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GEOPHYSICAL EXPLORATION IN BOLUNGARVÍK, NW-ICELAND AND IN ÁRSKÓGSSTRÖND, N-ICELAND

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ABSTRACT

For almost 50 years geophysical exploration has been applied in Iceland based on physical methods. A wide variety of exploration techniques are used. Among the most common methods in regional surveying are resistivity soundings and temperature gradient measurements in shallow drillholes, to test the range of change in temperature with depth. In Bolungarvík NW-Iceland, resistivity soundings were used in an attempt to find a low-resistivity anomaly that could indicate a geothermal system at depth. The results indicate that further exploration should be limited to the area around the warm spring in Sydridalur. In Árskógsströnd, N-Iceland, shallow drillholes were used to detect the thermal gradient in the area. A local anomaly of high thermal gradient indicates an up-flow zone of geothermal water. For further exploration it is suggested to drill at least a 300 m deep well.

Both in Bolungarvík and Árskógsströnd, the surveys were aimed at finding thermal water for district heating systems in nearby towns.

1. INTRODUCTION

Geophysics is the study of the parameters of the earth. In geothermal systems the main parameters are, temperature, permeability and chemical composition of the fluid. The above parameters affect the rocks, hence the physical properties of rocks can change. The electrical resistivity of rocks is dependent on those parameters, therefore electrical methods are used to delineate geothermal systems. TEM (transient electromagnetic) is a geophysical method which measures the underground resistivity from the surface. A magnetic field is created in the earth by sending current through a loop of wire on the ground. The current is cut off and the response signal from the decaying magnetic field is measured in the centre with a receiver. This signal can be interpreted on terms of the resistivity structure of the underlying earth. A thermal survey, measuring the temperature in shallow drillholes covering an area , is used to detect the thermal gradient in the area, thus determining positive thermal anomalies combined with geothermal up-flow zones.

This report deals with geothermal exploration using TEM soundings in Bolungarvík, NW-Iceland, comparing them with older Schlumberger dc-resistivity soundings. Also it covers a survey determining the temperature gradient by measuring the temperature in a set of shallow wells in the Árskógsströnd area N-Iceland. The aim of both surveys is to look for localized geothermal heat flow which might indicate a geothermal system in order to develop a central heating system in both areas. This project is the final part of the United Nations University Geothermal Training Programme undertaken by the author at Orkustofnun, Reykjavík, extending from April to October 1997.

2. GEOTHERMAL ENERGY IN ICELAND

Extra-ordinary volcanic magmatic activity in one part of the Mid-Atlantic ridge resulted in the Icelandic landmass. Iceland is a most spectacular geological phenomenon as it lies across the margin of the North American and Eurasian plates. The Mid-Atlantic ridge lies across Iceland from southwest to northeast and this zone is characterized by spreading and volcanism.

The deepest well drilled in Iceland confirms a minimum of 4 km thickness of the lava pile. The lava pile dips gently on a regional scale, generally toward the central part of Iceland, i.e. the active volcanic zone.

The predominantly volcanic pile of Iceland, which ranges in age back to about 16 m.y. is conventionally divided into four stratigraphic groups or series. This division is based on climatic evidence from inter lava sediments or volcanic breccia and/or palaeomagnetic reversal patterns supported by absolute age data (Saemundsson, 1979). The four groups are:

Postglacial: Last 9000-13000 years.
Upper Pleistocene: Back to 0.7 m.y., corresponding to the present normal magnetic epoch, Brunhes.
Plio-Pleistocene: 0.7-3.1 m.y., includes the Matuyama epoch and the Gauss epoch upward of the Mammoth event.
Tertiary: Rocks older than 3.1 m.y.

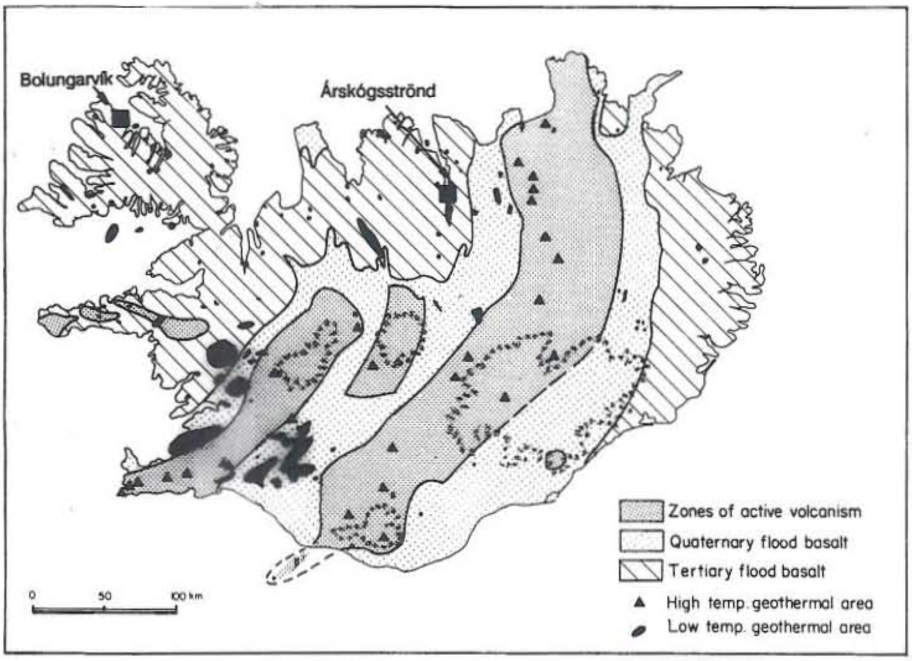


FIGURE 1: Geological structure of Iceland and location of high- and low-temperature fields

The magmatic activity and volcanism results in high heat flow in the crust of Iceland. Like at many constructive plate boundaries the mean heat flow decreases with increasing distance from the volcanic rift zone (Fridleifsson, 1979).

Figure 1 shows the main geological structures and location of geothermal fields in Iceland. Geothermal fields in Iceland are divided into two main groups, high-temperature and low-temperature fields.

2.1 High-temperature fields

The high-temperature fields lie within the active zones of rifting and volcanism. According to the plate tectonics theory, the highest heatflow on a constructive plate margin should be along the volcanic zone, which is the surface expression of the plate boundary. These areas are always associated with volcano-tectonic features such as volcanic fissure swarms or more commonly central volcanoes with intermediate and acid volcanic activity, fault swarms and sometimes calderas (Fridleifsson, 1979). The strata of the active high-temperature areas are, like the Plio-Pleistocene strata composed of layers of subaerial lavas intercalated by thick piles of subglacially erupted pillow lavas and hyaloclastites. The rate of temperature change or thermal gradient in the high-temperature areas is more than 200°C/km. The main heat source of a high-temperature field is the cooling of volcanic intrusions. Isotope studies indicate that hydrogeological cycle in the high temperature fields are much more localized than those of the low-temperature areas. Over the past few years many high-temperature fields have been delineated by the use of geophysical exploration methods.

A few geothermal power plants have been built in Iceland. An electrical power plant is in Krafla, a high-temperature field in N-Iceland, and co-generation plants are in the Nesjavellir and Svartsengi high-temperature fields, SW-Iceland that produce both electricity and hot water for central heating by heating cold groundwater with steam.

High-temperature areas in Iceland cover an area of almost 500 km^2. The total natural heat discharge of the high-temperature fields is not known, but has been estimated at about 4000 MW (Fridleifsson, 1979).

2.2 Low-temperature fields

The uppermost part of the Icelandic crust is composed of subaerial lavas and much subordinate airborne tuff in the Tertiary areas, but of subaerial lavas intercalated with moranic horizons and thick piles of subglacially erupted pillow lavas and hyaloclastites in the Plio-Pleistocene provinces, which flank the active volcanic zone (Fridleifsson, 1979). The low-temperature fields appear mainly in the Plio-Pleistocene or Tertiary volcanics toward two sides of the rift (Figure 1). In the sub-aerially erupted Tertiary volcanics, the flow channels for the geothermal fluid appear to be mainly dykes and faults, but to a lesser extent high-porosity stratiform horizons. In the Plio-Pleistocene strata, which are characterized by a succession of subaerial lavas intercalated with thick piles of subglacially erupted pillow lavas, hyaloclastites and detrital beds, potential flow channels are much more abundant. The main heat source in the low-temperature field is regional heat flow with maximum temperature of the fields in the uppermost 1 km $\leq$150°C. The geothermal fluids in the low-temperature fields are frequently confined within near-vertical aquifers. The general conceptual model is that water is heated up at depth and flows towards the surface along some permeable structure, such as a dyke, a fault or a fracture. The total natural discharge from all hot springs in low-temperature areas in Iceland with temperature higher than 15°C has been estimated to be around 2000 l/s and their estimated total thermal power equals 500 MW (Pálmason et al., 1985). The bulk of the hot springs has a flowrate of less than 5 l/s, but both in Plio-Pleistocene and Tertiary strata individual springs may have a flowrate of several tens of l/s.

Geothermal research in Iceland, both in high- and low-temperature fields, consists of geological studies and geophysical exploration supported by geochemical studies. The conventional methods for delineating a geothermal area are described in this report. One geophysical method measures the electrical resistivity of the earth by sounding electromagnetic signals and receiving their response signals. Another is the drilling of shallow drillholes and measuring the temperature in order to calculate the geothermal gradient in the area.

3. TEM SOUNDING SURVEY IN BOLUNGARVÍK

Bolungarvík is a small village with a population of little over 1000, located in the southern part of Ísafjardardjúp bay NW-Iceland upon Tertiary formations. A geothermal project was launched this year by Orkustofnun, (National Energy Authority of Iceland) in an attempt to find a geothermal system that could provide Bolungarvík with hot water for a district heating service. For this purpose Orkustofnun carried out geophysical studies using resistivity methods (TEM) in Bolungarvík and a temperature gradient survey in Ísafjördur in order to find a possible up-flow zone. The geophysical exploration in Bolungarvík is a part of this report. Furthermore, three older Schlumberger dc-soundings were compared to the results of the TEM soundings. The aim of this exploration was to find a low-resistivity anomaly that could indicate a geothermal system at depth and then define the location of a deep exploratory well. The area is shown in Figure 2.

Underground water in the area is controlled by the direction of geological structures, i.e. dykes and faults. The regional trend of faults and dykes is NE-SW, but there is also a local NW-SE trend found in the area. The main direction of geological structures in Bolungarvík is NE-SW. The only warm spring in the area appears along a dyke with this direction, in the Sydridalur valley and has a surface temperature of 27°C.

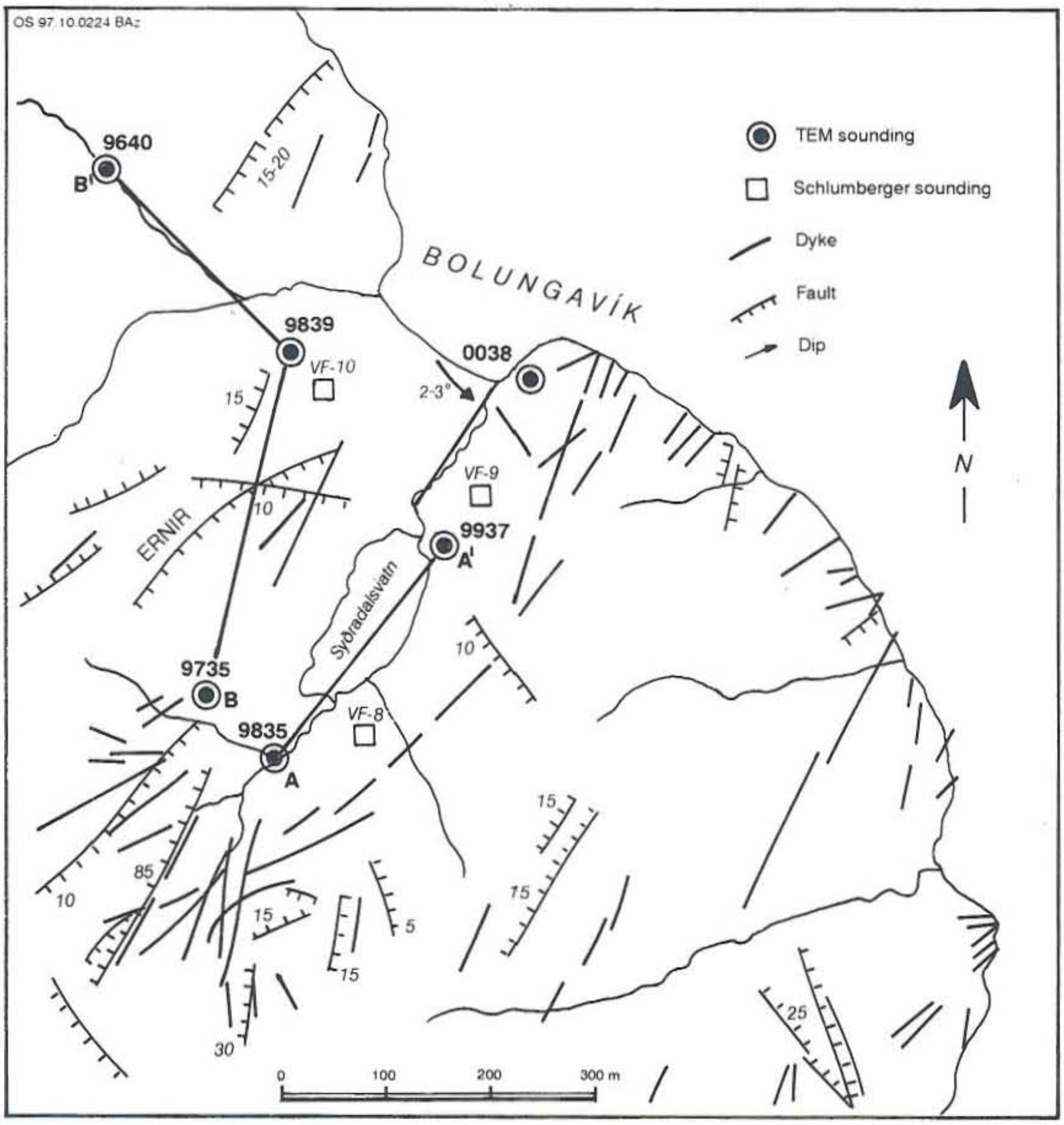


FIGURE 2: Geological structures and location of TEM and Schlumberger soundings in Bolungarvík

