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The Role of Electrical Interface Conduction in Geothermal Exploration

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ABSTRACT

In order to investigate the interface conduction in the basaltic rocks of the high temperature fields in Iceland we measured the electrical conductivity for the frequency range 0.1 – 100000 Hz versus pore fluid salinity of 12 selected samples of basaltic material from Iceland. These included 2 fresh and completely unaltered samples of recent basaltic lava, 5 samples of basalt and hyaloclastite from the smectite alteration zone and 5 from the chlorite zone.

About 2-5% reduction in conductivity is observed per decade in frequency. For the unaltered samples a linear relationship is found between the bulk conductivity and the pore fluid conductivity over almost the whole range of salinities, showing that the pore fluid conduction is always dominant and the interface conduction is negligible. The samples from the smectite zone show almost no dependence on the pore fluid salinity but considerable interface conduction as predicted, the value being from 20-300 μS/cm with the isoelectrical point at fluid conductivity in the range of 4000 – 6000 μS/cm.

In contrast to previous hypothesis, the samples from the chlorite zone show also significant interface conduction, 4 out of 5 samples show value in the range of 10-30 μS/cm but the isoelectrical point is lower than in the smectite zone, usually at fluid conductivity in the range 1000-3000 μS/cm.

Since the temperature dependence of conductivity is at least twice as high for the interface conduction as for the pore fluid conduction, our results imply that interface conduction is the dominant conduction mechanism for most high temperature geothermal fields regardless of their pore fluid salinity. Furthermore, the observed change in conductivity at the top of the chlorite zone is not due to change in dominant conduction mechanism, i.e. from interface conduction to pore fluid conduction, as has previously suggested, but probably rather due to reduced degree of interface conduction in the chlorite zone associated with the lower cation exchange capacity of chlorite compared to smectite.

As a consequence of this we present a revised version of the model for the electrical resistivity of the basaltic upper crust in Iceland.

1. INTRODUCTION

Resistivity soundings are one of the best prospecting tools for geothermal reservoirs. This is because of the high dependence of the electrical conductivity on physical parameters like temperature, porosity, pore fluid salinity, fluid saturation and the degree of interface conductivity. The last property depends on the stage and amount of the hydrothermal alteration of the rocks; the dominant actors being on the cation exchange capacity (CEC) of the alteration minerals involved (e.g. Waxman and Smiths, 1968, Pezard, 1990). These parameters affect the conductivity in different and complicated ways. In order to be able to use the observed resistivity structure of the earth to conclude about the geothermal parameters involved it is necessary to understand the dependence of the in-situ conductivity on these various parameters. This has been done mainly by three methods, in-situ observations (e.g. Flóvenz et al, 1985), laboratory measurements (e.g. Pezard 1990) and theoretical considerations (e.g. Bussian, 1983 and Revil and Glover, 1997).

High temperature fields in volcanic rocks are usually characterized by a low resistivity cap surrounding a resistive core (Árnason and Flóvenz, 1992, Árnason et al, 2000, Ussher et al. 2000). In active high temperature fields in Iceland the top of the resistive core has been found to correlate with a change in alteration mineralogy from smectites to mixed clays and chlorites (Árnason and Flóvenz, 1992, Árnason et al 2000). It has been postulated that this resistivity change is due to change from high interface conduction in the smectite zone to moderate pore fluid conduction in the mixed clay or chlorite zone. This change seems to occur at approximately 230°C. Thus, by mapping the surface of the high resistive core it is possible to map the 230°C isothermal surface. Since reservoir temperatures above 230°C are well suited for electricity production surface exploration with geoelectrical soundings is able to detect those areas. The disadvantage of this method is that the resistivity increase is frozen-in and does not change when the system cools down. Thus the top of a high resistive core can only be interpreted as having been exposed to 230°C during some period in the thermal history of the geothermal system. This hypothesis is based on large-scale field observations in Iceland but needs direct support from laboratory measurements. The purpose of the present work was to provide better insight into the processes by measuring the conductivity of selected samples of volcanic rocks from geothermal systems in Iceland as a function of the pore fluid salinity. Simultaneously the samples from the same wells were used to measure the temperature dependence of rock conductivity up to 230°C and those results are described in another paper (Kulenkampff et al, 2005).

3. BACKGROUND MATERIAL AND THEORY

Already in the middle of the twentieth century scientists became aware of the importance of clay minerals in electrical conductivity of porous rocks (e.g. Windsauer and McCardell, 1953, Hill and Milburn, 1956). Measurements of electrical conductivity in shaly sand showed peculiar behavior of the relationship between the bulk conductivity of the sample and the pore fluid conductivity. Waxman and Smith (1968) proposed a parallel resistor model to explain the behavior. The equation

$$\sigma = \frac{\sigma_w}{F} + \sigma_s \qquad (1)$$

where σ, σ_w σ_s and denote the bulk conductivity, pore fluid conductivity and surface (interface) conductivity respectively and F denotes the intrinsic formation factor of the sample, is frequently referred to as the Waxman-Smith equation, although it has been slightly modified from the original one (e.g. Rink and Schopper, 1976). In measurements of sample conductivity vs. pore fluid conductivity for sedimentary rocks Rink and Schopper (1976) observed that the bulk conductivity was almost constant for dilute pore fluid but showed linear relationship at high pore fluid conductivity. They explained this behavior by surface conduction in clay minerals. Similar results were obtained for volcanic rocks by Flóvenz et al (1985) by comparison of data from resistivity soundings, P-wave structure from refraction seismic and temperature and pore fluid measurements from geothermal wells in Iceland. Pezard (1990) got similar results from measurements of basaltic cores from the Ocean Drilling Program. Several other papers have recently been published on this subject e.g. Revil et al. (1996, 1997, 2002).

