

# RESISTIVITY METHODS IN GEOTHERMAL PROSPECTING IN ICELAND

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**Abstract.** Resistivity techniques have been used successfully to identify and delineate geothermal resources in Iceland. The most frequently used techniques include Schlumberger, central loop TEM and head-on profiling. Geothermal systems in Iceland are located both within and outside the active volcanic regions. Outside the active volcanic regions the temperature in the upper most kilometer of the geothermal systems is below 150° C whereas the temperature in the geothermal fields within the active volcanic regions exceeds 200° C. The resistivity of the rock in geothermal fields located outside the active volcanic regions ranges from about 10  $\Omega$ m to some hundreds of  $\Omega$ m, and are characterized by considerably lower resistivity than of the surrounding rocks. Most of the geothermal systems within the active volcanic regions, show common resistivity structure with low resistivity of 1–5  $\Omega$ m surrounding an inner core of higher resistivity. This increasing resistivity with depth is associated with a change in the conduction mechanism, from interface conduction to electrolyte conduction due to a change in alteration minerals at about 240° C. Examples of resistivity surveys of geothermal fields from both outside and within the active volcanic regions are discussed.

## Introduction

Among the physical parameters characterizing the porous crust, resistivity is among the most easily measurable from the surface. As resistivity is highly dependent on temperature, resistivity methods are very useful in finding and delineating geothermal systems. Since the first resistivity sounding in Iceland in 1947, several thousand resistivity soundings have been performed. The most frequently used methods for shallow depth exploration (<1 km) are Schlumberger and TEM soundings. Head-on resistivity profiling is used to locate vertical resistivity boundaries, such as narrow vertical aquifers and sharp resistivity boundaries. AMT and MT are used for deep crustal studies.

The purpose of this paper is to give an overview of how resistivity methods are used in exploring geothermal systems in Iceland. This is done by discussing results of resistivity surveys from geothermal fields both within and outside the active volcanic regions, but first we briefly discuss the relation of the geothermal activity to the main geology and the general resistivity structure of Iceland.

## Main geology

Figure 1 shows the main features of the Icelandic geology. The Mid-Atlantic ridge transects Iceland from southwest to northeast along the active zones of rifting and volcanism. These zones are characterized by many active central volcanos with

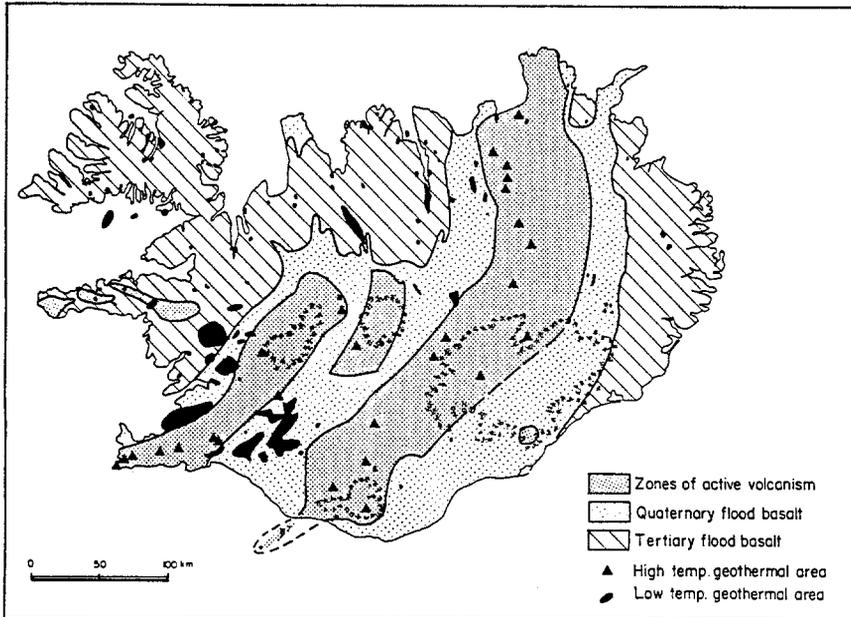


Fig. 1. Geothermal areas and the geological structure of Iceland. After Flóvenz *et al.* (1985).

fissure swarms, each 5–10 km in width and several tens of kilometers in length (Sæmundsson 1978, 1979). The geothermal systems within the active volcanic zones are found within volcanic centers or associated fissure swarms. The heat sources are cooling intrusions or magmatic bodies and temperatures up to 400° C have been found. They are called high-temperature systems in contrast to the low-temperature systems found outside the active volcanic zones, which have temperatures lower than 150° C at 1 km depth. Most of the low-temperature geothermal areas are located on the flanks of the active volcanic zones (Figure 1). Their heat sources are either cooling rocks of old volcanic centers which have drifted out from the active volcanic axis or high heat flow in young volcanic crust.