3.1 Geological setting

The northwest part of Iceland is a mountainous peninsula with mountains rising up to an elevation of 1000 m. This peninsula was constructed by erupting lavas from spreading centres of northeasterly strike. The total thickness of the basaltic lava pile in this peninsula is about 6-7 km and the strata dips gently 3-6° at sea level toward the southeast. The lava pile comprises mainly tholeitic flood basalt from 8 to over 15 million years old, middle-late Miocene (McDougall et al., 1984). The K-Ar dating studies on rock samples from near the town of Ísafjördur measured the average age of 14.1±0.2 ma.

The lava pile in this region (Saemundsson, 1979) consists of subaerial lava flows with scoriaceous flow tops, interbedded with minor sediments. It is affected by two relatively dense sub-vertical fault and dyke systems. The major fault system strikes NE-SW and the minor system strikes a NW-SE direction. Most hot springs and hot water manifestations in the area show these trends. In the exploration area, the main direction of the geological structures is NE-SW that is the regional direction (Figure 2). Observed vertical displacement of faults in this area ranges from 1 m up to 40 m. Fault zones are typically 1-5 km wide. The composition of dykes is mainly basaltic, and the main strike of dykes is NNE-SSW and NE-SW, and the thickness is 2-10 m. Most of the dykes are subvertical and strike in the direction of about N30°E. They are highly jointed and it is difficult to estimate dyke density and tectonic activity.

A NE-SW trending fault zone appears in the western part of the Sydridalur valley in Bolungarvík. The exposure of the fault zone lies in the mountains lining of Sydridalur valley to the south. This fracture zone turned out to be highly permeable when a tunnel was cut through the zone some years ago (Gudmundsson, 1991). In this exploration project, an attempt is made to see if low resistivity is combined with this fault zone. The warm spring in Sydridalur valley seems to lie close to the margin of the fault zone. On the other hand, it is very difficult to site a sounding close to the fault-zone due to steep mountains and difficult terrain.

3.2 Electrical resistivity methods in geothermal exploration

Electrical resistivity methods are among the most important geophysical methods used to delineate a geothermal area. The resistivity is directly related to parameters like salinity, temperature, alteration and porosity (permeability). The principle of electrical resistivity methods can be described by Ohm's law as follows:

$$E = \rho j \tag{1}$$

where

- E = Electrical field strength (V/m);
- j = Current density (A/m^2);
- ρ = Specific resistivity (Ωm), the reciprocal of resistivity is conductivity ($1/\rho = \sigma$).

Alternately, resistivity can be defined with the ratio of potential difference to the current as follows:

$$\rho = \Delta V / I \tag{2}$$

Electrical conductivity in the minerals and solutions takes place by the movement of electrons and ions. The common principle of all resistivity sounding methods is to induce an electrical current in the earth and monitor signals at the surface generated by the current distribution.

For water-saturated rocks (low salinity of fluid), the bulk resistivity of rocks is practically independent of the salinity and only to a small extent to the temperature of the fluids. Resistivity variations in the uppermost kilometre of the crust, therefore, primarily reflect variation in the amount of interconnected pores and fractures of the rock (Flóvenz and Georgsson, 1982). On the other hand the salinity of electrolytes, such as NaCl in water, affects resistivity in a nearly inversely linear manner (Keller and Frischknecht, 1966). For T = 0°C we have

$$\rho = 9.545\,C^{-0.937} \approx 10/C \tag{3}$$

At moderate temperatures, 0-200°C, the resistivity of aqueous solution decreases with increasing temperature. The reason is increased mobility of the ions caused by a decrease in viscosity of the water (Björnsson and Hersir, 1991).

Dakhnov (1962) has described the following relationship:

$$\rho_W = \frac{\rho_{W_o}}{1 + \alpha (T - T_o)} \tag{4}$$

where

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

It has been observed for many cases that resistivity of water-saturated rocks varies approximately as the inverse square of the porosity. This empirical relationship is called Archie's law. It describes how resistivity depends on porosity if ionic conduction in the pore fluid dominates other conduction mechanism in the rock

$$\rho = \rho_W\, a\, \phi_t^{-n} \tag{5}$$

where

- ρ = Bulk resistivity (Ωm);
- ρ_W = Resistivity of the pore fluids (Ωm);
- ϕ_t = Porosity in proportion of total volume;
- a = An empirical parameter, varies from less than 1 for intergranular porosity to over 1 for joint porosity (usually around 1);
- n = Cementing factor, an empirical parameter, varies from 1.2 for unconsolidated sediments to 3.5 for crystalline rocks (usually around 2).

It can be shown that a regional resistivity survey can be used to estimate the size of a geothermal system and to map the flow pattern of thermal waters from its origin in the highlands to the geothermal area in the lowlands (Flóvenz and Georgsson, 1982; Flóvenz et al., 1985; Eysteinsson et al., 1993).

3.3 The central loop TEM sounding survey

Time domain or transient electromagnetic (TEM) method has been used in geothermal exploration to define the resistivity structure of the earth. The popularity of this method in geothermal exploration has been increasing. The method has several advantages over the conventional Schlumberger DC-sounding (Árnason, 1989; Björnsson and Hersir, 1991), such as.

- The transmitter couples inductively to the earth and no current has to be injected into the ground; this is most important where the surface is dry and resistive like on dry fresh lava or in deserts.
- The monitored signal is a decaying magnetic field, not an electric field at the surface. This makes the results much less dependent on local conditions at the receiver site.
- This method is much less sensitive to lateral resistivity variation than the DC-methods.
- In DC-sounding the monitored signal is low when subsurface resistivity is low, like in a geothermal area, whereas in TEM soundings the situation is the reverse, the lower the resistivity the stronger the signal.
- The TEM method is sensitive to broad band electromagnetic noise. The method is, therefore, hard to apply close to electrical power lines and other similar devices.
- The central loop configuration is more downwardly focussed than other resistivity methods. This implies that resistivity structures with relatively strong lateral variation of resistivity can be mapped with 1-D inversion of central loop TEM soundings (Árnason and Flóvenz, 1992).
- Two main disadvantages of the TEM method are relatively slow speed of operation and bulky equipment, which necessitates accessibility of sounding sites by vehicles. In Iceland this is solved by performing the TEM sounding in late winter when the land is covered by snow and, therefore, easily accessible by snowscouters.

3.3.1 Principle of method and data processing

In the central-loop TEM sounding method the current in the ground is generated by a time varying magnetic field. This is done by injecting time variant current, normally a step function, into a grounded dipole or loop of wire at the surface. It is different from the magnetotelluric method (MT) in that the magnetic field is not randomly varying, but a field of a controlled magnitude generated by a source loop. A loop of wire is placed on the ground and a constant magnetic field of known strength is built up by transmitting a constant current into the loop.

After turning the current off abruptly, the decay of the magnetic field induces electrical current in the ground and the current distribution in the ground induces a secondary magnetic field decaying with time. The decay rate of this secondary field is monitored by measuring the induced voltage in the receiver coil at the center of the transmitter loop. Figure 3 shows the configuration of central loop TEM sounding.

Depth of exploration is also increased by increasing the source-receiver distance. The current distribution and the decay rate of the secondary

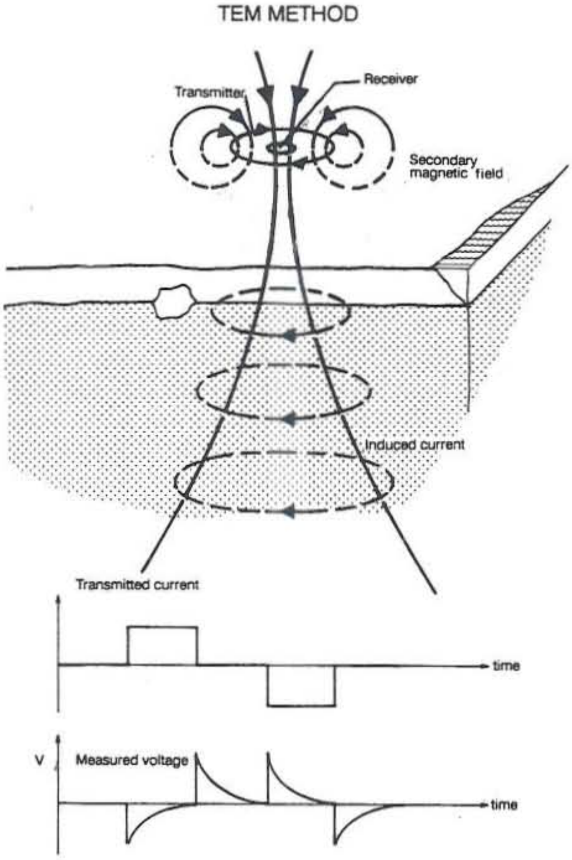


FIGURE 3: The central-loop TEM sounding configuration (Hersir and Björnsson, 1991)

magnetic field depends on the resistivity structure of the earth. In this method depth of penetration depends on the geoelectrical section and the time of induction in the receiver can be traced in time before it is drowned in noise.

After current, I, is turned off, the induced voltage in the receiver coil with an affective area A_r placed at the centre of a circular transmitter loop of radius r is given, as a function of time t, by the cosine transform integral

$$V(t,r) = \frac{2}{\pi}\int_0^\infty Re\left[\frac{V(\omega,r)}{-\omega i}\right]\cos(\omega t)\,d\omega \tag{6}$$

In the case of a one-dimensional (1-D) resistivity structure on a horizontally layered earth, the frequency domain response $V(\omega\,,t)$ can be expressed by the Hankel transform integral

$$V(\omega,r) = -IA_r i\omega\mu_0 r\int_0^\infty \lambda\frac{S_0}{S_0-T_0}J_1(\lambda r)\,d\lambda \tag{7}$$

The quantities S_0 and T_0 are determined recursively in terms of frequency and the resistivity and thickness of the successive layers. For a homogeneous half-space, the induced voltage in the receiver coil at the centre of the loop can be as follows:

$$V(r,t) = IA_r\frac{\mu_0 r^2}{20\pi^{\frac{1}{2}}}\frac{(\mu_0\sigma)^{\frac{3}{2}}}{t^{\frac{5}{2}}} \tag{8}$$

For large t, the sounding response is frequently expressed in terms of late time apparent resistivity which is defined by solving Equation 3

$$\rho_a(r,t) = \mu_0\left[\frac{IA_r\mu_0 r^2}{20\pi^{\frac{1}{2}}t^{\frac{5}{2}}V(r,t)}\right]^{\frac{2}{3}} \tag{9}$$

t	= Time elapsed, after the current in the transmitter loop is turned off (s);
A_r	= Cross sectional area of the receiver coil (m^2);
I	= Current (A):
μ_0	= Magnetic permeability in vacuum (Henry/m);
r	= Radius of transmitter loop (m);
$V(r,\,t)$	= Transient voltage (V).

At present, only a 1-D inversion of TEM sounding is commercially available, but as discussed earlier, the resolving power of a 1-D inversion of TEM sounding is often comparable with 2-D modelling of DC data. The apparent resistivity curve can be interpreted in a PC-computer using the inversion program TINV, developed at Orkustofnun (Árnason, 1989). The aim is to determine the layered earth model whose response reproduces the measured values as closely as possible. The program is a non-linear least square written in standard FORTRAN 77 and can be implemented on any machine supporting FORTRAN 77 (Árnason, 1989).

