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## **GEOPHYSICAL METHODS USED IN GEOHERMAL EXPLORATION**

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### **ABSTRACT**

Geophysical methods are one of the three main disciplines applied on the surface to explore geothermal resources, including also geology and the chemistry of thermal fluids. The most successful methods are aimed at parameters that are directly influenced by the geothermal activity, such as resistivity methods and thermal methods, and are usually referred to as direct methods. Other methods explore the physical parameters of the host rock, including magnetic properties, density and seismic velocity, and are usually referred to as indirect or structural methods. Short description is given on the most important methods, their strength and limits discussed and examples given on their use.

### **1. INTRODUCTION**

Geophysical exploration of geothermal resources deals with measurements on the physical properties of the earth. The emphasis is mainly on parameters that are sensitive to temperature and fluid content of the rocks, or on parameters that may reveal structures that influence the properties of the geothermal system. The aim can be to:

- delineate a geothermal resource;
- outline a production field;
- locate aquifers, or structures that may control aquifers in order to site wells; or
- assess the general properties of the geothermal system.

The important physical parameters in a geothermal system are:

- temperature;
- porosity;
- permeability;
- chemical content of fluid (salinity); and
- (pressure).

Most of these parameters cannot be measured directly through conventional geophysical methods applied on the surface of earth. On the other hand, there are other interesting parameters that can be measured which are linked with the parameters above and may thus give important information on the geothermal system. Among these parameters are:

- temperature (°C);
- electrical resistivity ( $\Omega\text{m}$ );
- magnetisation ( $\text{Vs/m}^2$  or T);
- density ( $\text{kg/m}^3$ );
- seismic velocity (km/s);
- seismic activity;
- thermal conductivity ( $\text{W/mK}$ ); and
- streaming potential (V).

A distinction is usually made between direct methods, and indirect or structural methods. The direct methods give information on parameters that are influenced by the geothermal activity, while the structural methods give information on geological parameters which may reveal structures or geological bodies that are important for the understanding of the geothermal system. The direct methods include thermal methods, electrical (resistivity) methods and self potential (SP), while the structural methods include magnetic measurements, gravity measurements, active seismic methods and passive monitoring of seismicity.

Different methods may be applied for the exploration of low-temperature fields compared to high-temperature resources. Furthermore, different methods are sometimes used from one country to another despite similar geothermal surroundings, based on the routines that have been developed at the different institutions. It is also important to combine different methods, as relying on the results of measurements of a single parameter usually does not give adequate information for good understanding of the geothermal system.

In the following sections the most important methods used in geophysical exploration of geothermal fields are described briefly, and examples given on their use. For a more thorough overview, see e.g. Hersir and Björnsson (1991) or Keary and Brooks (1992); and also Árnason and Flóvenz (1992).

## 2. THERMAL METHODS

Thermal methods include direct measurements of temperature and/or heat, and thus correlate better with the properties of the geothermal system than other methods. However, as a (near-) surface method they are limited to shallow levels. To measure temperatures close to the surface, in the uppermost metre or so, is fairly simple. Knowledge about status at deeper levels is based on the existence of wells, usually shallow gradient wells (e.g. 30-100 m deep), from which the thermal gradient can be calculated and possibly the depth to the exploitable geothermal resource. Drilling is though usually fairly expensive, and puts practical limits to the use of the method. Furthermore, shallow wells are not always adequate to get reliable values on the thermal gradient.

The heat exchange mechanism in the earth is important for interpretation of thermal methods. A distinction is made between:

- Conduction, which is based on atomic vibrations, and is important for transfer of heat in the earth's crust;
- Convection, which transfers heat by motion of mass, e.g. natural circulation of hot water; and
- Radiation, which does not influence geothermal systems.

The simplified geothermal relationship for thermal conduction is (conductive heat transfer only):

$$Q_{\text{cond-z}} = -k \Delta T / \Delta z$$

The parameter  $k$ , the thermal conductivity ( $\text{W/m}^\circ\text{C}$ ), is a material constant, which ranges between 1 and 5  $\text{W/m}^\circ\text{C}$ , with the low values usually associated with sedimentary formations and the higher for crystalline rocks. The thermal gradient,  $\Delta T / \Delta z$ , gives information on the increase of temperature with

depth, and its distribution can be important information for understanding and delineation of the geothermal resource, both on a regional scale and local scale. If the conductive heat transfer,  $Q$ , is 80-100 mW/m<sup>2</sup> or higher, it may indicate geothermal conditions in the subsurface. For convection, a distinction is made between free convection, which is driven by density gradients in the fluid (due to a difference in temperature), and thus the heat is transferred, while forced convection is driven by an external pressure gradient, such as hydrostatic head. Convection is a much more powerful way of transmitting heat, but relies on the permeability of the rocks. High-temperature geothermal systems are based on the free convection of the geothermal fluids, while geothermal systems in sedimentary basins are based on conductive heat flow. However, geothermal systems are often of a mixed type.

Despite their limitations and their dependence on information from wells, thermal methods are important in geothermal exploration. They include the following:

- Mapping of thermal distribution at the surface, including:
  - Detailed geothermal surface mapping (GPS);
  - Soil temperature measurements in the uppermost metre or so (Figure 1);
  - Airborne IR survey.
- Temperature measurements in 20-100 m gradient wells, used to delineate regional or local gradient anomalies (Figure 2).
- Heat flow surveys for regional assessment for assessing thermal output of areas; here thermal conductivity measurements may be necessary and possibly terrain correction of the gradient survey.

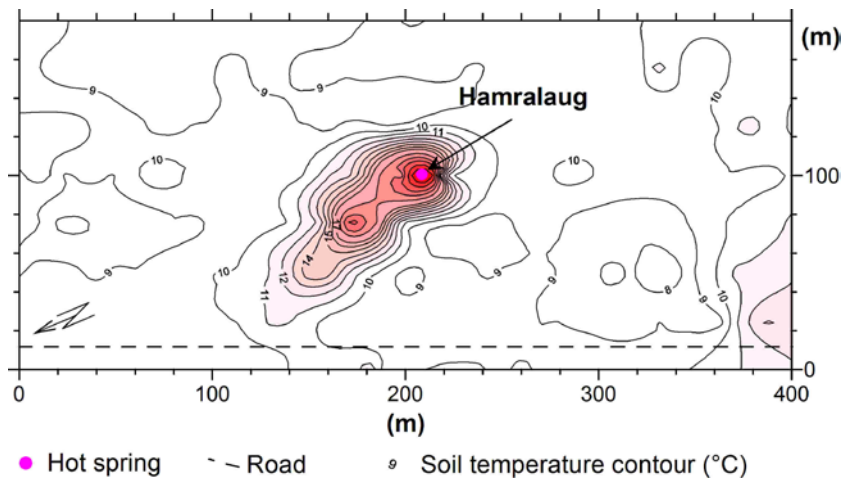


FIGURE 1: Soil temperature measurements at 0.5 m depth at Asgardur in Borgarfjordur, W-Iceland, where the temperature contours reveal a linear distribution, believed to be caused by a fracture (Ganbat, 2004)

