Proceedings World Geothermal Congress 2020+1
Reykjavik, Iceland, April - October 2021

The Relation Between Resistivity and Temperature in the Basaltic Crust of Iceland

Ólafur G. Flóvenz and Arnar Már Vilhjálmsson

ÍSOR, Iceland GeoSurvey, Grensásvegur 9, 108 Reykjavík, Iceland

ogf@isor.is, amv@isor.is

Keywords: resistivity, temperature, low temperature fields, high temperature fields, brittle ductile boundary, basaltic crust

ABSTRACT

The ability of resistivity measurements to provide information on the temperature distribution in the uppermost crust is vital for geothermal exploration. During the past few years effort has been made within various projects and by many researchers to understand the resistivity properties of basaltic rocks and how it is affected by temperature. These efforts cover various types of laboratory measurements of core samples including measurements at in situ conditions for high temperature fields. In this paper we review the past results of laboratory data and validate some of the results by applying them to real field data where logged boreholes and resistivity models from TEM and MT measurements are available. We predict, and show example, that a typical low temperature field in Iceland should produce a low resistivity anomaly at shallow depth but a high resistivity anomaly deeper in the reservoir. The main results of this paper are, however, that the laboratory measurements of resistivity of rock samples at in-situ conditions are crucial to draw conclusions about geothermal reservoir properties from resistivity soundings.

1. INTRODUCTION

Resistivity imaging of the subsurface involves four steps: data acquisition, pre-processing and preparation of data for inversion or modelling, the modelling of the processed data, and finally the interpretation of the subsurface resistivity model in terms of geothermal parameters. The last item is the main topic of this paper. The first three items are related to the creation of a reliable resistivity model of the underground. They are all subject to some basic assumptions and include uncertainties and ambiguities that must be recognized when the final models are interpreted in geothermal terms and evaluated. In this paper we assume that the resistivity models we use are correctly representing the true resistivity of the subsurface. Furthermore, we concentrate on the basaltic rock that forms the Icelandic crust.

Resistivity methods are the most important geophysical methods in geothermal exploration. Not only is the resistivity highly sensitive to temperature and geothermal alteration processes, but it depends also on several other physical parameters of the pore fluid and the reservoir rocks. It is a complicated task to interpret subsurface resistivity in terms of geothermal parameters and requires under-standing of the still debated conduction mechanisms in basaltic rock and its dependence on various parameters.

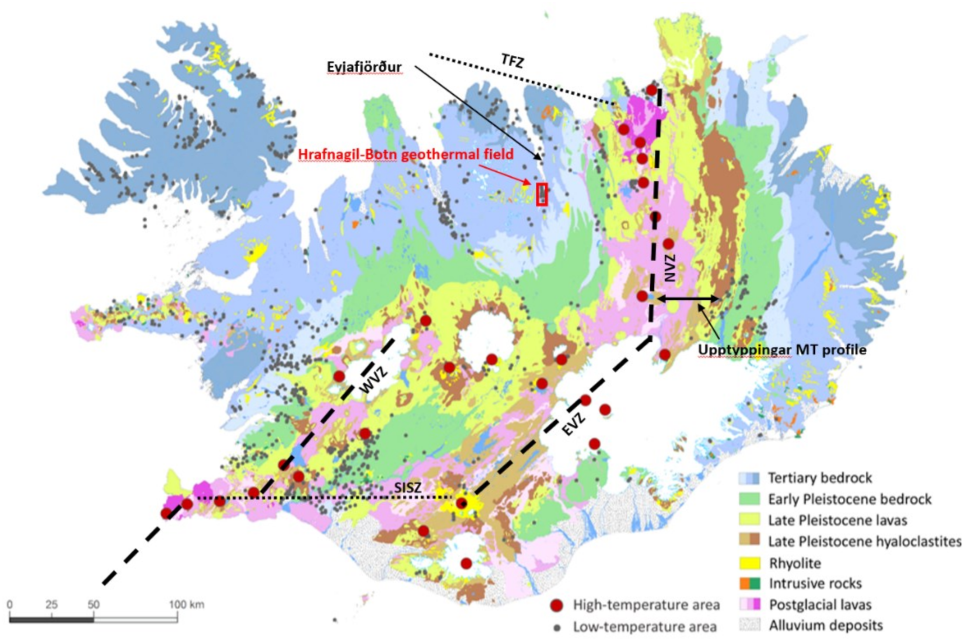

Figure 1. A simplified geological map of Iceland showing the increasing age of the crust away from the volcanic rift zone. The red dots show the location of known high temperature fields and the black dots the low temperature fields. The black pecked lines show approximately the active rift axis and the black dotted lines show the transform zones. WVZ is the Western Volcanic Zone, EVZ is the Eastern Volcanic Zone, NVZ is the Northern Volcanic Zone, TFZ is the Tjörnes Fracture Zone and SISZ is the South Iceland Seismic Zone.

2. FORMATION AND DEVELOPMENT OF THE ICELANDIC CRUST AND ITS GEOTHERMAL FIELDS

Iceland is the only place in the world where an active oceanic spreading ridge is above sea level. The reason for this is the presence of a low-density mantle hot spot, centred below central Iceland, that increases the magmatic production rate compared to normal oceanic ridges. This leads to abnormal crustal thickness beneath Iceland of 20-40 km (Bjarnason et al., 1993; Kaban et al., 2002) compared to normal oceanic crust of 10 km or less. The spreading axis of the Mid-Atlantic Ridge crosses the island as a zone of active spreading and volcanism, referred to as the axial volcanic zone (Figure 1). It is composed of several segments named the Western Volcanic Zone (WVZ), the Eastern Volcanic Zone (EVZ) and the Northern Volcanic Zone (NVZ). The measured half spreading rate in Iceland is close to 1 cm/year. The axis rises from sea on the Reykjanes peninsula on the south-west corner of the country and submerges again at the north-eastern coast (Figure 1). The axial volcanic zone does not form a straight line through the country. It is shifted about 150 km eastwards, close to the southern coast (The South Iceland Seismic Zone (SISZ)) and back in a westerly direction at the northern coast (The Tjörnes Fracture Zone (TFZ)). Throughout the almost 20 m.y. of the exposed geological history of Iceland the axial rift zone has shifted eastwards a few times, leaving behind traces of ancient spreading axis and transform tectonics, especially on the American plate (e.g. Sæmundsson, 1979).

