Resistivity Surveying in Geothermal Exploration with an Application to the Eyjafjordur Low-Temperature Area, North Iceland

Sarantsetseg Lkhagvasuren[1], Gylfi Pall Hersir[2], Knutur Arnason[3], Malfridur Omarsdottir[4]

Institute of Astronomy and Geophysics of Mongolian Academy of Sciences, 7th Building of MAS, Lkhagvasuren Street-42, Bayanzurkh district, Ulaanbaatar, Mongolia[1]

sarantsetseg@iag.ac.mn[1]

IcelandGeoSurvey, Grensásvegur 9, 108 Reykjavík, Iceland[2,3]

gylfi.pall.hersir@isor.is[2]
ka@isor.is[3]

United Nations University Geothermal Training Programme, Grensásvegur 9, 108 Reykjavík, Iceland[4]

malfridur.omarsdottir@os.is[4]

Keywords: resistivity, geothermal exploration, magneto-tellurics, geophysical methods.

ABSTRACT

Geothermal exploration involves geology, geochemistry and geophysics. In geophysical exploration, resistivity surveying plays the most important role in delineating the reservoir. The parameters that control the geothermal system show a strong response to electrical resistivity. The resistivity methods that are mostly used in geothermal exploration in Iceland are TEM (Transient Electromagnetics) and MT (Magneto Tellurics). The application of these methods is discussed in this report together with an example from the Eyjafjordur low-temperature area in N-Iceland. The resulting resistivity cross-sections and resistivity depth slices show a shallow lying low-resistivity layer and a deep lying low-resistivity anomaly towards the end of the cross-sections. The result of this work has been compared with results from Flovenz and Karlsdottir (2000) which interpreted TEM data from the same area. The results are also compared with borehole data and stratigraphy.

1. INTRODUCTION

Geothermal manifestations are commonly seen along plate boundaries and faults. A good example is Iceland which is sliced or crossed by the North America and Eurasian plate's boundaries. Geothermal systems are controlled by temperature, pressure, porosity, permeability and chemical composition of the fluid. These parameters give a response to electrical resistivity. Because of this, the resistivity methods are the most powerful method to find and delineate geothermal areas. The resistivity model of the Eyjafjordur low temperature geothermal area is presented through 1D joint inversion of the MT (Magnetotelluric) and TEM (Transient electromagnetic) data. The main part of this project work is the application of the MT and TEM methods in exploration of low-temperature geothermal areas. In TEM, an electrical current is injected into a source loop and then turned off abruptly. Afterwards, voltage is registered in a receiver loop as a function of time. MT uses natural fluctuations of the Earth´s electromagnetic field – the electric and magnetic fields are registered in orthogonal directions. Data processing gives an apparent resistivity curve and a resistivity model of the subsurface is obtained from an inversion process. Finally, the model is used for geothermal interpretation. These methods have been used effectively in geothermal exploration in Iceland for decades. Finally, the theory and application of the MT and TEM methods, respectively, are explained.

2. TEM (TRANSIENT ELECTROMAGNETIC) METHOD

The TEM method is an active method used widely in geothermal exploration in Iceland. A constant current is injected into a transmitter loop from a transmitter that is fed by a generator or batteries. According to Biot-Savart´s law, a constant magnetic field of known strength is created (Figure 1).

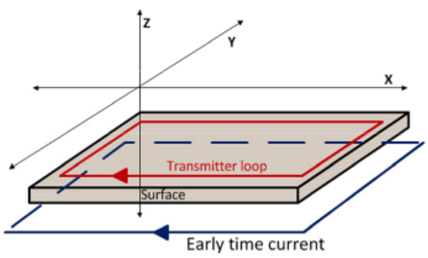

Figure 1: Current propagation at early times (from Badilla, 2011; modified from Rowland, 2002).

A receiver loop (big flexible loop and small coil) is connected to the receiver and placed at the center of the transmitter loop. Then the current is abruptly turned off. The current and magnetic field decrease downwards and outwards and the magnetic field induces a current in the resistive media that the current travels through at late times. (Figure 2).

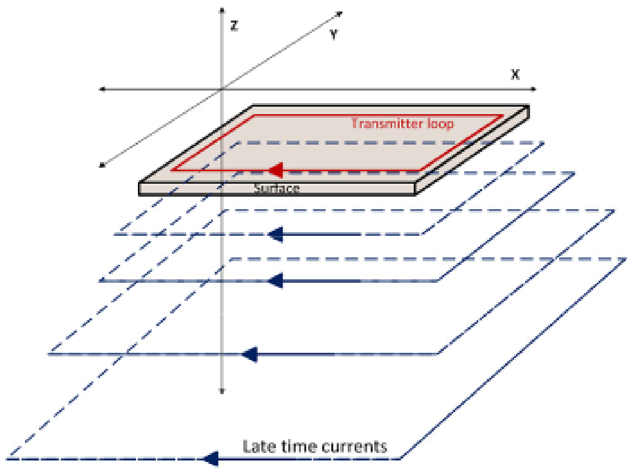

Figure 2: Current propagation at late times (from Badilla, 2011; modified from Rowland, 2002).

After the turn off (the current turns off linearly) measurements start. During the decay of the magnetic field, a current is induced in the ground. According to Faraday's law, induced voltage in the receiver coil creates a decaying secondary magnetic field. The rate and magnitude of the decaying secondary magnetic field depends on the resistivity of the ground. Resistivity of the ground can be estimated by measuring the voltage that is registered in the receiver coil. Measurements are done in the big flexible loop (effective area of 57,000 m^2) and small loop (effective area of 100 m^2) using a frequency of 2.5 Hz and 25 Hz, respectively. The current and frequency are fed by operator into the equipment. The turn off time is written down in the field book together with other information. Generally, the source loop is 200 m x 200 m square and the transmitted current is 20-25 A.

2.1 Depth of penetration

The TEM method depends on the resistivity beneath the sounding as well as on the setup geometry, the generated current and its frequency. The depth of penetration increases with time after the current turn-off (Flóvenz et al., 2012). The depth of penetration in the central loop TEM sounding depends on how long the induction in the receiver coil can be traced before it is drowned in noise. The induced voltage in the receiver coil in a homogeneous half space of conductivity σ at late time is given as (Árnason, 1989):

$$V(t,r) = I_0 \frac{C\,(\mu_0 \sigma r^2)^{3/2}}{10\pi^{1/2}\, t^{5/2}} \tag{1}$$

Where, $C = A_r n_r A_s n_s \frac{\mu_0}{2\pi r^3}$

n_r = Number of windings in the receiver coil;
A_s = Area of the transmitting loop (m^2);
n_s = Number of windings in transmitting loop;
A_r = Area of the receiver coil (m^2);
t = Time elapsed after the current in the transmitter is turned off (s);
μ_0 = Magnetic permeability in vacuum (Henry/m);
$V(t,r)$ = Induced voltage (V);
r = Radius of the transmitter loop (m)
I_0 = Current in the transmitting loop (A)

The apparent resistivity can be derived from Equation 1.

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

Using Equation 2, the apparent resistivity curve is calculated as a function of time after the current is turned off.

