



BOREHOLE GEOLOGY AND ALTERATION MINERALOGY OF WELL HE-11, HELLISHEIDI GEOTHERMAL FIELD, SW-ICELAND

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ABSTRACT

HE-11 is a 1652 m deep well in the Hellisheidi high-temperature field in the Hengill geothermal area in SW-Iceland. It was drilled in June 2004 as a production well for the utilization of electric power generation. The well is directional, aimed to intersect a 5000 years old northeast trending Postglacial volcanic fissure. The lithology consists predominantly of volcanic rocks of basaltic composition that includes lava flows and sub-glacial formations of hyaloclastites. Several basaltic intrusions intrude the hyaloclastites. The hyaloclastite formations make up about 90% of the whole stratigraphic section and include tuff, breccia and pillow lava. Hydrothermal alteration indicates four alteration zones; a zone of no alteration down to 486 m, the smectite-zeolite zone (486-700 m), mixed-layer clay zone (700-756 m), chlorite zone (756-950 m) and a chlorite-epidote zone (>950 m), indicating temperatures of <200°C, 200-230°C, 230-250°C and >250°C, respectively. The time related mineral depositional sequences show a geothermal system undergoing a progressive temperature increase over time. A correlation between the formation and alteration temperatures indicates that recent heating has occurred from about 500 m down to 1000 m but is in equilibrium from there to about 1200 m. Many of the aquifers in the production part of the well appear to be related to dyke intrusions, indicating the dominance of fracture permeability in the geothermal reservoir.

1. INTRODUCTION

1.1 Purpose of study

Drilling is essential in both exploration and exploitation of geothermal fields. Borehole geologists use cuttings and cores to determine subsurface lithology, hydrothermal alteration and the location and geological identity of probable aquifers. With such data, it is also possible to evaluate how the thermal evolution of a geothermal system has developed. In this paper the author describes the borehole geology and alteration mineralogy of HE-11 in the upper 1200 m. This report is mostly based on the study of available data from the time of drilling, and includes the study of circulation

fluid and drill cuttings. Other relevant data were used such as temperature, pressure and geophysical logs.

1.2 General information

Well HE-11 is situated within the Hellisheidi high-temperature area, which is a part of the Hengill central volcano. It was drilled down to 1652 m in June 2004. The Hengill central volcano covers an area of about 40 km² (Björnsson et al., 1986). Hellisheidi geothermal field is located in the southern part of Mt Hengill, and some 20 km south of the Nesjavellir high-temperature field (Figure 1). The extent of the present well field covers some 12 km². The first exploration well was drilled in 1985 at Kolvidarhöll (KhG-1) at the western boundary of the Hellisheidi field. A total of 17 wells has been drilled in Hellisheidi till September 2005, nine of them deviated. The depths range from near 1000 m to a maximum of 2808 m (HE-8), which is the second deepest high-temperature well drilled to date in Iceland. Plans are to complete 5 exploitation wells within the next year to sustain a new 80 MWe power plant for 30 year operation, to be commissioned in late 2006. At the present time the Hellisheidi geothermal power plant is being built by Reykjavík Energy in the southwest part of the Hengill area.

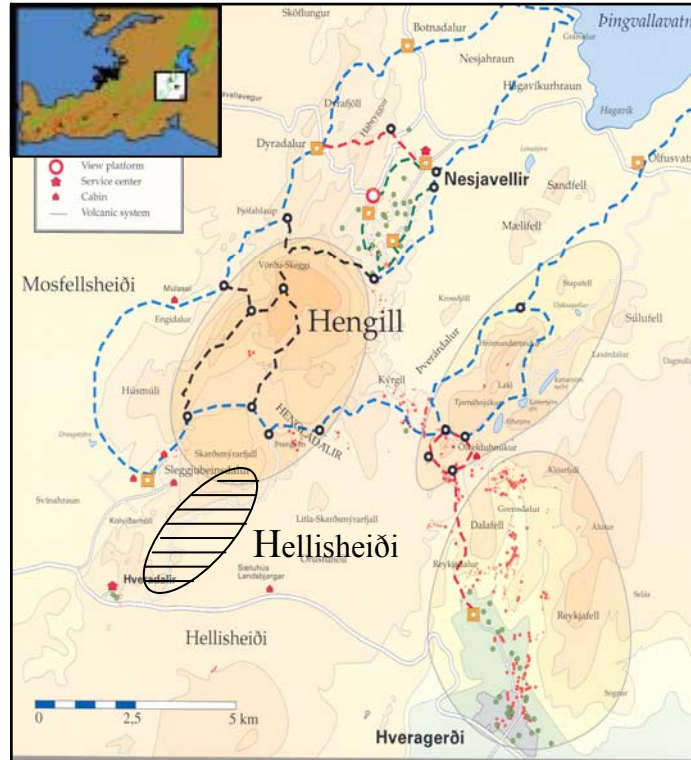


FIGURE 1: Location map of the study area (modified from Gíslason and Gunnlaugsson, 2003)

1.3 Description of work

The cutting samples are essential to enable the development of geological and hydrothermal alteration models of a geothermal reservoir. The study of geological structures and hydrothermal alteration by cutting analysis involves three main pieces of equipment, namely the stereo-binocular microscope, the petrographic microscope and the X-ray diffractometer (XRD).

Binocular microscope is used mainly during drilling to make the geological section and hydrothermal alteration, which is important in helping to make decisions during drilling, such as when setting casing shoes, in prediction of loss zones, or deciding on final drilling depth. Wellsite geology, thus, plays an important part during drilling. Well HE-11, the object of this study, was drilled to 1652 m. The binocular microscope used in the laboratory is an *Olympus SZX 12 Japan*, with a magnification of 6× - 80×.

Petrographic microscope is a useful instrument in the study of drill cuttings in thin section as it allows a more detailed analysis of the rock and in particular the alteration minerals, the study of the replacement mineralogy and amygdale and veins minerals. Petrographical analyses are also good for studying the sequence of secondary mineral evolution, especially in the veins and amygdales. The type of polarized microscope used in the petrography laboratory is an *Olympus SZX, Ortholux II POL-BK, Leitz Wetzlar* with a binocular magnification of 4/0.12P to 25/0.5P

X-ray diffractometer is mostly used for identifying clay minerals and for cross-checking observations made under both the polarized and binocular microscopes, especially where zeolites and other white minerals could not be identified individually. The procedure of sample preparation for clays and deposition minerals and also running the XRD machine can be seen in Appendix I.

Software Log Plot 2001 from Rock Ware, Inc. was used to manipulate all the borehole geology, logging and drilling data. It was used for drawing alteration mineral zonations, and different borehole geology data, showing lithology, the distribution of alteration minerals, alteration zones, possible intrusions and possible aquifers. Comparing data from geophysical logs with cutting analysis, thin-section analysis and XRD analysis is valuable especially for finding possible aquifers and possible intrusions.

2. GEOTHERMAL RESOURCES

2.1 Regional geology

Iceland is a geologically young country that lies on the Mid-Atlantic Ridge. Regions of active extension and volcanism in Iceland lie along volcanic rift zones that cross the island from northeast to southwest. This is the boundary between the North American and Eurasian tectonic plates. It is one of the few places on earth where you can see an active spreading ridge above sea level with the two plates moving apart on average about 1-2 cm per year in either direction (Saemundsson 1979). Figure 2 shows the general geology of Iceland along with main geothermal activity. The oldest parts are of Tertiary age (3.1-16 m.y.) east and west of the active volcanic rift zones. Here the volcanic succession is dominated by sub-aerial lavas which dip smoothly towards the active volcanic zone. The younger part is composed of alternating sequences of lavas and hyaloclastites that belong to Plio-Pleistocene age (0.01-3.1 m.y.). The main difference between these two units is the volcanic environment. The lavas are erupted in interglacial periods whereas the hyaloclastites are formed during the glacial periods. The youngest formations in the volcanic zones are sub-aerial lavas of Holocene age (younger than 10,000 years).

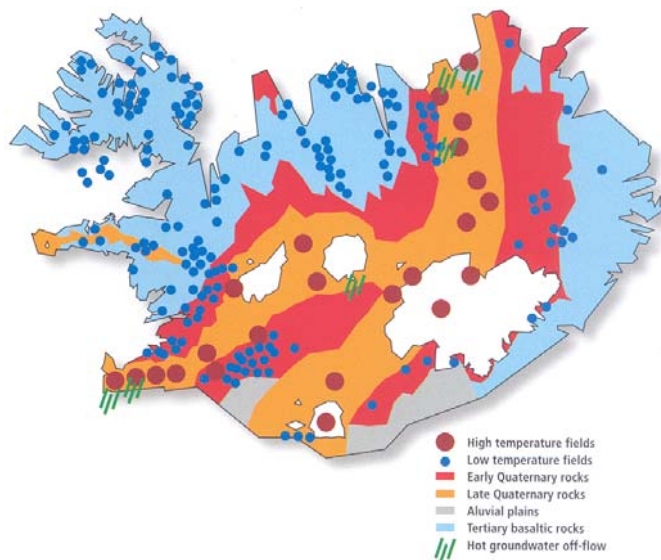


FIGURE 2: A simplified geological map of Iceland, see text for further explanations (Reykjavík Energy, 2005)

Most of the rifting occurs within fissure swarms (Saemundsson, 1979). These fissure swarms in the active volcanic zones are 5-10 km wide and 30-100 km long. Each zone consists of nested grabens, where near-vertical normal faults are exposed at the surface. High-temperature geothermal systems are associated with the young volcanic systems along the active tectonic rifts. They are expressed by thermal manifestations which colour the surface by alteration, mud pots, natural hot springs, fumaroles and geysers. There are well over 100 volcanoes in Iceland, exhibiting nearly every type of activity. At least 20 have erupted in historical times, with more than 150 recorded eruptions. Fissures often become long, thin volcanic vents. Iceland is the classic area for fissure eruptions in which lava often fountains out and forms lava floods of vast extent. Each fissure represents a small amount of crustal extension, perhaps about one metre on average (Saemundsson, 1979).