### General resistivity structure

The general resistivity structure of the uppermost 1 km of the Icelandic crust has been described by Björnsson (1976), and Flóvenz *et al.* (1985). Within the active volcanic zones it is characterized by high resistivity ( $10^4$ – $10^5$   $\Omega\text{m}$ ) in dry highly porous and fractured basalt lavas, decreasing to about 1000  $\Omega\text{m}$  below the groundwater level. At 600–800 m depth the resistivity decreases to 30–40  $\Omega\text{m}$ , compared to 1–10  $\Omega\text{m}$  within geothermal fields. Outside the active volcanic zones the resistivity is typically 100–500  $\Omega\text{m}$  down to a depth of some hundreds of meters. Below that the resistivity generally decreases to as low as 30–200  $\Omega\text{m}$ . The

geothermal fields in the low temperature areas (outside active volcanic regions) are characterized by resistivity lower than that of the surroundings; in some cases as low as 10  $\Omega\text{m}$  in an area with regional resistivity of the order of 100  $\Omega$ , and in other cases the resistivity within the geothermal systems can be as high as few hundreds of  $\Omega\text{m}$  in regional resistivity greater than 500  $\Omega$ .

There are two conduction mechanisms that reduce the resistivity of Icelandic basalts, namely pore fluid conduction (electrolyte conduction) and interface conduction. In basaltic lavas with no mineral alteration but saturated with pore fluid of resistivity  $\rho_w$ , the bulk resistivity can be described by Archie's law, i.e.,  $\rho = \rho_w \phi^{-n}$ , where  $\phi$  is the total porosity,  $\rho_w$  decreases with increasing temperature, and  $n$  is about 2. This is the pore fluid conduction. Minor alteration of the basalts leads to formation of a thin layer of conductive clay minerals like smectite at the interface of the rock matrix and the pore fluid. For typical pore fluid ( $\rho_w > 10 \Omega\text{m}$  at room temperature) this interface conduction dominates the pore fluid conduction and the bulk resistivity can be described by  $\rho \approx k \phi_f^m$ , where  $k$  decreases with increasing temperature,  $\phi_f$  is the fracture porosity rather than total porosity, and  $m$  is in the range of 1.0–1.2. In non-saline geothermal systems the interface conduction dominates the pore fluid conduction, but in saline geothermal systems this is reversed and the interface conductivity can almost be neglected. In geothermal systems of intermediate salinity both conduction mechanism have to be considered (Flóvenz et al. 1985, Árnason and Flóvenz 1992).

### Geothermal field in a low temperature area

The energy from the low temperature geothermal systems in Iceland is mainly used for space heating. Today about 85% (Pálmason 1992) of all district heating is by geothermal energy. Much effort has been put into locating and delineating those geothermal resources, and resistivity methods have been the main research tool. As a first step in a search for geothermal fields, sparse resistivity soundings are made covering a large area where geothermal activity is suspected. If that study indicates a low resistivity anomaly, further resistivity soundings are then made to delineate the geothermal field.

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, like a dyke, a fault or a fracture. A method called head-on resistivity profiling has been found to be capable of locating such vertical aquifers very precisely.

The Head-on profiling method is an extended version of a half Schlumberger array (Cheng 1980; Flóvenz and Georgsson 1982; Flóvenz 1984). In this method three current electrodes are used, A and B as in conventional Schlumberger array and a third electrode C, located perpendicular to the line joining A and B and at great distance compared to the distance between A and B. 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 between M and N, located on the line between A and B. Three apparent resistivities are calculated, the dipole (or Schlumberger) apparent resistivity,  $\rho_{AB}$  and two monopole apparent resistivities,  $\rho_{AC}$  and  $\rho_{BC}$ . 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, commonly with 25 m between measuring points. The same survey line is generally transversed with two different spacings between A and B, e.g., 500 and 1000 m, the latter for more depth resolution.