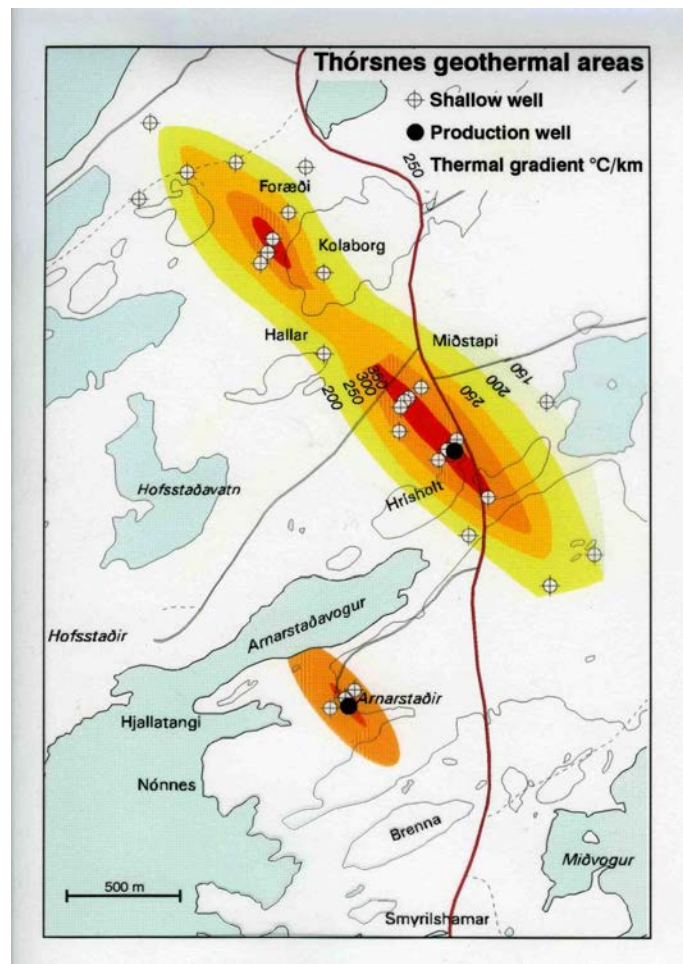


FIGURE 2: Gradient map of from the Thórsnes geothermal area, W-Iceland, where no geothermal activity was known on the surface; the gradient anomaly revealed two permeable fractures which proved very productive when drilled into (Saemundsson, 2008)

### 3. ELECTRICAL METHODS

#### 3.1 Introduction

Electrical methods or resistivity methods are the most important geophysical methods in the surface exploration of geothermal areas, and as such the main methods used in delineating geothermal resources and production fields. The parameter of interest is the electrical resistivity of the rocks which correlates both with the temperature and alteration of the rocks which are key parameters for the understanding of the geothermal systems.

The main principle is that electrical current is induced into the earth which generates an electromagnetic signal that is monitored at the surface. The basic relationship behind resistivity measurements is the Ohm's law, which states that:

$$\mathbf{E} = \Delta \mathbf{j}$$

where  $\mathbf{E}$  is the electrical field strength (V/m);

$\mathbf{j}$  is the current density ( $\text{A/m}^2$ ); and

$\Delta$  is the *electrical resistivity* ( $\Sigma m$ ), which is a material constant.

For a unit cube/bar, the relationship for resistivity is defined as:

$$\Delta = V / I$$

The reciprocal of resistivity is *conductivity* (S/m), thus it is also possible to talk about conductivity measurements. However, in geothermal, the tradition is to refer to electrical or resistivity measurements.

Electrical methods include many different types of measurements and varying setups or configurations for the different types. The most important types are:

- *DC methods*, where current is generated and injected into the earth through electrodes at the surface. The measured signal is the electrical field generated at the surface.
- *TEM*, where current is induced by a time varying magnetic field from a controlled source. The monitored signal is the decaying magnetic field at surface from the secondary magnetic field.
- *MT*, where current is induced by the time variations in earth's magnetic field. The measured signal is the electromagnetic field at the surface.

#### 3.2 Resistivity of rocks

Electrical resistivity of rocks in geothermal surroundings is a parameter which reflects the properties of the geothermal system, or its history. Thus, a good knowledge on the resistivity is very valuable for the understanding of the geothermal system. This relates to the fact that *the resistivity of rocks is chiefly controlled by parameters that correlate to the geothermal activity*, such as:

- Porosity and pore structure, where distinction is made between:
  - Intergranular porosity such as in sedimentary rocks,
  - Fracture porosity, relating to tension, fracturing or cooling of igneous rocks,
  - Vugular porosity which relates to dissolving of material (limestone) or gas content (in volcanic magma);
- Alteration of the rocks, lining the walls of the pores, often related to as water-rock interaction;
- Salinity of the fluid in the pores;
- Temperature;
- Amount of water, i.e. saturation or steam content; and
- (Pressure).

The four listed first, namely fracture (secondary) porosity, alteration, salinity of the fluid, and temperature, are the most important ones, and basic parameters for a geothermal system. This explains why the parameter “*resistivity of the rocks*” is so important in geothermal exploration, and especially in volcanic surroundings, and hence the important role of resistivity soundings.

Generally, it can be said that electrical conduction is mainly through interconnected water-filled pores. If the rocks are fresh, the conduction is mainly through the water, while the alteration lining the walls of the pores is the decisive factor, when created at temperatures between 50 and 200°C, due to its very conductive properties. At higher temperatures resistive alteration takes over, changing the decisive conduction carrying mechanism again to the water. Figure 3 summarizes this and shows how the resistivity of water-bearing rocks changes with the alteration and temperature.

High-resistivity core within a low-resistivity coat is typical for most high-temperature geothermal systems and has to do with the resistive alteration minerals that form at higher temperatures. Further discussion on resistivity of rocks and its correlation to the properties of geothermal systems will not be given here, but the reader is referred to Árnason et al. (2000) for further information.

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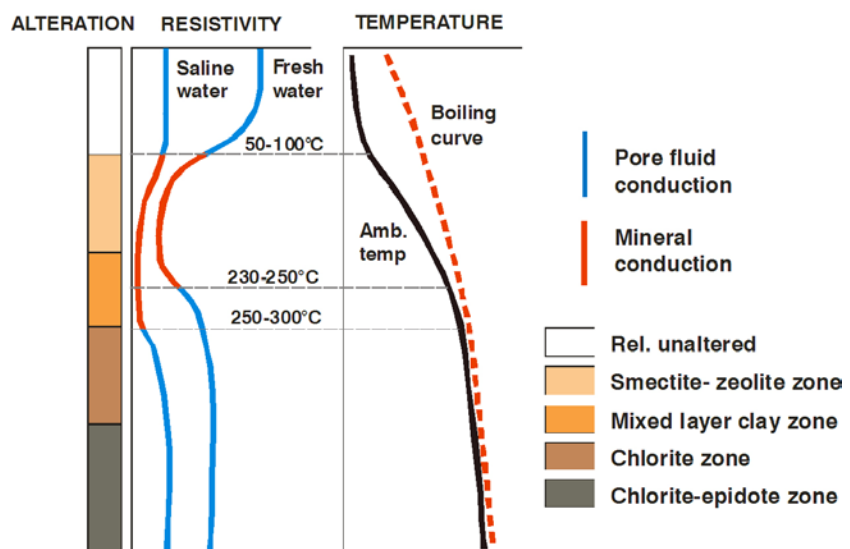


FIGURE 3: Resistivity of rocks and its association with alteration and temperature (Árnason et al., 2000)

### 3.3 DC-measurements

DC measurements are the oldest of the resistivity methods, at least as methods used for geothermal exploration and were the routine methods in the 1950s -1980s, but their application is not as common nowadays. Most of the configurations rely on *two pairs of electrodes* – one pair for current transmission and the other for measuring the potential difference. The geographical distribution defines the type. A distinction is made between sounding and profiling. A sounding, which is much more common to use, has a fixed centre and the varying of the electrode spacing is used for mapping resistivity changes with depth, while in profiling, the electrode distances are fixed, and the whole array moved along the profile line for mapping lateral changes. The most common configurations at least in geothermal exploration are:

- *Schlumberger sounding*, which has been widely used for a long time and is still the most popular one. The electrodes are on a line, and the set up is mirrored around the centre. The pair of potential electrodes is kept close to the centre, while the pair of current electrodes is gradually moved away from the centre, for the current to probe deeper and deeper into the earth.
- *Dipole sounding or profiling*; here various arrays exist, many used quite extensively in the 1970s into the 1980s.
- *Head-on profiling*, a successful method for locating near-surface vertical fractures or faults. It is really a variety of the Schlumberger profiling method with a third current electrode located far away at a right angle to the profile line.