2.1Crustal formation

The crustal accretion process in Iceland has been modelled and described by Pálmason (1973) (Figure 2). His model assumes a simple spreading axis and spreading rate, where new crust is partly formed by dyke injection into the existing crust and partly by surface volcanism. The eruptions cause the lava to accumulate at certain rate, normally distributed around the spreading axis, and the crust subsides by the same amount as the overlying erupted mass. This means that lava that solidifies on the surface moves horizontally away from the spreading axis, with half spreading rate of 1 cm/year, but simultaneously moves vertically down due to the load of lava that accumulates later at the surface. The trajectories of the lava successions are shown in Figure 2. The magma that cool on the surface at the spreading axis move vertically down with time, whilst those which accumulate outside the spreading axis move both laterally and downwards. When lava has left the volcanic zone it only moves laterally away from the spreading axis with time.

A consequence of this process is that the lava becomes reheated as it moves to greater depth, pores close due to external pressure and it undergoes hydrothermal alteration which finally makes the rock almost impermeable. Close to the spreading axis the subsiding crust can even be reheated to solidus of certain minerals, so it starts to melt partially and create silicic magma. This rifting process is important to understand the alteration pattern of the crust that has large impact on the bulk resistivity.

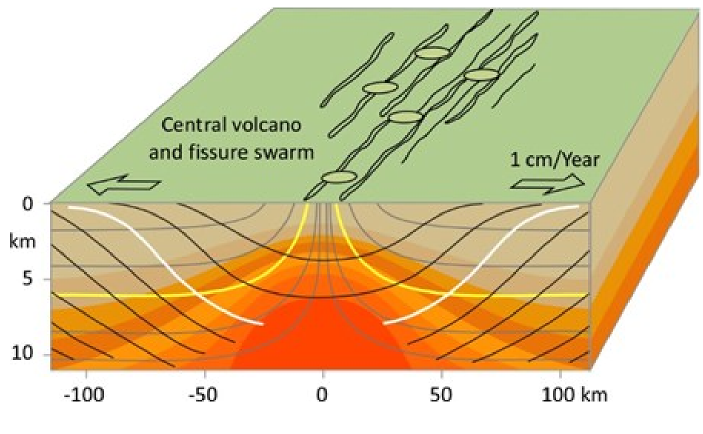

Figure 2. The model of the crustal accretion by Pálmason (1973). The crust forms at the rift axis though intrusions and subaerial eruptions and moves laterally away from the axis and subsides simultaneously due to subsequent lava accumulations on the surface. The bold yellow line and the thin grey lines show the trajectories of the lava material from the surface away from the rift axis and downwards. The bold white line and the thin black lines show the isochrones and the temperature is shown with the colour scale.

The subaerial volcanism of the country resulted in extensive eruption of flood basalts that characterizes the pre-glacial and interglacial periods. During the glaciation periods, when the country was mostly covered with ice, elongated hyaloclastite ridges or table mountains were formed and accumulated above the volcanic fissures. The glaciation and deglaciation furthermore lead to large vertical crustal movements that might have contributed to the formation of fracture dominated hydrothermal fields outside the rift zones (Böðvarsson, 1982).

Fresh lava on the surface has very high porosity (28 ± 13%) and permeability (27 mD) (Sigurðsson and Stefánsson, 1994). Therefore, the assumption can be made that there is practically no conductive heat transport in the near surface lava pile, all heat from below is removed by groundwater flow. As the lava becomes buried the porosity and permeability reduce with depth because of the burial pressure and the precipitation of secondary minerals from geothermal fluids. Temperature logs in boreholes in the volcanic zones in Iceland indicate that primary permeability has been reduced enough at around 1 km depth to let thermal conduction dominate the heat transport. A result of this is that the uppermost 1 km of the volcanic crust in Iceland should have very high permeability. But since repeated glacial erosion has removed the uppermost 1-2 km of the crust at present sea level outside the volcanic zone, the general permeability of the basaltic crust outside the volcanic zone is low.

2.2 Heat flow in Iceland

Heat flow in Iceland is high compared to continental areas. The heat flow is basically controlled by two processes. One is a background heat flow, originating from the cooling crust moving away from the spreading axis, like at the mid-oceanic ridges. The other is local high or low heat flow anomalies caused by convection of water in vertical fracture systems. The high values correspond to the upwelling part, while the low values relate to the down flow pattern. Since the crust in Iceland is rather homogeneous with respect to thermal conductivity, usually within 1.6-1.9 W/m°K, (Oxburgh and Agrell, 1980; Pálmason 1979), the near surface temperature gradient is frequently used as a proxy for heat flow. The typical background values of the temperature gradient in Iceland is 80-100°C/km at the border of the volcanic rift zone down to 40-50°C/km in the oldest crust that is farthest away from the rift axis. Within the volcanic zone, however, the uppermost 1 km of the crust consists of highly permeable young volcanics, where all conductive heat from below is transported away by large groundwater currents. Therefore, almost a zero-temperature gradient is observed in the uppermost 1 km within the volcanic rift zone, except for the high-temperature hydrothermal fields associated with the volcanic centres.

2.3 Geothermal fields

Geothermal areas in Iceland are basically of two different types, high-temperature fields and low-temperature fields. The locations of the high and low temperature fields in Iceland are shown in Figure 1. There are fundamental differences between high and low temperature fields. The high-temperature fields have reservoir temperatures of 200-340°C and are exclusively located within active volcanoes or recent post-glacial volcanism in the axial rift zone. Their surface manifestations are hot springs and fumaroles and high-temperature rock alteration, resulting in colourful and picturesque landscapes. The geothermal fluid is usually acidic, and the rather high chemical content prevents direct use of the fluid. However, apart from the western part of the Reykjanes peninsula, the reservoir fluid is of meteoric origin with relatively low salinity and electrical conductivity. Due to the crustal spreading and shift of the rift axis in time, many central volcanoes have migrated out of the volcanic rift zone where they originally formed and are now widely found as extinct and exhumed in the lava pile of the island.