The Urriðavatn low temperature geothermal field in eastern Iceland is a good example of the application of this method (Einarsson et al. 1983). A number of hot springs were known to exist in Urriðavatn lake, and from geological mapping and magnetic surveying, the locations of many dykes and faults were also known. It was however impossible to say which of the possible aquifers, if any, carried geothermal waters. Eleven head-on survey lines were performed across the lake as shown on Figure 2. The data from one of the survey lines (line 3) is shown in Figure 3, along with 2-D interpretation. The monopole apparent resistivities are displayed on the upper part of the figure as a difference from the dipole apparent resistivity i.e.  $\rho_{AC} - \rho_{AB}$  and  $\rho_{BC} - \rho_{AB}$ . The dipole (or Schlumberger) apparent resistivity is shown in the center of the figure as a function of the location along the survey line. The two monopole curves are seen to cross right over a conductive fracture. Similar results were found for other lines, leading to the location of an aquifer carrying geothermal fluids as shown on Figure 2. In addition there is generally lower resistivity of 50–200  $\Omega\text{m}$  around this conductive fracture as compared to the 300–500  $\Omega\text{m}$  of the surroundings. Most of the hot springs are located parallel to this conductive fracture, but at a distance of about 25 m west of it, indicating that the conductive fracture dips slightly to the east. This was further supported by temperature measurements in nearby wells (Einarsson *et al.*, 1983). A well was sited so that it would cut the aquifer at a depth of about 900 m. The drillhole (well no 8 on Figure 2) was hotter and much more productive than previous wells.

### The high temperature geothermal field at Nesjavellir

The main effort in geothermal prospecting in Iceland has been in studying the low temperature fields. Only a few of the high temperature fields have been studied in detail, and up to now only seven of them are utilized.

One of the best studied high temperature geothermal fields is the Nesjavellir field in southwest Iceland. It is a northeastwards extension, along a fissure swarm, from a geothermal system under the mountain Hengill, which in turn is a part of a still larger geothermal field in the Hengill area (see Figure 4). Hengill is an