Necessary equipment for DC soundings is a good current transmitter, a sensitive voltage receiver, wires on reels, electrodes, and a power source. By using (car) batteries as a source for the current



transmitter, the equipment is not very bulky, and usually possible to carry on foot for measurements in areas where access by car is not possible. Good DC equipment is marketed by many companies selling geophysical instruments.

The earth is not homogeneous so what we are looking at is a sort of average resistivity of the earth below the measuring site within a certain depth range. This “average” resistivity is usually referred to as *apparent resistivity*. For a sounding, apparent resistivity is plotted as a function of  $AB/2$  (the length of the current arm) with increasing electrode separation, usually on a bilogarithmic scale, where the length of the current arm relates to the depth penetration. Schlumberger soundings are usually extended to a maximum current arm,  $AB/2$ , of 1-2 km. Much longer current arms have been used, but the long wires for longer distances can be difficult in practice. 1D interpretation is simple but 2D or even 3D interpretation of the resistivity bodies often necessary, and much more complicated.

Figure 4 shows a regional resistivity map from Husavik, N-Iceland based on Schlumberger soundings reflecting the resistivity at 500 m below sea level. It shows well a defined low-resistivity anomaly associated with the powerful Hveravellir low-temperature area.

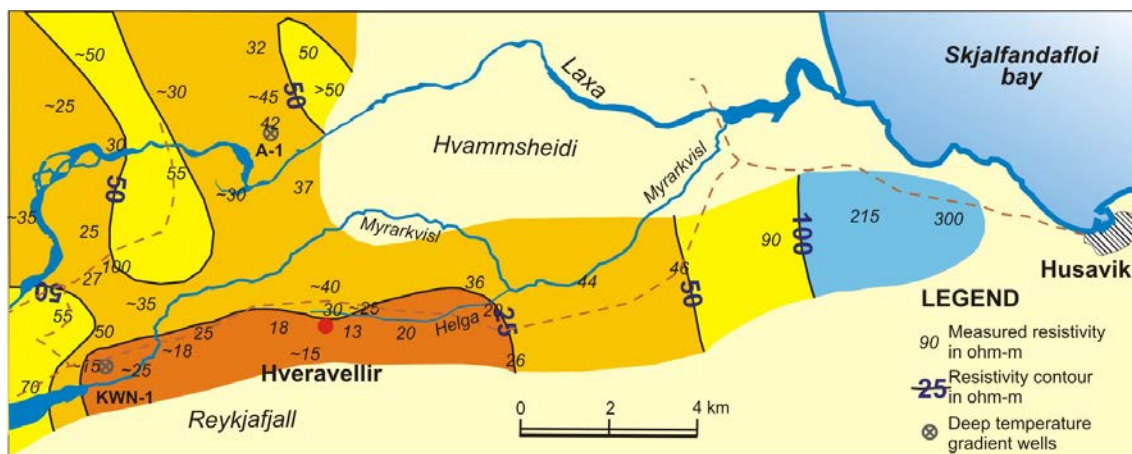


FIGURE 4: Resistivity map at 500 m depth b.s.l. from the Husavik area, N-Iceland, outlining the powerful Hveravellir low-temperature area (Georgsson et al., 2005)

### 3.4 TEM measurements

In the late 1980s, a new method started to make its impact in geothermal exploration and gradually it has replaced Schlumberger sounding as the routine method in the geothermal exploration of the uppermost 1-1.5 km of the earth. This is the TEM method (or Transient Electro-Magnetic method), which is an electromagnetic method which uses a controlled-source to create the signal to be measured. In the *Central loop TEM sounding method* (Árnason, 1989), referred to as TEM from now on, a constant magnetic field is built up by transmitting current  $I$  through a big loop (grounded dipole). Then the current is abruptly turned off. A secondary field is thus induced, decaying with time. This decay rate is monitored by measuring the voltage induced in a receiver coil in the centre of the loop on the surface. Current distribution and decay rate recorded as a function of time depend on the resistivity structure below the measuring site, and can be interpreted as such. The signal can also be based on a grounded dipole to create the primary magnetic field. TEM data are presented on a bilogarithmic scale as DC data, but here the apparent resistivity is plotted as a function of time after the current was turned off.

TEM equipment is sophisticated and relatively expensive, at least compared to DC equipment. Besides the receiver including a datalogger, a transmitter is required connected to a good electric generator able to generate high currents (the order of 10 A) and thus a strong magnetic field through the transmitting loop. The heavy generator makes it difficult to do measurements by carrying the

equipment on foot. In addition, a measuring coil and wires for the loops are needed with the current wires being able to carry heavy currents. In practice, the big transmitting loop is usually a square with a side of 150-300 m. With good conventional TEM equipment and a transmitter loop of 300×300 m, with TEM it is possible to study resistivity structures down to depth levels of 1-1.5 km.

TEM measurements have many advantages over Schlumberger measurements, such as in less need for manpower, both in data acquisition and interpretation, and in not being as sensitive for lateral inhomogeneities, and in improved resolution. Thus 1D interpretation of data is usually adequate, as the method is much more focussed than DC measurements.

Figure 5 shows a TEM resistivity cross-section from the large Hengill high-temperature area. The profile crosses an area where geothermal activity is not seen on the surface but the low-resistivity body with a high-resistivity core seen on its western (left) side indicates that this body has reached temperatures above 250°C, reflecting a suspected new production field. Figure 6 shows a TEM resistivity map of the Hengill area at a depth of 600 m b.s.l. The map delineates a wide spread low-resistivity layer but also the higher resistivities found in the central part of the geothermal system, indicating temperatures at or above 250°C in an area covering tens of square kilometres.

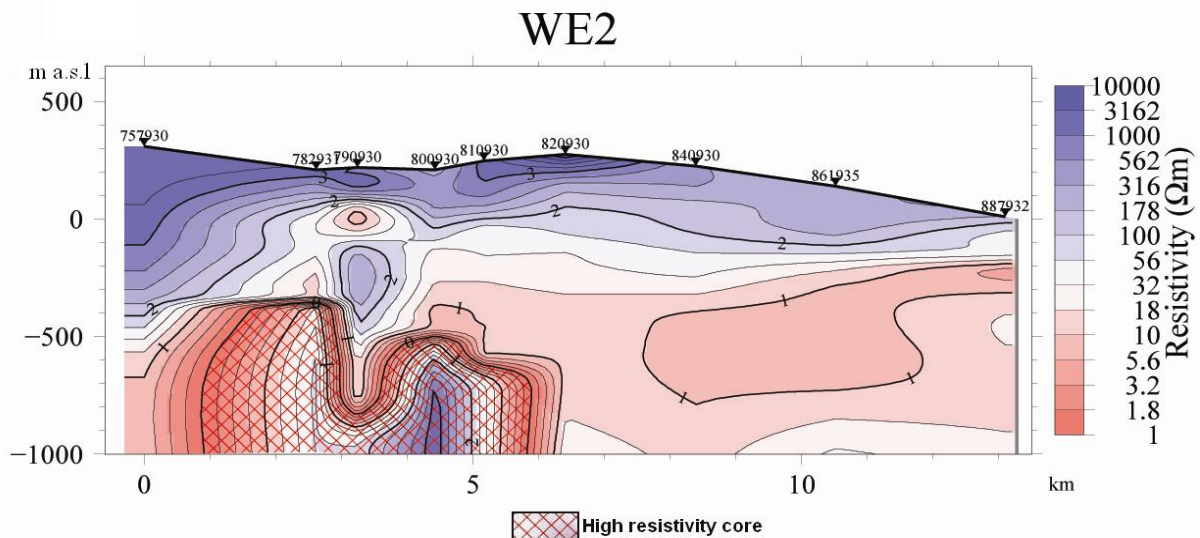


FIGURE 5: A W-E trending TEM resistivity cross-section from the southwest part of the Hengill high-temperature geothermal area, showing a resistivity anomaly that may indicate a new geothermal field where no geothermal activity is known on the surface (Tsend-Ayush, 2006)

