



UNITED NATIONS  
UNIVERSITY

**UNU-GTP**

Geothermal Training Programme



## RESISTIVITY OF ROCKS

**Gylfi Páll Hersir and Knútur Árnason**

ÍSOR – Iceland GeoSurvey

Grensásvegur 9, 108 Reykjavík

ICELAND

*gylfi.pall.hersir@isor.is, ka@isor.is*

### ABSTRACT

A brief summary of the importance of the electrical resistivity of rocks in geothermal exploration is given as well as some of its main relationships with some of the characteristics of geothermal reservoirs. Special emphasis is on the resistivity of rocks in high temperature geothermal fields where the dominant factor is the hydrothermal mineral alteration and how it relates to temperature. This review is to a great extent based on experience and work done through the years by Iceland GeoSurvey and its predecessor the NEA. In this review it should be pointed out that an overview of geophysical exploration for geothermal resources was given in a UNU GTP publication (Hersir and Björnsson, 1991). Other important sources here are two more recent articles from the World Geothermal Congresses (Árnason et al., 2000 and Flóvenz et al., 2005).

### 1. INTRODUCTION

Measuring the electrical resistivity of the subsurface is the most powerful prospecting method in geothermal exploration. Resistivity is directly related to the properties of interest, like salinity, temperature, porosity (permeability) and alteration. To a great extent, these parameters characterize the reservoir.

The specific resistivity,  $\rho$ , is defined through **Ohm's law**. The electrical field strength,  $E$  [V/m] at a point in a material is proportional to the current density,  $j$  [A/m<sup>2</sup>]:

$$E = \rho j \quad (1)$$

The proportional constant,  $\rho$  depends on the material and is called the (specific) resistivity, measured in  $\Omega\text{m}$ . The reciprocal of resistivity is conductivity ( $1/\rho=\sigma$ ).

Resistivity can also be defined as the ratio of the potential difference,  $\Delta V$  [V/m], to the current,  $I$  [A], across material which has a cross-sectional area of 1 m<sup>2</sup> and is 1 m long.

$$\rho = \frac{\Delta V}{I} \quad (2)$$

Electrical conductivity in minerals and solutions takes place by movement of electrons and ions. Most rocks near the earth's surface have low conductivity. Conduction of electricity is mostly through

groundwater contained in pores of the rocks and along surface layers at the contact of rocks and solution. The case of altered rocks in high temperature areas will be discussed separately.

The common principle of all resistivity sounding methods is to induce an electrical current into the earth and monitor signals, normally at the surface, generated by the current distribution. In conventional direct current soundings such as the Schlumberger soundings, this is done by injecting current into the ground through electrodes at the surface and the signal measured is the electric field (the potential difference over a short distance) generated at the surface. In magnetotellurics (MT) the current in the ground is induced by time variations in the earth's magnetic field, and the signal measured is the electric field at the surface. In transient electromagnetics (TEM) the current is also induced by a time-varying magnetic field, but in this case the current source is not the natural field, the source is of a controlled magnitude generated by current in a loop or a grounded dipole and the monitored signal is the decaying magnetic field at the surface.

## 2. CONDUCTION IN SOLUTIONS

In aqueous salt-solution the ions of the solid separate and are free to move independently in the solution. In an electric field cations are accelerated to the negative electrode and the anions to the positive one. A viscous drag force limits the velocity of the ions. The mobility (terminal velocity/electrical field) of the ions depends on temperature (viscosity) and on concentration.

Groundwater may have a variety of salts in the solution. Therefore equivalent salinity is defined as the salinity of a NaCl-solution with the same resistivity as the particular solution. Mobility of the ions does not vary widely and equivalent salinity is therefore close to true salinity. The right part of Figure 1 shows the nearly linear relationship between the salinity of electrolytes and conductivity ( $\sigma \approx C/10$ , where  $C$  [g/l] is the concentration of NaCl).

At moderate temperatures, 0-200°C, resistivity of aqueous solutions decreases with increasing temperature. The reason is increasing mobility of the ions caused by a decrease in the viscosity of the water. Dakhnov (1962) has described this relationship:

$$\rho_w = \frac{\rho_{wo}}{1 + \alpha(T - T_0)} \quad (3)$$

where  $\rho_{wo}$  = Resistivity of the fluid at temperature  $T_0$ ;  
 $\alpha$  = Temperature coefficient of resistivity, with  $\alpha \approx 0.023^\circ\text{C}^{-1}$  for  $T_0 = 23^\circ\text{C}$ ,  
 and  $0.025^\circ\text{C}^{-1}$  for  $T_0 = 0^\circ\text{C}$ .

