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GEOPHYSICAL EXPLORATION OF THE HELGAVATN LOW-TEMPERATURE FIELD, W-ICELAND AND THE ÁRSKÓGSSTRÖND AREA N-ICELAND

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ABSTRACT

Helgavatn, W-Iceland, and Árskógsströnd, N-Iceland, are two fracture dominated low-temperature geothermal fields in Iceland. Geophysical exploration there was aimed at locating the upflow zones of the geothermal water. At Helgavatn head-on resistivity profiling was chosen to try to detect permeable near-vertical fractures, which are seen as low-resistivity structures. The results of computer modelling of the three Helgavatn head-on profiles show good correlation with known faults and dykes and indicate that the main hot springs at Helgavatn are located at the intersection between a WNW-ESE trending fault and a NE-SW trending dyke. At Árskógsströnd a study of the general tectonics of the area was carried out, including a.o.t. study of air photos, an analysis of microearthquakes in the area and on-site inspection. Accompanying this was the drilling of 12 shallow gradient wells to locate areas of anomalously high temperature gradient where good fracture permeability might also be expected. Two interesting areas were located, one in Ytri-Vík in the southern part, and the other at well site ARS-6, which is located between the two small villages, Litli-Árskógssandur and Hauganes. Further gradient wells around ARS-6, followed by a deeper exploration well, are recommended in order to find hot water for the villages. Some low-temperature fields in Greece resemble the Icelandic fields dealt with here. The geophysical methods described could be useful in the exploration of these.

1. INTRODUCTION

Geothermal research in Iceland consists of geological studies and geophysical exploration supported by information on the properties of thermal fluid. Other available and relevant information such as analyses of microearthquakes in the prospect area can also be taken into account. The most widely used method in geophysical exploration is measuring the electrical resistivity of the earth. DC-soundings using the Schlumberger configuration and the TEM (transient electromagnetic) soundings are used to delineate geothermal systems in large areas. For detail studies head-on profiling is used for detection of near-surface near-vertical discontinuities, interpreted as upflow zones of geothermal water. Drilling of shallow explorations wells for temperature gradient studies has also been widely used to locate upflow zones.

At the Helgavatn low-temperature field W-Iceland, head-on profiling was used for the detection of vertical discontinuities of geothermal interest. The modelling of the measurements was carried out with computer programs developed at Orkustofnun. Based on a proposed starting model, the programs produce by inversion a best fitting model for the measured data lines.

At Árskógsströnd, N-Iceland, a geophysical exploration was carried out combining several methods in order to try to locate a feasible geothermal prospect for drilling. Two small communities instigated this project in order to try to find hot water for space heating. The emphasis was on drilling shallow gradient wells for measurements here, taking into account available geologic information on the dominant fracture systems in this area

Finally, a short discussion on the use of various geophysical methods in the exploration of low-temperature geothermal fields in Greece is given.

2. HEAD-ON PROFILING AT HELGAVATN, W-ICELAND

2.1 Geological background

The Helgavatn area is a part of the Borgarfjörður region in West Iceland. The basement consists of Late Tertiary basaltic lava flows. The axis of the Borganes anticline runs NE-SW. East of the anticline axis the lavas dip 6-10°SE towards the active Reykjanes-Langjökull rift zone. The anticline axis itself dips to the north and disappears underneath the Hredavatn unconformity (Saemundson, 1979). This unconformity is a major gap in the lava sequence (Figure 1). Below it, the flows are considered to be 13 million years old or more and were extruded within the former Snaefellsnes rift zone (Jóhannesson,

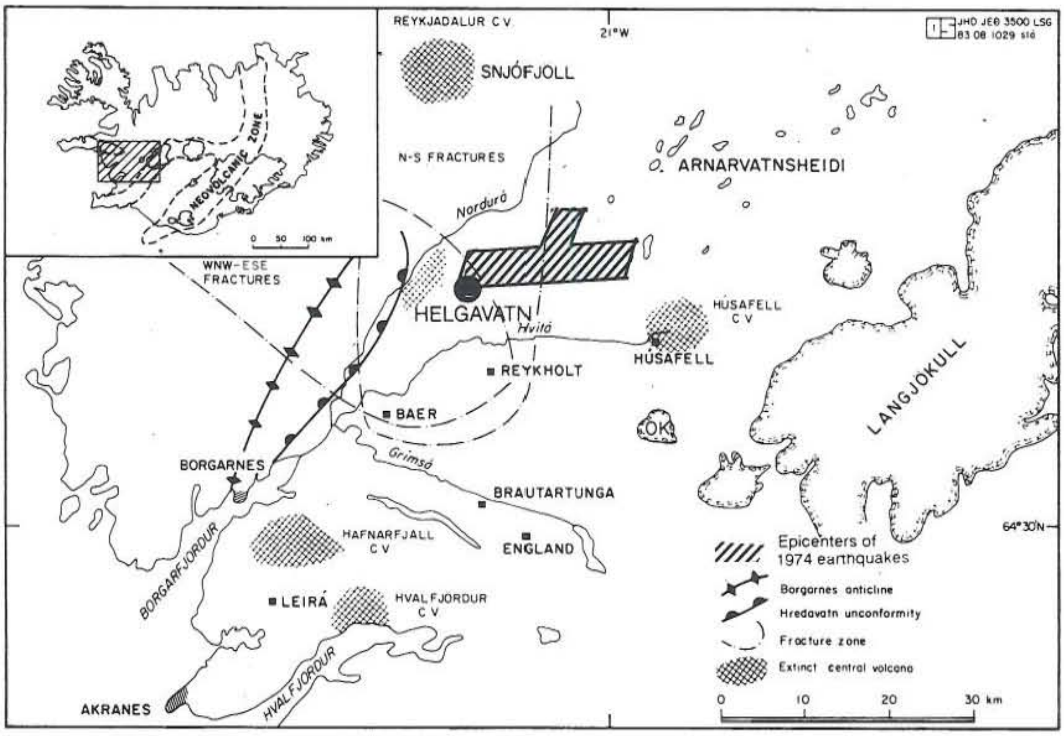

FIGURE 1: The main tectonic features in the Borgarfjordur region

1980). After the tilting and erosion of the lavas they were covered by the Hredavatn sediments which were subsequently covered by younger flows, 6-7 million years old, extruded within the Reykjanes-Langjökull rift zone. The flows become gradually younger to the east on approaching the rift zone.

The great variety of faults and fractures in the Borgarfjördur region are suggested as being of two different origins (Jóhannesson, 1980):

a) The NE-SW trending fault swarms correspond to the fissure swarms of the active rift zone. The fault swarms are accompanied by dyke swarms.
b) Faults which are formed in a stress field characterised by shear forces. These belong to the Snaefellsnes fracture zone which stretches from the Snaefellsnes peninsula in the west to the Borgarfjördur region in the east. The faults of the fracture zone can be divided into 3 groups, based on their trend and age, NW-SE, N-S and NE-SW. The fracture zone was mainly active about 8-13 m.y. ago. However, minor movements have continued up to Postglacial times in association with the current intraplate Snaefellsnes volcanic zone. The dykes and faults are usually vertical and those oriented parallel to the strike of the lavas usually transect the lavas at right angles. The easternmost recent volcanic activity seen at the surface and associated with the Snaefellsnes zone is in Nordurárdalur, a few kilometres west of Helgavatn.

Helgavatn lies in the northern part of the Borgarfjördur region. The fracture system at Helgavatn is composed of two subsystems of different age. The older one has a NE-SW direction and the younger a direction of NW-SE to E-W. As stated earlier, the younger fracture system is connected to the Snaefellsnes fracture zone.

2.2 Geothermal activity

2.2.1 General aspects of the Borgarfjördur thermal region

The Borgafjördur thermal region is the largest low-temperature region in Iceland. The second largest is in South Iceland and these two are adjacent to the Reykjanes-Langjökull axial rift zone bordering its western and eastern margin, respectively. It has been divided into five separate thermal systems mainly based on the results of a regional resistivity survey and the chemistry of the thermal water. One of these is the Reykholt thermal system, the largest, and includes Helgavatn. The Reykholt geothermal system covers an area of 250-300 km^2 in the valleys of the upper Borgafjördur region. The origin of thermal water is in the Arnarvatnsheidi highlands (Figure 1), where precipitation percolates down to a depth of 1-3 km and is heated by the regional heat flow. The thermal water flows laterally for about 50 km to the southwest, driven by the hydrostatic gradient. In the low-temperature areas in the lowlands, the main aquifers appear to be permeable northeasterly faults and occasional dykes. They are intersected by active northwest-trending fractures which enable hot water flow to the surface (Georgsson et al., 1984; 1985).

The Reykholt thermal system comprises the following major thermal fields: Klettur-Runnar, Deildartunga-Kleppjárnseykir, Hurdarbak-Sídumúli, Vellir-Sturlureykir, Reykholt-Kópareykir-Haegindi, Nordureykir-Háafell, Stóriás, Brúareykir, Lundar and Helgavatn. The natural discharge is equivalent to about 400 l/s of boiling water. The highest base temperature is at the Reykholt-Kópareykir thermal field where it exceeds 140°C. The temperature decreases in all directions from the centre.

The Reykholt thermal system can be separated from its three adjacent thermal systems in Borgarfjördur. The Baer system to the west, can be distinguished from it by the slightly different chemistry (higher salinity) and higher Cl/B ratio. The Húsafell system to the east, has a different chemical composition due to flow through acid rock associated with the extinct Húsafell central volcano, and has a different deuterium content. The Brautartunga system to the south, has a similar chemical composition and deuterium content as the Reykholt system but is separated from it geographically.

Arnórsson et al. (1983) showed that major elemental chemistry of Icelandic geothermal waters at equilibrium is fixed by two parameters, temperature and the concentration of chloride. As a result undissociated weak acids (H_2SiO_4, H_2CO_3, H_2S, H_2SO_4, HF) and the cation/proton ratios (Na^+/H^+, K^+/H^+, Ca^{++}/H^+, Mg^{++}/H^+) only depend on temperature. The chloride content of water from the Reykholt thermal system is almost constant. Detailed study of the major elemental composition of the water shows overall equilibrium conditions between water and minerals at the calculated chalcedony equilibrium temperature, assuming 20% steam loss for boiling hot springs.