The program assumes that the filed data is collected with equipment where the current transmitted in the loop is turned off linearly from maximum to zero and that the time values, at which the apparent resistivity values are given, are equally spaced in logarithm of time after the current has become zero. The program assumes that data is collected with a circular loop. If this is not the case, the actual transmitter loop is simulated by a circular loop having the same area.

3.3.2 Instrumentation

The central loop TEM sounding field equipment used in Iceland are Protem/EM 37 from Geonics Ltd., consisting of a current transmitter, receiver box, generator, receiver and transmitter loops. A small coil with an affective area of 100 m^2, a flexible loop with an affective area of 8112 m^2 and a square transmitter loop of 300 m side length were used. In some cases, a 200 m side length is thought to be sufficient. Dependent on the terrain the transmitted current is usually in the range of 20-23 A, and the transient signal is recorded in the time interval of 0.087-70.4 ms at 20 channels after the current turn-off. Both for transmitter and receiver, timing is controlled by synchronised high precision crystal clocks. The induced voltage is measured by the receiver each time the transmitted current is turned off. Data were recorded in the frequency domain (high-low). At high frequency the repetition rate of transmitted current signals is 25 Hz, in 10 ms time between current on and current off. At low frequency the repetition rate is 2.5 Hz, with current on and current off intervals of 100 ms. Then the data is edited to remove electromagnetic noise to obtain induced voltage and finally, apparent resistivity is calculated as a function of time.

3.4 Discussion of results

The TEM survey in Bolungarvík consists of 6 soundings covering an area of 20 km^2 carried out in July 1997 by Orkustofnun, and interpreted by the author simultaneously. Figure 2 shows the location of the soundings. Most of the soundings are located in the Sydridalur valley. The soundings were interpreted using the 1-D inversion program TINV. For a given number of layers, the program chooses the resistivity values for each layer. The soundings in this area required four or five layers to get a best possible fit. One has to bear in mind that the TEM equipment used here does not give an acceptable account of the uppermost 100 m of the soundings. The soundings are divided into two groups, 9835-9937-0038 and 9839-9735-9640. Table 1 shows the results from 1-D inversion of the TEM soundings.

TABLE 1: Results of TEM soundings in Bolungarvík

9835		9937		0038	
Resistivity (Ωm)	Thickness (m)	Resistivity (Ωm)	Thickness (m)	Resistivity (Ωm)	Thickness (m)
40.8	14.0	81.1	46.1	24.9	28.4
583.4	46.6	384.7	29.1	69.5	100.5
136.1	91.0	218.0	55.1	2.6	17.7
26.3	134.3	57.1	72.1	150.9	611.9
235.8		138.2		7.1	
9735		**9839**		**9640**	
87.6	59.3	116.3	74.8	1773.2	12.5
328.2	242.4	592.6	220.7	145.5	234.9
141.7	348.1	43.8	311.0	46.8	29.4
68.9		103.7		26.5	183.4
				(6651.3	

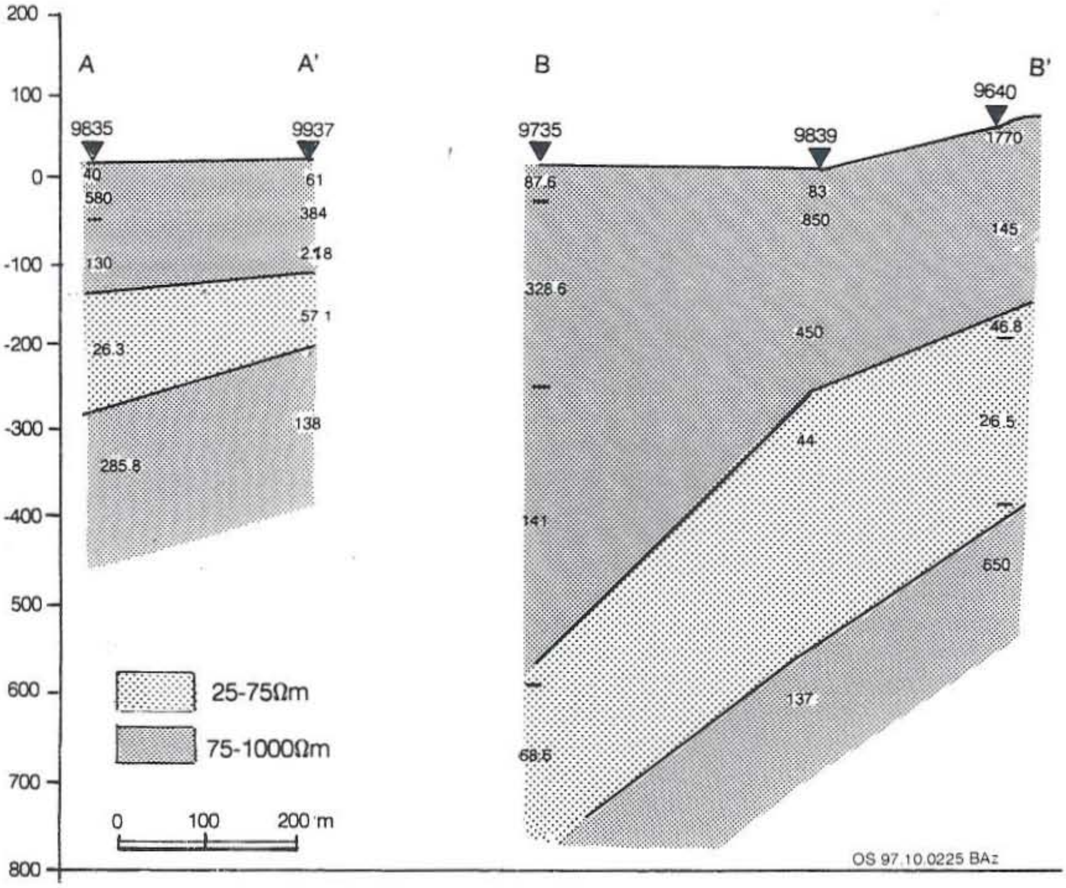


FIGURE 4: Resistivity cross-sections from the Bolungarvík area

The work started by data acquisition in the field in July 1997. As stated before, a good fit was acquired with four or five layers. The data in sounding 0038 is not good. Due to difficult terrain, a steep slope, a 200x200 m loop was used. The data turned out to be scattered and the most likely explanation is a high voltage power line nearby. The data could also be affected by the sea, which would cause a reduction in the measured resistivity. Appendix I shows the interpretation of these soundings.

Figure 4 shows the resistivity cross-sections AA′ and BB′ in the area. The location of the lines is shown in Figure 2. In the uppermost 100 m of the soundings the resistivity values and the thickness are not well defined. Still, it is clear that there is a thin layer at the surface with resistivity below 100 Ωm underlain by a layer with higher resistivity, usually 100-500 Ωm. This high-resistivity layer is 100-150 m thick in resistivity section AA' in the eastern part of the valley. It seems to thicken to the west and is 200-300 m in section BB' in the western part of the valley. In section AA', the high-resistivity layer is divided in two with higher resistivity at the top, 300-500 Ωm underlain by a slightly lower resistivity of 100-200 Ωm. Beneath these layers there is a distinctive low-resistivity layer in the whole area (25-75 Ωm). In section AA', this layer has a thickness of 100-200 m. It shows an increasing thickness to the west in section BB' (300-350 m). The resistivity is lowest in sounding 9835, which is close to the thermal manifestations on the surface (section AA'). The resistivity in this layer is higher, 40-70 Ωm, in all other soundings in the valley; the only exception is a 26.5 Ωm resistivity at depth in sounding 9640 in Hlídardalur. As we have no other soundings to support this extra low resistivity in Hlídardalur, that will be unexplained.

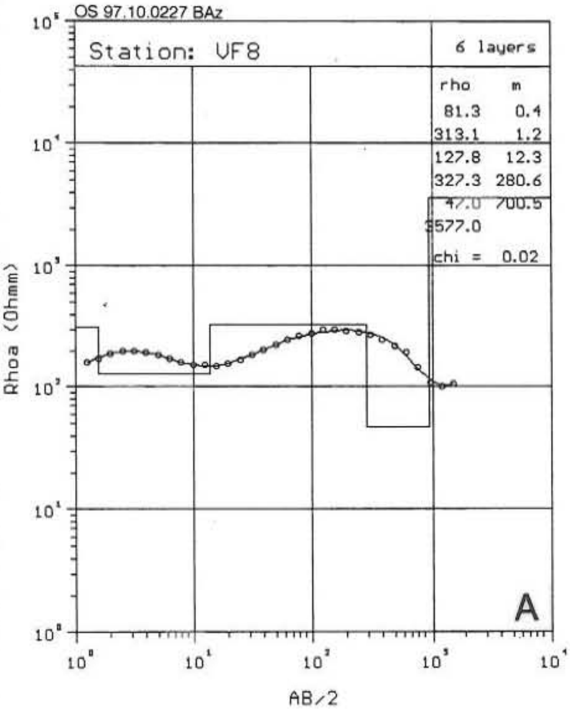
The low resistivity in sounding 9835 is possibly affected by the presence of a geothermal system. The geothermal up-flow is most likely confined to the fracture zone at the western margin of the Sydridalur valley. As the resistivity in the low-resistivity layer is clearly higher in all nearby soundings, the geothermal up-flow zone is probably very narrow. The low-resistivity layer is, in most soundings, underlain by a high-resistivity layer.

3.5 Comparison of TEM-results with Schlumberger DC sounding and borehole data

In DC methods such as Schlumberger, an electrical current is injected into the ground. The DC methods are more sensitive to resistivity changes at the surface. They give a better account of the resistivity in the uppermost 100 meters than TEM soundings which, on the other hand, have a better downward penetration. The Schlumberger soundings from Bolungarvík were interpreted by the use of the 1-D inversion program SLINV and the results compared to that of the TEM soundings nearby.

Figure 5 shows the comparison between inverted curves from two TEM and one Schlumberger sounding. The interpretation of the Schlumberger soundings in the area are shown in Appendix II. Both TEM and DC-Schlumberger soundings show the existence of a low-resistivity layer below 150 m depth. Table 2 summarizes the result of 1-D interpretation.

The first three layers in the Schlumberger sounding are in all 14 m thick and have 100-300 Ωm resistivity, which corresponds to the first layer in the TEM sounding with 15-45 m thickness and 40-80 Ωm resistivity. Layer 4 in the Schlumberger sounding, with 280 m thickness and 327 Ωm resistivity, corresponds to the second and third layers in the TEM sounding with



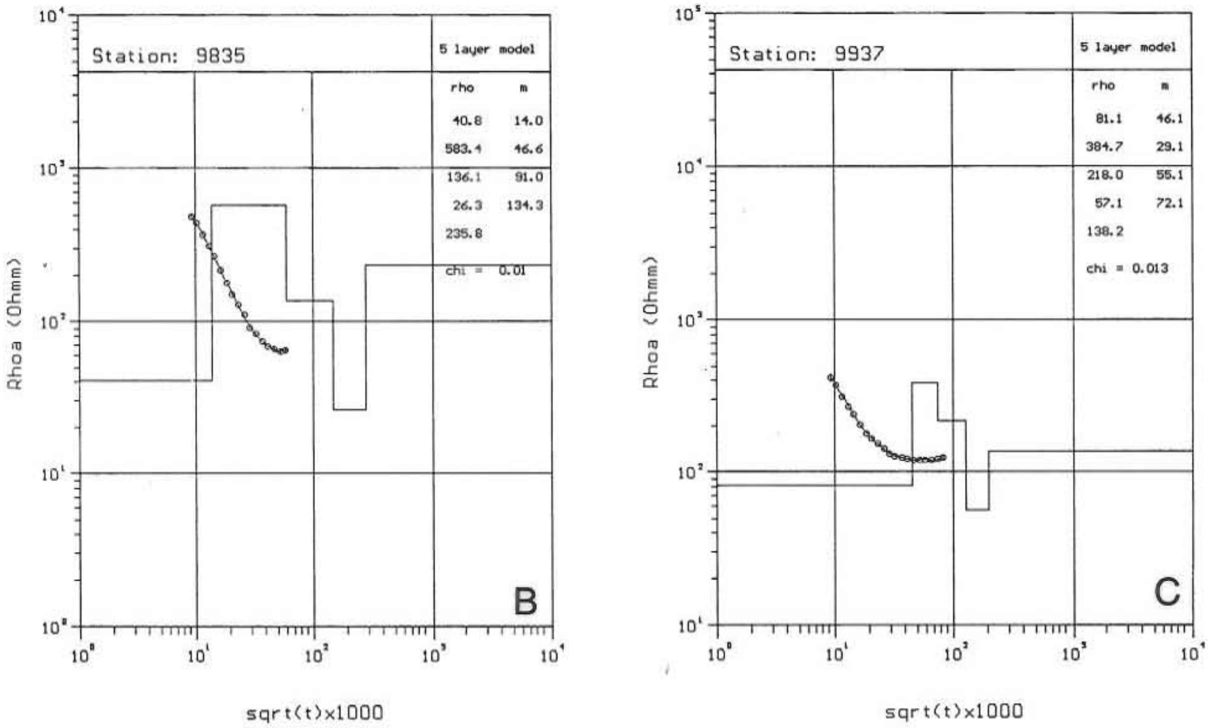


FIGURE 5: a) Schlumberger sounding VF-8, b-c) TEM soundings 9835 and 9937 in Bolungarvík with 1-D interpretation

30-45 m and 400-600 Ωm resistivity and 55-90 m and 135-220 Ωm respectively. The low-resistivity layer appears in both the DC and the TEM soundings. In the Schlumberger sounding the layer is 700 m thick and has a resistivity of 47 Ωm. The fifth layer in the TEM-soundings has 26-58 Ωm resistivity and 72-134 m thickness. The low- resistivity layer is underlain by high resistivity according to both methods.

