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Námafjall High Temperature Field in N Iceland. A 3D Resistivity Model Derived from MT Data

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ABSTRACT

Resistivity surveys have been performed in the geothermal field in Námafjall since nineteen sixties, first with DC methods that just penetrated the mere surface but recently with TEM/MT that give information on the resistivity structure down to 10 km depth. In 2013 a 3D resistivity model derived from MT data was presented (Karlsdóttir et al. 2013) compiled from 69 MT soundings that had been static shift corrected by TEM. The model confirms a conventional resistivity structure with a low resistivity cap covering the geothermal system, underlain by a high resistivity core. It also reveals low resistivity anomalies that arch up under the geothermal system from a deep seated low resistivity layer at depth. These low resistivity anomalies may indicate the up flow of geothermal fluid into the geothermal system.

1. INTRODUCTION

The geothermal field, Námafjall in N Iceland lies within the southern extension of the fissure/fracture swarm, related to Leirhnúkur volcano and Krafla geothermal field. The geothermal field takes name after the mountain Námafjall (the mountain by the mines). Sulphur was mined from Námafjall for centuries and exported until the middle of nineteenth century. The main surface manifestation, sulphur and hydrothermally altered rocks are present in the slopes of Námafjall, and to the east of it in Hveraröند and to the west in Bjarnarflag. Hveraröند has a great tourist attraction as it holds an area with colorful surface manifestations, mud pits and steam vents, many of which are old shallow drill holes now covered with stones. The main well field until now is in Bjarnarflag and therefore the name Bjarnarflag is commonly used when referring the geothermal field. In this report, the name Námafjall is used for the gross geothermal field but the name Bjarnarflag applies to the geothermal area and well field west of the Námafjall mountain.

2. RESISTIVITY IN VOLCANIC GEOTHERMAL SYSTEMS

The main factors influencing resistivity in rocks are water content, salinity and temperature of the fluid, and the type of hydrothermal alteration of the rocks due to geothermal activity. In essence, water-saturated rocks conduct electrical currents more readily than dry rocks and conductivity increases with increasing temperature up to about 300°C (Violay et al., 2010). Geothermal systems can be distinguished from the surroundings because the electrical conductivity (resistivity) of certain clay minerals (phyllosilicates, such as smectite) found in fractures in the rocks are strongly temperature dependent. The electrical resistivity of the rocks is only weakly influenced by the salinity of the fluid, unless the salinity is very high and approaches that of seawater (Flóvenz et al., 2005).

Surface resistivity surveys of high-temperature geothermal systems in the basaltic rocks of the volcanic zones of Iceland (and where the host rocks are volcanic) always seem to commonly reveal similar resistivity structure which correlates to the distribution of alteration mineralogy. A low resistivity cap is observed on the outer and the upper margins of the reservoirs and is 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. The low-resistivity cap coincides with the smectite-zeolite zone and the transition to the more resistive core occurs within the mixed layer clay zone. Within the resistive core chlorite and epidote are the dominant alteration minerals. The alteration mineralogy is mostly influenced by temperature. This has the consequence that if the alteration is in equilibrium with present formation temperature, the resistivity structure can be interpreted in terms of temperature. The upper boundary of the low-resistivity cap is found where the temperature is in the range of 50–100°C and the transition to the resistive core occurs at temperatures in the range of 230–250°C (Árnason et al., 2000).

The resistivity reflects the hydrothermal alteration caused by the heating of the rocks and reflects the peak temperature experienced by the system, being it at the present or in the past. Thus, resistivity measurements reveal the alteration but do not indicate whether cooling has occurred after the alteration was formed because the resistivity profile only captures the alteration in the formation, irrespective of any later cooling of the system. The resistivity structure reflects the temperature, provided there is equilibrium between alteration and present temperature. In case of cooling the alteration may remain and the resistivity will reflect the temperature at which the alteration was formed. Whether the resistivity (and the alteration) indicates the present temperature of the system will only be confirmed by drilling.