At high temperatures, a decrease in the dielectric permittivity of the water results in a decrease in the number of dissociated ions in the solution. Above 300°C, this starts to increase fluid resistivity (Quist and Marshall, 1968) as seen in Figure 1.

## 3. ELECTRICAL RESISTIVITY OF WATER-BEARING ROCKS

In most rocks near the earth's surface, the conduction is dominated by electrolytic conduction in aqueous solution of common salts distributed through the pores of the rock and/or at the rock-water interface. The rock matrix itself is normally an insulator. The electrical resistivity of rocks depends on:

- Porosity and the pore structure of the rock;
- Amount of water (saturation);
- Salinity of the water;
- Temperature;

- Water-rock interaction and alteration;
- Pressure; and
- Steam content in the water.

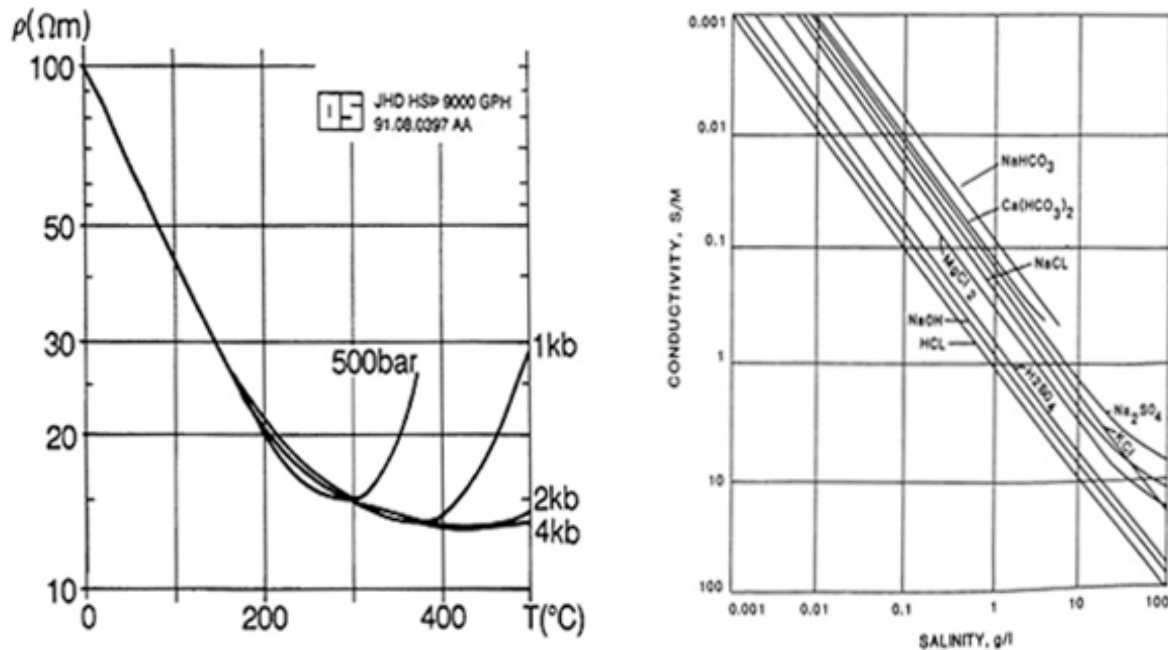


FIGURE 1: At left - electrical resistivity as a function of temperature at different pressures (from Hersir and Björnsson, 1991); at right - conductivity as a function of salinity

The most important factors are the porosity, temperature, salinity and water-rock interaction. In geothermal areas, the rocks are water-saturated. Ionic conduction in the saturating fluid depends on the number and mobility of ions and the connectivity of flowpaths through the rock matrix. Usually, the saturating fluid is among the dominant conductor in the rock and the degree of saturation is of great importance to the bulk resistivity. The pressure dependence is negligible compared to the temperature dependence, provided that the pressure is sufficiently high so that there is no change in phase.