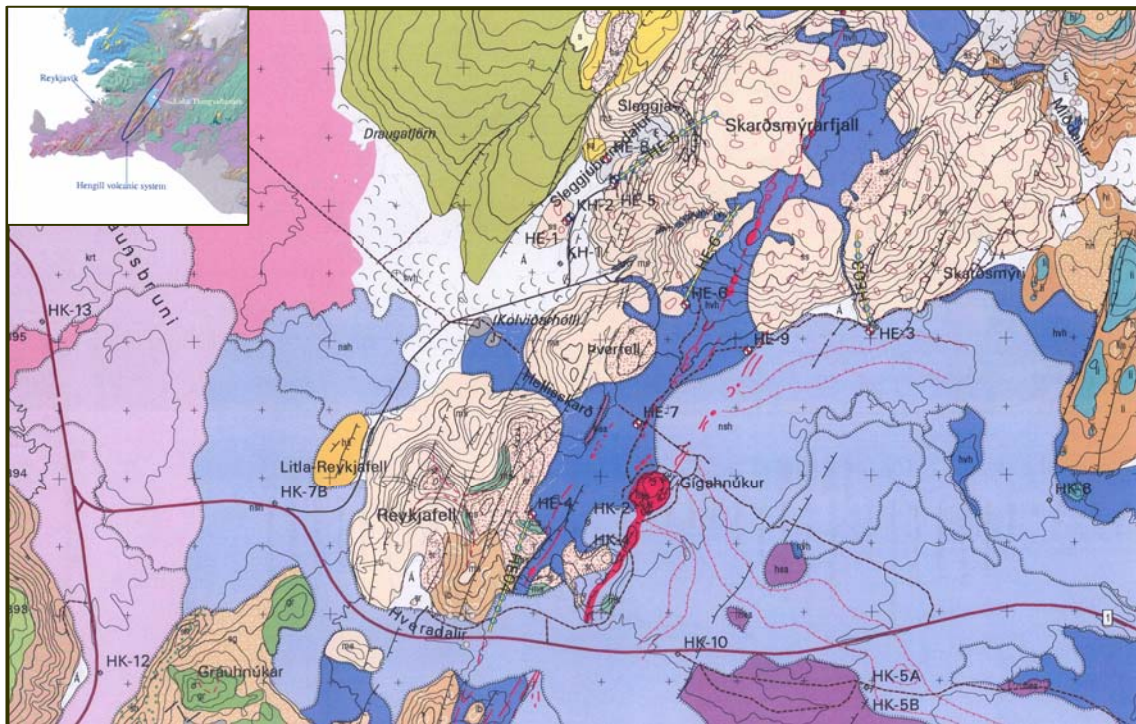


FIGURE 3: Geological map of Hellisheidi area (Saemundsson, 1995), the part shown extends over an area of approx. 9 km × 6 km; wells marked with HE are recently drilled high-temperature wells in the Hellisheidi field

2.2 Geology of Hengill area

The Hengill area, east of Reykjavik, is one of the largest high-temperature areas in Iceland, extending over some 50 km². The geothermal activity is believed to be connected to three volcanic systems. The geothermal area in Reykjadalur, Hveragerdi belongs to the oldest system, called the Grensdalur system. North of this is a volcanic system named after Mt. Hrómundartindur, which last erupted about 10,000 years ago. The geothermal area in Ölkelduháls is connected to this volcanic system. West of these volcanic systems lies the Hengill volcanic system, with intense tectonic and volcanic NE-SW fractures and faults extending from Lake Thingvallavatn to Nesjavellir and further to the southwest through Innstidalur, Kolvidarhóll, Hveradalur (hot spring valley) and Hellisheidi (Saemundsson 1979). A simplified geological map of the southern part of Hengill volcano, in which the Hellisheidi project lies, is shown in Figure 3.

A tensional stress system has developed in the zone of the active Hengill volcanic system. The tensional stress has opened northeast trending vertical fractures, faulting, graben and geothermal systems that occur along the fissure swarm, which provide highly permeable conditions for fluid flow (Björnsson et al., 1986), as shown in Figure 4. The fracture network is periodically activated, providing conduits for the episodic eruption of basalt and the emplacement of dikes. Magma moves into the shallow crust at temperatures on the order of 1200°C and supplies heat to the hydrothermal system (Bödvarsson et al., 1990). The Hengill area on the surface is almost entirely built up with volcanic rocks of late Quaternary and Holocene age (Árnason et al., 1967). The rocks are mostly sub-aerial basaltic flows and hyaloclastites but small amounts of intermediate and rhyolite compositions occur as well.

Hellisheidi geothermal system is located in the southeastern part of the Hengill mountain. The Hengill mountain itself was mostly accumulated in one or two large subglacial eruptions during the last glacial period. New geological data presented in 2002 suggest that the lower part of the mountain may have

formed during the 2nd last glacial period (Fridleifsson et al., 2003). Hyaloclastite deposition, a typical formation in Iceland, is a fine, glassy debris formed by the sudden contact of hot and coherent magma with either cold water or water-saturated sediments, usually associated with glaciers. Rapid heat loss from the magma to the ice sets up tensile thermal stress in the magma carapace as it cools, chills, and contracts, causing the glassy, chilled magma to fragment and form quenched-fragmented debris.

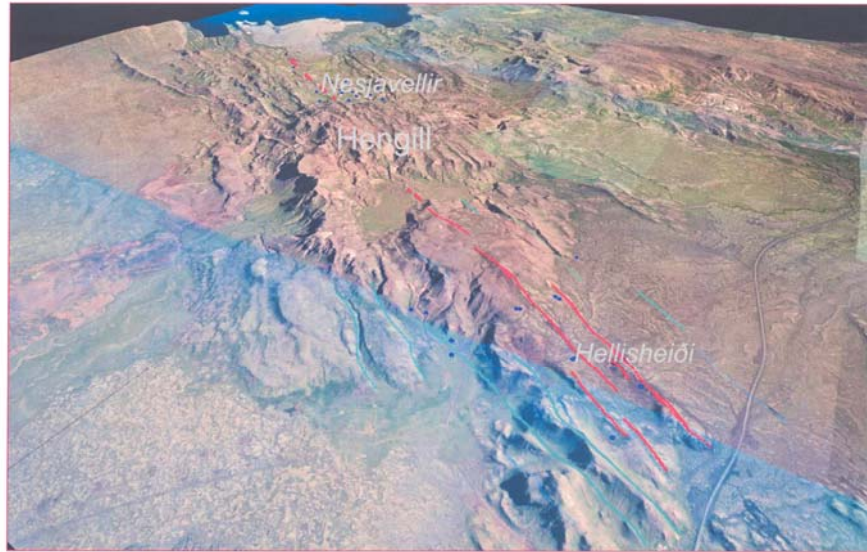


FIGURE 4: Structural activity in Hengill area in a 3-D view showing faulting, graben and the elongated fissure swarms

If the deposit remains in contact with water after its formation, the glassy debris may further hydrate to form palagonite. The Hengill triple junction sits between the Reykjanes Peninsula rift zone, the western volcanic rift zone, and the South Iceland Seismic (transform) zone, in SW-Iceland (Foulger, 1988). Geology of the Hellisheiði geothermal field is characterized by the presence of two kinds of common hyaloclastite formations and basaltic fissure lavas produced in an active fissure swarm.

Hyaloclastite formations: Subglacial fissure eruptions produce elongated ridges that are 1-5 km wide, up to tens of kilometres long, and a few hundred metres thick. The cores of these ridges consist of permeable pillow lavas, but the flanking hyaloclastite deposits can serve as aquitards. Subglacial eruptions are remarkable in that they are able to create both the reservoir and caprock in one volcano (Wohlitz and Heiken, 1992). Commonly, hyaloclastite formations in Iceland are created by the effect of volcanism under glaciers. The sequence of formation is shown in Figure 5. The most recent volcanism of this type was in Bárðarbunga 1996 in the glacier Vatnajökull, but the most famous episode was formed at the seafloor south of Iceland in 1963-1967, and formed the island Surtsey

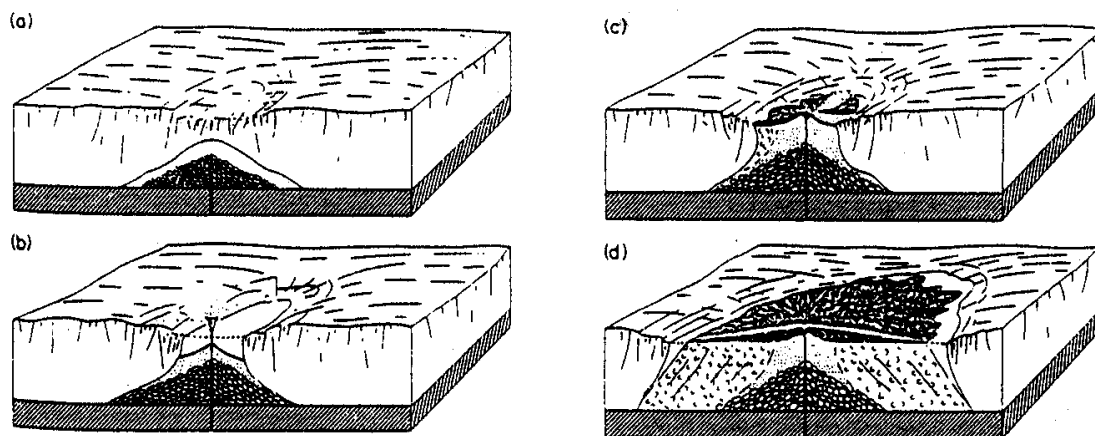


FIGURE 5: Growth of a subglacial, monogenetic volcano; a) a pile of pillow lava forms deep in melt water lake; b) slumping on the flanks of the pillow lava pile produces pillow lava breccia; c) hyaloclastite tuffs are erupted under the shallow water, d) a lava cap propagates across its delta of foreset breccia (Jones, 1969; Saemundsson, 1979)

which belongs to the Vestmannaeyjar archipelago. The morphological structure is named móberg (in Icelandic) or hyaloclastite, and the formation, table mountain. Hyaloclastite formations characterize the environment in the Hengill region and form mountains like Mt. Skardsmýrarfjall, Mt. Reykjafell and Mt. Hengill.

The fissure swarm: There are two main volcanic fissures trending NE-SW in the Hengill area of Holocene age that have fed the last volcanic eruptions in the area, extending from Lake Thingvallavatn in the northeast part of the Hengill area (Nesjavellir high-temperature field) to about 20 km southwest of the Hengill mountain (Hellisheidi). The age of the older one is about 5500 years and the younger one is about 2000 years old (Saemundsson, 1967). The lava flows are widespread and cover a large part of Hellisheidi. These eruptive fissures and parallel faults control up- and out-flow of hot water and steam from the centre of the Hengill system. Tectonic activity is episodic and accompanied by rifting and major faulting along the fissure swarm that intersects the Hengill central volcano and magma is injected into the fissure swarm. A row of small craters is marked along both the older and younger eruptive fissures.

