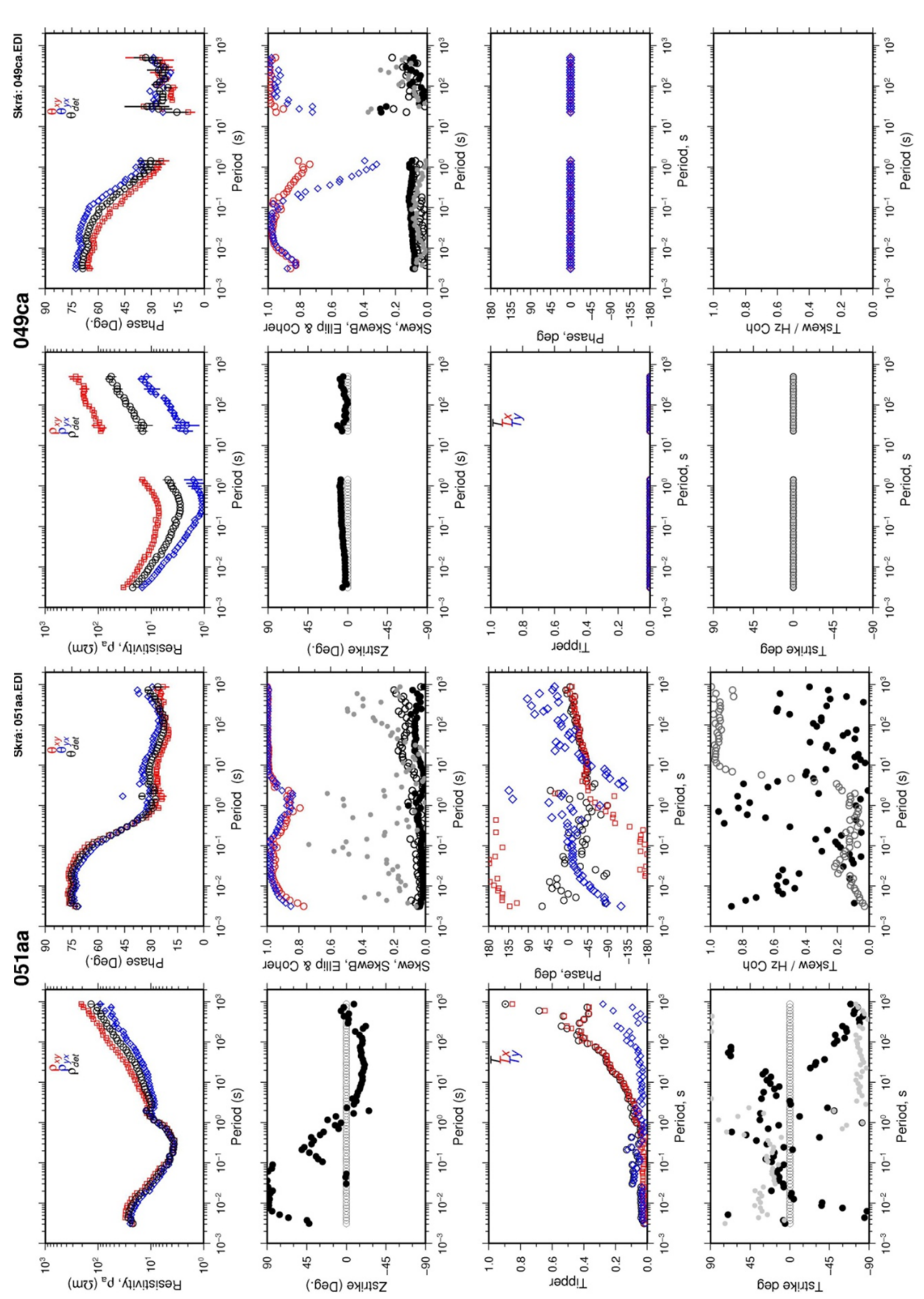
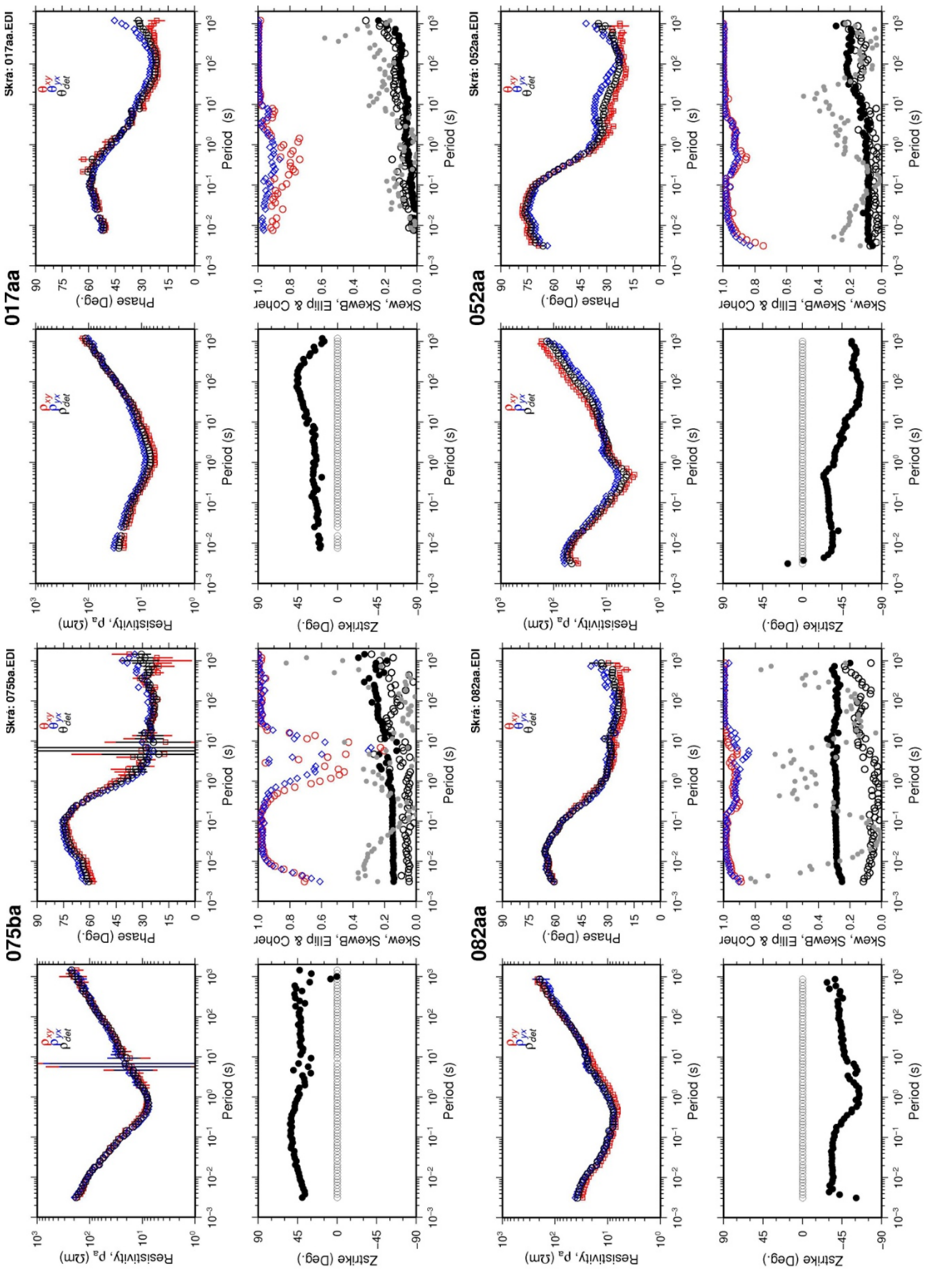
APPENDIX I: TEM data and 1-D inversion models from the TEMTD program