Wherever MT measurements have been conducted in the volcanic zones in Iceland, a deep-seated low-resistivity layer is seen at a 10–15 km depth. (Eysteinsson and Hermance, 1985; Árnason et al., 2010) The upper boundary (10 Ωm contact) of this low-resistivity layer arches up to a depth as shallow as 2–3 km beneath high-temperature geothermal systems, e.g., in the Krafla area (Mortensen et al., 2009). As the low-resistivity layer is thought to reflect very high temperatures, it is interpreted as providing information about upwelling of heat into geothermal systems. Plume-like low-resistivity anomalies in limited areas beneath the deep low-resistivity layer, as seen in TEM and MT measurements at Upptyppingar (Vilhjálmsson et al., 2008), also support the idea of active up-flow of hot material (magma?).

3. RESISTIVITY SURVEY IN NÁMAFJALL

The first resistivity survey was conducted in Námafjall in 1970 using DC methods; Schlumberger- as well as Wenner configuration. Results showed low resistivity due to thermal alteration in the uppermost 300 meters of the system. In 1996 and 2005 TEM (transient electromagnetic) survey was conducted, succeeded by MT (magneto telluric) survey at same locations in 2009. The results of 1D inversion of the TEM and MT measurements revealed a conventional resistivity configuration with a low resistivity cap underlain by a high resistivity core and a deep seated low resistivity layer under Námafjall geothermal field, arching up under the area where most surface manifestations are present, in Námafjall and Hveraröndd. It reaches the depth of about 3 km under Námafjall and Hveraröndd (Karlsdóttir, 2011). In 2011 and 2012 more MT and TEM soundings were added resulting in a total of 69 TEM/MT sounding pairs available for 3D inversion.

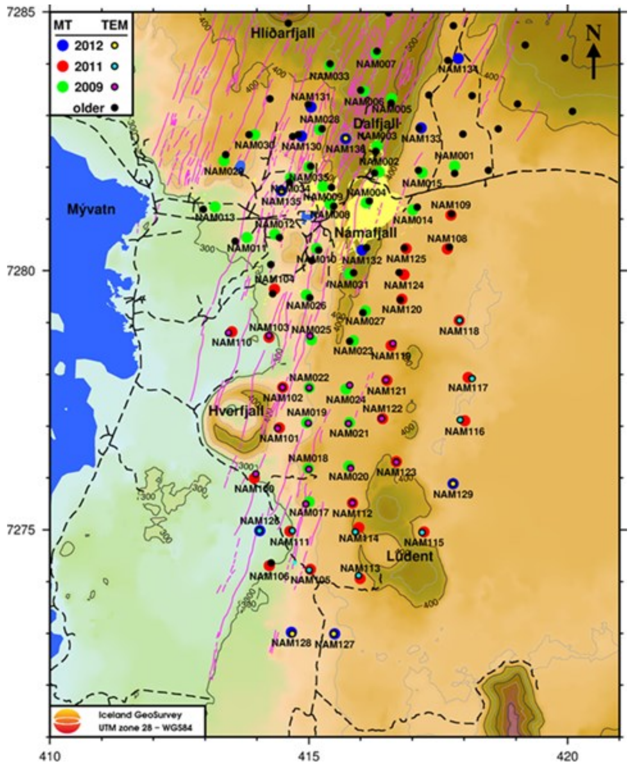

Figure 1 Námafjall. The survey area.