2.2.2 Helgavatn

The geothermal manifestations and the geology around Helgavatn have been mapped in detail (Figure 2). The highest measured temperature is 74°C in the main hot springs, about 200 m east of the Helgavatn farm. The natural discharge is high, about 30 l/s in the main hot springs alone. Fractures were mapped in all openings in the vicinity of the hot springs, both in the Laugarás hill and along the river Kjarrá. The fracture mapping indicates that a large E-W trending fault with downward faulting to the north controls the upflow close to the surface, possibly together with a northeasterly-trending dyke that is exposed a little to the southwest of the main hot springs. This kind of a fracture system is typical for the geothermal fields in the area (Georgsson et al., 1984; 1985). Further exploration has usually shown that the main geothermal upflow is connected to more than one fracture direction, and that the main fractures intersect

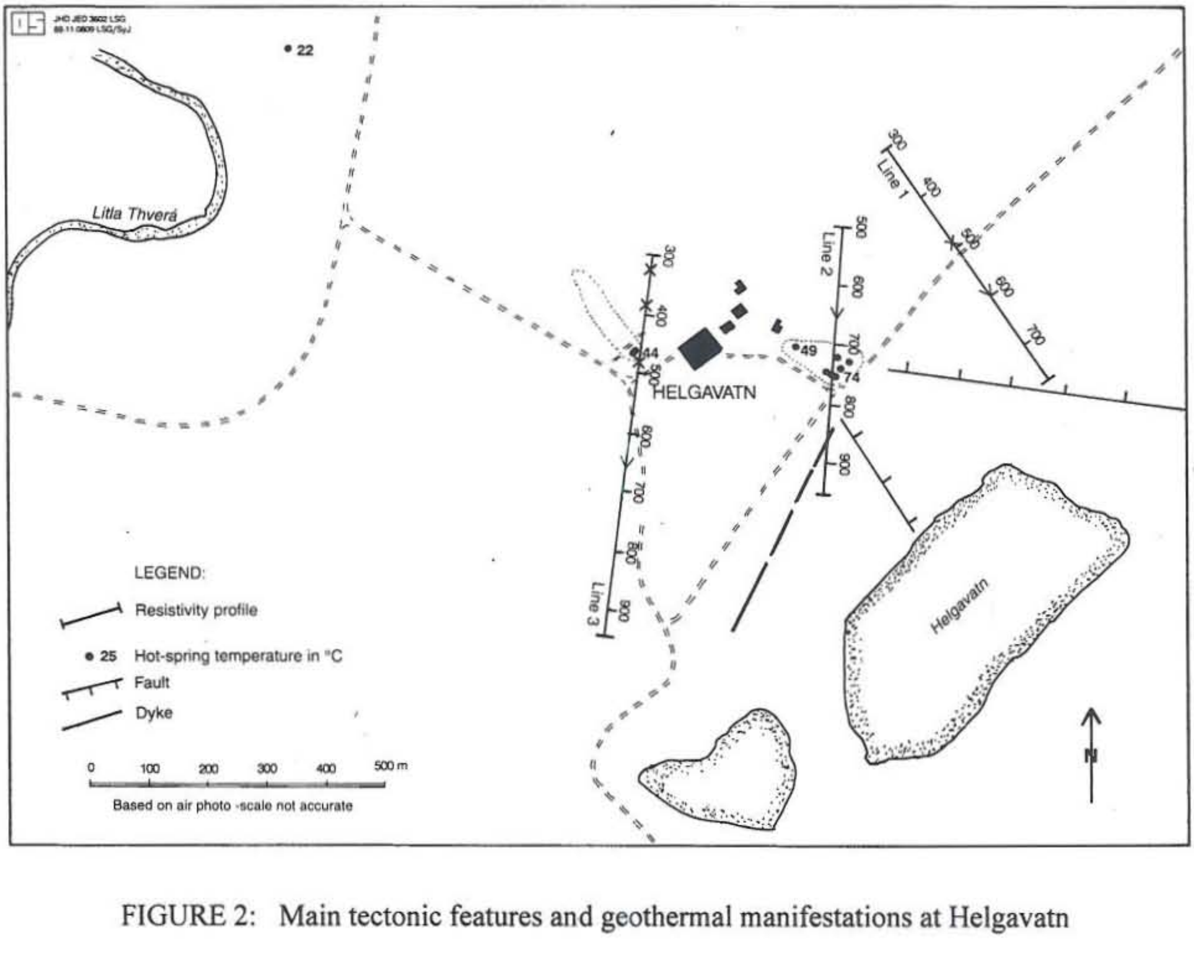

FIGURE 2: Main tectonic features and geothermal manifestations at Helgavatn

at the main hot springs. Also, through experience it is known that the upflow channels are very localized and that a small displacement in locating a borehole can influence the results in a significant way.

For many years the geothermal water has been used for house heating at the Helgavatn farm, but now some of the neighbouring farms are also getting hot water for heating their homes through two pipelines that were laid in the nineties. One of the pipelines is 3.5 km long and serves three farms south of Helgavatn, the other is 2 km long and serves two farms southeast of Helgavatn.

The close relationship between active faults and the flow of hot water was emphasized in the Borgarfjördur earthquakes in 1974. The active zone of epicentres was 15-20 km north of Reykholtsdalur valley and just west of the Helgavatn farm. It showed a westerly trend, but was intersected by a secondary zone with a northeasterly trend. The latter coincides with the northern part of the main low-resistivity anomaly of the Reykholt thermal system. The Deildartunga fault stretches to the northeast towards the same area. The effect of the earthquakes on thermal activity was most notable at Helgavatn, where the thermal water disappeared for three weeks, reappearing with substantially increased discharge and slightly higher temperature. In numerous other springs the water turned milky, gradually returning to normal in a few weeks time (Georgsson et al., 1984).

The active tectonics play an important role in the Reykholt thermal system. Northeasterly faults serve as the main flow channels for the system and the active fractures open the way for the water to the surface. The seismicity of the area keeps them open, as it counteracts low temperature zeolithization or sealing which otherwise would gradually fill the channels.

2.3 Geophysics

2.3.1 Resistivity methods

In geothermal exploration, the most widely used geophysical methods are resistivity sounding and profiling. In this report the emphasis is on the head-on resistivity profiling method which is considered especially suitable for the detection of near surface, near-vertical discontinuities such as faults or dykes. Generally, resistivity measurements in areas of geothermal potential show lower values than in the surrounding areas. The resistivity of water-saturated rocks depends on the porosity of the rocks, salinity of the pore fluid, temperature and alteration minerals. The above factors often interact in a complicated way, which is not fully understood. Some empirical equations have been proposed, which describe the influence of the different factors. These equations are based on measurements of resistivity in different rock samples under different conditions.

Flóvenz et al. (1985) tried to explore the relationship between the above mentioned prime factors and the resistivity of the rocks in the uppermost kilometre of the Icelandic crust outside volcanic zones. They found that for rocks saturated with water of low salinity, i.e. with resistivity higher than approx. 2-5 Ωm (at room temperature), the resistivity of the rocks is practically independent of the resistivity of the water but depends on the porosity and temperature. Furthermore, the electrical resistivity seemed to be very much connected to the alteration minerals.

More specifically, the following can be stated (Flóvenz et al. 1985; Árnason and Flóvenz, 1992; Árnason, 1993):

a) For Icelandic crust the resistivity is typically 100- 500 Ωm down to a depth of some hundreds of metres. Below that the resistivity generally decreases to 30-50 Ωm. The geothermal fields in the low-temperature areas outside active volcanic zones are characterized by lower resistivity than the surroundings, in some cases as low as 10 Ωm in areas with a regional resistivity on the order of 50 Ωm. In areas of higher regional resistivity, such as 300-500 Ωm, the geothermal systems can show a resistivity of approx. 100 Ωm.

b) The conduction mechanisms for electrical resistivity can be divided into two categories, namely pore fluid conduction (electrolyte conduction) and interface conduction. For the former the conduction is through the fluid volume, and in the other the walls of the fractures lined with conductive alteration minerals are the conducting media. Pore fluid conduction dominates in basaltic lavas with no mineral alteration and where the pore fluid is highly saline (ρ_w less than about 2 Ωm). For these cases, Archie's law describes the bulk resistivity

$$\rho = \rho_w \phi^{-n} \tag{1}$$

where

ϕ = Total porosity;
ρ_w = Pore fluid resistivity which decreases with increasing temperature [Ωm];
n = Empirical constant which is about 2.

Minor alteration of the basalts leads to formation of a thin layer of conductive clay minerals such as smectite at the walls of the fractures and pores in the rock matrix. For typical pore fluid, ρ_w above 5 Ωm (at room temperature), conduction along these interfaces dominates the pore fluid conduction. Generally, in non-saline geothermal systems the interface conduction dominates the pore fluid conduction. In saline geothermal systems this is different. There pore fluid conduction is stronger than interface conductivity. In geothermal systems of intermediate salinity, both conduction mechanisms have to be considered.

In Iceland, resistivity methods along with geological research comprise the first step in a geothermal survey. Resistivity soundings are distributed over a large area where geothermal activity is suspected. If this study indicates an area of anomalously low resistivity, further resistivity soundings are made to delineate the geothermal prospect.

In fracture-dominated low-temperature geothermal systems the geothermal fluid is confined within near-vertical aquifers. The general conceptual model is that the 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 head-on resistivity profiling method has been found to be very useful in precisely locating such vertical aquifers in the uppermost few hundred metres.

2.3.2 Background information on Reykholt thermal system

The bulk resistivity of Quaternary and Tertiary rocks in Iceland is practically indepented of the salinity and the temperature of the pore fluid as long as the resistivity of the fluid is above approx. 5 Ωm (Flóvenz et al., 1985). Therefore, resistivity variations in the uppermost kilometre of the crust primarily reflect variations in the amount of interconnected pores and fractures of the rock. Consequently, a regional resistivity survey can be used to estimate the size of geothermal systems and to map the flow pattern of thermal waters from its origin in the highlands to the geothermal areas in the lowlands. The regional resistivity survey in the Borgarfjördur region was initiated on this basis.

About 100 Schlumberger soundings have been made in the Borgarfjördur region covering the valleys of Borgarfjördur and the nearby highlands. The largest low resistivity anomaly coincides with the

Reykholt thermal system. It includes all major thermal fields in Reykholtsdalur and the neighbourhood except the Stóriás field, covering an area of 250-300 km². The area of most intense geothermal activity is also characterized by the lowest resistivity. The resistivity survey indicates two centres of thermal activity in Reykholtsdalur valley, the Reykholt-Kópareykir thermal field, which is characterized by its high base temperature and the Deildarunga-Kleppjárnsreykir thermal field, which is characterized by its enormous natural discharge. Even though it must be assumed that the main flow of thermal waters from the highlands towards Reykholtsdalur valley is at deeper levels, the NE-SW elongated shape of the anomaly suggests a recharge area in the Arnavatnsheidi region (Georgsson et al., 1984).

2.4 Head-on measurements at Helgavatn

2.4.1 Head-on profiling

The head-on profiling method is an extended version of a half Schlumberger array (Cheng, 1980; Flóvenz and Georgsson 1982 , Flóvenz 1984). Figure 3a shows a schematic drawing of the Schlumberger array, and Figure 3b the head-on array. Three current electrodes are used, A and B as in the convectional Schlumberger array and a third electrode C located perpendicular to the line joining A and B at infinity; or practically more than two times the distance AB away. When current is transmitted through B and C (A and C), B (A) can be considered as a monopole current source. Current is injected through each of these three pairs of electrodes and the potential is measured each time between M and N, located on the line between A and B. Three apparent resistivities are calculated, the dipole apparent resistivity, ρ_{AB} and two monopole apparent resistivities, ρ_{AC} and ρ_{BC}. The whole array, i.e., the current electrodes A and B and the potential electrodes M and N, is moved along a line perpendicular to the suspected aquifer, with 25 m between measurement points. The same survey line is generally traversed with two different spacings between A and B, e.g. 300 and 500 m, the latter for more depth resolution. The data presentation of head-on profiling is the plot of ρ_{AB} values with distance along the profile and the differences of ρ_{AC} - ρ_{AB} and ρ_{BC} - ρ_{AB} (denoted as $\rho_{AC\text{-}AB}$ and $\rho_{BC\text{-}AB}$) .