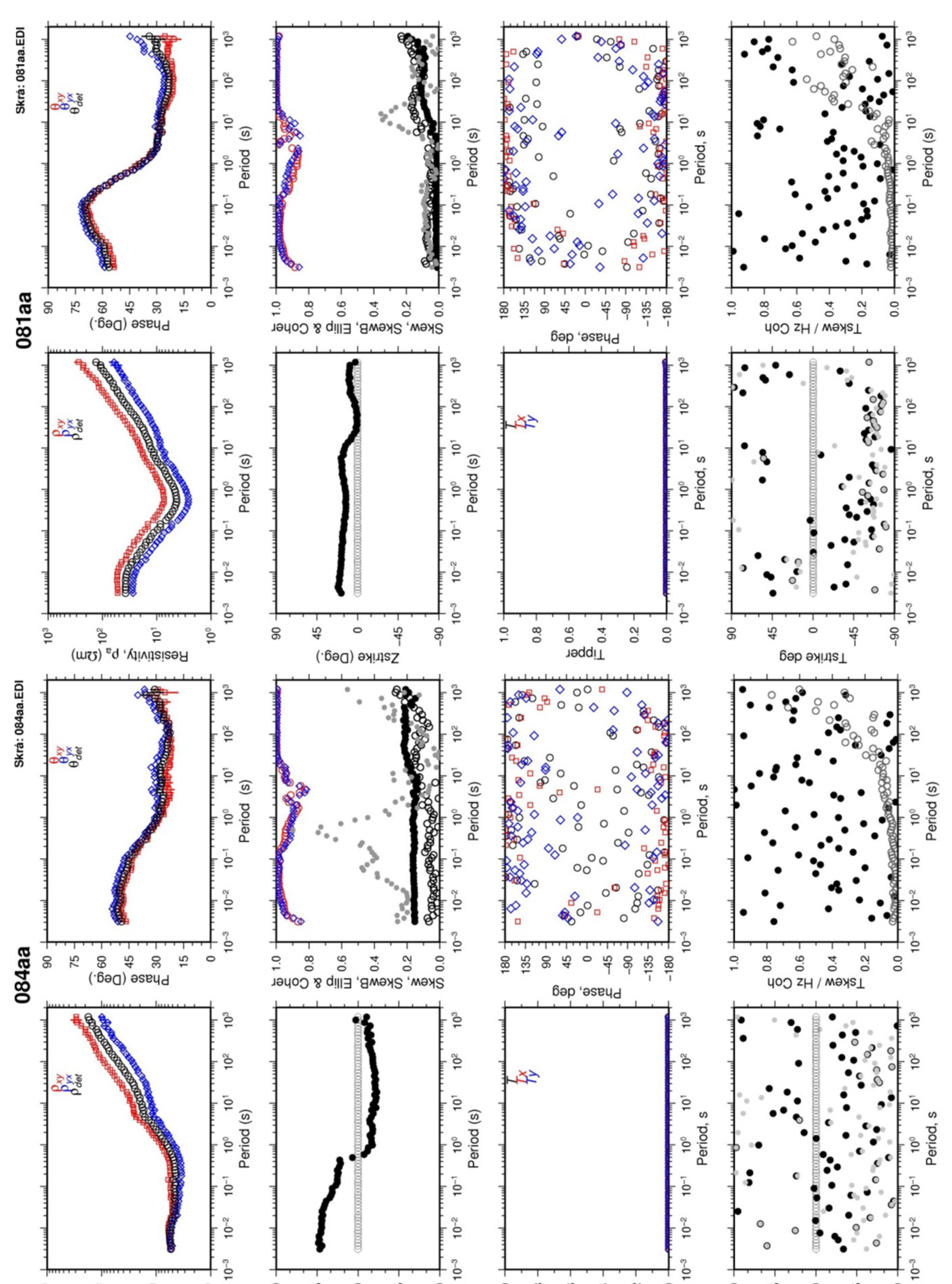


APPENDIX II: MT data (EDI files)