2. DATA

We selected 12 cores from different alteration zones of the Icelandic crust. Two samples came from fresh Holocene surface lavas that have never been submerged below the groundwater table. The ten remaining samples were selected from geothermal wells, five from the smectite alteration zone and five from the chlorite alteration zone. The wells are in the following geothermal fields: Krafla in N-Iceland, Nesjavellir and Hengill in SW-Iceland and the Reykjanes area in SW-Iceland where the central part of the Mid-Atlantic Ridge emerges from the sea. Table 1 gives an overview of the samples and a picture of them is shown in figure 1. The samples from Krafla are from NQ-size continuous coring wells but the cores from Nesjavellir-Hengill and Reykjanes are spot-cores taken during the drilling and the original diameter of the core was about 10 cm. For our measurements samples, 25 mm in diameter and up to 50 mm long, were drilled from the original core at the rock lab at GeoForschungZentrum, Potsdam (GFZ).

With the exception of the samples nos. 11 and 18 all the borehole core samples were taken from the well and kept in the core storage at Iceland GeoSurvey (ISOR) at room temperature until they were used for the measurements. This means that the samples have been kept dry for 10-25 years before our measurements were made. Cores nos. 11 and 18 were wrapped into a plastic foil immediately after their collection from the borehole and covered with paraffin to prevent the sample form drying out.

2.1 Porosity and density determinations

Prior to the conductivity measurements the porosity and the density of the samples were measured by the triple weighing method, i.e. the samples were dried in vacuum in an oven at 60°C to remove all water from interconnected pores. The samples were weighted in dry condition and then saturated with distilled water under vacuum and the sample weighted in air and submerged in water. From these measurements the density and porosity of each sample were calculated. The results are shown in table 2.

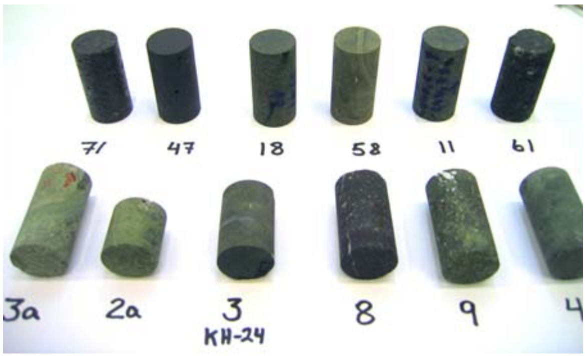


Figure 1. The samples used for the experiment.

2.2 Sample cleaning

Immediately after the porosity determinations, the samples were cleaned by putting them repeatedly into distilled water and the conductivity of the fluid measured as a function of time to control the amount of equivalent NaCl ions removed from the sample. They were cleaned in three groups of six, five and one sample. Figure 2 shows the progress of the cleaning procedure for two of the groups. One of them consisted of five cores from high-temperature fields of low salinity and these were cleaned together in the same container while the sample from the seawater saturated Reykjanes field was cleaned separately.

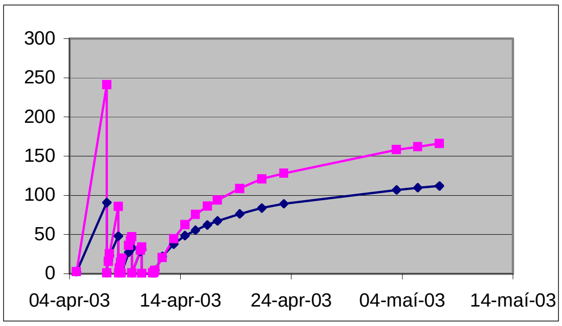


Figure 2. Cleaning fluid conductivity in µS/cm as function of time. The sample are saturated repeatedly with water with conductivity of 1,2 µS/cm. The red line shows the data for the core 8 from the Reykjanes field but the blue one is for five samples together from freshwater saturated geothermal fields.

When the samples dry in the core storage the dissolved solids in the initial pore fluid precipitate in the pores and form solid minerals. Some of these minerals, like silica might not easily dissolve again in contrast to soluble salts. By assuming that the conductivity of the initial pore fluid (table 1) is due to NaCl only, it is possible to calculate the amount of salt that should be dissolved from the pores during the cleaning process, using the measured porosity of the samples. It turns out that for the sample from Reykjanes (core 8) the removed amount of NaCl equivalent is close to the expected amount of the precipitation in the pores while much higher total amount was removed from the other 5 samples than expected. In order to account for the NaCl equivalent mass that was removed during the cleaning process, the pore fluid conductivity had to be about ten times higher than the in-situ pore fluid or alternatively the repeated addition of distilled water did cause dissolution of material from the sample. This could indicate that at pore water concentrations below 100µS/cm some dissolution occurs from minerals in the sample.

