RESISTIVITY SURVEYING AND ELECTROMAGNETIC METHODS

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ABSTRACT

The task involved in geothermal exploration is the detection and delineation of geothermal resources and the understanding of their characteristics, the location of exploitable reservoirs, and the siting of boreholes through which hot fluids at depth can be extracted. Geological and geochemical mapping are usually limited to direct observations on the surface and conclusions and extrapolation that can be drawn about the system and possible underlying structures. Geophysical surface exploration methods are different. They utilize equipment that measures directly some physical parameters on the surface that are directly created by physical properties or processes at depth. Geophysical exploration is a young scientific discipline that developed slowly in the first half of the twentieth Century.

Geophysical exploration methods can be classified into several groups like seismic methods, electrical resistivity methods, potential methods (gravity and magnetics), heat flow measurements, and surface deformation measurements. Resistivity methods are the most important geophysical methods in geothermal exploration. The reason is that the resistivity is highly sensitive to temperature and geothermal alteration processes and is directly related to parameters characterizing the reservoir. Here, a review will be given on resistivity surveying and electromagnetic methods; the transient electromagnetic (TEM) and magnetotelluric (MT) method as well as the DC method: Schlumberger soundings.

1. INTRODUCTION

Measuring the electrical resistivity, ρ, of the subsurface is the most powerful geophysical prospecting method in geothermal exploration and the main method used for delineating geothermal resources. The electrical resistivity (Ωm) is defined by the Ohm's law, $E = \rho j$, where E (V m^{-1}) is the electrical field and j (A m^{-2}) is the current density. Electrical resistivity can also be defined as the ratio of the potential difference ΔV (V) to the current I (A), across a material which has a cross-sectional area of 1 m^2 and is 1 m long ($\rho = \Delta V/I$).

The electrical resistivity of rocks is controlled by important geothermal parameters like temperature, fluid type and salinity, porosity, the composition of the rocks, and the presence of alteration minerals. The reciprocal of resistivity is conductivity, σ (Sm^{-1}). Therefore, it is also possible to talk about conductivity measurements. However, in geothermal research, the tradition is to refer to electrical or resistivity measurements.

There exist several different methods to measure the subsurface resistivity. The common principle of all resistivity methods is to create an electrical current within the Earth and monitor, normally at the surface, the signals generated by the current. There are two main groups of resistivity methods, direct current (DC) method and electromagnetic (EM) method (sometimes called AC soundings). In conventional DC methods, such as Schlumberger soundings, the current is injected into the ground through a pair of electrodes at the surface and the measured signal is the electric field (the potential difference over a short distance) generated at the surface. In EM methods, the current is induced in the Earth by an external magnetic field. In MT, alternating current is induced in the ground by natural oscillations in the Earth's magnetic field, and the measured signal is the electric field at the surface. In TEM, the current is created by a man-made time-varying magnetic field generated by a current in a loop on the surface or by a grounded dipole. The monitored signal is the decaying magnetic field at the surface caused by induced currents at depth. It is customary in geophysics to talk about passive and active methods, depending on whether the source is a natural one or a controlled (artificial) one. MT is an example of a passive method, whereas Schlumberger and TEM are active ones.

All geophysical exploration technologies involve four steps: data acquisition, processing of data as an input for inversion or modeling, the modeling of the processed data, and finally the interpretation of the subsurface resistivity model in terms of geothermal parameters. In the resistivity method, the term apparent resistivity, ρ_a, is used. This denotes the measured or calculated resistivity as if the Earth was homogeneous. It is a sort of an average of the true resistivity of the Earth detected by the sounding down to the penetration depth of the subsurface currents. The measured apparent resistivity is inverted to the true resistivity of the subsurface through modeling.

In all types of resistivity measurements, the final product of the data acquisition and the accompanying processing is a curve normally giving the apparent resistivity as a function of some depth-related free parameter. In the case of DC soundings, the free parameter is the electrode spacing; for TEM soundings, the time after turning off the source current; and the period of the EM fields in case of MT soundings.

Since the apparent resistivity does not show the true resistivity structure of the Earth, it has to be modeled in terms of the actual spatial resistivity distribution, that is, resistivity as a function of the two horizontal directions, x and y, and the vertical direction, z. This is the task of the geophysical modeler, transforming the measured apparent resistivity into a model of the true resistivity structure. The procedures are similar, whether considering DC soundings or EM soundings. In the modeling, geometrical restrictions of the resistivity structure are applied; the modeling is done in a one-, two-, or three-dimensional (1D, 2D, or 3D) way. In the 1D modeling, the resistivity distribution is only allowed to change with depth and is in general assumed to resemble a horizontally layered Earth (Figure 1).

The 2D modeling means that the resistivity distribution changes with depth and in one lateral direction, but is assumed to be constant in the other orthogonal horizontal direction. The last one is the so-called electrical strike direction, which is usually the direction of the main structure or the geological strike in the area. In a 2D survey, soundings are made along a profile line, which should be perpendicular to the strike. Good data density is needed, depending on the required spatial subsurface resistivity resolution, preferably a spacing of less than 1–2 km between soundings. The 2D modeling can account for variations in the topography.

