



UNITED NATIONS
UNIVERSITY

GEOHERMAL TRAINING PROGRAMME



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THE RESISTIVITY STRUCTURE OF HIGH-TEMPERATURE GEOHERMAL SYSTEMS IN ICELAND

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ABSTRACT

Electrical and electro-magnetic methods have been used extensively to identify and delineate high-temperature geothermal reservoirs in Iceland. All high-temperature systems, within the basaltic crust in Iceland, have a similar resistivity structure, characterized by a low resistivity cap at the outer margins of the reservoir, underlain by a more resistive core towards the inner part. This is found in fresh-water systems as well as brine systems, with the same character but lower resistivities in the brine systems. Comparison of this resistivity structure with data from wells shows a good correlation with alteration mineralogy. The low resistivity in the low-resistivity cap is dominated by conductive minerals in the smectite-zeolite zone in the temperature range of 100-220°C. At temperatures 220-240°C zeolites disappear and the smectite is gradually replaced by the resistive chlorite. At temperatures exceeding 250°C chlorite and epidote are the dominant minerals and the resistivity is probably dominated by the pore fluid conduction in the high-resistivity core. The important consequence of this is that the observed resistivity structure can be interpreted in terms of temperature distribution. A similar resistivity structure is to be expected in acidic rocks. Due to different alteration mineralogy, however, the transition from the conductive cap to the more resistive core presumably occurs at temperatures lower than 200°C.

1. INTRODUCTION

Resistivity methods have been used in geothermal surveying for decades in Iceland. From the mid sixties, DC-methods, mostly Schlumberger soundings, were used to identify and delineate high-temperature systems. In the mid eighties the DC methods were succeeded by central-loop TEM-soundings (Transient Electro-Magnetic). The TEM-soundings have proven to be more downward focused and have better resolution at depth than the DC-methods

As geothermal exploration progressed and resistivity data were obtained from different geothermal fields, a characteristic resistivity structure of the high-temperature geothermal system started to emerge. The fields have a distinctive low resistivity zone at their outer margins which is underlain by higher resistivity towards the interior of the reservoir. This resistivity structure was found to

contradict the conceptual model that the resistivity should generally decrease with increasing temperature. As data from geothermal wells became more abundant, a comparison of the resistivity structure with geological and geophysical well data, was made possible. This comparison showed that the resistivity structure could be correlated to the alteration mineralogy, which on the other hand, basically reflects the thermal conditions in the geothermal system.

The idea, that the resistivity was affected by the alteration came up within the geophysical group at Orkustofnun in the early eighties and was first presented in a workshop within the Geological Society of Iceland in Mars 1982. Increasing data have supported and clarified these observations (Arnason et al., 1987; Arnason and Flovenz, 1992). Recently, similar observations have been made in Japan (Uchida, 1995). In this paper we review and present several examples of this resistivity structure and discuss its causes and implications.

2. RESISTIVITY OF ROCKS

The resistivity of water-saturated rocks is in general dependent on many physical parameters such as porosity, the salinity of the saturating fluid, temperature, conductivity of the rock matrix, and thermal alteration. The interplay of these parameters is quite complex and in some respects not fully comprehended. Empirical equations have been proposed as to the influence of the different parameters, but they are usually based on measurements of resistivity in rock samples under different and often simplified conditions. A general formula describing the resistivity of saturated rocks is bound to have many free parameters in order to account for the different factors affecting the resistivity. The compilation of such a general formula is further hampered by the difficulty in controlling individual factors in a reproducible manner.

Many useful simplified formulas do exist, that can be good approximations under certain conditions. The most simple, and probably the most widely used, is Archie's law (Archie, 1942):

$$\rho = \rho_w \cdot a \cdot \Phi^{-m} \quad 1)$$

where ρ , ρ_w and Φ are the bulk resistivity, the resistivity of the saturating fluid, and porosity, respectively, and a and m are empirical coefficients. This formula seems to be a fairly good approximation when the conductivity is dominated by the saturating fluid. The empirical coefficients a and m are usually reported around 1 and 2, respectively. Another useful, but simplified, formula was put forward by Rink and Schopper (1976), where they, in addition to pore fluid conduction, as described by Archie's law, include interface conduction.