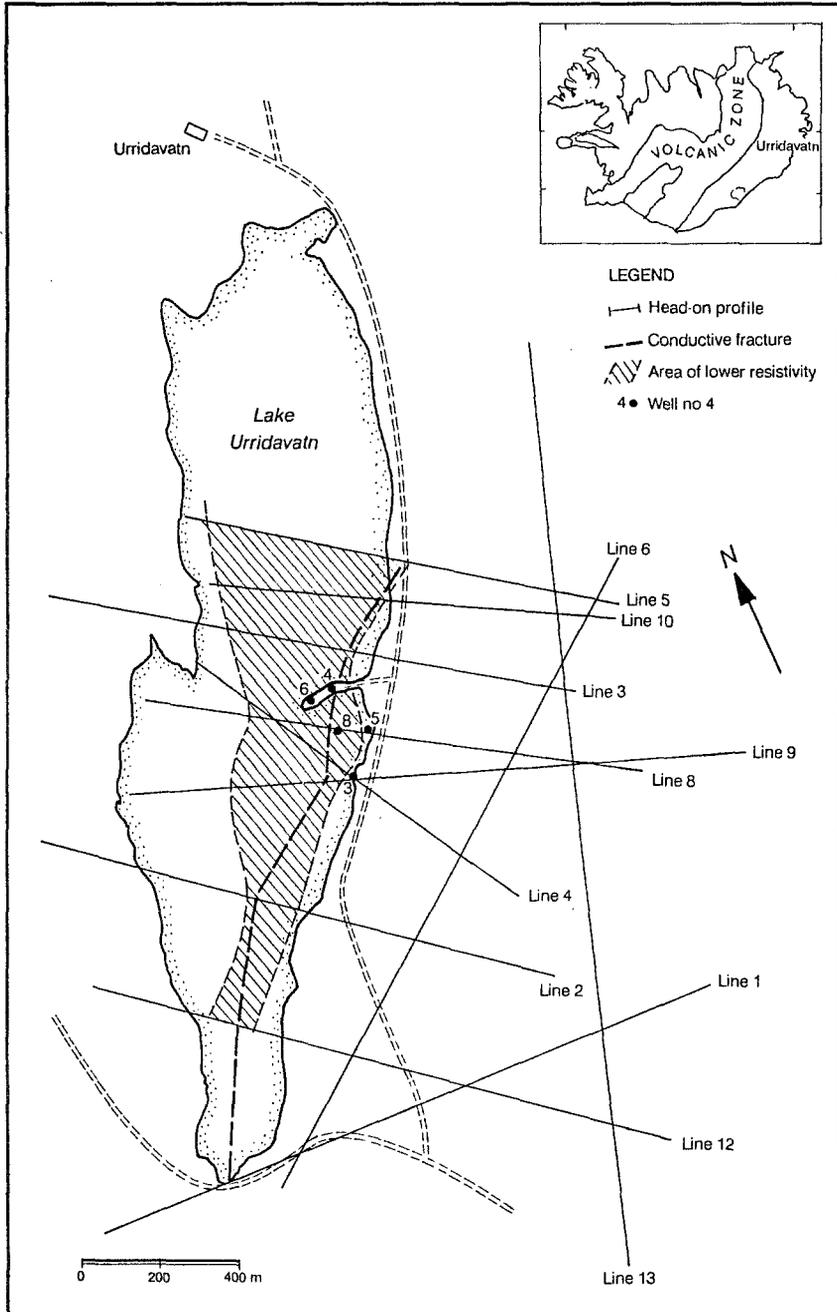


Fig. 2. The geothermal area of Urriðavatr. Redrawn from Einarson *et al.* (1983).

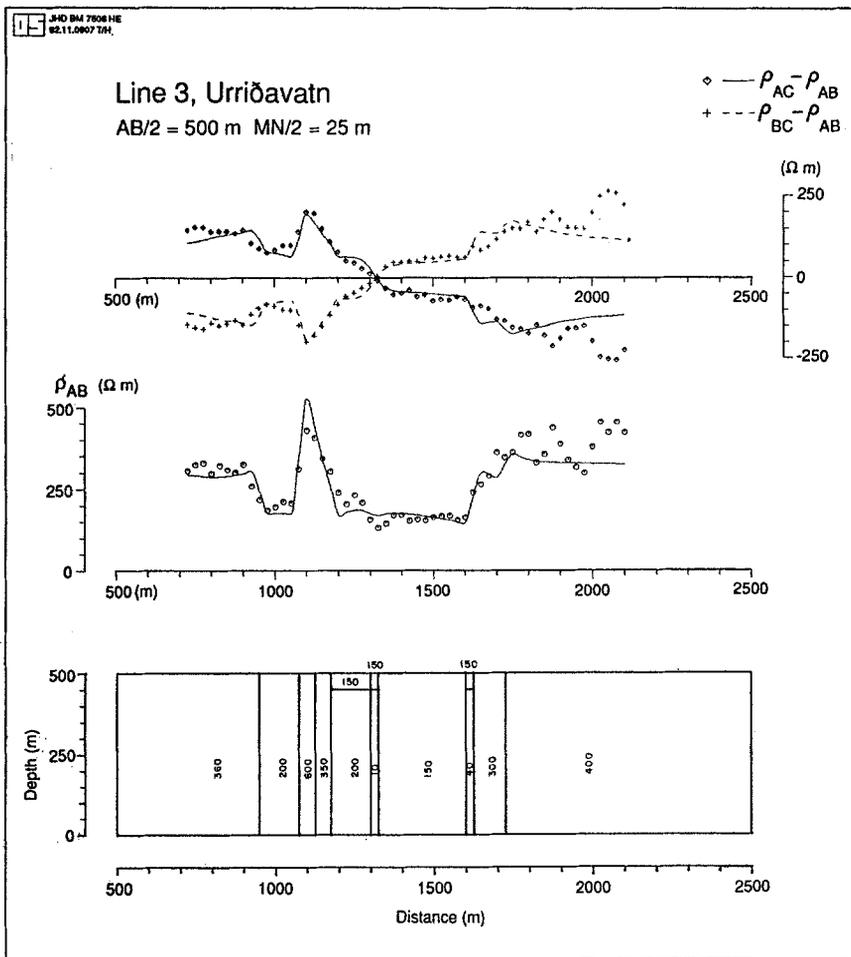


Fig. 3. Result of head-on resistivity profile, line 3. The circles are the data and the curves are the response of the 2-D model shown on the lower part of the figure. Redrawn from Einarson *et al.* (1983).

active central volcano in the western volcanic zone. The last eruption in Hengill occurred some 2000 years ago, and a rifting episode occurred 200 years ago (Árnason *et al.*, 1986).

