BOREHOLE STRATIGRAPHY AND ALTERATION MINERALOGY OF WELL HE-6, HELLISHEIDI, SW-ICELAND

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ABSTRACT

Well HE-6 is located in the Hellisheidi high temperature geothermal field in southwest Iceland. Drilled in 2002 to explore the conditions of the geothermal system, well HE-6 is now discharging at a rate of 29.4 kg/s with an enthalpy of 1950 kJ/kg. The stratigraphic units in well HE-6 are comprised of 4 distinct hyaloclastite formations, 3 lava units and 2 possible intrusions, one being of intermediate composition. There are four alteration zones: smectite-zeolite (376-730 m), smectite-mixed layer clay-chlorite (730-800 m), chlorite (800-1000 m) and chlorite-epidote (1000-1018 m). Fourteen aquifers were located along lithological contacts, intrusive boundaries, fractures and faults, indicating good, permeability zones. The distribution of hydrothermal alteration minerals, such as pyrite and calcite, also depict valuable permeable zones. Calcite forms a platy structure at boiling temperatures and the homogenization temperatures range from 240 to 275°C. This homogenization temperature of the platy calcite represents the highest alteration temperature range and conforms to the present formation temperatures in the well. However, the alteration temperatures are actually lower than the present formation temperatures in the well, indicating that the well may be heating up.

1. INTRODUCTION

Borehole geological logging is an important part of subsurface geothermal exploration which contributes to the understanding of the natural conditions in the geothermal system and its production characteristics prior to utilization (Steingrímsson and Gudmundsson, 2006). Before drilling exploration wells, the geothermal reservoir is broadly outlined from the initial geological, geophysical and geochemical exploration studies on the surface. Geothermal drilling has been carried out in Iceland for decades which makes it an ideal place to study borehole geological logging. Well HE-6, a production well in the Hellisheidi high temperature geothermal field, has been assigned to the writer for this purpose. The main resolution of the geological logging in boreholes is twofold: firstly to locate and characterize the permeability in the geothermal reservoir, and secondly to assess the condition of the system regarding alteration (Franzson, 2014).

1.1 Well HE-6

The Hellisheidi high temperature geothermal field is situated 45 km east of Reykjavik, Iceland (Figure 1). This geothermal field has been exploited and utilized by Reykjavik Energy for electricity and district heating and has a power plant presently producing 303 MW_e and 133 MW_{th} (Gunnlaugsson, 2012). Well HE-6 is situated at 384632E, 395257N (ISN 93 coordinate system) and 420 meters above sea level. It is a directional well striking N25°E with an inclination of 35° (Jónsson et al., 2002ab; Richter et al., 2002). The well was drilled to 2001 m in 2002 to explore the temperature and permeability conditions of the geothermal system. Well HE-6 is now discharging at a rate of 29.4 kg/s with an enthalpy of 1950 kJ/kg (Gunnlaugsson, 2014).
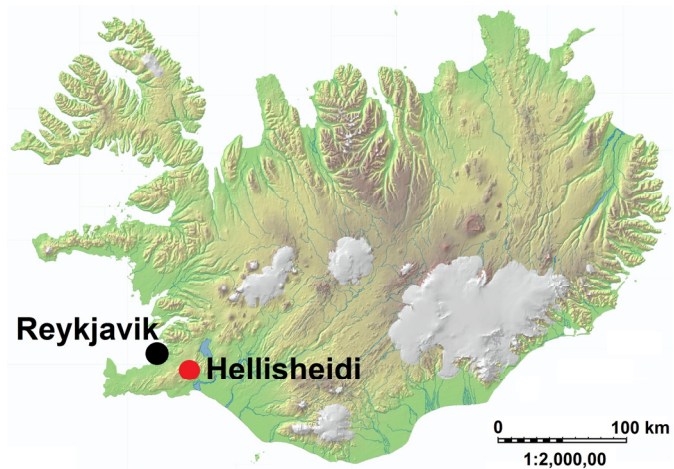
FIGURE 1: Location of Hellisheidi high temperature geothermal field in Iceland

2. GEOLOGICAL SETTING

2.1 Geology of Iceland

Iceland is located between 63°23´N to 66°30´N and 13°30´W to 24°30´W in the North Atlantic Ocean and south of the Great Arctic Circle. Formed entirely by volcanic activity and centred directly on the active diverging plates of North America in the west and Eurasia in the east, Iceland is spreading apart at a rate of 1 cm per year in both directions. This divergence occurs at the Mid Atlantic Ridge which runs northeast-southwest through the centre of Iceland, marked locally by the Reykjanes Ridge in the south and the Kolbeinsey Ridge in the north (Wolfe et al., 1997). The spreading is associated with a deep seated mantle plume directly under the Vatnajökull glacier, evident from the recent eruption of the Grímsvötn volcanic system on 4 September 2014. The geological map of Iceland (Saemundsson, 1979), first published over three decades ago, has been refined over the years and is shown in Figure 2.