The low-temperature fields in Iceland show quite a different character from those in the high-temperature fields. Their location is shown in Figure 1. They are almost all outside the axial rift-zone. Their reservoirs are fracture dominated in otherwise low permeability basaltic lavas or hyaloclastites. Frequently, they express an elongated surface heat flow anomaly that is 1-3 km long and a few hundred meters wide expressing the up-flow zone of the geothermal reservoir. The heat is extracted from the relatively high background temperature gradient by fluid convection in permeable fracture systems (Flóvenz and Sæmundsson, 1993). The chemical content of the geothermal fluid within the low-temperature system is usually quite low, typically with total dissolved solids of less than 300 mg/L. At a few places, the reservoir fluid is slightly seawater contaminated giving rise to total dissolved solids of over 1000 mg/L.

3. THE CAUSE OF RESITVITY VARIATIONS IN THE BASALTIC CRUST OF ICELAND

The crust in Iceland is composed of two main regimes, the upper crust mainly made of extrusive basaltic material with increasing proportion of dyke intrusions with depth and lower crust that is supposed to be made of almost 100% intrusives. The thickness of the upper crust seems to be 6 km or less depending on the degree of glacial erosion with exception of a part of SE-Iceland where it is deeper (Flóvenz, 1980). The lower crust is supposed to be of rather homogeneous lithology and the resistivity of it is most likely only affected by temperature variations.

The upper crust in the tertiary areas of Iceland is dominantly made of extensive layers of flood basalts with thin layers of scoria and sediments in between and intersected by near vertical basaltic dykes. Minor volumes of rhyolites also occur. TEM resistivity measurements show that the resistivity of different units of flood basalts can vary by factor of 2-5 (Flóvenz and Karlsdóttir, 2000). This physical cause of this difference has not been studied but it might be due to different system of microfractures affecting the formation factor. The upper crust in the quaternary part of Iceland is characterized by a mixture of hyaloclastites formed under glaciers and flood basalts erupted during interglacial periods. Therefore, the upper crust within the quaternary part of Iceland show much more topography than observed within the tertiary lava pile. This is reflected in the resistivity of the upper crust since hyaloclastites normally have much lower resistivity than the basaltic lavas.

The process of electrical conduction in basaltic crust has been investigated for a long time, both experimentally and theoretically, and still is (e.g. Levy et al., 2018 and references therein). The highlights of these research activities are the complex role of smectite and other clay minerals (e.g. Levy et al., 2018; Flóvenz et al., 1985; Árnason et al., 2000), the dominance of surface or interface conduction and its high temperature dependence (e.g. Kristinsdóttir et al., 2010; Flóvenz and Karlsdóttir, 2000) and the measurements of resistivity of porous rock above 300°C (e.g. Nono et al., 2018). To summarize, the following factors seems to be of the most importance for the crust in Iceland:

Lithology affects the resistivity in the upper crust. Hyaloclastites have considerably lower resistivity compared to basaltic lava at similar temperature and alteration stage. There is evidence for lower resistivity in sequences of thick porphyritic basalts and in sequences of olivine basalts compared to ordinary tholeiites (Flóvenz and Karlsdóttir, 2000) but this need to be investigated further. Since the lower crust is supposed to be composed of rather homogeneous material the lithology is not an important factor for the lower crust.

Clay content is one of the most important factors affecting the resistivity in the upper crust. This applies especially to the smectite content due to its high cation exchange capacity (CEC) (Levy et al, 2018) that causes surface or interface conductivity to be the dominant conduction mechanism rather than the pore fluid or mineral conductivity (Flóvenz et al, 1985). Massive hydrothermal alteration in high temperature geothermal fields results in large smectite content with relatively high conductivity. In a normal tertiary crust outside high temperature fields the clay content is not likely to vary considerably on regional scale.

Alteration increases generally with depth due to the crustal accretion process as described in Figure 1. This is a non-reversible regional process, controlled by the rate of spreading and subsiding as well as the regional temperature gradient. It includes formation of alteration minerals that fills up pores and fractures in the rock and causes the resistivity to increase with depth.

Pore fluid salinity is not affecting the resistivity seriously unless for completely fresh basalts close to the surface where no clay minerals have formed or where the pore fluid is of seawater salinity, e.g. in the outer part of the Reykjanes peninsula and in a few coastal areas in Iceland. This is because the conduction in the basalts is dominated by interface (surface) conduction but not by pore fluid conduction (Flóvenz et al., 1985; Kristinsdóttir et al., 2010; Nono et al., 2018).

Phase change of the pore fluid as well as entering the supercritical regime (Kummerow et al., 2018; Milsch et al., 2010; Nono et al., 2018) is reported to cause strong decrease in conductivity. The critical point of the pore water depends on the salinity. For the conditions in Iceland the critical point should be reached where the temperature exceeds 374 - 407°C below 3 km depth, depending on the fluid salinity.

Mineral conduction is very low unless at very high temperature. Measurements of basalt samples shows that it follows typical Arrhenius relationship (Nono et al., 2018). Furthermore, the mineral conduction seems to take over as the dominant conductor in the temperature range 400-600°C and control the resistivity until onset of partial melting.

