GEOTHERMAL TRAINING PROGRAMME
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IS-108 Reykjavík, Iceland

Reports 2003
Number 7

TRANSIENT ELECTROMAGNETIC RESISTIVITY SURVEY AT THE GEYSIR GEOTHERMAL FIELD, S-ICELAND

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ABSTRACT

A general description of the resistivity method is given, and the results of a TEM resistivity survey at the Geysir geothermal field, S-Iceland are presented. The effects of temperature, porosity, salinity, and water-rock interaction on resistivity of rocks are briefly described. Theoretical aspects of DC resistivity measurements are presented and the general principles and application of Transient Electromagnetic (TEM) soundings are discussed. Comparison of DC and TEM resistivity soundings is dealt with briefly.

The famous Geysir high-temperature area is in a shallow valley elongated N-S in the upper southern lowlands of Iceland. A total of 9 TEM stations were interpreted with the one-dimensional inversion program, CLTINV. Results reveal the resistivity distribution of the rock formation in the uppermost kilometre. The resistivity in the uppermost 100-400 m is quite high, ranging between 200 and 3000 ohm-m. Below that, the resistivity is low, and at some level it is obviously influenced by the vicinity of high-temperature geothermal activity, indicated by resistivity values of 1-10 ohm-m. The resistivity structure within the Geysir field shows the characteristics of a geothermal system with very low resistivity at depth, 1-5 ohm-m, around the Geysir area, but without confirmed evidence of high resistivity in the central part of the geothermal system. Provided there is an equilibrium between thermal alteration of the rock and present temperature, the resistivity survey indicates temperatures as high as 200-220˚C in the uppermost kilometre of the geothermal system, but not exceeding that, as there is no certain sign of a high-resistivity core. Chemical evidence indicates temperatures of 220-240˚C in the geothermal system, agreeing fairly well with the resistivity.

1. INTRODUCTION

Geophysics is a science that deals with the application of physics on the surface of the earth and beneath it. The physical parameters studied by geophysics are temperature, electrical resistivity, density, magnetization and magnetic susceptibility, radioactivity, seismicity, etc. The geophysical methods are broadly divided into two groups: active methods and passive methods. A passive geophysical method uses the natural fields of the earth. The natural field methods consist of gravity, magnetic, electrical, and electromagnetic methods (Keary and Brooks, 1992). The active geophysical methods are those that involve generation of artificial fields that may be used analogously to natural fields. The natural field methods provide information at greater depths of the earth than artificial source methods.

In geothermal exploration, the task of geophysics is related to detection and delineation of geothermal resources, locating exploitable reservoirs, and siting of drill holes through which hot fluids at depth can be extracted (Hersir and Björnsson, 1991). The most common geophysical methods in geothermal exploration are: Electrical resistivity (DC or AC) methods, magnetic measurements, gravity measurements, and temperature surveys (mapping, gradient and heat flow). In geothermal prospecting, the information sought includes low-resistivity/high-resistivity bodies influenced by geothermal activity, dykes, faults, irregularly shaped intrusions and volcanic plugs.

Of DC resistivity measurements, the Schlumberger array is the most preferred configuration for good vertical resolution at depths of up to about 1 km. The TEM method serves a similar purpose, but gives a better resolution at depth. For greater depth resolution, the magnetotelloric method is best. It is very common to apply several geophysical methods rather than a single one in geothermal exploration. The choice of the geophysical method during the exploration of a geothermal field depends on the objective and the cost of the survey. It should be noted that there is no single method or combination of different methods that can be called optimal in all cases. The most suitable method may vary between countries and for different geothermal fields. In Iceland, thermal methods, electrical resistivity and the TEM method, gravity and magnetic measurements stand out amongst others in the study of geothermal reservoirs (Hersir and Björnsson, 1991).

The main reason for using resistivity methods is that resistivity is more dependent on important geothermal parameters like temperature, porosity, and fluid salinity than any other physical parameters that can be measured from the surface. They are also cost-efficient in prospecting. The central loop TEM sounding method has several advantages over conventional DC sounding methods. The most important one is that the transmitter couples inductively to the earth, and no current has to be injected into the ground (Árnason, 1989).

A central loop TEM survey was conducted at the Geysir geothermal field and the surrounding area in the summer of 2003. Participation in the survey and interpretation of the soundings presented in this report served as the main project of the author during the six months training at the Geothermal Training Programme of the United Nations University (UNU-GTP). The survey was organized by ISOR (Iceland GeoSurvey, formerly Orkustofnun Geoscience Division), and the main purpose was to provide a basic knowledge of the Geysir geothermal field. The duration of the field trip was August 6-14, 2003. A total of 23 stations were mesured during the field session. A 1-D inversion program for TEM soundings was used for interpretation of the data, but here results of only 9 of them are presented. Also included in this report is a comparison of the results of the TEM resistivity survey and an old DC survey.

2. RESISTIVITY OF ROCKS AND MINERALS

2.1 Basic principles of resistivity

The electrical resistivity of material is defined as the electrical resistance in ohms between the opposite faces of a unit cube of material. For a conducting cylinder of resistance δR, length δL, and a cross-sectional area δA (Figure 1), the resistivity ρ is given by the equation:

$$\rho = \frac{\delta R\, \delta A}{\delta L} \tag{1}$$

The unit of resistivity is ohm-m. The reciprocal of resistivity is conductivity.

Equation 1 is used to determine the resistivity of a homogeneous material with a regular geometric shape like a cylinder or cube. The resistivity of a material is defined mathematically according to Ohms law that states that the electric field strength at a point in a material is proportional to the current density passing through that point (Keller and Frischknecht, 1966).

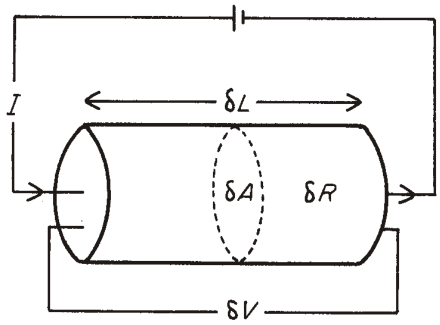

FIGURE 1: Parameters used to define resistivity (Keary and Brooks, 1992)

$$\vec{E} = \rho \vec{j} \qquad \text{or} \qquad \rho = |\vec{j}| / |\vec{E}|$$

where $\vec{E}$ = Electric field (V/m);
$\vec{j}$ = Current density (A/m^2);
ρ = Resistivity (ohm-m).

In all resistivity measurements, the surveying is conducted by measuring a signal from a naturally occurring or artificially induced current in the ground. In conventional DC soundings such as Schlumberger soundings, this is done by injecting current through electrodes at the surface and the measured signal is the electric field (potential difference over a short distance) generated at the surface. In magnetotelluric (MT) soundings, the current in the ground is induced by time variations in the earth's magnetic field, and the measured signal is the electric field at the surface as in the DC soundings (Árnason, 1989).

2.2 Factors affecting electrical resistivity of rocks

The resistivity of rock is dependent on various factors. The most important are porosity, temperature, water saturation, salinity of the saturating water and the rock's alteration mineralogy. In a geothermal environment, electrical conductivity is mainly due to movement of ions. This conduction of ions is related to the number and mobility of ions, and the open flow paths through the rock matrix.

2.2.1 Temperature

The relationship between resistivity and temperature of water is shown in Figure 2. At lower temperatures, a rise in the temperature of an electrolytic solution decreases the viscosity which leads to an increase in the mobility of ions and hence lower resistivity. At high temperatures, a decrease in the dielectric permittivity of water results in a decrease in the number of dissociated ions in a solution. Above 300°C, this starts to increase the fluid resistivity (Quist and Marshall, 1968).

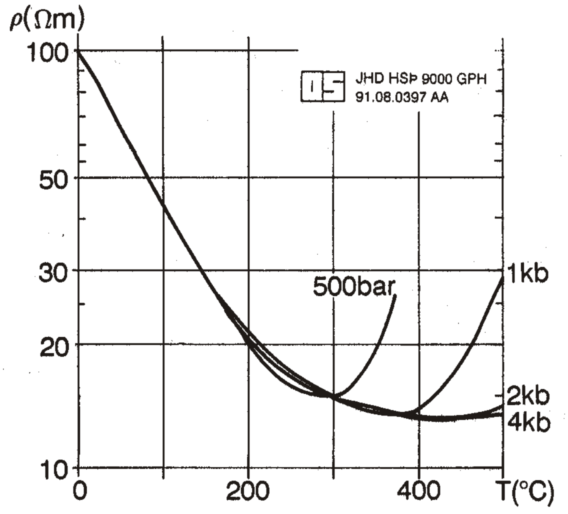

FIGURE 2: Electrical resistivity of water as a function of temperature (modified from Quist and Marshall, 1968) for different pressures

The relationship between the resistivity, ρ, and temperature, T, of the rock saturated with an electrolyte has been approximated by Dakhnov (1962):

$$\rho = \frac{\rho_{18^\circ}}{1 + \alpha_t(t - 18^\circ C)} \tag{3}$$

where ρ_{18° = Resistivity measured at the reference temperature 18°C (ohm-m);
T = The ambient temperature (°C);
α_T = Temperature coefficient of resistivity (°C^{-1}),
usually about 0.025°C^{-1} for most electrolytes in geothermal systems.

The effect of temperature variations is greatest at low temperatures (less than 100˚C), but decreases at higher temperatures when other factors, such as porosity, salinity and alteration mineralogy, become dominant factors (Flóvenz et al., 1985).