Iceland has experienced intermittent glacial periods through time due to its northerly location. Today, glaciers and ice caps cover about 11% of the land area which is 11,400 km² out of the 103,125 km² area of Iceland. The surface rocks are made up of post glacial and interglacial lavas and sub-glacial hyaloclastites. The youngest rocks are located in the centre of Iceland, in the active rift zones, Quaternary formations are found in and along the margins of these rift zones and older Tertiary formations occur away from the rift zones to the east and west (Figure 2). Fissure swarms and central volcanoes mark these zones (Saemundsson, 1979).

The Northern Rift Zone (NRZ), Eastern Rift Zone (ERZ), Western Rift Zone (WRZ) and Reykjanes Rift Zone (RRZ) in Figure 2 are large, 40 to 50 km wide rift zones made up of smaller, 5 to 15 km wide (up to 200 km in length) volcanic fissure swarms. High temperature geothermal fields are confined to the active rift zones characterized by reservoir temperatures of more than 200°C at 1 km depth, whereas low-temperature areas are located outside the volcanic rift zone, having lower than 150°C reservoir temperatures at 1 km depth.

The Hengill volcanic system, which hosts the Hellisheidi high temperature geothermal field, is one of these volcanic systems and lies directly on a triple junction between the RRZ, WRZ and the South
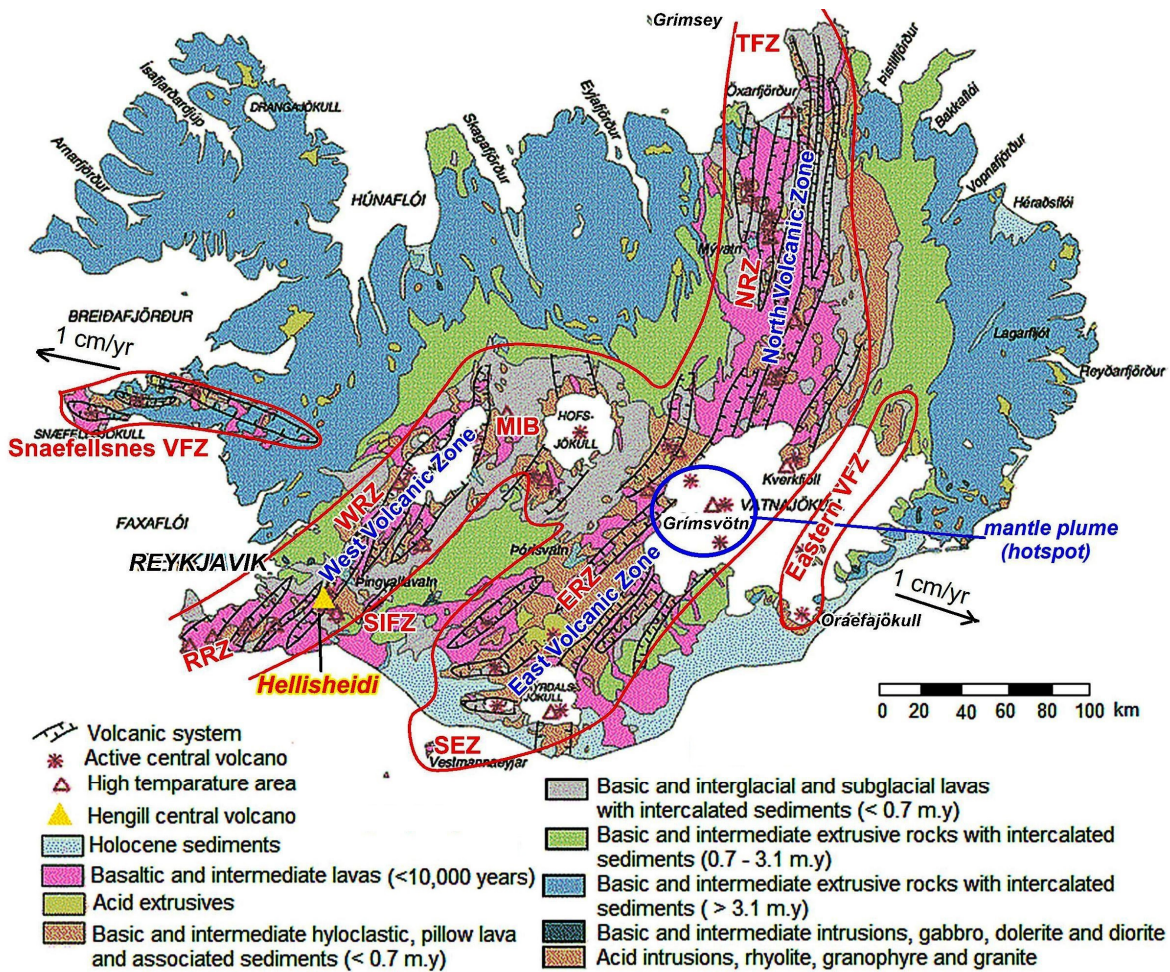
FIGURE 2: The regional geological map of Iceland (modified from Saemundsson, 1979)
showing the volcanic systems and fissure swarms, location of the mantle plume,
rift systems, direction of the spreading and the location of the Hellisheidi
high temperature geothermal field in the Hengill central volcano