3. MT (MAGNETOTELLURIC) METHOD

MT is a passive electromagnetic method that uses the natural magnetic field which provides a broad range of frequency between 10^{-4} and 10^3 Hz. The magnetic field changes with time and a corresponding electric field changes in the surface of the ground is measured to reveal subsurface resistivity distribution for great depth ranges. The signal sources are natural fluctuations of the Earth´s magnetic field. Those fluctuations or primary magnetic field (H_p) induce a current in the ground (Keary et al., 2002). A local conductive structure effects the density and distribution of the eddy currents. The current induces a secondary magnetic field (H_s). The resulting magnetic field H=($H_p + H_s$) is measured with induction coils in horizontal and orthogonal directions (H_x, H_y *and* H_z). The accompanied electric field (E_x and E_y) using voltmeter as a potential difference ΔU, between pairs of electrodes at distance L at the surface $\boldsymbol{E} = \Delta U/L$. The small amplitude geomagnetic time variations of the Earth´s field contain a wide spectrum, generated by two different sources. One is a low frequency signal (< 1 Hz, long period) which is created from ionospheric and magnetospheric currents caused by solar winds radiated from the sun and interacting with the earth´s magnetic field known as micropulsations. This is used for deep crustal and upper mantle structure investigations. The other one is high frequency (>1 Hz, short period) signal that is generated by thunderstorm activity near the equator, used for shallow crustal structure study. The magnetic **H** and electric field **E** are measured at the surface and reveal the apparent resistivity ρ_a as a function of ω (2π/T).

$$\boldsymbol{E}(w) = Z(w) \cdot \boldsymbol{H}(w) \quad (3)$$

Where Z is a tensor that depends on resistivity structure of the ground and ω is the angular frequency (Hz) (Hersir and Björnsson, 1991). The tensor equation can be written as:

$$\begin{bmatrix} E_x \\ E_y \end{bmatrix} = \begin{bmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{bmatrix} \begin{bmatrix} H_x \\ H_y \end{bmatrix} \quad (4)$$

For a 1D earth, conductivity changes with depth only. Therefore, the diagonal elements of the impedance tensors, Z_{xx} and Z_{yy} are equal to zero, while the off-diagonal components are equal in magnitude, but have opposite signs:

$$Z_{xy} = -Z_{yx} \quad (5)$$

From the impedances, the apparent resistivity of the earth ρ and phases θ for each period T are calculated by the following relationships:

$$\rho_{xy}(T) = 0{,}2T\left|Z_{xy}\right|^2 = 0.2T\left|\frac{E_x}{H_y}\right|^2; \quad (6)$$

$$\theta_{xy} = \arg(Z_{xy})$$

$$\rho_{yx}(T) = 0{,}2T\left|Z_{yx}\right|^2 = 0.2T\left|\frac{E_y}{H_x}\right|^2; \quad (7)$$

$$\theta_{yx} = \arg(Z_{yx})$$

As mentioned above, $Z_{xy} = -Z_{yx}$, for a homogeneous and 1D Earth as well as $\rho_{xy} = \rho_{yx}$ (Flóvenz et al., 2012). In the 1D inversion in this work, the determinant of the impedance tensor, which is rotationally invariant (the value of the resistivity does not change with rotation) are used. The determinant value is calculated by the following equation:

$$\rho_{det} = \frac{1}{w\mu_0}\left|Z_{det}\right|^2 = \frac{1}{w\mu_0}\left|\sqrt{Z_{xx}Z_{yy} - Z_{xy}Z_{yx}}\right|^2; \quad \theta_{det} = \arg(Z_{det}) \quad (8)$$

The determinant of the impedance tensor is like an average value of the apparent resistivity (see Figure 8).

3.1 Skin depth

Skin depth is determined as a depth where the electromagnetic fields have been reduced to e^{-1} of their original value at the surface. It is used like a scale length for the time changing field or an estimate of how deep such a wave penetrates into the earth.

$$\delta = 500\sqrt{\rho T} \quad (9)$$

Where δ – skin depth (m); T – period (s); ρ – resistivity (Ωm)

The skin depth depends on the resistivity and the period. Consequently, a low frequency signal penetrates deeper into the earth than the high frequency signal (Hersir and Björnsson, 1991).

3.2 Static shift

All resistivity methods that measure the electric field on the surface suffer the static shift or telluric shift problem that is manifested as an unknown multiplier of the apparent resistivity. In MT, the electrical field (**E**=ρ**j**) is measured at the surface. It is affected by current distortion and topography. When voltage difference is registered on the surface, and there is a high resistivity body in the layer, the current always flows through the low-resistivity part. It causes a distortion of current, assuming that the current flows along the horizontal layers. Because of the shape of the topography, at high elevation, current density is low and the curve is shifted downwards but at the low elevation, it would be high, and it is shifted upwards. Static shift problems of MT are fixed by comparing the MT data with a nearby TEM resistivity curve when doing joint inversion. The TEM method is only sensitive to the near surface resistivity structure and topography at early times. On the other hand, the TEM data collected are measured during late times.

4. TEM AND MT DATA ACQUISITION

4.1 TEM data acquisition

In central loop TEM soundings, the transmitter is fed by a generator or batteries and connected with a transmitter loop, which is usually a square with a variable effective area of 40,000-90,000 m^2 (but rarely less than 10,000 m^2 for shallower probing). Around 20–25 A current is injected from the transmitter into the transmitter loop. Additionally, even if the current is less than 10 A, one can get data with lower signal-to-noise ratio (SNR). The receiver is connected with the receiver loop that is 10 x 10 m square loop with several windings and circular loop of 1 m^2 with 100 windings (effective area is 100 m^2), respectively. Those receiver loops are used at 2.5 Hz while a small loop is used at 25 Hz or even higher frequency. The receiver loop is placed at the center of the transmitter loop.

Initially, the TEM receiver instrument is calibrated using a known signal source. Geonics PROTEM digital receiver can do its calibration automatically (Geonics, 1999). Time synchronization between the receiver and transmitter must have done precisely before the measurement is carried out because TEM works with very rapid signal transition. There are two ways to do synchronization, either through a reference cable or crystal clocks, which were used in this case.

4.2 MT data acquisition

In MT measurement, two lines are placed perpendicular to each other for the purpose of measuring the electric field (E_x and E_y) oriented N-S and E-W. The lines are connected to a data logger and an electrode that is usually composed of lead – chloride with a porous ceramic bottom for measuring telluric currents. The electrode must be in good contact with the ground. Magnetic coils are used to measure the magnetic field in three orthogonal directions (H_x, H_y and H_z). MT measurement setup is sketched in Figure 3.

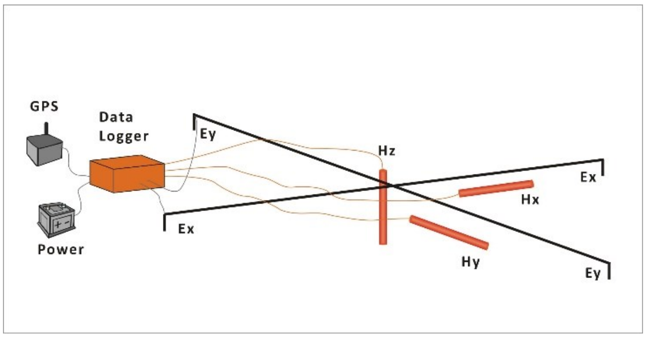

Figure 3: The setup of an MT sounding: Electrodes for measuring the electric field, coils for the magnetic field, acquisition unit for digital recording and GPS for synchronizing the data (Flóvenz et al., 2012)

For a homogeneous or layered earth, the electrical field is induced by its orthogonal source magnetic field (E_x correlates with H_y and E_y with H_x) (Hersir and Árnason, 2013). In order to prevent noise from the random or unnecessary movements, the coils are embedded into the ground. The polarity of the coils must be taken into account (Phoenix Geophysics, 2009). After the setup is completed, one should write down the direction of the setup and the coils used for each line and the distance between electrodes as this information is needed for later data processing. The data logger is configured to record at least for one day. During night time the signal strength is stronger than during day time (Simpson and Bahr, 2005). Before the operator downloads the data from the MT units for processing, MT data quality check has to be carried out at the site location. Example of time series data is shown in Figure 4.