2.2.2 Porosity and texture of rocks

Increasing porosity of rocks decreases the resistivity. Some crystalline rocks with negligible porosity are conductive along their cracks. A widely used relation between the bulk resistivity ρ of water-saturated rocks, the resistivity of the water and the porosity of the rock is (Archie, 1942):

$$\rho = a\phi^{-b}\rho_w \tag{4}$$

where ϕ = Porosity;
ρ_w = Resistivity of pore fluid (ohm-m)
a and b = Empirical constants.

Equation 4 is called Archie's law. It is valid with a ≈ 1 and b ≈ 2, if the resistivity of the pore fluid is of the order of 2 ohm-m or less, but not if the resistivity is much higher (Flóvenz et al., 1985).

2.2.3 Salinity

The salinity of a NaCl solution with the same resistivity as a particular solution is called equivalent salinity. The effect of salinity of sodium chloride solution on resistivity at various temperatures is presented in Figure 3. Figure 3 shows that the resistivity is approximately inversely proportional to salinity except at extremely high concentration. The conductivity of a NaCl solution increases seven fold when the temperature increases from 0 to 140˚C (temperatures in the range of 100-140˚C need to be accompanied by sufficiently high pressure to keep water in the liquid state).

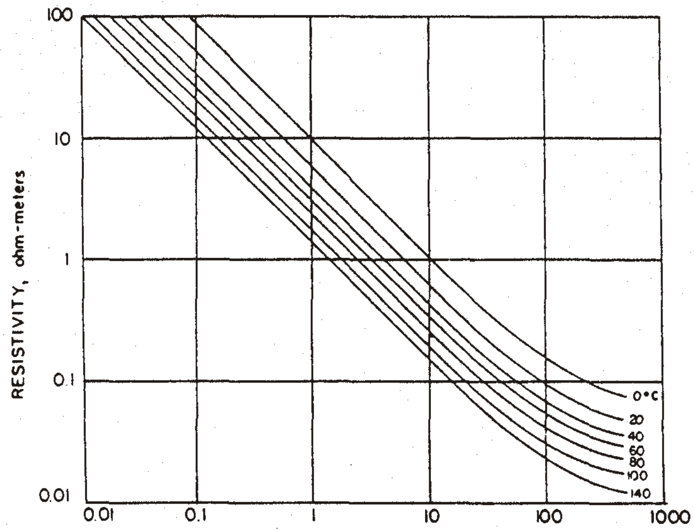

FIGURE 3: Resistivity of solutions of NaCl as a function of concentration and temperature (Keller and Frischknecht, 1966)

2.2.4 Water-rock interaction and interface conditions

Interface conductivity is caused by mineral alteration of the rock matrix. The type of alteration minerals formed depends upon the temperature and the chemical composition of the fresh rock and the saturating fluid. These alteration minerals lining the walls of the fractures seem, to a large extent, to control the electrical resistivity in the rocks up to temperatures of 200-250˚C. The following discussion is mainly based on results from Icelandic high-temperature geothermal areas, and an important part of that data comes from the Nesjavellir high-temperature geothermal field, SW-Iceland (Árnason et al., 1986,1987; Hersir and Björnsson, 1991).

Icelandic high-temperature geothermal fields have similar characteristics, with a low-resistivity cap at the outer margin of the geothermal reservoir that is underlain by more resistive rocks in the core of the system. The low-resistivity horizon is characterised by conductive minerals of smectite-zeolite type formed at temperatures of 100-220˚C. These minerals have loosely bound ions that form a conductive layer on the walls of the interconnected fractures that controls the resistivity of the rock. Smectite is a layered sheet silicate with abundant loosely bound cations, and hence is highly conductive. At the temperature range 230-240˚C, a mixed-layer clay zone is formed where zeolites disappear and smectite converts into chlorite. The clay mineral chlorite has similar structure as smectite, but in chlorite the cations are fixed in a lattice, and hence chlorite is resistive to electric current. When the temperatures exceed 250˚C, chlorite and epidote become the dominant minerals. They are both resistive and the resistivity is therefore probably dominated by conduction through the pore fluid in this resistive core of the geothermal system. The important consequence of this is that the observed resistivity structure can be interpreted in terms of temperature distribution. A similar resistivity structure is to be expected in acidic rocks. Due to different alteration mineralogy, however, the transition from the conductive cap to the more resistive core presumably occurs at temperatures lower than 200˚C (Árnason et al., 2000).

3. DC-RESISTIVITY METHODS

The DC-resistivity sounding measurements are the most widely used geophysical method in the study of subsurface electrical properties of the earth. In DC measurements, the current I is passed into the earth through electrodes and the resulting electrical potential V is recorded. Variations in resistivity of the rocks in the subsurface control the current flow in the earth and this, in turn, affects the potential distribution on the surface which is measured. By producing resistivity values from surface measurements, valuable information on resistivity structures at depth, and hence geological structures, are revealed.

3.1 Current flow in a homogeneous earth

Consider an element of homogeneous cylinder (Figure 4). The flow of current passing through the cylinder generates a potential $-\delta V$ between the ends of the cylinder. According to Ohm's law, the electrical potential $-\delta V$, the resistance δR, and the current I are mathematically related by the equation:

$$\delta V = -\delta R\, I \qquad (5)$$

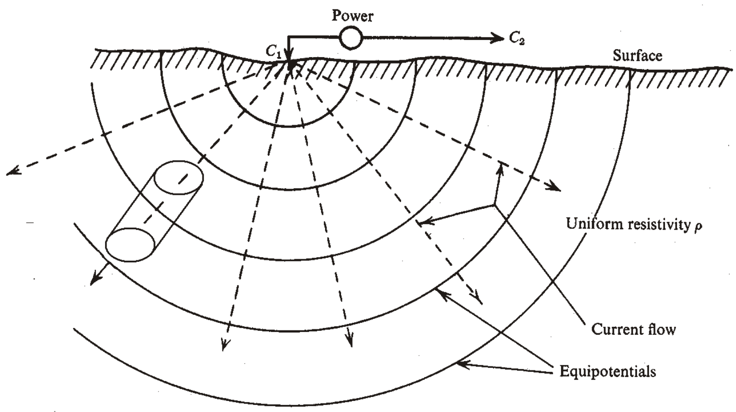

FIGURE 4: Point source of current at the surface of homogeneous medium (taken from Keary and Brooks, 1992)

Equation 1 can be expressed as:

$$\delta R = \frac{\rho\, \delta L}{\delta A}$$

By substituting this into Equation 5, the potential gradient, $\delta V/\delta L$ (V/m) can be solved for:

$$\frac{\delta V}{\delta L}=\frac{-\rho I}{\delta A}=-\rho\, j \tag{6}$$

where j = Current density (A/m^2)
I = Current (A)

For a homogeneous earth, the potential resulting from a single current source on the surface of the earth is shown in Figure 4. The current density j at any distance r from its source is given by:

$$j=\frac{I}{2\pi\, r^2} \tag{7}$$

By substituting this into Equation 6, the first derivative of the potential can be given as:

$$\frac{\delta V}{\delta L}=\frac{-\rho I}{2\pi\, r^2}=\frac{dV}{dr} \tag{8}$$

By integrating over r, Equation 9 is obtained:

$$V_r=-\int\frac{dV}{dr}ds=\int\frac{\rho I}{2\pi\, r^2}ds=\frac{\rho I}{2\pi\, r}+C \tag{9}$$

and by demanding that $V_r \rightarrow 0$ when $r \rightarrow \infty$, implies that $C = 0$.

If a current I is injected at one electrode, and the same amount of current extracted at a second electrode, the potential $V_1 + V_2$ at a point on the surface at the distance r_1 and r_2 from the first and second source, respectively, is given as:

$$V_1+V_2=\frac{\rho \mathrm{I}}{2\pi}\left(\frac{1}{r_1}-\frac{1}{r_2}\right) \tag{10}$$

If a second potential electrode is introduced at P_2, it is possible to measure the difference in potential ΔV between the two points P_1 and P_2

$$\Delta V=\frac{\rho \mathrm{I}}{2\pi}\left\{\left(\frac{1}{r_1}-\frac{1}{r_2}\right)-\left(\frac{1}{r_3}-\frac{1}{r_4}\right)\right\} \tag{11}$$

Such an arrangement corresponds to the many four-electrode arrangements normally used in resistivity field work (Telford et al., 1976).

3.2 Potential of point source on the surface of the layered earth

As for a homogeneous half-space, the potential due to a point current source on the surface of a layered half-space can be calculated analytically. For a horizontally-layered half-space, the following four specifications are done (Koefoed, 1979):

1. The subsurface consists of a finite number of layers separated from each other by horizontal boundary planes; the deepest layer extending to infinite depth, the other layers having finite thickness;
2. Each of the layers is electrically homogeneous as well as electrically isotropic;
3. The field is generated by a point source of current that is located at the surface of the earth;
4. The current emitted by the source is direct current (DC).

From Equation 2, and by using the relationship between the electric field and potential in Equation 12

$$E = -\vec{\nabla} V \tag{12}$$

the gradient of the potential can be represented as:

$$-\vec{\nabla} V = \rho \vec{j} \tag{13}$$

Taking the divergence and using the fact that there is only a point source and it is assumed to be at the origin of the coordinate system, the following equation can be derived:

$$\vec{\nabla}.\left(\vec{\nabla} V\right) = -\rho I \delta\left(\vec{x}\right) \tag{14}$$

where I is the induced current and $\delta\left(\vec{x}\right)$ is Dirac's delta "function".

Equation 14 can be solved for V_i in each layer, by separation of variables (see e.g. Koefoed, 1979).