Iceland Fracture Zone (SIFZ). The WRZ continues north-eastward from Hengill and connects to the NRZ through the Mid-Icelandic Belt (MIB), a transform zone (Einarsson, 1991). The NRZ continues north from the Vatnajökull glacier. The SIFZ in the south and the Tjörnes Fracture Zone (TFZ) in the north are transform zones which connect the volcanic rift zones to the segments of the Mid-Atlantic-Ridge. Due to rift relocations, the WRZ is apparently a failing rift zone while the ERZ will eventually become the main spreading zone in South Iceland (Saemundsson, 1979; Hardarson et al., 1997).

2.2 Hellisheidi high temperature geothermal field

The Hellisheidi high temperature geothermal field, centred by the Hengill central volcano is situated in the WRZ on a triple junction where two active rift zones meet a seismically active transform zone (Helgadóttir et al., 2010) (Figure 2). The geology of Hellisheidi has been mapped in detail by Saemundsson (1995) (Figure 3). The dominant rock formations in the Hellisheidi field are hyaloclastites which formed subglacially when magma cooled during an eruption into the base of the glacier, forming pillow basalts, breccias and tuffs. Occasionally these hyaloclastites are disturbed by lava successions which flowed to the lowlands during interglacial and interstadial periods. Hellisheidi is within the northeast-southwest fault zone and graben of the Hengill central volcano (Franzson et al., 2005).
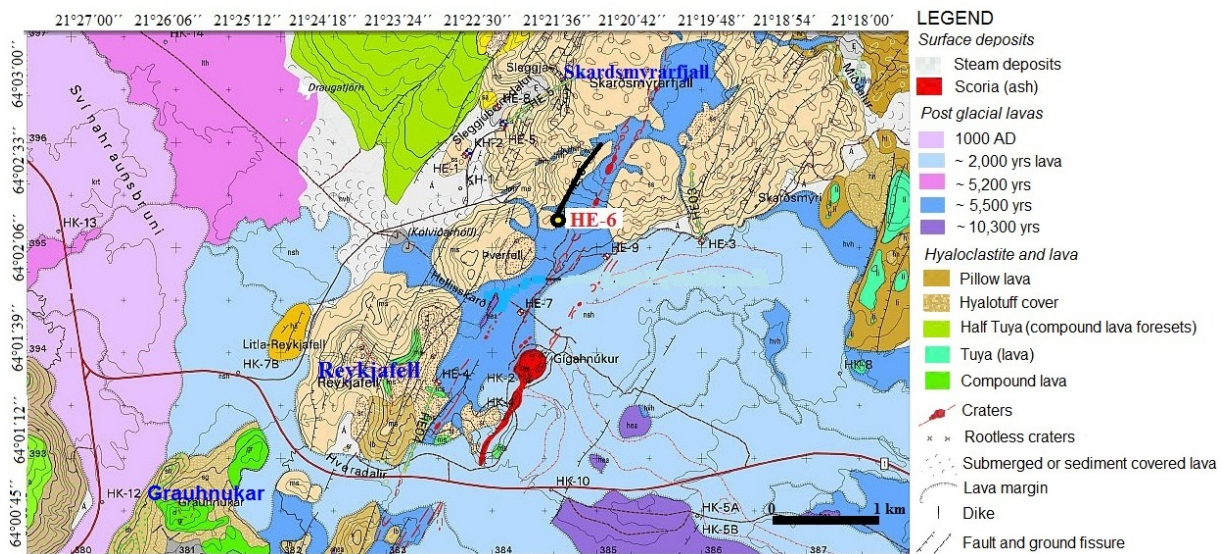
FIGURE 3: Geological map of the Hellisheidi high temperature field (modified from
Saemundsson, 1995) showing the location of well HE-6 drilled
north-eastward towards Mt Skardsmýrarfjall