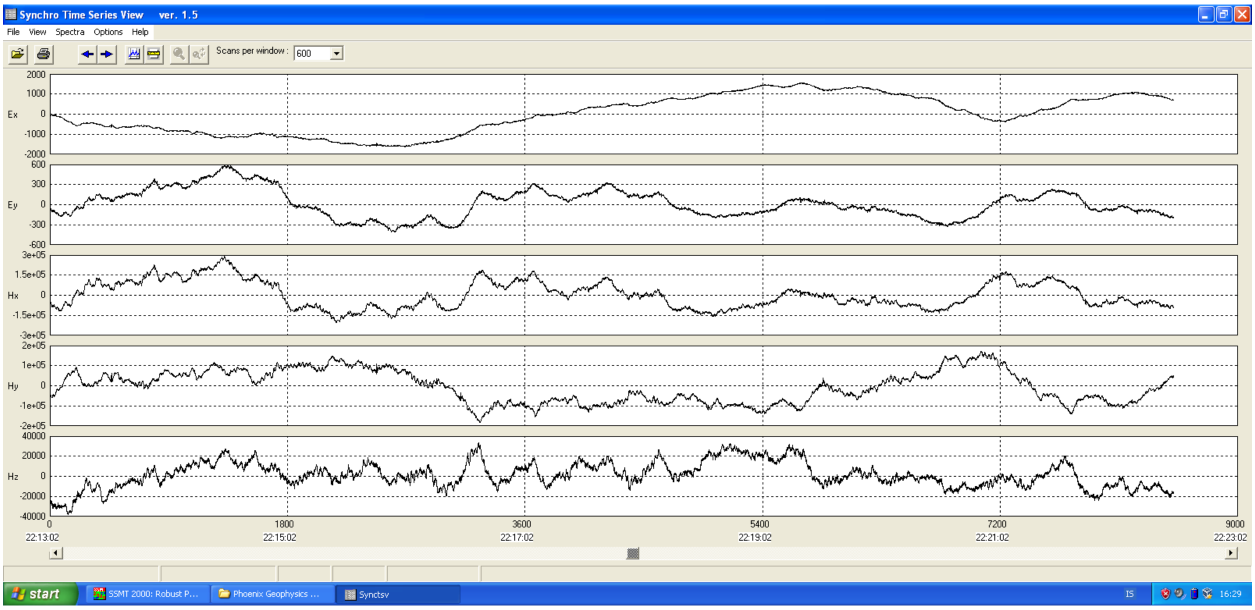

Figure 4: Time series data recording

Data in Figure 4 are from the study presented here. The time series data are viewed by time series viewer.

5. PROCESSING OF TEM AND MT DATA

5.1 TEM data processing

TEM raw data are (extension .fru) files, which contain the induced voltage values measured during the late times in 20 gates (Figure 5) in the TEM receiver for different frequencies (25 Hz and 2.5 Hz). The raw data are an input for TemX (Árnason, 2006b) which is a Linux-based software. It supports stacking of the measured values and excludes outliers. The format of the output file is a text file (extension .inv) that includes the apparent resistivity. During one period one can measure 2 times and for the 25 Hz frequency one will have 50 measurements that are stacked in the program and for 2.5 Hz frequency, 5 measurements are stacked.

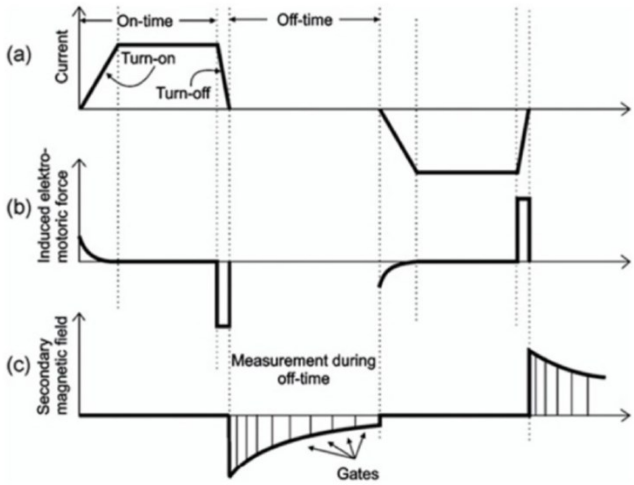

Figure 5: Time gates (Gichira, 2012)

5.2 MT data processing

Raw MT data are time series data that contain two electric and three magnetic field readings. Data processing is completed by SSMT2000 software provided by Phoenix Geophysics in Canada. The Fourier transform in its various forms decomposes a time series data into its frequency components. In MT, raw time series data are processed using calibration and site parameter files. The Fourier coefficients are calculated and then reprocessed with data from base stations, using robust routines. The output MT plot file contains multiple cross powers for each of the frequencies analysed (Africa, 2013). MTEditor is used for iterative weighting of residuals to identify and exclude points. Finally, the resistivity and phase curves are plotted. Now we have data that is Fourier transformed and minimized misfit. Then, there are some parameters of the plot files which are displayed, such as tipper magnitude, coherency between channels, and strike direction (Figure 10). These files are converted into industry-standard Electrical Data Interchange (edi) format (SEG, 1991) for the TEMTD (Árnason, 2006b).

6. INVERSION OF THE DATA

The inversion problem consists of obtaining physical parameter that can explain the measured values. Modelling of resistivity soundings is achieved using forward modelling and, thereafter, an inversion process. In forward modelling, an apparent resistivity curve is calculated from a guessed initial model by the geophysicist. After the forward calculation, the measured and the calculated data are compared. Then, the model is changed and the process repeated until the best solution is found. Inversion works with data and initial model to calculate the best solution of the forward calculation. The initial model is gradually enhanced through an iterative process by calculating adjustments to the model from the difference between the measured data and the response of the model (Figure 6), until a satisfactory fit is reached.

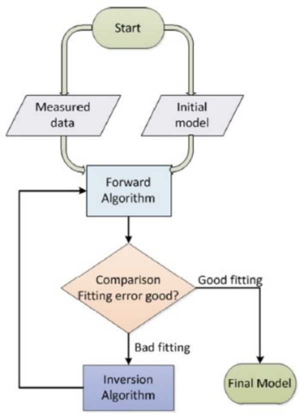

Figure 6: The inversion process

7. CASE EXAMPLE – EYJAFJORDUR LOW TEMPERATURE AREA, NORTH ICELAND

7.1 The study area

Iceland lies at the intersection of the Mid-Atlantic Ridge and the Wyville-Thomson Ridge, a seismically inactive transverse ridge crossing most of the north-eastern branch of the North Atlantic between East Greenland and the Faroe Islands. The North America and Eurasian plate boundary crosses Iceland from southwest to northeast. Active volcanoes and earthquakes are sign of the plate boundaries in Iceland, showing where active plate movement is taking place. Productivity of volcanic material has been shown to be anomalously high on this part of the mid-oceanic ridge system (Vogt, 1971).

Eyjafjordur is the longest fjord in Iceland. It is located in the central north of the country (Flóvenz and Karlsdóttir, 2000). This is the second most populous area of Iceland with the town of Akureyri in its center with almost 20,000 inhabitants. The mountains of Eyjafjordur are built of 3-10 m.y. old Tertiary volcanic formations. The rocks are mainly basaltic lavas with thin scoraceous and sedimentary inter-layers. The lava pile dips about 4-7°C to the south and southeast. Thick sedimentary beds are found locally. At least three extinct central volcanoes are buried in the basaltic lava pile around Eyjafjordur (Figure 7). Map of the Eyjafjordur area is presented in Figure 8 with locations of the TEM and MT soundings.