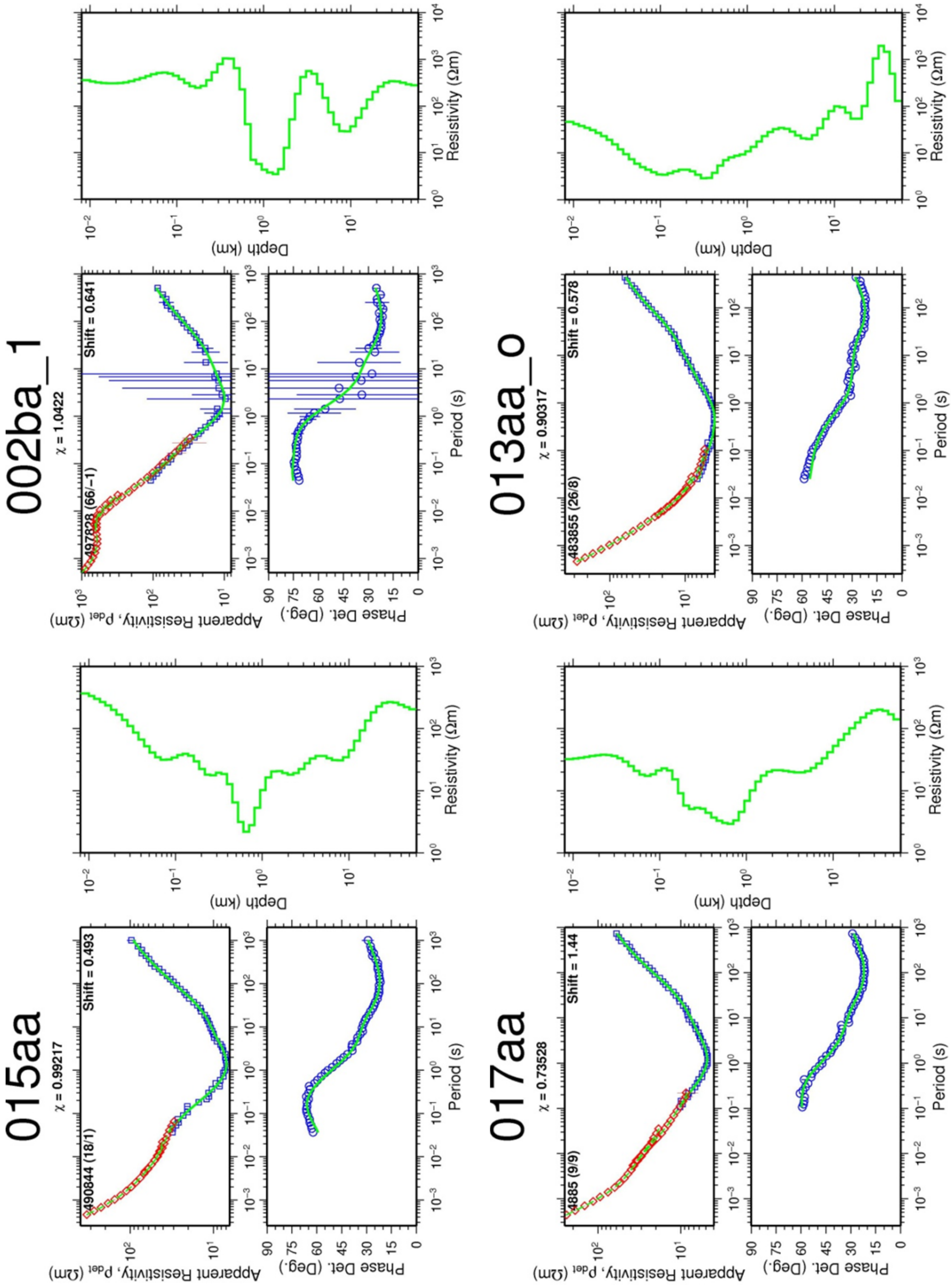


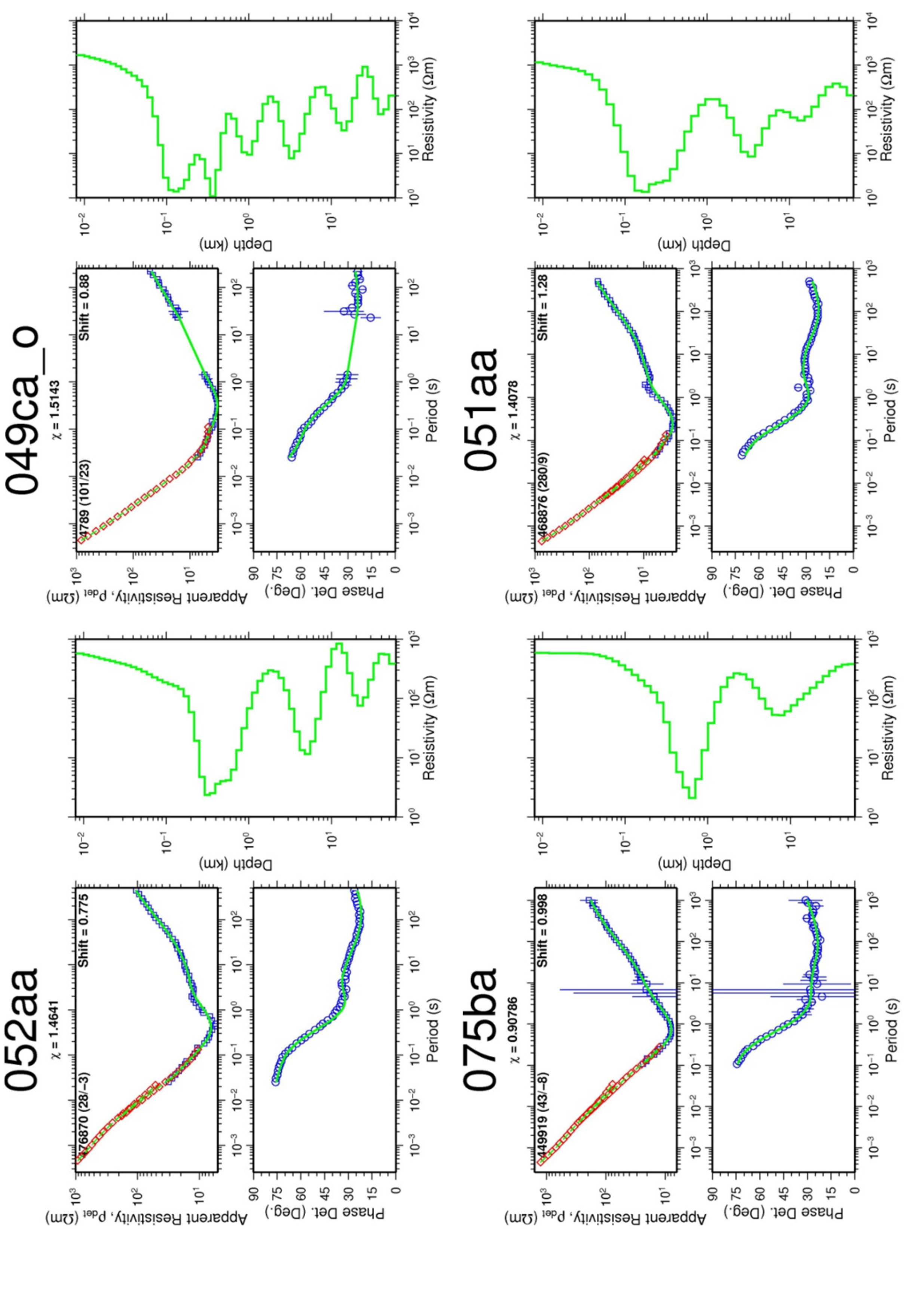

051aa
Skrá: 051aa.EDI
049ca
Skrá: 049ca.EDI
Resistivity, ρa (Ωm)
Phase (Deg.)
Zstrike (Deg.)
Skew, SkewB, Ellip & Coher
Tipper
Phase, deg
Tstrike deg
Tskew / Hz Coh
Period (s)
Period, s