The first measurements were carried out after about a month of cleaning when the cleaning solution had reached a semi-stabile concentration. After each resistivity measurement the samples were dried in an oven under vacuum at 60°C for 6 - 24 hours. Following that, the samples were put into a exsiccator and vacuum of 0.1 mbar created. Then the samples were saturated with a NaCl solution of desired conductivity under the vacuum, which should enable the saturating fluid to penetrate all the interconnected pores. After the saturation the conductivity of the fluid, in which the samples were submerged, was monitored and measurements done when stability was reached. The time for reaching the stability increased by increasing salinity of the saturation fluid from 1-2 days to a week or so.

3. EQUIPMENT

The equipment used is the simple measurement cell shown on figure 3. The sample was wrapped on its sides with an insulating Teflon thread seal and put into a hollow 50mm long Plexiglas cylinder. Then a fluid chamber with current and potential electrodes was tightly attached to each end of the cylinder and filled with the saturating fluid that comes in contact with the bare ends of the sample. The current electrodes are at the outer ends of the fluid chamber while the potential electrodes are at the contact to the sample ends. Then the Zahner IM-6 electrochemical workstation was used to measure the impedance spectrum of the sample over the frequency range 0.1 to 10^5 Hz.

Figure 3. The equipment used for measurements of the conductivity as function of pore fluid salinity.

4. MEASUREMENTS

4.1 Porosity and density determinations

The porosity and density data of the samples are shown in table 2. The relationship between the porosity and the density of the twelve samples is shown on figure 4. It shows all the samples on the same plot but the samples from different alteration zones are shown by data points of different shape and color. It appears that the samples from the different alteration zones fall into different areas of the plot. The unaltered samples seem to have the highest density for given porosity, the samples from the smectite alteration zone the lowest and the samples from the chlorite zone lie in-between. This probably reflects the alteration of the rock matrix, i.e. in the smectite zone the alteration process of the matrix reduces the density while at higher degree of alteration denser alteration minerals replace lighter ones. In figures 5 and 6 the samples from the chlorite and smectite zones are plotted separately and a linear relation between the porosity and density is calculated. The linear regression gives matrix density of 3.0 +/- 0.04 and 2.95 +/- 0.05 for the density at zero porosity and the gradient is 0,028 +/- 0.003 and 0.032 +/- 0.005. The errors are too large to conclude that there is a significant difference between these groups of samples, more data is needed for such conclusion.

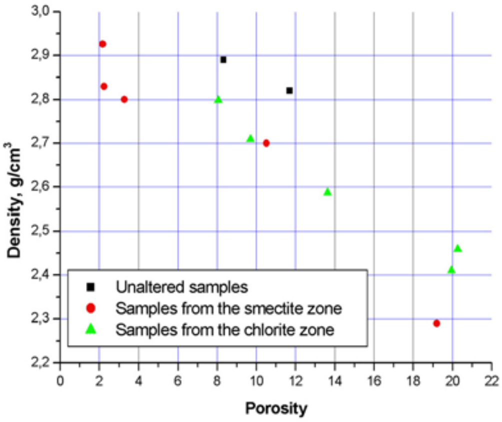


Figure 4. The density-porosity relationship for all the twelve samples

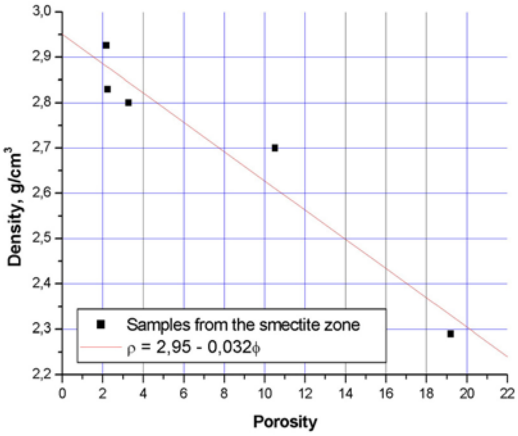


Figure 5. The density-porosity relationship for the samples from the smectite zone

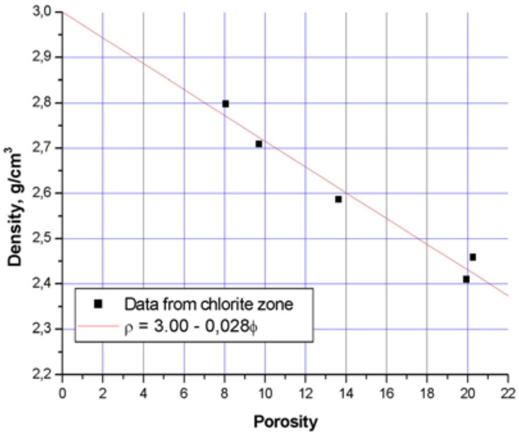


Figure 6. The density-porosity relationship for the samples from the chlorite zone.

3.3 The conductivity versus pore fluid salinity

A typical example of a full frequency spectrum of the samples is shown in Figures 7 and 8. These are the resistance (fig.7) and phase angle (fig.8) as function of the frequency. The results show a slight dependence of the resistance on the frequency, i.e. a slightly increased conductivity with increasing frequency. Above 10 kHz there is a rapid increase in conductivity and of the negative phase angle for some of the samples. This high frequency behavior is principally caused by stray capacities and lead inductivities. The picture is more or less the same regardless of the salinity. For the most practical cases in geothermal exploration the frequencies below 100 Hz are of most importance. When geoelectrical data are converted into resistivity versus depth, frequency dependence of the resistivity is generally not taken into account. Our measurements show that within this low frequency range, the decrease in resistivity is of the order of 2- 5% per decade in frequency. It follows that it will not have serious effects on the final resistivity models of the earth if the frequency dependence of resistivity is omitted during inversion of geoelectrical data.