### 3.5 MT-measurements

MT or natural-source electromagnetic use the earth's natural electromagnetic field as its power source. The variable natural magnetic field induces electrical currents in the conductive earth. By measuring the signal of the fluctuating magnetic field and the electrical currents (i.e. the electrical field) on the surface of the earth, it is possible to correlate this to the resistivity of the earth below the measuring site. The frequency of the signal relates to its probing depth, with low frequencies reaching deeper levels. Thus, frequencies of 0.00001 - 10 Hz are used for deep crustal investigations, while higher frequencies, like 10 - 1000 Hz, for the upper crust.

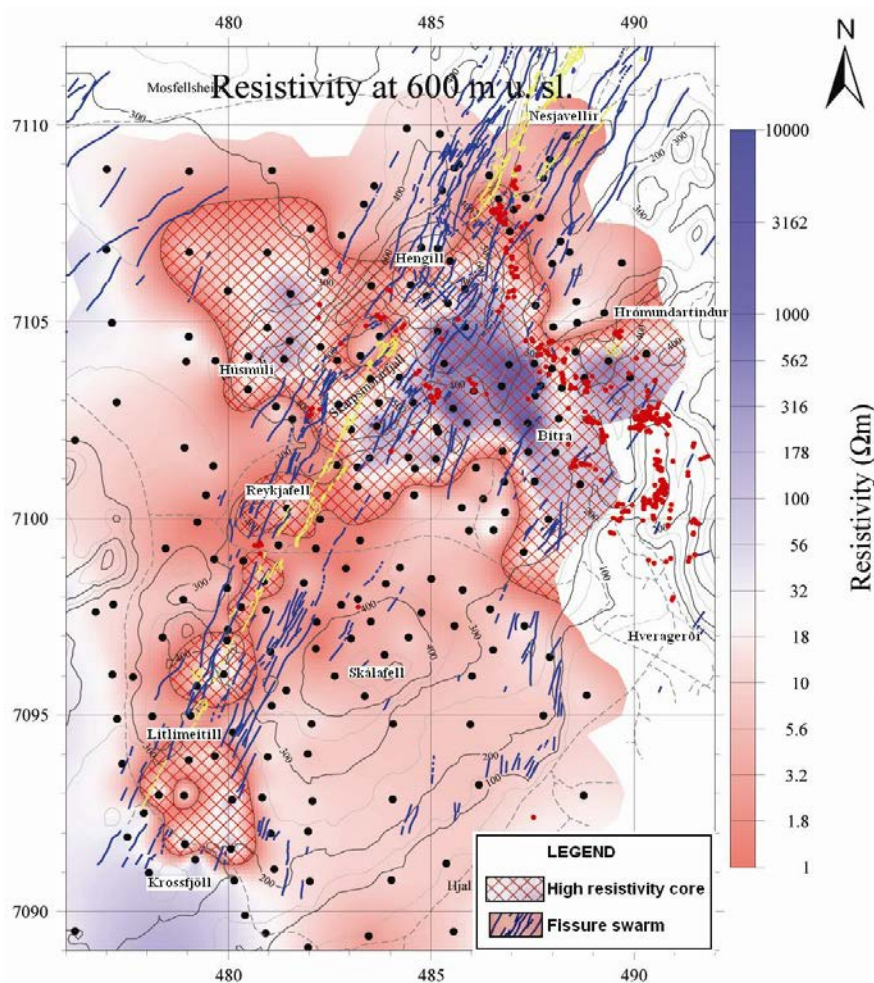


FIGURE 6: The Hengill high-temperature area, SW-Iceland, resistivity map based on TEM soundings showing the resistivity at 600 m below sea level (Tsend-Ayush, 2006)

The MT equipment is fairly simple and portable with the electromagnetic field monitored through magnetic coils and electric dipoles, through a data acquisition system, usually connected to remote-reference station for electro-magnetic noise.

MT is a powerful method to probe deep resistivity structures, which gives it an advantage compared to the other main electrical methods. The equipment is portable and the data collection is simple, involving measurement magnetic field components  $\mathbf{B}$  and the induced electrical field  $\mathbf{E}$ , both as a function of time for several hours, at each site. MT measurements are, however, quite sensitive to cultural noise (power lines etc.). Similarly, the measurements probe a large volume of rocks and are therefore sensitive to 3-D resistivity variations. Detailed interpretation can therefore be difficult and may require 3-D interpretation. More recently, the method has routinely been used in combination with TEM, with the TEM measurements used for mapping of the uppermost kilometre in detail in order to enhance the interpretation of the MT measurements and thus leading to better information at deeper levels. This way, good information on the distribution of the resistivity into the deeper parts of the geothermal system can be collected, reaching to 5-10 km depth.

Figure 7 shows a resistivity cross-section from the Menengai high-temperature geothermal system in Kenya down to about 8 km depth. It shows low resistivity in the uppermost kilometre inside the Menengai crater, but higher resistivity at deeper levels. Figure 8 shows a resistivity cross-section from the Krafla high-temperature system in N-Iceland. It reflects well the low resistivity close to the surface and the higher resistivity below associated with the exploitable geothermal reservoir at 1-4 km depth. Below that an extensive low-resistivity body may reflect partially molten magma.



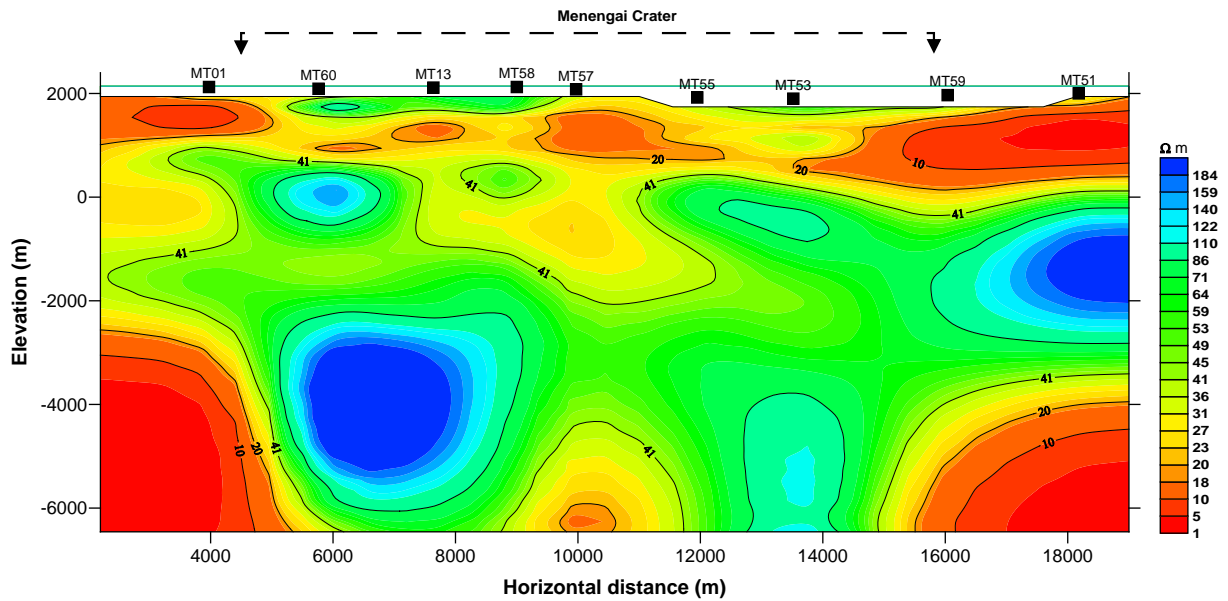


FIGURE 7: MT cross-section from the Menengai high-temperature area and crater in the Kenyan rift (Wameyo, 2005)