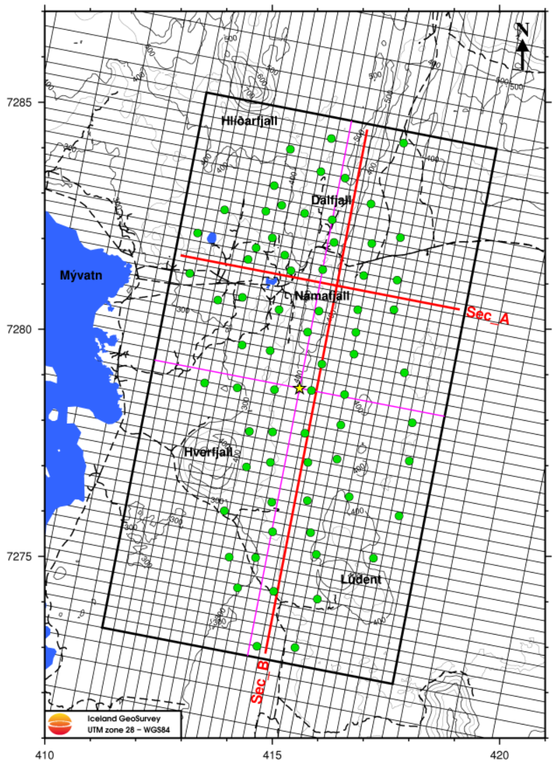

Figure 1 Námafjall. The survey area.

Figure 2 The model grid

The MT instruments used in this campaign are from Phoenix Ltd. in Canada (MTU type) and can measure the MT signals in the frequency range up to 1000 Hz and down to 0.00002 Hz (Phoenix Geophysics, 2009). Four sets of MT equipment were used in the field work. One served as a base station for remote reference processing of the data and was located outside the survey area. The remaining three MT units were operated in the investigation area, i.e. moved to a new location every day for recording till the next day, recording for 16-22 hours at each site. Three of the MT units measure five components (E_x, E_y, H_x, H_y and H_z); the base station and two of the others. The fourth station measures only the electric field, i.e. E_x, E_y. The two component unit is always set up close to a five component site (around 1 km away), and the magnetic field at that site used for the data processing. This approach is chosen if the magnetic field is almost identical at the two stations, valid for short distances. A reference station was set up at a distance from the survey area (10 – 300 km).

4. THE 3D MODEL

The 3D inversion was performed using the inversion program WSINV3DMT written by Prof. Weerachai Siripunvaraporn (Siripunvaraporn et al., 2005; Siripunvaraporn and Egbert, 2009). WSINV3DMT uses finite difference forward algorithm and utilizes a formulation of the inverse problem in the data-space rather than in the model-space. This reduces the dimensionality of the problem dramatically and makes 3D inversion of MT data attainable.

When running a 3D inversion an initial model is required to start the procedure. The initial model can influence the final result but that is further discussed in Section 4.3 and Chapter 5. The process of 3D inversion of MT data is a highly underdetermined problem, i.e. the number of unknown resistivity values is much higher than the number of data values. In the present case the number of data points is 8556 (69 soundings x 31 periods x 4 real and imaginary off-diagonal tensor elements, see below) but the model has 151536 unknown resistivity values (in the 74 x 52 x 28 blocks, see below) or almost 13 times the number of data points. The inversion therefore needs to be regularized by imposing constraints on the model (mathematically this means to make the model parameters interdependent in such a way that the number of the actually free parameters is reduced). This can be done by constraining the model parameters to vary smoothly, often referred to as minimum structure or Occam inversion (Constable et al., 1987). Another way of regularizing is using a reference or "prior" model and constrain the inversion model not to deviate too much from the prior model. Using a prior model offers the possibility of fixing some of the model parameters to a priory known values. The inversion code used here aims to minimize a penalty function that consists of a combination of these regularization methods, i.e. minimizes a weighted sums of the 1) difference between measured data and calculated response, 2) the roughness of the model and 3) the deviation from the prior model. Initially, the inversion process quickly adjusts the model to reduce severe misfit of the data. Later on, changes that would further reduce the data misfit are rejected because they make the model deviate too much from the prior model.