The 3D modeling allows the resistivity to vary in all three directions. For a meaningful 3D interpretation, high data density is needed with a good areal coverage of the soundings, again depending on the required spatial subsurface resistivity resolution, preferably on a regular grid, for example, 1 km between sites. Soundings located well outside the prospected area are necessary to constrain the 3D subsurface resistivity model.
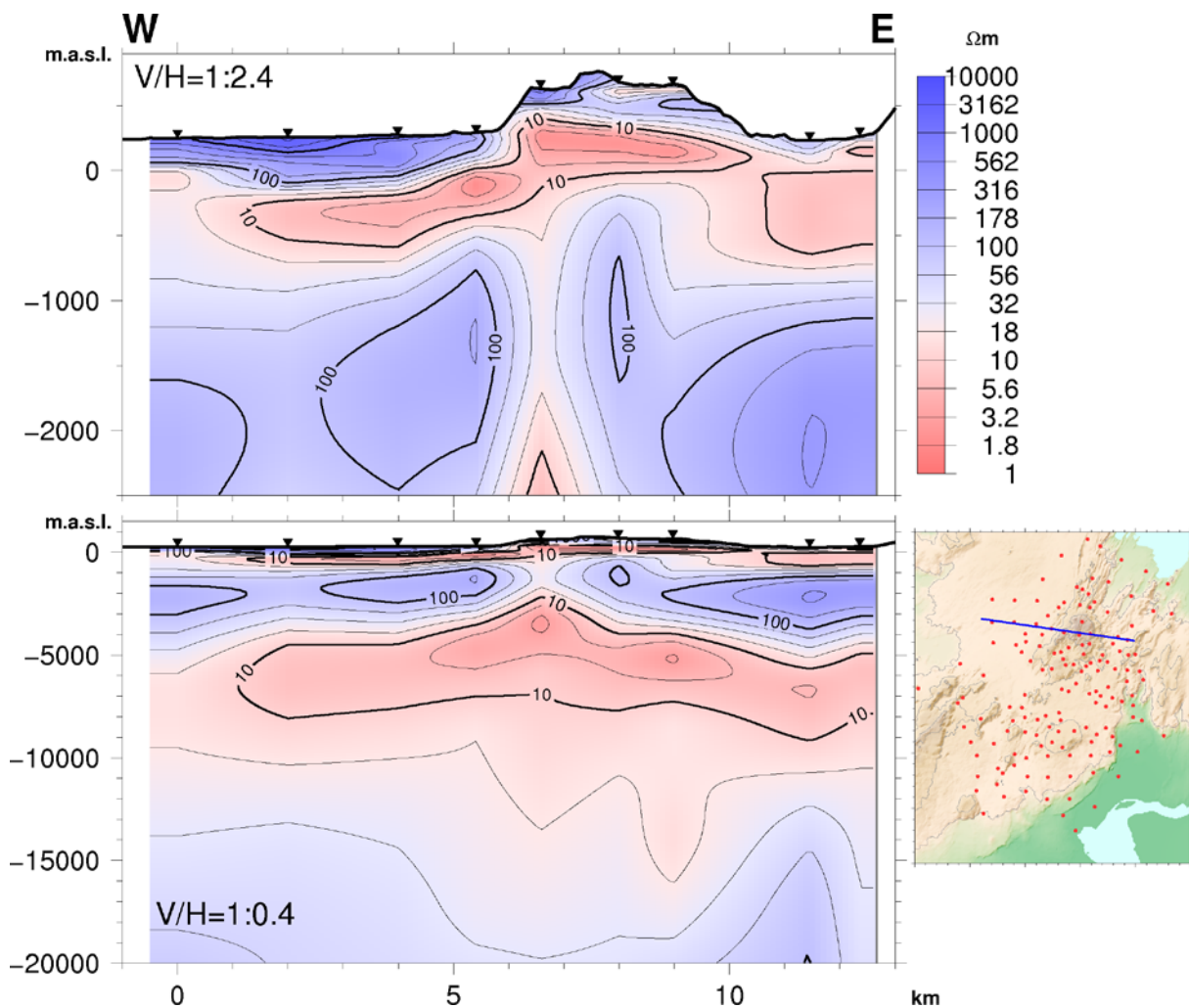
FIGURE 1: Resistivity section across the Hengill high-temperature geothermal area, Southwest Iceland, for two different depth ranges obtained from stitched joined 1D inversion of TEM and determinant MT data. Inverted triangles denote MT stations, and V/H is the ratio between vertical and horizontal axes. Section location is shown as a blue line in the map to the right. Red dots in the map denote MT stations (Árnason et al., 2010).

In 1D modeling, data from one and only one sounding are supposed to fit the response from a given model. Although a 1D Earth is assumed, in practice there are different 1D models for different soundings within the same survey area – every sounding has its own 1D model. In 2D modeling, data from all the soundings on the same profile line are supposed to fit the response from the same 2D model. In 3D modeling, data from all the soundings in the survey or modeled area are supposed to fit the response from the same 3D model.

The review given here on, „Resistivity surveying and electromagnetic methods", is to a great extend based on previous work by the authors (Hersir and Björnsson, 1991; Flóvenz et al., 2012).

2. SCHLUMBERGER SOUNDINGS

In DC methods, direct current is injected into the ground through a pair of electrodes at the surface. The current in the Earth produces an electrical field in the surface that is related to the resistivity of the underlying ground. The electrical field is determined from the measured potential difference between a pair of electrodes at the surface.

Schlumberger soundings have been widely used through recent decades in geothermal prospecting. The electrode configuration is shown in Figure 2. The electrodes are on a line, and the setup is symmetric around the center. A pair of potential electrodes (usually denoted by M and N) is kept close to the center, while a pair of current electrodes (usually denoted as A and B) is gradually moved away from the center, for the current to probe deeper into the Earth. The distance between the current electrodes is commonly increased in near-logarithmic steps (frequently 10 steps per decade) until the scheduled maximum separation has been reached. In principle, the distance between the potential electrodes MN should be small and fixed, but in practice, it needs to be enlarged a few times as the spacing between the current electrodes is increased. This is to increase the voltage signal and to maintain an acceptable signal to noise ratio.
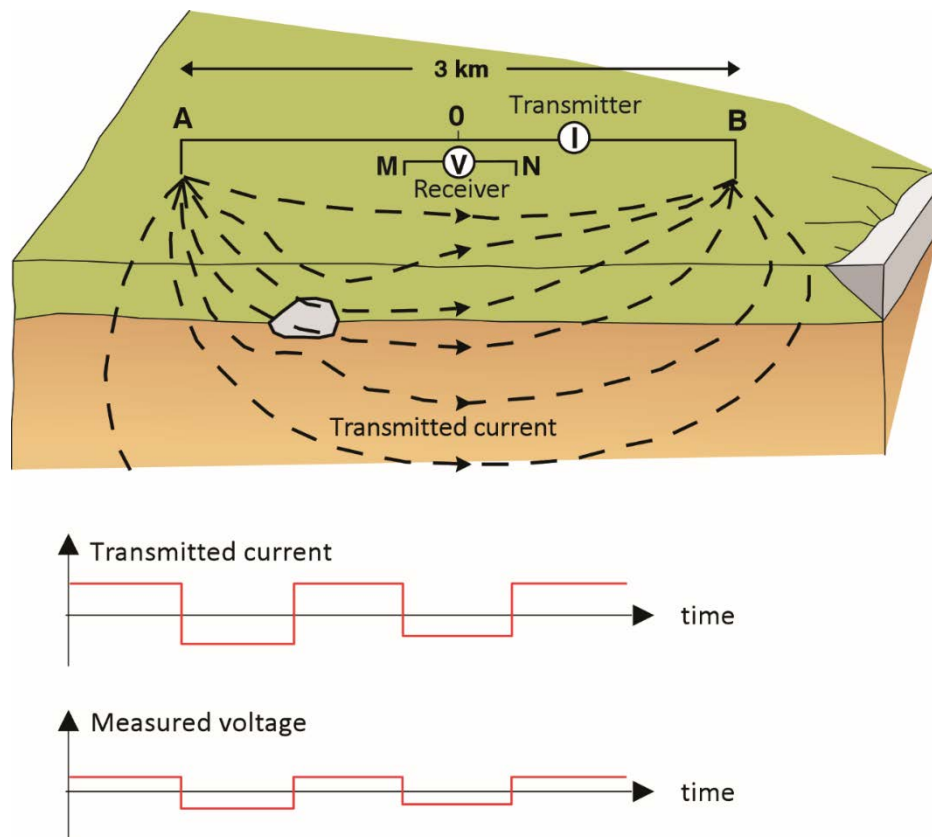