**Porosity** is defined as the ratio between the pore volume and the total volume of a material. There are primarily three types of porosity: *Intergranular*, the pores are formed as spaces between grains or particles in a compact material (sediments, volcanic ash). *Joints-fissures*, a net of fine fractures caused by tension and cooling of the rock (igneous rocks, lava). *Vugular*, big and irregular pores, formed as material is dissolved and washed away, or pores formed by gas (volcanic rocks, limestone). Pore spaces must be interconnected and filled with water in order that fluid conduction takes place. In all types of porosity there are larger voids, called storage pores, and finer connecting pores. Most of the resistance to electric current flow (and fluid flow) is met in the connecting pores.

It has been observed for many cases that resistivity of water-bearing rocks varies approximately as the inverse square of the porosity. This empirical law, called **Archie's law** (Archie, 1942) describes how resistivity depends on porosity if ionic conduction in the pore fluid dominates other conduction mechanism in the rocks.

$$\rho = \rho_w a \Phi_t^{-n} \quad (4)$$

where  $\rho$  = Bulk (measured) resistivity;  
 $\rho_w$  = Resistivity of the pore fluid;  
 $\Phi_t$  = Porosity in proportions of total volume;

- $a$  = Empirical parameter, varies from  $<1$  for intergranular porosity to  $>1$  for joint porosity, usually around 1; and
- $n$  = Cementing factor, an empirical parameter, varies from 1.2 for unconsolidated sediments to 3.5 for crystalline rocks, usually around 2.

Archie's law is valid if the resistivity of the pore fluid is of the order of  $2 \Omega\text{m}$  or less, but doubts are raised if the resistivity is much higher (Flóvenz et al., 1985). However, Archie's law seems to be a fairly good approximation when the conductivity is dominated by the saturating fluid (Árnason et al., 2000).

Conduction in the rock matrix is normally negligible. The main contributions to the conduction mechanisms are: Conduction by dissolved ions in the pore fluid, pore fluid conduction. Conduction by adsorbed ions on the pore surface, surface conduction. Conduction in alteration minerals, mineral conduction. Figure 2 illustrates these main conduction mechanisms.

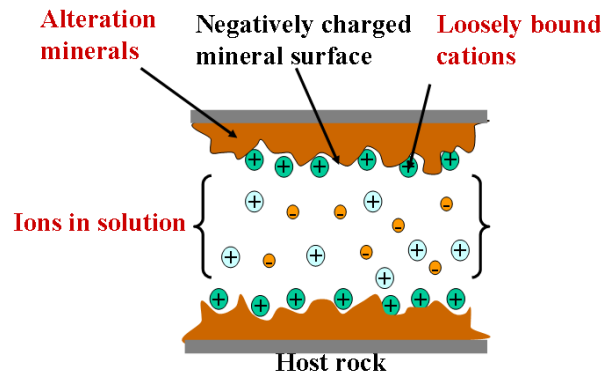


FIGURE 2: Conduction mechanisms

Figure 3 shows the relationship between the bulk resistivity and the pore fluid resistivity for different porosities and temperatures for rocks in the uppermost kilometer of the Icelandic crust outside the volcanic zone. It is based on a model put forward by Flóvenz et al. (1985). The model includes both electrolytic and surface/mineral conduction. For pore fluid resistivities less than  $2 \Omega\text{m}$  the dominant conductivity is pore fluid conductivity, Archie's law applies. For rocks saturated with fluids, having resistivity higher than about  $2 \Omega\text{m}$  and at room temperature, the bulk resistivity is practically independent of the resistivity of the fluid, but dependent on porosity and temperature, the dominant conductivity is mineral and/or surface conductivity.