The user can adjust the smoothing criteria, but not the weight of the deviation from the prior model. This can constrict the inversion from fitting the measured data adequately, especially if a model that deviates considerably from the prior model is required. Running the inversion in steps, where the initial and prior models are updated at each step (the model that gave the best fit in the previous step), can facilitate the processes of fitting the data. In this way the limitation of the prior model is gradually relaxed until the data fit can no longer be improved.

WSINV3DMT assumes flat surface. This seems to be a limitation, but prior to the inversion, the MT data are corrected for static shift and this correction removes topographic effects in the data to a large extent (Árnason, 2008). The inversion is performed for the complex off-diagonal elements of the MT impedance tensor, i.e. 4 numbers (2 real and 2 imaginary parts) for each period of each sounding. The misfit measure is the RMS misfit of the observed and calculated tensor elements, weighted by the variance of the measured values.

4.1 The model grid

The model grid is set out so that the dense part of the grid covers the main area of data coverage. The model used for the inversion was run with mesh with 250 m grid pane spacing in the central area of the data coverage.

The 3D model consists of resistivity cubes in a 3D grid mesh defining the internal coordinate system. The mesh has 75 vertical grid planes in the x-direction and 53 vertical in the y-direction and 29 horizontal grid planes. The grid is dense in the area of interest with grid plane spacing of 250 m in the area of the data coverage. Outside the dense area the grid spacing increases exponentially to the edges at ±137,268 km and ±134,518 km in the x- and y-directions, respectively. Figure 2 shows a horizontal slice of the central part of the model grid mesh and the location of the MT soundings in the grid. Red star shows the origin (middle point) of the grid. The horizontal grid planes are likewise dense at shallow depth but eventually with exponentially increasing spacing to the bottom at the depth of 160,684 km. The shallowest layer thicknesses are 16, 26, 36, 50, 76, 100, 158, 200 m etc.

4.2 Initial and prior models

3D inversion of MT data is a highly underdetermined problem. The inversion is started from an initial model and in order to regularize the inversion, a prior model is used to constrain the deviation of the resulting model from the prior model. As a consequence of this the resulting model will depend on the initial and prior model.

Some components of the prior model can, in some cases be assumed to be known a priory and even fixed in the inversion. In the case of the Námafjall area, the proximity of the sea (ca. 40 km) will probably not have a great influence but will be taken into account as it may have some impact in the MT data, at long periods. The sea has therefore to be taken into account in the inversion and the resistivity of the model cells in the sea were assigned the average resistivity of seawater (0.3 Ωm) and the inversion is forced to keep it fixed. The proximity to the Mývatn Lake was not considered in the inversion, as the lake is not saline and is very shallow.

To investigate the influences of the initial model on the resulting models, inversions with five distinct initial models were done. In all cases the same model was used as initial and prior model. Out of these five inversions, three results will be discussed in this report:

(1) A model compiled from joint 1D inversion of individual TEM/MT sounding pairs (Karlsdóttir, 2011).

(2) A homogeneous half-space model with resistivity 70 Ωm.

(3) A model compiled from the resulting 3D model from (1) down to 3196 meters depth and a homogeneous half-space of 71 Ωm below that (referred to as “Final 1D / Homog. Earth, 71 Ωm” in chapter 4).

As discussed above, a step procedure was used during all the inversion. The inversion was allowed to run for 5 iterations at each step and restarted. This stepwise inversion (and relaxation of the prior model) was continued until the data fit could not be improved any more (4–7 steps).

4.3 The inversion

The inversion program was executed using a parallel processing version of the WSINV3DMT code using the Message Passing Interface (MPI) parallel computing environment (Siripurvaraporn, 2009). It was executed on a 32 core computer with 132 GB memory. The inversion is a very heavy computational task and each iteration with the 250 m grid spacing, took about 7-8 hours and the total computing time was more than 250 hours.

The data misfit is defined as the RMS (Root-Mean-Square) of the difference between the measured and calculated values, weighted by the variance of the measured values. In all cases the off diagonal tensor elements (real and imaginary parts) were used.