Temperature is probably the most important geothermal exploration parameter to obtain from resistivity soundings. However, the conductivity dependence on temperature is quite complex. Laboratory measurement show that for the temperature range 25-200°C the conductivity increases strongly with temperature, and the temperature increase can be described by the equation

$$\sigma(T) = \sigma_O(1 + \alpha(T - T_0)) \qquad (1)$$

where $\sigma(T)$ is the conductivity as function of temperature T, σ_O is the conductivity at the reference temperature T_0 and α is the respective temperature coefficient. The coefficient α would be 0.023 if the conductivity would be controlled by pore fluid conductivity only. However, most experiments on basalts show much higher α-values for samples saturated with fluid of typical salinity for the Icelandic crust and its geothermal reservoirs where interface (surface) conductivity prevails. Laboratory measurements of 25 samples of Icelandic hyaloclastites, basalts and dolerites (Kristinsdóttir et al, 2010, Nono et al 2015) from different alteration zones gives average value of 0.050 for all the samples while the hyaloclastites give average value of 0.35 with standard deviation of 0.12 and basalt and dolerite samples have average value of 0.055 with standard deviation of 0.019 when one out-layer is ignored. Similar values are reported in Flóvenz et al. (1985), Revil et al. (1998) and Kulenkampff et al. (2005). According to laboratory experiments of Nono et al. (2018) the rapid increase in conductivity versus resistivity flattens out somewhere between 200 and 300°C and start to decrease sharply when temperature exceeds 300°C and drops until the temperature has reached 400-600°C where mineral conduction becomes dominant until the onset of melting (see Figure 11 in Nono et al., 2018).

In the following chapters we use the above-mentioned relationship to interpret two resistivity sections from different areas in Iceland.

4. RESISTIVITY OF THE LOW TEMPERATURE FIELDS OUTSIDE THE VOLCANIC ZONE

Since 1970 hundreds or thousands of resistivity soundings have been collected outside the volcanic zone all over Iceland as a part of geothermal exploration for low temperature systems. Most of these soundings are DC soundings with Schlumberger arrangement but since 1990 those were replaced by TEM soundings. Typical depth resolution of these soundings is 300-1000 m, depending on the sounding configuration and the actual resistivity.

The general resistivity structure of the uppermost crust in Iceland was described by Flóvenz et al. (1985). In the tertiary lava pile the resistivity depends on two main parameters, the temperature gradient and the depth of burial, with the associated secondary mineralization, including formation of smectites. Both parameters lower the resistivity with depth but below a certain level the increased mineralization increases the resistivity with depth and counteracts the effect of increased temperature with depth. In general, it is assumed that a low temperature field should appear as a low resistivity anomaly.

The convective nature of the low temperature fields is recognized through comparison of temperature logs inside a low temperature field with similar temperature logs outside the field where conductive background heat flow dominates. This is further explained in Figure 3. From the figure we can conclude that the low temperature fields should appear as low resistivity anomalies at shallow depth but high resistivity anomalies below some depth limits. The depth limit depends in each case on the reservoir temperature and the background temperature gradient.

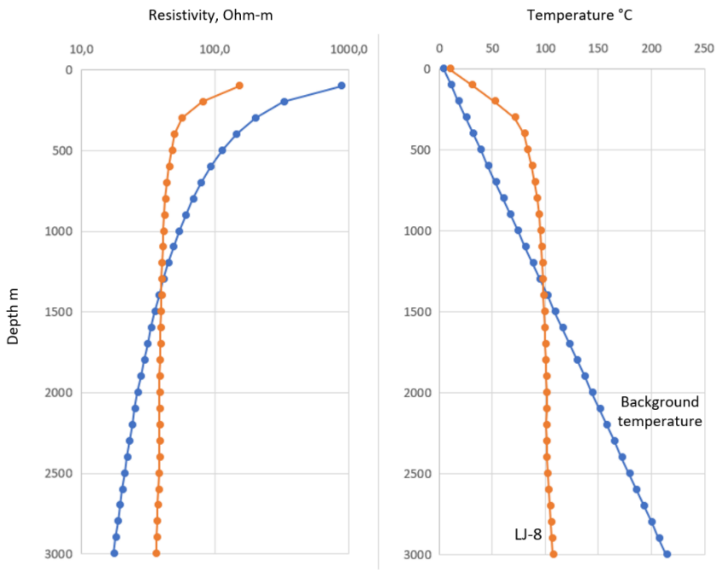

Figure 3. The graph to the right shows an equilibrated temperature log from well LJ-8 (orange dots) in the Laugaland low temperature field in the tertiary lava pile of Eyjafjörður, N-Iceland (Figure 1), compared to the temperature derived from the background temperature gradient in the surrounding area (blue dots). Above ca 1400 m the temperature within the geothermal field is considerably higher than expected from the background temperature gradient but lower below. This shows clearly the convective heat extraction process within the low temperature geothermal fields, heat has been brought from the deeper part of the system to the shallow part and to the surface through hot springs. The left graph shows the corresponding resistivity change with depth if temperature is the only parameter that changes with depth. The resistivity is calculated from equation 1 by assuming the resistivity to be 200 Ωm at 25°C and the coefficient α to be 0.055 °C^{-1}.

Let us now look at a combined TEM/MT resistivity section crossing the Hrafnagil-Botn geothermal field in Eyjafjörður, N-Iceland (Figure 1), based on joint 1D inversion of MT and TEM data (Figure 4). The well GG-1 in the section is about 2 km west of well LJ-8 (see Figure 3). The figure is taken from Vilhjálmsson and Karlsdóttir (2015) but the upper part of this section, i.e. based on TEM soundings alone, was discussed previously by Flóvenz and Karlsdóttir (2000). The crust in the area is formed of slightly south dipping widely extensive sequences of basaltic layers. Well LJ-8 was drilled through the same layers as shown in the section. The highlights of the section are (the numbers refer to the yellow numbers in Figure 4):