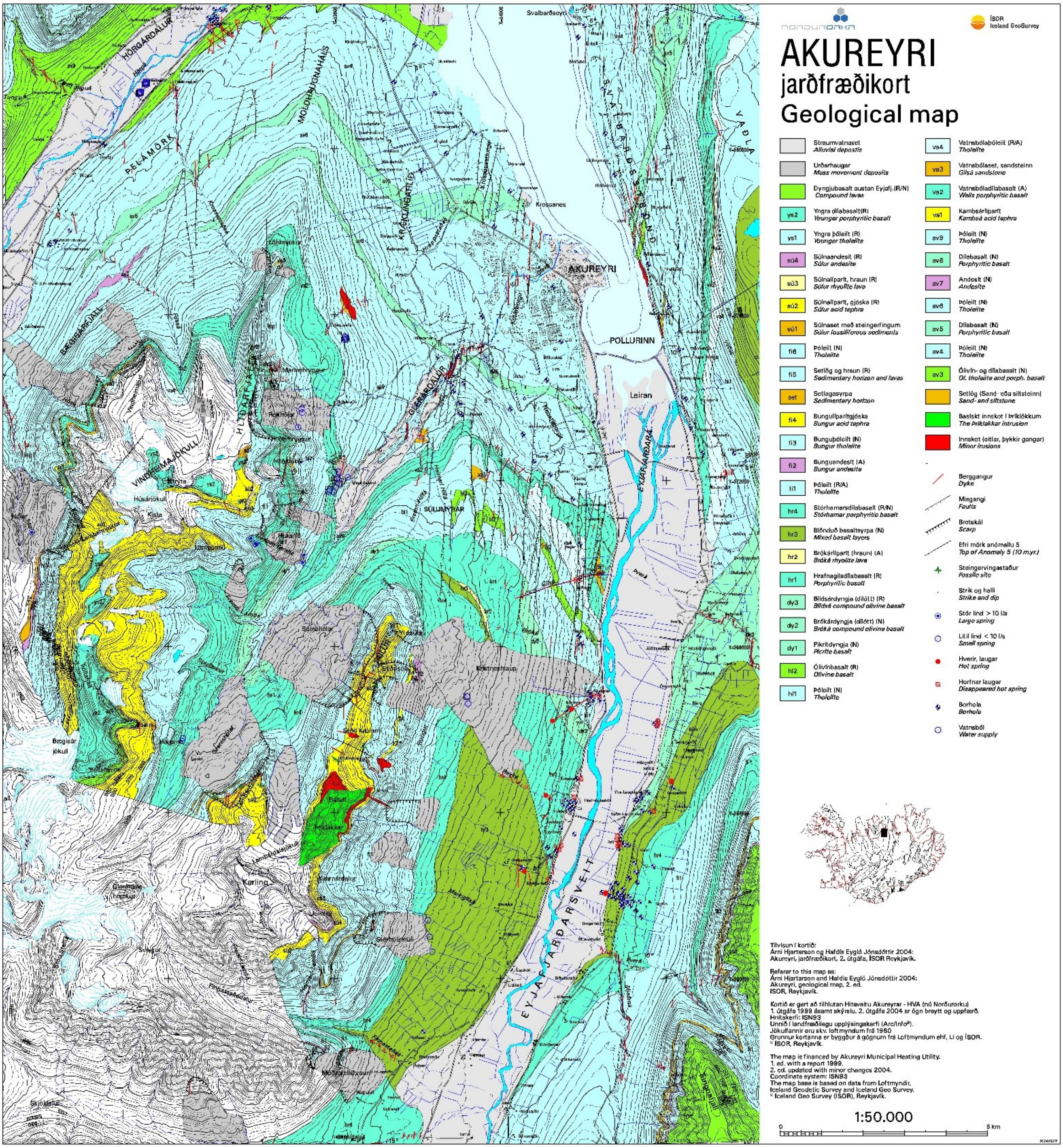

Figure 7: Geological map of the Eyjafjordur area (Hjartarson and Jónsdóttir, 2004)

7.2 TEM and MT soundings – data collection and processing

In framework of this work, 38 TEM soundings and 25 MT soundings are processed and inverted separately. The location of 23 of them are used here for the joint inversion **(**Figure 8). The stretched 2 lines shown in Figure 8 are locations of the resistivity cross-sections which were drawn down to different depths (see further Chapter 8.1).

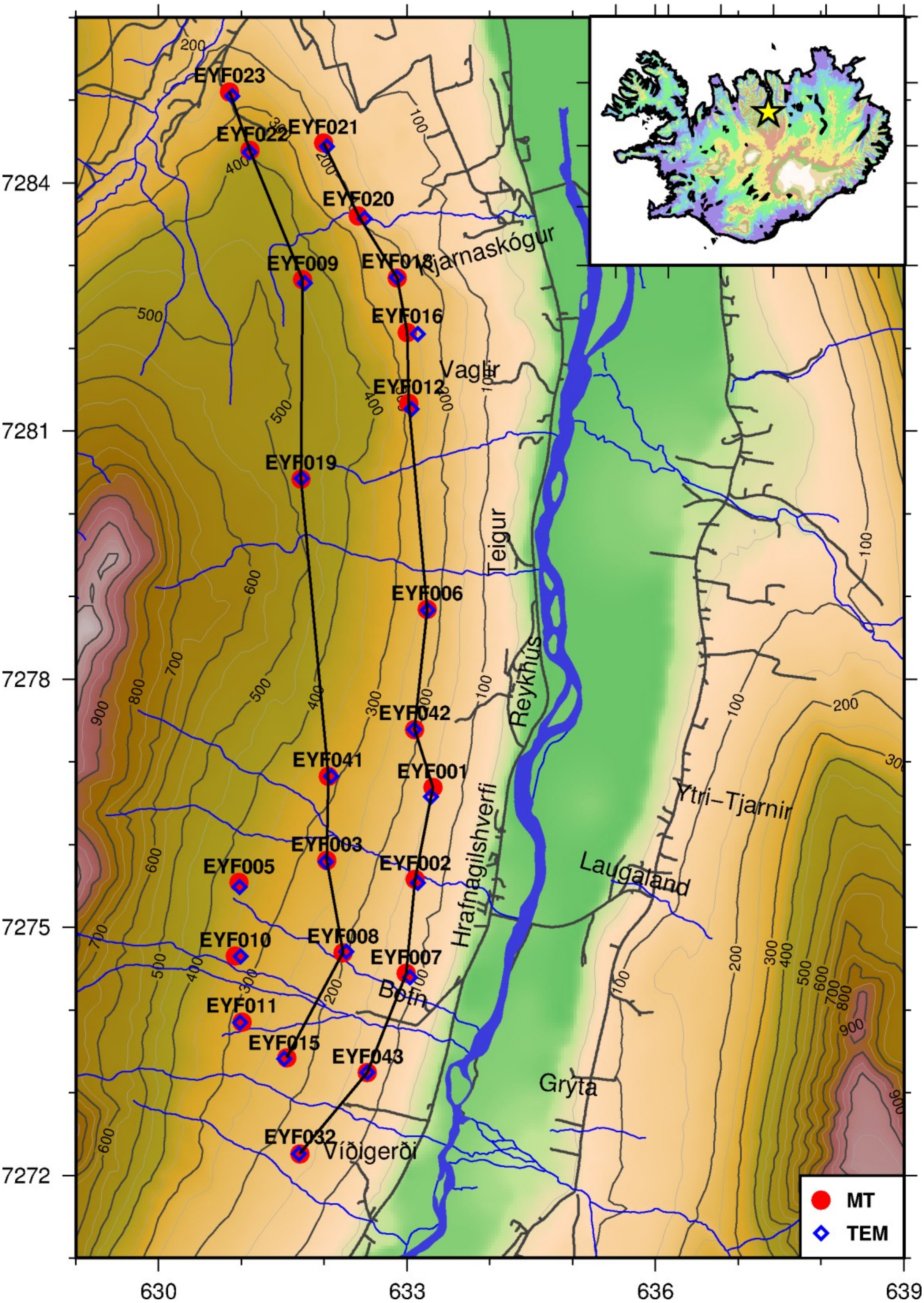

Figure 8: Location map for the TEM and MT soundings in Eyjafjordur. The yellow star on the Icelandic map on the right top corner of the figure shows the location of Eyjafjordur low temperature geothermal area; locations of two cross-sections are shown as black lines.

7.2.1 1D inversion of TEM data

Input files for the inversion program, TEMTD (Árnason, 2006a) are .inv files that contain the apparent resistivity and a guessed model of resistivity and layer-thickness values using a layered model. The purpose of the inversion is to obtain the true resistivity for each layer. The output file includes resistivity changes with depth (Figure 9**)**. The inversion algorithm that is used in the program is the non-linear least-squares inversion of the Levenberg-Marquardt type (Árnason, 2006a). The misfit function is the root-mean-square difference between measured and calculated values, weighted by the standard deviation of the measured values. In Occam inversion, which is used here in the inversion, layer thicknesses are kept fixed, equally spaced on a log scale, and the conductivity distribution is forced to be smooth by adjusting damping parameters.