By applying the boundary conditions that the potential must satisfy at layer boundaries (continuity of V and the vertical component of the current density), at the surface (no current through the surface except at the source) and at infinite depth ($V \to 0$ as $z \to \infty$), the potential on the surface, at the distance r from the source can be expressed as

$$V(r) = \frac{\rho_1 I}{2\pi} \int K_1(\lambda) J_o(\lambda r) d\lambda \tag{15}$$

where the Kernel function $K_1(\lambda)$ is determined by the recursion relation in Equation 16.

In layer i, a function K_i is defined in terms of K_{i+1} in the layer below. This applies for the resistivity of layer i +1 and the resistivity and thickness of layer i. These functions obey the following recurrence relation:

$$K_i = \frac{K_{i+1} + \frac{\rho_i}{\rho_{i+1}} \sinh(\lambda d_i)}{\frac{\rho_i}{\rho_{i+1}} + K_{i+1} \sinh(\lambda d_i)} \tag{16}$$

In the infinite basement, $K_N = 1$. By starting in the basement and calculating K_N, K_1 in the top layer can be calculated recursively by using Equation 16.

3.3 Definition of the apparent resistivity

The main objective of a resistivity survey is the determination of resistivity distribution in the subsurface. In the theoretical case of a homogeneous earth, the expression of the potential difference on the surface of the earth that is caused by point sources of current is given by Equation 11. The resistivity of a homogeneous earth can be calculated from this equation for any type of configuration. If the above mentioned formula is applied for the non-homogeneous earth, the "average" resistivity that is computed for the respective electrode spacing is called apparent resistivity.

3.4 Different electrode configurations in resistivity measurements

There are various types of electrode configurations which are commonly used. Of the many configurations that have been used in resistivity prospecting at various times, not more than six have survived to any extent (Telford et al., 1976). These are the Wenner array, Schlumberger array, the three point (gradient) system, double-dipole system, Lee partition method, and the line-electrode configuration.

In all the configurations, four electrodes are in use where two are for current, conventionally called A and B, and two are potential electrodes, M and N, for measuring the potential difference. At present, the most common type is the Schlumberger configuration.

3.5 Schlumberger sounding and the horizontally layered earth

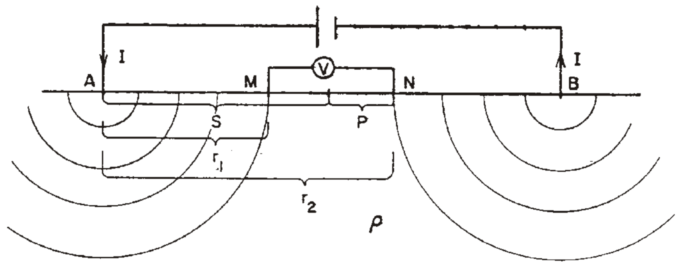

FIGURE 5: Schlumberger electrode configuration

In the Schlumberger configuration, all four electrodes are on a straight line and symmetric about the sounding centre (Figure 5). Ideally, the potential electrodes are kept fixed and close to the centre. The current electrodes are moved stepwise away from the centre, normally with the electrode spacing equally distributed on log-scale. Increased distance between the current electrodes increases the depth of penetration of the current and that of the sounding. As the spacing between the two current electrodes is increased, the signal for the potential difference becomes lower. Gradually, the potential signal gets too weak for measurements to be taken. In this case, the spacing of the potential electrodes needs to be enlarged.

For a homogeneous earth, the potential difference in a Schlumberger sounding (see Figure 5) is given as:

$$\Delta V = \frac{I\rho}{2\pi}\left\{\left(\frac{1}{AM}-\frac{1}{BM}\right)-\left(\frac{1}{AN}-\frac{1}{BN}\right)\right\} \tag{17}$$

By solving Equation 17 for ρ and using the variables S and P defined in Figure 5 as:

$$AM = BN = S - P\,; \quad AN = BM = S + P$$

the apparent resistivity can be expressed as:

$$\rho = \frac{\Delta V}{I}\pi\frac{S^2 - P^2}{2P} \tag{18}$$

4. TEM SOUNDINGS

4.1 Basic principles

A naturally occurring or artificially induced alternating current is used in electromagnetic resistivity surveys. With an artificial source, the primary electromagnetic field may be generated by passing alternating current through a large loop of wire or grounded dipole. The response of the ground is the generation of secondary electromagnetic fields and the resultant fields may be detected by induced voltage in a receiver coil by the process of electromagnetic induction (Keary and Brooks, 1992)

Electromagnetic methods are divided into frequency domain and time domain methods, depending on whether the source signal is harmonic (sinusoidal) or not. Time domain electromagnetic methods are also divided into two types, the grounded dipole and the loop source methods.

Figure 6 shows the survey layout of the central-loop TEM sounding method. Here, a loop of wire is placed on the ground and constant current is generated in the loop. A constant magnetic field of known strength is then built-up. The current is then abruptly turned off and the decaying magnetic field induces an electrical current in the ground as shown in Figure 6. 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 the receiver coil at the centre of the transmitting loop.

The central-loop TEM method was first tested for geothermal exploration in Iceland in the summer of 1986 (Árnason et al., 1987), and for mapping of saline groundwater in the summer of 1988. The depth of penetration is dependent on how long the induction in the receiver coil can be traced in time before it is drowned in noise. For a transmitter loop with an area about 100,000 m^2 and current about 20 A, the induction can be measured up to about 100 ms after the current is turned off. This gives penetration depth of 500-1000 m, depending on the resistivity structure. This is similar to that of Schlumberger soundings with a maximum distance of 3 km between the current electrodes.

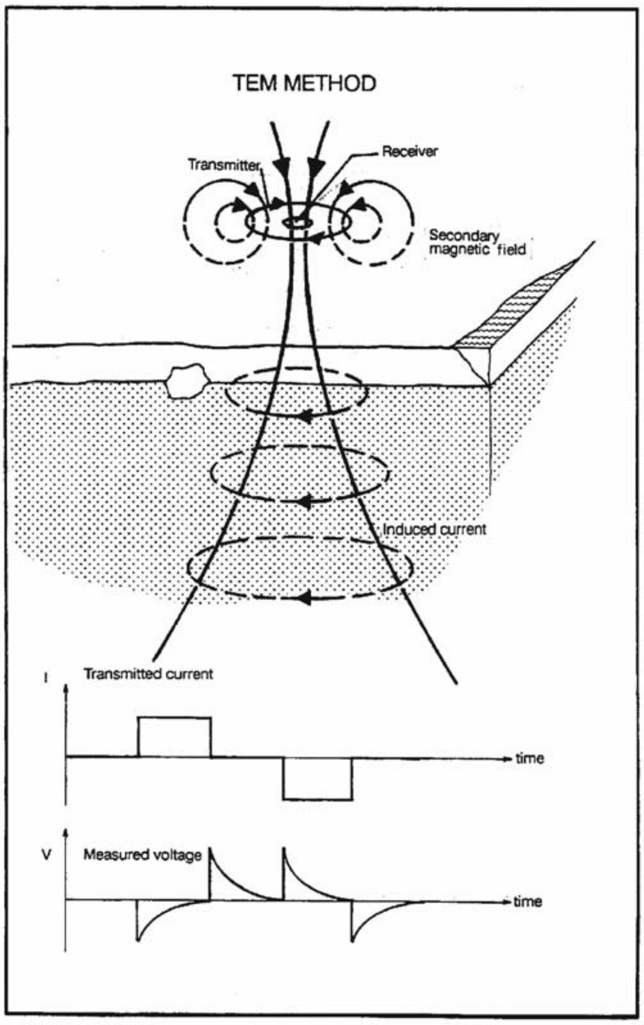

FIGURE 6: The central loop TEM sounding configuration (Hersir and Björnsson, 1991)

For the case of a horizontally layered earth, the induced voltage in a receiver coil at the centre of a circular source loop of radius *r* can be calculated analytically (Árnason, 1989). For a constant current turned off at $t = 0$, it is given as:

$$V(t,r) = I_0 C \int_0^\infty \mathrm{Re}\left[E(\omega,r)\right]\cos(\omega t)d\omega \tag{19}$$

where C $= A_r\, n_r\, A_s\, n_s \dfrac{\mu_0}{\pi^2 r^3}\, \mu_o$

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

The complex frequency domain response is given as:

$$E(\omega,r) = 2r^2 \int_0^\infty \lambda \frac{S_o(\lambda,\omega)}{S_o(\lambda,\omega) - T_o(\lambda,\omega)} J_1(\lambda r)\, d\lambda \tag{20}$$

The functions $S_o(\lambda,\omega)$ and $T_o(\lambda,\omega)$ are determined by recursion relations, similar to Equation 16 and dependent on layer resistivities and thicknesses.

For a homogeneous earth, the transient voltage in the receiving coil can be divided into three stages. These are the early stage, the intermediate stage, and the late stage. In the early stage, the induced voltage is constant with time, but it starts to decline with time in the intermediate stage. At late times, the induced voltage in the receiving coil on a homogeneous half-space of conductivity σ is given approximately by (Árnason 1989):

$$V(t,r) = \mathrm{I}_0 \frac{C(\mu_0 \sigma r^2)^{3/2}}{10\pi^{1/2} t^{5/2}} \tag{21}$$

If the logarithm of the voltage is plotted as a function of logarithm of time, Equation 21 gives a straight line with the slope -5/2.