TABLE 2: Comparison of results of two TEM soundings and one Schlumberger sounding, more or less in the same area in Bolungarvík

9835		9937		VF-8	
Resistivity (Ωm)	**Thickness (m)**	**Resistivity (Ωm)**	**Thickness (m)**	**Resistivity (Ωm)**	**Thickness (m)**
40.8	14.0	81.1	46.1	81.3	0.4
583.4	46.6	384.7	29.1	313.1	1.2
136.1	91.0	218.0	55.1	127.8	12.3
26.3	134.3	57.1	72.1	327.3	280.6
235.8		138.2		47.0	700.5
				3577	

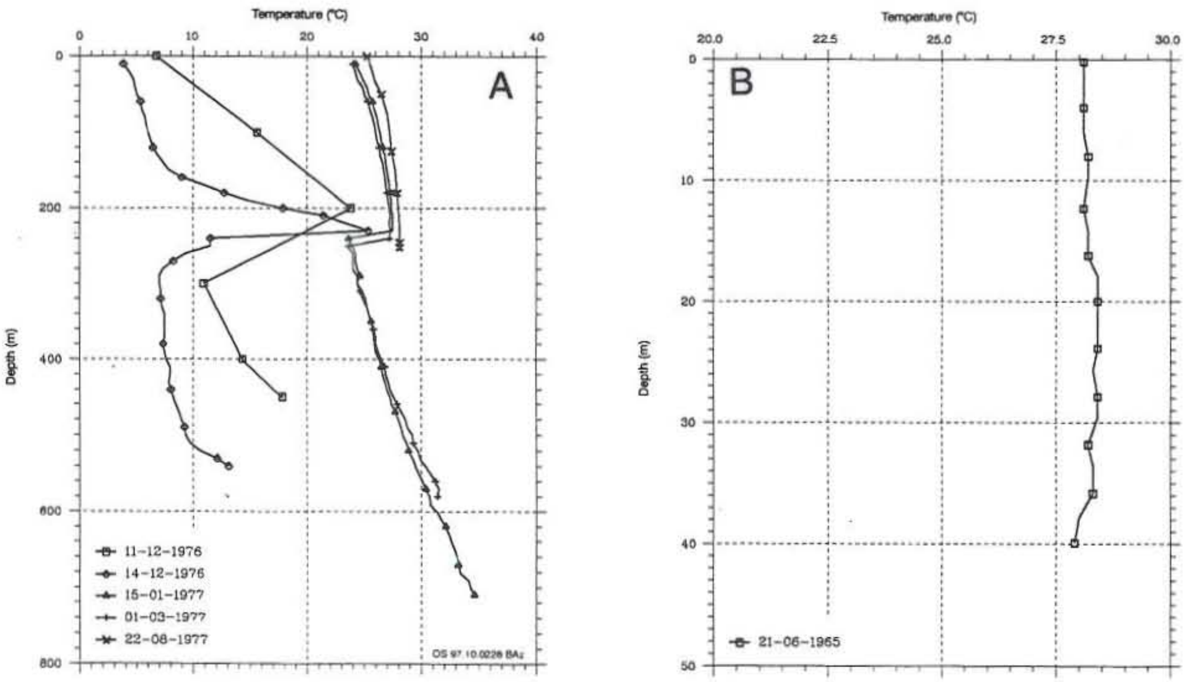


FIGURE 6: Temperature logs from a) well BH-3 and b) well BH-1 in Bolungarvík

Two boreholes, BH-1 (1965) 40 m deep and BH-3 (1976) 700 m deep, have been drilled in the Sydridalur valley close to the warm spring. Figure 6 shows the temperature measurements for the two boreholes. There is a good correlation between the temperature curve and the resistivity data at around 150-250 m.b.s.l depth, where the temperature curves from BH-3 clearly show an aquifer. BH-1, drilled in 1965 to 40 m depth, does not show a variation in temperature because of a flow of 28°C water from an aquifer near the bottom. Even though neither of the wells is good for an accurate assessment of the temperature gradient, it can be estimated to be around 44°C/km from well BH-3.

3.6 Conclusions and recommendations

In the Bolungarvík area, there is a warm spring in Sydridalur valley with a temperature of 27°C. TEM survey in the area was carried out in July 1997, consisting of 6 soundings in order to get good information on the distribution of resistivity at depth. The sounding area was about 20 km^2. The soundings were interpreted one-dimensionally by using a computer program TINV; most of the soundings had four or five layers. The resistivity cross-sections show the change in resistivity as a function of depth. The results of the TEM sounding were compared with three older Schlumberger soundings in the same area. The following can be said:

1. TEM soundings in Bolungarvík show a low-resistivity layer 25-75 Ωm at depth in all Sydridalur valley. It is 100-150 m thick in the eastern part, and 200-300 m thick in the western part of the valley. A clear resistivity low in sounding 9835 can be interpreted as a probable geothermal system at depth as the only geothermal warm spring in the area is close by.

2. Data acquisition in sounding 0038 failed due to difficult terrain and nearby high-voltage power lines.

3. By looking at a geological map of the area, it is possible to see the dip of faults at both sides of Sydridalur. The dip in the eastern part is opposite to that of the western part, so a graben is formed. And the geothermal up-flow is probably controlled by the fracture zone.

4. On the grounds of this survey it is recommended to limit further exploration to the area around the warm spring in Sydridalur (at sounding 9835). It is recommended to drill an exploratory drill-hole in the vicinity of the warm spring, but to try to get as close to the fracture zone in the mountain slope as possible.

5. A temperature gradient survey will be carried out in Ísafjördur in October 1997, the results will be compared to the results from the Bolungarvík survey before a decision is made on drilling a possible production well.

4. GEOPHYSICAL EXPLORATION IN ÁRSKÓGSSTRÖND

Árskógsströnd (Figure 7) is located in the western part of Eyjafjördur in North-Iceland. A geothermal project started in Árskógsströnd in 1996 aimed at developing a heating system from geothermal resources between Árskógssandur and Hauganes. Geothermal manifestations in the area were limited to Hálshöfdi, a warm spring with a temperature of 35.5°C, but a high temperature gradient was also found in a temperature gradient survey in well ARS-3 at Ytri-Vík. As both Hálshöfdi and Ytri-Vík are at a considerable distance from the villages at Árskógssandur and Hauganes, a decision was made to continue the temperature gradient survey to look for a high temperature gradient closer to the villages. For this purpose Orkustofnun advised on the drilling of many shallow drill holes to test the temperature gradient and try to locate a zone of high heat flow in the area as close as possible to both villages. By 1996, 16 shallow wells had been drilled (Flóvenz and Smárason, 1997). In this report an interpretation of the distribution of earthquake epicenters is compared to lineaments detected from aerial photos, to see if they connect to geothermal manifestations. An isotemperature map, based on the shallow wells in the area, suggested a closer investigation of an area around well ARS-15 which showed a high temperature gradient anomaly of 200°C/km. As a part of the Geothermal Training Programme final project in 1996, Konstantinos Velegrinos (1996) prepared a report based on information from 12 wells and found the highest temperature gradient in the vicinity of well ARS-6, showing a temperature gradient anomaly of 145°C/km.

This year the project continued by making a magnetic map of the area surrounding wells ARS-6 and ARS-15 and then by drilling wells ARS-17 to ARS-29. This report is based on the interpretation of

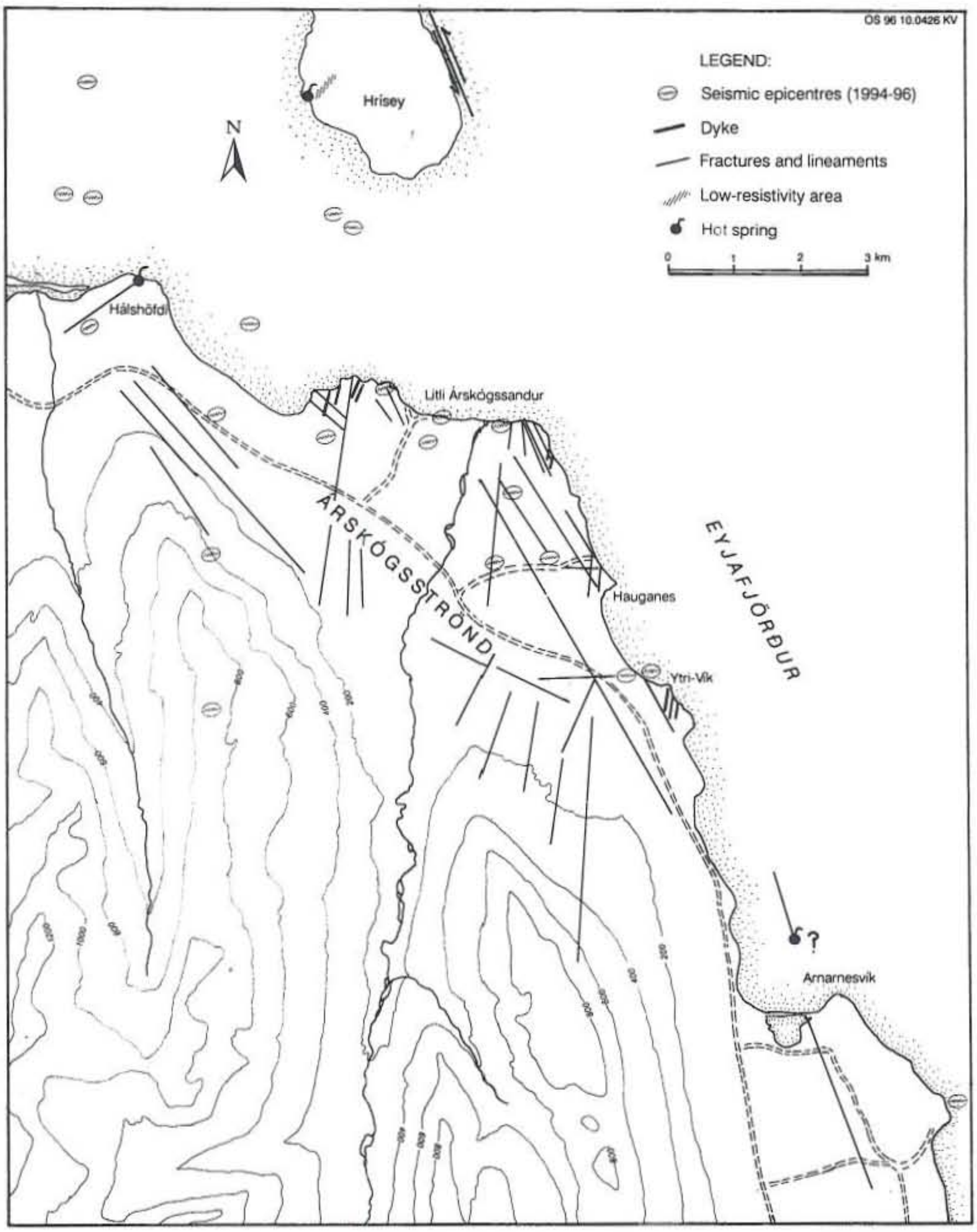


FIGURE 7: The earthquake epicenters and lineaments found on aerial photos (Velegrinos, 1996)

temperature data from wells ARS-13 - ARS-29 in making an isotemperature map, comparing that to the information given by the magnetic measurements to give new data on hydrothermal activity in the research area, and finally to suggest the location of a deep exploratory drillhole to test the temperature gradient to a greater depth.

4.1 Geological setting

Árskógsströnd is located upon Tertiary flood basalts of an age in the range 6-10 m.y. The composition of lava piles in the area is mainly tholeiitic basalt with a thin layer of interbedded scoria and sediments. The lava thickness varies from a few metres to about 30 m, intersected by vertical dykes and faults. A large number of dykes of variable thickness and type occur in the area trending SSW-NNE. The thickness of the dykes varies from 0.5 to 5 m with mostly basic composition and they are more or less vertical (Stebbing, 1963).