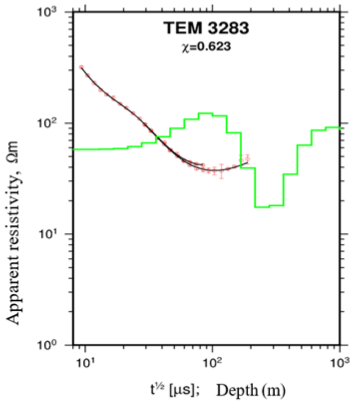

Figure 9: TEM inversion

7.2.2 1D inversion of MT data

The magnetotelluric sounding method for the determination of subsurface electrical conductivity was first proposed by Cagniard (1953). For an anisotropic or laterally inhomogeneous earth, the impedance becomes a tensor quantity. For the inversion program, the input is .edi-files from the MT soundings that includes impedances or apparent resistivity and phase values. In the 1D case, only the determinant of the impedance tensor is used which is rotationally invariant. Figure 10 presents the apparent resistivity and phase derived from the xy (red) and yx (blue) components of the impedance tensor and the determinant invariant (black), the Z-strike or Swift angle (black dots), and multiple coherency of xy (red) and yx (blue) and ellipticity (gray dots).

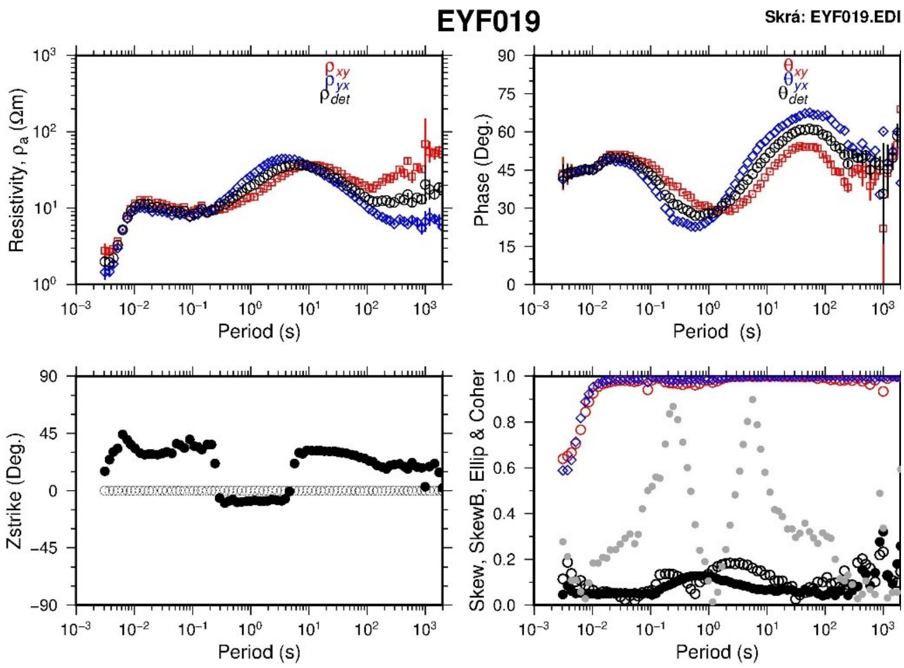

Figure 10: Processed data for MT sounding 019 from the Eyjafjordur geothermal low temperature area, North Iceland.

7.3 Joint inversion of TEM and MT data

For the joint inversion, the input is .inv files for the TEM data and .edi files for the MT data. Joint inversion of TEM and MT data is executed using the TEMTD program. For 1D inversion of TEM data the program inverts for time and voltage, and the apparent resistivity. The software is used to invert for the MT apparent resistivity and phase obtained from the rotationally invariant determinant of the tensor elements.

TEM soundings play an important role in correcting for static shift problem of MT data by jointly inverting both TEM and MT data. An example of the result of the jointly inverted TEM and MT data is presented in Figure 11.

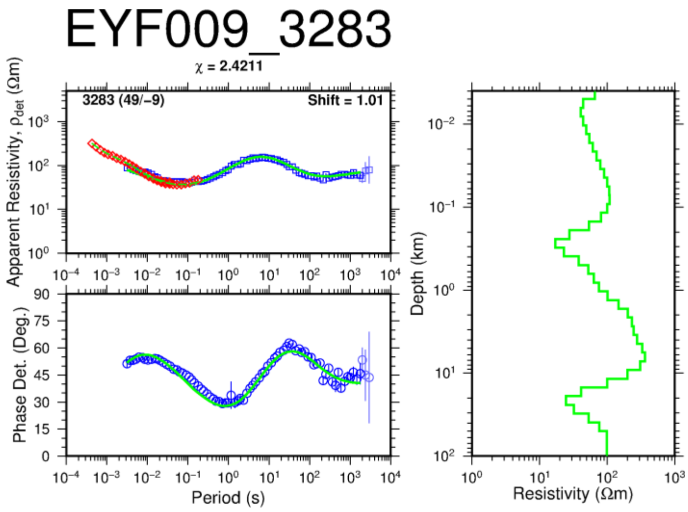

Figure 11: 1D joint inversion result of TEM sounding (3283) and MT sounding (EYF009). Here, red diamonds are TEM apparent resistivity points; blue squares are MT apparent resistivity and blue circles are MT apparent phase points which are derived from the determinant of MT impedance tensor, Green lines on the right represent 1D joint inversion model. On the top, the name of figure is called by Mt and TEM soundings that are located near from each other and misfit value of the model; The shift value executed to MT data in order to the TEM data is 1.01.

8. RESULTS

8.1 Resistivity cross-sections and resistivity depth slices

Figure 12 and Figure 13 show two resistivity cross-sections reaching down to different depths in the Eyjafjordur area, based on the 1D joint inversion of the TEM and MT data. Their location is given in Figure 8. A total of 8 resistivity depth slices are shown in Figure 14 and Figure 15.

The resistivity cross-sections show a shallow lying low-resistivity anomaly extending from the surface down to a depth of around 1,000 m. The low-resistivity is dipping upwards towards the north. At greater depth of several km, another low-resistivity body is seen in both cross-sections doming up in the northern part of the sections. These resistivity structures are even more visible in the depth slices. The deep lying conductive body is seen in one sounding only in the northwest part of the survey area at 7,000 and 10,000 m b.s.l. but extends over the northern part of the area at 15,000 m b.s.l. and in particular at 20,000 m b.s.l.