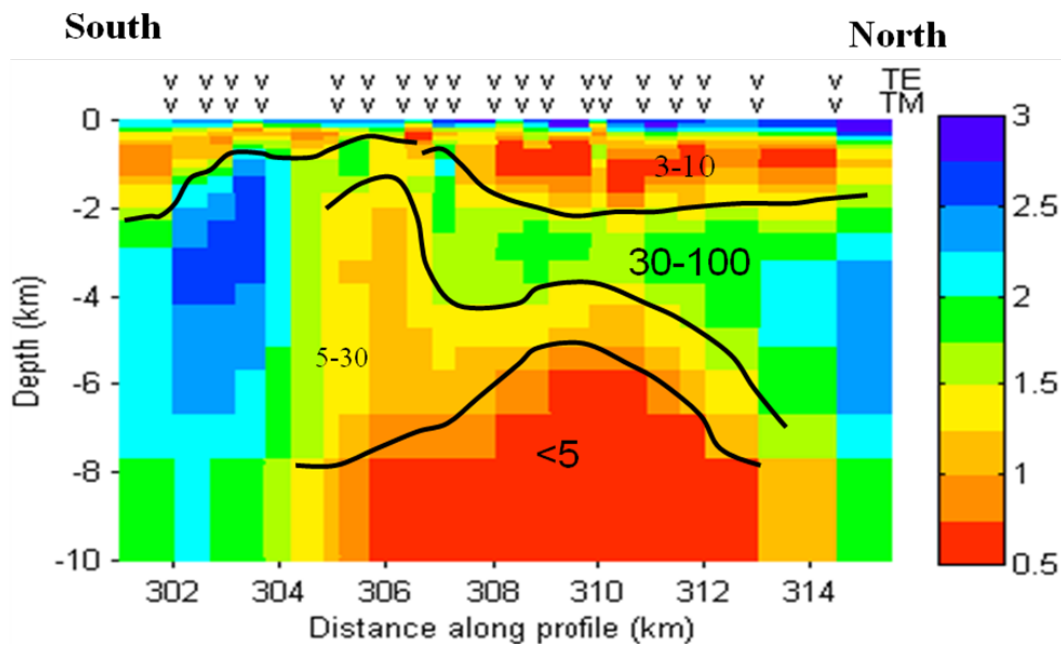


FIGURE 8: MT cross-section from the Krafla high-temperature field in N-Iceland (Árnason, 2007)

### 3.6 SP measurements

Self potential or SP is a special type of electrical measurements aimed at mapping the DC-component of the earth's natural electrical potential. Significant anomalies can be associated with geothermal activity. The anomalies may relate to temperature difference creating thermo-electrical potential, or fluid flow of hot or cold water, but may also be influenced by conductive mineral deposits or chemical variations in rock or pore fluid. The method is not widely used as the interpretation of the data is difficult to quantify, but it still can give important information. The most important applications are for reconnaissance studies, mapping boundaries and tracing faults. In Japan the method has also been used for reservoir monitoring especially for liquid-dominated geothermal systems, assuming that the main cause of the SP anomaly is streaming potential (Yasukawa et al., 2005).

#### 4. MAGNETIC MEASUREMENTS

Magnetic methods are widely used in geothermal exploration, often together with gravity measurements and seismic refraction, in mapping geological structures. Magnetisation of rocks is of two types:

- Induced magnetisation  $M_i$ , which has the same direction as the ambient earth's field; and
- Permanent magnetisation  $M_p$ , which often predominates in igneous rocks and depends upon their properties and history.

A magnetic anomaly is a local or regional disturbance caused by a change in the magnetisation. It is characterized by the direction and magnitude of the effective magnetisation and the shape, position, properties and history of the anomalous body. The magnetic field strength is usually presented in  $\gamma$  (gamma) or nT (nanotesla). Magnetometers are fairly simple equipment and easy to use.

In geothermal exploration, magnetic measurements generally aim mainly at locating hidden intrusives and possibly estimating their depth, or at tracing individual buried dykes and faults. They may also aim at finding areas of reduced magnetization due to thermal activity. The measurements for local structures such as individual faults or dykes are done on the ground by regular measurements along parallel profiles or in a grid (Figure 9). Measurements aimed at larger anomalies such as mapping of deeper intrusions or outlining sedimentary basins are done through aeromagnetic surveys, where the height and spacing of the profiles relies on the preferred data density (Figure 10). In geothermal

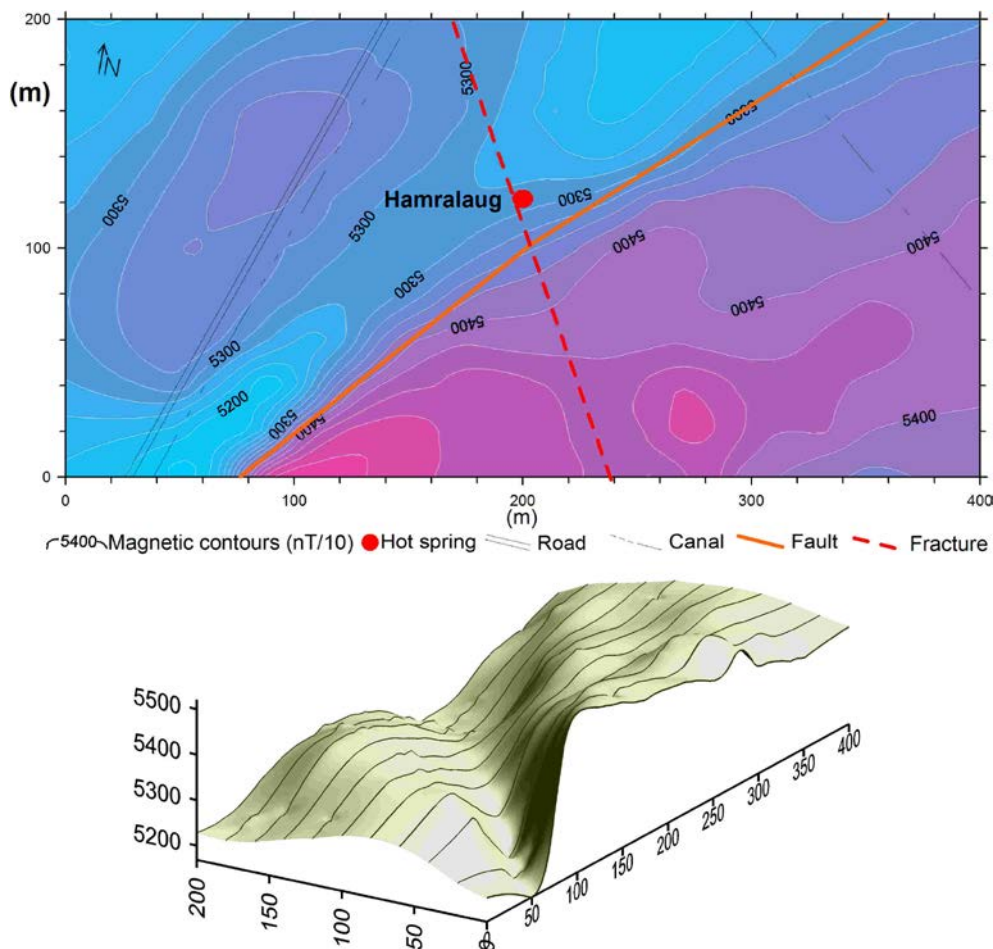


FIGURE 9: Magnetic measurements at the Ásgardur geothermal field in W-Iceland, presented both in a map and through 3D surfaces. The main anomaly is typical for a fault, resulting from different magnetization of the rocks on each side of the fault (Ganbat, 2004)