The northeast-southwest faults and fractures which form part of the 5 to 15 km wide and 40 to 50 km long fissure swarm bisects the Hengill central volcano. Major up- and outflow zones in the field are largely related to volcanic fissures of 2000 and 5000 years of age (Saemundsson 1995; Björnsson 2004; Franzson et al., 2005). The fissure swarm has been the drilling target in the Hellisheidi field (Helgadóttir et al., 2010). Well HE-6 is located directly southwest of Mt Skardsmýrarfjall (Figure 3) and 1 km west of the 5000 year old row of fissures and was drilled directly northeast towards the core of the Hengill geothermal system (Jónsson et al., 2002a).

3. ANALYTICAL METHODS

3.1 Binocular and petrographic microscope analysis

An Olympus binocular microscope was used to analyse 509 cutting samples collected at 2 m intervals during drilling, representing a total of 1018 m of the subsurface rocks from well HE-6. Thirty thin sections were analysed using a Leica petrographic microscope. Logging procedures for binocular and petrographic analyses, from the Iceland GeoSurvey (ISOR), were used to establish the subsurface stratigraphy and alteration mineralogy. This was done based on the visual abundances of the primary and secondary minerals and their characteristics in each sample.

A preliminary binocular microscope analysis was carried out at the drill-site as the rock cuttings were collected during drilling. This is critical to assess the alteration temperatures of the geothermal reservoir because the drilling progress is heavily dependent on this preliminary assessment. Further detailed analyses were carried out at the ISOR labs in Reykjavik.

Petrographic microscope analysis is more precise in establishing the mineral assemblage of each stratigraphic unit and alteration mineralogy. Samples were selected based on the binocular microscope analysis and prepared for thin-section at an external laboratory. The thin-section preparation takes a fair amount of time and is, therefore, undertaken at a later stage as detailed understanding of the reservoir characteristics becomes important for future predictions of the geothermal system. The time relationships of the alteration mineral depositions are established during the petrographic study.

3.2 Use of the geophysical measurements and drilling data

During the drilling process a number of measurements are taken to estimate different reservoir parameters which are important to assess several different parameters and conditions of the well. Steingrímsson and Gudmundsson (2006) and Steingrímsson (2011) describe in detail the techniques involved in the acquisition of different types of well logging measurements in Iceland. These well logs make it possible to study the well, its geometry, the rock formation and fractures, the reservoir temperature and fluid pressures and feed points connecting the well to the geothermal reservoir. The geophysical measurements used during this study are: neutron logs, resistivity logs, gamma logs, caliper logs and the temperature logs which were acquired from the ISOR database and from the drill reports (Jónsson et al., 2002ab; Richter et al., 2002). Below is the description of what each of these logs was used for in this study:

i. *Temperature logs* were used to locate aquifers or feed points. The effects, of a hot or cold flow into or out of the well from the formation, cause a sudden increase or decrease in the temperature, providing important clues to the location of aquifers in the borehole.
ii. *Natural gamma logs* were used to refine lithological boundaries including alteration zones. As different rock types contain different amounts of radioactive isotopes (K, Th, and U), gamma logs can reveal if lithological units are of evolved chemical compositions. Gamma rays also generally show low values in the fresh rocks and a general increase in the altered zones. They can also be used to locate aquifers since alterations zones are often connected to aquifers.
iii. *Neutron logs* were also used to locate aquifers. Similar to the natural gamma log, the rise and fall of the measurements assist in locating a porous formation which generally gives a high spike. Water saturated basalts (aquifer hosted basalts) show moderate to high spikes, indicative of porous zones.
iv. *Resistivity logs* were used to refine the lithological boundaries (or thickness) and the alteration zones. Fresh rocks have higher resistivity compared to altered rocks due to the infiltration of water, lowering the resistivity.
v. *Caliper logs* were also used alongside the other geophysical measurements to locate and confirm aquifers in the well. The diameter of the well expands, indicated by a spike in the caliper reading when a possible permeable zone is encountered. Caliper logs thus reveal wash-out and caves in the well. These logs are also used to estimate the volume of cement needed for casing operations.

Some important information in the drilling reports (Jónsson et al., 2002ab; Richter et al., 2002) was used to assist in this study. This information included the circulation losses and the depths of the casings. The circulation losses reliably assisted with the identification of large, medium and small feed zones.