The main structural event is an anticline with an axis trending NNE-SSW in the western part of the area. It goes through Hálshöfdi and divides the area into two parts, west and east of the anticline. In the eastern part of the research area, the strata dip is 6-8° to the southwest (Saemundsson, 1979) but it is difficult to measure accurately.

The geology in the area is affected by lateral strike slip movement faults, connecting the northeast volcanic zone to the Kolbeinsey ridge north of Iceland. The trend of these transform faults is NW-SE. In the northern part of Árskógsströnd an active seismic zone crosses the area, the Dalvík lineament. It is a transform fault system, which is a part of the fault zone connecting the northeastern volcanic zone to the Kolbeinsey ridge. The lineament is seismically active but no movements have ever been observed. The only strike slip fault observed in the area is north-northwest striking a left lateral fault along the east coast of the Hrísey island in Eyjafjördur (Fridleifsson, 1989).

4.2 Previous studies

Geothermal research in the Árskógsströnd area has been based on seismic data and epicenters of earthquakes, defining probable lineament on aerial photos, analyses of features and fractures in field inspection, and the drilling of a set of shallow wells to find the location of a possible up-flow zone. Figure 7 shows the location of main earthquake epicenters in the period 1994-1996 and the lineament of aerial photos produced by Velegrinos (1996). The lineaments on aerial photos can be put into three groups, 350-10°, 310-330°, and 270-290°. Field measurements show the direction of fracture systems as 330-340°, 270-280° and 10°. The question was if any lineaments with high temperature gradient would connect to geothermal manifestations, i.e. between the hot springs in Hrísey island (surface temperature 51°C), and a shallow borehole in Ytri-Vík. This would support the existence of geothermal system and probable geothermal manifestations in Arnarnesvík. The geothermal gradient wells were sited to answer this. Figure 8 shows the temperature gradient based on data from the 16 shallow wells.

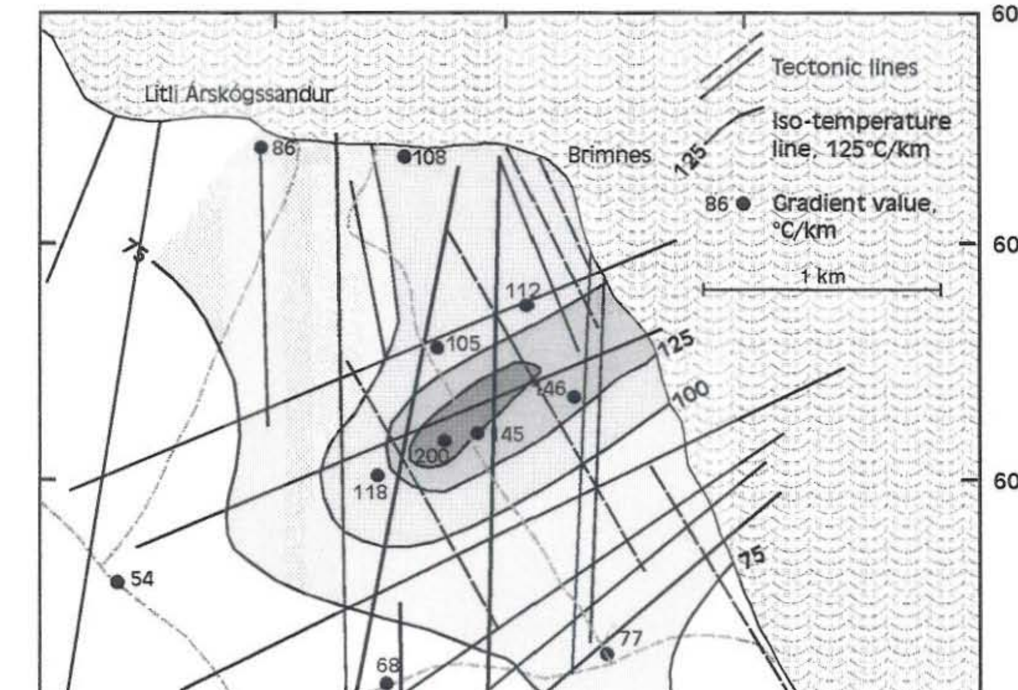


FIGURE 8: The distribution of temperature gradients close to wells ARS-6 and ARS-15 in Árskógsströnd (Flóvenz and Smárason, 1997)

The results show two anomalies in the Árskógsströnd region. One around the Ytri-Vík area shows an anomaly around 317°C/km, and another close to the wells ARS-6, 145°C/km and well ARS-15, 200 °C/km. The latter is the area suggested for further exploration by drilling more boreholes because of the short distance to the Árskógssandur and Hauganes villages. Note that on the isogradient map we can find the temperature gradient anomaly around well ARS-6 and ARS-15. It seems to be elongated in a northerly direction and coincides with a lineament observed from the aerial photos.

4.3 Magnetic survey

Based on previous studies, it was decided to make a magnetic map of the area of interest surrounding drillholes ARS-6 and ARS-15. The magnetic measurements were carried out in May 1997, using a proton magnetometer, measuring in 5 m intervals along lines located 20 m apart. The measured area covered approximately 0.6 km^2.

In the low-temperature areas in Iceland, where dykes and faults are often traced, it is common to do a magnetic survey with 20 m distance between profile lines and 5 m between measuring points on each line. The magnetic sensor is on a stick 2-4 m above the surface, or in a backpack, if only some 5-10 nT accuracy is required (Hersir and Björnsson, 1991).

After plotting the data and producing the profiles of magnetic anomalies, the data is transported onto a magnetic contour map (Figure 9), as the

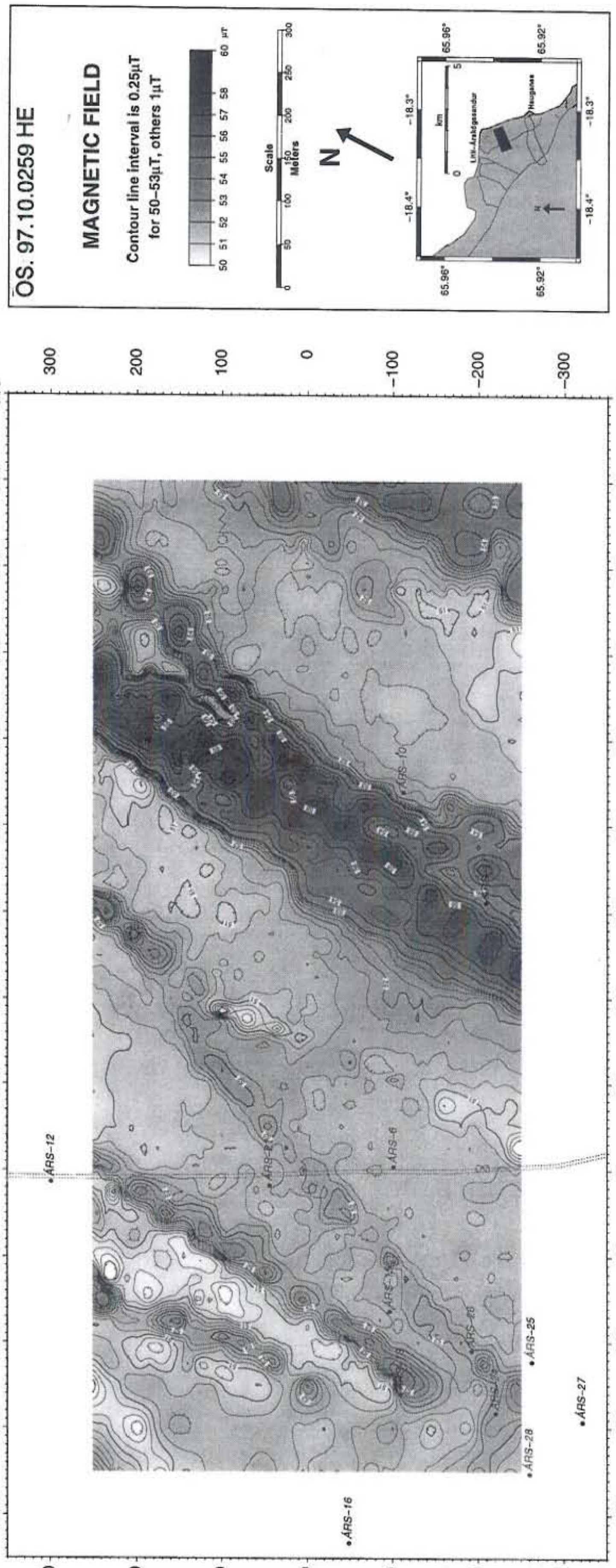


FIGURE 9: The magnetic contour map for the Ytri-Vík area, Árskógsströnd (in nT)

trend of probable lineaments on a magnetic contour map can show the direction of dykes and faults. Finally we can produce the map of dyke lineaments in the area. Figure 10 shows the direction of dykes derived from the magnetic contour map. The map also shows the location of the shallow drill holes in the area.

The main direction of dykes is SSW-NNE and is almost parallel with the Dalvík lineament. The density of dykes in different parts of the map is not equal and we can more or less define three swarms of parallel dykes from west to east. Drillholes ARS-6, 15, 16, 18, 19, 21, 24, 25, 26, 27, 28 and 29 are within the westernmost dyke swarm and wells ARS-10 and 17 are situated in the middle one. Probably each of the dyke sets is injected into a fracture zone and the density of dykes corresponds to a higher permeability. The trend of the magnetic anomalies is generally north to northeast. The highest temperature gradient (200 °C/km) from the previous studies was found in well ARS-15. The magnetic map shows that ARS-15 is situated within the westernmost dyke swarm derived from the magnetic measurements. Therefore the area around drillhole ARS-15 was chosen for further exploration by drilling more exploratory wells in order to delineate a possible thermal gradient anomaly of ≥200 °C/km.

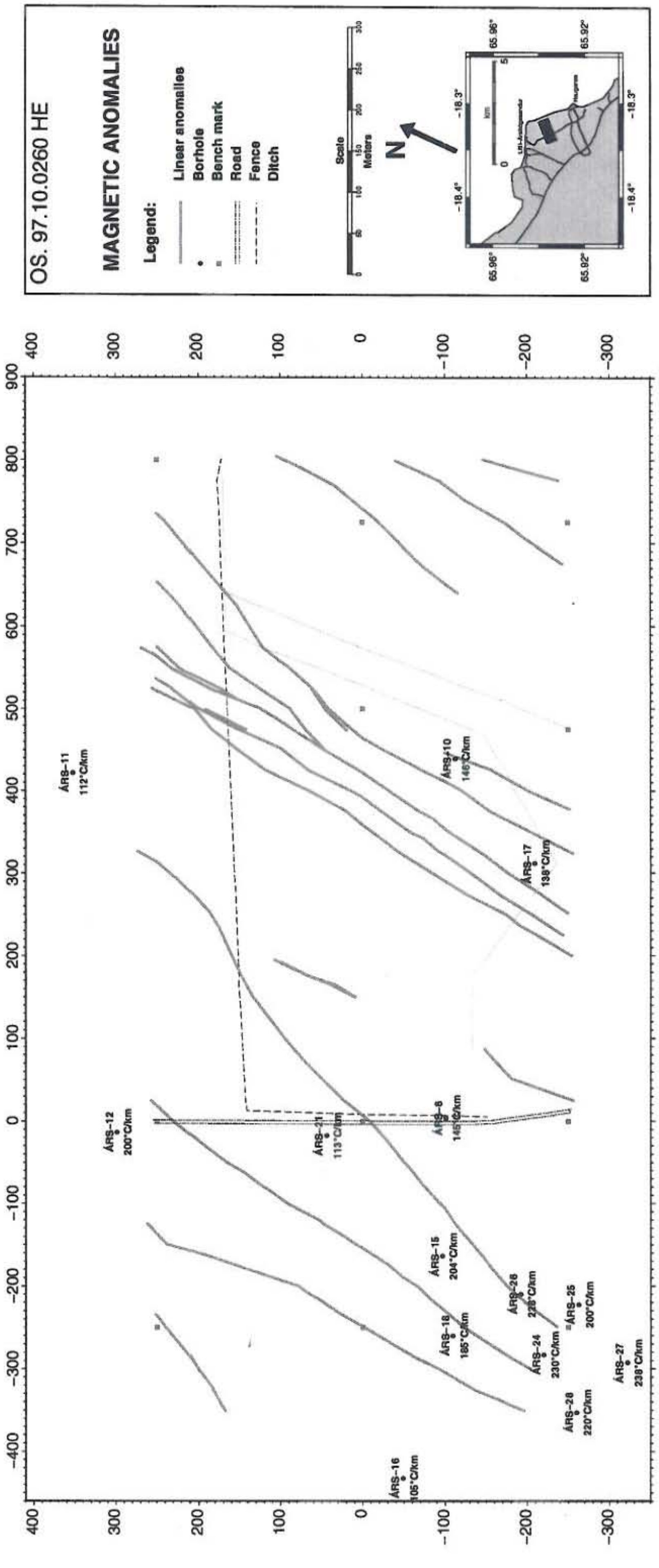


FIGURE 10: Direction of dykes in the Ytri Vík area, Árskógsströnd, extracted from the magnetic map, and location of shallow drillholes

4.4 Thermal survey and temperature measurements

4.4.1 Principle of method

Thermal survey or temperature gradient measurement is a direct geophysical method for measuring temperature in shallow drill holes of 20-200 m depth. The method has a good correspondence with the geothermal systems. The holes are often drilled with air-hammer drilling in 1-2 days and are, therefore, relatively low in cost. The main objective of a temperature well is to obtain the subsurface temperature. Usually the wells need several weeks to obtain a constant thermal situation and equilibrium. Measurements in temperature gradient wells are made with a resistivity thermometer every 2-5 m along the well with an accuracy of about 0.01°C or only 0.1°C where high anomalies are present, like in geothermal areas. The Tertiary low-temperature areas in Iceland are fracture-dominated systems which derive their heat from the hot crust by active and localized convection in near vertical fractures (Bödvarsson, 1982; Björnsson et al., 1990). Away from these fracture zones, the bedrock is quite impermeable and heat transfer is dominated by conduction. It is suggested that at a shallow depth, the temperature is much higher within the geothermal field than outside but at greater depth the situation is reversed. That is because of the heat transferred from the deeper part of the systems to the shallower part (Flóvenz and Saemundsson, 1993).