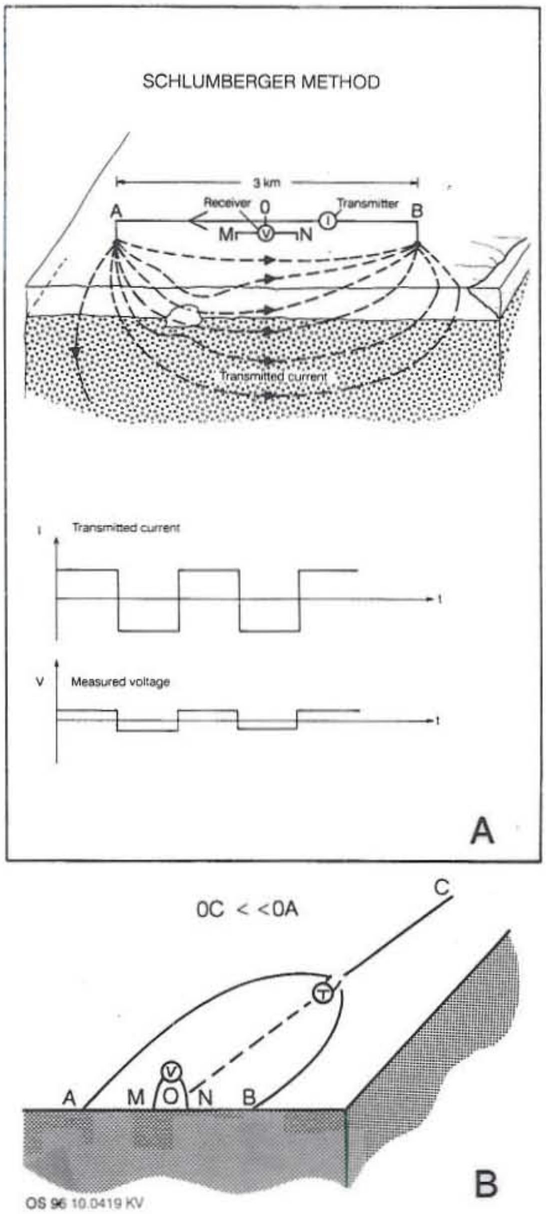

FIGURE 3: a) Schlumberger configuration; b) Head-on configuration

In the case of homogeneous isotropic earth, all three resistivity values are identical. But in the presence of near-vertical resistivity contrasts they change. The electrical current always tends to flow along a conductive fracture. Therefore, the current density will increase, compared to homogeneous earth, in the area between the current source (A or B) and the conductive fracture, but decrease elsewhere. Since the electrical field is proportional to the current density by Ohm´s law and the apparent resistivity is proportion to the electrical field in the vicinity of the potential electrodes MN, an increased apparent resistivity will be measured when MN are between the current source and the conductive fracture, with

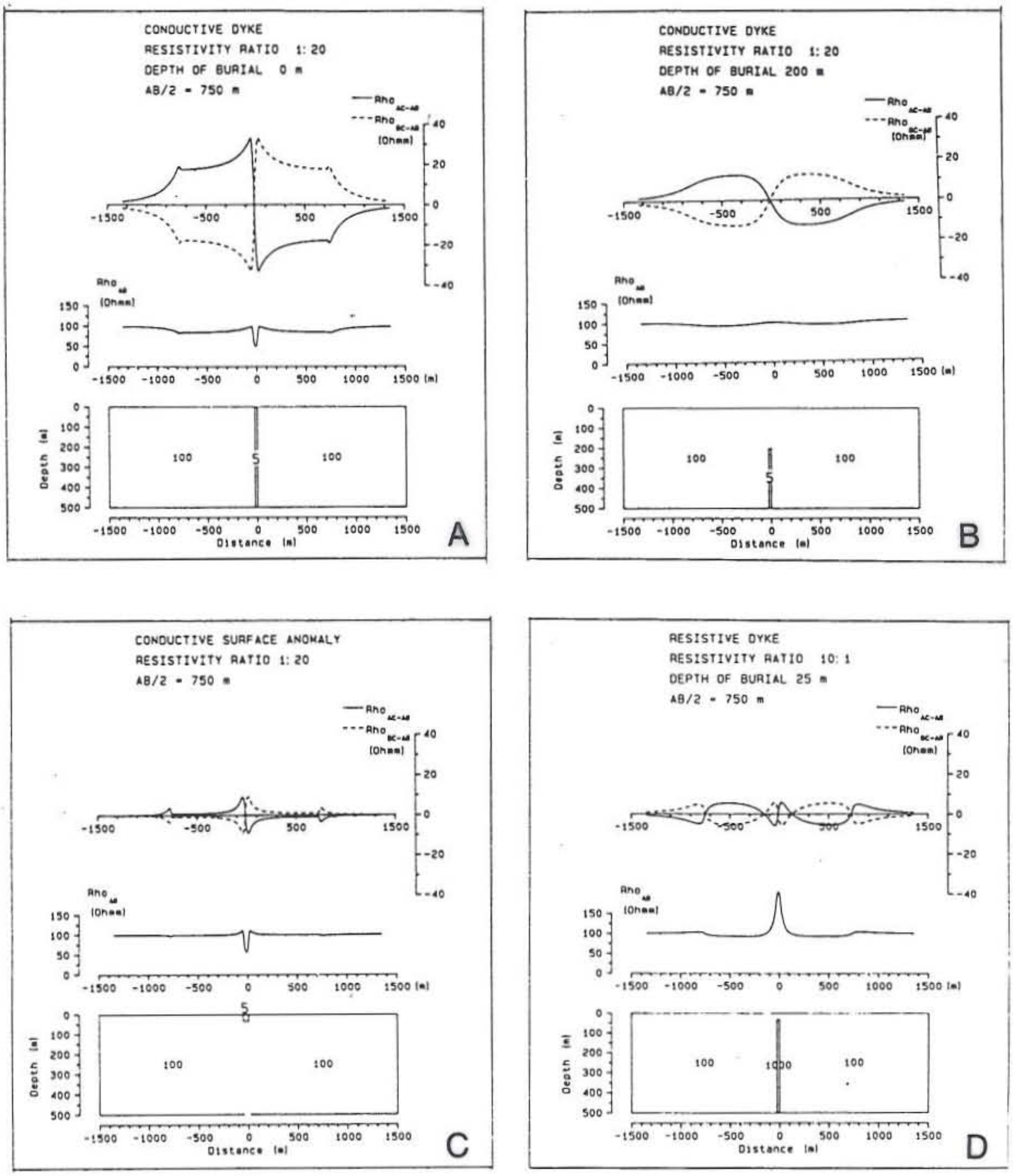

FIGURE 4: Theoretical curves which show the effect of different vertical resistivity structures in a homogeneous isotropic field, a) Low-resistivity body (dyke) in high-resistivity surroundings close to the surface; b) Low-resistivity body in high-resistivity surroundings, top at 200 m depth; c) Conductive body close to the surface; d) Non-conductive body which reaches up to the surface

a decrease elsewhere. The effect in head-on profiling of different vertical structures in a homogeneous isotropic field is shown in Figure 4 by theoretically calculated curves (Flóvenz, 1984).

2.4.2 The head-on data from Helgavatn

At the Helgavatn three head-on profiles have been measured across the main geothermal direction. The measurements were carried out with a 300 m current arm (AB/2 = 300 m), but in addition one of the profiles was also measured for a current arm of 500 m. The potential arms were kept at 25 m (MN/2 = 25 m) and measurements made every 25 m along the profile line. The location is shown in Figure 2.

Profile 1 is measured in an area where no geothermal manifestations are found on surface nor are there any visible fractures except for the westerly trending main fault near the end of the line which is at its downthrown side. Profile 2 was measured with two different current arms, 300 m and 500 m. This line crosses the area of the main geothermal field and crosses a dyke close to a fault line ending there. Profile 3 crosses an area of minor geothermal manifestations. The development of the current arm is 300 m.

2.4.3 Interpretation

The interpretation of the measurements is based on two-dimensional resistivity modelling. It is done with the help of the FELIX computer program developed at Orkustofnun, which is composed of two separate programs, LIKAN and TULKUN. LIKAN is for the construction of the resistivity models in the form of a cross-section composed of rectangular and triangular resistivity blocks, each with different resistivity. This model is used as an input file in TULKUN, which divides each block into many mini-triangles and calculates the potential at the edges of these triangles by the fine element method, taking into consideration the relative position and the resistivity of each block, and computing resistivity values by simulating the actual electrode configuration. Inversion procedures with the help of the program INVKIN include comparison between the data and the computed results, yielding the best fitting resistivity values for the proposed structural model. If the fit is not good enough, a better structural model must be suggested and the calculations repeated.

To create models of the three resistivity profiles it was postulated that the regional resistivity is 45-60 Ωm and that the composition should include two layers with their discontinuity at 80-100 m below the surface. This assumption was based on the values of Schlumberger soundings (Figure 5) carried out in

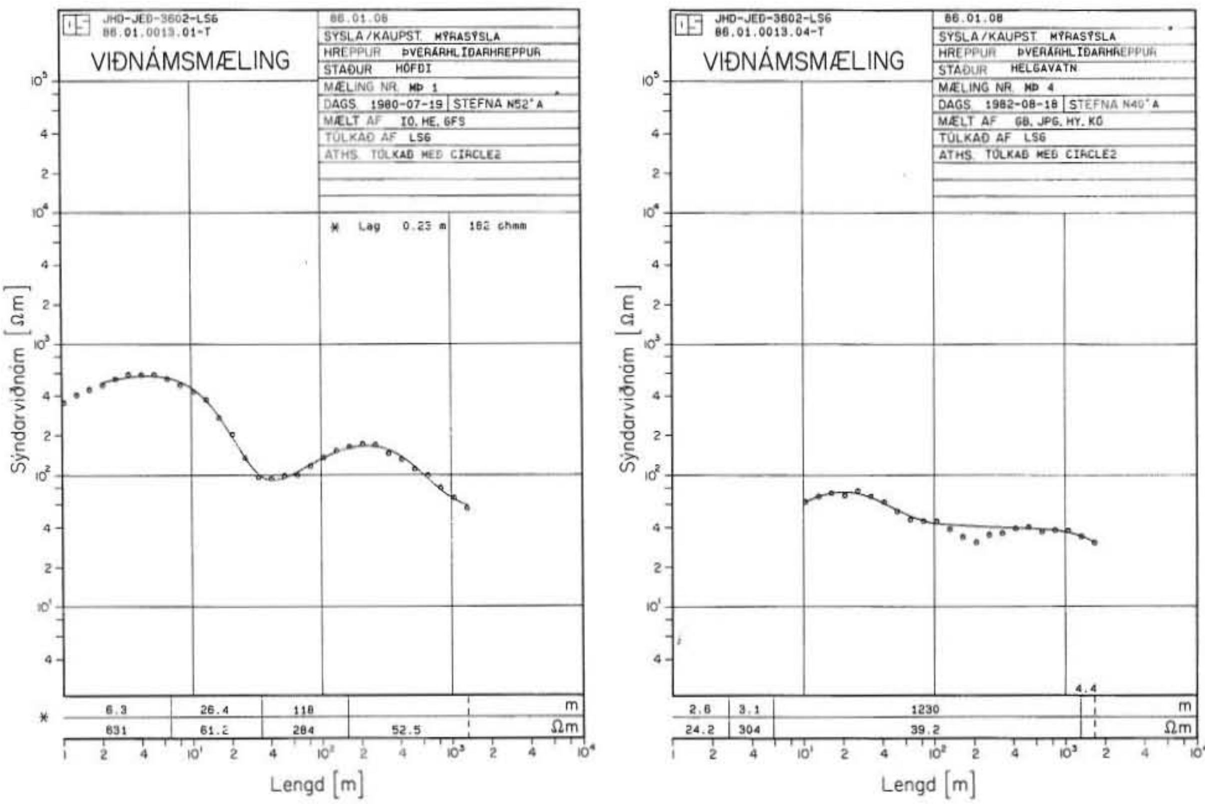

FIGURE 5: Schlumberger soundings near Helgavatn and interpreted models

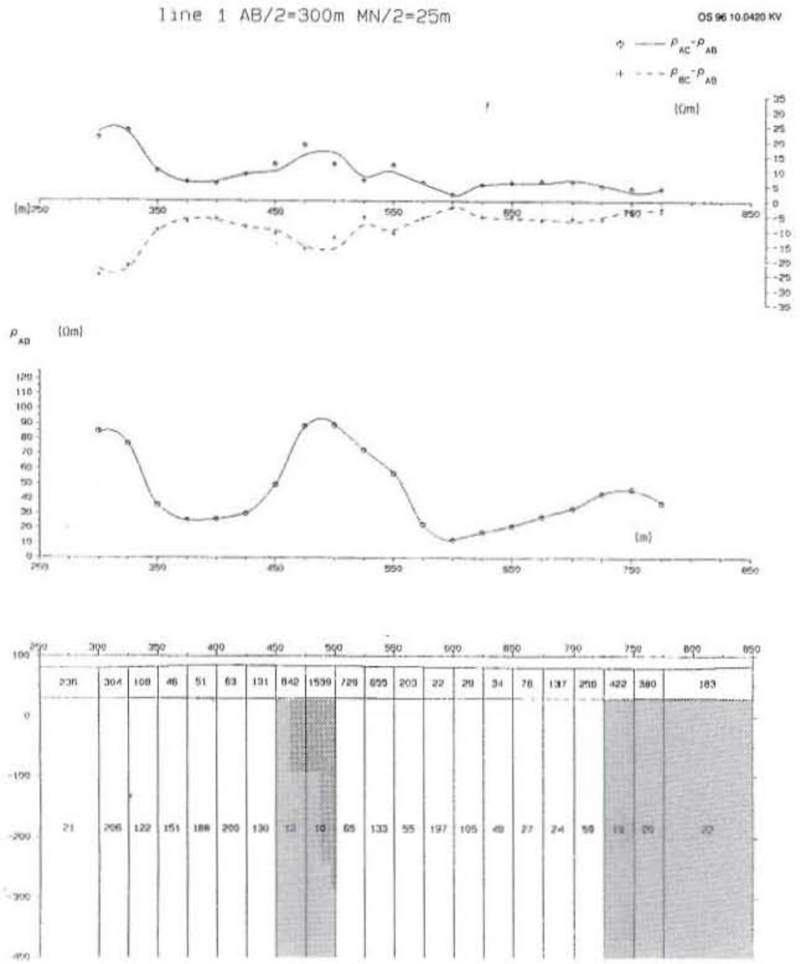

FIGURE 6: The data, model and calculated curves by the inverted resistivities for profile 1