A resistivity survey was carried out in the Nesjavellir area with the objective of delineating the geothermal system with high resolution in both vertical and lateral directions. This was achieved by applying both Schlumberger soundings (for depth resolution) and head-on profiling (for lateral resolution) along survey lines oriented perpendicular to the suspected resistivity boundaries, and by performing joint 2-D modeling of the whole data set (Árnason *et al.*, 1986, 1987a).

Figure 5 shows a simplified 2-D resistivity model from a NW-SE trending

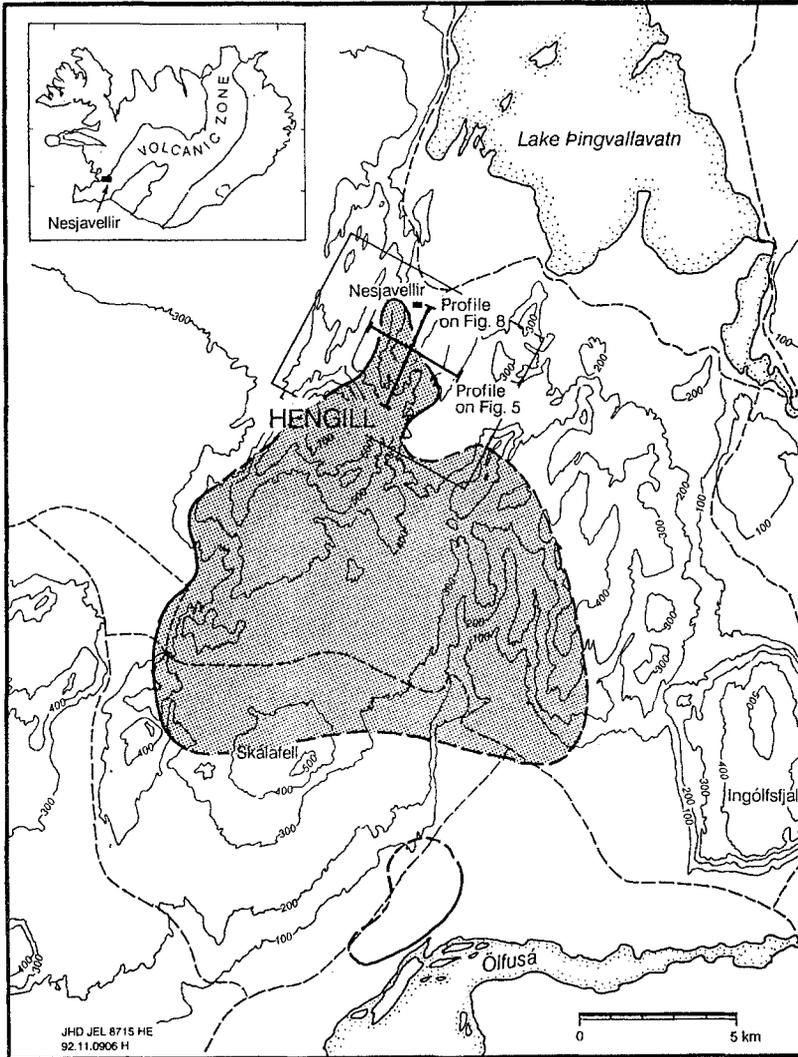


Fig. 4. The Hengill geothermal field in SW Iceland. Shaded area marks the geothermal region where resistivity is lower than  $10 \Omega\text{m}$  or high resistivity is seen below low resistivity at 400 m below sea level. Locations of two cross sections (Figures 5 and 8) from the Nesjavellir geothermal field are shown. Redrawn from Hersir *et al.* (1990).