Flovenz et al. (1985) explored the relationship between the bulk resistivity, fluid resistivity, porosity, and temperature, for rocks in the uppermost kilometre of the Icelandic crust, outside the volcanic zones. From field data and core-sample measurements, they compiled a semi-empirical relation, based on the so-called double porosity model (Stefansson et al., 1982). They found that for rocks, saturated with fluids with resistivity higher than about 2 Ωm , at room temperature, the bulk resistivity is practically independent of the resistivity of the fluid, but dependent on porosity and temperature. Flovenz et al. concluded that, except for rocks saturated with highly saline waters (sea-water), electrical conduction in the Icelandic crust, outside the volcanic zones, is mainly controlled by alteration minerals (clay minerals and zeolites).

3. THE RESISTIVITY STRUCTURE REVEALED

The first application of the resistivity method on high-temperature fields in Iceland was carried out in the early seventies on Reykjanes geothermal field (Bjornsson et al., 1972), and Krisuvik geothermal field (Arnorsson et al., 1975).

The first large-scale resistivity survey for a high-temperature geothermal exploration was performed in the early seventies in the Krafla geothermal field, NE Iceland (Karlisdottir et al. 1978). DC-methods, mainly Schlumberger soundings, were applied. A well defined low resistivity anomaly was detected in relatively resistive host-rocks. All soundings within the low-resistivity anomaly showed, however, increasing apparent resistivity in the datapoints for the largest electrode spacing. This was found to be in contradiction with the conception that the resistivity should generally decrease with increasing temperature at depth. Attempts were made to explain the increasing apparent resistivity by lateral resistivity variations or electromagnetic effects. These possibilities were ruled out by theoretical calculations and field tests, showing, that the resistivity did indeed increase again below a relatively thin low-resistivity anomaly.

The next extensive resistivity survey was a reconnaissance survey in the Svartsengi-Eldvorp-Reykjanes geothermal fields on the outer part of the Reykjanes peninsula, SW Iceland, using Schlumberger soundings. The outer peninsula is penetrated by sea-water and the geothermal systems were manifested by low resistivity anomalies (0.5–2 Ω m) in a relatively conductive host rock (5–10 Ω m) (Georgsson and Tulinius; 1983). No clear evidence was found indicating increasing resistivity towards the centre of the geothermal systems as was observed in the fresh-water saturated geothermal system in Krafla.

In the years of 1985 to 1987 a detailed resistivity survey was carried out in Nesjavellir geothermal field, in a NNE-SSW trending fissure swarm, north of the Hengill central volcano in SW Iceland (Arnason et al.; 1986, 1987). DC-resistivity methods, both Schlumberger soundings and half-Schlumberger head-on resistivity profiling, were used to collect large data sets on several profiles, designed for a joint 2D-modelling of the Schlumberger and head-on data. The 2D-modelling resulted in highly constrained and detailed resistivity sections through the uppermost one kilometer of the reservoir. The models showed a well defined low resistivity layer of 3–5 Ω m on the outer margins of the reservoir, and underlain by, about an order of magnitude higher resistivity deeper in the geothermal system..

At that time, numerous wells had already been drilled into the Nesjavellir geothermal system, and abundant geological and geophysical data were available. The resistivity model for each section was compared to geological and geophysical data from nearby wells (within 100m from the profile). No obvious correlation was observed between lithology and resistivity. A good and clear correlation was, on the other hand, found between the alteration mineralogy and resistivity. Comparison with porosity logs did not reveal any obvious correlation. The porosity is, however, strongly correlated to lithology, as is to be expected.

The resistivity models correlated well with the resistivity logs in the shallower parts of the wells, but at depth, the resistivity logs had qualitatively similar pattern as the 2D models but showed considerably lower values. This was at first considered to mean that the increasing resistivity towards the inner part of the reservoir was due to partial boiling. This was later ruled out, based on pressure- and temperature logs. A closer inspection of the resistivity logs showed some inconsistencies in the data and that the resistivity logs only gave relative information at depth. The problem turned out to be an instrumental problem of the resistivity-logging tool. This was later solved, and the discrepancies were no longer present.