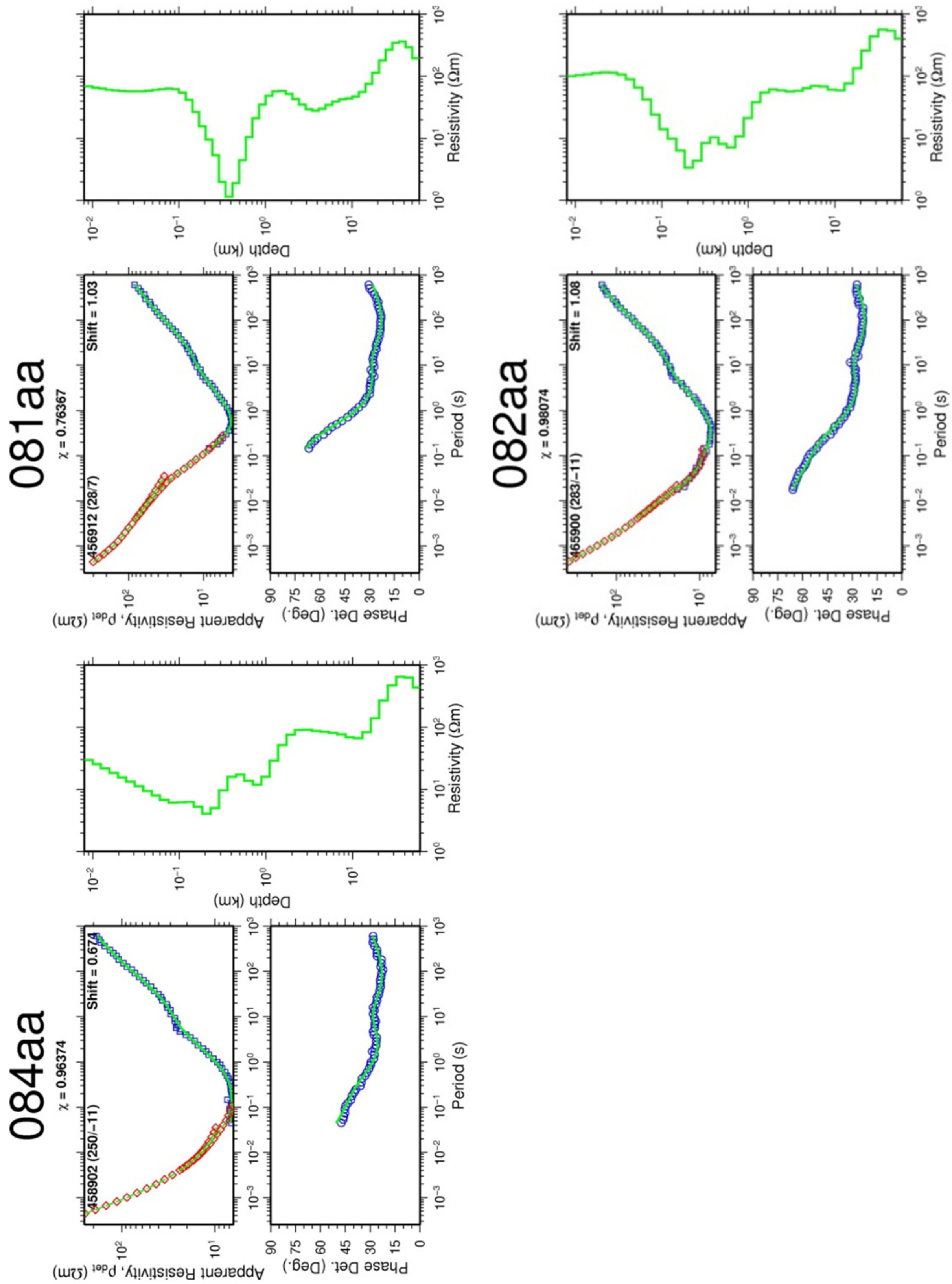

075ba
Skrá: 075ba.EDI
017aa
Skrá: 017aa.EDI
082aa
Skrá: 082aa.EDI
052aa
Skrá: 052aa.EDI
Resisitivity, ρa (Ωm)
Phase (Deg.)
Zstrike (Deg.)
Skew, SkewB, Ellip & Coher
Period (s)