FIGURE 2: Schlumberger configuration: As the spacing between the current electrodes, A and B, is increased the current penetrates deeper into the subsurface and the measured potential difference at the surface (between M and N) is affected by the resistivity of deeper lying layers. The lower part of the Figure shows the typical half duty square wave current and the corresponding potential signal. (With courtesy of ÍSOR).

The measured resistivity value is the apparent resistivity, ρ_a, a sort of an average resistivity of the material through which the current passes. By using Equation 1, the apparent resistivity can be easily derived from the measured current and potential difference and the geometrical setup parameters (Figure 2) as follows:

$$\rho_a = \frac{\Delta V}{I} \cdot (S^2 - P^2)\frac{\pi}{2P} \tag{1}$$

where S = AB/2, P = MN/2, and K is a geometrical factor.
During the measurement, the apparent resistivity obtained from Equation 1 is plotted as a function of AB/2 on a bilogarithmic scale and then inverted into a resistivity model. For a single sounding, it is done in 1D way, traditionally by assuming that the Earth is made of horizontal homogeneous and

isotropic layers with constant resistivity. The apparent resistivity curve can be inverted to estimate the resistivity and thicknesses of the layers. An example of an apparent resistivity curve and the result of the 1D inversion are shown in Figure 3.
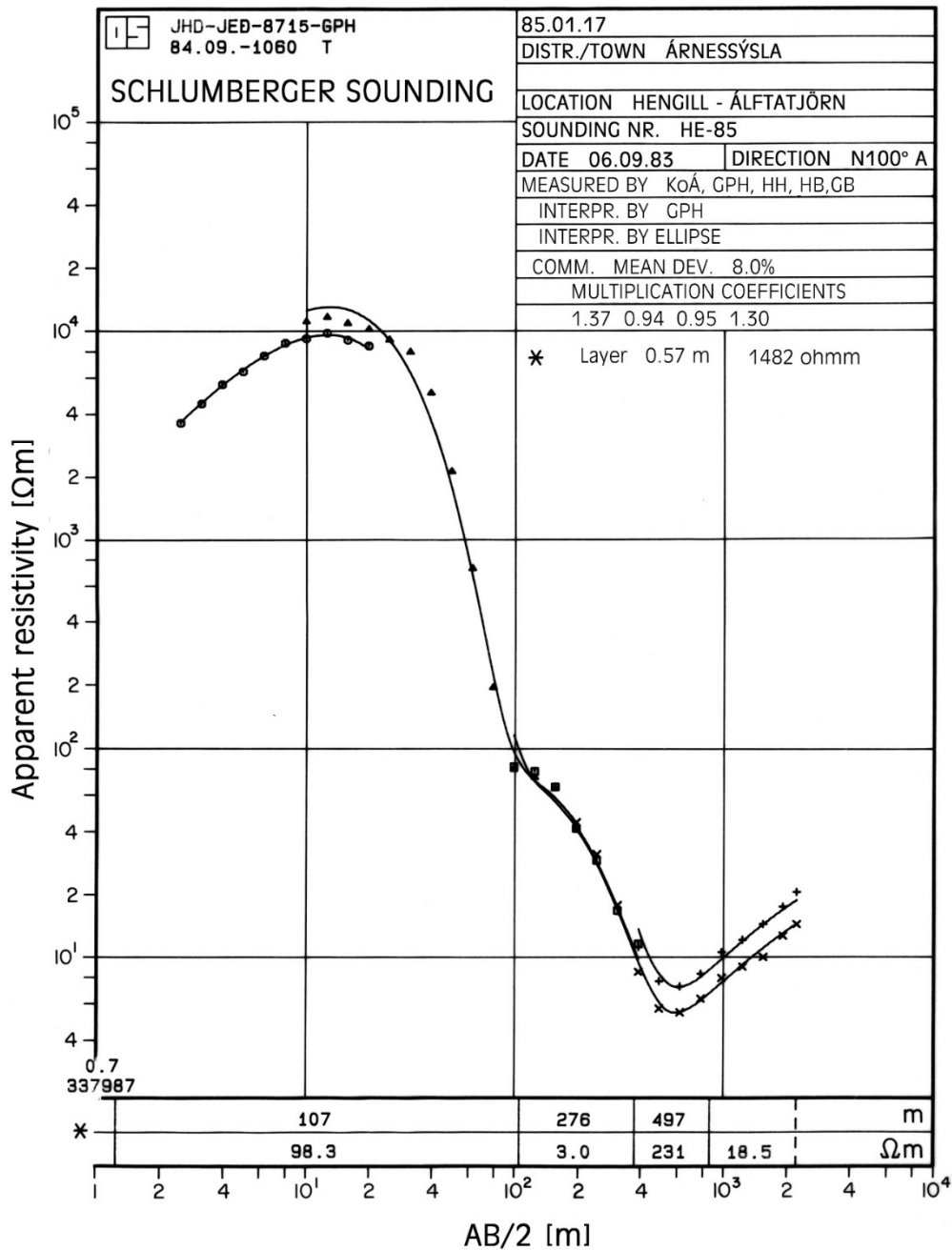

FIGURE 3: One dimensional inversion (layered earth model) of a Schlumberger sounding from the Hengill high-temperature geothermal area, Southwest Iceland (Hersir et al., 1990). The data points are black dots and the calculated curve from the final model (the response) is shown as black lines. The final model parameters (resistivity and thickness of layers) are shown at the bottom.

Typical soundings for geothermal exploration have a maximum current electrode spacing AB/2 of 1–3 km, but much longer spacing have been used. In practice, very long wire distances can be difficult to handle and the injected current has to be quite large, otherwise the voltage signal will drown in noise. For the depth penetration of the sounding, a rule of thumb says that it reaches down to about one-third of the distance AB/2, but is in fact dependent on the actual resistivity structure.