Figure 1 shows a smoothed 2D model for one of the profiles from Nesjavellir, perpendicular to the fissure swarm (so 2D assumption is well justified). A clear resistivity anomaly is seen, with a cap of resistivities of the order of $5\Omega\text{m}$ at the margins and higher resistivity deeper in the reservoir. The reservoir is confined by dykes and faults in the fissure swarm and has very sharp near-vertical boundaries and some lateral flow near the surface. Three wells are close to the profile. On Figure 1, the wells are projected onto the section, showing the zones of dominant alteration minerals. Formation temperature isotherms, based on temperature logs from the wells are also shown. The figure shows very good correlation between the resistivity and temperature. The resistivity is high in the cold, unaltered rocks outside the reservoir and decreases strongly at the onset of geothermal alteration, in the smectite-zeolite zone, when the temperature has reached about 100°C . It is low, generally lower than $5\Omega\text{m}$, down to the mixed-layered clay zone, where it increases considerably again and stays relatively high in the chlorite and chlorite-epidote zones at temperatures exceeding 250°C .

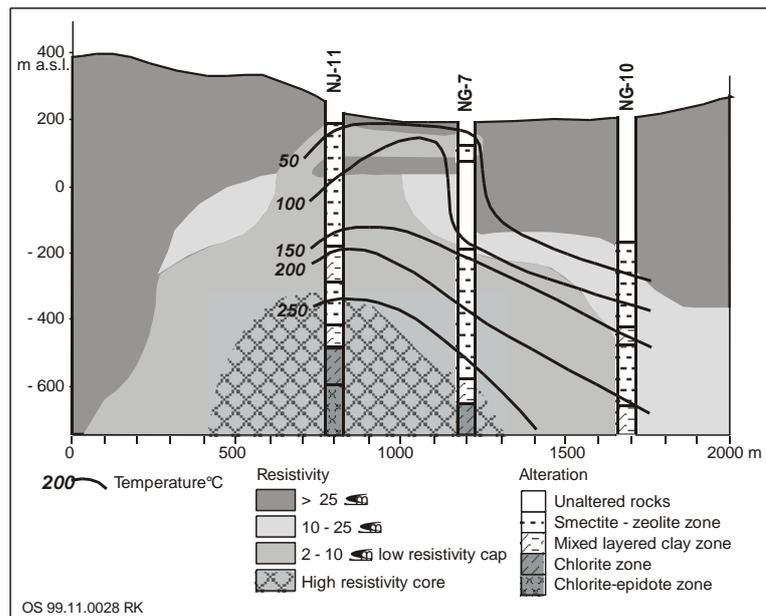


FIGURE 1: Resistivity cross-section from Nesjavellir geothermal field, alteration zoning in wells and temperature

Shortly after the survey in the Nesjavellir field, the DC-methods were succeeded by central-loop TEM-soundings. Since that time several resistivity surveys have been carried out at various fresh-water saturated high-temperature geothermal fields. During the same period, several wells have been drilled and more well data have become available for comparison with the resistivity structure. All these surveys revealed basically the same general resistivity structure. In all cases where well data are available for comparison, the resistivity structure correlates with the alteration mineralogy, but no obvious correlation is found with lithology. A good correlation is generally found between resistivity and temperature, but with some important exceptions. These exceptions are found where parts of the reservoir have recently been cooled down and the alteration mineralogy is no longer in equilibrium with the temperature.