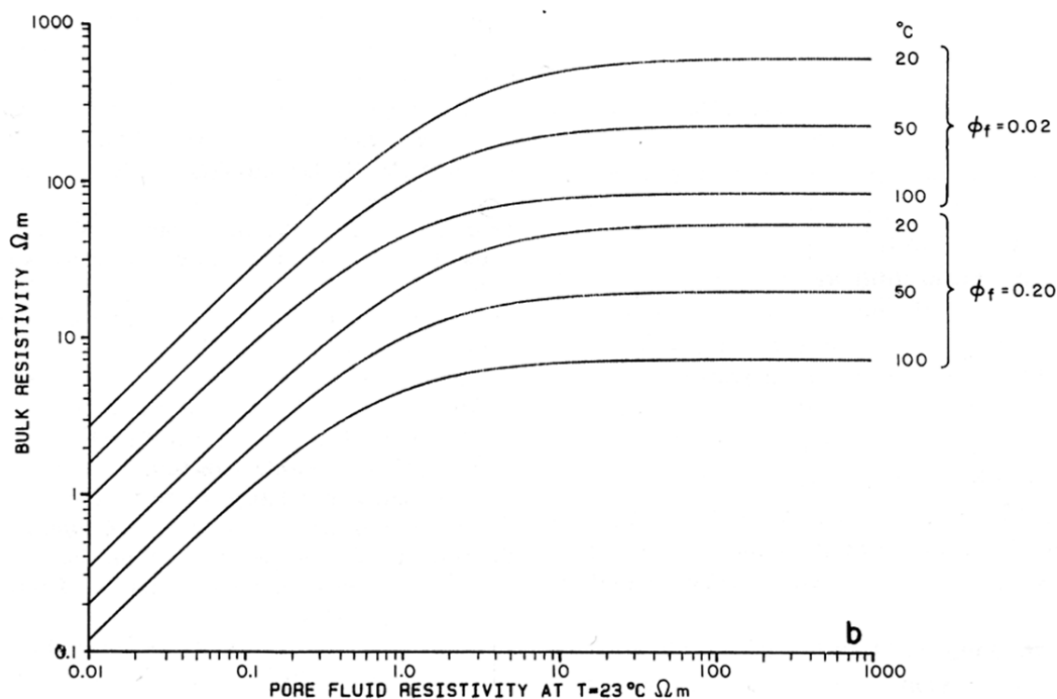


FIGURE 3: Bulk resistivity as a function of pore fluid resistivity for different temperatures and porosities (Flóvenz et al., 1985)

#### 4. RESISTIVITY STRUCTURE OF HIGH TEMPERATURE GEOTHERMAL FIELDS

All high temperature geothermal fields in Iceland and in general where the host rocks are volcanic not sedimentary, have a similar resistivity structure. They are characterized by some kind of a convex structure. At a certain depth, a low resistivity (conductive) cap or zone domes up (the outer margin of the reservoir) and is underlain by higher resistivity, a resistive core (Figure 4). Not so many years ago most scientists found it hard to believe that the resistivity would increase at this depth since the temperature obviously increases with depth. This was a kind of a puzzle and many different theories were put forward. In Iceland it was not until a detailed multi-method DC resistivity survey was conducted in the Nesjavellir area, SW-Iceland in 1985 and 1986 that a reasonable explanation was put forward. The joint 2D interpretation of the different datasets showed a detailed picture of the resistivity structure in the uppermost 1 km under Nesjavellir, SW-Iceland that could be compared to data from nearby boreholes (Árnason et al., 1987a and 1987b).