2.3 Geophysical surveys of the Hengill area

Extensive resistivity surveys have been done in the Hellisheidi geothermal field including Schlumberger and TEM resistivity measurements (Björnsson et al., 1986; Árnason and Magnússon, 2001). Aeromagnetic and gravity surveys have furthermore been done in the Hengill geothermal area, including Hellisheidi (Björnsson et al., 1986). The DC-resistivity surveys delineated a 110 km² low-resistivity area at 200 m b.s.l., and the magnetic survey showed a negative and transverse magnetic anomaly coherent with the most thermally active grounds. Transient electro-magnetic soundings (TEM) revised the resistivity map. The results of this study suggested that the resistivity anomaly is complex and affected by processes such as faulting, shearing and spreading (Figure 6). The electrical resistivity also provides information on the degree of alteration and therefore hydrothermal activity, including both fossil and

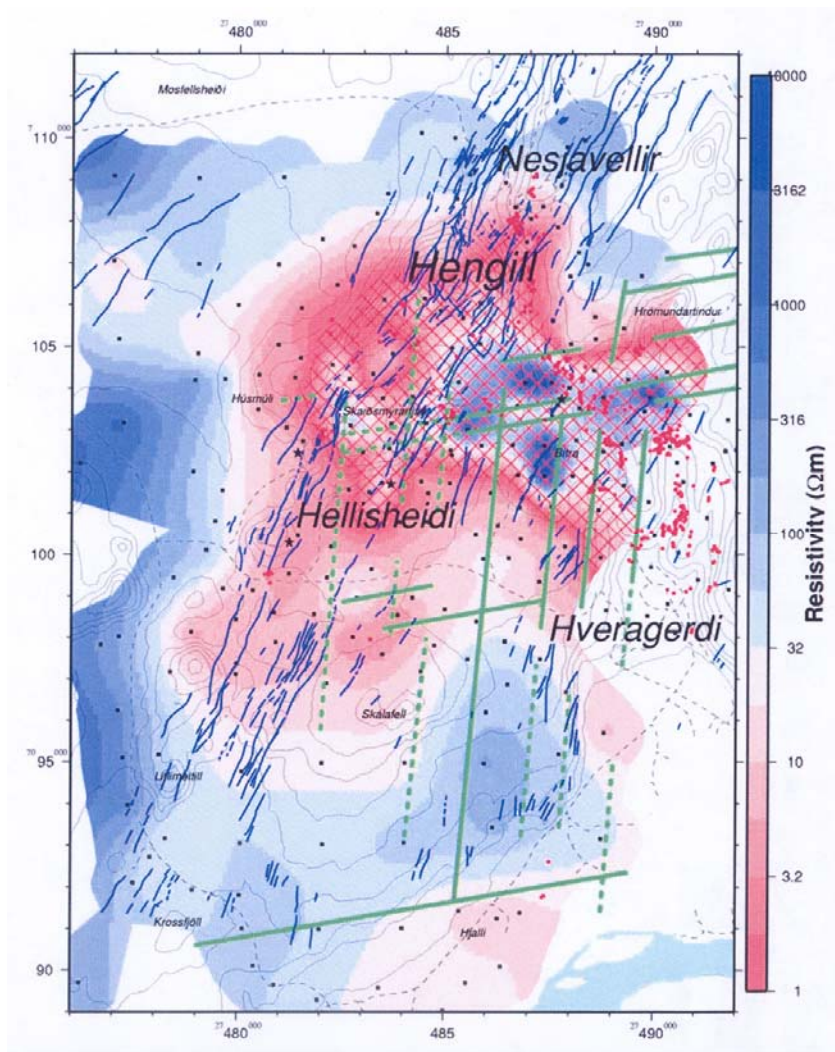


FIGURE 6: Hengill area, resistivity at 100 m b.s.l. according to a recent TEM survey, also showing faults (Árnason and Magnússon, 2001)

recent alteration. A low-resistivity area covering 110 km² measured at a depth of 400 m, indicates roughly the extent of the high-temperature fields. All surface manifestations like fumaroles and altered grounds are within this area (Hersir et al., 1990). The main feature is a high-resistivity zone (50-500 Ωm) which is observed beneath the low-resistivity layer. A correlation between resistivity, rock temperature and alteration at Nesjavellir (northeast of Hellisheidi) shows high-resistivity values close to the surface which can be attributed to fresh unaltered rocks. The low-resistivity values (1-5 Ωm) are connected with the smectite-zeolite belt at temperatures between 50 and 200°C. But below the low resistivity there is a high-resistivity core associated with a high-temperature alteration zone. This high-resistivity core is related to the chlorite-epidote zone that is located under a chlorite zone, indicating temperatures of more than 240°C (Árnason et al., 2000).

3. BOREHOLE GEOLOGY

The purpose of exploration drilling is to gather all available data about the geothermal reservoir. At the drill site, borehole geologists focus on analysing the cuttings and cores and make a lithological section of the drilled section, as well as identifying the alteration minerals and mapping their distribution.

3.1 Drilling

Well HE-11 is located at the south-western margin of Stóra Skarðsmýrarfjall (SK) mountain. The coordinates (Lambert-system) for the well are: X = 384645.948 and Y = 395233.318 and the altitude about 420 m a.s.l. This well is at the same wellpad as well HE-6, to minimize environmental impacts. Well HE-11 is directional, and was aimed at intersecting the 5000 year old volcanic fissure in Hellisheidi geothermal field (c.f. Figure 7). The well was directed to N 55° ± 10° as shown in Figure 7. The kick-off point was at 245 m depth where the deviation from vertical started with a building rate averaging 2°/30 m. During drilling the inclination and direction were measured regularly by using MWD-tools (Measure With Drilling) placed in the collar close above the bit, and confirmed with gyro surveys as needed.

The drilling was carried out by the Icelandic drilling company, Jarðboranir Ltd, started the 21st of June 2004, and finished after completion and injection tests, 40 days later. The drilling was done in four separate stages: a 22 ½" anchor casing was cemented down to 98 m, a 18 ⅝" safety casing down to 269 m, and a 13 ⅜" production casing down to 765 m depth. The production part of the well was drilled using a 12 ¼" bit down to 1652 m depth and cased with a 9 ⅝" slotted liner down to 1564 m depth. The schematic design of the well is shown in Figure 8. Mud circulation was used down to the production casing depth, but freshwater in the production part, to keep it clean and cool and to protect the well's aquifers against sealing.

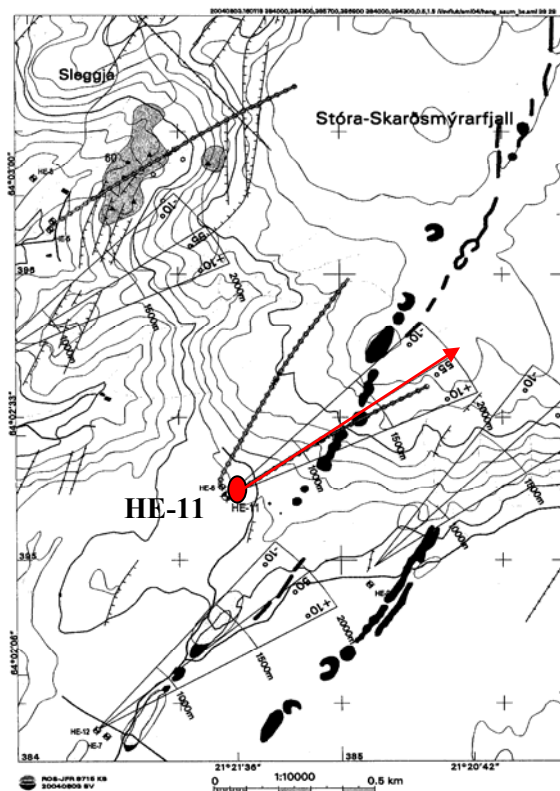


FIGURE 7: Enlarged map of the Hellisheidi well field, location and direction of the directional well HE-11

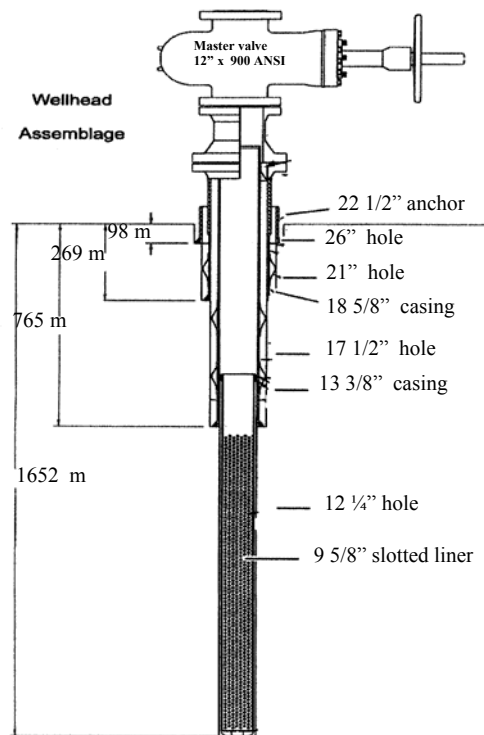


FIGURE 8: Schematic design of well HE-11 (modified from Jónsson et al., 2004)

About 600 samples of drill cuttings were picked at 2 m intervals during the drilling of the well down to 1200 m where total circulation loss occurred. The rock cuttings were collected in 125 ml plastic containers and labelled appropriately. The sampling method of collecting drill cuttings every 2 m in the well was based on the fact that lithology, alteration and fault/fracture intensity in geothermal systems can be very complex and can change abruptly over small distances. Therefore, it is of great value to achieve the best possible representative samples of the intervals. Every sample was analyzed to determine the rock type and to identify secondary minerals. For further study, 34 thin sections were prepared for investigation using a petrographic microscope and 18 samples to be run through X-ray diffraction (XRD).

By studying all of the cutting samples, thin sections and geophysical logs and using binocular and petrographic microscopes and X-ray diffractometry methods, stratigraphy, alteration mineralogy, permeable zones, alteration zonation and aquifers were

determined. In fact by studying all available evidence, it is possible to define the geological characteristics of the well and also find how these properties can affect the hydrothermal fluid flow through the well's aquifers.

3.2 Stratigraphy

All the rock formations in the well are basaltic, according to the drill cutting analysis and the available geophysical logs. A slightly simplified lithological profile is shown in Figure 9, along with several geophysical logs. The basaltic rocks are rather uniform in composition, either constituting coarse-grained olivine tholeiites or fine-grained tholeiites. The lithofacies, however, are more variable springing from sub-aerial lava flows between intercalating sub-glacially extruded hyaloclastites. The hyaloclastite is divided into hyaloclastite tuff (predominantly of volcanic glass), hyaloclastite breccia (made up of similar amounts of glass and partially crystallized basalt), and glassy basalt/pillow lava (dominantly partially crystallized basalt, with a minor amount of glass). These three lithological types of hyaloclastites vary in appearance from one depth to another, to some extent depending upon increased alteration. The hyaloclastite formations occur commonly as heaps, ridges or clastic sediments close to their vents.

Intrusive rocks of basaltic composition are common, and are both of olivine tholeiite and tholeiite composition. Commonly they are coarser-grained than the lavas of same composition, due to slower cooling. Narrow intrusions may still be fine- to medium-grained. They are often assumed to have intruded sub-vertically into the hyaloclastites. The intrusions are usually of lower porosity and lower permeability and of younger age than the surrounding rocks. The location of intrusive rocks are shown in a separate column alongside the lithology profile in Figure 9.