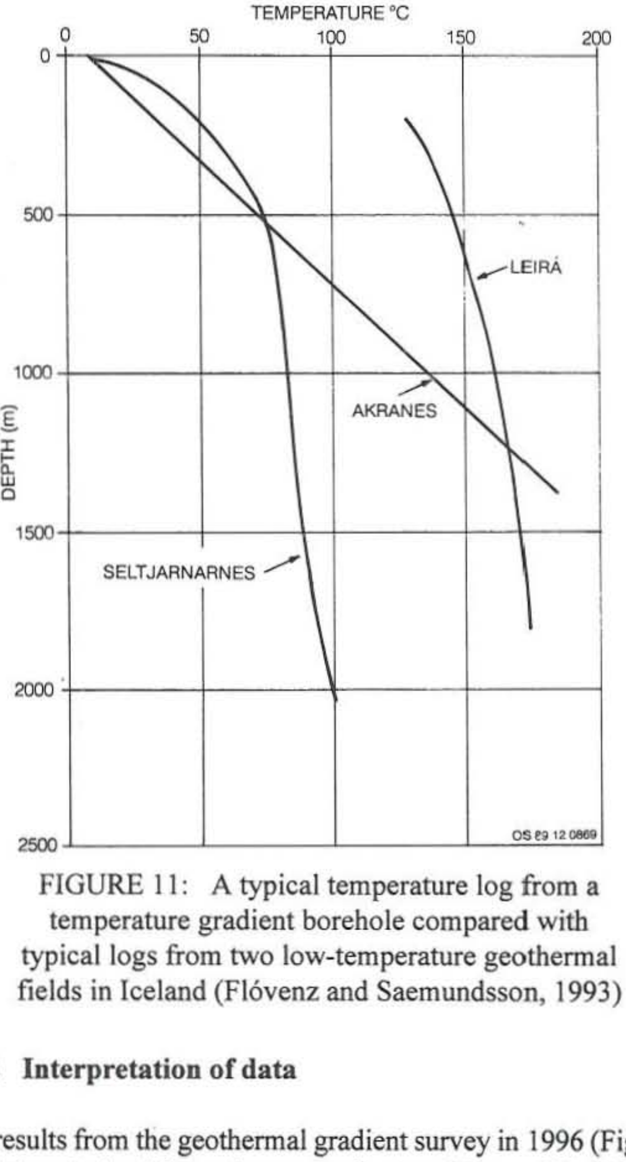


FIGURE 11: A typical temperature log from a temperature gradient borehole compared with typical logs from two low-temperature geothermal fields in Iceland (Flóvenz and Saemundsson, 1993)

Figure 11 shows typical temperature logs from boreholes in two low-temperature fields compared with the temperature profile from a borehole outside a geothermal area (Flóvenz and Saemundsson, 1993).

4.4.2 Interpretation of data

The results from the geothermal gradient survey in 1996 (Figure 8) have been drawn on a map along with the dykes derived from the magnetic map (Figure 10). Also included on the map, are the geothermal gradient studies from wells ARS-17 to ARS-29. The results are shown in Figure 12. Temperature logs were made just as the drilling of each well was finished and repeated to see if thermal equilibrium had been reached. Temperature gradient estimates can be considered quite reliable from all wells except for well ARS-29, which was still being drilled when this was written. The temperature gradient for ARS-29 must, therefore, be considered a minimum estimate as the temperature logs are affected by the cooling water. Table 3 shows the temperature gradient from wells ARS-17 to ARS-29 at Árskógsströnd. The temperature logs from the wells are shown in Appendix III.

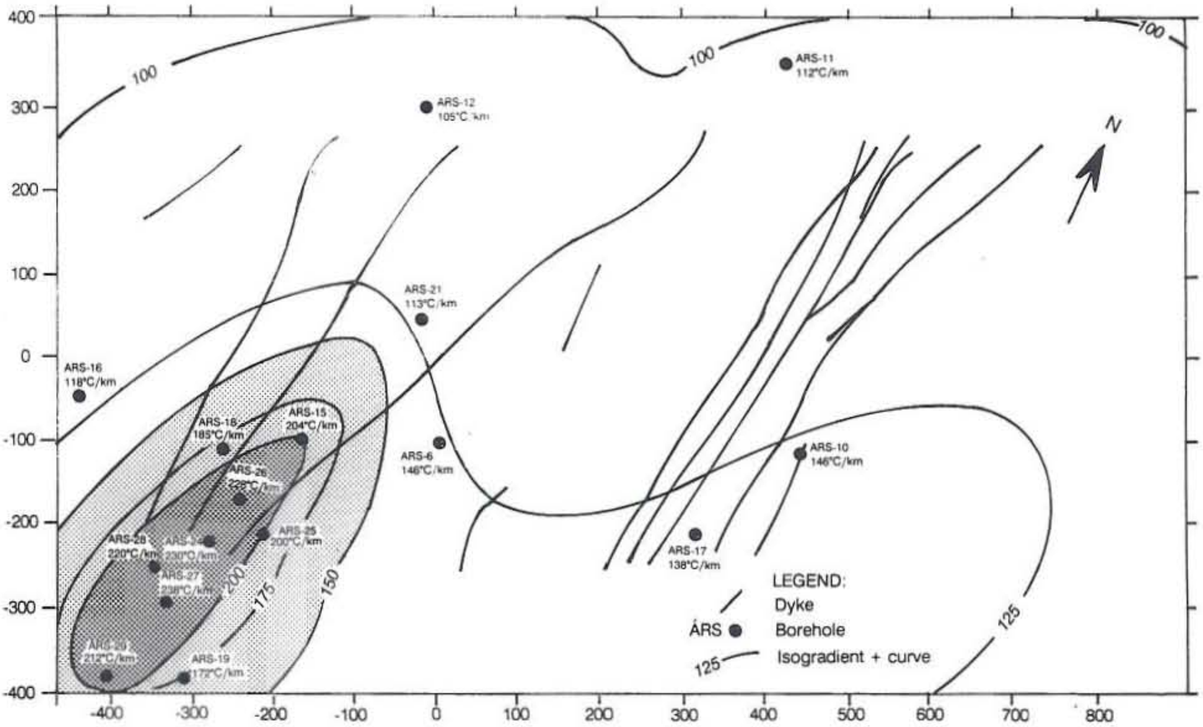


FIGURE 12: Temperature contour map of Árskógsströnd area with data from 28 shallow wells

The temperature gradient map in Figure 12 shows a distinct anomaly. This geothermal gradient anomaly exceeds 200°C/km, and that seems to be the most promising point of attention. The isogradient curve interval is 25°C. The temperature gradient in ARS-15 was 200°C/km. Wells ARS-26, 24, 27 and 19 are on a line to the south. The four wells ARS-24, 26, 27 and 28, show the highest temperature gradient, of more than 230°C/km. It is very interesting to note that the water table in all the wells is at 5-10 m depth except for the wells limited to the fracture zone, where it is at 30 m depth. In this part, the density of dykes is very high and probably in this fracture zone the injection of several dykes make it a high-permeability zone. The elongation of the geothermal anomaly is in a north-northeasterly direction; it is parallel to the direction of the dykes. The highest temperature gradient in the area seem to be between ARS-24 and ARS-27, hence it is the most promising area.

TABLE 3: Measured temperature gradient in Árskógstrrönd area

Well number	Temperature gradient (°C/km)	Well number	Temperature gradient (°C/km)
ARS-17	138	ARS-24	230
ARS-18	185	ARS-25	200
ARS-19	172	ARS-26	228
ARS-20	88	ARS-27	238
ARS-21	113	ARS-28	220
ARS-22	102	ARS-29	≥212
ARS-23	108		

4.4.3 Summary and conclusions

The location of the high temperature gradient anomaly is in an area where the density of dykes is clearly high and there probably is an important fracture zone. The direction of the isogradient curves is almost parallel to the direction of the dykes as seen from the magnetic map. The temperature gradient

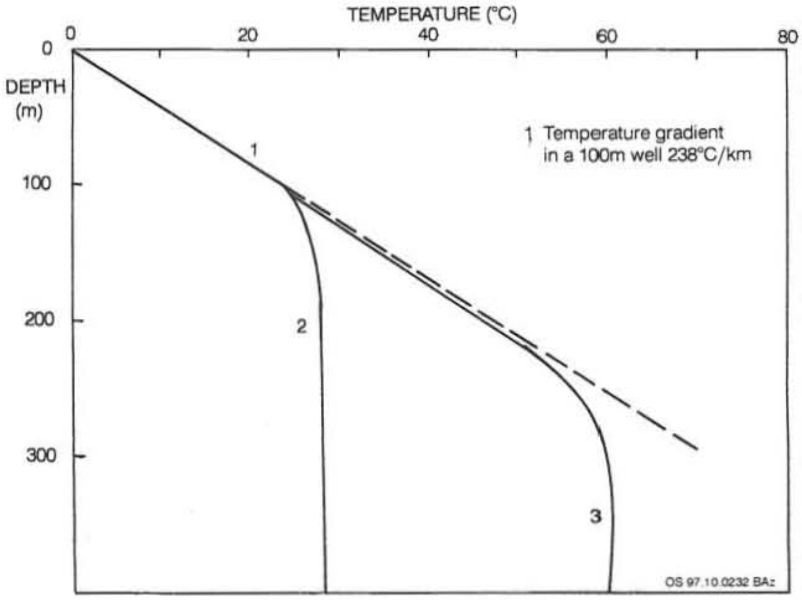


FIGURE 13: Probable temperature curves from a deep exploration well close to ARS-27, Árskógsströnd, depending on temperature in geothermal system

measurements show the highest gradient in well ARS-27 (240°C/km), which seems to be the best area to locate a deep drill hole. In a deep drillhole, it is hoped to get water of 60-80°C, which would be very good for a central heating system for Árskógssandur and Hauganes. But it is common in the low- temperature areas in Iceland for the temperature gradient as determined by the shallow wells, not to continue with depth. In the research area the water table is at a depth of 5-6 m, but in the wells within the fracture zone close to the high temperature anomaly, it is around 30 m below surface. This could indicate a high permeability within the fracture zone.

In conclusion, a deep exploration well should be sited between ARS-24 and ARS-27. The well should be drilled to at least 300 m depth to see if a continuous temperature gradient of 240°C/km will be observed. The situation could, however, be different as indicated in Figure 13. In a deep exploration well (Figure 13) we can expect the following causes:

1. The temperature gradient to continue, as seen in the shallow wells, which would be the best result. In that case, we could expect hot water of 80°C at 500 m depth.
2. The temperature gradient to be reduced much at about 100 m depth.
3. The temperature gradient to be somewhere between these two cases. It is a likely situation, but will hopefully give 60-80°C water at 500 m depth if the fracture zone acts like an up-flow zone.

The best procedure is to drill an exploratory well to 250-300 m depth without casing. If the temperature gradient seems to be continuous, the well should be cased, the drilling continued and the well, if successful, can be used as a production well later.

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At last my appreciation for the clear example of my late father and thanks to my kind mother and my family for giving me the opportunity for education and for helping me throughout my life.