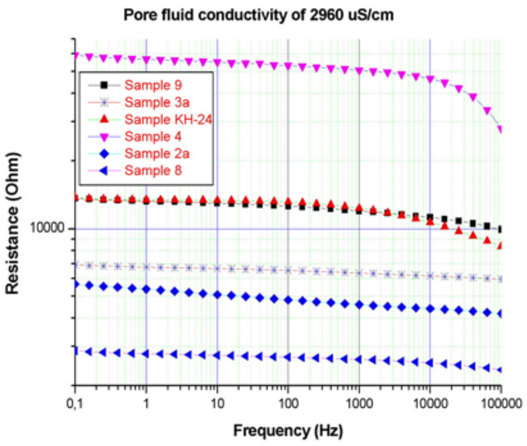


Figure 7. Resistance of six of the samples saturated with pore fluid of 2960 µS/cm as function of frequency.

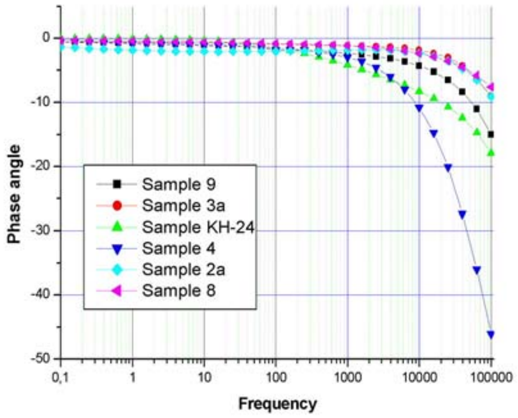


Figure 8. Phase angle of the same samples as in fig. 7 as function of frequency. Note the slightly increasing negative phase angle up to 500 to 10000 Hz and then very rapid increase in the negative value.

The results of the measurements of sample conductivity as function of pore fluid conductivity are shown on log-log plots in figure 9 for all the twelve samples used. Figure 10 shows the same data but they are now normalized according to the conductivity value at the lowest pore fluid conductivity. According to equation (1) the typical curves in figs. 9 and 10 should consist of a flat part where interface conduction is dominant and a inclined part with gradient equal to one where pore fluid conduction dominates. The curvature from the flat part to the constant unity gradient is the interval where both conduction types are involved. The isoelectrical point is defined where the contribution from the both conduction mechanisms are equal. Inspection of figs. 9 and 10 shows that both the conduction ways are present in the samples from the smectite and the chlorite zones but the isoelectrical point moves to higher value for the samples from the smectite zone. For the unaltered zone pore no interface conduction is seen for the one of the two sample but minor for the other one.

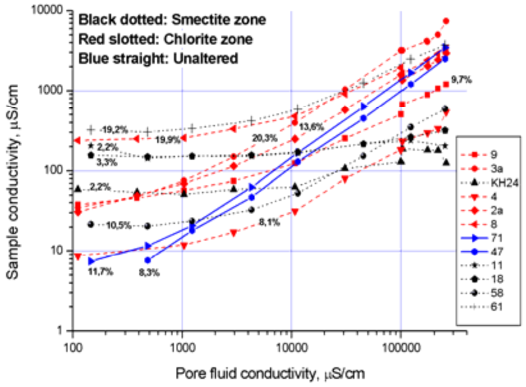


Figure 9. The result of the measurements of sample conductivity at 1 Hz versus pore fluid conductivity. Data from all the different alteration zones are shown on the figure. Note that on the average the interface conductivity is higher in the smectite zone than in the chlorite zone and almost absent in the unaltered samples. The porosity value for each sample is shown. Note that sample 8, which is reported to be from the chlorite zone, behaves like a sample from the smectite zone. For explanation of sample numbers, see tables 1 and 2.

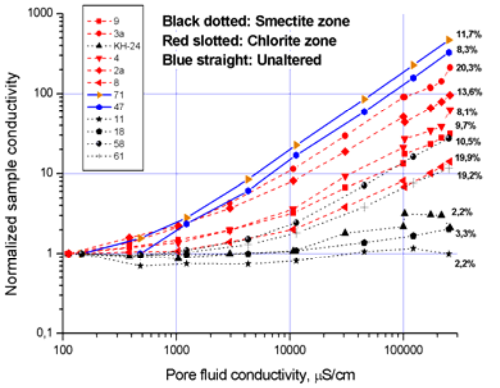


Figure 10. Normalized values for the data in figure 9. It shows clearly how the data from different alteration zones groups into different region of the plot. The porosity value for each sample is shown. For explanation of sample numbers, see tables 1 and 2.

5. DISCUSSIONS

Figure 11 (at the end of the paper) shows the linear plots of the sample conductivity versus pore fluid conductivity for the high salinity range of the measurements, i.e. where pore fluid conduction is dominant and interface conduction can be discarded as negligible. For sample 24, 11 and 18 it was not possible to make such plots due to the near constant bulk conductivity over the whole range of the pore fluid conductivity. This is caused by a high formation factor in combination with a high surface conductivity, making the contribution of the pore fluid conductivity negligible. A linear relationship should appear with a slope equal to the inverse of the intrinsic formation factor and interception equal to the value of the interface conductivity. Note that the error in the latter parameter is very high and is much better defined by the low salinity part of the curve in figure 9.