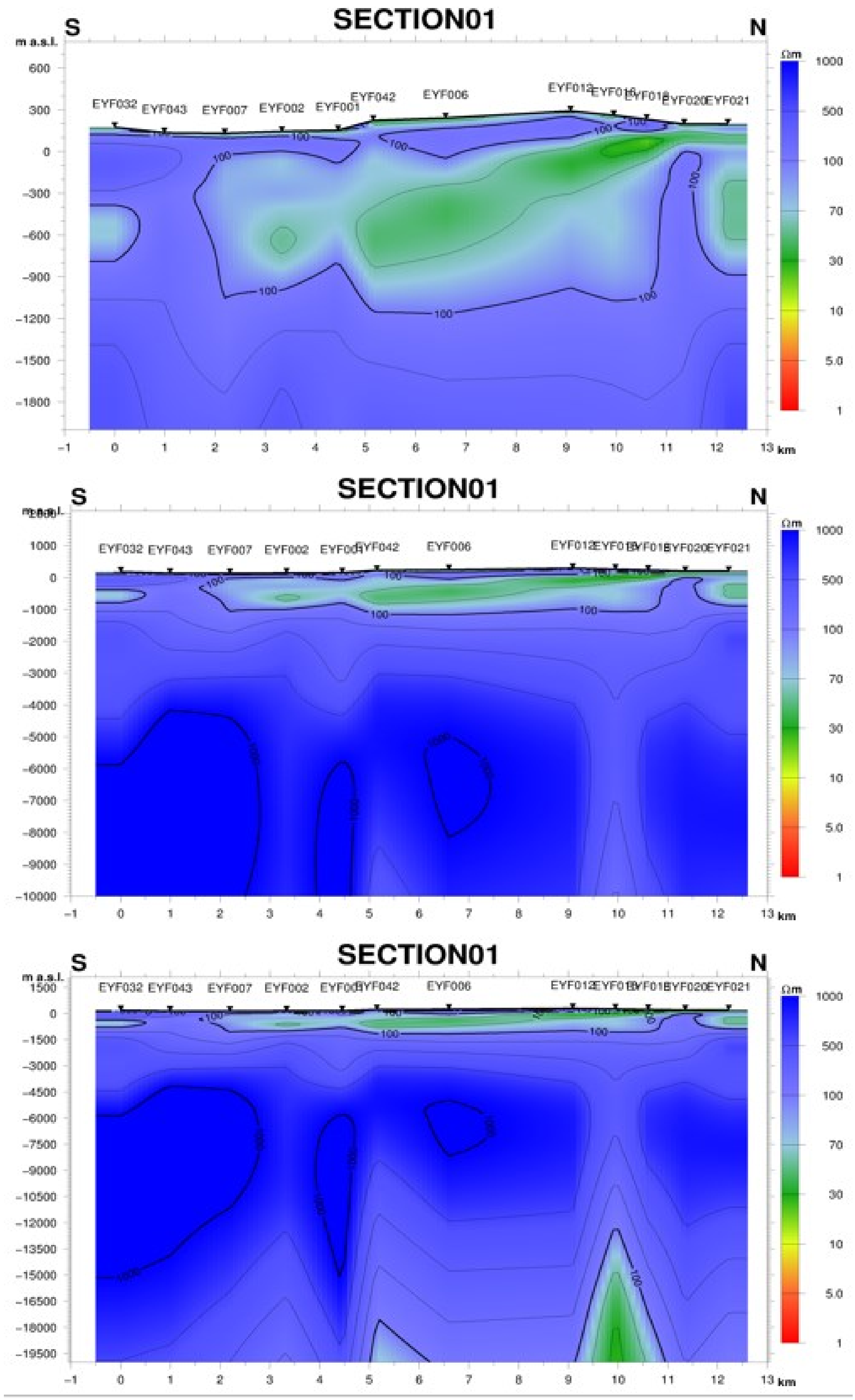

Figure 12: Resistivity cross-section N01 down to three depth levels, 2,000, 10,000 and 20,000 m b.s.l.; black triangles show the location of the MT soundings; the location of the cross-section is given in Figure 8.

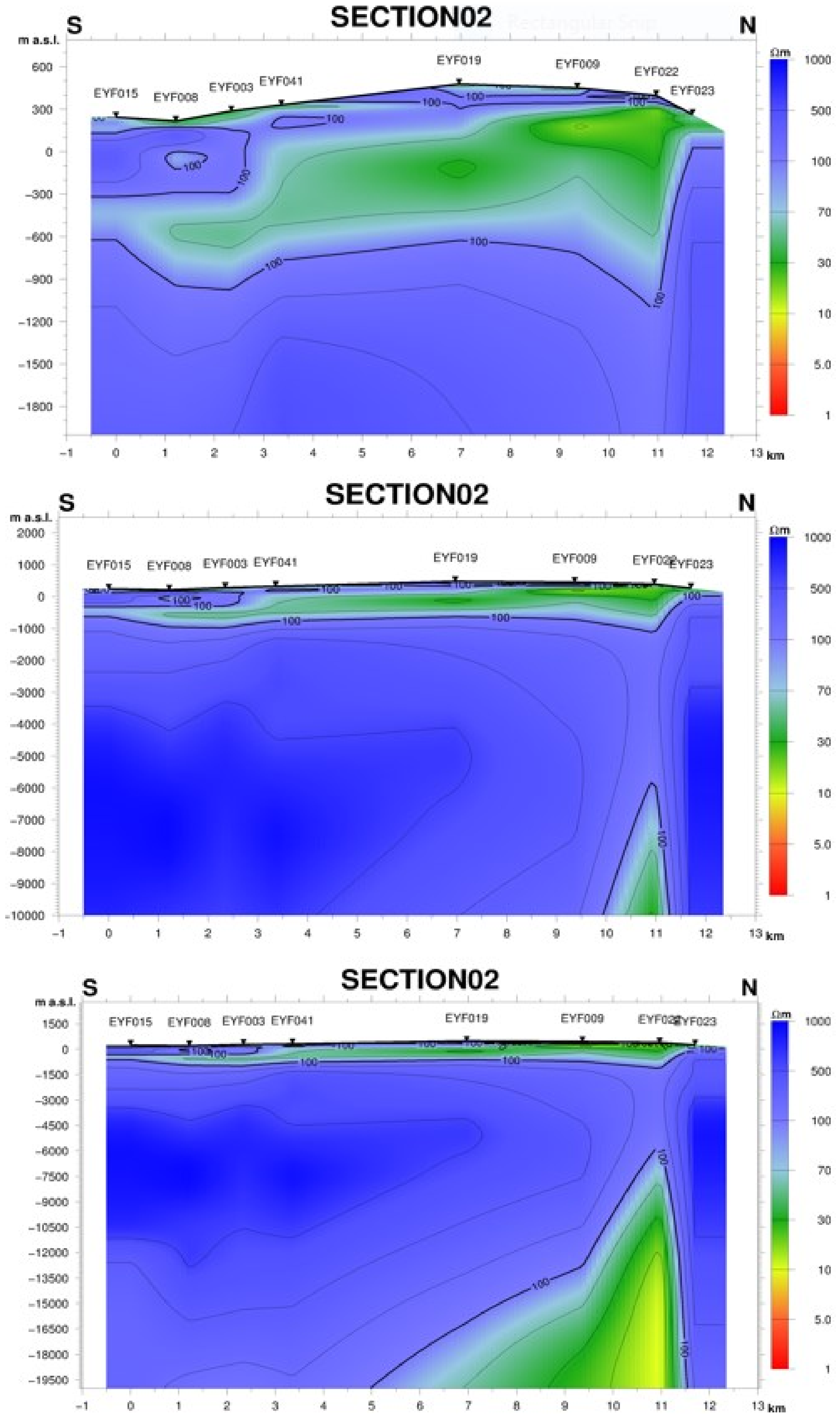

Figure 13: Resistivity cross-section N02 down to three depth levels, 2,000, 10,000 and 20,000 m b.s.l.; black triangles show the location of the MT soundings; the location of the cross-section is given in Figure 8.