For the initial model compiled from the 1D model, the initial misfit was 5.32 and the lowest obtained misfit was 1.74. This model obtained the best fit to data of all models and was selected as the resulting model.

5. THE 3D RESISTIVITY MODEL

The 3D inversion model reveals a conventional resistivity structure of a geothermal field in a volcanic rift zone i.e. a low resistivity cap covering a high resistivity core and a low resistivity anomaly at depth, indicating the heat source of the geothermal system.

Figures 3 - 7 display the resistivity structure of Námafjall, both as resistivity maps and as resistivity sections. Location of the resistivity cross sections is shown on Figure 2.

Figure 3 shows the resistivity at 200 m a.sl. It displays the top of the low resistivity cap at 300 m depth under surface where it surrounds the top of the high resistivity core. The low resistivity cap emerges up to the surface in Námafjall mountain, as can be seen on the resistivity cross section at Figure 5 but at 200 m a.sl. the top of the high resistivity core is starting to emerge. From this depth down to approximately 1000 meters depth under surface (500 m b.sl.) the low resistivity cap tilts down to all directions. The resistivity in the cap is < 10 Ωm to all directions except to the north where the low resistivity cap of Námafjall meets that of Krafla geothermal system. The low resistivity caps connect in a layer with somewhat higher resistivity than typical of a geothermal low resistivity cap, but still underlain by higher resistivity as seen on Figure 7 (Karlsdóttir 2002).

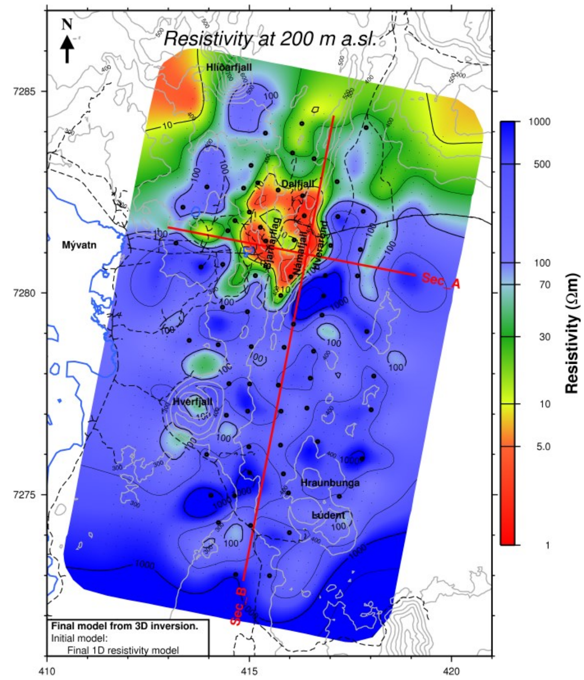

Figure 3 Resistivity at 200 m a.sl.

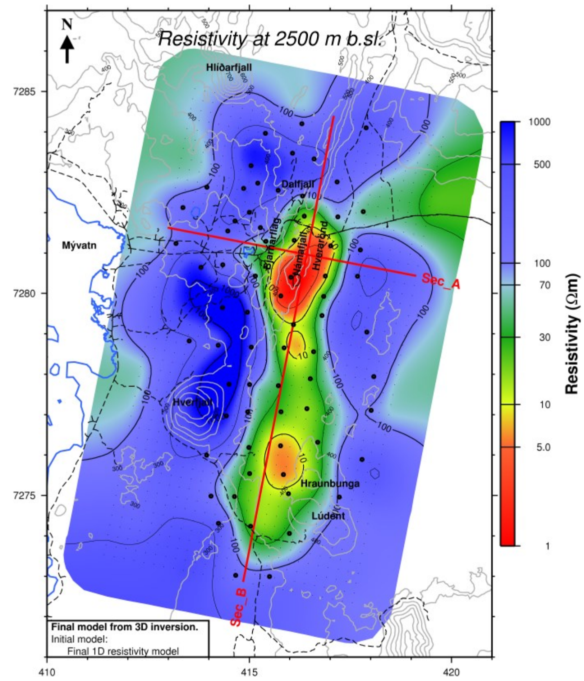

Figure 4 Resistivity at 2500 m b. sl.