084aa
Skrá: 084aa.EDI
081aa
Skrá: 081aa.EDI
Resistivity, ρa (Ωm)
Phase (Deg.)
Period (s)
Zstrike (Deg.)
Skew, SkewB, Ellip & Coher
Tipper
Phase, deg
Period, s
Tstrike deg
Tskew / Hz Coh

APPENDIX III: Joint 1-D TEM and MT inversion models from the TEMTD program




052aa
χ = 1.4641
Shift = 0.775
049ca_o
χ = 1.5143
Shift = 0.88
075ba
χ = 0.90786
Shift = 0.998
051aa
χ = 1.4078
Shift = 1.28
Apparent Resistivity
Phase Det. (Deg.)
Period (s)
Depth (km)
Resistivity (Ωm)


084aa
χ = 0.96374
458902 (250/−11)
Shift = 0.674
081aa
χ = 0.76367
456912 (28/7)
Shift = 1.03
082aa
χ = 0.98074
465900 (283/−11)
Shift = 1.08
Apparent Resistivity, ρdet (Ωm)
Phase Det. (Deg.)
Period (s)
Depth (km)
Resistivity (Ωm)

APPENDIX IV: Tables on mineral alteration

TABLE 1: Summary on mineral alteration and temperature in drillholes

Well no.	Temperature (°C)	Depth (m)	Alteration compared to the temperatures
TR-01 (2307 m)	100	Near the surfaces	Alteration is fairly significant and increases with depth;
	190	150	Low resistivity found in the upper most crust is probably controlled by smectite-zeolite alteration;
	250	500	The estimated temperature in the mid-zone of the well, from 1100 down to 1600 m, is somewhat lower than the alteration temperature. This is probably caused by some cooling;
	320	At the bottom	
TR-02 (2280 m)	100	At 120 m	At 600-800 m, circulation losses occurred during drilling operations - analysis was not available for this section of the well; Formation temperature and alteration temperature showed a cooling to have taken place down to around 1200 m depth. The alteration temperature is somewhat lower than the formation temperature below that depth. This indicates that the alteration has not yet progressed in line with temperature, indicating possibly still a rather young system.
	150	200 m	
	200	1200 m	
	250	1300 m	
	300	1550 m	
	330	Below 2100 m	
KR-02 (1200 m)	197	At 350 m	Drilled in 1960; well KR-02 is not well documented.
KR-08 (933 m)	190	Max. temperature at around 400 m	Well KR-08 is located at Ketill beneath Sveifluháls. Alteration is considerably less pronounced than in other wells; Smectite-zeolite zone is found at 200 m depth; while in the other wells it is found very close to the surface.
	167	Temperature slowly decreases to the bottom at 1000 m	

TABLE 2: Classification of mineral alteration with depth in drillholes

Mineral alteration	TR-01 (2307m)	TR-02 (2280 m)	KR-02 (1200 m)	KR-08 (933 m)
Smectite-zeolite	Close to surface at few m depth	25-50 m depth	Surface-140 m depth	200–400 m depth
Mixed-layer clays	150-500 m	Takes over at 150 m depth		300-700 m depth
Chlorite	500 m depth		Starts at 380 m depth	From 400 m all the way down
Ubiquitous	150-500 m			
Chlorite-epidote	700 m	520 m - higher at 800 m and still higher deeper	737 m depth	800 m down to the bottom
Epidote-amphibole	2000 m depth	Epidote at 520 m depth		