3. TEM SOUNDINGS

The TEM method is a fairly recent addition to the resistivity methods used in geothermal exploration, developed and refined since the late 1980s. This is mainly because the TEM response covers a very large dynamic range and advances in electronics were needed, and second, because interpretation of the data is intensive and relatively large computers were needed.

The principle of the TEM method is shown in Figure 4. 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 in the loop. The current is abruptly turned off. The magnetic field is then left without its source and responds by
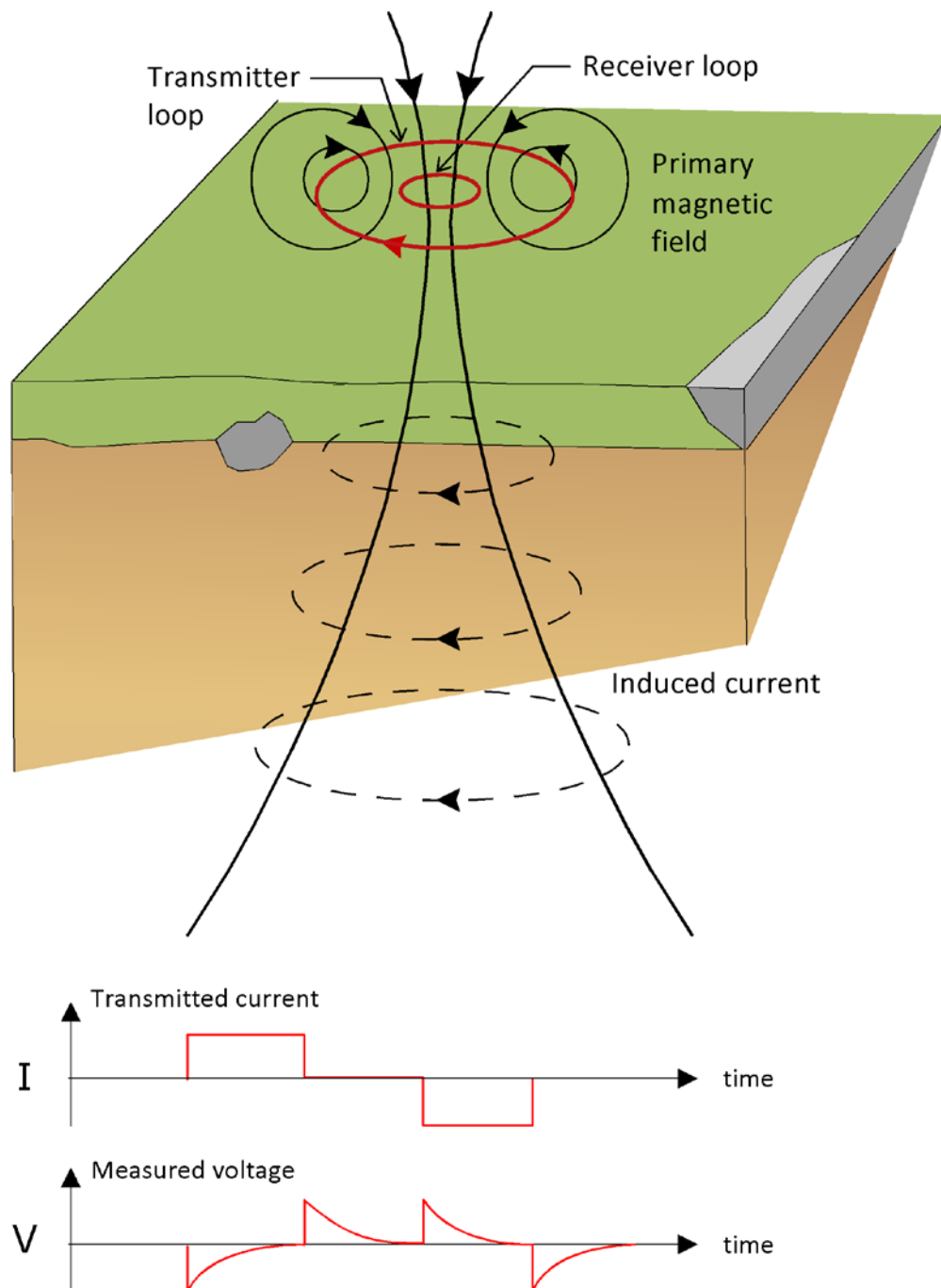
FIGURE 4: TEM sounding setup; the receiver coil is in the center of the transmitter loop. Transmitted current and measured transient voltages are shown as well. (With courtesy of ÍSOR).

inducing an image of the source loop in the surface. With time, the current and the magnetic field decay and again induce electrical currents at greater depths in the ground. The process can be visualized as if, when the current is turned off, the induced currents, which at very early times are an image of the source loop, diffuse downwards and outwards like a smoke ring (Figure 4). The decay rate of the magnetic field with time is monitored by measuring the voltage induced in a receiver coil at the center of the transmitting loop as a function of time, normally at prefixed time gates equally distributed in log time. The decay rate of the magnetic field with time is dependent on the current distribution which in turn depends on the resistivity structure of the Earth. The induced voltage in the receiver loop, 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 transmitter and receiver are synchronized either by connecting them with a reference cable or by high-precision crystal clocks so that the receiver gets to know when the transmitter turns off the current. Turning off the current instantaneously would induce infinite voltage in the source loop. Therefore, the transmitters are designed to turn off the current linearly from maximum to zero in a short but finite time called turn-off time. The zero time of the transients is the time when the current has become zero and the time gates are located relative to this. This implies that the receiver has to know the turn-off time. To reduce the influence of EM noise, the recorded transients are stacked over a number of cycles before they are stored in the receiver memory.

The depth penetration of the TEM method depends on the resistivity beneath the sounding as well as on the equipment and the field layout used (i.e., the setup geometry and the generated current and its frequency). The depth penetration increases with time after the current turn-off. Different frequencies of the current signal are therefore used, high frequencies for shallow depths and low frequencies for deep probing. For typical geometries and frequencies, the penetration depth is of the order of or somewhat < 1 km, depending also on the subsurface resistivity.

Typically, the size of the source loop is a 300 m x 300 m square loop (sometimes 200 m x 200 m, reducing the depth of penetration to around 500 m), and the transmitted current is a half-duty square wave (Figure 4) of around 20–25 A with frequencies of 25 and 2.5 Hz (50 Hz electrical environment).