the neighbourhood (Georgsson et al., 1984). Hence, all models for the three profiles contained a surface layer, reaching down to about 30 m a.s.l., i.e. with a thickness of about 50-100 m, and a regional resistivity of 40-60 Ωm at the outer boundaries. The process of the interpretation was as accurate as the data allowed. The model for profile 1 (Figure 6) was based on the assumptions outlined above. For profile 2 (Figure 7) the modelling followed the same concepts. However, the boundary between the two layers needed to be at a deeper level, around -30 m. Similar can be said for the model of profile 3 (Figure 8).

2.4.4 Results

Compared to the available data, the models used were perhaps a little bit too complicated. This resulted in the models being very sensitive to minor changes, which delayed the interpretation procedures. However, this was necessary in order to to be able accurately locate the interesting structures.

Profile 1 (Figure 6): Under a surface layer of 50 m, profile 1 shows two low-resistivity structures. At 450-500 m there is a well defined structure with a resistivity of 10-13 Ωm (lowest at 475-500 m). Another low-resistivity structure, of approx. 20 Ωm, is seen at 725-850 m, but it is not as well defined partly due to it being close to the end of the measuring range.

Profile 2 (Figure 7): In the interval 650-775 m, a broad low-resistivity structure is shown at the surface. Below it there are two low-resistivity structures, one with a very distinct resistivity low at 725-775 (6-15 Ωm), where $\rho_{AC} - \rho_{AB}$ and $\rho_{BC} - \rho_{AB}$ cross clearly. The other has a resistivity 10-17 Ωm and is seen at 850-925 m. Profile 2 is measured with two different current arms (300 and 500 m). Therefore, the results of the inversion are supported by more data and hence are more reliable. Both low-resistivity structures coincide with known geological structures.

Profile 3 (Figure 8): Under a surface layer three narrow low-resistivity structures are seen at 300-325, 375-400 and 475-500 m. The structure at 300-325 m is not well defined as it is close to the end of the measuring range. The low-resistivity structure at 375-400 m coincides with the crossing of data lines, $\rho_{AC} - \rho_{AB}$ and $\rho_{BC} - \rho_{AB}$.

2.4.5 Conclusions and recommendations

Figure 9 shows the location of the low-resistivity structures revealed by the head-on survey and the most probable connections between the profile lines. The results turned out reasonably well and fit the geothermal manifestations and the known geological structures. In profile 1 (Figure 6) the lowest resistivity coincides with the extension of the NE-SW dyke shown on the map (Figures 2 and 9). The low-resistivity structure at the end of the line suggests connection with the active W-E trending fault that crosses the Helgavatn hot springs. In profile 2 the broad low resistivity seen at the surface coincides with the main hot springs of the area, and the deeper low-resistivity structure definitely can be correlated with the W-E trending fault. The other deeper low-resistivity structure seems to coincide with the NE-SW trending dyke. The geothermal interpretation of profile 3 is not as clear. The thin low-resistivity structures seem to correlate with geothermal manifestations which the profile crosses, and the one furthest to the south (Figure 9) coincides with the location of a warm spring. It is probably the continuation of the W-E trending fault.

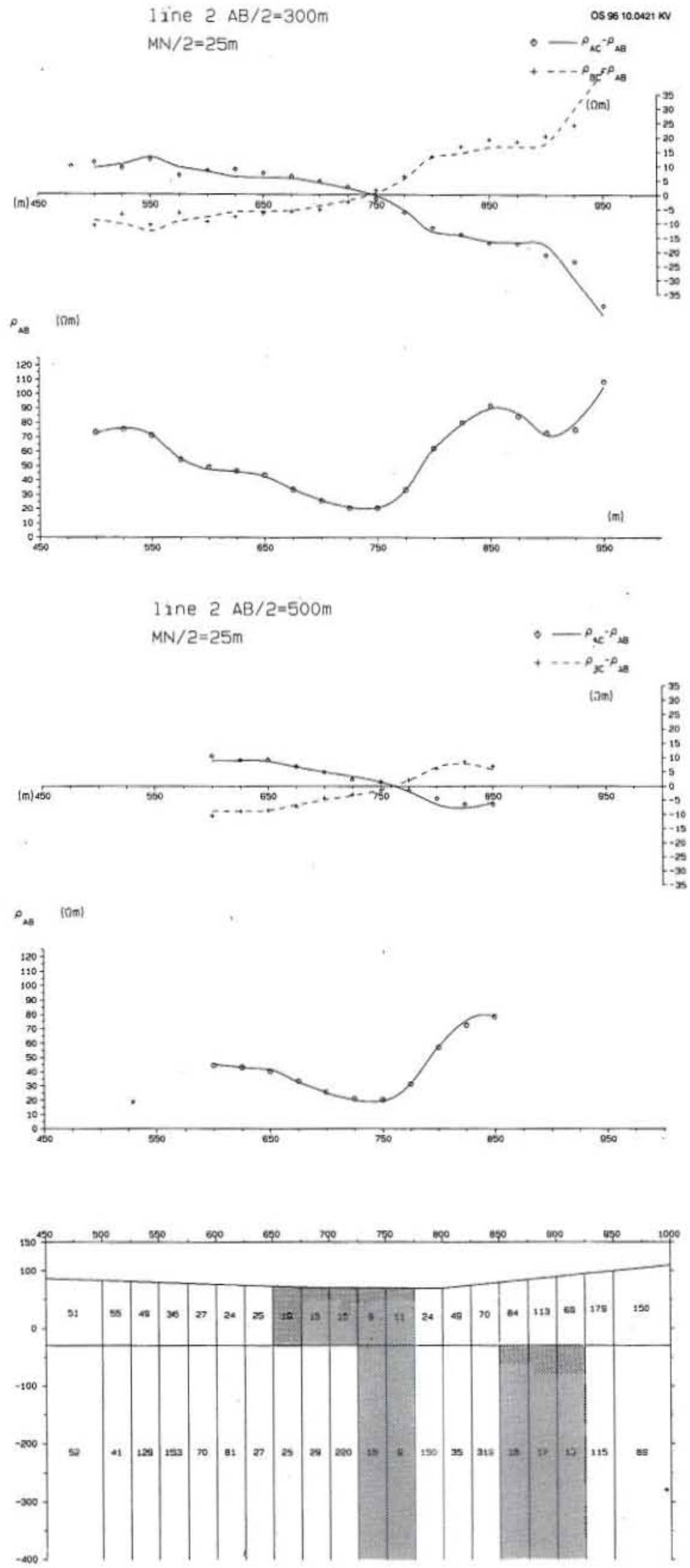

FIGURE 7: The data, model and calculated curves by the inverted resistivities for profile 2

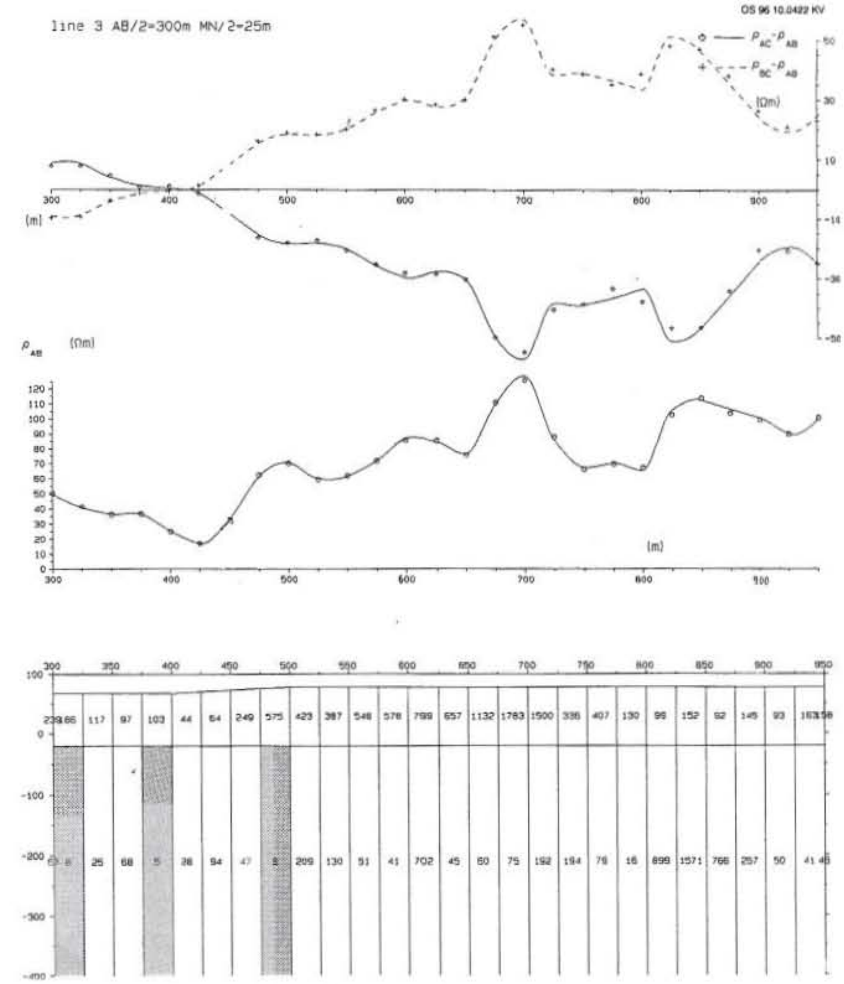

FIGURE 8: The data, model and calculated curves by the inverted resistivities for profile 3

In conclusion, it can be stated that the upflow of geothermal water at Helgavatn is controlled by two vertical structures with different directions, a W-E trending fault and a NE-SW trending dyke. The fault, which is seen clearly on the surface, is manifested as low resistivity both in profiles 1 and 2, and the upflow of the main hot springs of the area can definitely be correlated with the fault. The other fault seen close to the hot springs striking NW-SE does not seem do have any obvious correlation with the upflow of hot water. More probably, it is the NE-SW trending dyke that is also influential on the upflow. The dyke seems to be manifested as low resistivity in both profiles 1 and 2, and the main hot springs are located close to the intersection between the dyke and the fault. Georgsson et al. (1984; 1985) have stated that for the various geothermal fields of the Reykholtsdalur geothermal system, it is common that the upflow of geothermal water is related with, on one hand, NE-SW dykes or faults and, on the other hand, active N-S to W-E trending fractures or faults. The NE-SW trending structures seem to carry the water to the fields while the active fractures and faults open the way to the surface. The biggest hot springs are usually at the intersections. The results of the head-on survey at Helgavatn seem to indicate a similar status there with the main hot springs at the intersection between the fault and the dyke. Whether it is the dyke or the fracture that leads the water to Helgavatn is not as clear. With regard to the location of Helgavatn, both cases can be argued for and there is no data to conclude which is more probable.