Equation 21 can be solved for the resistivity $\rho = 1/\sigma$ giving the resistivity of the half-space in terms of induced voltage at late times after the source current is turned off:

$$\rho_a = \frac{\mu_0}{4\pi} \left[\frac{2\mathrm{I}_0 A_r n_r A_s n_s}{5t^{5/2} V(t,r)} \right]^{2/3} \tag{22}$$

4.2 Definition of late time apparent resistivity

It is customary to present the TEM sounding data by "late time apparent resistivity" as a function of time by using Equation 22. If the transient response is presented as late time apparent resistivity rather than induced voltage, the response will, in the case of homogeneous earth, approach the true resistivity of the earth for late time. In the case of layered earth, the late time apparent resistivity will approximately reach the true resistivity of the intermediate layers for some interval of time. This happens if the layers are thick enough so that the voltage response approximately reaches, for some time interval, the linearly decreasing late stage behaviour corresponding to homogeneous earth with the resistivity of the respective layers.

4.3 1D inversion of TEM soundings

The program CLTINV for interpretation of TEM sounding data is used in this report. It is a program used to perform 1D inversion in terms of horizontally layered models. The inversion algorithm used is the non-linear least-squares inversion of the Levenberg-Marquardet type (Árnason, 1989). 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 with exponential current turn-on and linear current turn-off. The program offers the possibility to keep models smooth, by performing a so-called minimum structure (Occam's) inversion. In this case, the layer thickness is fixed, equally spaced on log scale, and the conductivity distribution is forced to be smooth.

The program is run by typing CLTINV at the system prompt, followed by information on the measured data file, model and various optional parameters. The program starts iteration of the input data, and at the end of the iteration process it writes the measured data, the calculated response of the final model, the final model, and the Chisq misfit.

5. CENTRAL LOOP TEM RESISTIVITY SURVEY AT THE GEYSIR FIELD

5.1 Location and accessibility

The Geysir geothermal field is located in S-Iceland, about 110 km east of Reykjavik, the capital of Iceland, at the approximate elevation of 100 m. It is characterised by a variety of thermal manifestations within an area elongated in a NNE-SSW direction. An areal coverage of these thermal manifestations is about 3 km^2. In the vicinity are three mountains, Laugarfjall, Bjarnarfell and Sandfell with heights of 187, 727 and 610 m, respectively. An asphaltic main road runs from Reykjavík to the Geysir area being one of the main tourist attractions in Iceland. Geysir is a special kind of hot spring, which from time to time erupts a column of water above the ground. The temperature of the erupting water is close to boiling or near 100˚C. In the area around the Geysir high-temperature field (Figure 7), farming is important, mainly animal farming. Reykholt and Laugarás are the two nearest villages, both with large reservoirs of geothermal water used to heat greenhouses. In Reykholt, there is a health-care centre, a store, and accommodation facilities. The combined population is 286.

The Geysir area was for a long time the property of an individual, but he donated it to the Icelandic nation in 1953. Then, a special group, the Geysir committee, was put in charge of the area. In the following year, the area was fenced off. The fence keeps the livestock out and since then the vegetation has recovered considerably. In 1960, some shrubs were planted and now some 125 species of higher plants and 20 species of moss are found inside the fenced area. There are high restrictions on movement within the Geysir area due to environmental protection and also the possibility of stepping on altered ground and being burned (Torfason, 1985).

5.2 Geological setting

Figure 8 shows a simplified geological map of the Geysir area, and is characterized by a variety of geological formations and hot springs. The geothermal area is located at the eastern margin of the western volcanic rift zone of Iceland and is thought to obtain heat from intrusive bodies at a few kilometres depth

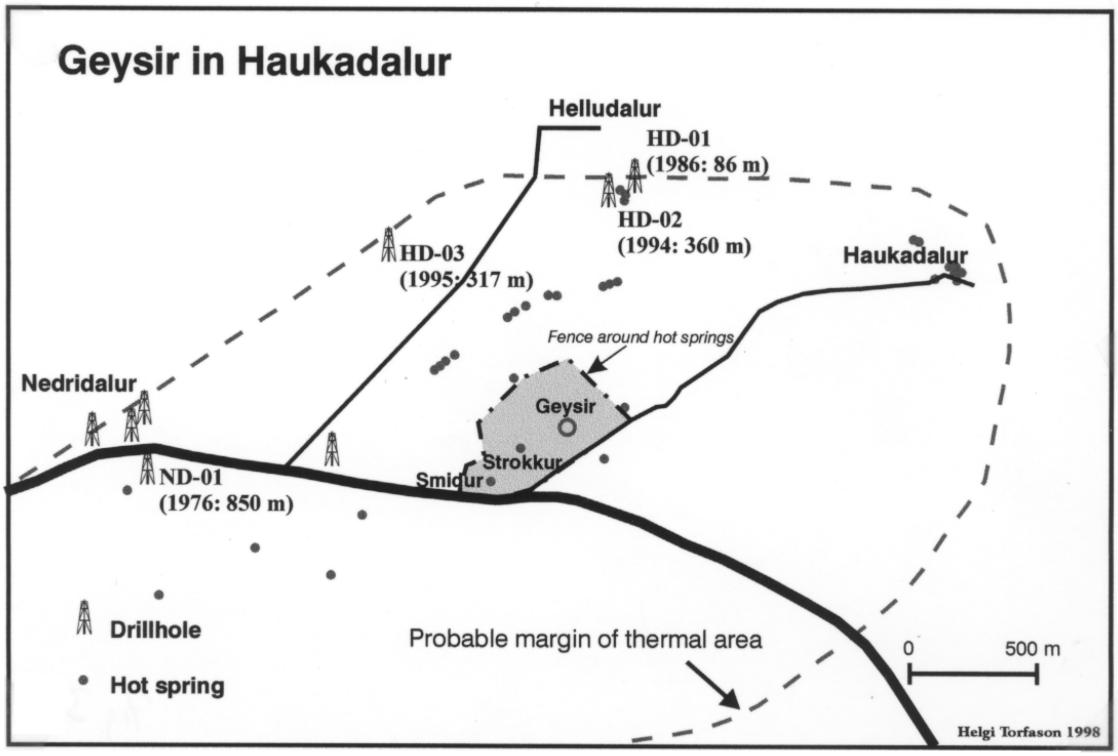

FIGURE 7: Location map of the Geysir geothermal field (Helgi Torfason, personal comm.)

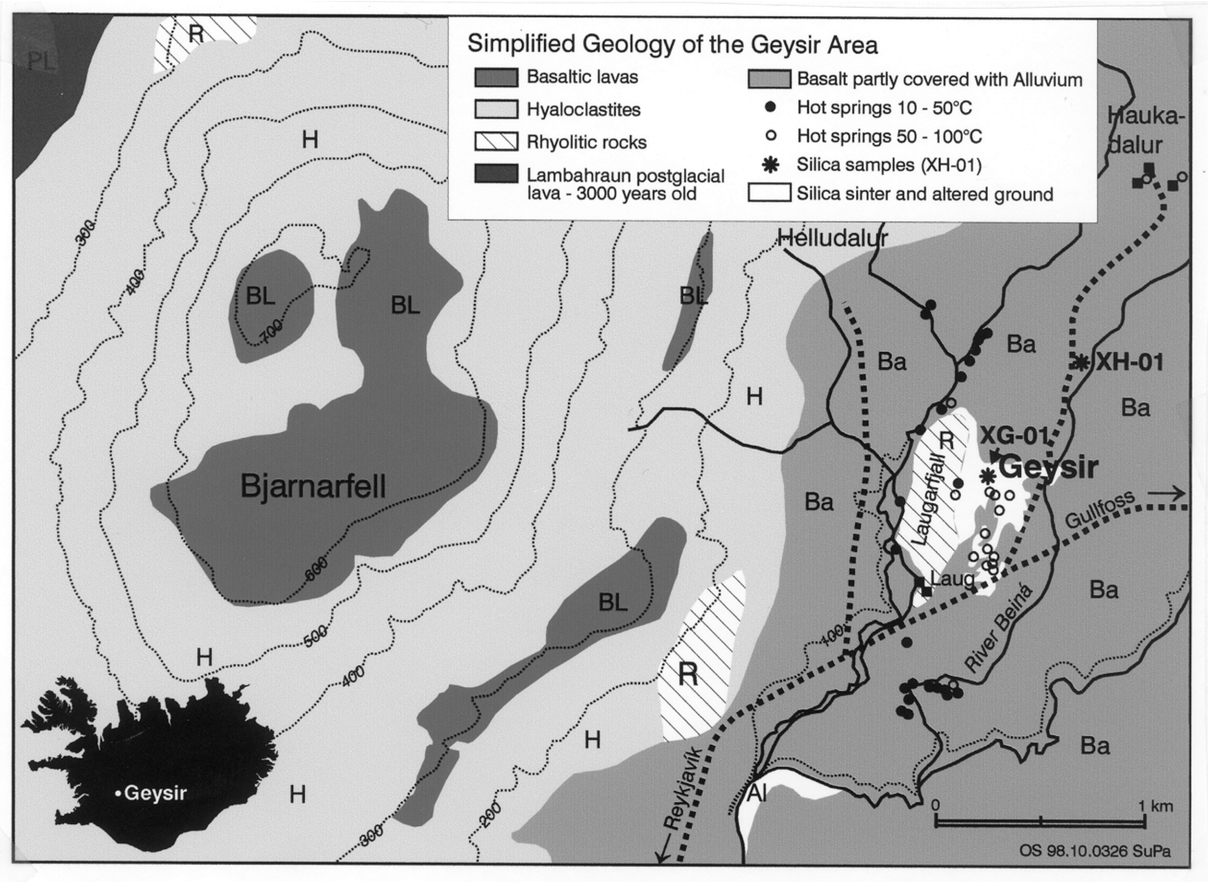

FIGURE 8: A simplified geological map of the Geysir geothermal area and its surroundings (Torfason, 1985)

associated with a recently extinct volcanic system or central volcano. Rhyolitic rocks are found at Laugarfjall, and east of there an alteration zone and quite widespread deposits of silica sinter are observed (Torfason 1985).