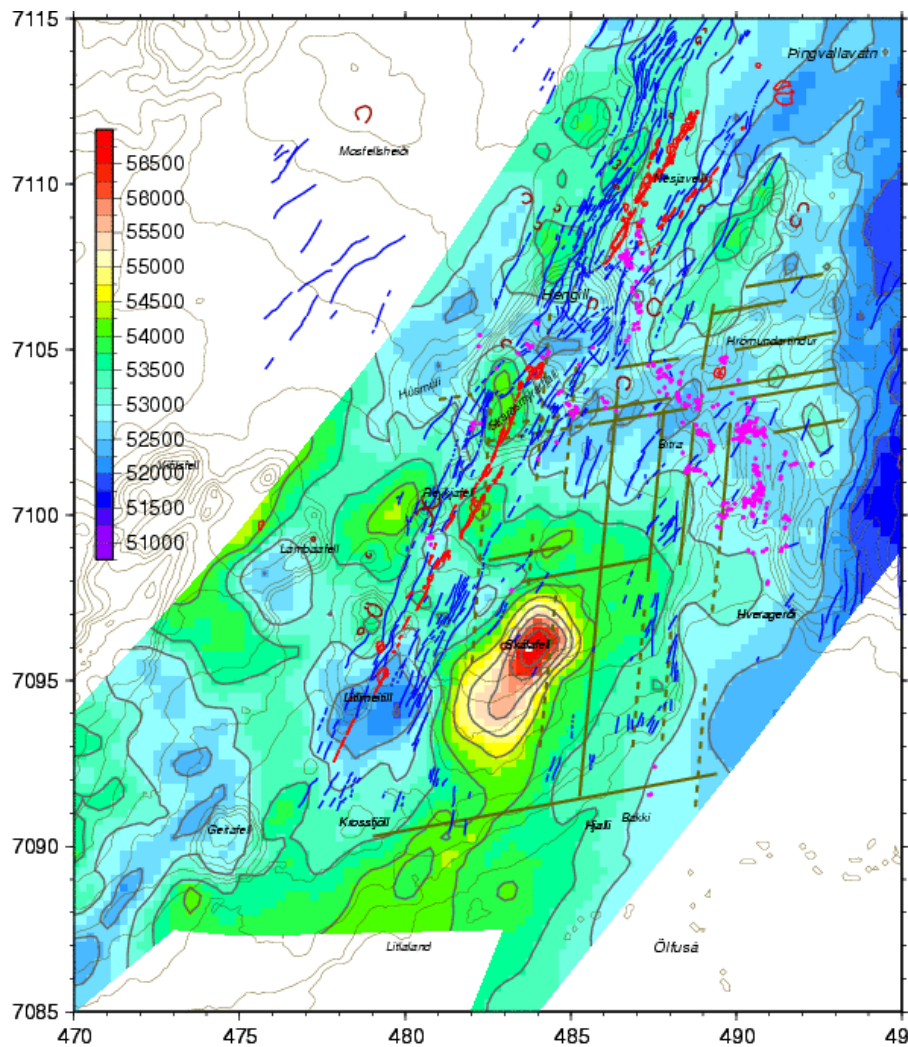


FIGURE 10: Magnetic map of the Hengill high-temperature area, showing also main tectonic features, Postglacial volcanic eruption sites, and the distribution of hot springs and fumaroles; the main geothermal activity correlates with low magnetic intensity (Árnason, 2007)

exploration an elevation of about 100 m above ground and 100 m spacing between profile lines is often used, but this is also influenced by the topography below. Data presentation is through contour maps or profile maps (Figure 9), and quite advanced filtering techniques can be used to distinguish anomalies in the data.

## 5. GRAVITY MEASUREMENTS

Gravity measurements are used to detect geological formations with different densities. The density contrast leads to a different gravitational force which is measured, and usually presented in mgal or  $10^{-3} \text{ cm/s}^2$ . The density of the rocks depends mainly on the rock composition and its porosity, but partial saturation of the rocks may also influence the values. Normally the density is between  $\sim 2$  and  $3 \text{ g/cm}^3$ . Generally, sedimentary rocks are lighter than crystalline rocks. The raw data needs to be corrected for several factors. The results are usually presented or shown in Bouguer maps based on the Bouguer anomaly  $\Delta g_B$ , where corrections have been done on the measured value  $g_M$  for tidal effects, elevation, local topography, latitude and drift in the gravimeter.

Methods of data interpretation are quite similar to those for magnetic measurements. However, the gravity method is somewhat limited due to the ambiguity that theoretically an infinite number of

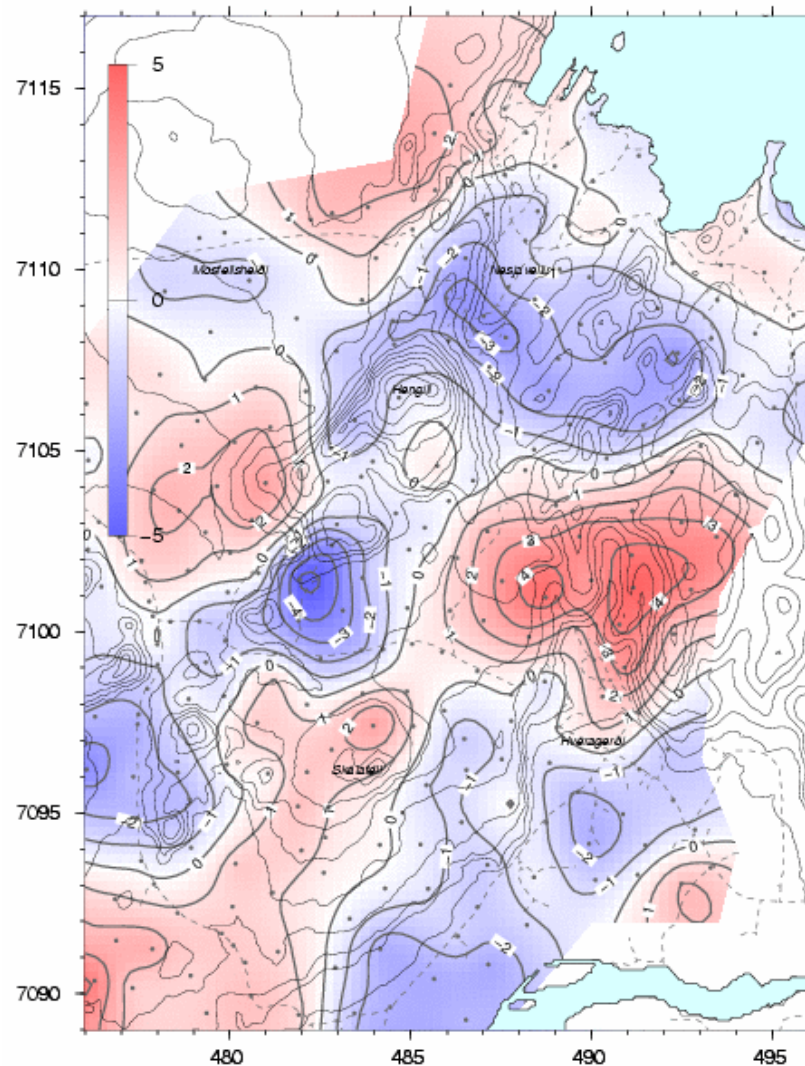


FIGURE 11: Bouguer gravity map of the Hengill high-temperature area; areas of high gravity may indicate intrusions at deeper levels (Árnason, 2007)

density distributions fits a given gravity field. Therefore, other information is needed to be able to fix values, e.g. thicknesses of formations etc.