- The two low resistivity layers (nos. 1 and 2) coincide respectively with series of thick olivine tholeiitic lava and series of 20-30 m thick porphyritic tholeiitic lavas interbedded with a few meter sedimentary layers in-between (Flóvenz and Karlsdóttir, 2000). The main aquifers in the boreholes are found to be within the porphyritic lava units.
- The higher resistivity coincides with series of compound basaltic lavas e.g. no. 4 in Figure 4.
- The lowest resistivity within the porphyritic units are found beneath the hot spring area between wells BN-1 and RWN-7 where the temperature is highest. It is noteworthy how the resistivity in the two low resistivity layers (nos. 1 and 2) increases sharply (nos. 7 and 8) south of the well BN-1 in line with lowering of the temperature. The southern limits of the geothermal field are just south of well BN-1. The resistivity increase is of similar order as would be expected from Eq. 1
- There is an up-doming high resistivity body (no. 3) directly below the lowest resistivity in the porphyritic layer (no. 2). Although the increase in the resistivity is probably somewhat exaggerated due to poor resolution of resistivity below the deeper conductor (no. 2), this is most likely an indicator of the heat extraction part of the geothermal system like is described in Figure 3. Calculation based on Eq. 1 using the same parameters as in Figure 3 indicate the resistivity in the up-doming body should be 50-100% higher than in the undisturbed surroundings.
- It should be noted here that in addition to the temperature effect on resistivity the general increase in degree of regional alteration increases the resistivity with depth. This is well demonstrated with a general increase in resistivity close to 2.5 km depth. According to mineral analysis from well LJ-8 the transition to the chlorite zone is at that depth (Kristmannsdóttir et al., 1977) where sharp increase in resistivity is expected (Árnason et al., 2000).
- In general, the resistivity picture presented in Figure 3 is easily explained by the combined temperature effect of a typical low temperature activity in the tertiary lava pile combined with the regional alteration processes. The exception is the low (no. 6) and high (no. 9) resistivity anomalies in the northern part of the section which is not easily explained.

A mentioned earlier, the above results are based on 1D inversion which fit well to other available data presented. 3D inversion also exists but we use the 1D model since resistivity logs from the boreholes fit much better to the 1D than the 3D sections.

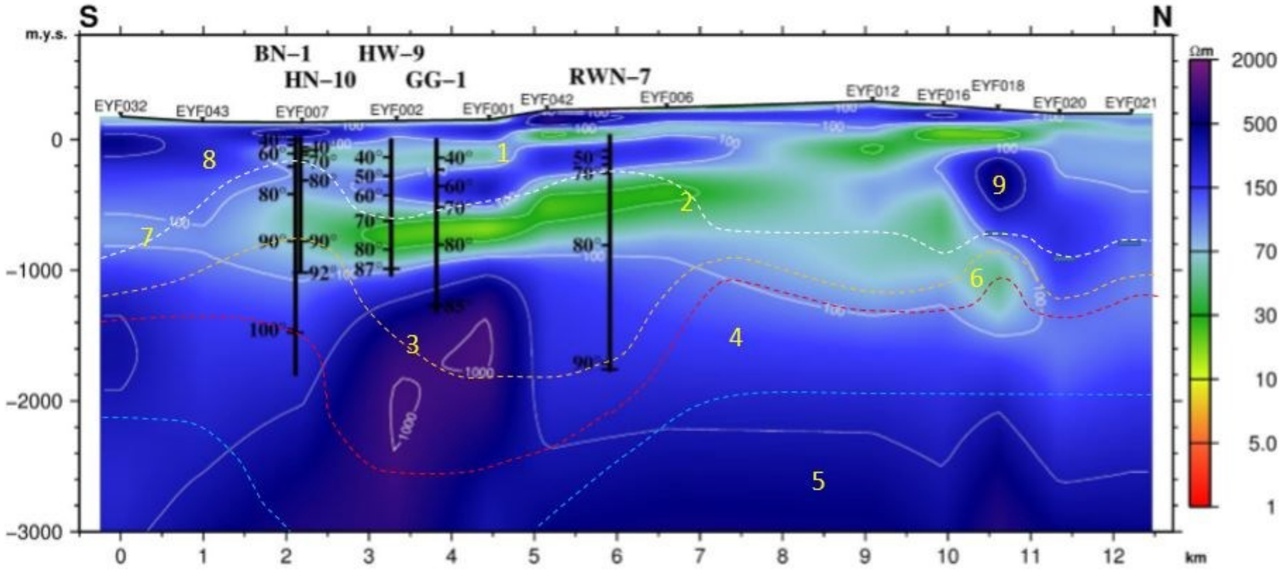

Figure 4. A resistivity and temperature cross section across the low temperature geothermal field in Hrafnagil-Botn in Eyjafjörður, N-Iceland (Figure 1). The resistivity model is from Vilhjálmsson and Karlsdóttir (2015) based on joint 1D inversion of TEM and MT data. The background colours denote the resistivity, the broken lines show the temperature, deep wells (BN-1, HN-10, HW-9, GG-1 and RWN-7) are shown by vertical black lines, and the ID-numbers of the MT soundings are marked as EYF032 etc. The yellow numbers refer to discussions in the main text. Surface geothermal activity (hot springs) is observed at several places between wells BN-1 and RWN-7. In addition, there is a single hot spring about 1 km south of MT sounding EYF012. The temperature in the wellfield is derived from temperature logs the deep wells, while the temperature north and south of it are extrapolated from 60-100 m deep heat flow wells.

5. RESISTIVITY INSIDE THE VOLCANIC RIFT ZONE

Most of the known high temperature fields in Iceland have been mapped by resistivity soundings while systematic resistivity mapping of the volcanic zone outside the high temperature field is scarce. They resistivity of the high temperature fields show generally a very typical structure of a shallow up-doming highly conducting layer covering a core of high resistivity. Next to the surface but outside hydrothermal activity very high resistivity is observed in fresh and unaltered lava at the surface. Outside the known high temperature fields but within the volcanic rift-zone, another layer of low resistivity has been observed at 5-10 km depth, deepening further away from the rift axis (e.g. Beblo and Björnsson, 1980).

The shallow low resistivity layer has been explained by high smectite content of the basaltic rock caused by hydrothermal activity. The actual resistivity of this layer depends on the amount of smectite. In the case of intense high temperature hydrothermal activity in the temperature range of 100-200°C, a large part of the rock mass is altered and the content of smectite is very high as displayed by high CEC values (Levy et al., 2018). The observed resistivity values can be as low as a few Ωm. In the case of a normal crustal accretion, as modelled by Pálmason (1973), and outside geothermal fields the smectite concentration is only modest and the typical resistivity values in the smectite zone is in the order of few tenths of Ωm. Very low resistivity values within the smectite zone (<20 Ωm) are therefore a sign of past or present hydrothermal activity.