profile (see Figure 4) perpendicular to the fissure swarm. The section also shows isothermal lines and geological information obtained from nearby wells. The resistivity ranges from few thousands of  $\Omega\text{m}$  on the surface to about  $1 \Omega\text{m}$  in the outer core of the geothermal system. The resistivity of the inner core of the geothermal system is about an order of magnitude higher than the outer core. The figure shows that the geothermal reservoir has sharp vertical boundaries, especially on

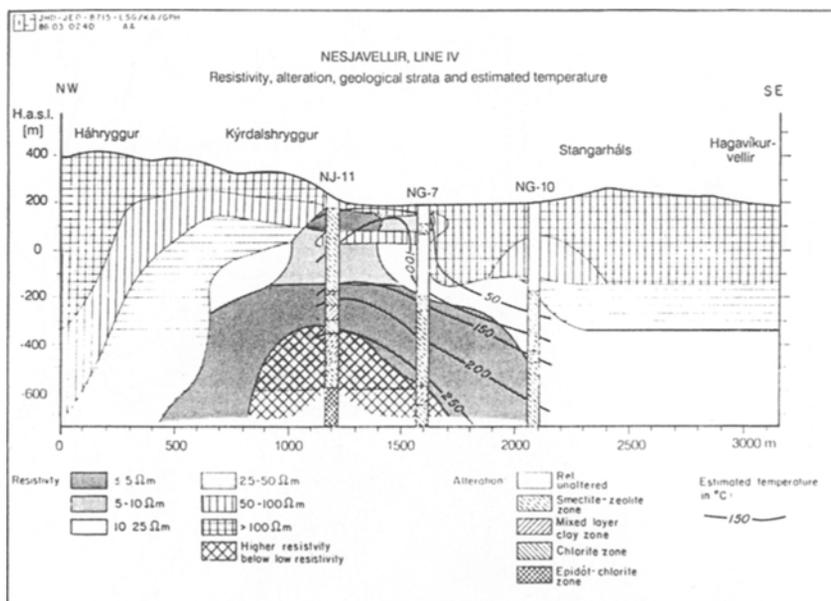


Fig. 5. Simplified resistivity cross section of the NW-SE profile shown on Figure 4, with isothermal lines and geology from nearby wells. After Árnason *et al.* (1986).

the NW side. Other survey lines perpendicular to the fissure swarm also showed sharp boundaries, indicating that the fluid is highly controlled by faults and fissures. There is a good correlation between resistivity and temperature. The resistivity is low ( $<10 \Omega m$ ) where temperature is higher than  $50^\circ C$  and is less than  $5 \Omega m$  for temperature higher than  $100^\circ C$ . But when the temperature exceeds about  $250^\circ C$ , the resistivity increases again. Similar resistivity structure is seen below the Hengill area as shown on Figure 4, where the shaded area shows the areal extent of resistivity lower than  $10 \Omega m$  or where high resistivity is found below low resistivity, at 400 m below sea level. This area marks the lateral extent of the Hengill geothermal field at this depth (Björnsson and Hersir 1981, Björnsson *et al.*, 1986; Hersir *et al.*, 1990).

The 2-D resistivity cross sections made in the Nesjavellir area were used to compile iso-resistivity maps at different depths. Figure 6 shows resistivity at sea level (200–400 m below surface). Low resistivity is found under Hengill volcano extending towards NE along the fissure swarm. Relatively low resistivity extends towards SE indicating connection to the geothermal activity east of Hengill. Figure 7 shows resistivity at 500 m below sea level (700–900 m below surface). An elongated low resistivity anomaly is seen 2–3 km NE of mount Hengill and an area of relatively high resistivity which is overlain by low resistivity occurs in the S and W, extending under Hengill. The cross section on Figure 5 (and other sections) shows that areas with resistivity lower than  $5 \Omega m$ , as well as areas with relatively



Fig. 6. Resistivity map of the Nesjavellir geothermal field at sea level. The location is shown on Figure 4. After Árnason *et al.* (1987).

high resistivity underlying low resistivity, are to be interpreted as geothermal reservoir.

The elongated low resistivity anomaly at 500 m below sea level has an NNW-SSE extension, making an angle of  $45^\circ$  with the presently active faults and fissures. Stratigraphic studies based on analyses of cuttings from wells strongly indicate the existence of N-S oriented faults at depth in this area, dated as younger than the last interglacial period. These faults seem to constrain fluid flow at this depth and define the NE boundary of the reservoir.