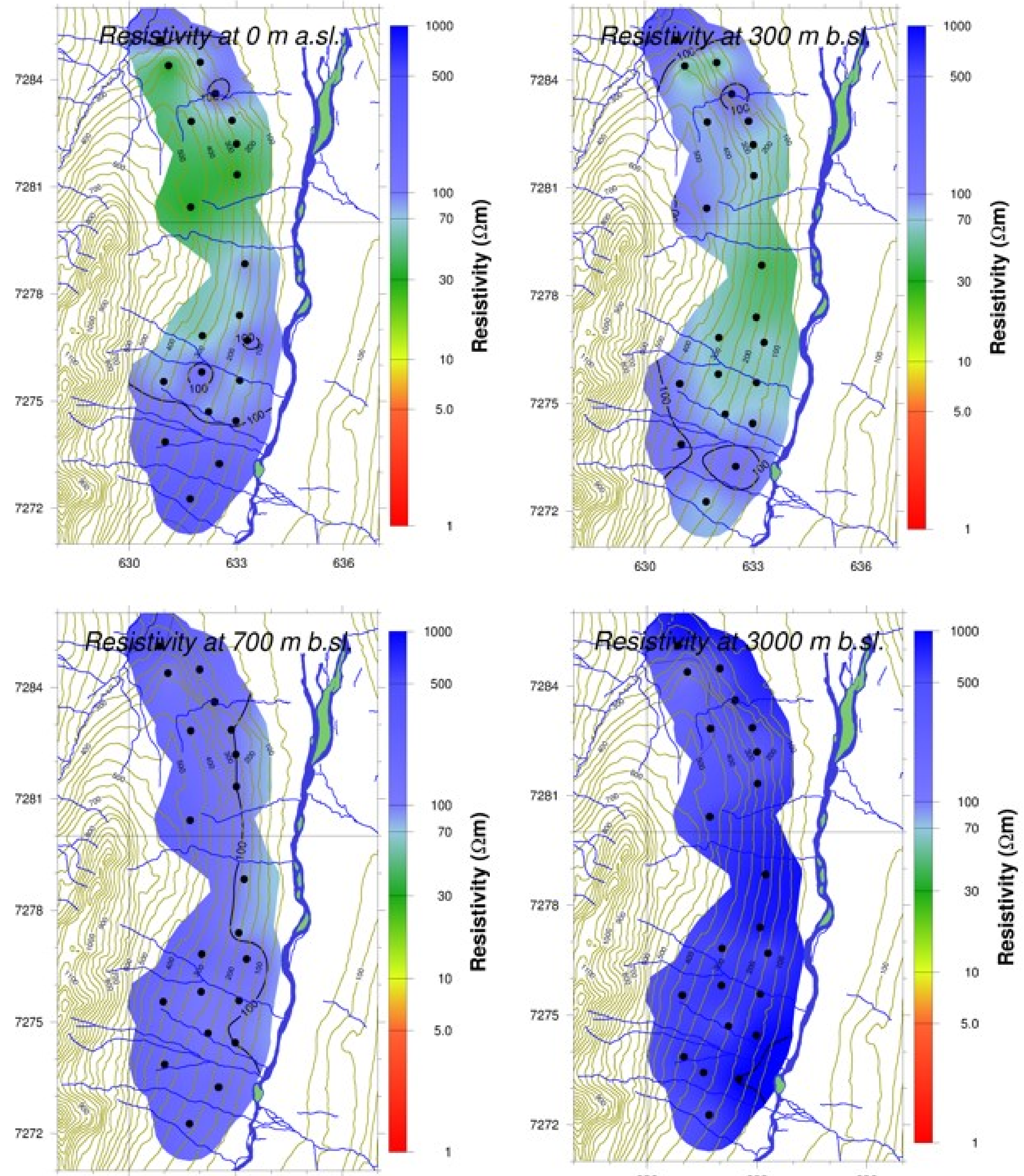

Figure 14: Resistivity depth slices at sea-level and at 300, 700 and 3000 m b.s.l.; black dots show locations of MT soundings.

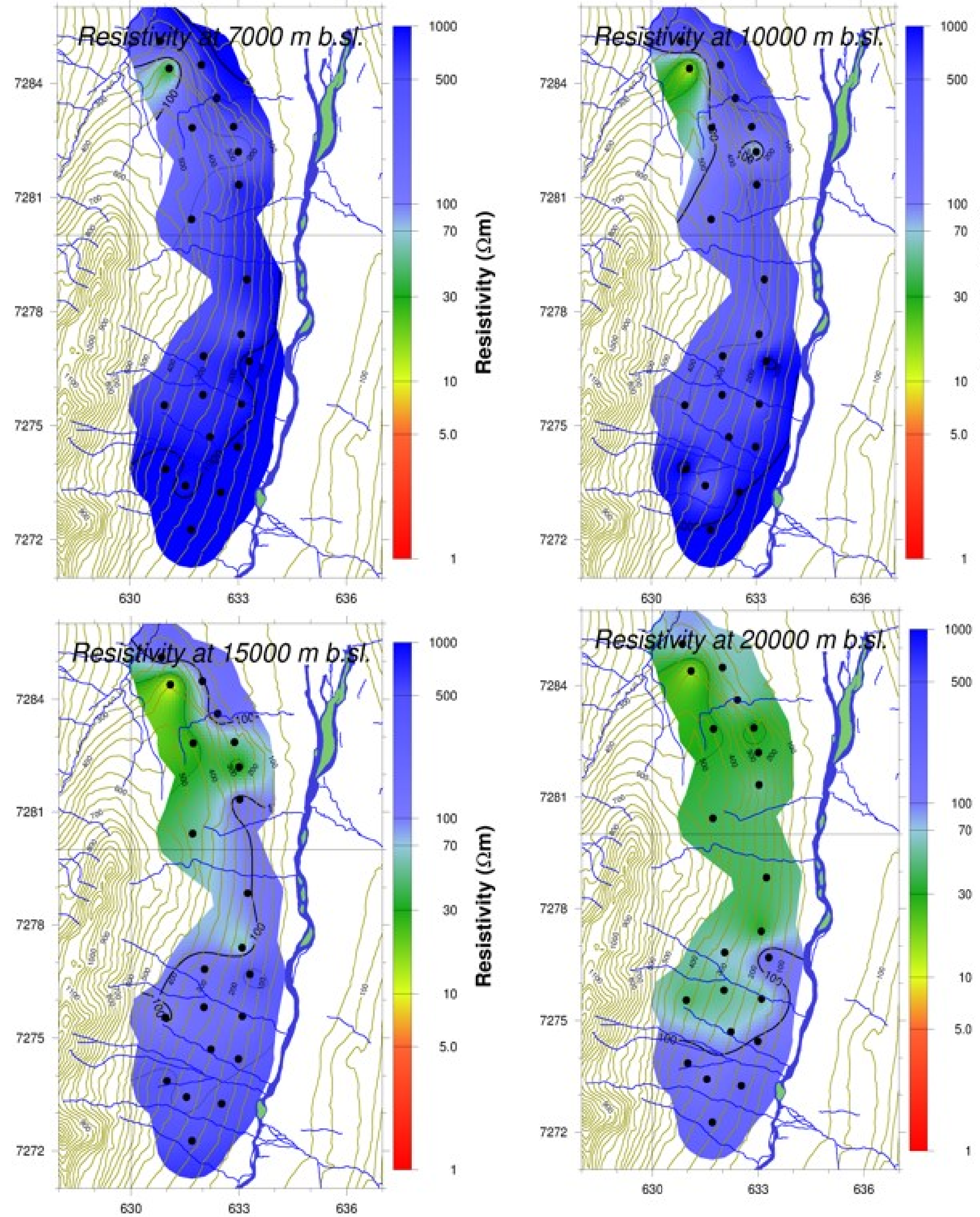

Figure 15: Resistivity depth slices at depth of 7,000, 10,000, 15,000 and 20,000 m b.s.l.; black dots show locations of MT soundings.

9. CONCLUSION

Geophysical methods are mostly used in combination with other investigations. Here, a case study was done in Eyjafjördur low-temperature geothermal field where TEM and MT methods were applied. The TEM and MT data have been processed and 1D inverted. The 1D joint inversion of TEM and MT data was performed using the TEMTD inversion program. The resistivity depth slices and cross-sections have been mapped from the results of the joint inversion.

A resistivity cross-section along the same N-S profile has been done previously by Flóvenz and Karlsdóttir (2000) and compared with isotherms from deep boreholes in the south part of the cross-section (Figure 16). From a depth of around 300-800 m, a low-resistivity layer extends up to the surface along the dipping lava pile layers. To the left, a shallow lying low-resistivity anomaly is coinciding with a phorphyritic and tholeiitic lava interbedded with sedimentary layers, a higher-resistivity layer coincides with a series of compound tholeiitic lavas. The low-resistivity anomaly observed in this work coincides with the temperature anomaly, which is interpreted as the main up flow zone of the geothermal activity according to Flóvenz and Karlsdóttir (2000). From the resistivity cross-sections and depth slices, a low-resistivity layer is seen extend upwards in a northerly direction. It reaches the surface at 11.5 km in cross-section 02 at 200 m a.s.l. (Figure 16).

A deep laying low resistivity is seen in cross-section 02 at 6,000 m b.s.l. and at 12,000 m in cross-section 01. It is seen in the subsurface in Eyjafjördur low-temperature area as in most of Iceland. It is presumably somehow connected to the heat source. According to Flóvenz and Karlsdóttir (2000), the heat source is around Botn area, close to MT soundings 007 and 008 (see Figure 10). When the results of this study are compared with those from Flóvenz and Karlsdóttir (2000), they are quite similar.

The low-resistivity anomaly is composed of porphyritic lavas, which temperature measurement analyses have shown to be the main up flow zone of the geothermal activity. The higher-resistivity host rock is presumably composed of a series of compound basaltic lavas.