The description of the stratigraphic column is mainly based on the binocular stereoscope observations, aided by petrographic analysis of thin sections. The succession is mainly divided into hyaloclastite

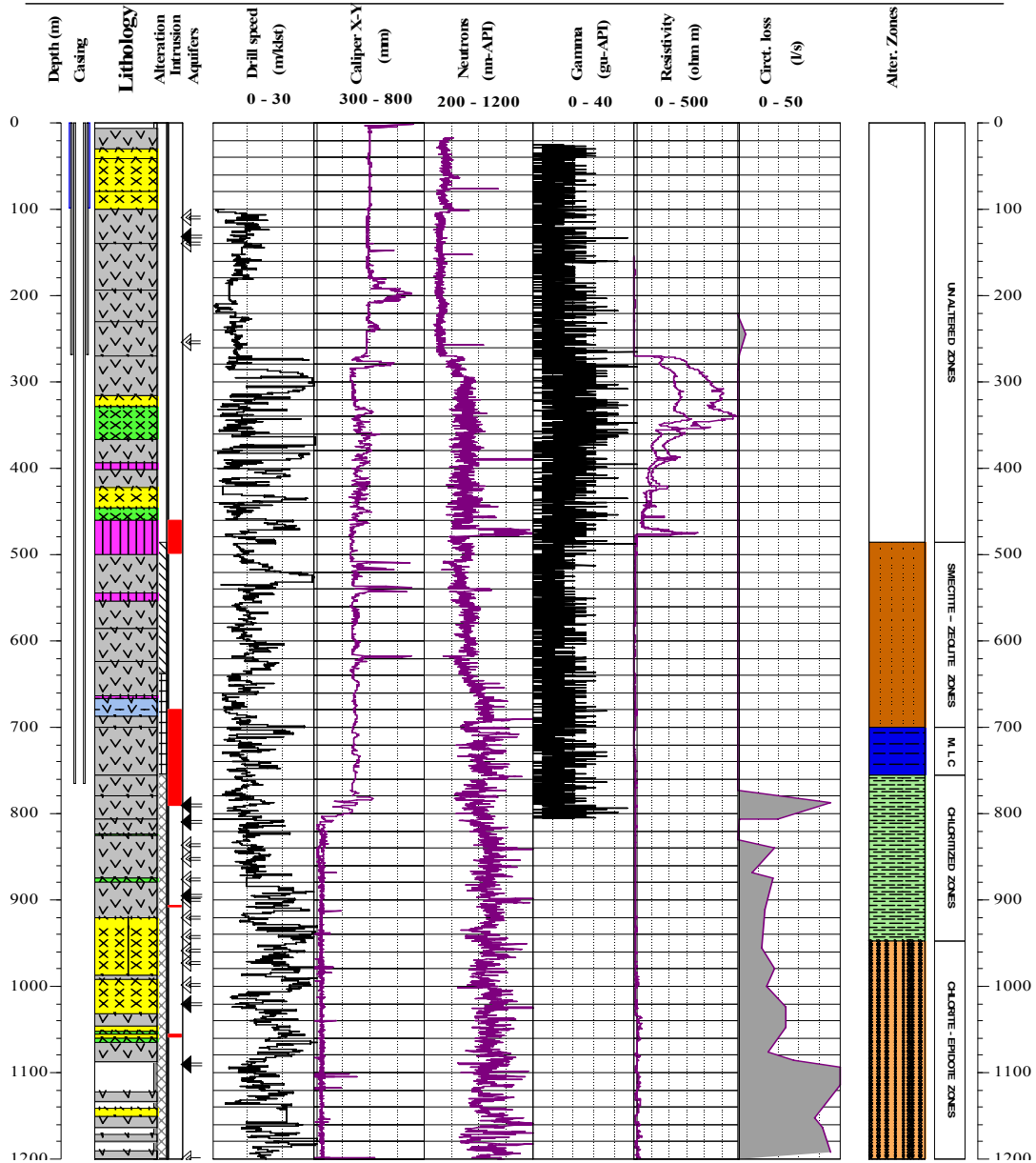


ISOR
ISLENSKAR DRUKUNANSOKNIIR

Drilling data Hellisheidi

HE-11

Place: Hellisheidi Drill: Jöttunn Circulation fluid : Mud and water Location nr.: 95111
Well: HE-11 Depth interval: 0-1200 m Project : Geologist :



LEGEND :

Rock Types		Alteration zone		Intrusions	
	Basaltic tuff		Un Altered zone		Intrusion
	Basaltic breccia		Smectite-zeolite zone	Alteration	
	Glassy Basalt		Mixed layer clay zone		Little alteration
	Fine-medium grained basalt		Chloritized zone		Medium alteration
	Tuffaceous sediment		Chlorite epidote zone		Much alteration
	No cuttings			Aquifers	
					Small-medium aquifer
					Large aquifer

FIGURE 9: A simplified stratigraphical section of well HE-11, with penetration rate and location of feed points, circulation losses, geophysical logs and alteration zones

formations formed under sub-glacial conditions and some apparent lava series which are described below; Intrusive rocks that dissect the well are dealt with in Section 3.3.

Hyaloclastite I (6 - 30 m)

This formation is red brown coloured, heavily oxidized, and contains limonite. It is plagioclase/olivine porphyritic. The formation (24-30 m) is underlain by glass-rich fresh hyaloclastite tuff and breccia.

Hyaloclastite II (30 - 100 m)

This formation is mostly made up of vesicular pillow lava of probable olivine tholeiite composition. It is plagioclase and sparsely olivine porphyritic. It is divided into fine- to medium-grained basalt (30-42 m), glassy basalt (42-80 m) and fine- to medium-grained basalt (80-100 m). Hydrothermal alteration is not seen but oxidation and limonite precipitation as an indication of groundwater percolation is considerable.

Hyaloclastite III (100 - 316 m)

This formation is dominated by basaltic tuff. It is fresh and brown coloured and highly porous. The tuff is plagioclase and olivine porphyritic. The vesicles occasionally contain iron oxide (limonite), aragonite and opal. The tuff in the top part of the formation may be reworked tuff sediment, but becomes more homogeneous and primary looking below.

Basaltic pillow lava I (316 - 328 m)

This formation is inferred as a pillow lava, as it consists mostly of glassy basalt to poorly crystallized basalt. Most of the vesicles and pores of the glassy basalt texture contain some siderite in the upper part (316-320 m depth) and radial carbonate in the deeper part (326-328 m depth). The sequence/unit consists of a mixture of poorly crystallized basalt and glassy basalt and basaltic tuff. The proportion of tuff in the unit decreases with increasing depth.

Hyaloclastite IV (328 - 394 m)

This formation consists of basaltic breccia which is a mixture of glassy basalt fragments and tuff. The rock is black to dark-green in colour. The vesicles contain some iron oxide, siderite, radial carbonate and opal. The plagioclase (crystal white) and olivine (yellow-greenish) are occasionally seen fresh as phenocrysts set in the microcrystalline to glassy groundmass. In the deeper part (384-394 m depth) fresh black glass is seen forming the outer margin of basaltic pillow lava.

Basaltic lava II (394 - 402 m)

The thickness of this layer is about 8 m, and includes relatively vesicular fresh fine- to relatively coarse grained basalt lava. It appears to be plagioclase, pyroxene and olivine porphyritic. Precipitation of limonite, siderite, and carbonate is seen in vesicles.

Hyaloclastite V (402 - 422 m)

The formation is dominated by brown-black-greenish basaltic tuff. The tuff is vesicular, which is to a large degree filled by secondary minerals of radial carbonate, iron oxide and siderite.

Hyaloclastite VI (422 - 500 m)

This formation is quite heterogeneous. The upper part is a mixture of glassy basalt breccia and light coloured tuff. The glassy vesicular basalt is to some extent filled by siderite and opal. Opal has commonly colloform and botryoidal texture. In thin section, at 486 m depth, clay alteration of the glass and olivine is noticeable, and clay precipitation is also seen in the vesicles. Pyrite starts to appear on grains and is disseminated in the groundmass at 494 m depth. This formation is mixed with a relatively fresh fine- to medium-grained basalt, containing plagioclase/pyroxen phenocrysts and is interpreted as an intrusion. Breccia dominates in the lower part of the succession.

Hyaloclastite VII (500 - 544 m)

The formation is dominated by a vesicular basaltic tuff, with a green to dark colour. The tuff is plagioclase-pyroxene-olivine porphyritic, set in microcrystalline glass. The majority of the glass and olivine have been altered to clay (smectite) and calcite. The vesicles in the tuff are filled by smectite, zeolite (scolecite/mesolite, stilbite) and calcite, while veins are commonly filled by calcite and pyrite. At 516 m depth, the tuff becomes slightly lighter in colour due to increased alteration, and pyrite becomes abundant, commonly disseminated in the groundmass. In the lower part this formation changes to tuff-breccia, with low to medium alteration.

Basaltic pillow lava III (544 - 554 m)

The thickness of this layer is about 10 m, and consists of dark-grey to brownish green fine- to medium-grained vesicular basaltic lava, probably pillow lava. The veins are mostly filled with calcite, pyrite, clay and minor iron oxide. Vesicles are to some extent filled by calcite, clays and zeolites (stilbite and mesolite). Pyrite is still abundant and becoming more dominant.

Hyaloclastite VIII (554 - 806 m)

The upper part of this formation is dominated by plagioclase porphyritic basaltic tuff, from dark- to light-green in colour in line with medium-high alteration. Vesicles in the basalt tuff are commonly filled by calcite, clay, zeolites (mesolite, stilbite, scolecite, laumontite) and occasionally pyrite. Veins are commonly filled with calcite, quartz, and pyrite. The tuff at 632-666 m depth shows higher alteration as indicated by the lighter colour. At 666-688 m depth, tuff appears to be reworked, with rounding and sorting of the grains and finely layered. Probable amorphous chalcedony starts at 686 m depth observed in the vesicles in this formation, and at 700 m depth, coarsely crystalline clay becomes highly pleochroic as observed in the petrographic microscope. Zeolites are not seen below 760 m, being taken over by wairakite and chlorite. The tuff is quite light coloured due to alteration but oxidation is also seen in places. At 664-666, 680-690, 762-772, and 778-792 m depth, the tuff mixes with fine- to medium-grained relatively fresh basalt, which in addition to the surrounding oxidation is interpreted as small intrusions. At the bottom of this unit wairakite becomes abundant from 780 m to the bottom of the analyzed section (1200 m depth) and is commonly associated with crystalline quartz. Pyrite remains abundant to the bottom on this formation either as big individual crystals or disseminated in the groundmass.

Hyaloclastite IX (806 m - 1192 m)

This formation is predominantly tuff down to about 920 m and is succeeded below by breccias and pillow basalt (glassy basalt). The rock is plagioclase porphyritic. The rock is highly altered, with abundant clay, chlorite, calcite and quartz. Several intrusions could be seen in this formation, fine- to medium-coarse-grained, and relatively unaltered compared to the surrounding rock. Prehnite, epidote, and wairakite appear in this formation. Vesicles are filled with quartz, calcite, chlorite and sometimes epidote, wairakite, prehnite and pyrite. Wollastonite starts to appear at 1088 m depth concomitant with the decrease of calcite.