The increase in resistivity below the conductive cap corresponds to the top of the more resistive chlorite alteration zone (e.g. Árnason et al. 2000) where smectite with is high CEC is replaced by chlorite with low CEC (Levy et al., 2018). It is known that this change in mineralogy occur at temperature close 230°C. Therefore, we can map the 230°C isothermal surface in the high temperature system if it has not cooled at later time.

In order to map the resistivity structure of the volcanic rift zone in more details a small part of the volcanic rift zone of NE-Iceland has been mapped by TEM/MT soundings (Vilhjálmsson and Flóvenz, 2017). A resistivity profile from this survey is presented in Figure 5. The profile extends eastwards from the Askja central volcano (Fig 1.) and its high temperature system in the middle of the volcanic rift zone to the eastern border of the volcanic rift-zone.

The resistivity section can be interpreted in the following way:

From surface to approximately 200 m depth: The resistivity is very high or of the order 103–104 Ωm. The thickness of this layer is variable from one place to another and it is interpreted as fresh lava, close to the surface and only partially saturated with water. The basaltic rocks have very high permeability in this depth range.

From 200 m to 800–1000 m depth: The resistivity is decreasing slowly with depth because of formation of expandable clay minerals on the walls of the pore space of the rock. Although there are no boreholes in the area, comparison with data from boreholes in similar geological settings lead to the conclusion that this depth range is still highly permeable and all heat brought form below through thermal conduction will be removed by strong groundwater flow (Flóvenz and Sæmundsson, 1993).

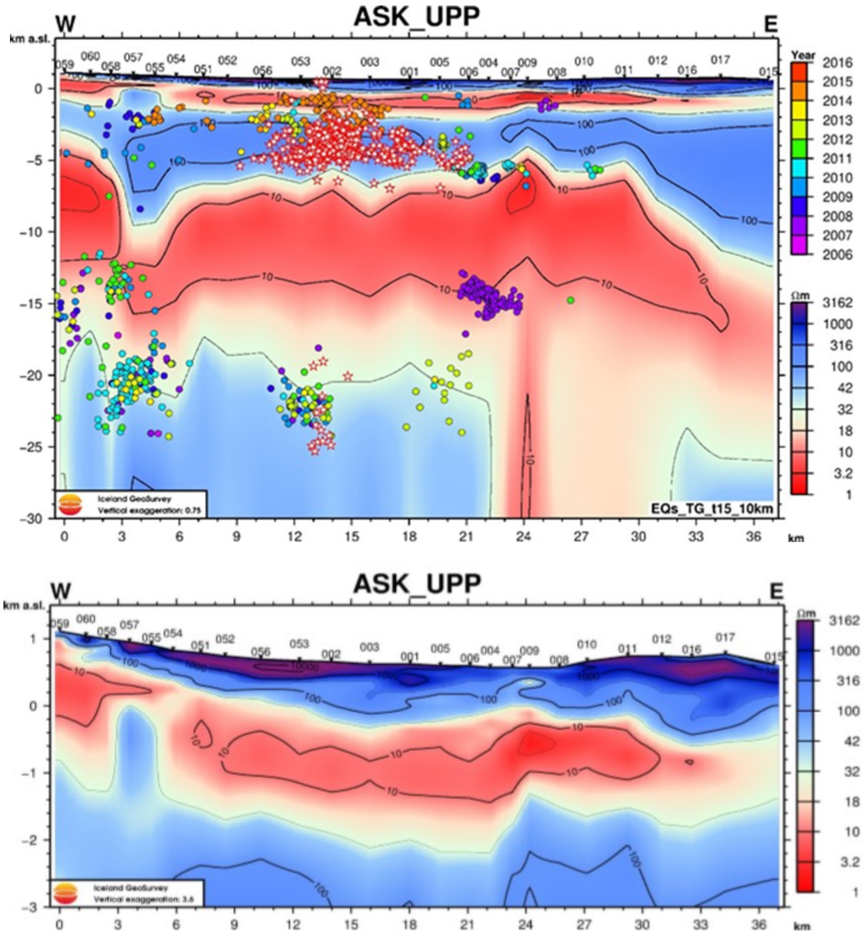

Figure 5. A West-East resistivity cross section from the Askja high temperature field located in the middle of the volcanic rift zone to the eastern border of it. The lower section is a blow-up of the shallower part. The colour coded dots on the figure denote hypocentres of microearthquakes from Greenfield et al. (2016) and the stars hypocentres from Soosalu et al. (2009). The deep conductive layer appears aseismic but at the same time S-waves from location beneath the layer propagate normally through it. This can be interpreted as the layer is close to the brittle ductile boundary but below the solidus of the material. The estimated temperature of the top of the deep conductive layer is therefore likely to be close to or above 500°C (Vilhjálmsson and Flóvenz, 2017). Hence, mapping the top of the deep conductive layer puts constraints on the temperature in the upper crust of the volcanic rift-zone. Note the up-doming of it under the high temperature field in Askja. Figure from Vilhjálmsson and Flóvenz (2017).

From 800–1000 m to 1900–2100 m depth: This depth range is characterized by sharp lowering of resistivity to values less than 10 Ωm at approximately 1400–1600 m depth indicating presence of hydrothermal activity. The layer is relatively flat but domes up almost to the surface within the Askja volcano where high temperature alteration is at the surface. This is the smectite layer that forms the cap rock covering the resistive core in the high temperature geothermal fields. The temperature necessary to form this layer is 100–230°C. This relatively flat layer suggests gradual subsidence on geological timescale of the whole area. At the eastern boarder of the volcanic zone the resistivity of the layer increases, suggesting either lower CEC due to less hydrothermal alteration or lower temperature in this part of the section.

At 1900–2100 m depth the resistivity starts to increase considerably. This corresponds to the boundary where chlorite with low CEC replaces the smectite with high CEC. This boundary is probably not very sharp but occurs through a zone of mixed clays where smectite and chlorite coexist. This change in mineralogy is estimated to occur at or around 230°C.

From 2–4.5 km depth the resistivity increases slowly according the resistivity model. In addition to the change in clay minerals at 2 km this increase is as predicted by the laboratory work of Nono et al. (2018), probably caused by onset of supercritical conditions.