A deep laying low-resistivity is seen in cross-section 02 at 6,000 m b.s.l. and at 12,000 m in cross-section 01. It is seen in the subsurface in Eyjafjordur low temperature area as in most of Iceland. It is presumably somehow connected to the heat source. According to Flóvenz and Karlsdóttir (2000), the heat source is around Botn area close to MT soundings 007 and 008 (Figure 8). When the results of this study are compared with those from Flóvenz and Karlsdóttir (2000), they are quite similar. The low-resistivity anomaly is composed of phorphyritic lavas, which temperature measurement analyses have shown to be the main up flow zone of the geothermal activity as seen in Figure 16. The higher-resistivity host rock is presumably composed of a series of compound basaltic lavas.

Using resistivity surveying methods, the resistivity structure of subsurface of the Earth is mapped as a function of depth. In geothermal areas, resistivity is prone to the parameters that can reveal the character of the geothermal reservoir. Resistivity surveying on the surface can directly probe deep structures in the subsurface of the Earth. It has many attributes for predicting the conditions of the geothermal reservoir. The main significance of the resistivity surveying is diagnostic of geothermal activity and it adds to the understanding of geothermal reality in the subsurface.

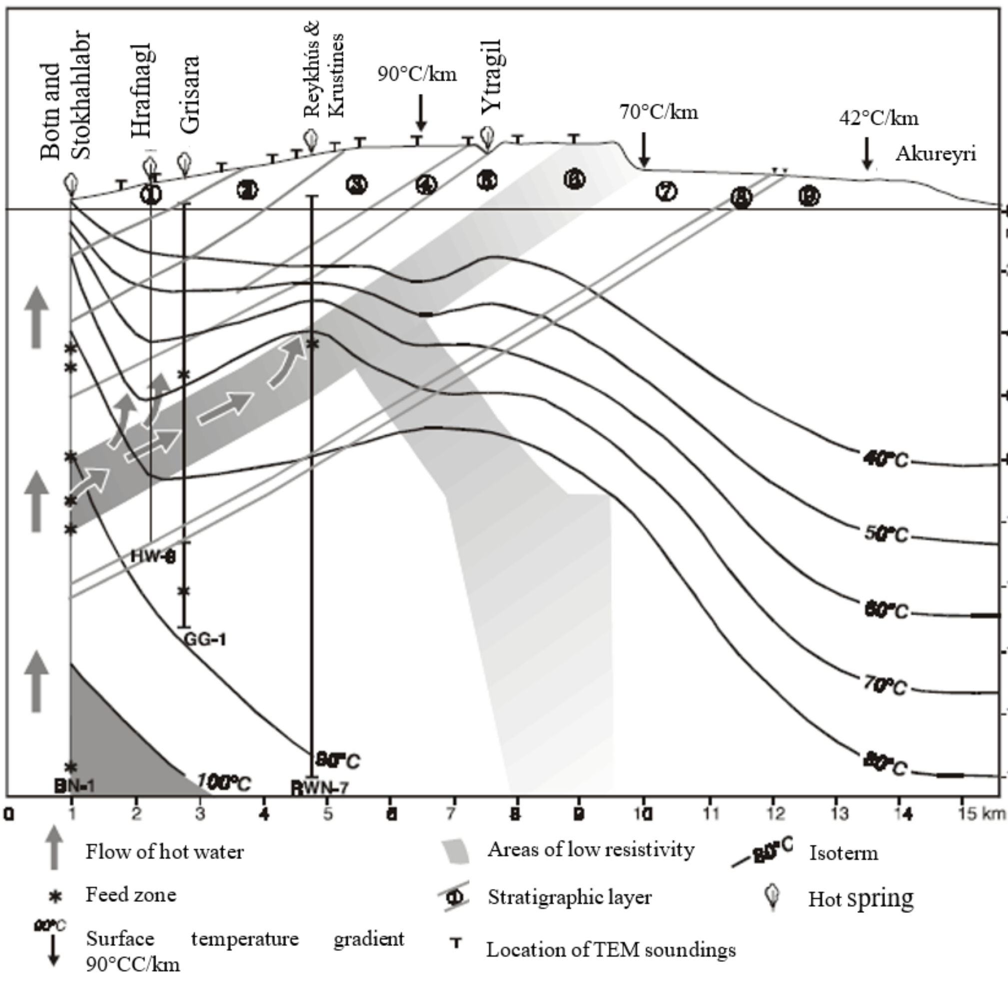

Figure 16: Geothermal interpretation of the Eyjafjordur low-temperature geothermal area (Flóvenz and Karlsdóttir, 2000)

REFERENCES

Africa, J.R.: 1D inversion of MT and TEM data with application of soundings from Krýsuvík, SW-Iceland and a review of MT/TEM data from Bacman geothermal project, Central Philippines. Report 5 in: *Geothermal training in Iceland 2013*. UNU-GTP, Iceland, (2013), 1-34.

Árnason, K.: *Central-loop transient electromagnetic sounding over a horizontally layered earth*. Orkustofnun, Reykjavík, report OS-89032/JHD-06, (1989), 129 pp.

Árnason, K.: *TemX short manual*. ÍSOR – Iceland GeoSurvey, Reykjavík, report, (2006a), 17 pp.

Árnason, K.: *TEMTD (Program for 1D inversion of central-loop TEM and MT data)*. ISOR – Iceland GeoSurvey, Reykjavík, short manual, (2006b), 16 pp.

Badilla E., D.: Resistivity imaging of the Santa Maria sector and the Northern zone of Las Pailas geothermal area, Costa Rica, using joint 1D inversion of TDEM and MT data. Report 8 in: *Geothermal training in Iceland 2011*. UNU-GTP, Iceland, (2011), 85-118.

Cagniard, L.: Basic theory of the magneto-telluric method of geophysical prospecting. *Geophysics, 18,* (1953), 605-635.

Flóvenz, Ó.G., Hersir, G.P., Saemundsson, K., Ármannsson, H., and Fridriksson Th.: Geothermal energy exploration techniques. In: Sayigh, A. (ed.), *Comprehensive renewable energy, vol. 7.* Elsevier, Oxford, (2012), 51-95.

Flóvenz Ó.G., and Karlsdóttir, R.: TEM-resistivity image of a geothermal field in N-Iceland and the relation of the resistivity with lithology and temperature. *Proceedings of the World Geothermal Conference 2000, Kyushu – Tohoku, Japan*, (2000), 1127-1132.

Geonics, Ltd.: *Operating manual for PROTEM 67 D*. Geonics Ltd, Ontario, (1999), 58 pp.

Gichira, J.M.: Joint 1D inversion of MT and TEM data from Menengai geothermal field, Kenya. Report 11 in: *Geothermal training in Iceland 2012*. UNU-GTP, Iceland, (2012), 137-167.

Hersir, G.P., and Björnsson, A.: *Geophysical exploration for geothermal resources. Principles and applications*. UNU-GTP, Iceland, report 15, (1991), 94 pp.

Hjartarson, Á., and Jónsdóttir, H.E.: *Akureyri, geological map* (2. ed.). ÍSOR – Iceland GeoSurvey, Reykjavík (2004).

Kearey, P., Brooks, M., and Hill, I.: *An introduction to geophysical exploration*. Blackwell Scientific Publications, Oxford, (2002), 254 pp.

Phoenix Geophysics: *Data processing. User's guide*. Phoenix Geophysics, Ltd., Toronto 2009.

Rowland, B.F.: *Time-domain electromagnetic exploration.* Northwest Geophysical Associates, Inc., report, (2002), 6 pp.

SEG: *MT/EMAP data interchange standard*. Society of Exploration Geophysicists, (1991), 112 pp.

Simpson, F., and Bahr, K.: *Practical magnetotellurics*. Cambridge University Press, Cambridge, UK, (2005), 254 pp.

Vogt, P.R.: Asthenosphere motion recorded by the ocean floor south of Iceland. *Earth Planet. Sci. Lett., 13,* (1971), 153-160.