Figure 2 shows a simplified resistivity cross-section through the eastern part of the Krafla geothermal system, NE Iceland, based on detailed central-loop TEM survey (Arnason and Karlsdottir, 1996). The zoning of alteration mineralogy in nearby wells and estimated reservoir isotherms has been projected onto the cross-section. The figure shows a very consistent correlation between the alteration mineral zones and resistivity. The low-resistivity cap (resistivity lower

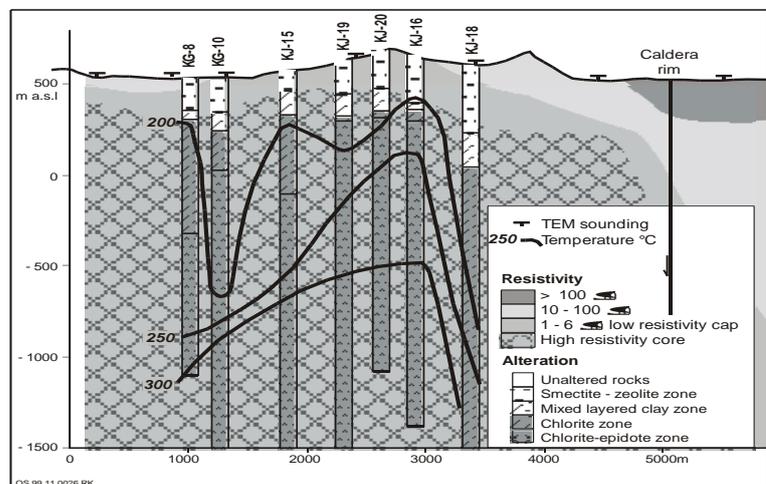


FIGURE 2: Resistivity cross-section from Krafla geothermal field, alteration zoning and temperature

than $10\Omega\text{m}$) coincides with the smectite-zeolite zone, which extends to the surface in the well field, and the increase in resistivity with depth very consistently follows the top of the mixed layered clay zone. The correlation with temperature is, however, not as good. It is evident from Figure 2 that the alteration mineralogy is not in equilibrium with present temperature in the system (the relation between reservoir temperature and alteration mineralogy of basaltic rocks will be summarised in a later section). Two distinct anomalies in the temperature-alteration relation are found, i.e. in well KG-10 in the western part of the section and in well KJ-18 in the eastern. The Krafla geothermal system has undergone several phases with heating and recooling in different parts of the reservoir. The hypothesis has been put forward (Saemundsson, 1991; Árnason and Karlsdóttir, 1996) that 2000 years ago, a dyke was injected and blocked the flow of geothermal fluid from west, resulting in considerable cooling around well KJ-18. A similar phenomenon was observed in the Nesjavellir geothermal system.

In 1988, Orkustofnun conducted a resistivity survey in the Asal Rift, Djibouti, East Africa, using central-loop TEM-soundings (Árnason and Flovenz, 1995). The Asal Rift is an active spreading zone with basaltic volcanism and hosts a geothermal system with highly saline fluids. The TEM-soundings show high resistivity from the surface and down to about 100m above the water-table, where the resistivity drops below $2\text{-}12\Omega\text{m}$, which was explained by partial saturation. At the water table the resistivity decreased further, but in most of the soundings the resistivity increased again at depth. Well data was sparse from the survey area, but comparison could be made with data from two wells. In both cases a distinctive lowering the resistivity coincided with the water table, as was to be expected, but the increasing resistivity with depth coincided with the top of the chlorite zone in both of the wells. Figure 3 shows the comparison for the well Asal-4 and the nearby TEM-sounding DJ-11. Due to the high salinity brine, the electrical conduction was thought to be dominant in the saturating fluid, and that resistivity variations were mainly due to differences in saturation and porosity as described by Archie's law. The increased resistivity at the top of the chlorite zone was therefore thought to reflect decreasing porosity due to mineral precipitation.