Applications in geothermal exploration include mapping of: basement depth variations in sedimentary areas, intrusive rocks sometimes associated with a possible heat source, fault or dyke systems, and alteration/cementation due to thermal effects. Figure 11 shows a Bouguer gravity anomaly map of the Hengill area. An important additional application has been the use of precision gravity measurements for monitoring mass extraction in geothermal systems with production (Figure 12).

## 7. SEISMIC METHODS

Seismic methods measure sound velocity distribution and anomalies in the earth as well as attenuation of the sound waves. They are divided into two groups, active methods where an external source is used to create sound waves, such as explosions or hammer devices, while passive methods detect the seismic activity in the earth and use that to get information on parameters that may be influential for the geothermal system.



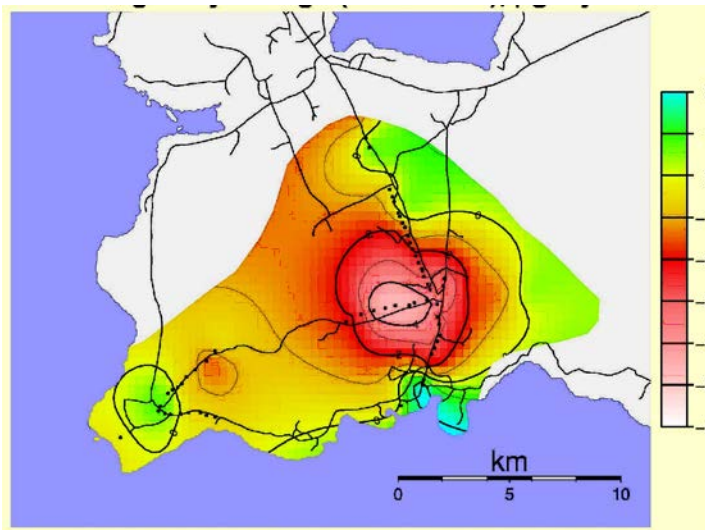


FIGURE 12: Map of the outer Reykjanes Peninsula, SW-Iceland based on precision gravity measurements showing the mean gravity change (reduction) in  $\mu\text{gal}/\text{year}$  from 1975 to 1999 due to production in the Svartsengi geothermal reservoir (Eysteinnsson, 2000)

Seismic methods rely on elastic waves which have different velocities when travelling through different rock types, and are refracted or reflected at discontinuities in or between formations. There are two types of elastic body waves:

- P-waves*, or pressure waves, where the wave movement is in the travel direction; and
- S-waves*, where the material movement is perpendicular to the wave direction; due to the wave movement, S-waves do not penetrate through fluids, but only solid rocks.

In active measurements a distinction is made between refraction measurements and reflection measurements, depending on the procedures. Active measurements give information on the density of the formations, the porosity and texture, boundaries and discontinuities and fluid-filled zones and thus even temperature. The measurements are quite expensive, especially good reflection measurements. And despite being the most used of all geophysical methods as a key method in oil exploration, active seismic measurements, and especially reflection measurements, are not used routinely in geothermal exploration. High-temperature geothermal activity is mainly found within or at least associated with active central volcanoes, which are usually dominated by complicated bodies of crystalline rocks. The information can therefore be quite difficult to interpret and depth penetration not high, and the method thus not cost-effective. The strength of the active methods is in sedimentary surroundings (Figure 13) as is to be expected for the basic methods used for oil carrying formations.

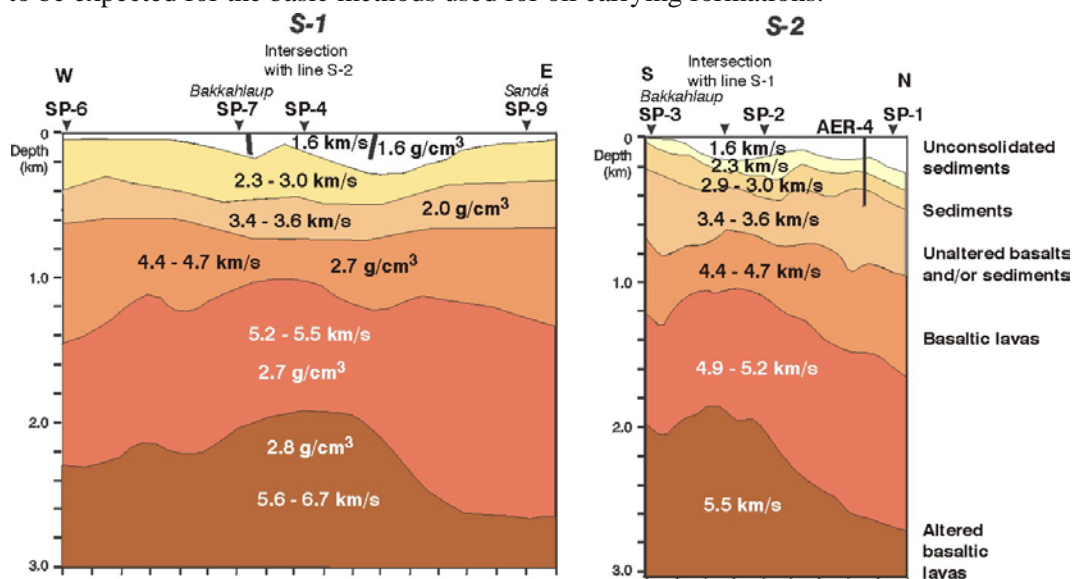


FIGURE 13: Cross-sections through the Oxarfjörður area in NE-Iceland, where geothermal activity is found in sedimentary surroundings, showing formation layers based on refraction measurements and gravity measurements. The up-doming of the high-velocity layers in the central part is the result of the geothermal activity (Georgsson and Fridleifsson, 2000)

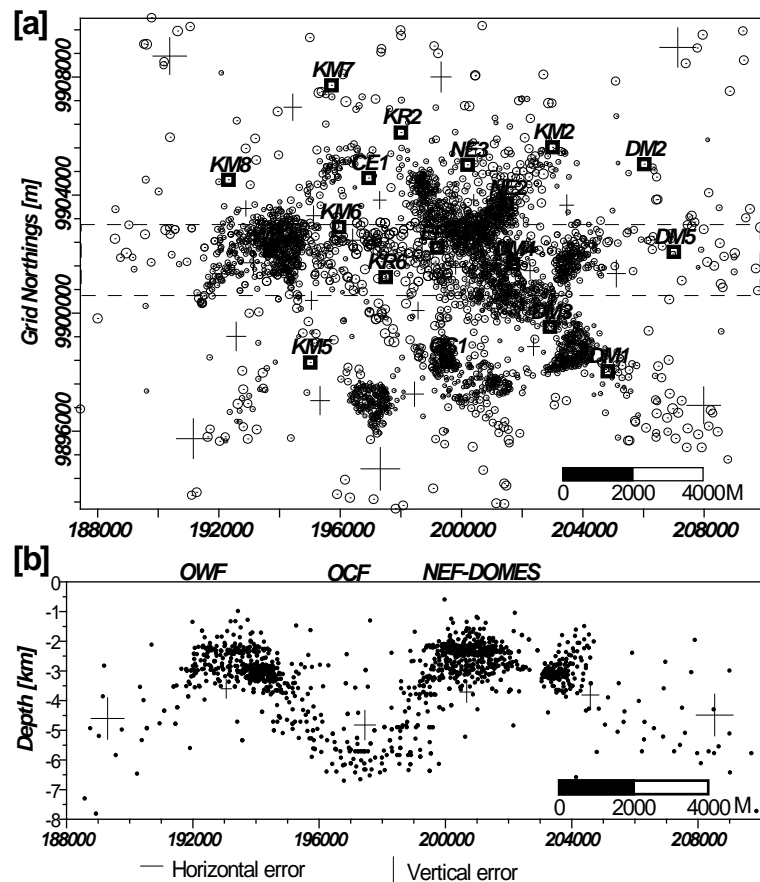


FIGURE 14: Distribution of earthquakes from the Olkaria field in Kenya. The cross-section below shows that the main production fields are associated with relatively high elevation of the brittle/ductile boundary, indicating high temperatures and shallow depth to heat source (Simiyu, 2007)