3.3 Intrusions

Intrusions are younger, much less porous, less altered, darker coloured and often coarser-grained than the surrounding rocks. Narrow intrusions may be fine- to medium-grained. Geophysical logs are also useful in identifying possible intrusions, as they often show peaks in neutron-neutron, and resistivity logs, and usually also have a lower rate of penetration. They occur mostly at four different depth intervals.

Intrusive rock at 460 - 500 m

Fine- to medium-grained basalt sparsely plagioclase porphyritic. The basalt is fresh and unaltered. A high peak in the neutron-neutron log and low penetration rate further confirms this.

Intrusive rock at 680 - 792 m

In this interval the highly altered tuff mixes with fine- to medium-grained, dark coloured, relatively fresh basalt. Three high peaks in neutron-neutron log indicate the presence of three thin intrusions, especially at 690 m, 710 m, 760 m depth.

Intrusive rock at 906 - 910 m

This is medium- to coarse-grained, relatively fresh basalt. Plagioclase and minor pyroxene are seen as phenocrysts. A high peak in neutron-neutron logs is a further indication of this minor intrusion.

Intrusive rock at 1056 - 1060 m

Medium- to coarse-grained and relatively fresh basalt. Plagioclase and minor clino-pyroxene are seen as phenocrysts in the petrographic microscope. A high peak in neutron-neutron is a further indication of this being a small intrusion.

3.4 Comparison with geophysical logs

Geophysical logging during the drilling of a well is made firstly for the safe operation and control of the well, and secondly to obtain information on the structure and the physical properties of the geothermal reservoir penetrated by the well, both with respect to the rock formations and the geothermal fluids (Stefánsson and Steingrímson, 1990).

During and after the drilling of well HE-11, 4 types of geophysical logs, i.e. caliper, gamma ray, neutron-neutron and resistivity, were run. All geophysical logs of well HE-11 are shown in Figure 9. Due to the greater diameter in the upper part of the well, it was difficult to correlate the lithological logs with the lithology, but in deeper parts where the well diameter is smaller, proper correlation could be done. The main advantages of the geophysical logs provided in well HE-11 are described below:

Caliper log, which shows the diameter of the well, is necessary for correcting lithological logs. It investigates soft formations or cavities in the well, and may be used to locate intrusive contacts and fractures, e.g. at 200, 510, 540, 620, and 1110 m depth.

Neutron-neutron log gives information on the porosity of the rock. It is a direct measure of the H₂O content in the rock formation, where high amplitude reflects dense rock with low water content. It shows high values at around 390, 480, 680–760, 840, 906–910, 960, and 1056–1060 m depth intervals, which match with dense rocks which probably are intrusions.

Natural gamma ray log reflects the chemical composition of the rocks, (sensitive to ⁴⁰K, U, Th), and is especially used for detecting evolved rocks. The log of HE-11 does not show high values, indicating that there is no acid rock in the well.

Resistivity log reflects the porosity of the rock, alteration and fluid temperature. Therefore, resistivity is high alongside the dense intrusive rocks, but low in porous more altered hyaloclastites and highly permeable zones within the well. It shows high values at 270–480 m depth, indicating relatively low alteration of the rocks and very low resistivity below as an indication of highly altered hyaloclastite rocks.

4. HYDROTHERMAL ALTERATION

When the geothermal fluid ascends through a reservoir to the surface a water-rock interaction will occur striving towards equilibrium condition. This interaction produces a suite of secondary minerals whose identity and abundance depends on the prevailing physical and chemical conditions. To

measure how completely a rock has reacted with the fluid, the intensity of alteration is used (e.g. Browne, 1978). The intensity of alteration around well HE-11 as seen in Figure 9 shows that the degree of alteration increases with depth. The important processes involved in hydrothermal alteration are replacement of primary or older hydrothermal minerals, direct deposition in veins, dissolution or leaching of primary minerals and re-crystallisation. Overall factors affecting hydrothermal alteration are temperature, permeability (related to gas contents and hydrology of a system), fluid composition (pH value, gas concentration, vapour- or water-dominated, magmatic, meteoric), initial composition of rocks, duration of activity (immature, mature), number of superimposed hydrothermal regimes (overprinting of alteration), and hydrology (variable) (Reyes, 2000). In basaltic formations in Icelandic high-temperature systems, the alteration intensity temperatures are lower than 200°C, and low-temperature zeolites and smectite (clay) are formed. The range where low-temperature zeolites and smectite are abundant is called the smectite-zeolite zone. In the temperature range above 200°C, the low-temperature zeolites disappear and the smectite is transformed towards chlorite in a transition zone, the so-called mixed-layered clay zone where smectite and chlorite coexist in a mixture. The chlorite zone is characterized by the appearance of chlorite and the progressive disappearance of smectite, occurring at temperatures around 230°C. At higher temperatures of about 240-250°C, epidote appears which marks the so-called chlorite-epidote zone. Comparison of estimated formation temperatures deduced from temperature logs, and the alteration mineralogy can be used to tell about present cooling or heating or if it has occurred recently (Árnason et al., 2000).

4.1 Rock alteration

The most common volcanic rock in Iceland is basalt and the primary mineral compositions are: Plagioclase, olivine, pyroxene and opaques (magnetite-ilmenite). Under the influence of hydrothermal fluids, primary minerals are gradually replaced by alteration minerals as follows:

1. Glass alters mostly to clays, sphene (titanite), epidote, calcite, zeolites and quartz.
2. Olivine alters to clays, sphene, calcite or sulphides.
3. Pyroxene is relatively resistant to hydrothermal alteration but mainly alters to clays.
4. Plagioclase is also relatively resistant to alteration but starts to alter to clay minerals and in accordance with increasing temperature to albite, calcite, sphene, chlorite, epidote and adularia.
5. Iron ore minerals alter to sphene and sulfides (usually pyrite).

In general, the order of decreasing susceptibility to alteration is glass, olivine, plagioclase and pyroxene.

The rocks in well HE-11 remain remarkably fresh down to 486 m depth, as observed where glass remains fresh to some extent down to that depth. Glass is completely altered to relatively dark-green clay at 486 m depth, but become lighter green in colour with increased alteration. Zeolite, quartz, calcite, clay and sphene alteration of the glass is common. The primary plagioclase phenocryst composition is commonly labradorite. It starts to alter at 590 m depth (<10%), with partial alteration into clays and chlorite. Medium to high alteration intensity (50–80%) is found at 830–1192 m depth, where it alters mainly to clay, calcite, quartz, albite, wairakite, chlorite, wollastonite and epidote. Pyroxene, the mineral most resistant to alteration, starts altering at 756 m depth into clay (<10%). Medium to high alteration intensity (<50–80%) is found at 1010–1192 m depth, altered mainly to clay, calcite and chlorite. Opaque minerals, (ilmenite/magnetite) commonly a part of the groundmass show a variable alteration in the well. At several locations it is altered by pyrite (494 m depth) and sphene (780 m depth). Geothermal systems within Hengill central volcano contain dilute fluids (Fridleifsson et al., 2003).

4.2 Distribution of hydrothermal alteration

In Icelandic geothermal systems, glass and olivine are the first minerals that alter at temperatures about 50°C; and at temperatures above 200°C both of them are usually completely altered (Figure 10). The results from cuttings, petrography and XRD analyses show that clay and calcite are the most common secondary alteration minerals down to 510 m depth in well HE-11. Siderite is also a common secondary mineral at depths between 182 and 480 m indicating relatively cold groundwater, with temperatures less than 50°C (Hjalti Franzson pers. com.). A sharp change to high-temperature alteration occurs at about 650 m depth in the well indicating a temperature increase to above 200°C. Various alteration minerals are found in this well. Figure 11 shows the distribution of these minerals within the stratigraphical formations described earlier in Section 3. They mostly occur as vesicle and vein infillings, and as replacement minerals within the rock matrix. Calcite, clays and pyrite are by far the most common followed by zeolites, quartz and other minerals. The hydrothermal alteration minerals identified from the cutting analysis are described below.

Carbonates are very common minerals in this well, showing a wide range of occurrences and being especially abundant within hyaloclastite series, but less common in dense basaltic bodies. Above 450 m depth they mostly occur as radial siderite, undefined radial carbonate, aragonite, or scattered dogtooth calcite. Calcite becomes the dominant carbonate below 490 m depth, first as vesicular infillings, then also as vein infilling and replacement mineral of mostly glass and later of plagioclase phenocrysts. It is easily identified in the cuttings by reacting with HCl acid. In thin-section it first appears at 486 m depth replacing glass and as a vein infilling. Platy calcite was observed filling up fractures and as an intergrowth with other minerals, common from about 780 m depth down to the bottom. It is suggestive of boiling in a geothermal system (Hjalti Franzson pers. com.).

Opal is pale yellow-brown coloured, transparent to sub-translucent, and in thin section often found as vesicle linings, in veinlets and as cavity fillings. It was found between 170 and 486 m depth, indicating a temperature below 100°C.

Chalcedony is identified in binocular microscope at 600, 612, 666 and 690 m depth occurring as cavity and vesicular linings at places associated with clay, quartz and zeolites. It also forms an aggregate structure and gradually changes to quartz at higher temperatures.

Quartz was rarely found as groundmass alteration but commonly as open space filling, often as crystal growth in the clay-lined or calcite-filled vesicles and is often found intergrown with wairakite (e.g. at 950 and 1044 m depth). It was first identified in a cutting analysis from 686 m depth and persists to the bottom of the well. In thin section it is seen as a replacement of plagioclase. Euhedral prismatic quartz crystals are abundant in veins and vesicles of the hyaloclastite formations, but is hardly seen in the groundmass as a secondary mineral.