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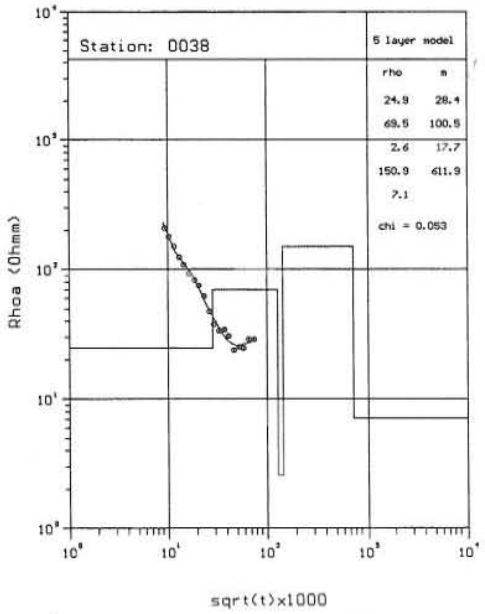
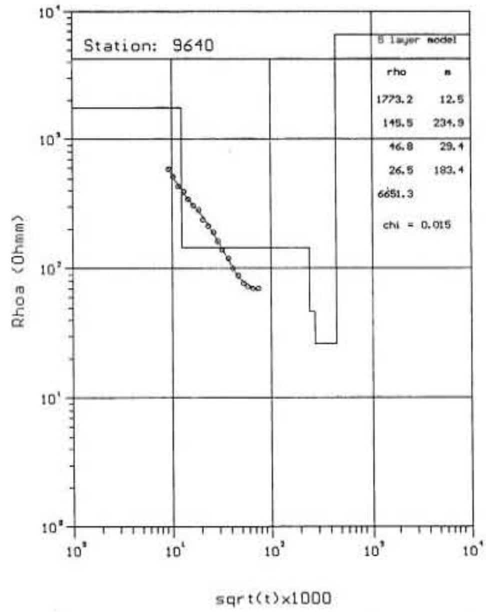
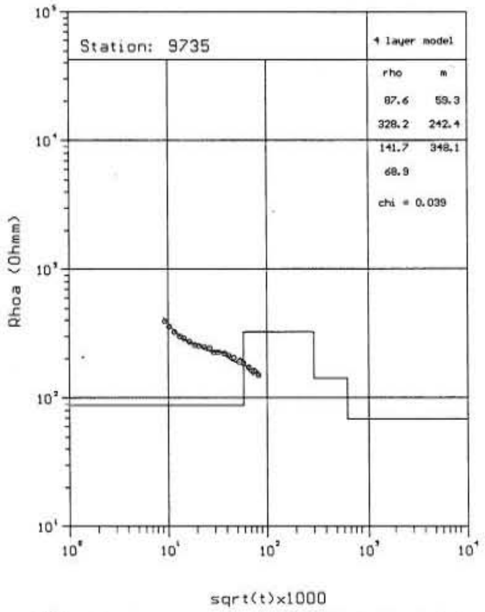
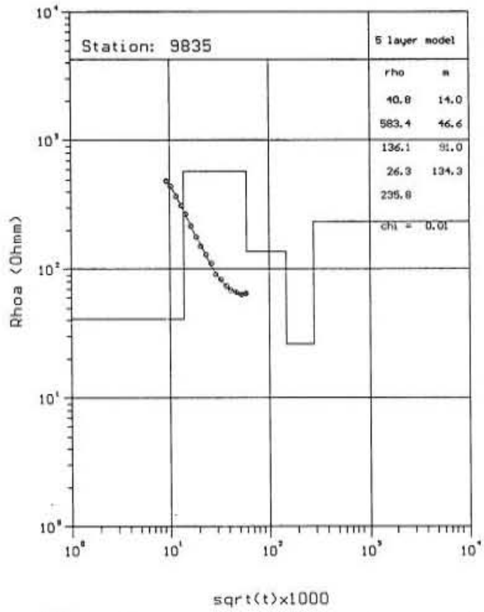
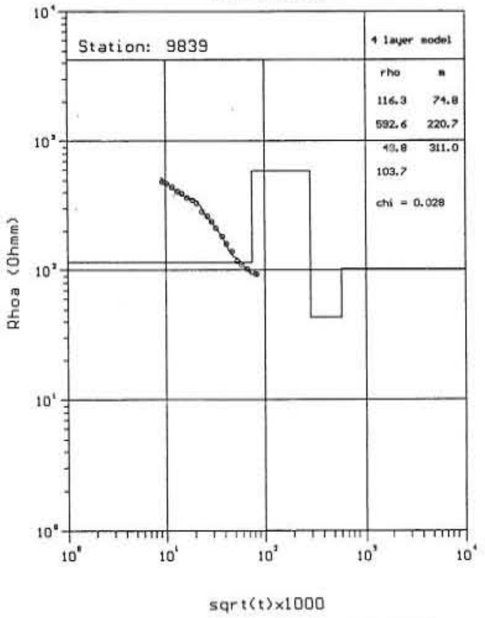
APPENDIX I: TEM soundings in the Bolungarvík area, data, interpreted curves and models

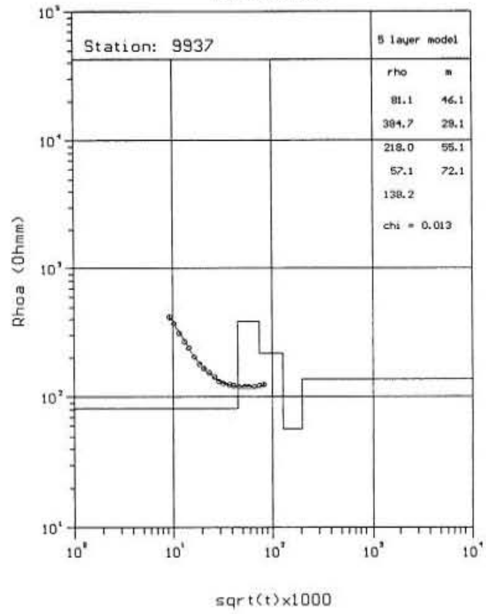




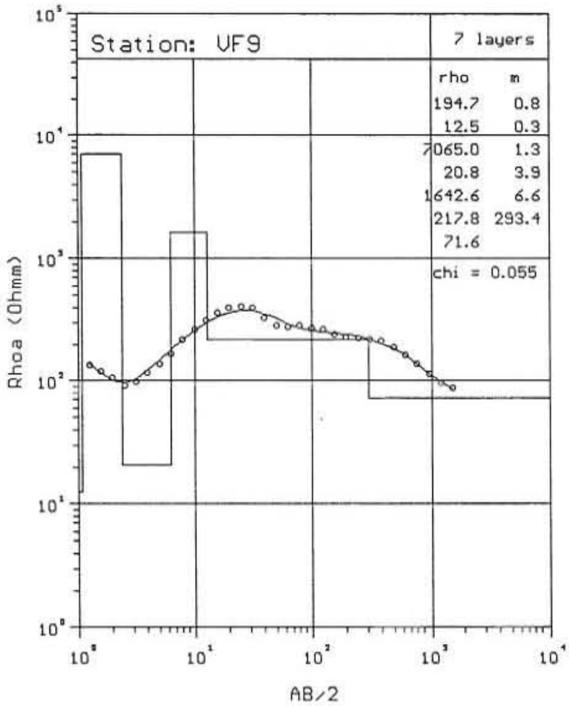


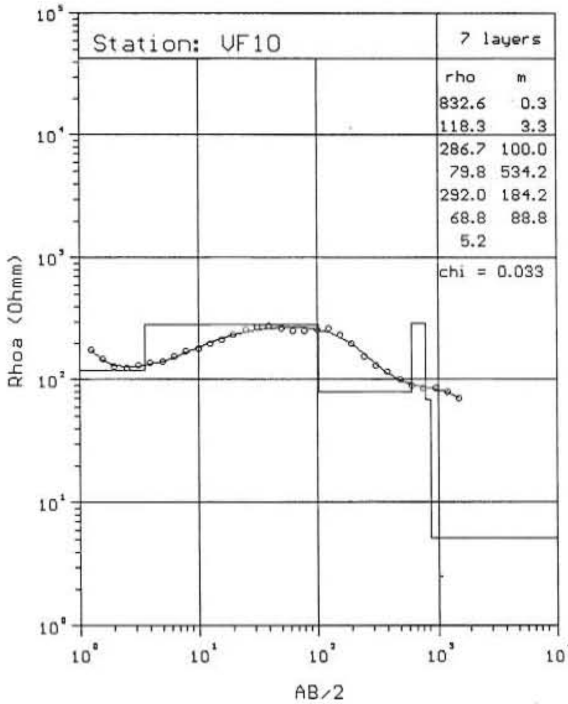




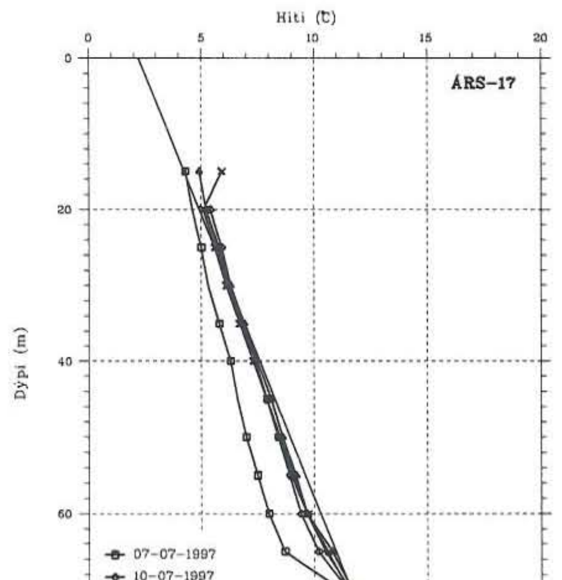


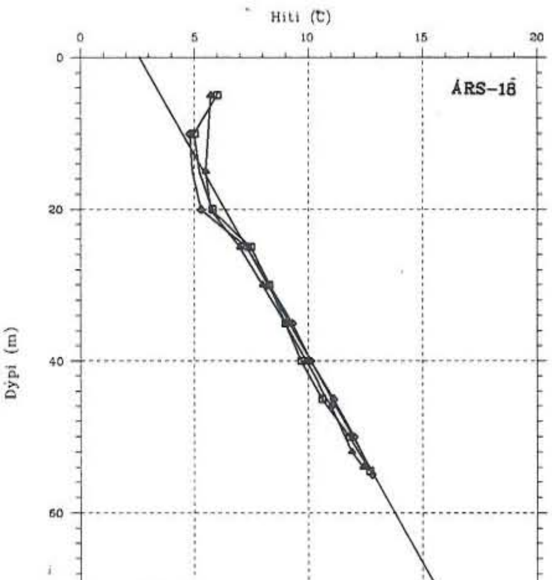
APPENDIX II: Schlumberger soundings in the Bolungarvík area, data, interpreted curves and models (for VF8 see Figure 5A)

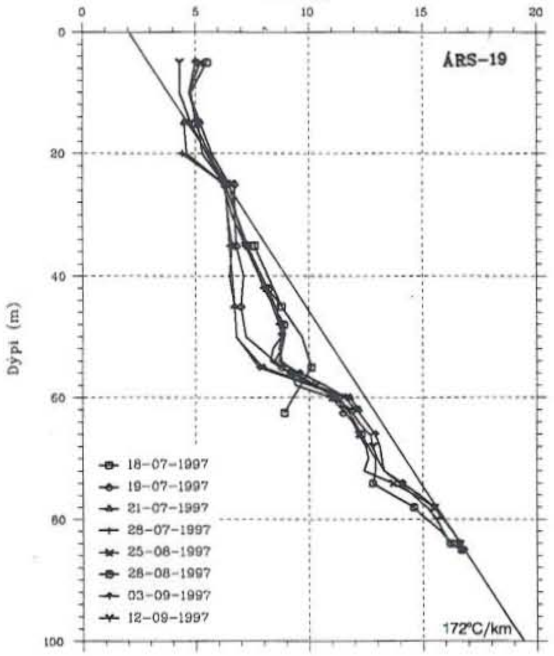


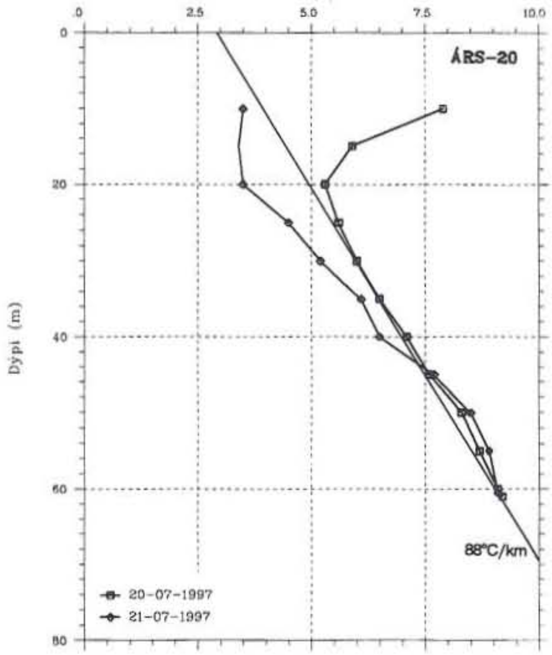


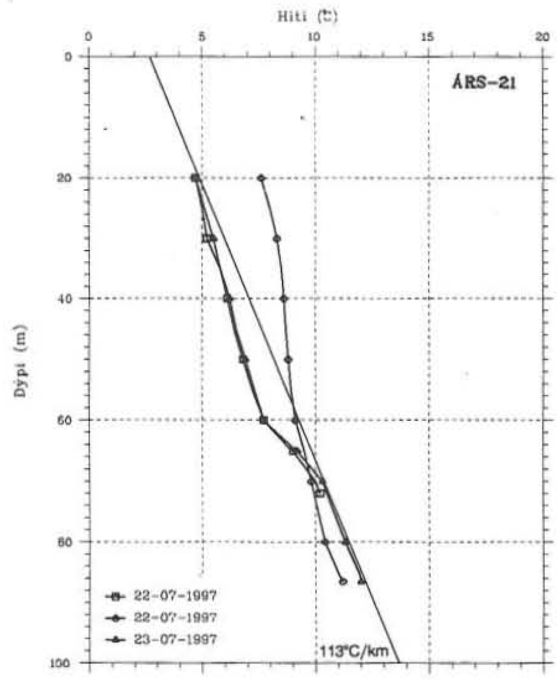
APPENDIX III: Temperature gradient measurements from shallow wells at Árskógsströnd