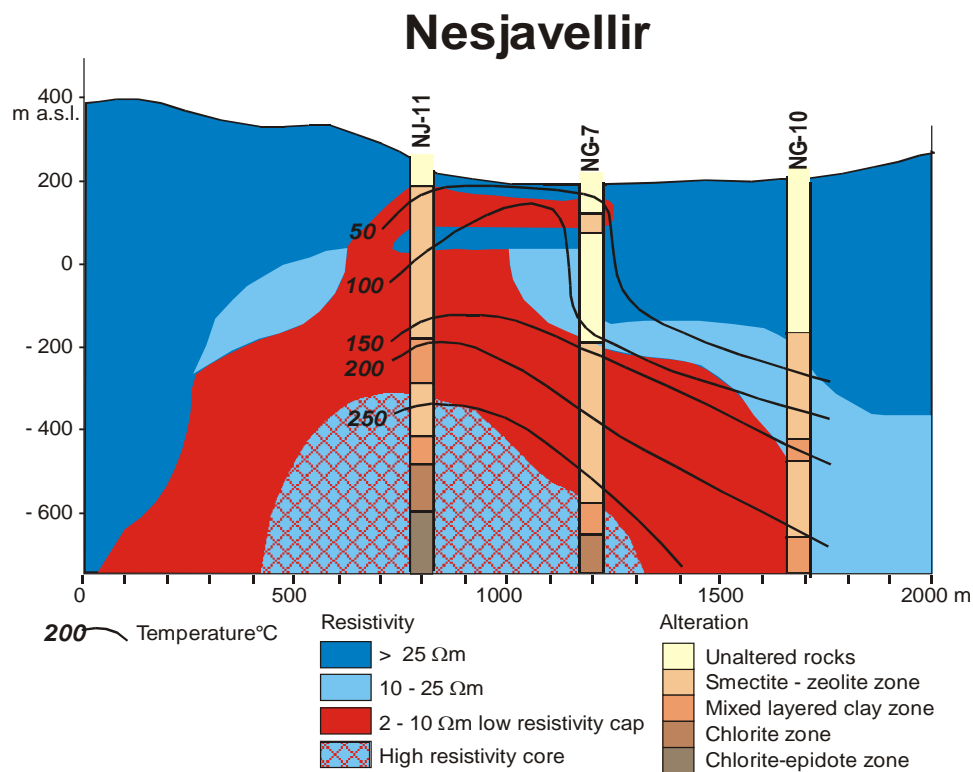


FIGURE 4: Resistivity cross-section from Nesjavellir geothermal field, SW-Iceland, alteration zoning and temperature (Árnason et al., 1987a)

The resistivity structure at Nesjavellir was compared to lithology, alteration mineralogy and temperature. The resistivity was not found to correlate with the lithology, but a fairly clear and consistent correlation was found with the alteration mineralogy. Figure 4 shows a simplified resistivity cross-section from Nesjavellir and the zones of dominant alteration minerals and temperature in nearby wells. The figure shows that the resistivity is high in the surrounding cold unaltered rocks. At temperatures of 50-100°C pronounced geothermal alteration sets in with smectite and zeolites as the dominant alteration minerals and the rocks become conductive. At higher temperature, in the range of 220-240°C, smectite and zeolites are gradually replaced by chlorite as the dominant alteration mineral in the so-called mixed layered clay zone (Kristmannsdóttir, 1979) and the resistivity increases again. At still higher temperatures epidote becomes abundant.

An overview of this relationship between alteration and temperature is given in Figure 5. The smectite and the zeolites minerals have loosely bound cations that make these minerals conductive, while in the

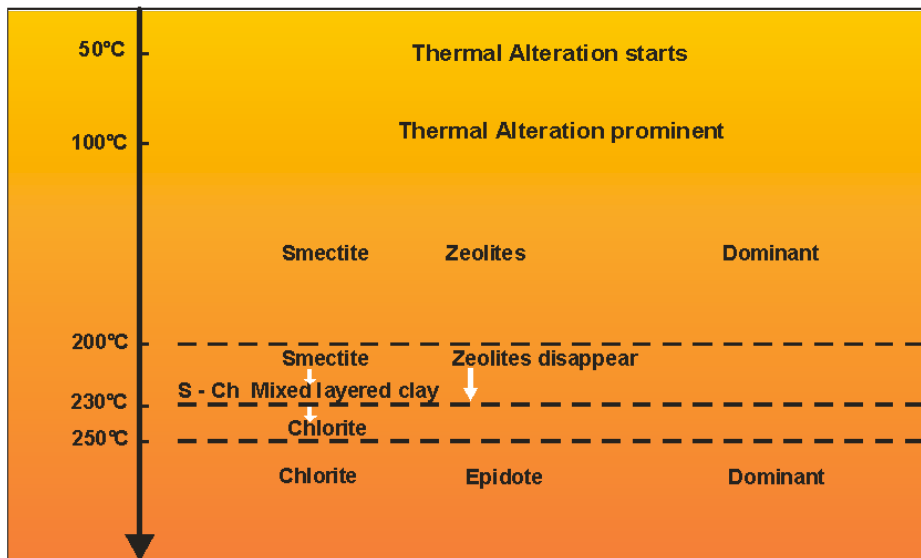


FIGURE 5: Alteration mineralogy and temperature

chlorite mineral these ions are bound in a crystal lattice (Deer et al., 1962), which makes the mineral more resistive. Similar correlation has been found in all the explored and drilled high-temperature geothermal fields in Iceland as well as in many high-temperature areas in other countries (Árnason et al., 2000).