At roughly 4.5 km depth the resistivity reaches maximum value of 100-200 Ωm and starts to decrease downwards. This corresponds to the change in conduction mechanism from fluid and surface conduction to mineral conduction that follows the Arrhenius law. This change in conduction mechanism is likely to take place at 400–600°C according to the laboratory results. If we assume the maximum resistivity value to be 150 Ωm resistivity at 4.5 km depth, the temperature to be 400°C and the activation energy to be 50 kJ/mol, based on lab measurements of Nono et al. (2018) on dolerites, we get the temperature of almost 700°C at the top of the deep low resistivity layer where the resistivity has dropped to 10 Ωm. These calculations are of course very rough, there are uncertainties in the resistivity values of the model in addition to the uncertainties in the parameters obtained by the lab measurements. But the calculations show that results from TEM/MT surveying combined with laboratory experiments on cores at simulated in-situ conditions and results from heat flow and seismic results can give a sensible coherent picture of the underground.

Around approximately 6–7 km depth the resistivity lowers rather sharply at the top of the deep conductive crustal layer. This depth corresponds also to the lower limits of earthquakes, interpreted as the brittle ductile boundary for strain rates typical for normal crustal stress accumulation (White et al., 2011). This might be a co-incidence. However, it raises the interesting question if the top of the deep conductive layer in Iceland marks the brittle ductile boundary. Other explanations like dehydration of amphibole might also apply as laboratory measurement indicate (Nono et al., 2018). The brittle ductile boundary has been estimated to occur at 750 ± 100°C in the Icelandic crust (Ágústsson and Flóvenz, 2005) so it marks the minimum value of the temperature of the conductive layer. Results from analysis of deep earthquakes in the area (Soosalu et al., 2009) show normal S-waves passing through the conductive layer which implies that the temperature of it is not close to partial melting. The IDDP-2 well at Reykjanes was drilled into an apparent aseismic volume in 2016 (Guðnason et al., 2016) but the circulation loss during the drilling clearly induced seismicity within this volume. It indicates that the rock in this volume deforms without faulting at normal strain rate but act as an elastic material when the stress accumulation is high as in the case of sudden cooling due to circulation loss. The real temperature at the bottom of the well at 4600 m depth has not be measured directly but was estimated 536°C from short term heat recovery measurements and >600°C from petrological and fluid inclusion studies (Elders et al., 2018, Bali et al., 2020, Zierenberg et al., 2020).

At 13–15 km depth the resistivity starts to increase again. This boundary is, however, very badly resolved due to the equivalence nature of the conductive layer; it might be thinner with lower resistivity or thicker with higher value. At similar depth, earthquakes start to occur again beneath the conductive layer. This suggests either very high strain rates that produce earthquakes despite the high temperature or, more likely, it represents different and stronger rocks and probably more resistive and denser than above.

6. CONCLUSIONS

We have applied to real exploration data on the results of earlier studies of the factors that influence the resistivity of the basaltic crust in Iceland and its geothermal fields. We show examples how resistivity section from high and low temperature fields in Iceland can be explained.

- We summarize the parameters that affect the resistivity of the basaltic crust according to earlier results and experiments.
- We predict that a typical low temperature field in Iceland should produce a low resistivity anomaly at shallow depth but a high resistivity anomaly deeper in the reservoir. This is based on temperature logs in deep boreholes in a low temperature area, estimation of the background heat flow and results of laboratory measurements of rock samples. We show a real example of this behaviour.
- We show how a resistivity section across the volcanic rift zone in N-Iceland can be explained in terms of lithology and resistivity variations with temperature. We propose that the highest resistivity in between the two low resistivity layers marks the chance from pore fluid and interface (surface) conduction in the rock to mineral conduction.
- We apply the laboratory measurements of resistivity of dolerite versus temperature by Nono et al. (2018) to estimate the temperature at the top of the brittle/ductile boundary that coincide with the top of the deeper low resistivity layer. The results give similar values as other estimates of the temperature at the brittle-ductile transition.
- This paper shows only a few examples how laboratory measurements can be used to convert observed resistivity values into important geothermal parameters that are valuable in geothermal exploration. But there is still a long way to go.
- The main results of this paper are, however, that the laboratory measurements of resistivity of rock samples at in-situ conditions are crucial to draw conclusions about geothermal reservoir properties from resistivity exploration with TEM and MT.

REFERENCES

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

Ágústsson, K. and Flóvenz, Ó.G.: The Thickness of the Seismogenic Crust in Iceland and its Implications for Geothermal Systems. *Proceedings of the World Geothermal Congress*. Antalya, Turkey. (2005), 9p.

Bali, E., Aradi, L.E., Szabo, A., Zierenberg, R.A., Friðleifsson, G.Ó. and C. Szabo, C.: Fluid Inclusions in the Deepest Part of the IDDP-2 Drill Core, Reykjanes Peninsula, Iceland. *Proceedings World Geothermal Congress*, (2020), (submitted).

Beblo, M. and Björnsson, A. 1980. Model of Electrical Resistivity Beneath NE-Iceland and Correlation with Temperature. *Journal of Geophysics*, **47**, (1980), 184-190.

Bjarnason, I.Th., Menke, W., Flóvenz, Ó.G. and Caress, D.: Tomographic Image of the Mid-Atlantic Plate Boundary in Southwestern Iceland. *J. Geophys. Res*., **98**, (1993), 6607–6622.

Böðvarsson, G.: 1982. Glaciation and Geothermal Processes in Iceland. *Jökull* **32**, 21-28.

Elders, W.A., Shnell J., Friðleifsson, G.Ó., Albertsson, A. and Zierenberg, R.A.: Improving Geothermal Economics by Utilizing Supercritical and Superhot Systems to Produce Flexible and Integrated Combinations of Electricity, Hydrogen, and Minerals. *GRC Transactions*, 42, (2018).

Flóvenz, Ó. G.: Seismic Structure of the Icelandic Crust Above Layer Three and the Relation between Body Wave Velocity and the Alteration of the Basaltic Crust. *Journal of Geophysics*, **47**(1–3), (1980), 211–220.