The Laugarfjall mountain, west of the geothermal area, is characterised by rhyolite and small patches of rhyolite are found in southeastern and northern Bjarnarfell. The Bjarnarfell mountain which rises to the west, consists of basaltic lava flows and brownish, crumbly rock type known as hyaloclastite, which is responsible for the rugged landscape of the volcanic zone. The hyaloclastites, were formed in volcanic eruptions beneath the glacier that covered the whole of Iceland periodically during the Ice-age (Pleistocene). When an eruption starts beneath a glacier, the magma is quenched very rapidly as it comes into contact with the ice and melt water. The magma solidifies so quickly that crystals are not formed and it turns into an amorphous volcanic glass formation. The glass is shattered into fragments during the eruption and becomes mixed with lumps of solid lava. Fresh volcanic glass is black, but with weathering and alteration, it turns to brown, i.e. to hyaloclastite.

The Geysir field is most known for its many geysers, especially the Great Geysir from which the English word geyser is derived,. The Icelandic word geysir means spouter or erupter. At present, Strokkur is the only geyser very active in the area. A number of thermal manifestations are seen in the area surrounding Geysir. Hot springs, fumaroles, solfataras or steam vents, and mud-pots are spread around the area. The fumaroles discharge steam and non-condensable gases, and the grey or reddish colour in the mud pots results from iron compounds. Hot springs emit only hot water without spouting, but the geyser Strokkur spouts 20-30 m columns every 3-5 minutes. In the beginning of the 18th century, Geysir erupted three times a day, but its activity declined and an interval of several days or weeks between eruptions became common. For the last few decades, Geysir erupted very irregularly or not at all, unless treated with soap. When Geysir is at its best, eruptions reach 70-80 m with thundering noise (Torfason, 1999). Tectonic fractures strike NE-SW, but there is some evidence of other fracture systems connected with the geothermal area, probably related to the extinct central volcano. Earthquakes in southern Iceland in June 2000 caused some tectonic movements in the Geysir area and reactivated the Great Geysir to some extent. Now there are again occasional eruptions of Geysir even though they do not reach the magnitude of the great eruptions of past.

5.3 Instruments

For the TEM resistivity survey, the Geonics Ltd. Protem 67, 20 gate model was used. The major components of the instruments are a generator and a current transmitter, box, receiver and transmitter loops, and a computer. During the current-off portions of the current waveform (Figure 6) the receiver measures the time derivative of the vertical component of the magnetic field using both a small coil with an effective area of 100 m^2, and a flexible loop with an effective area of 5613 m^2. The transmitter loop was a single turn square loop with 300 m side length.

The maximum transmitted current is usually in the range of 20-24 A, and the transient signal is recorded in the time interval of 0.087-70.4 ms at 20 logarithmically spaced channels after the current is turned off. Both the transmitter and the receiver timing are controlled by synchronized high-precision crystal clocks. The induced voltage is measured by the receiver each time after the transmitter is turned off. Data were recorded for two transmitter frequencies. At high frequency, the repetition rate of the transmitted current signal is 25 Hz, with 10 ms current-off segments. At low frequency, the repetition rate is 2.5 Hz, with current-off segments of 100 ms. Repeated transients are stacked and stored in the computer memory of the receiver and later downloaded to a PC computer.

5.4 Data processing and interpretation

The recorded data are edited to remove electromagnetic noise to obtain induced voltage and finally, apparent resistivity is calculated as a function of time, and plotted as such. A computer program called "a one-dimensional inversion program for central loop transient electro-magnetic sounding" or CLTINV was used to interpret the field data (see section 4.3). The inversion program simulates the actual current waveform and the geometry of the field setup.

The final results of the interpretation are presented by two output files, the output list file and the output plot file. In the output list file, the iteration number, the changes in the model parameters affecting the calculated apparent resistivity, eigenvectors and eigenvalues can be read. The program also writes the correlation matrix into the output file.

In the output plot file, the measured data points, calculated apparent resistivity values, the final model and the misfit chi-square sum are written. The results can be plotted both on screen and on a printer. The measured apparent resistivity values are plotted against the square of time, measured in micro-seconds, as small circles on a double logarithmic plot and the calculated apparent resistivity curve is drawn as an unbroken line. The resistivity model is displayed both numerically as resistivity values (ohm-m), and layer thicknesses (m); and also in a histogram where the x-axis shows the depth in m, and the y-axis the resistivity values in ohm-m.

5.5 Results

The results of the TEM survey are discussed in this section. Interpretation was done on 9 TEM soundings along a cross-sections oriented NE-SW. The locations of the soundings and the cross-section are shown in Figure 9. The observed and interpreted values of the soundings of the survey are presented. One of them is found below but others are presented in Appendix I. The NE-SW cross-section is described in detail in the following section. Finally, results from Schlumberger soundings in the area are compared to the results of the TEM soundings.

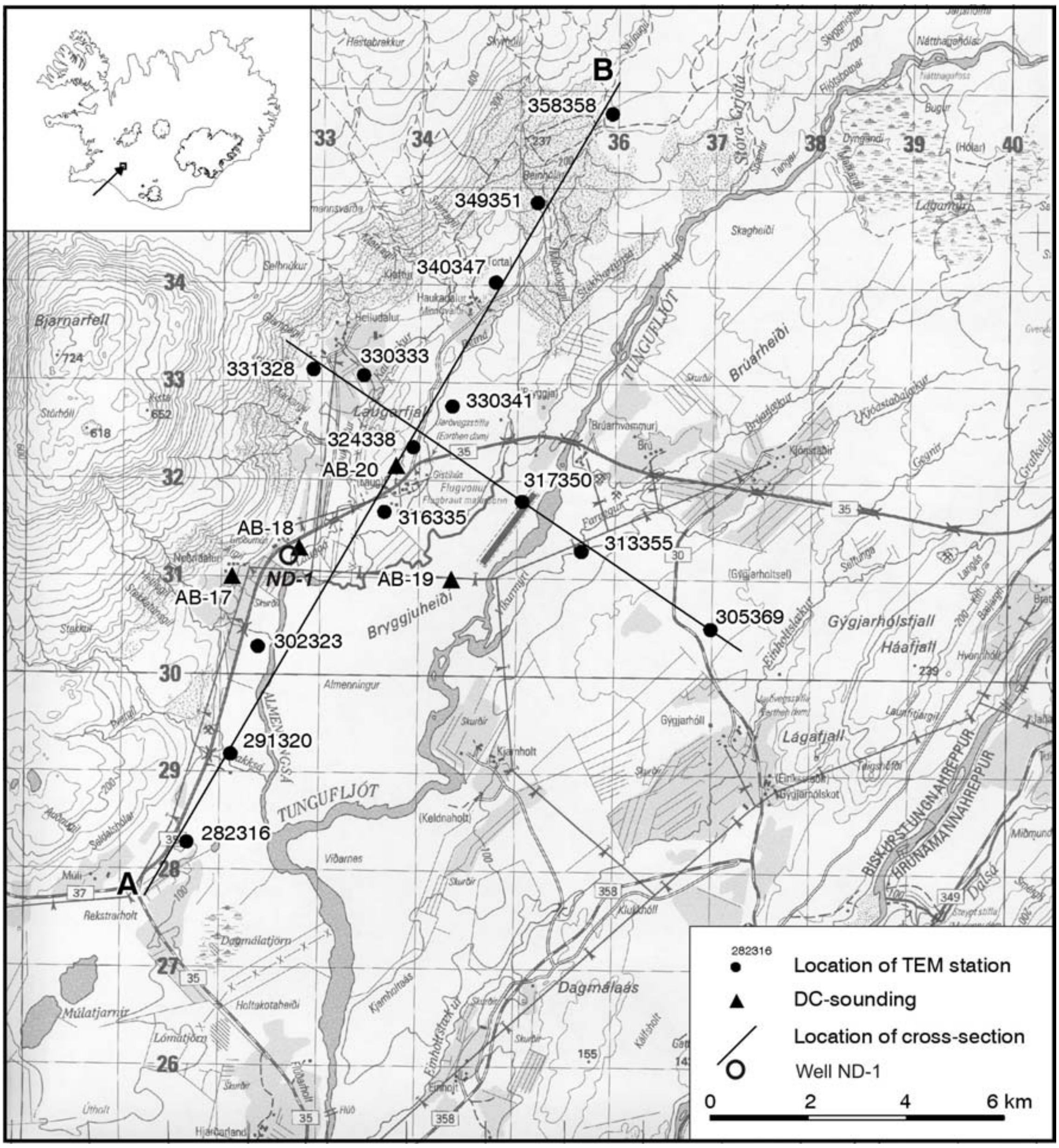

FIGURE 9: Location map of TEM and Schlumberger soundings and cross-section AB in the Geysir area

5.5.1 TEM soundings

An example of a TEM sounding and its interpretation is given in Figure 10 that shows TEM station 282316. The interpretation of the sounding assumes five layers. Two layers of relatively high resistivity reach down to almost 200 m depth, reflecting unaltered rocks close to the surface. Below 200 m depth there is a layer of moderate resistivity (30 ohm-m), the decrease in resistivity most likely due to increased low-temperature alteration, and temperature. At approximately 325 m depth, a thick (574 m) low-resistivity layer of 6.5 ohm-m is seen. This is the low resistivity defining the high-temperature field. Below a little rise in resistivity with depth is indicated, though still within 10 ohm-m.