Figure 6 shows measured normalised sample conductivity as a function of pore fluid conductivity for different salinity from different alteration zones in Iceland. This study was done in cooperation between scientist at Iceland GeoSurvey and GFZ in Potsdam, Germany (Flóvenz et al., 2005). For unaltered rocks the conductivity is dominantly pore fluid conductivity for all salinities. On the other hand for smectite altered rocks the figure shows near constant values of conductivity for different salinities (except for one sample), showing dominantly mineral conduction. In the case of chlorite alteration, the

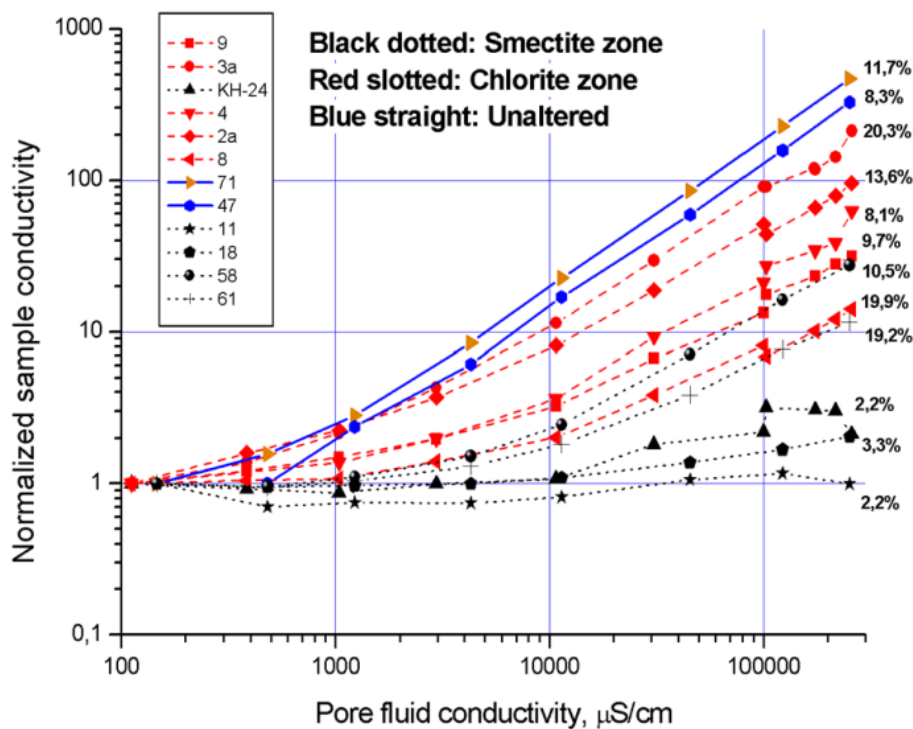


FIGURE 6: Normalized resistivity as a function of saturating fluid conductivities for different alteration zones in Icelandic rocks (Flóvenz et al., 2005)

conductivity curve is between these two, showing dominantly pore fluid and surface conduction. This is also in good agreement with the result in figure 4 showing the correlation between the subsurface resistivity and hydrothermal alteration minerals as revealed in cuttings from boreholes.

Figure 7 shows the general resistivity structure of the basaltic crust in Iceland based on in-situ measurements and laboratory data. The uppermost and unaltered part has relatively high resistivity and the conduction is mainly pore fluid conduction. The resistivity decreases as the smectite-zeolite zone is reached and mineral or surface conduction becomes the dominant conduction mechanism. Because of increasing

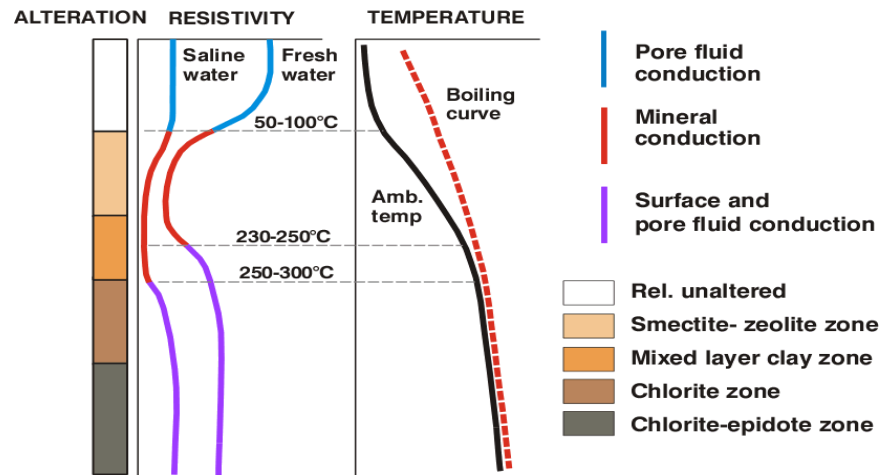


FIGURE 7: The general resistivity structure of the basaltic crust in Iceland summarized (taken from Flóvenz et al., 2005); the depth scale is arbitrary.