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

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*, The World Geothermal Congress, (2000), 1127-1132.

Flóvenz, Ó. G. and Sæmundsson, K.: Heat Flow and Geothermal Processes. *Tectonophysics*, 225, (1993), 123–138.

Greenfield, T., White, R. S. and Roecker, S.: The Magmatic Plumbing System of the Askja Central Volcano, Iceland, as imaged by Seismic Tomography**.** *Journal of Geophysical Research*: Solid Earth, **121**(10), (2016), 7211–7229

Guðnason, E. Á., Ágústsson, K. and Gunnarsson, K.: Seismic Activity on Reykjanes December 2015 – November 2016. *Iceland GeoSurvey, Report*, **ÍSOR-2016/090**, 45 pp.

Kaban, M. K., Flóvenz, Ó. G. and Pálmason, G.: Nature of the Crust-Mantle Transition Zone and the Thermal State of the Upper Mantle beneath Iceland from Gravity Modelling. *Geophysical Journal International*, **149**(2), (2002), 281–299.

Kristinsdóttir, L.H., Flóvenz, Ó.G., Árnason, K., Bruhn, D., Milsch, H., Spangenberg, E. and Kulenkampff, J.: Electrical Conductivity and P-wave Velocity in Rock Samples from High Temperature Icelandic Geothermal Fields. *Geothermics*, **39**, (2010), 94-105.

Kristmannsdóttir, H., Guðmundsson, Á. and Steingrímsson, B.: Borun við Ytra Laugaland í Eyjafirði, hola LJ-8. *Orkustofnun Report* **OS JHD 7724**, (1977), 71p.

Kulenkampff, J., Spangenberg, E., Flóvenz, Ó., Raab, S. and Huenges, E.: Petrophysical Parameters of Rocks Saturated with Liquid Water at High Temperature Geothermal Reservoir Conditions. *Proceedings World Geothermal Congress*, Antalya, Turkey (2005), 9p.

Kummerow, J., Raab, S., Schuessler, J.A. and Meyer, R.: Non-reactive and Reactive Experiments to Determine the Electrical Conductivities of Aqueous Geothermal Solutions up to Supercritical Conditions, *J. Volcanol. Geotherm. Res*. (2018), https://doi.org/10.1016/j.jvolgeores.2018.05.014.

Levy, L., Gibert, B., Sigmundsson, F., Flóvenz, Ó.G., Hersir, G.P., Briole, P. and Pezard, P.: The role of Smectites in the Electrical Conductivity of Active Hydrothermal Systems: Electrical Properties of Core Samples from Krafla Volcano, Iceland. *Geophys. J. Int.* **214**, (2018), 1558-1552.

Milsch, H., Kristinsdóttir, L.H., Spangenberg, E., Bruhn, D. and Flovenz, Ó.G.: Effect of the water steam phase transition in porous rocks on their electrical conductivity. *Geothermics*, 39, 1, (2010) pp. 106—114. DOI: http://doi.org/10.1016/j.geothermics.2009.09.001

Nono, F., Gibert, B., Parat, F., Loggia, D., Cichy, S.B. and Violay, M.: Electrical conductivity of Icelandic deep geothermal reservoirs up to supercritical conditions: Insight from laboratory experiments. *J. Volcanol. Geotherm. Res*., (2018) 15p. https://doi.org/10.1016/j.jvolgeores.2018.04.021.

Oxburgh, E.R. and Agrell, S.O.: Thermal Conductivity and Temperature Structure of the Reydarfjordur Borehole. *J.Gephys. Res*., **87** (B8), (1982), 6423-6428.

Pálmason, G., 1973. Kinematics and Heat Flow in a Volcanic Rift Zone, with Application to Iceland. *Geophys. J. R. Astr. Soc*., **33**, (1973), 451–481.

Pálmason, G. and Sæmundsson, K.: Summary of Conductive Heat Flow in Iceland. In: V. Cermák and L. Rybach (ed), *Terrestrial Heat Flow in Europe*, Springer Berlin, (1979), 219-220.

Revil, A. and Glover, P.W.J.: Nature of Surface Electrical Conductivity in Natural Sands, Sandstones and Clays. *Geophys. Res. Lett*., **25** (5), (1998), 691-694.

Soosalu, H., Key, J., White, R.S., Knox, C., Einarsson, P. and Jakobsdóttir, S.S.: Lower-crustal earthquakes caused by magma movement beneath Askja volcano on the north Iceland rift. *Bulletin of Volcanology*, **72**(1), (2009), 55.

Sæmundsson, K.: Outline of the Geology of Iceland, *Jökull* **29**, (1979), 7-28.

Sigurðsson, Ó. and Stefánsson, V.: Forðafræðistuðlar. Mælingar á bergsýnum. *Orkustofnun, Report*, **OS-94049/JHD-28 B**, (1994), 37p.

White, R. S., Drew, J., Martens, H. R., Key, J., Soosalu, H. and Jakobsdóttir, S. S. Dynamics of Dyke Intrusion in the Mid-Crust of Iceland. *Earth and Planetary Science Letters*, **304(3),** (2011), 300–312 https://doi.org/https://doi.org/10.1016/j.epsl.2011.02.038

Vilhjálmsson, A.M. and Flóvenz, Ó.G.: Geothermal Implications from a Resistivity Survey in the Volcanic Rift Zone of NE-Iceland and Comparison with Seismic Data." *Iceland GeoSurvey, Report,* **ÍSOR-20017/065**, (2017), 46p.

Vilhjálmsson, A.M. and Karlsdóttir, R.: MT og TEM í Eyjafirði Viðbótarmælingar 2013 og 3D úrvinnsla. *Iceland GeoSurvey Report* **ÍSOR-2015/074**, (2015) 81p.

Zierenberg, R.A., Friðleifsson, G.Ó., Elders, W.A., Schiffman, P., Fowler, A.P.L. and Reed M.: Active Basalt Alteration at Supercritical Conditions in IDDP-2 Drill Core, Reykjanes, Iceland. *Proceedings World Geothermal Congress*, (2020), (submitted).