The result of the estimate of the intrinsic formation factor, interface conductivity and rough estimate of the isoelectrical point is given in table 2.

Figure 12 shows the formation factor of the samples plotted against the porosity. Again we see how the values from the three different alteration zones fall into different parts of the F-ϕ plot. The samples from the unaltered zone have the lowest formation factor relative to porosity but the smectite samples the highest. Sample 8, which is believed to be from the chlorite zone show as before similar behavior as the samples from the smectite zone. To establish an Archie type porosity-formation factor relationship,

$$F = k \cdot \phi^{-m} \qquad (2)$$

it is necessary to treat the data from the three different alteration zones separately. Since the data points are very few it is only meaningful to look at the relationship for the chlorite zone and exclude sample 8. This is done on figure 13. It gives a good linear relationship on the log-log plot.

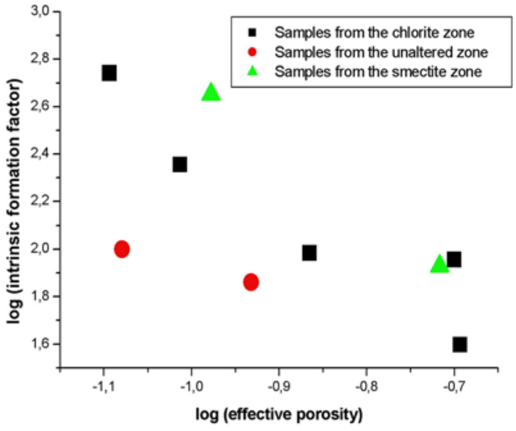


Figure 12. The intrinsic formation factor versus effective porosity. The formation factor is derived from the slope of the high salinity part of the sample conductivity as function of pore fluid conductivity. Note the different grouping of the data points from the different alteration zones.

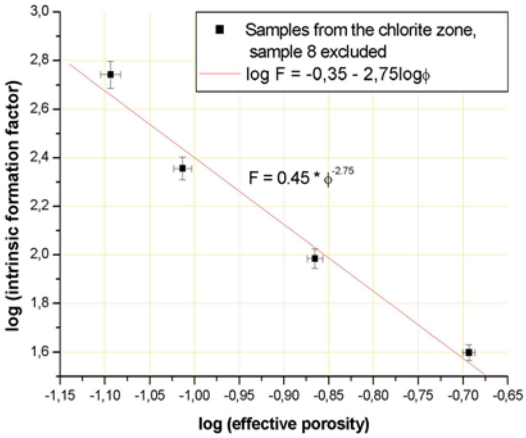


Figure 13. The intrinsic formation factor versus effective porosity for the samples from the chlorite zone. A cementation factor of 2.75 appears.

The cementation factor is 2,75 which would be interpreted as “vagular” or “moldic” porosity (Doveton, 1986) with respect to the classification of oil and gas reservoirs based on the cementation factor. Values between 1 and 2, that frequently are reported, often result from the disregard of the surface conductivity. Similarly the value of 0.44 for the constant k is abnormally low compared to published values from ocean basalts and dikes (Pezard 1990, Revil et al 1996).

Figure 10 show clearly how the data from the different alteration zones group together when the data are normalized. This shows the similarity in behavior between these groups although the absolute value of the conductivity is different. For the unaltered zone the pore fluid conductivity is obvious with little or no interface conductivity while the samples from the smectite zone shows the highest isoelectrical point and therefore interface conductivity over wider range of salinity than the samples from the chlorite zone.

From figure 9 it is clear that for the unaltered zone the pore fluid conductivity is dominant over most of the range of fluid salinities. Only at very low salinities, below 200 µS/cm signs of surface conduction is found, especially in sample 71. This means however that pore fluid conduction is dominant in practically all cases in unaltered zone since the pore fluid conductivity is usually higher than 100 µS/cm. Rough estimate of the value of the interface conductivity gives σ_s ~ 5 µS/cm for the unaltered zone.

Three of five samples from the smectite zone (nos. KH-24, 11 and 18) in figure 9 have very low effective porosity (2-3%). These samples show almost constant conductivity regardless of the pore fluid conductivity. There is only a small and irregular increase in bulk conductivity with pore fluid conductivity. It cannot be ruled out that this could be due to incomplete saturation of these low porosity samples. Note however that in these cases the interface conductivity is rather high or 50-150 µS/cm. The two other samples from the smectite zone (nos. 58 and 61) show a very typical interface conduction behavior but the value of the interface conductivity is different by an order of magnitude; being very low, 20 µS/cm in sample 58, which is a dense basalt core but 300 for sample 61 which is a highly altered hyaloclastite, the average value being 135 µS/cm. Rough estimate shows that the isoelectrical point for these two samples is at pore fluid conductivity of 4000-6000 µS/cm

In contrast to the existing hypothesis all the five samples from the chlorite zone show significant interface conduction (figure 9). One sample, sample 8 from the Reykjanes area, shows very similar behavior to the samples from the smectite zone, i.e. high value of interface conductivity (250µS/cm) and isoelectrical point at pore fluid conductivity close to 5000 µS/cm. The remaining four samples from Nesjavellir, Hengill and Krafla all show considerable interface conduction but the value is an order of magnitude lower than for the smectite zone (10 – 30 µS/cm) with isoelectrical point at σ_w < 4000 µS/cm. Keeping in mind that the pore fluid conductivity in most geothermal systems in Iceland is lower than 1000 µS/cm at 20°C and the temperature dependence of interface conduction is at least double as high as for the pore fluid conduction (Flovenz et al., 1985, Kulenkampff et al., 2005) it can be concluded that surface conduction is also the dominant conduction mechanism in the chlorite zone. This in disagreement with older ideas (e.g. Árnason et al. 1992).