There is a sharp change in resistivity from the low resistivity cap to the high resistivity core, where the resistivity exceeds 100 Ωm. A low resistivity layer is seen at depth (8 – 13km) that arches up under the geothermal system. This low resistivity anomaly is considered as the indication of the heat source of the geothermal system. It arches up to approximately 1500 meters b. sl. or approximately 2000 meters below surface under Námafjall mountain (Figure 5, 6 and 7). The anomaly has a more sharp boundary to the east, see Figure 6, than to the west, where there is an extension of the low resistivity anomaly to the south west under an area named Jarðbadshólar. This may indicate more up flow into the geothermal system to the immediate west of Námafjall mountain and indeed this is confirmed by comparison to temperature logs.

Another up-doming of the low resistivity anomaly is encountered within the fracture system of the geothermal system approximately 5 km to the south of Námafjall mountain in the vicinity of Hverfjall mountain. The anomaly arches up to 3000 m.b.sl. or about 1500 meters lower than the one under Námafjall (Figure 7) and indicates a second heat source or up flow into the geothermal system. The resistivity map at 2500 m b. sl. on Figure 4 shows the extent of the up flow zones at this depth i.e. the main up flow under Námafjall and Jardbadshólar and a potential up flow east of Hverfjall.

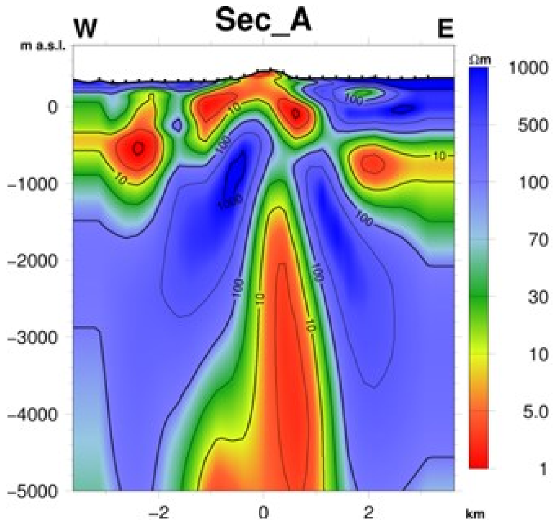

Figure 5. EW resistivity cross section A through Námafjall.

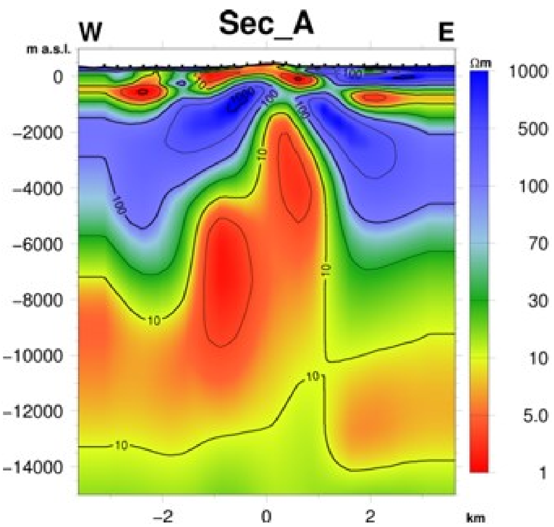

Figure 6. EW resistivity cross section A through Námafjall.