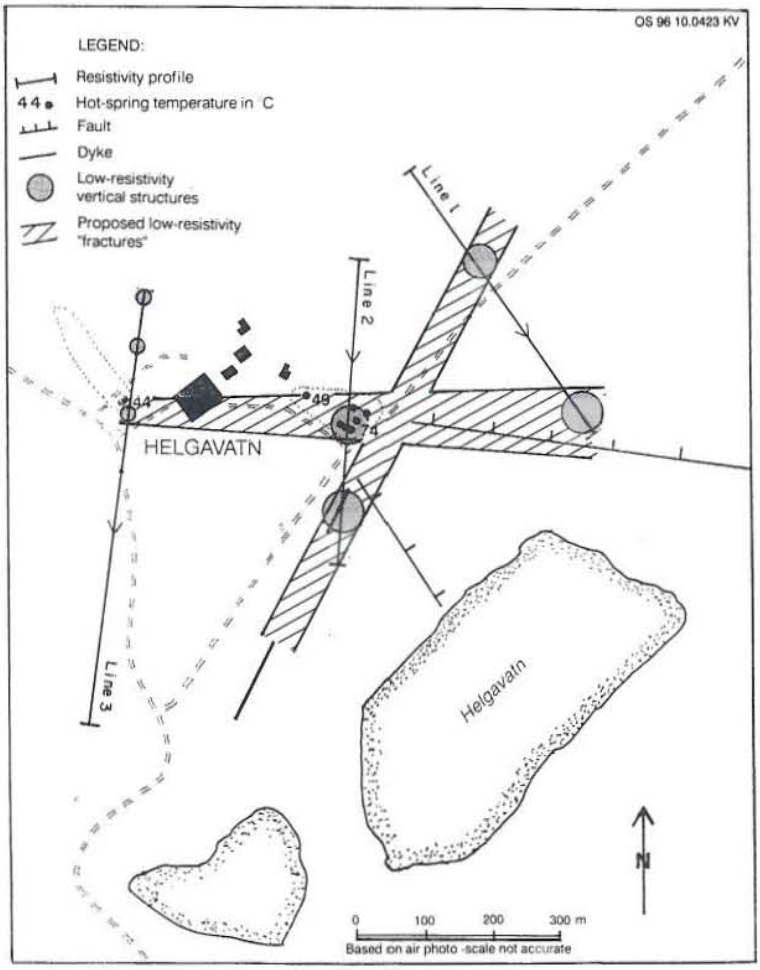

FIGURE 9: Low-resistivity structures seen in the head-on profiles at Helgavatn

3. GEOPHYSICAL EXPLORATION OF THE ÁRSKÓGSSTRÖND AREA

Árskógsströnd is a small community of only a few hundred people at the western side of Eyjafjördur, a long fiord in N-Iceland. Two small fishing-villages are at Árskógsströnd, called Litli-Árskógssandur and Hauganes, and the distance between them is about 5 km (Figure 10). In addition some common service buildings and a school are located close to the main road approximately 3 km from the villages. Presently, the houses at Árskósströnd are heated by electricity but since the price of electricity for space heating is considerably higher than from geothermal resources in Iceland, a geothermal exploration project was launched this year. A part of the data obtained is presented here and interpreted.

Geothermal surface manifestations were already known at two locations in the area, at Hálshöfdi, where there is a small hot spring, and at Ytri-Vík where a temperature gradient well has shown clear signs of geothermal activity (Figure 10). Both these sites are rather distant from the villages and the market is too small to pay for such long transmission pipelines. Therefore, exploration was aimed at the area between the villages and their immediate vicinity.

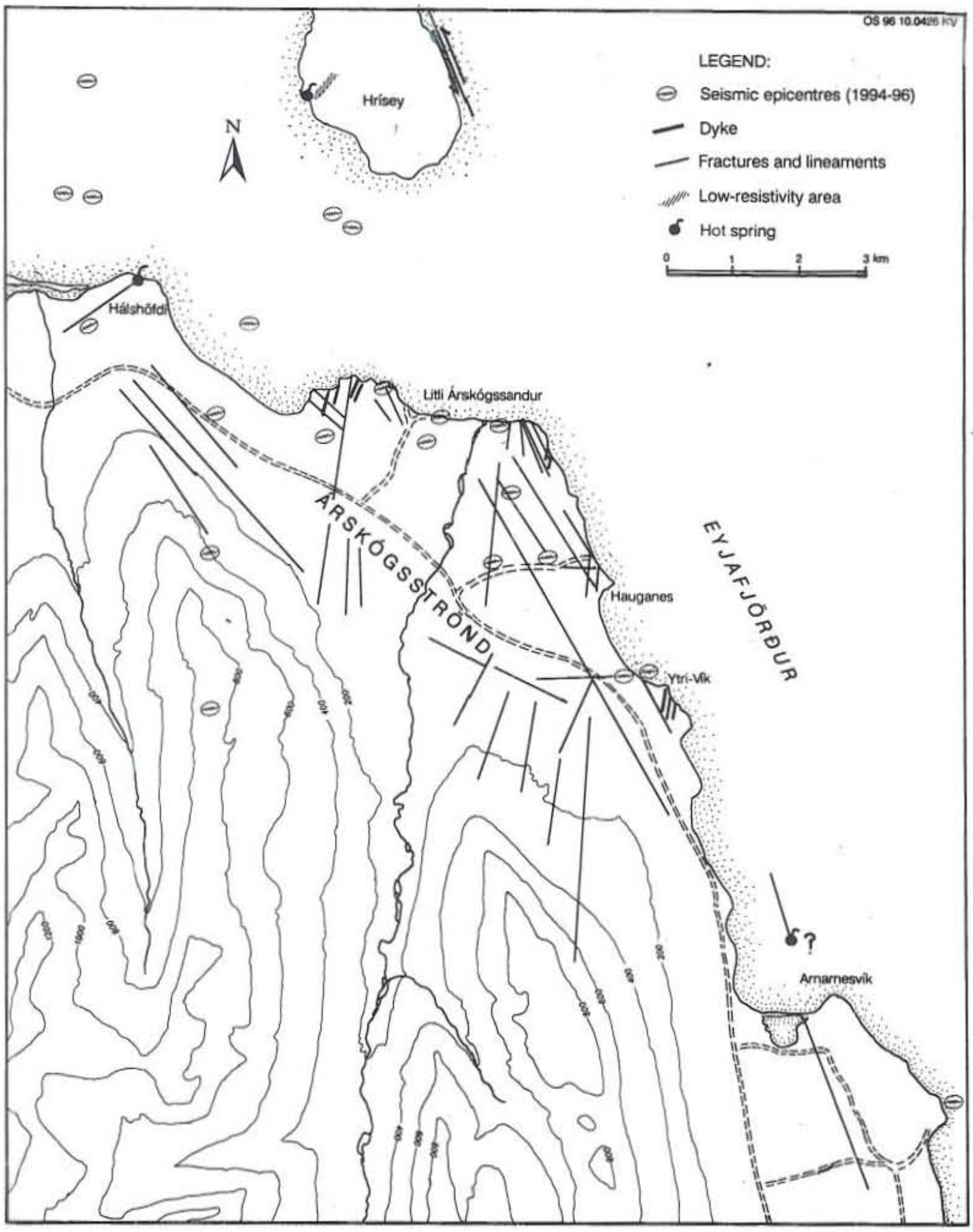

FIGURE 10: The Árskógsströnd area, the distribution of earthquake epicentres and main tectonic features

3.1 Geology and tectonics

The research area, Árskógsströnd, is approximately 60 km west of the present spreading axis of North Iceland and the age of the crust there is 6-10 m.y. It consists of sequences of flood basalt which were originally formed close to the spreading axis but have now moved to the present location through the processes of ocean floor spreading. At the western border of the research area a NNE-SSW trending anticline axis goes through Hálshöfdi. West of the anticline, the dip of the lava pile is 3-4° to the SW, but east of the anticline, at the location of the research area, the dip is 6-8° to the southeast (Saemundsson, 1979). The basalt lavas are of tholeitic composition and some are porphyritic. The lava thickness varies from a few metres to about 30 m and the flows are interbedded by very thin layers of scoria and sediments. The lava pile is intersected by numerous near-vertical dykes and normal faults.

A few kilometres north of Árskógsströnd, the WNW-trending Dalvík linearment crosses Eyjafjörður. It is an active seismic zone, and proposed to be an old transform fault connecting the northeastern volcanic zone to the Kolbeinsey ridge north of Iceland. A right lateral strike slip movement should be expected but no such movements have ever been observed. The only strike slip fault observed in this area is a north-northwest striking left lateral fault along the east coast of the Hrísey island in Eyjafjördur (Fridleifsson, 1989). The Dalvík lineament is seismically active. In the year 1934 the town of Dalvík was badly damaged an earthquake, of magnitude 6 on the Righter scale, located at this lineament.

Glacial erosion of the Ice age has removed about one kilometre of the original lava pile at Árskógsströnd and brought to the surface lavas that have been buried at that depth and reheated there to about 100°C during the spreading process. During this process the lavas suffered alteration of zeolitic type with mesolite and scolecite as the dominant secondary minerals. As a result the permeability of the rocks is quite low because the pores and fractures, which compose the primary permeability, are full of the secondarily-formed minerals. However, recent tectonic activity, possibly related to postglacial rebound or the present seismic zone, has created fractures in the rocks creating secondary permeability and, thus, opening the way for the formation of convective geothermal fields.

3.2 Exploration strategy

In order to look for possible geothermal fields within the research area the following methodology was chosen.

a) Since the area is close to an active seismic zone and a seismic station is run at Árskógsströnd it was decided to map the distribution of seismic events after the station was installed in 1994. The purpose was to try to find some lineament in the distribution of seismic epicenters that could indicate active fractures and thus the possibility of geothermal activity.
b) By careful inspection of aerial photos, possible lineaments were mapped.
c) Since the basement rock is clearly exposed at the shore of Árskógsströnd, fracture lineaments and dykes at the shoreline were investigated and compared to other observations.
d) A network of shallow boreholes for temperature measurements was drilled. The siting of the boreholes was decided with respect to the fracture analysis in the above mentioned terms.

3.3 The local seismicity

Data for the period of January 1994 to June 1996 were used from the seismic data bank of the Icelandic Meteorological Office, which runs the seismic network in Iceland. The data were inspected in different ways by use of simple Unix shell programs. Several maps of seismic events of different magnitude and at different depth levels were created and analysed. Figure 11 shows a map of all seismic events of central North Iceland for this period. The main seismicity is located along the transform faults which connect the northeastern volcanic riftzone to the active Kolbeinsey spreading ridge north of Iceland. The main direction of these transform faults is 307° for the one northeast of Árskógsströnd and 303° for the next one with similar direction. At Árskógsströnd a linear distribution of epicentres is seen striking at 295° (see also Figure 12). This distribution possibly belongs to a transform fault but it was not as active as the two others during the observed period. A fact which enforces this attitude is the Dalvík earthquake. Figure 12 shows the distribution of epicentres in the area around Árskógsströnd and possible lineaments deducted from it. These lineaments are of two main directions, one of which is around 20°, or north-northeasterly, and the other around 330°, or northwesterly.