In 1996 and 1997 a second resistivity survey was carried out on the outer part of the Reykjanes peninsula. This time the central-loop TEM method was applied. The result of this survey clearly demonstrated that the central-loop TEM method has much better resolution and more penetration depth than the DC-method. The TEM data revealed a clear resistivity image of the brine high-temperature geothermal systems in the peninsula (Svartsengi-Eldvorp and Reykjanes). The surrounding rocks have resistivity of the order of $5\text{-}15\Omega\text{m}$ and the geothermal systems appear as a low-resistivity cap, with resistivities ranging from $0.5\text{-}3\Omega\text{m}$, with an underlying high-resistivity core with resistivities in the range of $7\text{-}15\Omega\text{m}$ (Karlsdóttir, 1997, 1998). This is clearly seen in Figure 4, which shows a resistivity section from the Reykjanes in the west and to Eldvorp and Svartsengi in the east. Several wells have been drilled into the geothermal systems and the zones of dominant alteration minerals have been projected onto the section. Here again the resistivity layering shows an obvious correlation with the alteration mineralogy, but no obvious correlation was found with lithology. Figure 5 shows the alteration zones, lithological and resistivity logs from well SJ-18 in the Svartsengi geothermal field as well as a resistivity model from a nearby TEM-resistivity section.

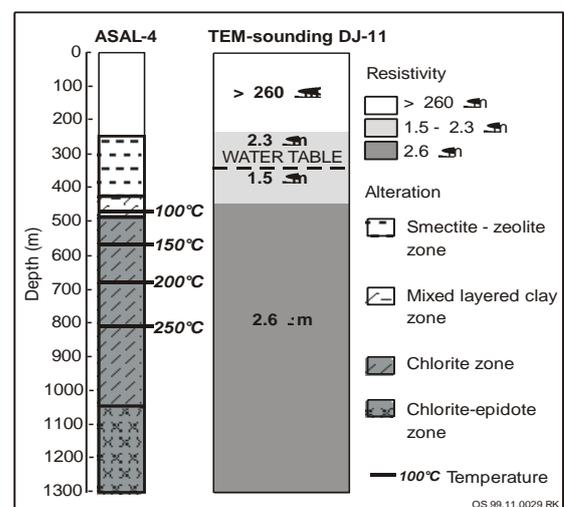


FIGURE 3: Alteration and temperature in well Asal-4 in Djibouti and resistivity from a nearby TEM-sounding

Figure 4 shows roughly the same correlation between the resistivity and alteration minerals as in the fresh-water systems, but there are some minor differences. In the fresh-water systems, the boundary between the low resistivity cap and the resistive core correlates with the boundary between the smectite-zeolite zone and the mixed-layered clay zone. According to Figures 4 and 5, this boundary seems to be within the mixed-layered clay zone. Well EG-2 stands out, indicating that the low resistivity cap is well within the chlorite zone. This is not significant because the well is to the side of the section, where the resistivity is steeply dipping, perpendicular to the section. The survey on the Reykjanes peninsula therefore indicates a slightly different correlation between alteration and resistivity. The correlation found in the highly saline system in the Asal Rift (Figure 3) supports the hypothesis that the transition from the low-resistivity cap to the resistive core is moved towards the chlorite zone. This can possibly be explained by slightly different alteration zoning in the saline systems.