Passive methods use the natural seismic activity to delineate active faults and permeable zones (shear wave splitting) or to locate the boundary between brittle and ductile crust, which can be indicative of depth to heat source. This can be important for the understanding of the geothermal activity in the measured area (Figure 14). Finally, S-wave shadows can be used to predict locations of bodies of partial melt, or magma chambers.

## 8. COMBINING METHODS

Good geophysical surveys are usually based on a combination of methods. It is necessary in order to get a better understanding of the geothermal system. Usually, routine methods are to be preferred, but if they do not suffice, then other methods or possibly improvised methods need to be taken into account. Figure 15 shows a simple model of the Asgardur geothermal system, based on the temperature measurements shown in

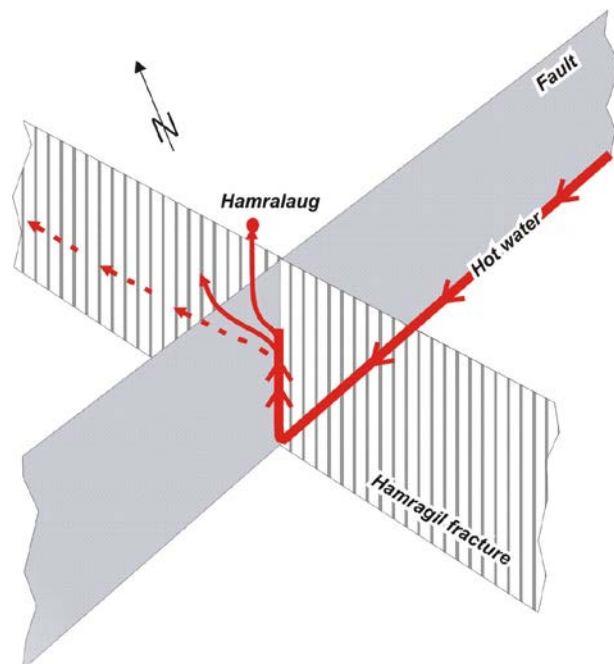


FIGURE 15: A simple model of the Asgardur low-temperature system in W-Iceland, based on the results of soil temperature measurements and magnetic measurements (Ganbat, 2005)

Figure 1 and the magnetic measurements in Figure 9. The NE-SW trending fault mapped by the magnetic measurements is believed to carry the water towards the geothermal field from the highlands, but the geothermal upflow (convection?) is at its intersection with a northwesterly fracture delineated through the soil temperature measurements.

Figure 16 shows maps of the large Hengill high-temperature area in SW-Iceland, where resistivity data has been combined with gravity data, information on active faults and the distribution of the geothermal activity. The maps have provided important information for understanding of the geothermal system and revealed possible new production fields.

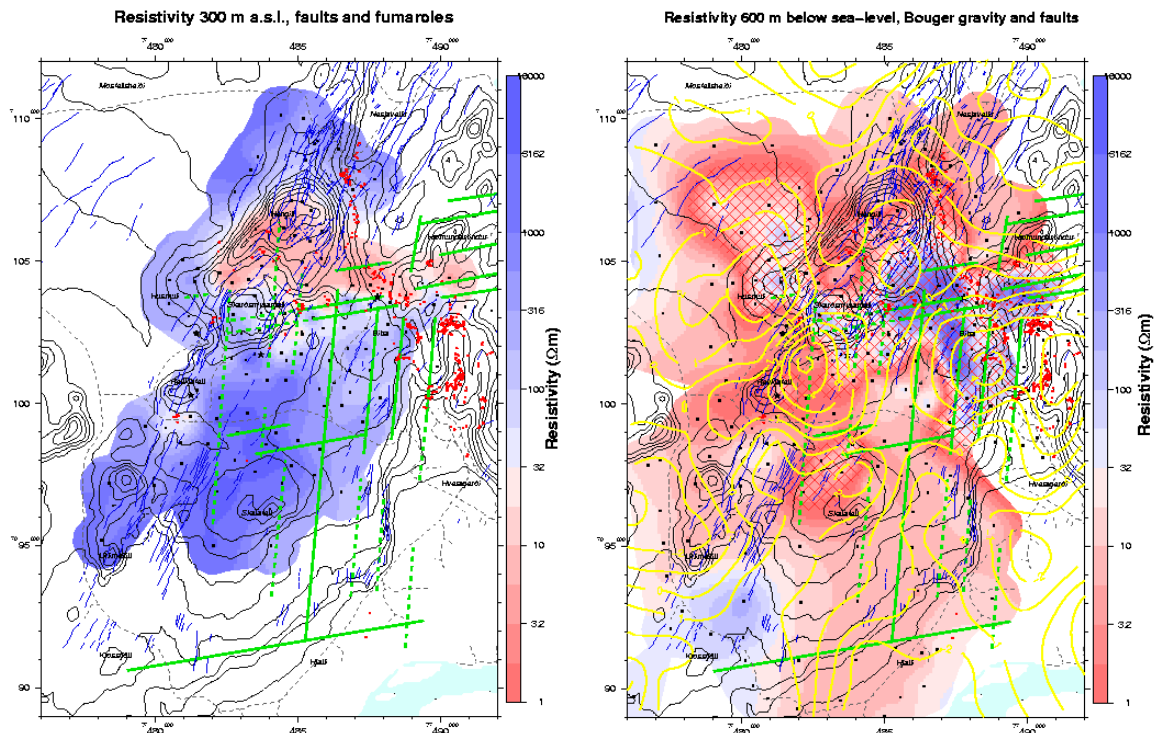


FIGURE 16: Resistivity maps at 300 m a.s.l. and at 600 m b.s.l. combined with the distribution of the geothermal activity, a Bouguer gravity map and information on active faults from seismic measurements. The geophysical data have given important information on known production fields but also indicated new ones where no geothermal activity is seen on surface (Árnason, 2007)

## 9. FINAL REMARKS

Economic exploitation of a geothermal system is usually limited to the uppermost 1-3 km of the crust, and the actual target depth or the selection of a drill rig may be important for the selection of methods. It is sound to apply methods with which you are well acquainted if it is believed that they suit the purpose of the survey. However, it is also important to have the knowledge and capacity to be able to improvise or adapt new methods if the conventional methods are not providing the information that is needed. Finally, the geophysicist must always remember that he does not stand alone. His results need to be integrated with information on the geology of the geothermal area and the chemistry of the geothermal fluids, for good understanding of the geothermal system.

A successful survey hopefully leads to the development of the resource, and if so, its success can be measured by the time, effort and money it has saved, for example by production of electricity or hot water for heating from a renewable energy source instead of oil or coal. Or if the measurements have managed to spare the drilling of one failed production well, they surely have saved a lot of money.

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