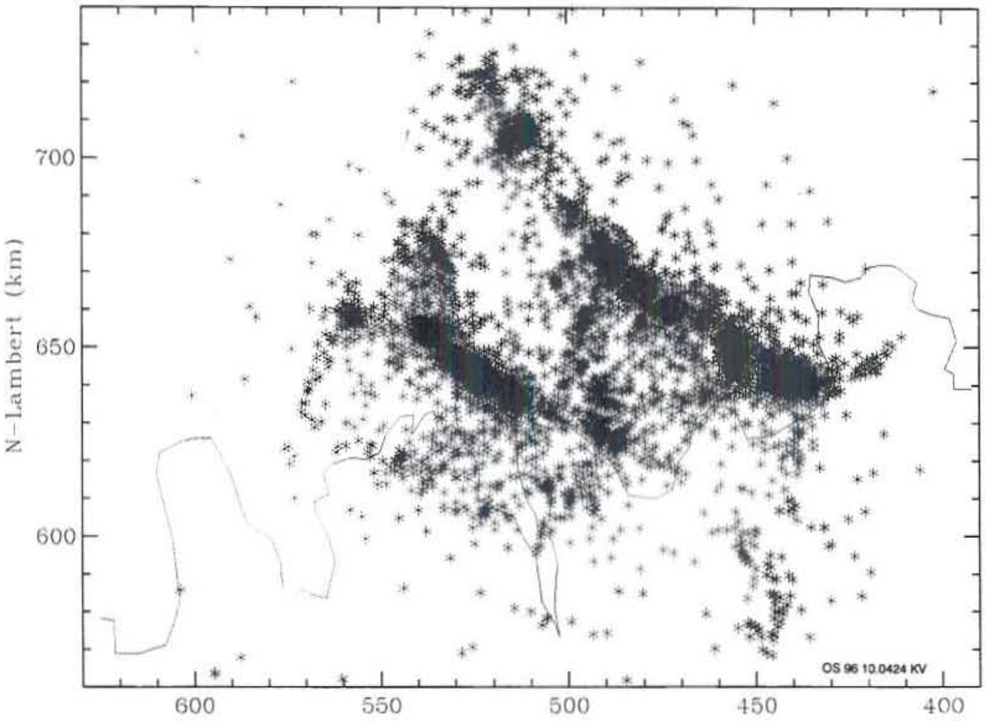

FIGURE 11: Map with the distribution of earthquake epicentres in N-Iceland

3.4 Mapping of lineaments on aerial photos

By looking carefully at the black and white aerial photos in stereoscope, a lot of possible lineaments could be identified. These are shown in Figure 10 together with the location of epicenters. Some of the events shown in Figures 11 and 12 are not included in Figure 10. These are shallow events, mainly in Hálshöfdi, and are most likely due to mining explosions.

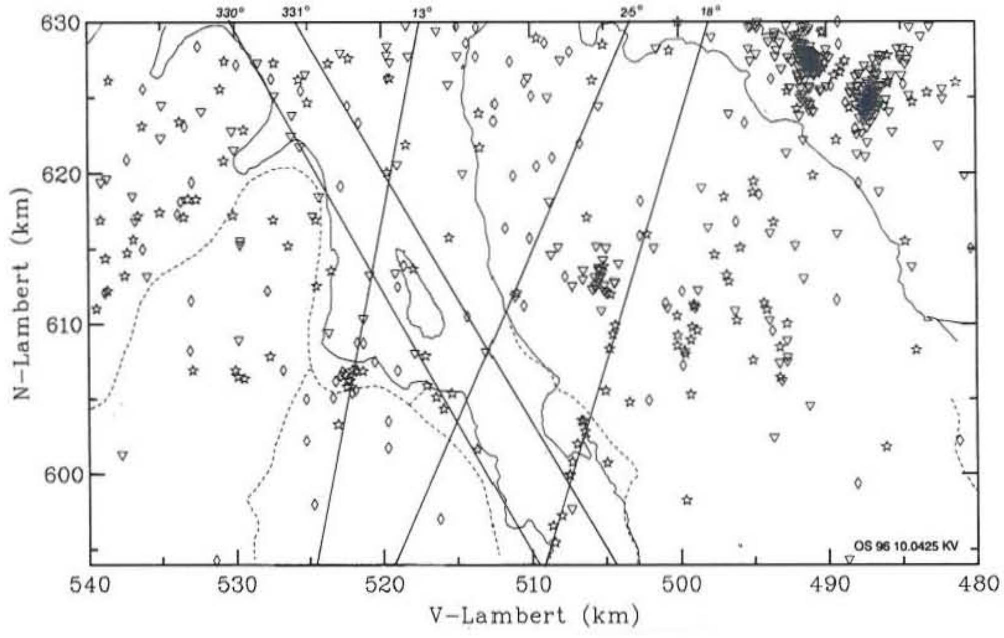

FIGURE 12: Map with the distribution of earthquake epicentres at Árskógsströnd, also shown are possible lineaments deducted from it

The lineaments on the aerial photos can be put into three groups according to their strike directions. These are 350-10°, 310-330° and 270-290° (Figure 10). To investigate the lineaments further, a one day visit to the site was organized and exposures at the coast were inspected. At the visited area the strike of the fracture systems was measured as well as the dip of the visible dykes. According to the measurements there are three main directions, 330-340°, 270-280° and 10°. Open fractures are observed at 330-340°, which seem to be the most recent fractures. The dip of the dykes is close to vertical, or about 80-85°, the thickness varies from 0.5 to 5 m and the strike is at 340-350° or at 10-20°.

In addition to the lineaments, hot springs and other signs of geothermal activity have been marked on the map on Figure 10. At the Hrísey island there is a 76°C hot geothermal field which provides water for a small village there. The original temperature in the hot spring was 51°C. This geothermal field is located at the intersection of a north-northwesterly trending fracture and another lineament striking northeast. In Merkisvík in Hálshöfdi, a hot spring of 35.5°C was found with silica temperatures of 45°C. A shallow borehole in Ytri-Vík has proven the existence of a geothermal field there with a silica temperature of 65°C. Another probable geothermal manifestation is in the Arnarnesvík bay where air bubbles have been detected by fishermen (Figure 10). Three of these geothermal manifestations, Hrísey, Ytri-Vík and Arnarnesvík seem to be linked to north-northwesterly trending fractures which are directly exposed on the north coast of Árskógsströnd, seen as a lineament on the aerial photos and as an active seismic lineament in the distribution of seismic events. This lineament was, therefore, used as the main target in the siting of the geothermal gradient wells.

3.5 Temperature gradient measurements

Altogether 12 boreholes have been drilled at Árskógsströnd. The first four, ARS-1, ARS-2, ARS-3 and ARS-4 were drilled in 1994 and 1995, but the others were drilled between August and October 1996. The location of the wells is shown in Figure 13 and the temperature logs in Appendix I. In all the wells, the temperature gradient has been calculated and the results are summarised in Table 1. The isolines for the temperature gradient are drawn on Figure 13.

For these calculations the average annual temperature is considered as 3.0-3.5°C. For well ARS-1 the surface is shifted down 12 m, because the surface layer is extremely permeable and its pores filled with water, hence the conductive zone which the calculations refer to is below that depth. At ARS-3 well there is an upwelling of hot water from the bottom to a depth of 125 m where the existence of an aquifer causes the temperature to decrease. The temperature profiles of wells ARS-4 and 5 show an upflow of relatively hot water at least from its bottom. In these cases the bottom hole temperature was used to determine the temperature gradient.

TABLE 1: Measured temperature gradient in the Árskógsströnd wells

Well sites	Temperature gradient (°C/km)	Well sites	Temperature gradient (°C/km)
ARS-01	85	ARS-07	70
ARS-02	72	ARS-08	71
ARS-03	307	ARS-09	113
ARS-04	111	ARS-10	113
ARS-05	77	ARS-11	54
ARS-06	138	ARS-12	115

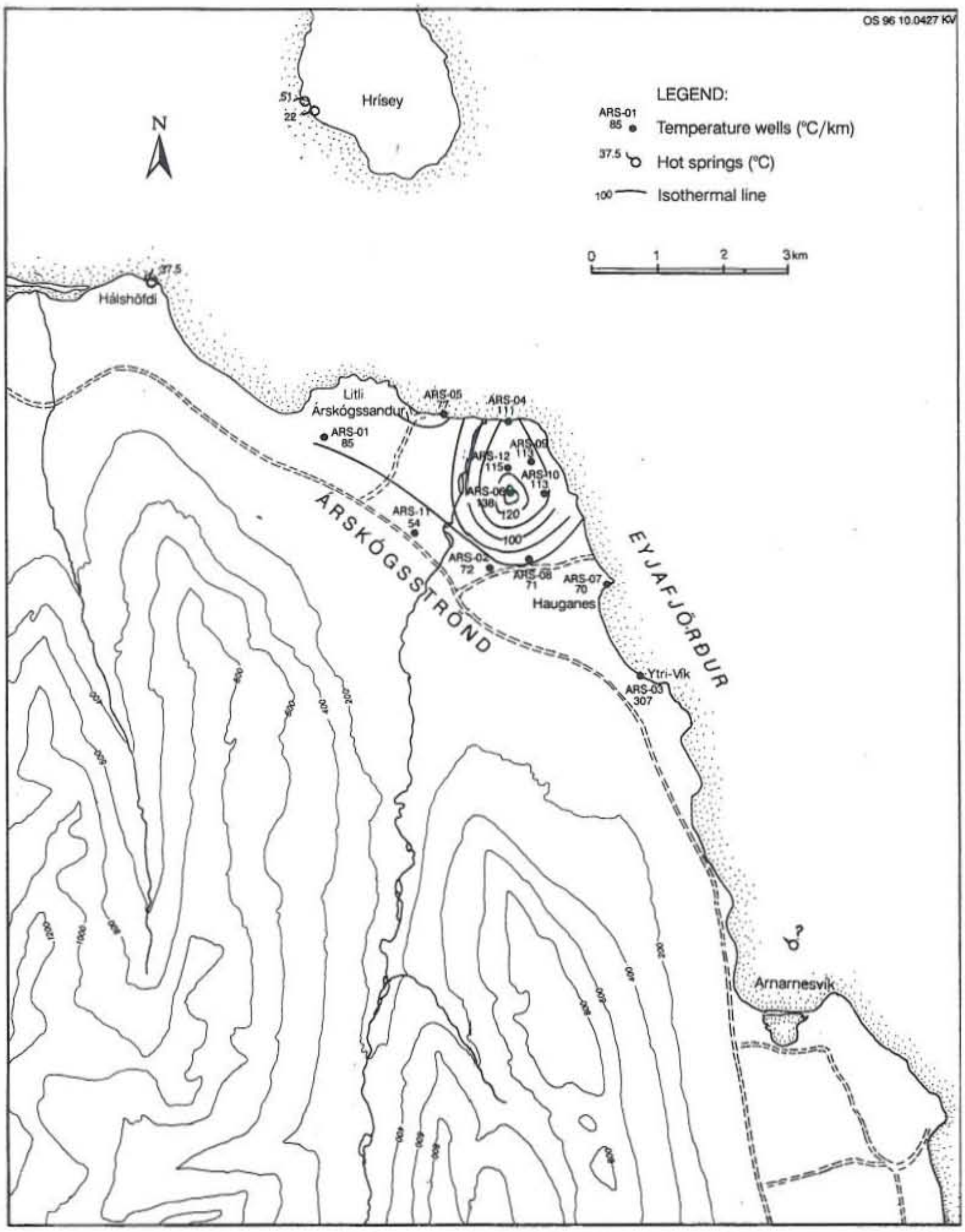

FIGURE 13: Isogradient map of Árskógsströnd and the location of geothermal manifestations

The temperature gradient wells reveal two anomalies in the Árskógsströnd region superimposed on a regional gradient of approximately 60°C/km. Apart from the exceptionally high gradient at Ytri-Vík, which certainly implies a hydrothermal system close to the well, the highest measured anomaly is 138°C/km in well ARS-06. The location (Figure 13) is good with regards to the two villages which are looking for warm water.