temperature and increasing alteration at greater depth, the resistivity decreases furthermore. Below, the mixed clay zone becomes dominant and the resistivity increases again, most likely due to strongly reduced cation exchange capacity of the clay minerals in the mixed clay and chlorite zone. Here the surface and pore fluid conduction probably dominate as the mineral conduction is diminished. The transition from smectite to mixed layer clay happens at a temperature around 230°C. At still greater depths the resistivity decreases slowly which is considered to be caused by slowly increasing temperature with depth in the reservoir.

The subsurface resistivity structure in high temperature geothermal fields therefore reflects the hydrothermal alteration. The primary minerals in the host rock matrix are transformed into different minerals because of water-rock interaction and chemical transport by the geothermal fluids. Alteration minerals depend on the type of primary minerals and the chemical composition of the geothermal fluid but mainly on temperature, but porosity and permeability also control the intensity of the alteration. If the alteration and temperature are in equilibrium the subsurface resistivity structure reflects not only the alteration but also which temperature to expect. This was an important finding, because if the temperature that produced the alteration mineralogy prevails, the resistivity structure can be used to predict the temperature. But if cooling occurs, the alteration remains and so does the resistivity structure. The resistivity structure can therefore in most cases be regarded as a “maximum thermo-meter”. However, it has occurred that alteration minerals have indicated lower temperature than measured in the wells. This has been interpreted as being due to a young system being heated up and the alteration is lagging behind, still not in equilibrium with the temperature.

Figure 8 shows a fine correlation between the subsurface resistivity and alteration zoning for all boreholes on a profile from Krafla geothermal field, NE-Iceland. Correlation with temperature is fair as well except for boreholes KG-10 and KJ-18. It has been proposed that 2000 years ago, a dyke was injected and blocked the flow of geothermal fluid resulting in considerable cooling around well KJ-18 (Sæmundsson, 1991; Árnason and Karlsdóttir, 1996). Therefore great care should be taken when predicting the temperature from the subsurface resistivity structure.