Results of the TEM sounding survey at the Geysir geothermal field reveal the sequence of the resistivity distribution of the rock formation at depth. The resistivity in the uppermost 100-400 m is generally quite high, ranging between 200 and 3000 ohm-m except within the thermally altered surface area where it is lower. As has been discussed previously, the high resistivity is correlated with fresh basalts and un-altered rhyolitic rocks near Laugarfjall. Thermal alteration is minimal in these rocks. Below that, the resistivity is generally low to very low, and at some level it is obviously influenced by the high-temperature geothermal activity, with resistivity on the order of 1-10 ohm-m.

5.5.2 The resistivity structure of the Geysir area

The characteristic resistivity structure of a high-temperature field in Iceland is a low-resistivity cap underlain by a high-resistivity core. The low resistivity is defined by resistivity values in the range of 1-10 ohm-m in fresh water systems. The high-resistivity core has resistivity at least an order of a magnitude higher than that of the low-resistivity cap. The outer margin of the low-resistivity cap delineates the high-temperature field within a crust not influenced by high-temperature alteration.

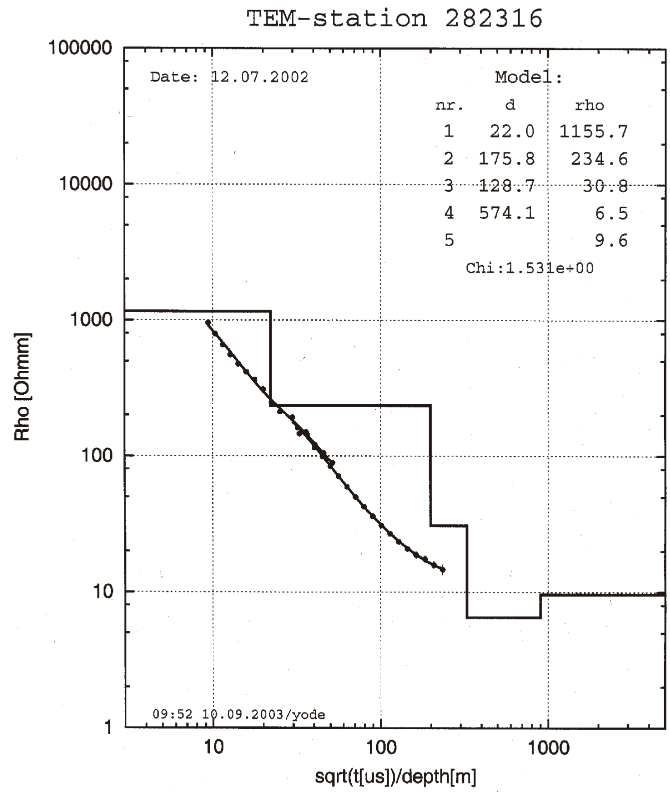

FIGURE 10: TEM sounding curve for station 282316 with 1D inverse interpretation

Figure 11 shows cross-section AB through the Geysir high-temperature field from the TEM survey. The location is along line A-B oriented NE-SW in Figure 9. It shows the characteristics of a geothermal field with very low resistivity, 1-10 ohm-m, in the vicinity of the geothermal area, but without a certain sign of a high-resistivity core down to at least 800 m depth. Low resistivity of 5-10 ohm-m, is found at the outskirts and at upper levels, but underlain by resistivity of < 5 ohm-m. Three soundings, 324338, 330341, and 340347 show a similar character with a body of high resistivity found between low-resistivity layers. This feature could have various explanations. One is a dense intrusion with less porosity than the surrounding rocks, but that cannot be verified as there are no drillholes in the area. Being only seen in soundings on this cross-section, there is no knowledge of the extent of this high-resistivity body. This could even be explained by a 3-D effect, but verifying that would require more soundings and computer software not within the task of this report. Outside the reservoir delineated by the low resistivity, the cross-section shows a resistivity structure of unaltered rock at upper levels, and slightly altered rocks at deeper levels.

The resistivity structure of most high-temperature fields reflects the thermal alteration. The low-resistivity "cap" with resistivity < 10 ohm-m in fresh water-systems corresponds to the smectite-zeolite zone. The resistivity increases in the mixed-layered clay zone due to increasing chlorite and decreasing smectite. The high-resistivity core corresponds to the chlorite-epidote zone. Provided there is equilibrium between alteration and temperature, the resistivity structure relates directly to temperature.

In the Geysir area, no high-resistivity core is seen, or it is found at deeper levels than seen by the TEM soundings presented here. Chemical evidence states that the Geysir area is definitely a high-temperature system with a temperature of 220-240˚C (Pasvanoglu, 1998). The resistivity structure seen in cross-section A-B (figure 1) and in other soundings of the TEM-survey at the Geysir field presented here seem to indicate that temperatures in excess of 220-230˚C are not reached in the uppermost kilometre.

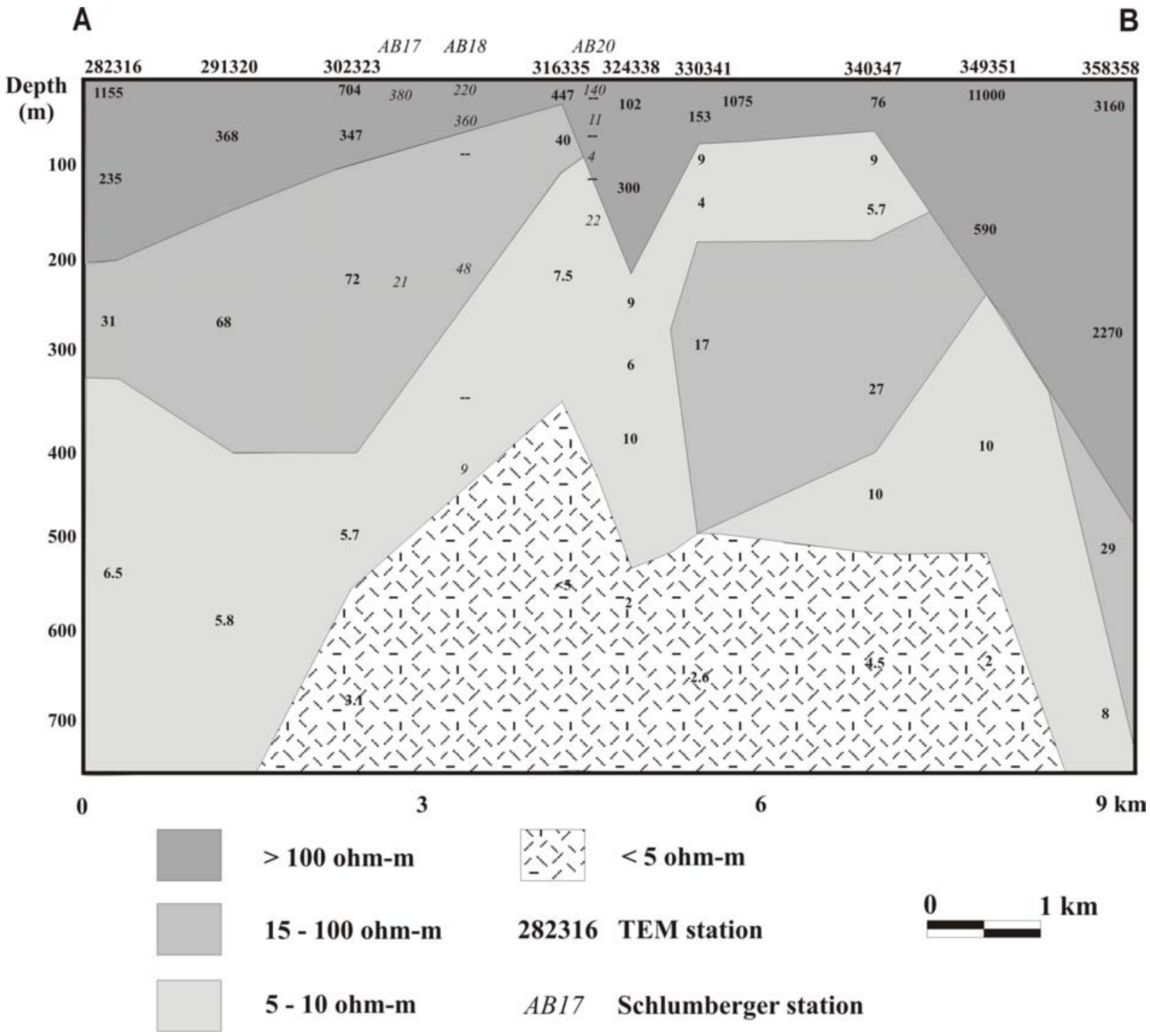

FIGURE 11: TEM cross-section A-B (for location see Figure 9)

5.5.3 Results of DC soundings

For comparison, results from an old minor DC resistivity survey were also looked at. The survey was conducted in 1973 by the Geothermal Division of Orkustofnun (The National Energy Authority of Iceland). Four Schlumberger soundings were measured at or close to the Geysir high-temperature geothermal area (AB17-20). It is interesting to compare the results of these with the TEM soundings. The soundings are shown in Figure 12 and in Appendix II.

Sounding AB20 (Figure 12) is located close to Geysir and hence to TEM sounding 324338. It shows the same characteristics as the TEM sounding, i.e. a low resistivity layer at 65 m depth but with slightly increasing resistivity underneath, around 20 ohm-m. This might be interpreted as indication of the high-resistivity core, but at TEM station 324338 the resistivity values are lower, i.e. below 10 ohm-m.