The results of a TEM sounding is, similarly to a Schlumberger sounding, expressed as an apparent resistivity, ρ_a (or more correctly the so-called late-time apparent resistivity), as a function of time after the current turn-off. At late times after the current turn-off, the induced currents have diffused way below the surface and the response is independent of near-surface conditions. The apparent resistivity is a function of several variables, including the induced voltage (V) measured at the time, t, elapsed after the current in the transmitter loop has been turned off; r which denotes the radius of the transmitter loop, the effective area (cross-sectional area times the number of windings) of the transmitter loop ($A_s n_s$), and the receiver coil ($A_r n_r$); and the current strength (I_0) and magnetic permeability (μ_0). The apparent resistivity, ρ_a, is given as follows in Equation 2:

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

The apparent resistivity curve is then inverted in terms of a horizontally layered Earth model with homogeneous and isotropic layers to give a 1D model below each sounding. For all the layers in the inversion, both the resistivity and layers' thicknesses can vary. An alternative method for 1D interpretation, and a more commonly one used today, is Occam (minimum structure) inversion (Constable et al., 1987). It is based on the assumption that the resistivity varies smoothly with depth rather than in discrete layers. In the Occam inversion, the smooth variations are approximated by numerous thin layers of fixed thickness and the data are inverted for the values of the resistivity (Figure 5).
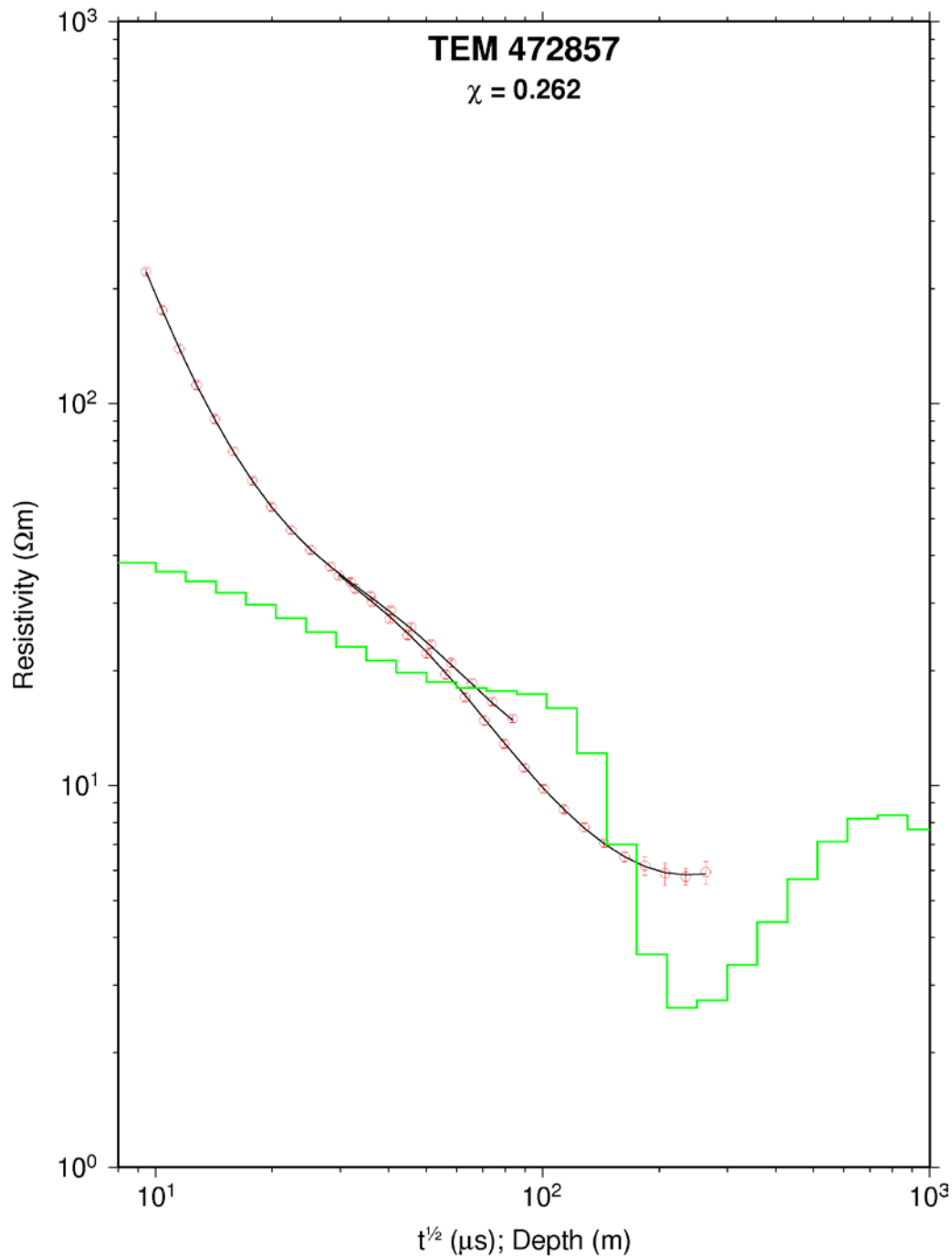
FIGURE 5: A TEM sounding and its 1D Occam inversion from high-temperature geothermal area Krýsuvík, Southwest Iceland (Hersir et al., 2010). Red circles: measured late-time apparent resistivity (different datasets for different receiver loop sizes and current frequencies); and black line: apparent resistivity calculated from the model shown in green. At the top of the Figure is the misfit function; the root-mean-square difference between the measured and calculated values, $\chi = 0.262$.

4. MT SOUNDINGS

MT is a passive method where the natural time variations in the Earth's magnetic field, the so-called micropulsations and sferics, are the signal source. Variations in the magnetic and the corresponding electric field in the surface of the ground are registered, which are used to reveal the subsurface resistivity distribution.

The MT method has the greatest exploration depth of all resistivity methods, some tens or hundreds of kilometers, depending on the recording period, and is practically the only method for studying deep resistivity structures. Similar to the TEM method, the MT method has, due to developments in the electronic and data industry in recent years, improved tremendously, on both the acquisition side (equipment and the measurement techniques) as well as on data analysis and the inversion of the data. MT has become a standard tool in surface exploration for geothermal resources.