Zeolites. The zeolite group includes hydrous calcium aluminium silicates that commonly occur as secondary minerals in the rock cavities, especially basalt (Kerr, 1959). The zeolite group has been identified by binocular and petrographic microscope, and commonly indicates low-temperature conditions (<100-120°C), except laumontite (120-200°C) and wairakite (200-300°C). In well HE-11,

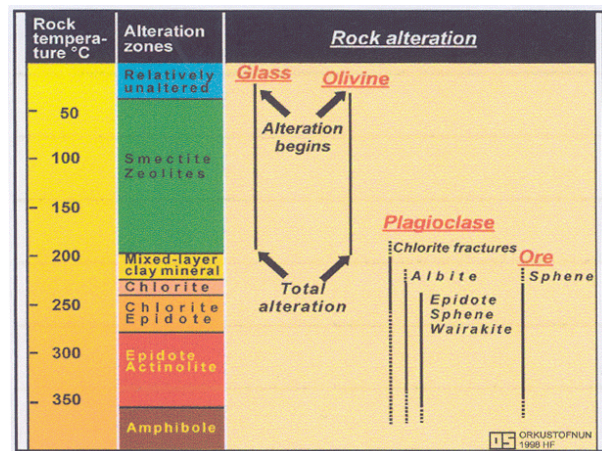
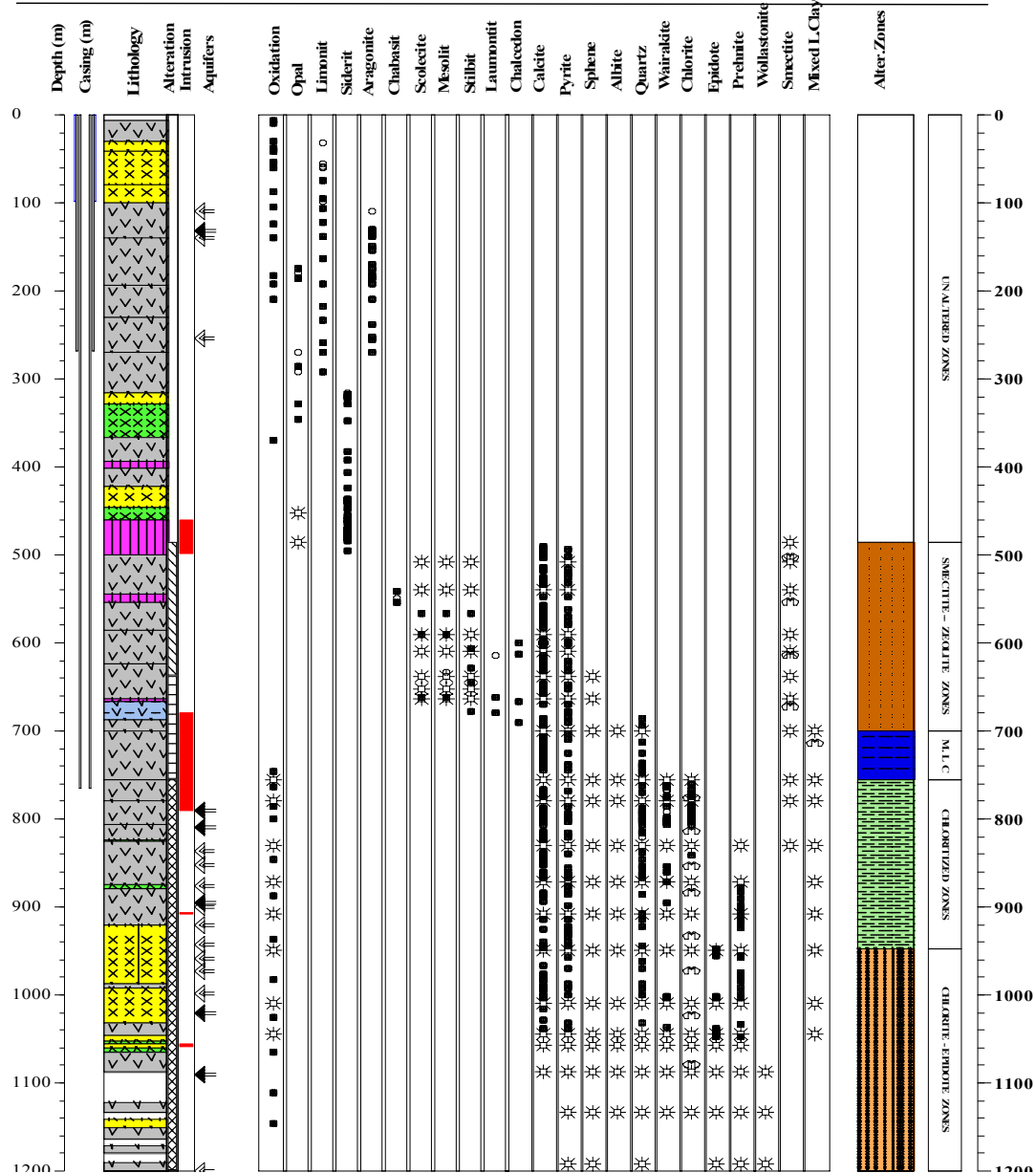


FIGURE 10: Temperature related alteration zones and rock alteration in Icelandic geothermal systems (Franzson, 1998)



Drilling data Hellisheidi HE-11

Place: Hellisheidi Drill: Jötunn Circulation fluid: Mud and water Location nr.: 95111
 Well: HE-11 Depth interval: 0-1200 m Project: Geologist:



LEGEND :

Rock Types

- Basaltic tuff
- Basaltic breccia
- Glassy Basalt
- Fine-medium grained basalt
- Tuffaceous sediment
- No cuttings

Alteration zone

- Un Altered zone
- Smectite-zeolite zone
- Mixed layer clay zone
- Chloritized zone
- Chlorite epidote zone

Intrusions

- Intrusion

Alteration

- Little alteration
- Medium alteration
- Much alteration

Aquifers

- Small-medium aquifer
- Large aquifer

FIGURE 11: Secondary mineral deposition and mineral alteration zones in well HE-11
 (■ from binocular, ☼ thin section, ∞ XRD)

the identified zeolites are stilbite, chabazite, laumontite, mesolite, scolecite and wairakite. Zeolites with the exception of wairakite are found above 700 m depth. The following describes zeolite groups as encountered in the well:

Chabasite. It is the lowest temperature zeolite, identified by nearly cubic-shaped crystals. This mineral was found at 542, 550, and 554 m as vesicle filling in basaltic lava.

Stilbite. This mineral was detected in thin sections from 500 to 700 m depth. Stilbite indicates temperature of 80–120°C.

Mesolite/Scolecite are characterized by radial texture and are commonly found both in binocular and thin-section analyses at depths of 500-700 m, indicating minimum temperatures of 90°C.

Laumontite. It has a prismatic and fibrous structure and is mainly found as cavity fills. Laumontite was found at depth of 610-688 m in vesicles of basaltic tuff, and indicates temperatures of 120-200°C.

Wairakite is the calcium analogue of analcime and occurs in a distinctive trapezohedral crystal form. It has a very low relief in thin section, a dull dark-gray colour and characteristic cross-hatching twinning. Fluid inclusions in its crystal structure usually give wairakite a typical cloudy look. It is first seen at 756 m depth and becomes abundant from 780 m depth to the bottom of the sampled section of the well. The appearance of wairakite suggests temperatures above 200°C and has been shown to exist at temperature as high as 300°C (Kristmannsdóttir, 1979).

Prehnite in thin-section is colourless, looks similar to epidote except that it has no pleochroism. It occurs from 830 m depth to the bottom of the well, often associated with epidote, inside vesicles of hyaloclastite tuff and breccia.

Epidote is yellowish–green crystal in cuttings and famous in high-temperature geothermal system, because it is easily identified and indicates temperature above about 250°C. In thin-section it appears colourless to yellowish green and is slightly pleochroic in granular to columnar aggregates. It was first identified at 950 m depth and extends to the bottom of the well. It was found to be associated with wairakite, chlorite, quartz, and calcite and was often grown together with prehnite. Precipitation of epidote in appreciable quantities indicates temperature above 250°C (Kristmannsdóttir, 1979).

Wollastonite is a calcium silicate, and is identified in thin-section as fibrous aggregates with fairly high relief and weak birefringence. In the binocular microscope it is seen as a whitish woolly mineral. It was first identified at 1088 m depth as finely radiating crystals in close association with quartz and epidote. It appears to form hydrothermally at rock temperatures near to and above 270° C.

Albite is found in basaltic rock as an alteration product of Ca-rich plagioclase (labradorite). In thin-section albite can be distinguished from primary plagioclase by their lower refractive index and extinction angle. It was first identified at 700 m depth as a replacement of plagioclase, and extends to the bottom the well. Albite is observed at rock temperatures about 150° C.

Sphene/titanite is generally formed as a replacement of opaque minerals (magnetite, ilmenite), but sometimes also as a replacement of glass, pyroxene and plagioclase. It was identified in petrographic microscope, where it is seen as red-brown and somewhat cloudy coloured crystals. It is first seen at 638 m depth as replacement glass and at 1044 m depth. Sphene is first seen in altering opaque mineral. This mineral indicates poor permeability in the Philippines (Reyes, 1990)

Iron oxides, commonly siderite and limonite, were found in the upper part or down to 450 m depth, often showing reddish tints covering the wall and spheroidal deposition inside vesicles (limonite). It indicates relatively cold groundwater, with temperatures less than 50°C. It is also found at deeper

levels between 740 and 1144 m depth, and is there interpreted as a thermal metamorphic effect from a nearby intrusion.

Pyrite is the most common iron sulphide mineral occurring as disseminations and fracture-fillings, together with calcite and quartz. It becomes a common mineral below 494 m depth, and extends to the bottom of the well. Pyrite is both found as a deposition and as a replacement of opaques. It is very easily identified in the binocular microscope, where it shows distinctive brass yellow colour and metallic cluster. Abundant pyrite is found to be indicative of permeability.

Chlorite in the cuttings is commonly light green, and often associated with high-temperature alteration minerals like epidote, prehnite and wollastonite. In thin-section it is coarse-grained and light green coloured in plane polarized light in the pores, grey-blue in cross polarized light, shows tiny green needles texture and sometimes radial forms. In thin-section it was first identified at 756 m depth and extends to the bottom of the well. In XRD it shows peaks of 14.4 Å and 7.1 Å, at 782 m depth, which do not change upon glycolation and heating (see graph in Appendix III).

Clay minerals are very important alteration the most common product of basaltic glass, olivine, pyroxene and plagioclase. The clays in the rocks of well HE-11 are both fine-grained clays (smectite) and coarse-grained clays (mixed layer clays and chlorites) observed in vesicle and vein fillings.

Smectite, is green to yellow-brownish. In thin-section it has light green colours in plane-polarized light, slightly pleochroic and is found as vesicular linings and alteration of glass. Smectite is formed in low-temperature environment and associated with zeolite minerals indicating temperature less than 200°C. In thin-section it starts to appear at 486 m and continues down to 700 m depth. Smectite in a XRD analysis from 510 m depth, shows peaks at about 15 Å for untreated samples and 17 Å and 10 Å for glycolated and heated samples, respectively.

Mixed layer clays, are found as an inter-layering of smectite and chlorite clay minerals indicating temperatures between 200 and 230°C. In thin-section at 700 m depth the clay shows pleochroism from light to darker green colours in plane polarized light, but in cross-polarized light it shows strong yellow brown colours. It appears as a coarse-grained clay mineral, coarser than smectite. In XRD analysis at 720 m depth, the sample shows peaks about 14.2 Å for untreated samples, 14.2 Å for glycolated ones, and 10 Å and 14.2 Å for heated samples.