The other soundings are all located south or southwest of the Geysir area. Sounding AB18 is approximately 3 km to the southwest from AB20 along the cross-section (Figure 11). It shows the same character as the nearest TEM sounding 316335, high resistivity at the surface, then 40 ohm-m down to 250 m depth where it reaches 9 ohm-m. However, it does not have the same resolution at depth as the

TEM sounding. Sounding AB17, still further to the south, shows resistivity of around 20 ohm-m below 70 m depth, but its resolution is poor due to its character and it does not detect the low-resistivity at 400 m depth as shown in the nearby TEM sounding 302323. Finally, AB19 that is about 3 km south of Geysir, shows relatively high resistivity in the uppermost 400 m, but indicates very low resistivity at deeper levels, at least below 10 ohm-m, similar to the resistivity structure seen to the west.

The comparison of the results of two methods shows that the TEM-method has a much better resolution at depth, whereas the DC soundings show a more detailed layering close to the surface (< 100 m).

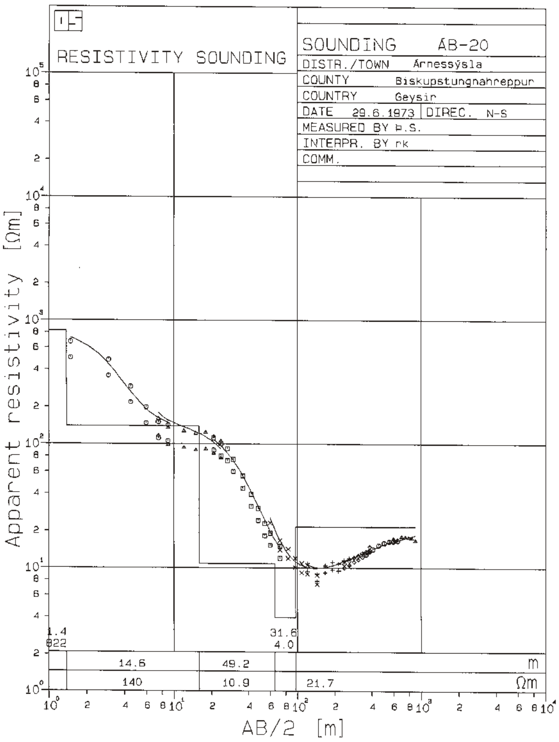

FIGURE 12: Schlumberger sounding AB20 with interpretation

5.5.4 Relationship between temperature and the resistivity structure

One of the most important aspects of the TEM resistivity survey at the Geysir geothermal field is to define the reservoir and its resistivity in general and try to establish the relationship between resistivity and temperature of the reservoir in particular. The temperature, and hence the thermal alteration, increases with depth. By water-rock interaction and chemical transport by geothermal fluids, the primary minerals in the host rock matrix are transformed, or altered into different minerals. The alteration process and the resulting type of alteration minerals are dependent on the type of primary minerals, chemical composition of the geothermal fluid, and temperature. The intensity of alteration is dependent on the temperature, but also on time and texture of the host rocks (Figure 13). Geothermal fluids can be divided into low-salinity fluids or fresh water, and saline fluids. The stability and formation of alteration minerals is mainly dependent on tem-perature (Árnason et al., 2000). The Geysir area has relatively fresh fluids.

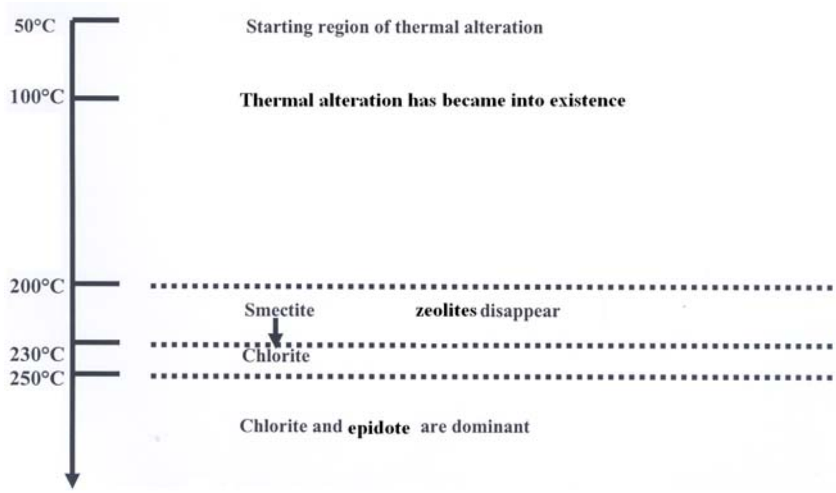

FIGURE 13: Diagrammatic representation of mineral alteration in a high-temperature geothermal system

No deep wells have been drilled inside the Geysir geo-thermal field, but well ND-1 at Nedridalur is 850 m deep and is found approximately 2 km southwest of the field. Figure 14 shows temperature log from the well. The well is artesian and yields 5 l/s of 70˚C water with the main aquifers at around 350 m depth. Below that, it has a very high temperature gradient, 220˚C/km, and the temperature at the bottom is around 170˚C. This high gradient obviously reflects the vicinity of the high-temperature field. The TEM soundings reveal the sequence of the temperature distribution of the rock formation at depth as indicated by resistivity and alteration, provided there is equilibrium between thermal alteration and temperature at present. The closest sounding is station 302323 that shows resistivity of 70-700 ohm-m in the uppermost 400 m of the well, where the well indicates temperatures below 70˚C. As has been discussed in the previous sections, this resistivity indicates low temperature and usually correlates with relatively fresh rocks, here basaltic lavas and hyaloclastite and un-altered rhyolite. Below 400 m depth, should be the smectite-zeolite zone, as indicated by temperatures of 100-200˚C, and resistivity below 10 ohm-m.

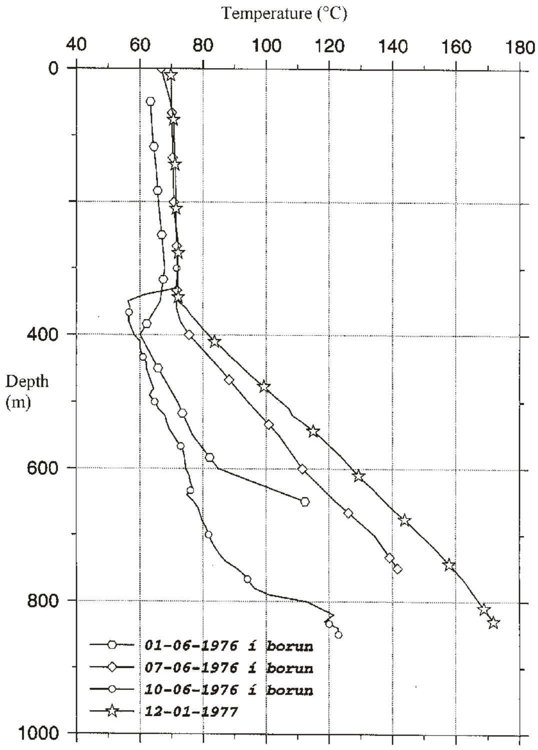

FIGURE 14: Temperature logs in well ND-1 at Nedridalur 2 km southwest of the Geysir area

6. ADVANTAGE OF THE TEM METHOD OVER CONVENTIONAL DC-METHODS

Based on experience and comparison of methods, various advantages of the TEM sounding method over conventional DC resistivity methods can be summarised as follows:

- The transmitter couples inductively to the earth and no current has to be injected into the ground. In the places where the surface is dry and resistive, this is essential.
- The monitored signal is the decaying magnetic field, not the electric field, which makes the TEM method much less dependent on local resistivity conditions at the receiver site. Distortion due to local resistivity inhomogeneities at the receiver site can be a severe problem in DC measurements.
- With respect to the DC method, TEM is less sensitive to lateral resistivity variations.
- In DC soundings, the monitored signal is low when subsurface resistivity is low, as in geothermal areas, whereas in TEM soundings the situation is the reverse. The lower the resistivity, the stronger the signal.
- Central-loop TEM soundings are more downward focussed than DC soundings. To increase the depth of penetration in DC measurements, it is necessary to increase the electrode spacing and this involves a large volume of rocks that can affect the monitored electric field significantly. This makes one-dimensional inversion better justified in the interpretation of central-loop TEM soundings, than in DC soundings.
- Performing a TEM sounding takes less manpower than performing a DC-Schlumberger sounding.
- With the instruments used in this survey, the DC soundings have a better resolution in the upper most 100-200 m, whereas the TEM soundings have a better resolution at depth.

7. SUMMARY AND CONCLUSIONS

A total of 15 stations were interpreted from a TEM resistivity survey conducted at and in the vicinity of the Geysir geothermal field. Results of a previous DC resistivity survey and data from well ND-1 at Nedridalur are also used for comparison. The data interpretation of the TEM survey was done by the one-dimensional inversion program CLTINV. The interpreted data are presented in the form of a cross-section and data curves and models.

Results of the TEM sounding survey at the Geysir geothermal field reveal the sequence of the resistivity distribution of the rock formation in the uppermost kilometre. The resistivity in the uppermost 100-400 m is quite high, ranging between 200 and 3000 ohm-m. This high resistivity close to the surface is correlated with fresh basalts and un-altered rhyolitic rocks. Thermal alteration is minimal in these rocks. Below that, the resistivity is low, and at some level it is obviously influenced by the high-temperature geothermal activity, indicated by resistivity of 1-10 ohm-m. The lowest values are found close to the Geysir area.