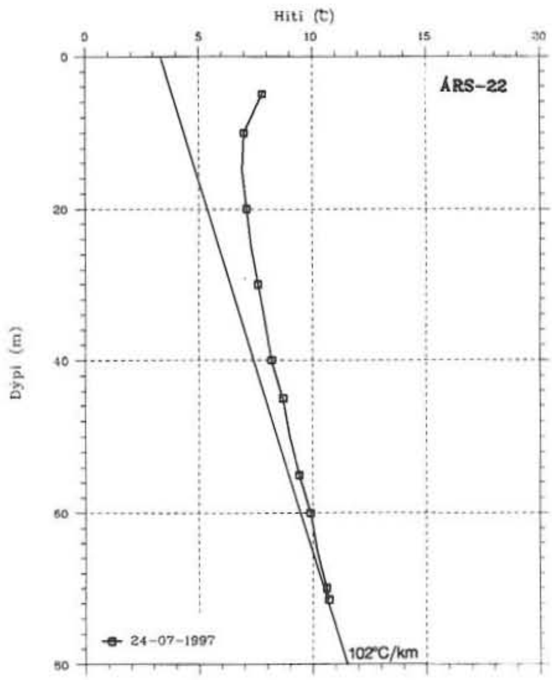


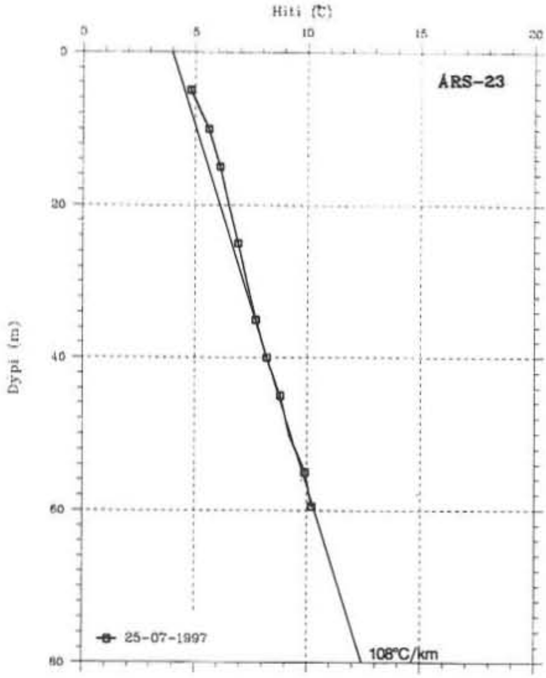


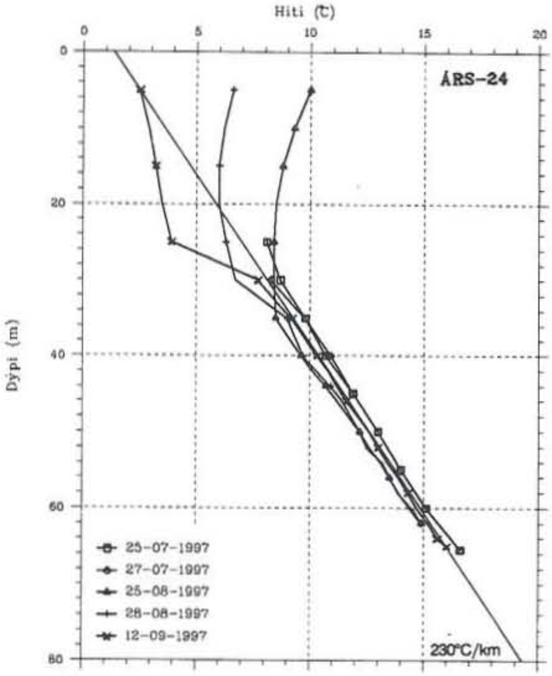

Hiti (°C)
Dýpi (m)
ÁRS-19
18-07-1997
19-07-1997
21-07-1997
28-07-1997
25-08-1997
28-08-1997
03-09-1997
12-09-1997
172°C/km

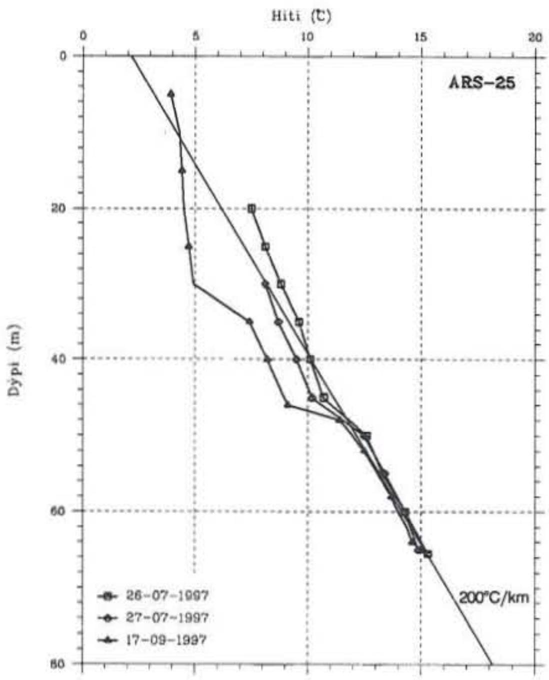

Hiti (°C)
Dýpi (m)
ÁRS-20
88°C/km
20-07-1997
21-07-1997

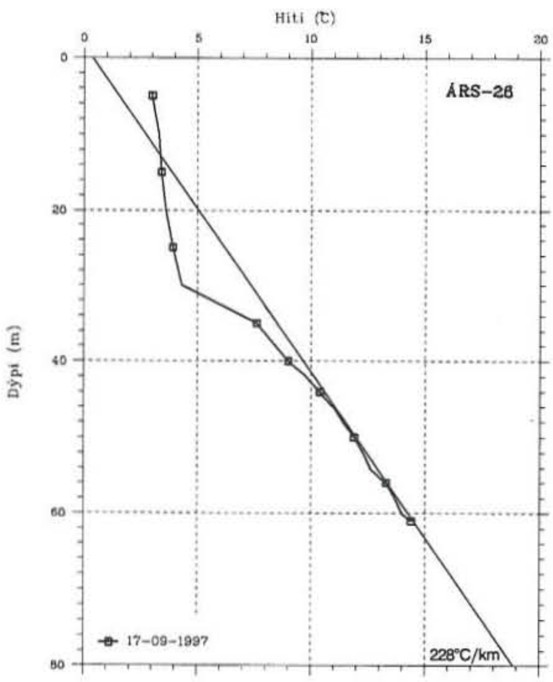

Hiti (°C)
Dýpi (m)
ÁRS-21
22-07-1997
22-07-1997
23-07-1997
113°C/km

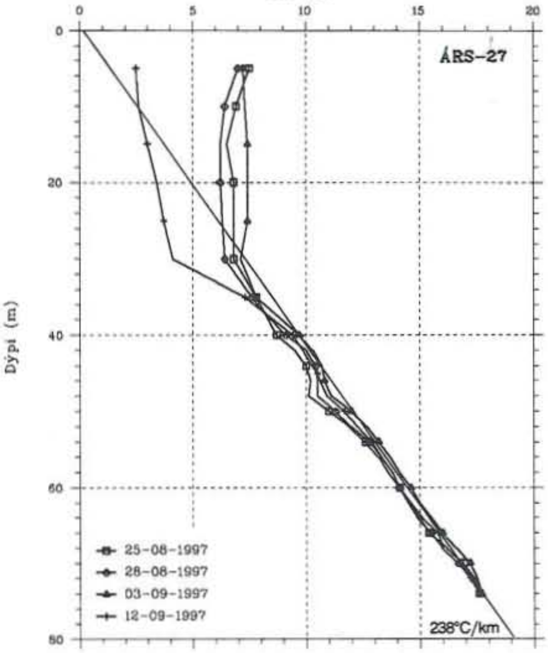

Hiti (°C)
Dýpi (m)
ÁRS-22
24-07-1997
102°C/km

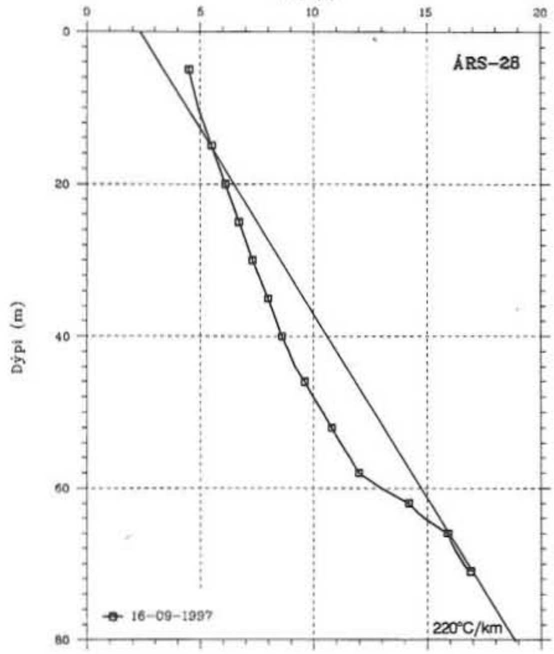

Hiti (°C)
ÁRS-23
Dýpi (m)
25-07-1997
108°C/km

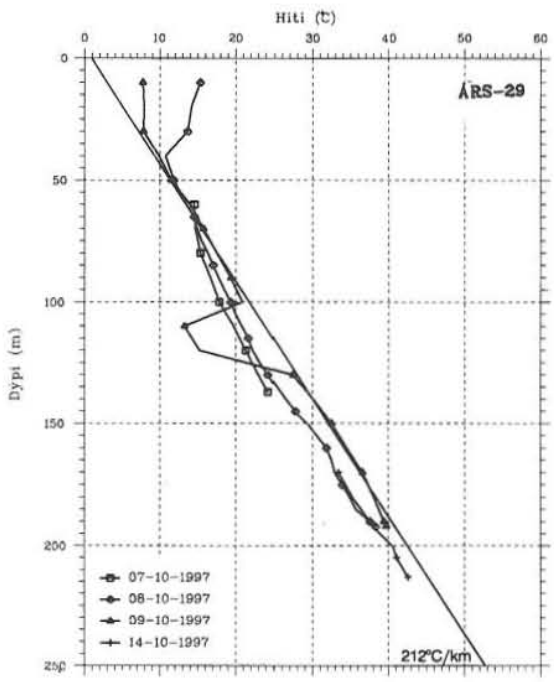

Hiti (°C)
ÁRS-24
Dýpi (m)
25-07-1997
27-07-1997
25-08-1997
28-08-1997
12-09-1997
230°C/km


Hiti (°C)
ÁRS-25
Dýpi (m)
26-07-1997
27-07-1997
17-09-1997
200°C/km


Hiti (°C)
ÁRS-26
Dýpi (m)
17-09-1997
228°C/km


Hiti (℃)
Dýpi (m)
ÁRS-27
25-08-1997
28-08-1997
03-09-1997
12-09-1997
238°C/km


Hiti (℃)
Dýpi (m)
ÁRS-28
16-09-1997
220°C/km


Hiti (℃)
Dýpi (m)
ÁRS-29
07-10-1997
08-10-1997
09-10-1997
14-10-1997
212°C/km