5.1 The resistivity model and well logs

The available resistivity well logs have been presented along with the 3D resistivity model in PETREL visualization software. Other data incorporated into Petrel are: Data on petrology from the cuttings, data on alteration minerals/zones from X-ray diffraction, temperature distribution from temperature logs as well as inferred data from the inspection of the wells, such as location of feed zones, fractures and faults. Figure 8 shows EW section through Námafjall and the well field, Bjarnarflag from Petrel for comparison of the resistivity model and alteration in wells as well as temperature logs from wells. On the left hand side the well trajectories show the alteration in the wells derived from X-ray diffraction. There is good accordance between the alteration in wells and the 3D resistivity model. The smectite zone falls clearly within the low resistivity cap and the mixed layered clay zone marks the boundary between the low resistivity cap and the high resistivity core beneath. The chlorite/epidote zone marks the high resistivity core. On the right hand side the well trajectories show the temperature in the wells. Highest temperature, 335°C, is encountered in well BJ-13 that aims at the top of the main low resistivity anomaly. Temperature as high as 330 °C is encountered in well BJ-14, aimed above the anomaly under Jarðbaðshólar, adjacent to the main anomaly (Figure 6 and 8). These two wells are the ones with highest temperature in the well field and are the ones aimed at the low resistivity anomalies that are thought to be the indicators of geothermal up flow into the geothermal system.

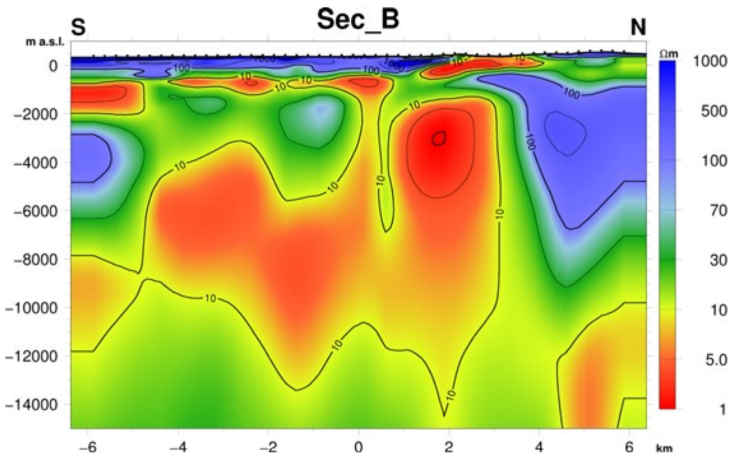

Figure 7. NS resistivity cross section B through Námafjall. Location see Figure 2.