4.3 Alteration mineral zonation

The degree of hydrothermal alteration in the rocks depends upon certain factors such as rock type, permeability, temperature, chemistry of invading fluids and duration of geothermal activity. According to Browne (1978, 1984), minerals commonly used as geothermometers are zeolites, clays and amphiboles. In Icelandic geothermal fields, most zeolites are common before 100°C and disappear around 200°C, where they are transformed to wairakite or quartz or dissolve. Quartz is found at temperatures above ca. 180°C, and calcite appears to be stable up to temperature around 280-290°C. Wairakite forms at temperature above 200°C, epidote >250°C, prehnite >250°C, wollastonite >270°C, while temperatures of <200°C are seen for smectites, 200-230°C for mixed layer clays, and chlorite starts forming at temperatures above 230°C. Actinolite appears at temperatures above 280°C. The correlation of alteration zones with rock temperature in active high temperature areas in Iceland is shown in Figure 10, (Kristmannadóttir, 1979, Browne, 1984, Franzson, 2000.). By using the XRD for analysing 18 clay samples, 4 alteration zones were defined. The results for the HE-11 borehole clay samples are shown in table in Appendix II.

On the basis of the index mineral and associated mineral assemblages seen in well HE-11, four alteration zones are identified. These zones are identified by the first appearance of the index minerals. The following is a description of the alteration zonation with temperature approximations.

Unaltered/fresh zone. This zone is partly based on the assumption that a very low-temperature condition must have prevailed prior to hydrothermal geothermal activity at the Hengill central volcano. Such a condition would include a 400-500 m thick cold groundwater layer underlain by a succession suffering a much higher thermal gradient. The alteration minerals could include ferro-oxidized, limonite, siderite, opal and radial carbonate crystals.

Smectite-zeolite zone. This zone is marked by the presence of smectite (fine-grained clay) and zeolites, mainly chabazite, stilbite, scolecite/mesolite and laumontite. The temperature is below 200°C. This zone extends from 486 to 700 m depth. The overlying 486 m Postglacial basalt lavas are generally still fresh and unaltered.

Mixed-layer clays (MLC) zone is made up of an assemblage of smectite, and mixed layer clays (coarse-grained clays) along with pyrite, calcite, quartz, albite, and wairakite. It ranges from 700 down to around 756 m depth, where chlorite first appears. The characteristic temperature range of this alteration zone in Iceland is about 200–230°C.

Chlorite zone. This zone is indicated by the presence of chlorite and supported also by calcite, pyrite, wairakite, sphene, albite and prehnite. The top of the chlorite zone was found at 756 m and extends down to 950 m depth. The temperature range of this zone extends from 230 to about 250°C.

Chlorite-epidote zone. It is clearly marked by the presence of epidote and chlorite. It is also supported by sphene, prehnite, pyrite, wairakite, albite and wollastonite. It extends from 950 down to 1192 m depth where total loss circulation occurred in this well. Temperature in this zone is above 250°C.

4.4 Mineral times sequences

The sequence at which deposition minerals precipitate in veins and vesicles informs on the changes occurring in the geothermal system through its life. Furthermore the degree of homogeneity of the mineral sequence within a system also informs on whether the system acts as one reservoir or whether it divides into subsystems. Mineral sequences are partly observed in the binocular stereomicroscope, but largely through the petrographic microscope.

The depositional sequences observed in thin-section are shown in Table 1. The mineral sequence can be divided into two depth ranges; above and below about 650 m depth, which is also the division between the groundwater - caprock above and the true high-temperature reservoir below. In the upper part the sequence starts with fine-grained clay lining the vesicles followed by zeolites (scolecite+stilbite) and then coated by calcite.

The overall mineral sequence in the lower part starts with platy calcite, followed by coarse-grained clays (MLC-CHL) and then succeeded by quartz+wairakite. Later precipitation of prehnite is succeeded by epidote and lastly by wollastonite. Calcite is only in one instance seen overprinting quartz, but its relation to prehnite, epidote and wollastonite is not known. Overall the sequence shows that minerals with higher temperatures tend to form later, and that would suggest that the reservoir is heating up with time. More data is, however, needed from other parts of the geothermal reservoir in order to confirm this evolutionary trend.

5. AQUIFERS

An aquifer is defined as a zone in the well where an unquantifiable volume of circulating groundwater is stored. Groundwater movement is controlled by a hydraulic gradient, formation permeability,

5.1 Aquifers at shallow depth

These aquifers extend from the top to about 765 m depth, to the bottom of the production casing, 13 $\frac{3}{8}$ " was set. The aquifers can be described as follows:

Aquifer 1

Located at 100–140 m depth and relates to contact between fine- to coarse-grained basalt and basaltic tuff and breccia which is highly porous, more permeable and contains higher percolating water than the upper tuff. This is probably a relatively small aquifer.

Aquifer 2

Located at 245–269 m depth, where 1–4 l/s circulation loss occurred. It is a small aquifer and occurs within a layer of hyaloclastite tuff.

5.2 Aquifers in the production part of the well

Many aquifers are located in the production part of the well which extends from 780 m to the bottom. Circulation losses start to occur below 810 m depth and become total below 1192 m depth. The circulation losses are shown in Figure 9. About 15 l/s loss is measured at 810 m and increases to about 45 l/s at 1090 m. At 1192 m total circulation occurred and remained so until 1652 m where drilling was terminated. It is clear from the temperature profile in Figure 12 that an inflow of hotter water, starts at 780 m, and further points of inflow are shown at several depths below that. The well was sited and directed towards the 5000 y old volcanic fissure. The feed points at about 800–1200 m depth are about 50–100 m away from the proposed volcanic feeder (dyke) if one assumes that the latter is vertical. However, the bottom aquifer at 1500 m, beyond the scope of this study, coincides roughly with the likely position of the volcanic feeder. The relatively high permeability of the zone below 780 m coincides with relatively high mineral deposition and rock alteration, such as abundant calcite, pyrite, albite, quartz, oxidation (which might indicate the proximity of intrusions), wairakite, prehnite, epidote and wollastonite.

A brief description of each aquifer in the production part of the well is given below:

Aquifer 1

Located at 780–810 m depth. The formation is highly altered, vesicular, hyaloclastite basaltic tuff, with abundant big cubic pyrite and albite. A relatively high circulation loss of 20 l/s occurred at 780 m depth which increased to 45 l/s at 800 m depth. Notable inflow into the well as indicated from the temperature log in Figure 12. Notable oxidation occurs below 760 m depth (c.f. Figure 11) which may indicate nearby intrusives, possible contributors to the relatively high permeability.

Aquifer 2

Located at 837 m depth. The formation is very high altered hyaloclastite basaltic tuff especially by oxidation and albite. Where the rock is more vesicular deposition is characterized by big pyrite, calcite and chlorite. The high oxidation may imply a connection to an intrusive. A circulation loss of 18 l/s occurred there and a temperature maxima on the temperature curve.

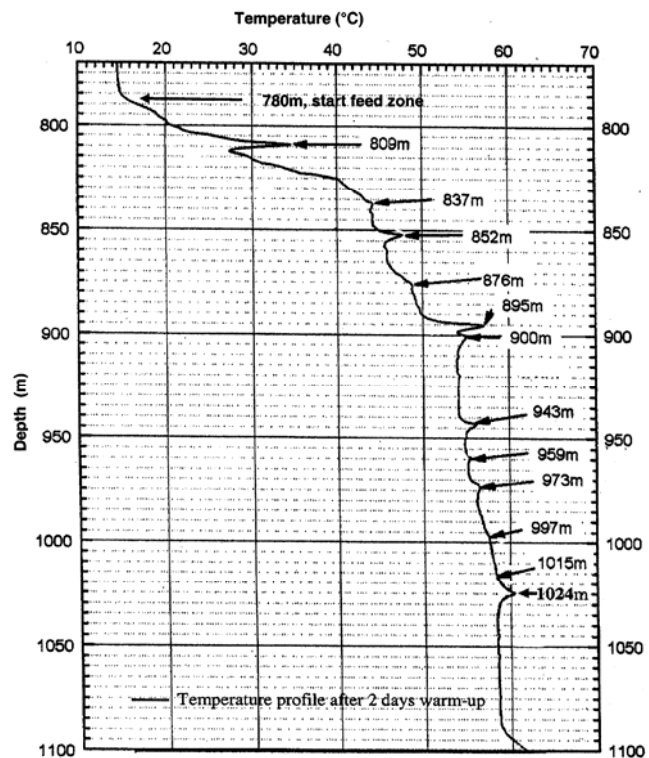


FIGURE 12 : Temperature log after 48 hours warm up, indicating many possible aquifers (Jónsson et al., 2004)

Aquifer 3

Located at 869–876 m depth. The location coincides with a boundary between highly altered vesicular basaltic tuff and breccia. Abundant vesicles are filled with chlorite, calcite, pyrite, and albite replacement of plagioclase phenocryst. Increased circulation loss was not observed coinciding with this aquifer, but the location is set by the temperature log (Figure 12).

Aquifer 4

Located at 895 and 900 m depth. The vesicular basaltic tuff is very highly altered and filled with chlorite, quartz, calcite, pyrite, and albite. No additional circulation losses were noted at this depth.

Aquifer 5

Located at 943 and 959 m depth, and coincides with an intrusion at that depth. No additional circulation loss was measured but its presence is indicated by a minor temperature maxima as shown in Figure 12.

Aquifer 6

Located at 973–997 m depth according to the temperature log. The formation is a highly altered vesicular glassy basalt and underlain by a basaltic tuff. No new circulation loss occurs at this depth. Abundant vesicles are filled with chlorite, wairakite, pyrite and calcite.

Aquifer 7

Located at 1015–1024 m depth. The formation is a highly altered vesicular glassy basalt, where vesicles are filled with chlorite, wairakite and pyrite. Albite replacement of plagioclase is very common. An increase of 5 l/s circulation loss was measured from about 18 l/s at 1015 m depth to 23 l/s at 1024 m depth.

Aquifer 8

Located at 1076–1094 m depth. It occurs within a highly altered basaltic tuff, abundant with vesicles filled with epidote and pyrite, chlorite and wairakite. An increased circulation loss occurs from 1076 down to 1094 m depth of some 12 l/s, while indications in temperature logs are not clear.

Aquifer 9

Located at 1115 m depth. A total circulation loss occurred at 1115 m depth (>30 l/s). The drilling continued with total loss of circulation until 1652 m, the final depth of this well. A further indication of the aquifer is a distinct cave shown by the caliper log (c.f. figure 9) at 1100-1105 and another smaller one at about 1120 m depth. It is difficult to ascertain the geological connection of the aquifer due to the lack of cuttings.

Aquifer zones at deeper levels

This study is focussed on the borehole geology above 1200 m depth. However, it must be pointed out that temperature logs indicate that the inflow of the aquifers above 1100 m flow into and down the well and disappear into the formations below 1400 m depth. Breakpoints in temperature logs indicate that the aquifers are located at about 1420, 1520 (most notable), 1580 and 1605 m depths. These aquifers appear to coincide with the proposed location of the volcanic feeder of the 5000 years old eruption.