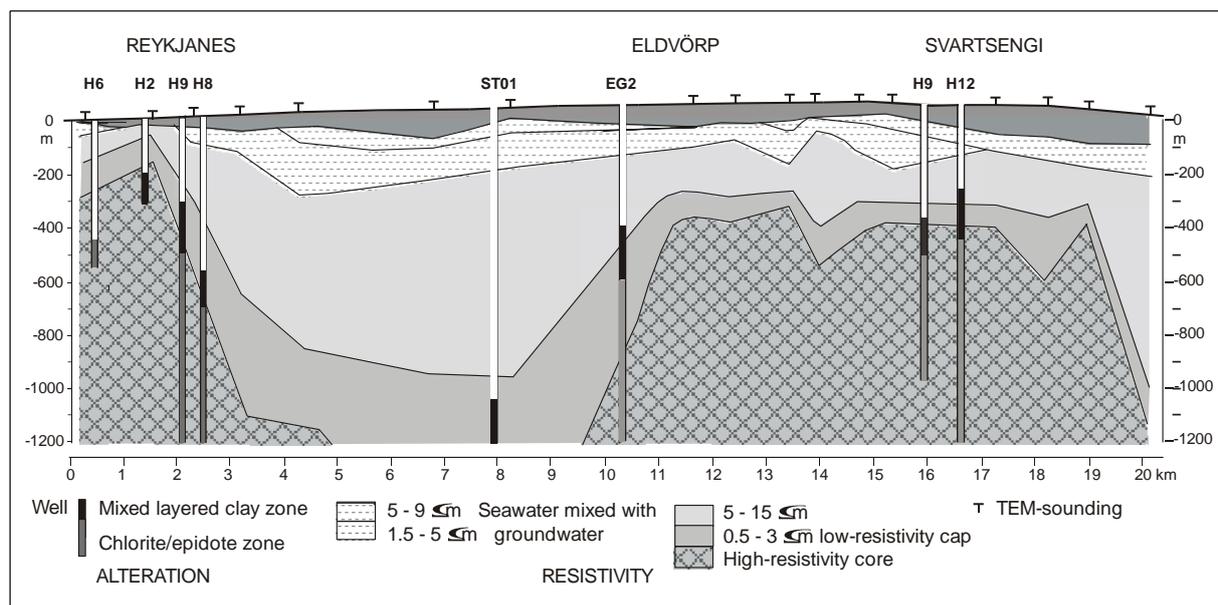


FIGURE 4: Resistivity cross-section from Reykjanes peninsula and alteration zoning in wells