In the MT method, the natural fluctuations of the Earth's magnetic field are used as a signal source. Those fluctuations induce an electric field and hence currents in the ground, referred to as eddy currents, which are measured on the surface in two horizontal and orthogonal directions (E_x and E_y). The eddy currents are also called telluric currents, derived from the Latin word tellus, which means Earth. The magnetic field is measured in three orthogonal directions (H_x, H_y, and H_z). A typical setup for an MT sounding is shown in Figure 6. It is customary to have the x direction to the magnetic north. For a homogeneous or layered Earth, the field is induced by its orthogonal source magnetic field (i.e., E_x correlates with H_y and E_y with H_x). For more complicated resistivity structures, these relations become more complex. The magnetic field is usually measured with induction coils and the electrical field by a pair of electrodes, filled with solutions like copper sulfate or lead chloride. The electrode dipole length is in most cases 50-100 m. The electric field equals the potential difference between the two electrodes divided by its length. A GPS unit is used to synchronize the data. The digital recording of the EM fields as a function of time is done through an acquisition unit and the time series are saved on a memory card.
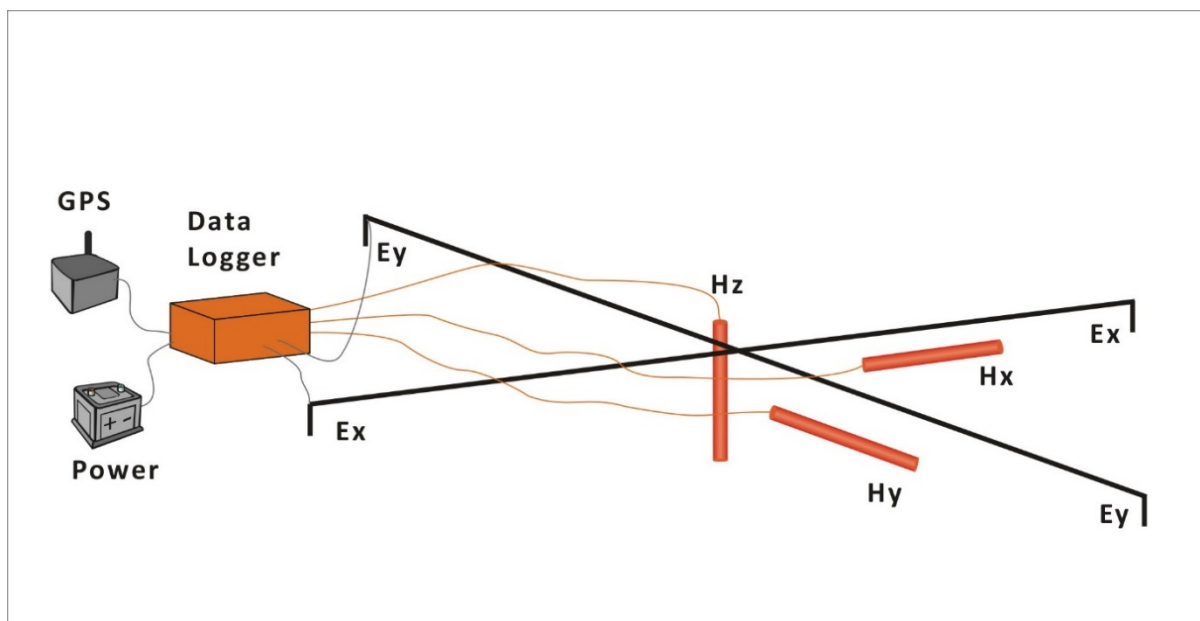

FIGURE 6: The setup of a magnetotelluric sounding. (With courtesy of ÍSOR).

MT generally refers to recording time series of electric and magnetic fields of wavelengths from 0.0025 s (400 Hz) to 1000 s (0.001 Hz) or as high as 10.000 s (0.0001 Hz). Audio magnetotellurics (AMT) refers to 'audio' frequencies, generally recording frequencies of 100 Hz to 10 kHz. Long-period magnetotellurics (LMT) generally refers to recording from 1.000 to 10.000 s or even much higher (to 100.000 s). Note that in MT, it is customary to talk both about wavelengths, measured in seconds (s) and its transformation, the frequency (1/T) measured in Hertz (Hz).

The small-amplitude geomagnetic time variations of Earth's EM field contain a wide spectrum generated by two different sources. The low frequencies (long periods) are generated by ionospheric and magnetospheric currents caused by solar wind (plasma) emitted from the sun interfering with the Earth's magnetic field known as micropulsations. Higher frequencies, >1 Hz (short periods), are due

to thunderstorm activity near the Equator and are distributed as guided waves, known as sferics, between the ionosphere and the Earth to higher latitudes.

MT signals are customarily measured in the frequency range downwards from 400 Hz. Typically, each MT station is deployed for recording one day and picked up the following day. This gives about 20 h of continuous time series per site, and MT data in the range from 400 Hz (0.0025 s) to about 1000 s. The short-period MT data (high frequency) mainly reflect the shallow structures due to their short depth of penetration, whereas the long-period data mainly reflect the deeper structures. Data qualities are to be inspected before moving to the next site and the measurement is redone for another day, if the data qualities are unacceptable – in particular, if the reason is found.

Following the data acquisition, the digitally recorded time series are Fourier transformed from the time domain into the frequency domain, cross- and auto-powers of the fields are calculated to give the apparent resistivity and phase as a function of the period/frequency of the electromagnetic fields. The 'best' solution that describes the relation between the electrical and magnetic field is found through the following equation:

$$\begin{bmatrix} E_x \\ E_y \end{bmatrix} = \begin{bmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{bmatrix} \begin{bmatrix} H_x \\ H_y \end{bmatrix} \tag{3}$$

or in matrix notation:

$$\vec{E} = Z\vec{H} \tag{4}$$

where E and H are the electrical and magnetic field vectors (in the frequency domain), respectively, and Z is a complex impedance tensor which contains information on the subsurface resistivity structure. The values of the impedance tensor elements depend on the resistivity structures below and around the site. For a homogeneous and 1D Earth, $Z_{xy} = -Z_{yx}$ and $Z_{xx} = Z_{yy} = 0$. For a 2D Earth, that is, resistivity varies with depth and in one horizontal direction, it is possible to rotate the coordinate system by mathematical means, such that $Z_{xx} = Z_{yy} = 0$, but $Z_{xy} \neq -Z_{yx}$. For a 3D Earth, all the impedance tensor elements are different.

From the impedances, the apparent resistivity (ρ) and phases (θ) for each period (T) are calculated according to the following equations:

$$\rho_{xy}(T) = 0.2T|Z_{xy}|^2; \ \theta_{xy} = \arg(Z_{xy}) \tag{5}$$

$$\rho_{yx}(T) = 0.2T|Z_{yx}|^2; \ \theta_{yx} = \arg(Z_{yx}) \tag{6}$$

As noted above, $Z_{xy} = -Z_{yx}$, for a homogeneous and 1D Earth, and hence, the xy and yx parameters, ρ_{xy} and ρ_{yx}, and θ_{xy} and θ_{yx} are equal.