6. DISCUSSION

The well HE-11 is located in the central part of Hellisheidi field, within the Hengill central volcano, where volcanism is most intense and where hyaloclastite formations have formed highlands. Interglacial lavas, when erupting in the highlands will flow down hill and accumulate in the lowlands surrounding the volcano and, for that reason, the stratigraphy of well HE-11 is predominantly of

hyaloclastites or about 90%. These are basaltic tuff, basaltic breccia and glassy basalt (pillow lava) and formed during sub-glacial eruptions. The boundary of the lava series is inferred to lie about 1000 m b.s.l., where an age assessment of the boundary was about 400,000 years. This would suggest that the age of the Hengill central volcano may be somewhat older or around 400,000 year (Franzson et al., 2005).

The geological and hydrothermal alteration study shows that the degree and intensity of rock alteration and alteration mineral distribution of well HE-11 increases with depth (Figure 9 and 11). Temperature is assessed from two data sources, first by investigation of hydrothermal alteration minerals during the drilling period (alteration temperature) and second by formation temperature, calculated from temperature logs during the warm-up period. The distribution of hydrothermal alteration minerals show that the uppermost 450 m are controlled by a groundwater system of relatively low temperature with negligible deposition of minerals. Below that there is a rapid change towards high-temperature condition as shown by the alteration assemblage. The time related mineral depositional sequence shows that the geothermal system where well HE-11 is drilled indicates a progressive heating with time, as far down as geological information reaches. An interesting relation is observed when comparing the alteration and assessed formation temperature curves, where the measured temperature is considerably higher than the alteration temperature, (c.f. Figure 13), implying that the geothermal system in the upper part of this well has recently heated up, and that the hydrothermal alteration has not yet reacted to that change. It is proposed that the heating up relates to the renewed geothermal outflow along the 5000 years old volcanic fissure eruption. An approximate equilibrium exists between alteration and formation temperature below 1000-1200 m depth. A comparison below 1200 m is not possible due to a lack of cuttings. Data from other wells in the neighbourhood show that the formation temperature decreases below the alteration temperature, indicating cooling (Franzson et al., 2005).

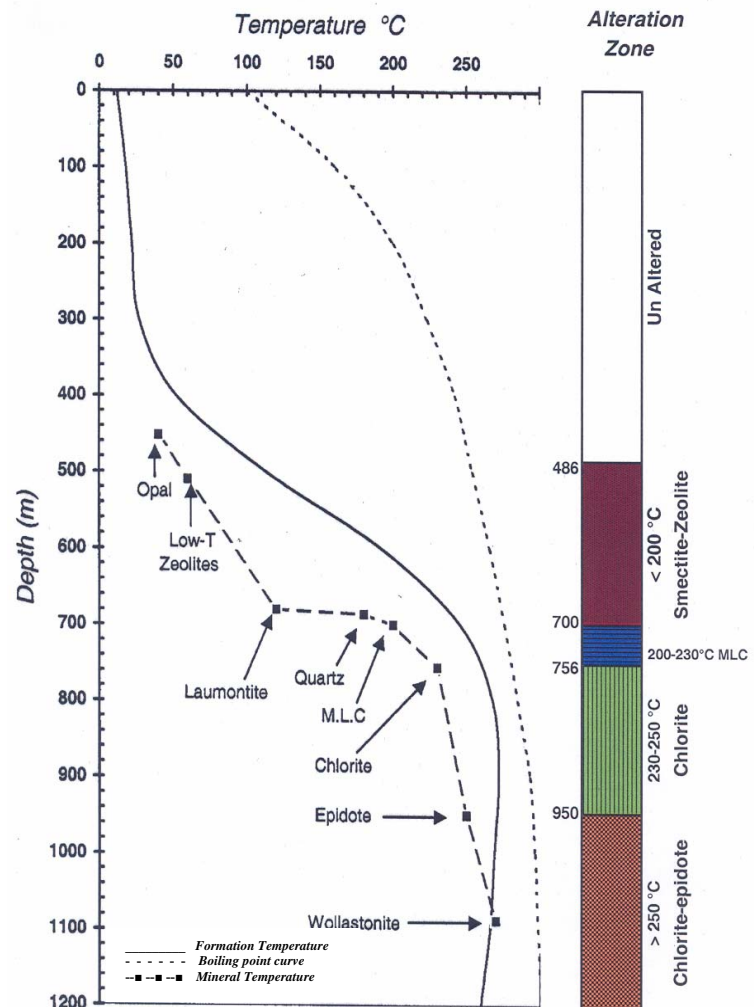


FIGURE 13: Correlation of hydrothermal alteration, formation temperature and boiling point temperature in well HE-11 (modified from Jónsson et al., 2004)

The Hellisheidi high-temperature anomaly is surrounded by a colder groundwater system. A consequence of permeability change at such a location may essentially be twofold. If the pressure of the geothermal system is higher than the surrounding cold groundwater system, the permeability change will result in an outflow of the former and heating up will occur. If the pressure difference favours the surrounding groundwater system, it will intrude into the geothermal system and cause

cooling. Such heating/cooling will be more rapid and extreme at locations of high permeability (Franzson, 2000).

Many aquifers are located in production part from 780 down to 1020 m depth. Thin volcanic intrusions occur at least at 3 locations, and common oxidation (as an indication of neighbouring intrusions) in the same interval, would suggest a connection of permeability with dyke intrusions. These are, however, 50 -100 m away from the location of the main volcanic feeder of the 5000 years old eruptive fissure. The bottom aquifers at 1400-1600 m depth are on the other hand believed to coincide with the aforementioned fissure. Thus the majority of the aquifers encountered in the wells can be directly related to margins of intrusions indicating the dominance of fracture permeability in the geothermal reservoir.

7. CONCLUSIONS

The following conclusions can be deduced based on geothermal evidence from the study of the uppermost 1200 m of well HE-11:

- The general stratigraphy of well HE-11 is predominantly hyaloclastites, consisting of basaltic tuff, breccia and glassy basalt (pillow lava), erupted during glacial times/sub-glacial. A few basaltic lavas are believed to exist, probably pillow basalts bodies. Several intrusions were also encountered.
- According to the distribution of alteration minerals, five alteration zones were found, a zone of no alteration (from the top to 486 m), a smectite-zeolite zone (486–700 m), a mixed layer clay zone (700–756 m), a chlorite zone (756–950 m) and a chlorite-epidote zone (950–1192 m). Their relevant temperature ranges are < 40, 40–200, 200–230, 230–250, > 250°C, respectively.
- The time related mineral depositional sequence shows a geothermal system above 1200 m is undergoing a gradual progressive temperature increase with time.
- Comparison of estimated formation temperatures in the hydrothermal system, with fossil temperatures in the rock formation as suggested by the secondary mineral study, implies that the hydrothermal system has recently heated up in the upper part but is at equilibrium below about 1000-1200 m depth. Data from other wells suggest that cooling may be taking place below about 1500 m depth.
- Many aquifers are located in the production part of the field and can be directly related to margins of intrusions indicating the dominance of fracture permeability in the geothermal reservoir.

ACKNOWLEDGEMENT

I would like to express my thanks to Dr. Ingvar B. Fridleifsson and Mr. Lúdvík S. Georgsson, Director and Deputy Director of the UNU Geothermal Training Programme for giving me the opportunity to participate in this special course and Mrs. Gudrún Bjarnadóttir for her efficient help and kindness during the training course.

My special thanks go to Dr. Hjalti Franzson my adviser, for his patient guidance and shared knowledge and experience in borehole geology, as well as for the assistance and supervision of this work. Thanks to UNU lecturers and ISOR staff, especially in the Geoscience division to Dr.

Gudmundur Ó. Fridleifsson and Mr Ásgrímur Gudmundsson for sharing their knowledge and valuable discussions regarding the report subject. Many thanks also to Anette K. Martensen for her kind and generous help sharing knowledge to analyse the cuttings and to Mr. Sigurdur Sveinn Jónsson for his kind help in the preparation of clay samples for XRD analysis.

I would also like to give thank to the management of Geothermal PERTAMINA Energy for support and granting permission to attend training in this programme.

Thanks are not complete if no mention is made of the deepest admiration and love of the author for his wonderful wife, Marthatia Utami Dewi and children, Nurul Annisa Shafira Maheswari and Muhammad Satria Bima Yudha, who await my arrival after six long months of absense from home. It would have been impossible for me to go through this training without your support, prayers and encouragement. And lastly, to the Almighty God, who made all of these things possible and worth remembering.

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APPENDIX I: Preparation of sample minerals for analysis by the XRD technique

1. Place approximately two teaspoons of drill cuttings into a test tube, wash out dust with distilled water. Fill the tube 2/3 with distilled water and plug with rubber stopper. Place the tube in a mechanical shaker for 4-8 hours, depending on the alteration grade of the samples.
2. Remove the tube from the shaker and allow to settle for 1-2 hours, until particles finer than approximately 4 microns are left in suspension. Pipette a few mm from the tube, below the level of the sample, and place about ten drops on a labelled glass slide. Avoid having the samples thick. Make a duplicate for each sample and let dry at room temperature overnight.
3. Place one set of samples in the desiccator containing Glycole ($C_2H_6O_2$) solution and the other set in a desiccator containing $CaCl_2 \cdot 2H_2O$. Store at room temperature for at least 24 hours. Thick samples will need a longer time in the desiccator, at least 48 hours.
4. Run both sets of samples from 2-14° on the XRD.
5. Place one set of samples (normally the glycolated one) on an asbestos plate and heat in a preheated oven at 500-550°C. The exact location of individual samples on the asbestos plate must be known before heating because labelling will disappear during the heating process. Cool the samples sufficiently before further treatment.
6. Run the samples from 2-14° on the XRD.

APPENDIX II : Results of the XRD analysis of clay minerals

Depth (m)	Untreated (Å)	Glycolated (Å)	Heated (Å)	Results
240	-	-	-	No Clay
340	-	-	-	No Clay
390	-	-	-	No Clay
410	-	-	-	No Clay
450	-	-	-	No Clay
510	15	17	10	Smectite
560	14.8	16.9	9.9	Smectite
620	12.8	13.5	9.9	Smectite
680	15.1, 12.7	13.4	9.9	Smectite
720	14.2	14.2	10.1, 14.2 (low)	MLC, Chlorite
782	14.3	14.3	14.3	Chlorite
820	14.3	14.5	14.1	Chlorite
860	-	-	-	No Clay
890	14.3	14.3	14.3	Chlorite
940	14.3	14.3	14.3	Chlorite
980	14.3	14.3	14.3	Chlorite
1030	14.3	14.3	14.3	Chlorite
1086	14.3	14.3	14.7	Chlorite

APPENDIX III: Characteristic XRD patterns for the clay minerals of well HE-11