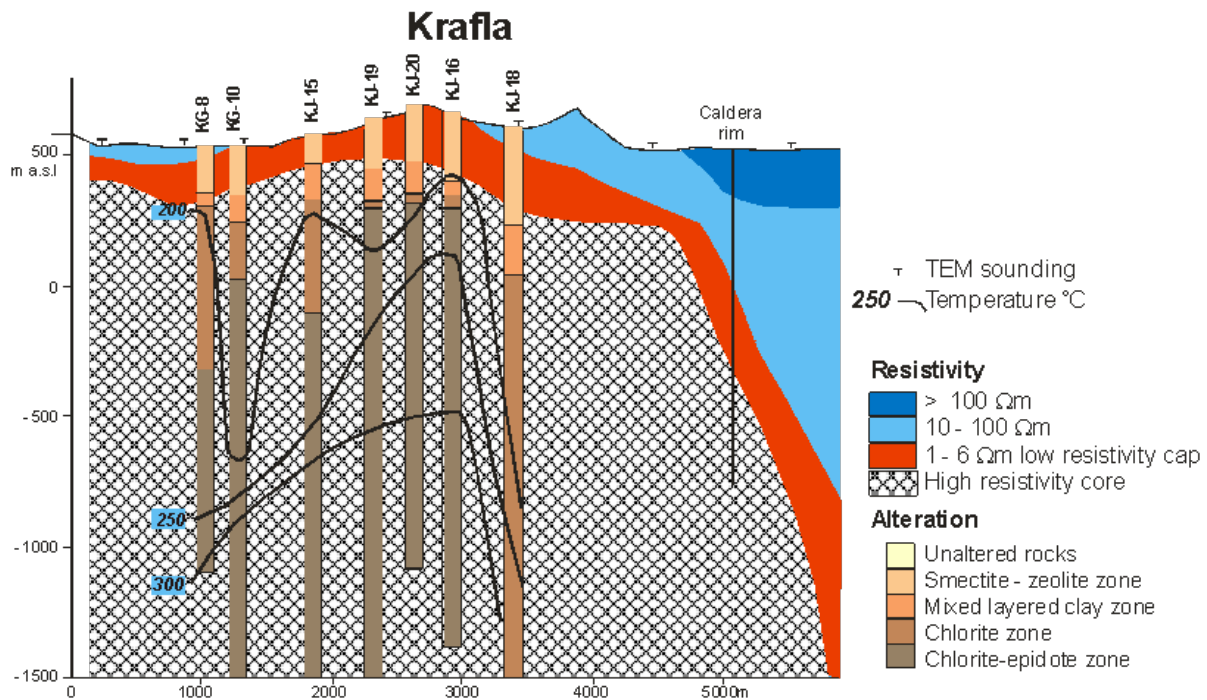


FIGURE 8: An example of cooling - resistivity cross-section from Krafla geothermal field, NE-Iceland, with alteration zoning and temperature (Árnason and Karlsdóttir, 1996)

## REFERENCES

- Archie, G.E., 1942: The electrical resistivity log as an aid in determining some reservoir characteristics. *Tran. AIME*, 146, 54-67.
- Árnason, K., Flóvenz, Ó., Georgsson, L., and Hersir, G.P., 1987a: Resistivity structure of high-temperature geothermal systems in Iceland. *International Union of Geodesy and Geophysics (IUGG) XIX General Assembly*, Vancouver, Canada, abstract V, 477.
- Árnason, K., Haraldsson, G.I., Johnsen, G.V., Thorbergsson, G., Hersir, G.P., Saemundsson, K., Georgsson, L.S., Rögnvaldsson, S.Th., and Snorrason, S.P., 1987b: *Nesjavellir-Ölkelduháls, surface exploration 1986*. Orkustofnun, Reykjavík, report OS-87018/JHD-02 (in Icelandic), 112 pp + maps.
- Árnason, K. and Karlsdóttir, R., 1996: *A TEM resistivity survey of the Krafla high-temperature field*. Orkustofnun, Reykjavík, report OS-96005/JHD-03 (in Icelandic), 96 pp.
- Árnason, K., Karlsdóttir, R., Eysteinnsson, H., Flóvenz, Ó.G., and Gudlaugsson, S.Th., 2000: The resistivity structure of high-temperature geothermal systems in Iceland. *Proceedings of the World Geothermal Congress 2000*, Kyushu-Tohoku, Japan, 923-928.
- Dakhnov, V.N., 1962: Geophysical well logging. *Q. Colorado Sch. Mines*, 57-2, 445 pp.
- Deer, W.A., Howie, R.A., and Zussman, J., 1962: *Rock-forming minerals, vol. 3. Sheet silicates*. Longmans, Green and Co. Ltd., London, 270 pp.
- Flóvenz, Ó.G., Georgsson, L.S., and Árnason, K., 1985: Resistivity structure of the upper crust in Iceland, *J. Geophys. Res.*, 90-B12, 10136-10150.



Flóvenz, Ó.G., Spangenberg, E., Kulenkampff, J., Árnason, K., Karlsdóttir, R., and Huenges E., 2005: The role of electrical conduction in geothermal exploration. *Proceedings World Geothermal Congress 2005*, Antalya, Turkey, 9 pp.

Hersir, G.P., and Björnsson, A., 1991: *Geophysical exploration for geothermal resources. Principles and applications*. United Nations University Geothermal Training Programme, Reykjavík, Iceland, report 15, 94 pp.

Kristmannsdóttir, H., 1979: Alteration of basaltic rocks by hydrothermal activity at 100-300°C. In: Mortland, M.M., and Farmer, V.C. (editors), *International Clay Conference 1978*. Elsevier Scientific Publishing Co., Amsterdam, 359-367.

Quist, A.S., and Marshall, W.L., 1968: Electrical conductances of aqueous sodium chloride solutions from 0 to 800°C and at pressures to 4000 bars. *J. Phys. Chem.*, 72, 684-703.

Saemundsson, K., 1991: The geology of the Krafla system. In: Gardarsson, A., and Einarsson, Á. (editors), *The nature of Lake Mývatn* (in Icelandic). The Icelandic Society of Natural Sciences, Reykjavík, 25-95.