The depth of penetration of MT soundings depends on the wavelength of the recorded EM fields and the subsurface resistivity structure. The longer the period T, the greater is the depth of penetration and vice versa. The relation is often described by the skin depth or the penetration depth (δ), which is the depth where the EM fields have attenuated to a value of e^{-1} (about 0.37) of their surface amplitude.

$$\delta(T) \approx 500\sqrt{T\rho} \ \ (m) \tag{7}$$

where ρ is the average resistivity of the subsurface down to that depth.

The xy and yx parameters, both resistivity and phase, are seldom the same, and in cases where they are not, they depend on the orientation of the setup of the measurement. In 1D inversion (layered Earth or Occam inversion), it is possible to invert for either xy or yx parameters, and there have been different opinions through the years which one to use. Nowadays, it is becoming more customary to invert for some rotationally invariant parameter, that is, independent of the sounding setup, which is defined in such a way that it averages over directions. Therefore, one has not to deal with the question of rotation. Three such invariants exist:

$$Z_B = \frac{Z_{xy} - Z_{yx}}{2} \tag{8}$$

$$Z_{det} = \sqrt{Z_{xx}Z_{yy} - Z_{xy}Z_{yx}} \tag{9}$$

$$Z_{gm} = \sqrt{-Z_{xy}Z_{yx}} \tag{10}$$

All these parameters give the same values for a 1D Earth response. For 2D, Z_{det} (determinant) and Z_{gm} (geometric mean) reduce to the same value, but Z_B (arithmetic mean) is different. For 3D responses, all these parameters are different.

There are different opinions on which of the three invariants, if any, is best suited for 1D inversion. However, based on the comparison of model responses for 2D and 3D models, it has been suggested that the determinant invariant is the one to use in 1D inversion (Park and Livelybrook, 1989) The apparent resistivity and phase calculated from the rotationally invariant determinant of the MT impedance tensor as a function of the period is shown in Figure 7.

The MT method, like all resistivity methods that are based on measuring the electric field on the surface, suffer from the so-called telluric or static shift problem (Árnason et al., 2010; Árnason, 2008; Sternberg et al., 1988). It is caused by resistivity inhomogeneity close to the electric dipoles. There are mainly three processes that cause static shift, that is, voltage distortion, topographic distortion, and current channeling (Jiracek, 1990). Voltage distortion is caused by a local resistivity anomaly, resistivity contrast at the surface. Topographic distortion takes place where the current flows into the hills and under the valleys, affecting the current density, and current distortion (current channeling) is where the current flow in the ground is deflected when encountering a resistivity anomaly. If the anomaly is of lower resistivity than the surroundings, the current is deflected (channeled) into the anomaly, and if the resistivity is higher, the current is deflected out of the anomaly. If the anomaly is close to the surface, this will affect the current density at the surface and hence the electric field. Like for the voltage distortion, this effect is independent of the frequency of the current.

These three phenomena are common in geothermal areas in volcanic environment where the surface consists of resistive lavas. Geothermal alteration and weathering of minerals can produce patches of very conductive clay on the surface surrounded by very resistive lavas, producing severe voltage distortion. Similarly, if the conductive clay minerals dome up to shallow depth but not quite to the surface, they can result in extensive current channeling.

The problem is that the amplitude of the electric field on the surface and, consequently, the apparent resistivity is scaled by an unknown dimensionless factor (shifted on log scale) due to the resistivity heterogeneity in the vicinity of the measuring dipole. Static shift can be a big problem in volcanic geothermal areas where resistivity variations are often extreme. Shift factors of the apparent resistivity have been observed as low as 0.1, leading to 10 times too low-resistivity values and about 3 times too small depths to resistivity boundaries (Árnason, 2008). In the central-loop TEM method, the measured signal is the decay rate of the magnetic field from the current distribution induced by the

current turn-off in the source loop, not the electric field. At late times, the induced currents have diffused way below the surface and the response is independent of near-surface conditions.

Therefore, TEM soundings and MT soundings can be jointly inverted in order to correct for the static shift of the MT soundings. The shift multiplier may be used in multidimensional inversion of MT soundings, and has been proven in high-temperature geothermal areas to be a necessary precondition. MT data cannot be used to correct themselves for static shifts. Interpretation of MT data without correction by TEM cannot be trusted except, may be, in areas where it is known that little or no near-surface inhomogeneity is present (e.g., thick homogeneous sediments). Figure 7 shows an example of a 1D joint Occam inversion of TEM and MT data.
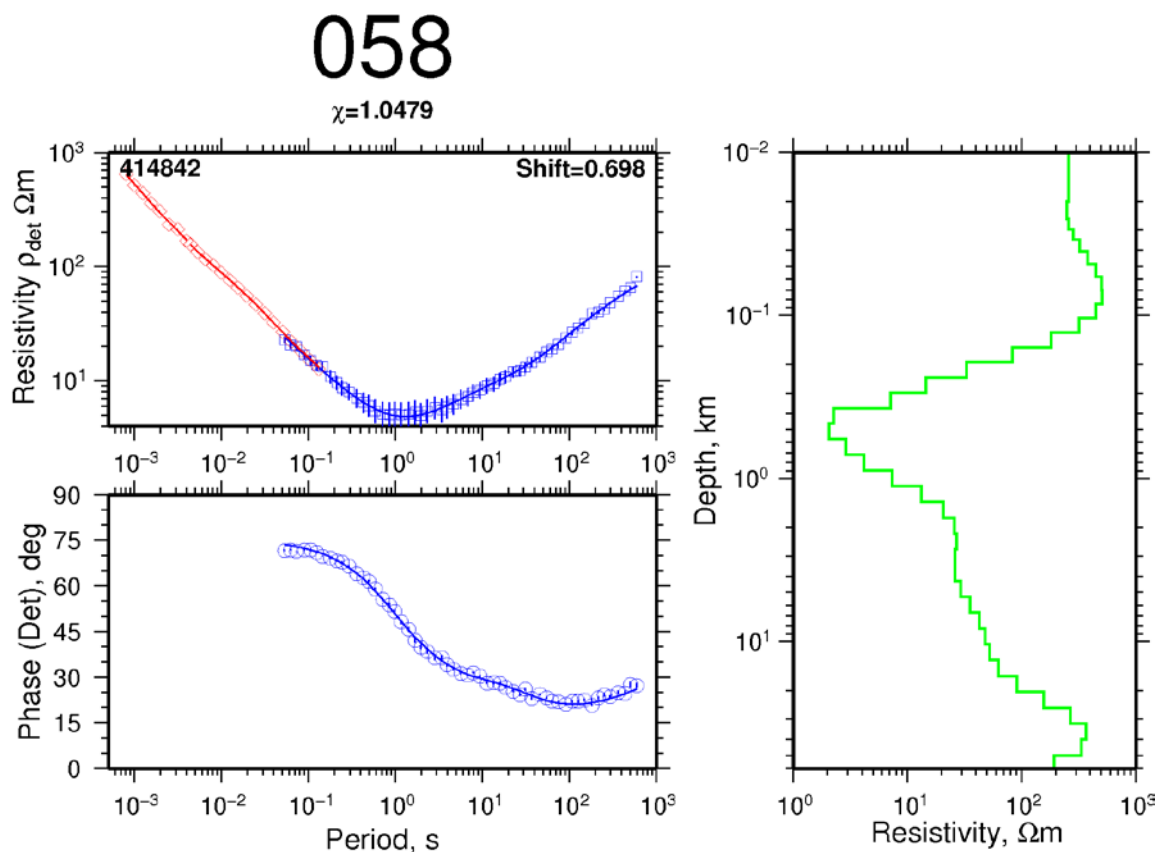