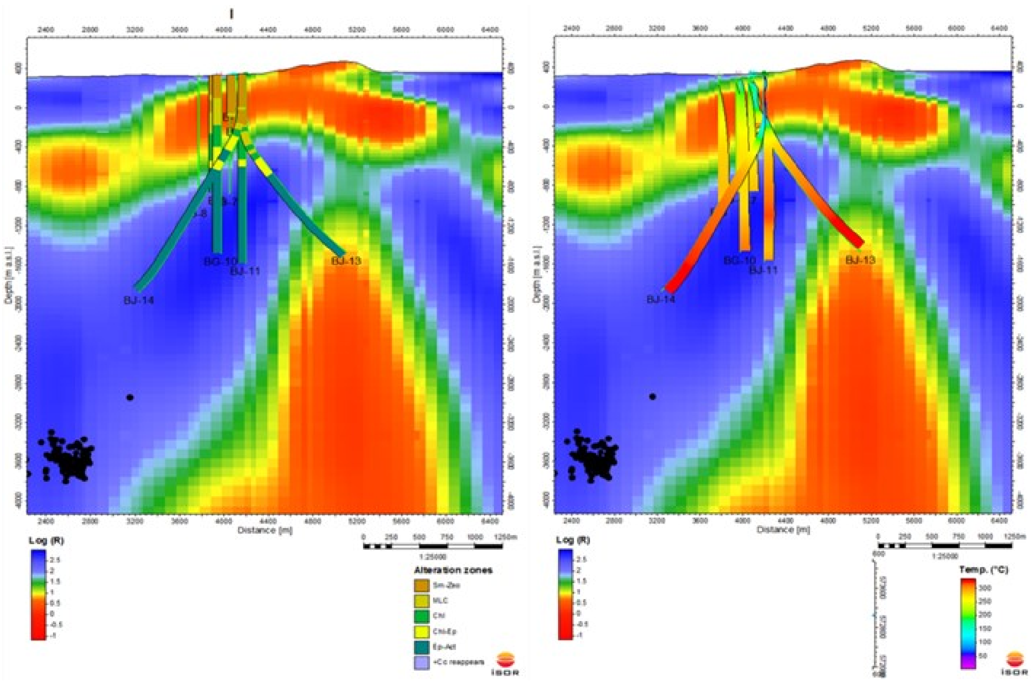

Figure 8 A visualization of the 3D resistivity model and data from well logs from Petrel. Resistivity cross section A and nearby wells.

CONCLUSION

The TEM/MT survey in Námafjall reveals a conventional resistivity structure of a high temperature field i.e. a low resistivity cap underlain by a high resistivity core and a deep seated low resistivity body at depth indicating the heat source.

- A low resistivity cap that reflects the zeolite/smectite zone covers the whole area. It reaches surface in Námafjall, Bjarnarflag and Hveraröndal. It tilts down towards all directions except to the north. Its upper boundary levels at approximately 500 m b.sl. To the south, where the data coverage is best, the cap levels at 500 m b.sl. between Námafjall and Hverfjall south of which it tilts further down to approximately 1000 m b.sl.

- To the north, the low resistivity cap connects to a layer with low resistivity (12 – 20 Ωm) in the area where Námafjall and Krafla Geothermal fields meet. This low resistivity layer has a character of a low resistivity cap with an underlying higher resistivity that connects to the high resistivity cores of both geothermal systems.

- A high resistivity core reflecting the chlorite/epidote zone underlies the low resistivity cap. The margin between the two reflect the 230 – 240°C temperature boundary provided there is a thermal equilibrium between thermal alteration and temperature at present within the geothermal system. The high resistivity core rises highest to 150 to 200 m a.sl. below Námafjall mountain.

- In all MT surveys conducted in Iceland, a deep seated low resistivity layer has been encountered at 8 to 15 km depth b.sl., except in Reykjanes peninsula and close to the coast line. Most MT surveys have been conducted in geothermal fields and within the volcanic zone. The deep seated low resistivity layer domes up to a shallower depth beneath high temperature fields. At Námafjall it can be seen at 1500 m b.sl. or at approximately 2 km depth below surface. The up doming of the deep low resistivity layer is considered to indicate the main up flow of heat into the geothermal system.

- An up flow of heat into the system is indicated by two anomalies, one under Námafjall and another south east of Hverfjall. The Námafjall low resistivity anomaly domes up to 1500 m b.sl. under Námafjall mountain with an adjacent anomaly under Jardbadshólar doming up to 4000 m b. sl. The Námafjall anomaly seems to indicate the main up flow zone into the geothermal system. The Hverfjall low resistivity anomaly domes up to 3000 m b.s.l. just south east of Hverfjall and may indicate a second up flow zone into the geothermal system.

- The 3D model run with the 1D input model had the best fit of the data (RMS fit 1.76). Running input models with different initial resistivity did not give resolution at depth below 3 km b.sl. and did not get as good RMS data fit.

- All available resistivity well logs have been presented along with the 3D resistivity model in PETREL visualization software. Other data now being incorporated into Petrel are: Data on petrology from the cuttings, data on alteration minerals/zones from X-ray diffraction, temperature distribution from temperature logs as well as inferred data from the inspection of the wells, such as location of feed zones, fractures and faults. Comparison of the 3D resistivity model to well data from well logs shows good accordance between resistivity and alteration of the rocks, as well as between resistivity and temperature as displayed in 5.1.

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