### Resistivity structure of high temperature geothermal systems

Resistivity surveys have shown that most high temperature geothermal systems in Iceland have a similar resistivity structure to the one found at Nesjavellir. The outer part is characterized by a low resistivity cap of  $1-5 \Omega m$ . Below this cap, in the core of the system, the resistivity increases and is generally about an order of magnitude higher. Exception from this are high temperature systems on the outer part of the Reykjanes peninsula, which are saturated with seawater. These saline systems are characterized by low resistivity anomalies of  $2-4 \Omega m$  in an environment of about  $10 \Omega m$  (Georgsson 1984). Unlike the non-saline systems, no resistivity increase is observed towards the core of those systems.

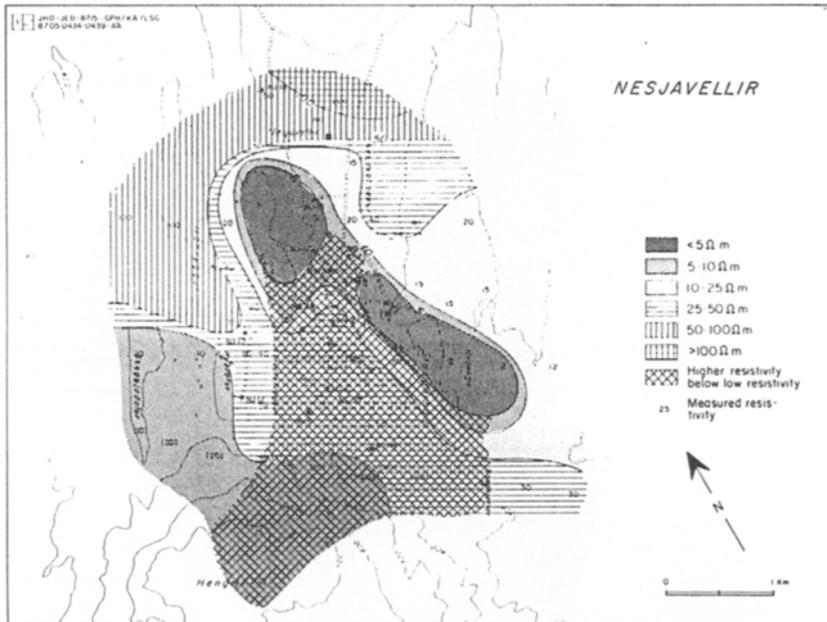


Fig. 7. Resistivity map of the Nesjavellir geothermal field at 500 m below sea level. The location is shown on Figure 4. After Árnason *et al.* (1987).

Comparison has been made between the resistivity structure of the non-saline systems, the resistivity of the geothermal fluids, lithology and alteration mineralogy (Flóvenz *et al.*, 1985, Árnason *et al.*, 1987b). This comparison shows that conduction in the low resistivity outer cap is mainly due to conductive clay minerals, such as smectite, rather than conduction in the saturating fluid.

The transition from the conductive outer cap to the more resistive inner core of the non-saline geothermal systems coincides with the 240–250° C isotherms (see Figure 5). At these temperatures, clay minerals are replaced by none conductive minerals like chlorite and epidote, and the pore fluid probably becomes the dominant conductor (Árnason and Flóvenz 1992).

In the brine systems the high conductivity of the fluid dominates conduction due to alteration minerals and therefore no major resistivity increase is associated with the transition from clay to chlorite-epidote alteration zones.

#### A comparison of DC and TEM soundings

The use of TEM soundings in geothermal exploration in Iceland has increased rapidly over the last few years. The central loop TEM method has proven to be just as effective as the combined Schlumberger and head-on profiling method, as