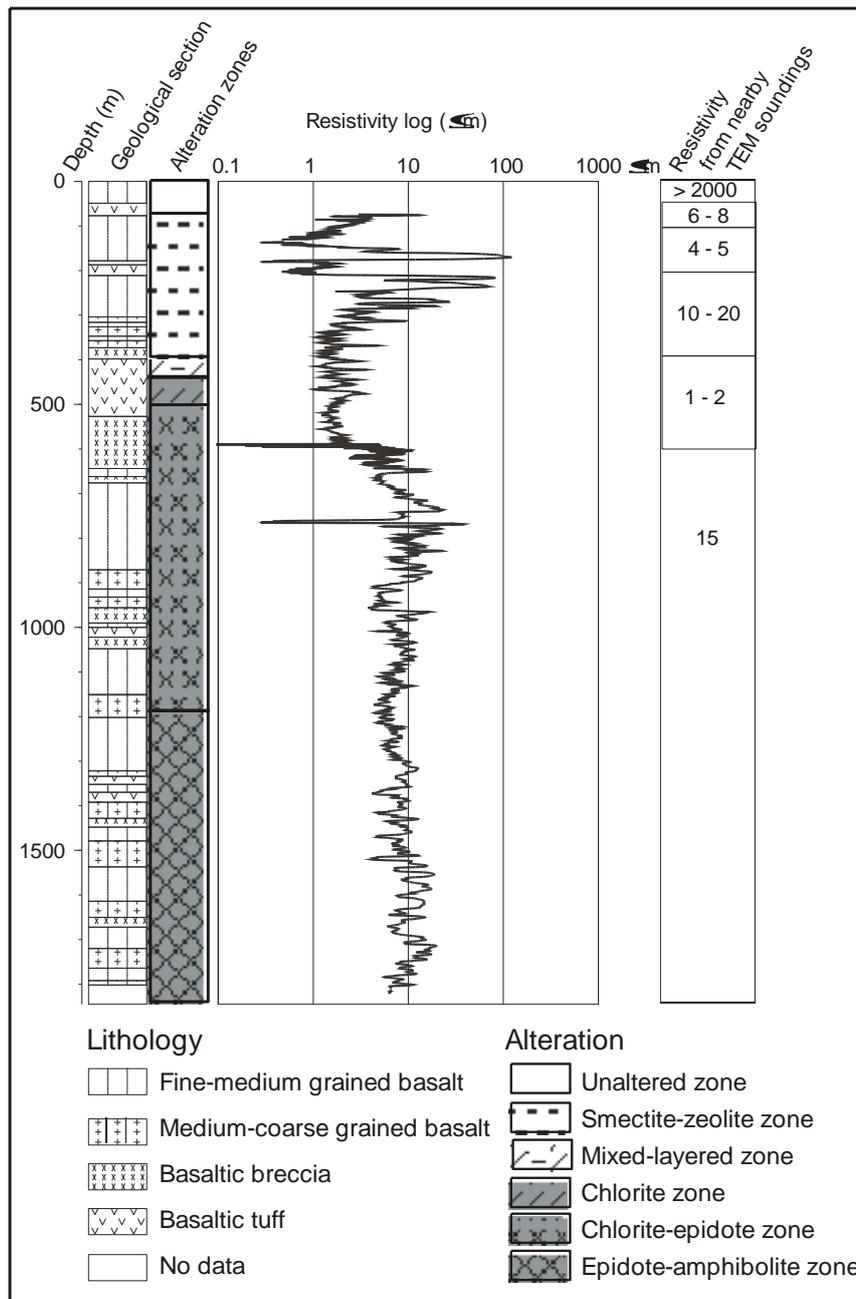


FIGURE 5: Geological section, alteration zoning, resistivity logs in well SJ-18 in Svartsengi and resistivity from nearby TEM-soundings

4. ALTERATION OF BASALTIC ROCKS

Due to water-rock interaction and chemical transport by the geothermal fluids, the primary minerals in the host rock matrix are transformed, or altered, into different minerals. The alteration process and the resulting type of alteration minerals are dependent on the type of primary minerals, chemical composition of the geothermal fluid and temperature. The intensity of the alteration is furthermore dependent on the temperature, but also on time and the texture of the host rocks. The alteration process and the resulting alteration mineralogy of the basaltic rocks in high-temperature geothermal systems in Iceland have been studied quite extensively (Kristmannsdóttir, 1979). The primary mineralogy of the basaltic host rocks in the volcanic zones of Iceland is relatively homogeneous. The geothermal fluids can be divided into two relatively homogeneous types, i.e. low salinity or fresh-

water and saline fluids. Due to this homogeneity, the stability and formation of alteration minerals is mainly dependent on temperature.

There is no room here for a lengthy discussion so we will only discuss the main features and the dominant minerals or mineral classes. At temperatures lower than 220°C, low-temperature zeolites and the clay mineral smectite are formed. The alteration intensity is normally low for temperatures below 50-100°C. The range where low temperature zeolites and smectite are abundant is called the smectite-zeolite zone. In the temperature range from 220°C to about 240-250°C, the low temperature zeolites disappear and the smectite is transformed into chlorite in a transition zone, the so-called mixed layered clay zone, where smectite and chlorite coexist in a mixture. At about 250°C the smectite has disappeared and chlorite is the dominant mineral, marking the beginning of the chlorite zone. At still higher temperatures, about 260-270°C, epidote becomes abundant in the so-called chlorite-epidote zone. This zoning applies for fresh water systems. In brine systems, the zoning is similar but the mixed layered clay zone extends over a wider temperature range or up to temperatures near 300°C.

A similar alteration zoning, dependent on temperature, is observed in geothermal systems in acidic rocks, but with somewhat different alteration minerals. (Kristmannsdottir, 1985).

This correspondence of different stable and dominant alteration minerals with different temperatures is used extensively in geothermal exploration and drilling. Analyses of drill-cuttings during drilling are used to estimate the unperturbed formation temperature. Comparison of estimated formation temperature from temperature logs, and the alteration mineralogy, can be used to tell whether present temperatures are in equilibrium with the alteration, or if cooling or heating has recently occurred. Such a comparison clearly shows that the rocks around wells KJ-18 and KG-10 on Figure 2 have recently been cooled.

5. CONDUCTION MECHANISMS

Although the relevant conduction mechanisms behind the observed resistivity structure of the high-temperature geothermal systems are not known in details, they can be qualitatively understood in terms of the structural and physical properties of the different dominant alteration minerals.

If pore fluid is the dominant conductor, measured resistivities of the geothermal fluids of the fresh-water systems (10-15Ωm at room temperature) and reasonable values for porosity (10-15%) and for the coefficients, a and m , in Archie's law (eq. 1), give the resistivity in the range of 15 to 50Ωm for temperatures in the range of 200 to 250°C. This is higher than the observed resistivity in the low-resistivity cap, by a factor of 3 to 10, showing that a different conduction mechanism is dominant. The smectite clay mineral is an obvious candidate. The smectite and chlorite minerals are closely related. Both are so-called sheet silicates (Deer *et al.*, 1962). Smectite has hydrated and loosely bound cations between the silica plates, making the mineral conductive and with a high cation exchange capacity. In the chlorite mineral the cations are on the other hand fixed in a crystal lattice, making the mineral resistive.