Since the top of the high resistive boundary coincides usually with the change from smectite to mixed clay / chlorite alteration the resistivity increase has to be related to the properties of the alteration minerals. The interface (surface) conductivity has been shown to be directly proportional to the cation exchange capacity (CEC) of the minerals involved (Revil et al., 1998). The CEC differs from one mineral to another, smectite has CEC of 0.8-1.5 meq/g but chlorite 0.01 meq/g. Thus smectite has up to two orders of magnitude higher CEC than the chlorite that can explain the resistivity increase at the top of the mixed clay/chlorite zone. However, our results show that the surface conductance in the chlorite zone is still dominating the pore fluid conduction at room temperatures and at most common salinities for Icelandic geothermal reservoirs.

Figure 14 shows a modified version of the main resistivity structure of the basaltic crust in Iceland and what processes are involved. It is both based on in-situ measurements and our laboratory data. The uppermost part is unaltered, with negligible interface conduction, showing relatively high resistivity, depending on the pore fluid salinity. Below this zone the zeolite-smectite zone is entered where interface conductivity becomes dominant and the resistivity is strongly reduced. Further decrease in resistivity with depth follows, partially due to increased temperature and partially due to increased alteration. Below where the mixed clay zone is reached the resistivity increases again, probably due to strongly reduced cation exchange capacity of the clay minerals in the mixed clay and chlorite zone. The transition from the smectite to mixed clays seems to happen at temperature close to 230°C. Further below a slow decrease in resistivity is expected and is usually seen in MT measurement. This decrease is considered to be due to slowly increasing temperature with depth in the reservoir.

From the present data it is also obvious that samples from the same alteration zone can have very different interface conductivity, which again is a decisive parameter for the bulk resistivity for the sample at reservoir salinities. It has also been observed that different types of basalt lava units can have quite different bulk resistivity even though they are in the same alteration zone and at the same temperature (e.g. Flóvenz and Karlsdóttir, 2000). This of course leads to the conclusion that factors like amount of alteration minerals and matrix microstructure of the formation must influence the bulk resistivity heavily. Systematic studies of the observed resistivity in volcanic rock, including CEC analysis is necessary for further understanding of the resistivity structure of geothermal system in volcanic environment.

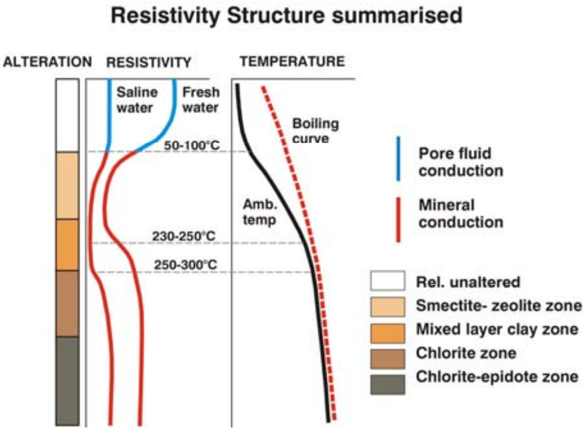


Figure 14. The general resistivity structure of the basaltic crust in Iceland. The depth scale is arbitrary; the actual scale will depend on the past and present temperature profiles.

6. CONCLUSIONS

The main conclusions can be summarized as follows.

We have measured the conductivity of twelve samples of volcanic rock from different alteration zones in Iceland as function of the pore fluid conductivity at room conditions. Unaltered samples show typical pore fluid conductivity over almost all the range of salinities while the samples from the smectite and chlorite zone show near constant value of conductivity over wide range of pore fluid salinity. Only at high salinities, i.e. over 1000 µS/cm for the sample from the chlorite zone and 3000-5000 µS/cm for the samples from the smectite zone, the pore fluid conduction starts to become of significance. This means that for almost all freshwater saturated high temperature fields the interface conduction is the dominant conduction mechanism, both in the chlorite and the smectite zone. We conclude that the observed increase in resistivity at the top of the mixed clay/chlorite zone in many high temperature geothermal reservoirs worldwide is most likely due to much higher cation exchange capacity in the smectites than in the alteration minerals below.

REFERENCES

Árnason, K., and Flóvenz, O.G.: Evaluation of physical methods in geothermal exploration of rifted volcanic crust. *GRC Transactions*, vol. 16, 207-217, 1992.

Árnason, K., Karlsdottir, R., Eysteinsson, H., Flóvenz, O.G., and Gudlaugsson, S.T.: The Resistivity Structure of High-Temperature Geothermal Systems in Iceland, *Proceedings, World Geothermal Congress*, Kyushu-Tohoku, 923-928, (2000).