FIGURE 7: Joint 1D Occam inversion of TEM and MT data for sounding 058 from high-temperature geothermal area Krýsuvík, Southwest Iceland (Hersir et al., 2010). Red diamonds: TEM apparent resistivity transformed to a pseudo-MT curve; blue squares: measured apparent resistivity; blue circles: apparent phase – both derived from the determinant of MT impedance tensor; green lines: on the right are results of the 1D resistivity inversion model and to the left are its synthetic MT apparent resistivity and phase response. Note the shift value being as low as 0.698 for the MT data to tie in with the TEM data. The misfit function; the root-mean-square difference between the measured and calculated values is $\chi = 1.0479$.

Besides 1D inversion, MT data can also be inverted in 2D and particularly in 3D, which in recent years is becoming more practical (Siripunvaraporn, 2010). In 3D inversion, the responses from the 3D model should fit reasonably well with the data from all the MT soundings from the modeled area, both xy and yx parameters. The data should be static shift-corrected prior to the inversion. Figure 8 shows an example of 1D inversion versus 3D inversion. Clearly, 1D inversion reproduces the basic resistivity structures but smears them out, whereas the 3D inversion sharpens the picture considerably. Additional information from other investigations has been added to the figure to ease further geothermal interpretation.
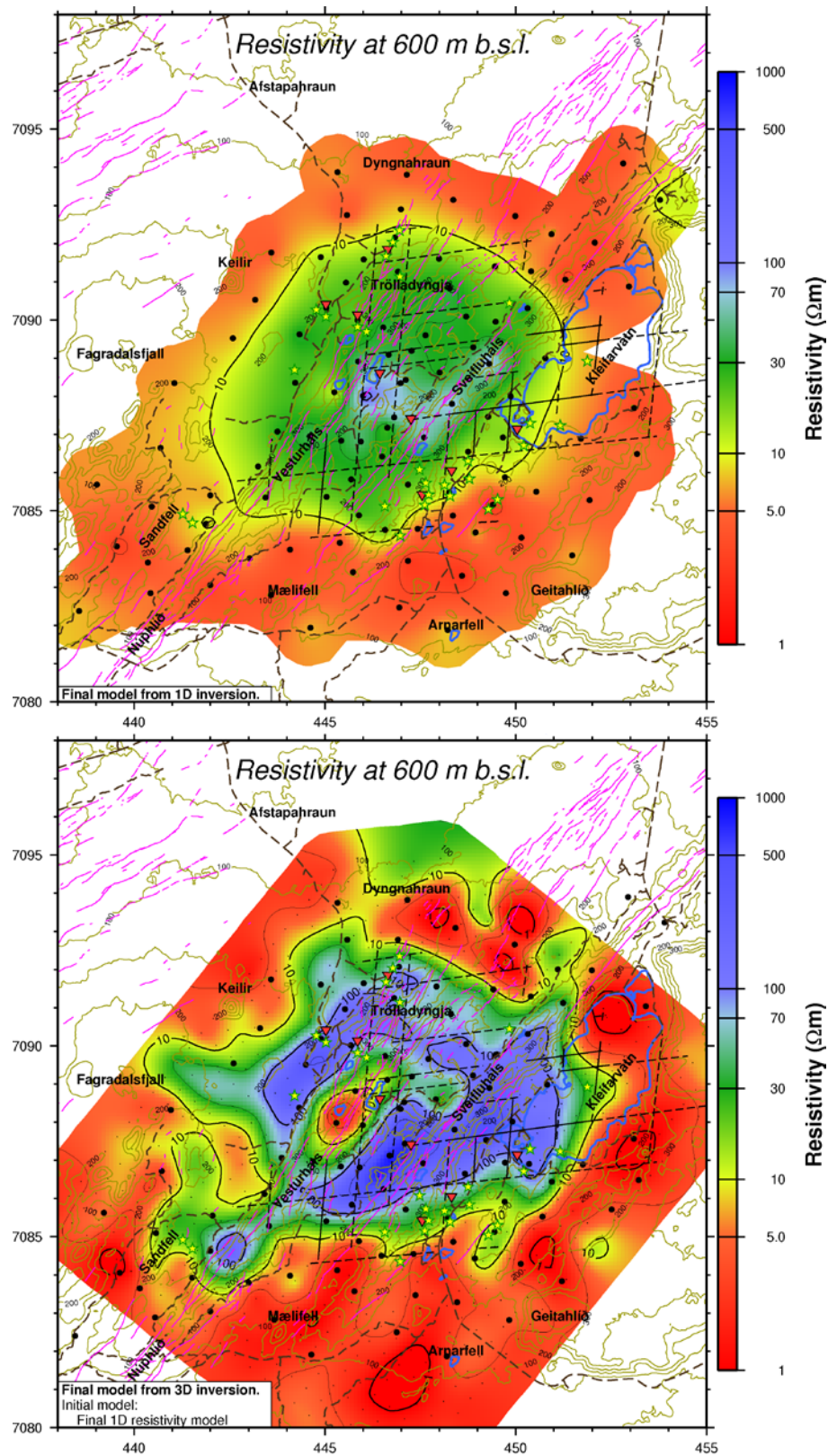
FIGURE 8: Resistivity models from Krýsuvík, Southwest Iceland (Hersir et al., 2013): based on 1D model (upper panel), and 3D model using the 1D model as an initial model in the 3D inversion (lower panel). Black dots are MT soundings, wells are denoted by red-filled triangles, fumaroles by green/yellow stars, and fractures and faults by magenta lines. Solid and broken black lines show possible fracture zones, inferred from seismicity.

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