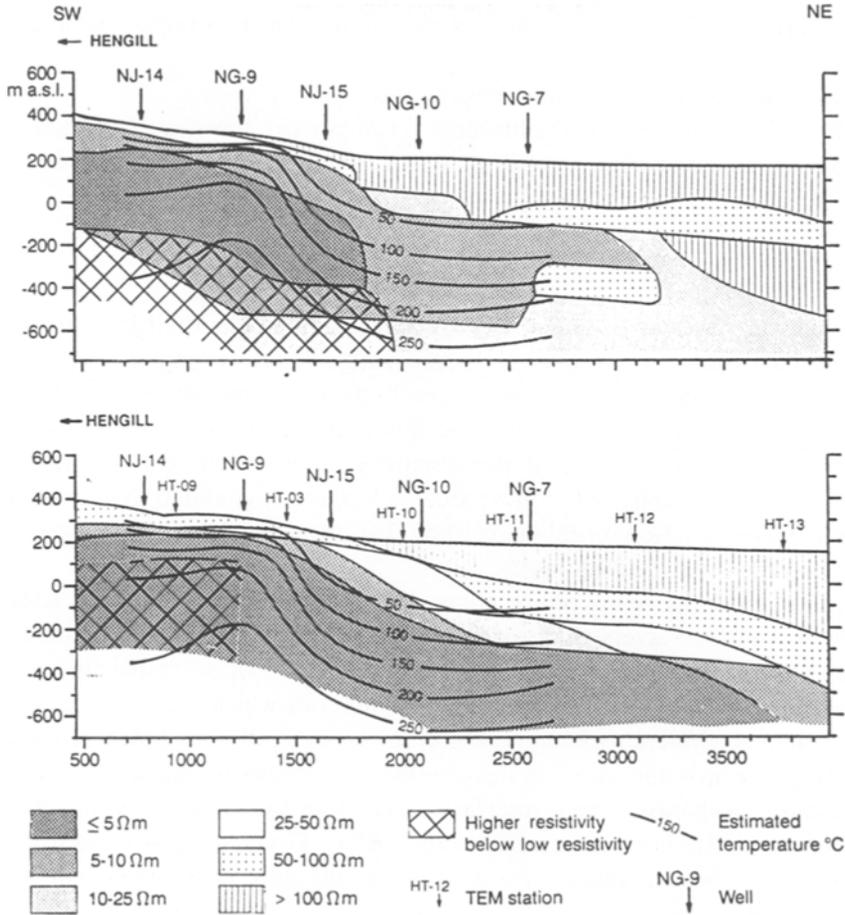


Fig. 8. Resistivity cross section from the SW-NE profile shown on Figure 4, from the Nesjavellir geothermal field. The upper section is a result of 2-D modeling of DC data, and the lower section is 1-D inversion of TEM data along the same profile. After Árnason and Flóvenz (1992).

shown on Figure 8. The figure shows that 1-D interpretation of six central loop TEM stations has similar resolution to the joint 2-D interpretation of Schlumberger and head-on profiling data. The cross section is along a SW-NE profile from the Nesjavellir geothermal field (see Figure 4). Yet the cost of the TEM cross section was only 20% of the combined Schlumberger and head-on profile. Also shown on Figure 8 are temperature isolines as measured in nearby wells. The correlation between resistivity and temperature is clear.

The main advantages of TEM-soundings as compared to DC-soundings can be summarized as follows:

- Data collection is faster and less expensive.

- No current is injected directly into the ground which is important in areas of highly resistive surface layer.
- Less distortion over local inhomogeneities.
- 1-D inversion is more justifiable since TEM is more downward focused.
- Rough areas are better accessible by performing measurements on snow.

### Conclusion

Resistivity methods have proven to be most valuable in identifying and delineating geothermal systems in Iceland. In some cases, resistivity surveying has led to identification of previously unknown geothermal systems where no surface geothermal manifestations are found. The low temperature geothermal systems are characterized by a lowering of the resistivity to less than 50% of the regional value. The geothermal fluid in these areas is frequently confined to narrow vertical aquifers. The head-on profiling method has proven to be quite successful in locating such water bearing faults.

Most of the high temperature areas show the same resistivity characteristics. Below the surface layers there is an outer cap with resistivity lower than 5  $\Omega\text{m}$  which is underlain by a more resistive inner core. These resistivity variations correlate with zones of different alteration minerals which are at thermodynamic equilibrium at ambient rock temperature. It has been postulated that the increasing resistivity towards the core of the geothermal system is due to change in the conduction mechanism, from interface conduction to electrolyte conduction due to a change in alteration minerals at about 240° C. Geothermal systems with saline fluids do not show increased resistivity at depth, since conduction is dominated by the highly conductive fluid.

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