Bussian, A.E, Electrical conductance in a porous medium, *Geophysics*, 48, 1258-1268, (1983).

Doveton, J.H.: Log analysis of subsurface geology, John Wiley & Sons New York, (1986).

Flóvenz, Ó.G., Georgsson, L.S., and Árnason, K.: Resistivity Structure of the Upper Crust in Iceland, *J. Geophys. Res.*, 90, 10136-10150, (1985).

Flóvenz, Ó.G., and Karlsdóttir, R.: TEM-resistivity image of a geothermal field in N-Iceland and the relation of the resistivity with lithology and temperature.

Proceedings, World Geothermal Congress, Kyushu-Tohoku, 1127-1132, (2000).

Hill, H.J. and Milburn, J.D.: Effect of clay and water salinity on the electrochemical behavior of reservoir rocks, *Trans. Am. Inst. Min. Metall. Pet. Eng.,* 207, 65-72, (1956).

Kulenkampff, J., Spangenberg, E., Flovenz, O., Raab, S., Huenges, S.: Petrophysical parameters of rocks saturated with liquid water at high temperature geothermal reservoir conditions. Proceedings World Geothermal Congress, (2005).

Pezard, P., Electrical Properties of Mid-Oceanic Ridge Basalt and Implications for the Structure of the Upper Oceanic Crust in Hole 504B. *J. Geophys. Res., 95*, 9237-9264, (1990).

Revil, A., Darot, M., Pezard, P.A., and Becker, K., Electrical Conduction in Oceanic Dikes, Hole 504B, in *Proceedings of the Ocean Drilling Program, Scientific Results,* edited by J.C.Alt, H.Kinoshita, L.B. Stokking and P.J.Michael, vol 148, 297-305, (1996).

Revil, A. and P.W.J. Glover, Theory of ionic-surface electrical conduction in porous media. *Phys. Rev*. B., 55, (3), 1757-1773, (1997).

Revil, A. and P.W.J. Glover, Nature of surface electrical conductivity in natural sands, sandstones and clays. *Gephys. Res. Lett.,* 25 (5), 691-694, (1998).

Revil, A., L.M.Cathles III, S. Losh and J.A. Nunn, Electrical conductivity in shaly sands with geophysical applications, *J. Geophys. Res*. 103 (B10), 23925-23936, (1998).

Revil, A., D. Hermitte, E. Spangenberg, J.J. Cochemé, Electrical properties of zeolitized volcaniclastic materials. *J. Geophys. Res*. 107 (B8), doi10.1029/2001JB0005999, (2002).

Rink, M., and Schopper, J.R.: Pore structure and physical properties of porous sedimentary rocks, *Pure Appl. Geophys.,* 114, 273-284, (1976).

Ussher, G., Harvey, C., Johnstone, R., and Anderson, E.: Understanding the Resistivities Observed in Geothermal Systems, *Proceedings, World Geothermal Congress, Kyushu-Tohoku,* (2000).

Waxman, M.P. and Smits, L.J.M.: Electrical Conductivities in Oil-Bearing Shaly Sands, Soc. *Pet. Eng., 8*, 107-122, (1968).

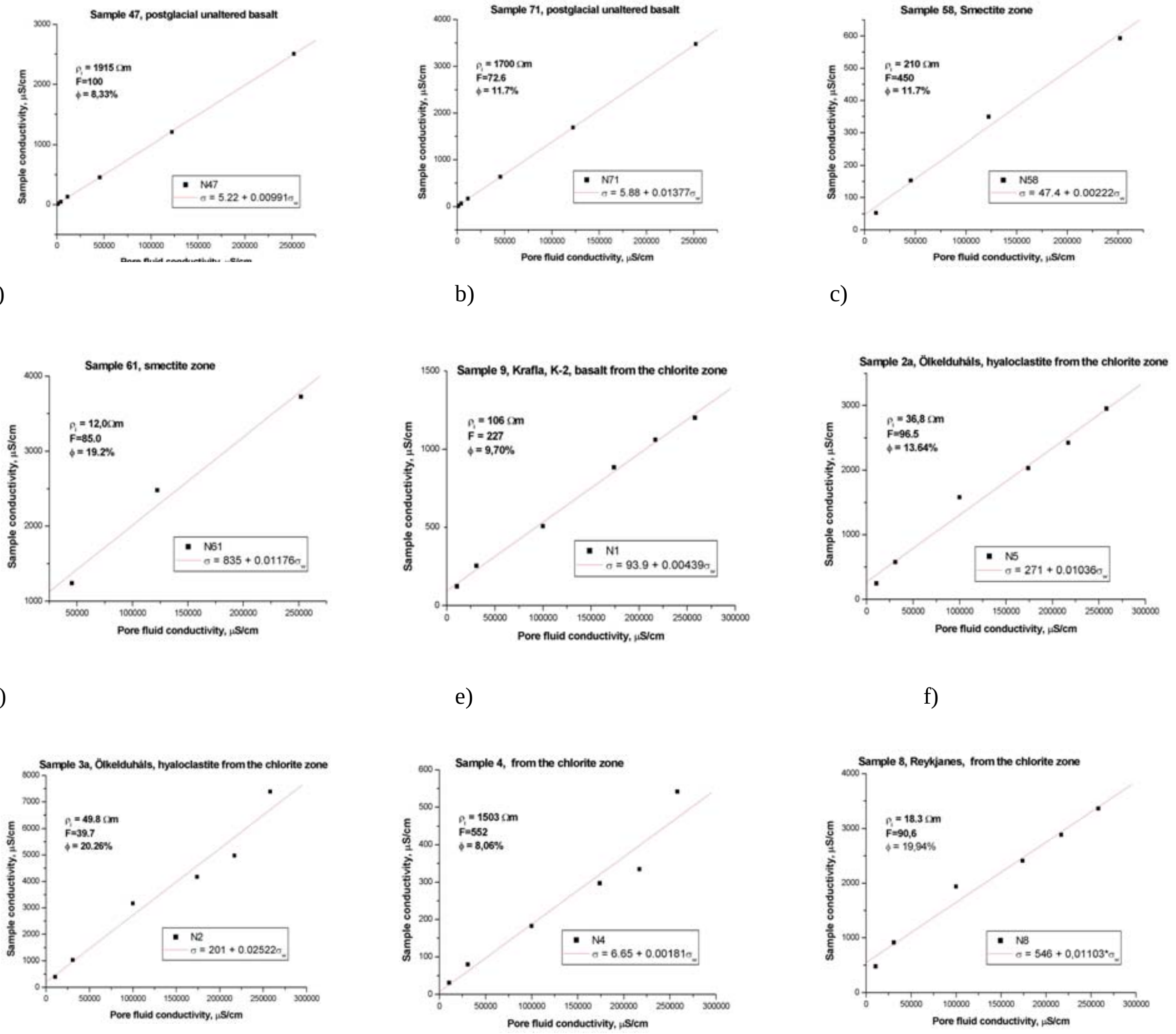
Windsauer, W.O., and McCardell, W.M.: Ionic double-layer conductivity in reservoir rocks, *Am. Inst. of Mining and Metal. Eng.*, 198, 129-134, (1953).