In the case of fresh-water geothermal systems, the conduction in the surrounding rocks is probably due to low intensity and low temperature alteration minerals and/or pore fluid conduction in very fresh rocks. In the low resistivity cap the conduction is dominated by the highly conductive alteration minerals. The above estimate for the contribution of pore fluid conduction at high temperatures roughly agrees with the values found in the resistive core, indicating that the pore fluid conduction is dominant.

In the saline systems the resistive inner core has resistivity of the order of $10\Omega\text{m}$. If it is assumed that this is, like in the case of fresh-water systems, mainly due to porefluid conduction, it must follow, that the alteration minerals contribute significantly to the conductivity in the low resistivity cap.

6. RESISTIVITY AS A THERMOMETER

The correlation between the resistivity structure of high-temperature geothermal systems in basaltic rocks and alteration mineralogy can be summarised as follows: The resistivity is relatively high in cold unaltered rocks outside the reservoir. The smectite-zeolite zone forms a low resistivity cap on the outer margins of the reservoir. The resistivity increases again towards the interior of the reservoir at the top of, or within, the mixed layer clay zone.

This observation is of great importance, because the temperature dependence of the alteration mineralogy makes it possible to interpret the resistivity layering in terms of temperature, provided that the temperature is in equilibrium with the dominant alteration. The upper boundary of the low-resistivity cap corresponds to temperatures in the range of $50\text{-}100^\circ\text{C}$, depending on the intensity of the alteration. The transition from the low resistivity cap to the resistive core corresponds to temperature in the range of $230\text{-}250^\circ\text{C}$. Thus, if alteration is in equilibrium with temperature, the mapping of the resistivity structure is in fact mapping of isotherms.

It is evident from Figure 2, that the resistivity reflects the alteration, but not the present temperature, if cooling has recently taken place. In this case the resistivity is to be considered as a maximum thermometer. The dominant high temperature minerals, like chlorite and epidote, are stable at lower temperatures and do not degenerate to lower temperature minerals. Under prolonged cooling, higher temperature alteration can, however, probably be so heavily overprinted by low-temperature minerals that the resistivity reflects the new thermal conditions, but no data exist, at present, which conclusively confirm this. If the reservoir, or parts of it, is heated up, lower temperature minerals like smectite can transform to chlorite and mixed layered clays, and it is believed that the dominant alteration and the resistivity can adjust relatively quickly to increased temperatures.

In acidic rocks, a structural transition of dominant alteration minerals, similar to the smectite-chlorite transition in basaltic rocks, occurs with temperature, but at temperatures lower than 200°C (Kristmannsdottir, 1985). A transition from the low resistivity cap to a more resistive core is therefore expected in acidic rocks, at lower temperatures than in basaltic rocks.

7. CONCLUSIONS

Surface resistivity surveys of high-temperature geothermal systems in the basaltic rocks of the volcanic zones of Iceland always seem to reveal basically the same resistivity structure. A low resistivity cap is observed on the outer margins of the reservoirs and underlain by a more resistive core. Extensive comparison of this resistivity structure to well data has revealed a consistent correlation to the zones of dominant alteration minerals, where the low-resistivity cap coincides with the smectite-zeolite zone and the transition to the more resistive core occurs at the boundary, or within the mixed layer clay zone. The alteration mineralogy is, on the other hand, mostly predicted by temperature. This has the important consequence that, the resistivity structure can be interpreted directly in terms of temperature, if the alteration is in equilibrium with temperature. The upper boundary of the low-resistivity cap is found where the temperature is in the range of $50\text{-}100^\circ\text{C}$ and the transition to the resistive core occurs at temperatures in the range of $230\text{-}250^\circ\text{C}$.

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