3.6 Conclusions and recommendations

From the comparison of the fractures and dykes observed in the field, lineaments on aerial photos, distribution of the seismic events and the isolines of the temperature gradient (Figure 14), several conclusions can be drawn. The observed fractures can be divided into three main directions, NW-SE (330-340°), WNW-ESE (270-280°) and NNE-SSW (10°). The west-northwesterly trending fractures are parallel to the "weak" transform fault close to Dalvík. The other two fracture directions seem to the

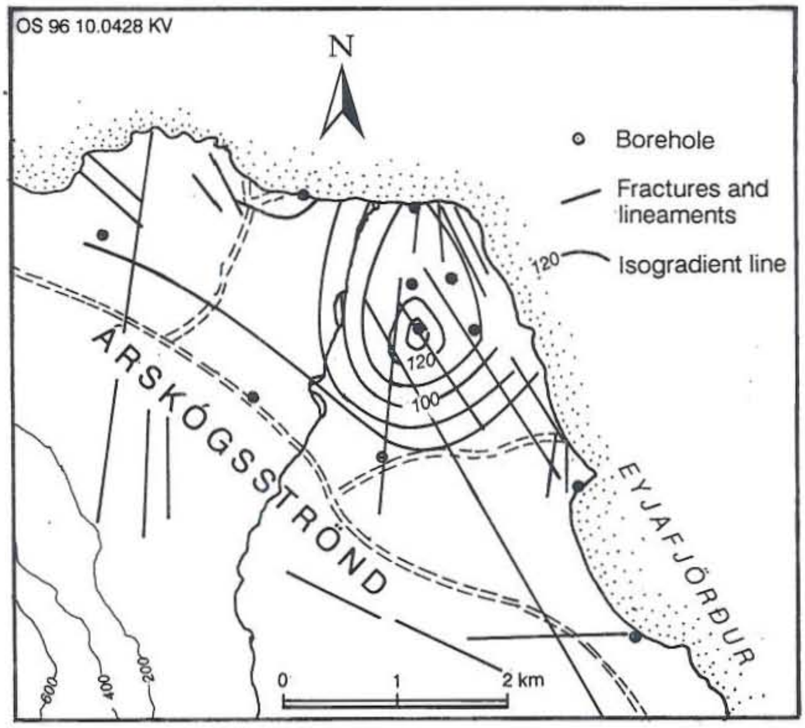

FIGURE 14: Correlation of fracture system and iso-temperature-gradient lines

result of the shear movements of the transform fault system due to the intensive field which is generated by it. However, further study on the tectonics of the area is needed to study this in detail.

The temperature gradient anomaly around well ARS-6 seems to be elongated in a northerly direction and coincides with a lineament observed from the air photos. However, more boreholes are needed around ARS-6 to determine the direction of the anomaly. It also coincides with a proposed line observed by earthquake epicentre distribution and, indeed, two earthquake epicentres happen to be in the vicinity of well ARS-6. This seismic line has north-northwesterly direction which corresponds to the youngest fracture system which appears to be open. At the extrapolation of this line to Hrísey a visible fracture striking 335° is observed close to the geothermal field on the southwestern part of the island (Björnsson and Flóvenz, 1985). It also crosses the projection on the surface of some earthquake epicenters. Furthermore, it is noteworthy that the hot springs in Hrísey are at the intersection of this northwesterly trending structure and another structure striking N20°A. This seems also to be the case at the geothermal field in Ytri-Vík, where the high temperature gradient is observed at the intersection of a northwesterly striking fracture and northeasterly and west-northwesterly striking dykes. Similar conditions appear to be present where the anomaly appears near well ARS-6 an intersection of a northwesterly-trending fracture and a northeasterly-striking dyke. This information indicates that the geothermal fields tend to be created where the northwesterly striking fracture system, which is tectonically active, intersects the north-northwesterly or west-northwesterly striking dyke systems.

Considering the above information, the area near the vicinity of the well ARS-06 is the most appropriate place to drill for hot water apart from Ytri-Vík which has been considered to be too far away from the villages for exploitation. Further research is, however, necessary before a deeper well can be sited. The anomaly has not been sufficiently well mapped, and at it is rather unlikely that a maximum geothermal gradient has been found. The maximum observed gradient is presently in well ARS-6, close to 140°C/km. If we extrapolate downward the temperature profile in this well, we will find that a temperature of 70°C can be expected below 500 m depth, if the reservoir temperature is that high. If such high temperatures exist, it will certainly be enough for economical house heating in the villages at Árskógsströnd. Therefore, the next step in the exploration at Árskógsströnd is to map the gradient anomaly in more detail with more shallow wells to define the direction and maximum value of the anomaly more precisely, map the dykes and possible faults close to the anomaly by a ground magnetic survey and, finally, drill a 400-500 m deep exploration well to test the temperature of the underlying reservoir. Alternatively, a head-on resistivity survey could partly replace the additional temperature gradient wells.

4. GEOPHYSICAL EXPLORATION USED IN LOW-TEMPERATURE FIELDS IN GREECE

Geophysical exploration can be carried out by several public and private companies, institutions, universities, etc., in Greece. They all are highly equipped and include specialists with long term experience and background. For geothermal exploration the equipment is the same, but different geology and other factors may infer that assumptions have to be specified for each area of interest.

The low temperature fields in Greece cover many wide regions of the country, especially in the northern part of the mainland and in the islands in Central-North Aegean sea where the old volcanic arch exists. The classical methods of head-on profiling and TEM can easily be used, as well as shallow drilling for gradient wells. And generally a lot of money is provided for drilling purposes. Similarly, a net of earthquake monitors has been developed by the University of Athens and the University of Thesalonikis which covers almost all of the country and can, hence, be used to follow the microseismicity for geothermal purposes.

Recently, the Greek government approved a plan for investments of the utilization of renewable forms of energy where geothermal energy is included. The investments would approximately reach the level of one billion dollars. Unfortunately, it has not yet been well defined when or how this plan will be implemented. However, this is promising for geothermal utilization as it is only a matter of time until it will be started.

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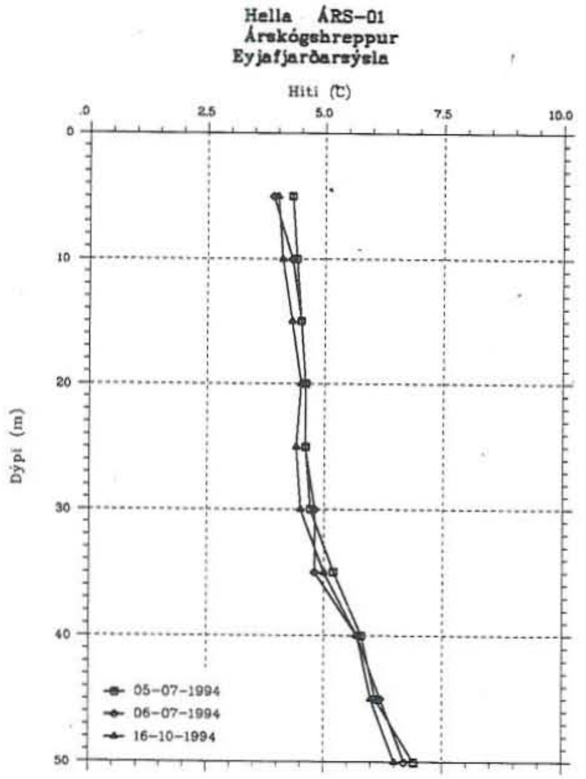
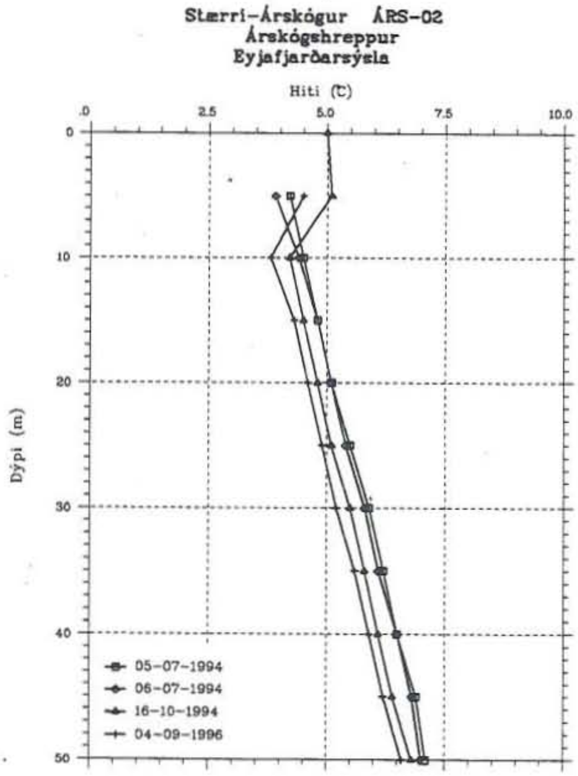
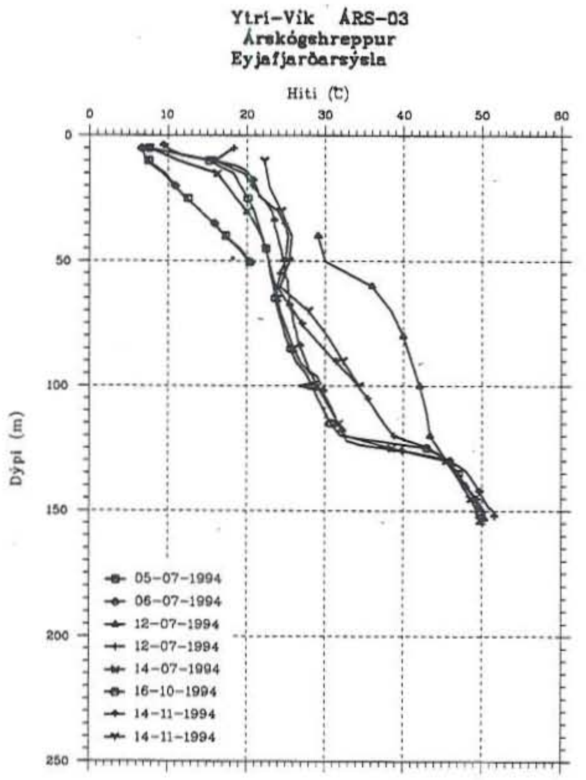
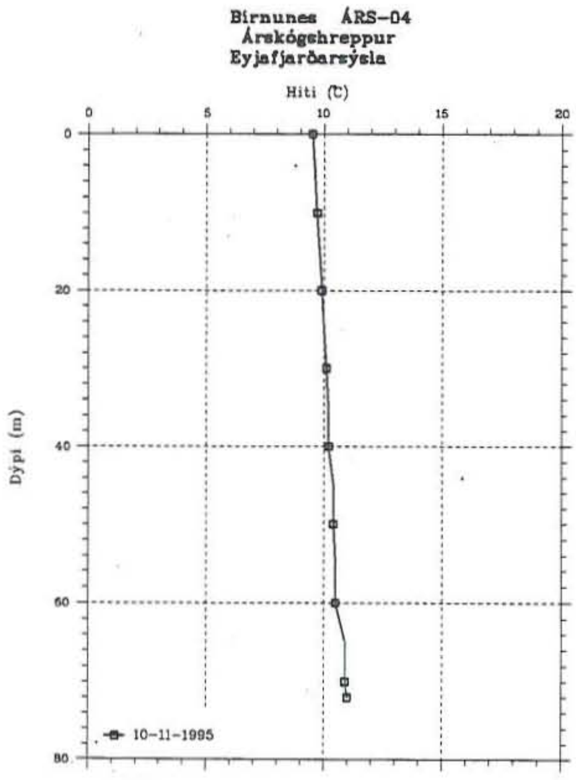
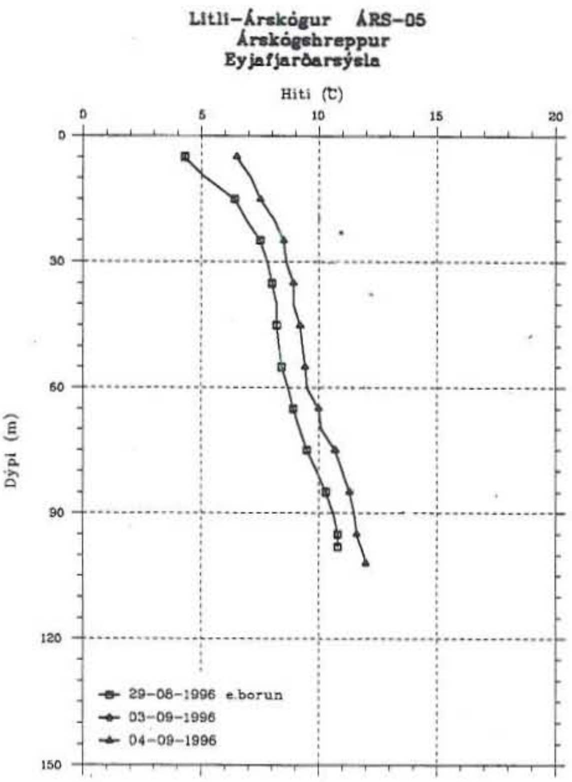
APPENDIX I: Temperature measurements of shallow gradient wells at Árskógsströnd