The resistivity cross-section through the Geysir high-temperature field shows the characteristics of a geothermal system with very low resistivity at depth, 1-5 ohm-m, around the Geysir area, but without confirmed evidence of high resistivity in the central part of the geothermal system. Provided there is equilibrium between thermal alteration of the rock and present temperature, the resistivity survey might indicate temperatures as high as 200-220˚C in the uppermost kilometre of the geothermal system, but not exceeding that, as there is no sign of a high-resistivity core. Chemical evidence indicates temperatures of 220-240˚C in the geothermal system, agreeing fairly well with the resistivity.

The well ND-1, about 2 km southwest of the Geysir fields, is located close to TEM station 302323 and shows that the temperature gradient is very high, or around 220˚C/km. This indicates that below 350-400 m depth (temperatures above 70˚C), the smectite-zeolite zone has been reached. This alteration zone is associated with quite low resistivity, here < 10 ohm-m. The well penetrates the outskirts of the geothermal system that results in this high thermal gradient.

TEM soundings are more effective in the survey of geothermal fields than the conventional DC method. There may be various reasons for this conclusion, however, the main one is that TEM has more depth penetration than the DC method, and a better resolution at depth.

ACKNOWLEDGEMENTS

I would like to express my deepest acknowledgement to Dr. Ingvar B. Fridleifsson and Mr. Lúdvík S. Georgsson, for giving me a chance to be one of the UNU Fellows of the year 2003. I am happy to convey my gratitude to Mrs. Gudrún Bjarnadóttir for her assistance and support during my study in Iceland. My sincere gratitude also goes to my advisors Knútur Árnason and Ragna Karlsdóttir for their guidance, comments and critical review of this report. I would also like to extend my thanks to the geophysics division of ISOR and my special thanks to Hjálmar Eysteinsson, Karl Gunnarsson and Ingvar Thór Magnússon for the their valuable lectures and for sharing their experience. I am also grateful to the Geological Survey of Ethiopia for supporting my stay in Iceland.

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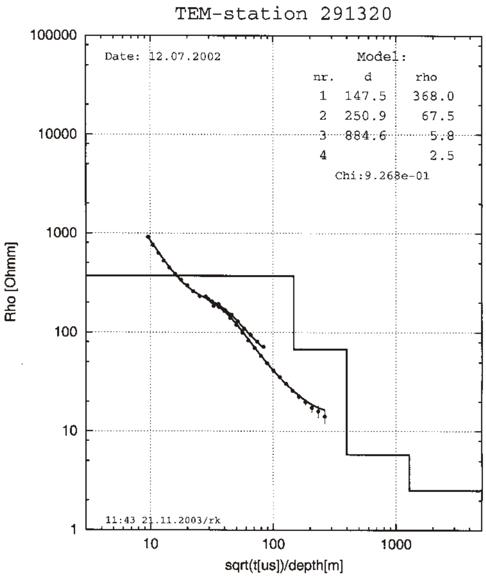
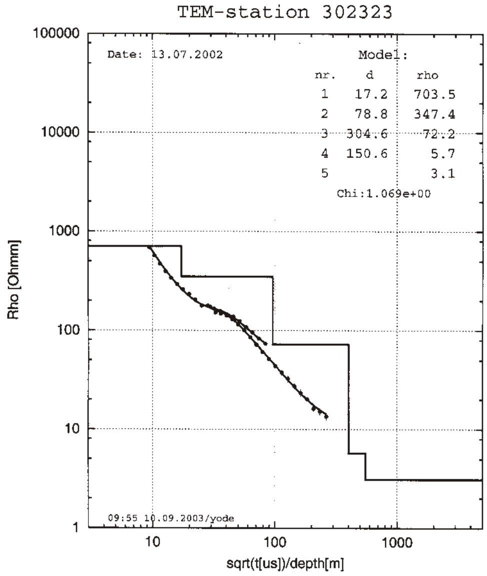
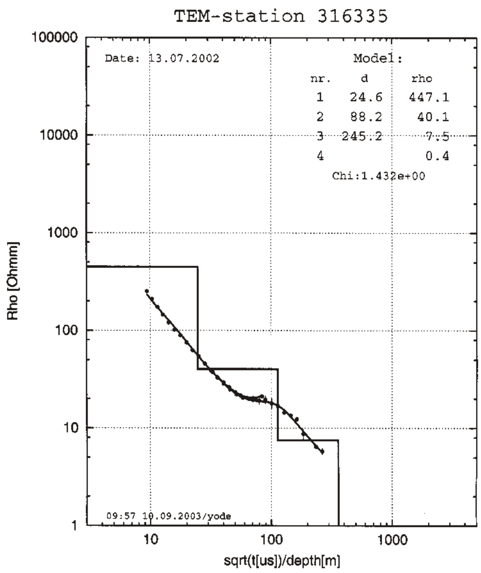
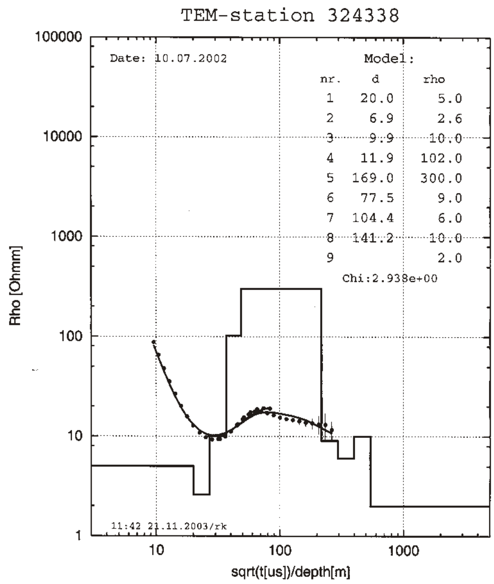
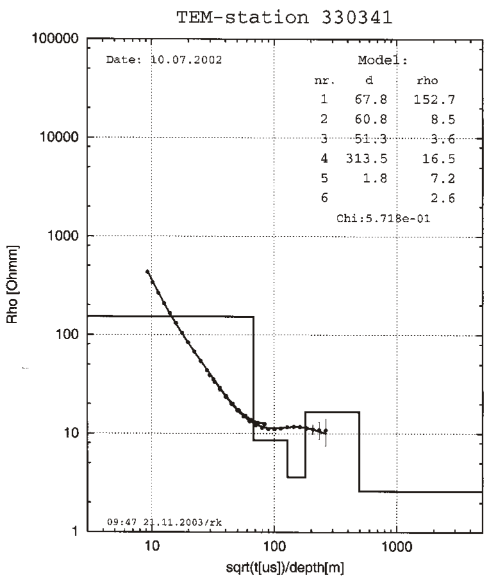
APPENDIX I: TEM soundings in the Geysir area and their 1D interpretation

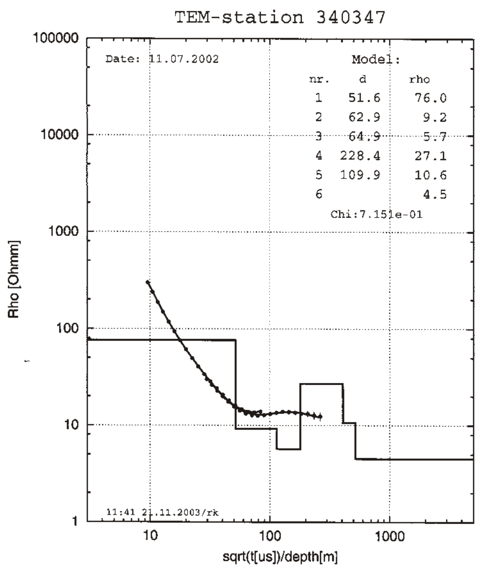


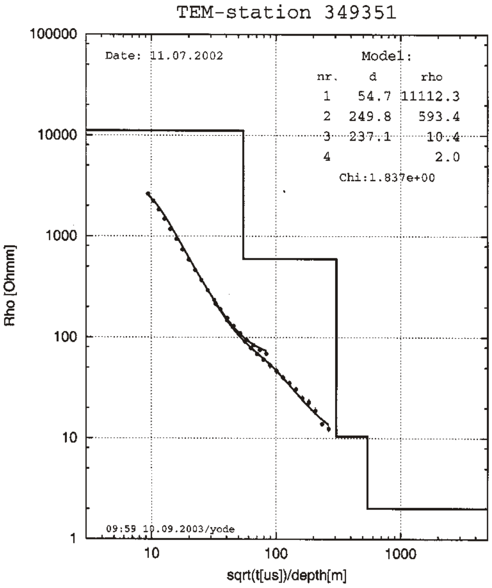


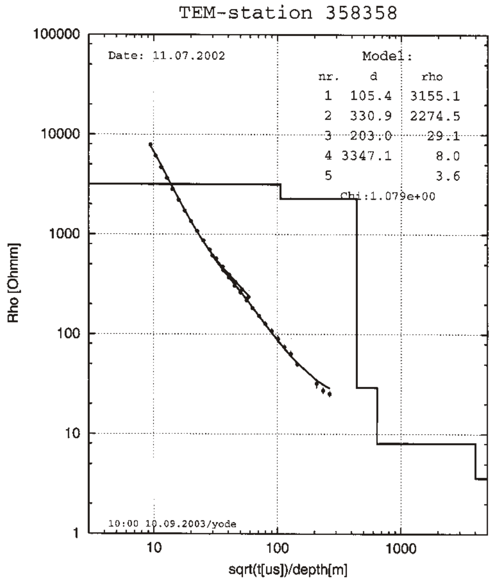












APPENDIX II: Schlumberger soundings in the Geysir area and their 1D interperation