Table 1. An overview of the core samples used in the measurements and the conditions in the boreholes of the geothermal field from which they derive.

Sample no.	Location	Well no.	Year of sampling	Material	Alteration zone	Sample depth (m)	In-situ temperature °C	Estimated in-situ fluid conductivity at 25°C in µS/cm
2a	Hengill	ÖJ-1	1994	Hyaloclastite	Chlorite	794,5	200	808
3a	Hengill	ÖJ-1	1994	Hyaloclastite	Chlorite	795	200	808
3 (KH-24)	Krafla	KH-1	1991		Smectite			780
4	Nesjavellir	NJ-17	1986		Chlorite	~876-7	190-200	751
8	Reykjanes	RN-8	1969	Basaltic tuff	Chlorite	~1005	260-270	45500
9	Krafla	K-2	1974	Basalt	Chlorite	540	200	909
11	Krafla	KS-3	2002	Basalt	Smectite	240		
18	Krafla	KS-3	2002	Basalt	Smectite	380		
47	Heiðmörk	surface	2003	Basalt	Unaltered	0	5	Not saturated
58	Krafla	KH-1	1991	Basalt	Smectite	187		780
61	Krafla	KH-1	1991	Hyaloclastite	Smectite	195	170	780
71	Reykjanesskagi	surface	2003	Basalt	Unaltered	0	5	Not saturated

Table 2. The result of measurements of porosity and density for the core samples as well as the derived values for the intrinsic formation factor and the interface conductivity.

Sample no.	Location	Well no.	Year of drilling	Material	Alteration zone	Porosity %	Density g/cm^3	Formation factor	Interface conduct. µS/cm
2a	Hengill	ÖJ-1	1994	Hyaloclastite	Chlorite	13,6	2,59	96,5	20
3a	Hengill	ÖJ-1	1994	Hyaloclastite	Chlorite	20,3	2,46	39,7	30
3 (KH-24)	Krafla	KH-1	1991		Smectite	2,18	2,93	-	50
4	Nesjavellir	NJ-17	1986		Chlorite	8,06	2,80	552	8
8	Reykjanes	RN-8	1969	Basaltic tuff	Chlorite	19,9	2,41	90,6	250
9	Krafla	K-2	1974	Basalt	Chlorite	9.7	2.71	227	30
11	Krafla	KS-3	2002	Basalt	Smectite	2,24	2,83	-	150
18	Krafla	KS-3	2002	Basalt	Smectite	3,27	2,80		150
47	Heiðmörk	surface	2003	Basalt	Unaltered	8,33	2,89	100	5
58	Krafla	KH-1	1991	Basalt	Smectite	10,5	2,70	450	20
61	Krafla	KH-1	1991	Hyaloclastite	Smectite	19,2	2,29	85	300
71	Reykjanesskagi	surface	2003	Basalt	Unaltered	11,70	2,82	72,6	5



a) b) c)

d) e) f)

g) h) i)

Figure 11. The sample conductivity plotted against the pore fluid conductivity for the high salinity range (> 10000µS/cm) where pore fluid conduction is dominant. Note that with exception of the Reykjanes peninsula in SW Iceland, the pore fluid conductivity in the upper crust and in the geothermal fields is usually in the range of 200 - 1000 µS/cm at room temperature. A linear relationship should exist in the high salinity range with a slope equal to the inverse intrinsic formation factor and interception equal to the interface conductivity. Note the latter is very inaccurate from the present dataset and is much better determined from the low salinity range.