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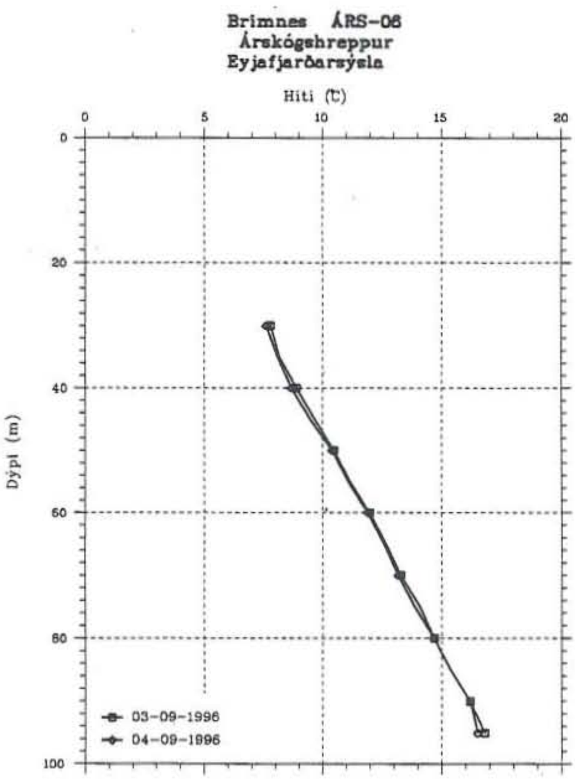
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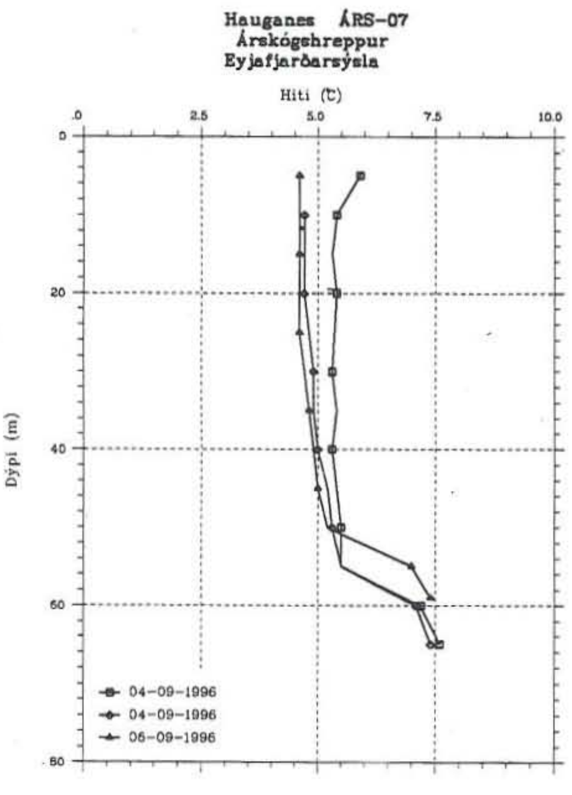
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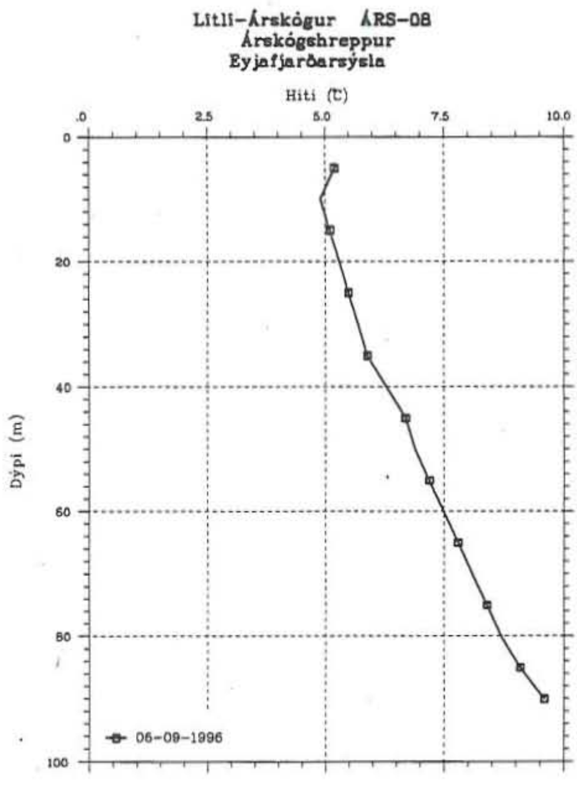
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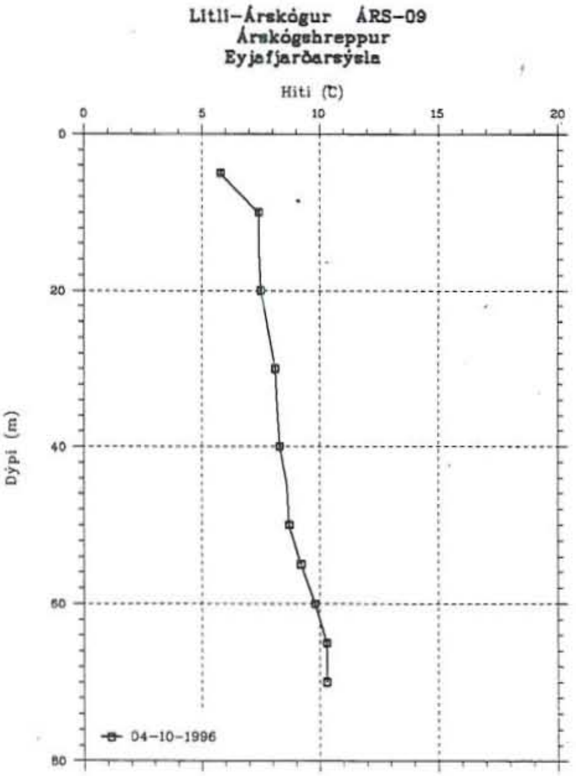
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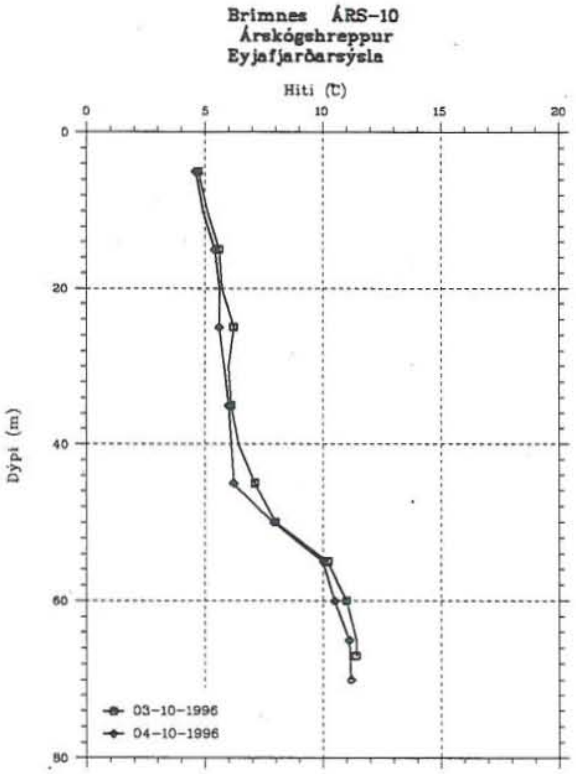
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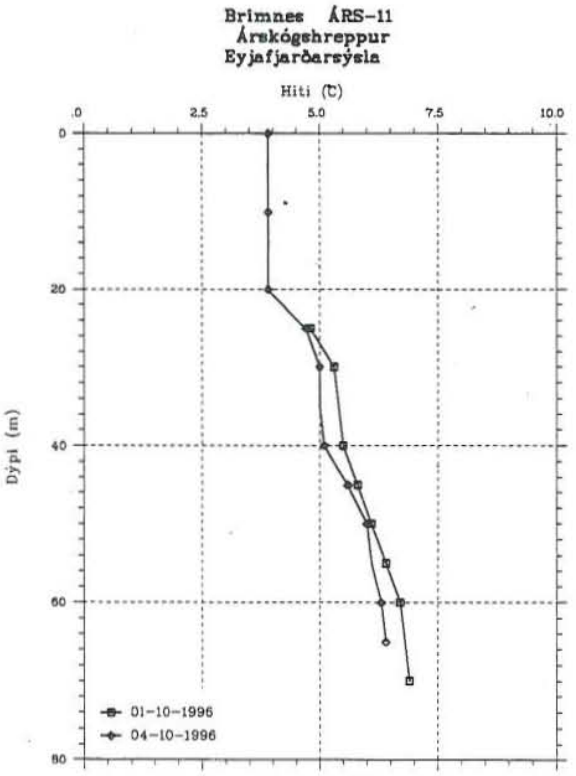
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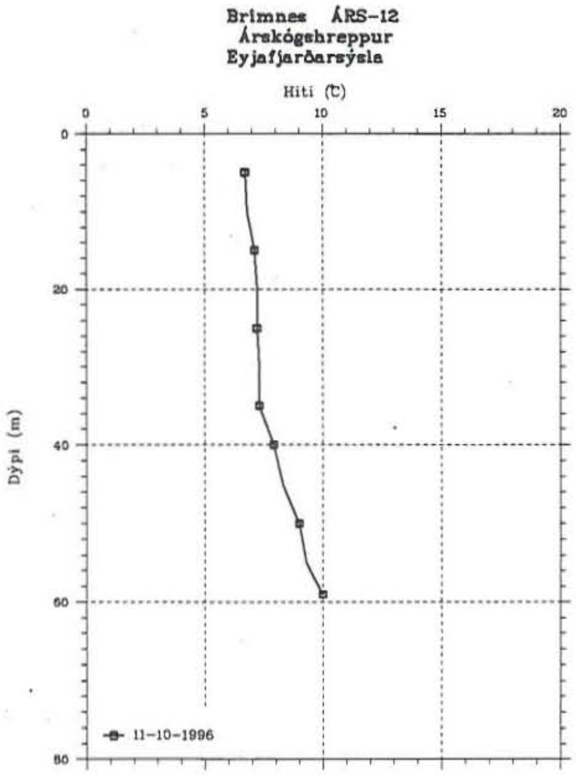
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