3.3 Fluid inclusion analysis

Fluid inclusion temperatures of platy calcite crystals sampled at 814 m, 840 m and 906 m were analysed using a microthermometric heating and cooling stage under a Laborlux 12 Leitz microscope. These crystals represent the altered and permeable zones of the geothermal reservoir and were picked under the binocular microscope and prepared using the standard procedure for microthermometry of fluid inclusion at ISOR (Gudfinnsson, 2011).

The fluid inclusion analysis is a reliable measure of the thermal history of the geothermal reservoir and is usually undertaken after the binocular microscope analysis. Fluid inclusion temperature measurements deduce the original temperatures of the fluids trapped within the crystal during its growth and recrystallization. The platy calcite crystals contain inclusions of liquid aqueous solutions and gas bubbles; the gas bubble forming process in each crystal is reversed by heating to determine the temperatures when they were trapped. The crystals are heated until the fluid homogenizes in a single phase (i.e. bubble disappears) and the temperature is measured. On cooling, the vapour bubble will reappear as it is trapped within the crystal.

3.4 X-ray diffractometer analysis

The X-ray diffractometer (XRD) analysis of the clay compositions was conducted on 40 selected rock cutting samples in a XRD machine; the model type being Bruker AXS D8 Focus. Each sample was selected based on binocular microscope analysis to aid the interpretation of the alteration zones within the geothermal reservoir. The clay minerals contain large percentages of water trapped between their crystal lattices. The clays are silicate sheets, part of a general but important group within the phyllosilicates. Most clay minerals are chemically and structurally analogous to other phyllosilicates but contain different amounts of water. The clays are important temperature dependent minerals and give information on the alteration temperatures of the formation in the geothermal reservoir. However, it is difficult to identify specific types of clay under both the binocular and petrographic microscopes. Therefore, the XRD analysis is very important because it permits a detailed determination of the clays.

A standard XRD procedure practised by ISOR was used for the preparation of the samples. It is important to note that two types of sample were prepared from each depth; one sample was treated with glycol ($C_2H_6O_2$) solution while the other was untreated but placed in a desiccator. Both these samples were stored for 24 hours prior to the analyses, however, thicker samples took over 48 hours in the desiccator. The analyses of these samples were made on the XRD machine. However, a third measurement was taken, after heating the glycolated sample on an asbestos plate to completely dehydrate the sample. A measurement is recorded for each sample as untreated, glycolated and heated.

4. RESULTS

4.1 Stratigraphy of well HE-6

The stratigraphy of well HE-6 correlates well with the preliminary stratigraphic observations (Jónsson et al., 2002ab; Richter et al., 2002**),** although a few minor differences were observed. The rocks are predominantly of basaltic hyaloclastite formations, some basaltic lavas and two intrusions, one being basaltic, the other chemically more evolved. The classification of each rock unit is based on the rock compositions and their crystallinity. The primary minerals of the basaltic rocks are glass, olivine, plagioclase, pyroxene and opaque minerals, mostly magnetite and ilmenite. Palagonite is the first stable alteration product of volcanic glass and has replaced some glass during the formation of the hyaloclastites sub-glacially. The classification of the different units (pillow lava, breccia, and tuff) within the hyaloclastite formation is based on the visual abundance of glass in the primary mineral assemblages. The hyaloclastite formation is distinguished from both the lavas and intrusions based on the abundance of glass. The intrusions are aphyric and often less altered than the host rock. The lavas can be either crystallized or aphyric.

Four hyaloclastite formations were intersected in well HE-6, three lava units and two minor possible intrusions (Figure 4). The hyaloclastite formations were encountered at: 0-88 m (Hyaloclastite I), 118-436 m (Hyaloclastite II), 440-454 m (Hyaloclastite III) and 474-1018 m (Hyaloclastite IV). Lava units, or crystallized basalts, were encountered at: 88-118 m (Lava I), 436-440 m (Lava II), and 454-474 m (Lava III). The two possible basalt intrusions were encountered at 690-692 and 742-748 m. Figure 4 shows a stratigraphic profile of well HE-6 including the alteration intensity and zones, aquifers, formation temperature and the geophysical measurements taken during the drilling.

Hyaloclastite I (0-88 m): This formation consists of two units, pillow basalts from 0-80 m and basaltic tuff from 80-88 m.

0-80 m - pillow basalt: This unit consist of approximately 10% glass, partially crystalline basalt with phenocrysts of olivine, pyroxene and plagioclase. The glass and basaltic fragments have many vesicles. Pillow basalts occur at the surface and down to 70 m. From the surface to 70 m, the pillow basalt unit is
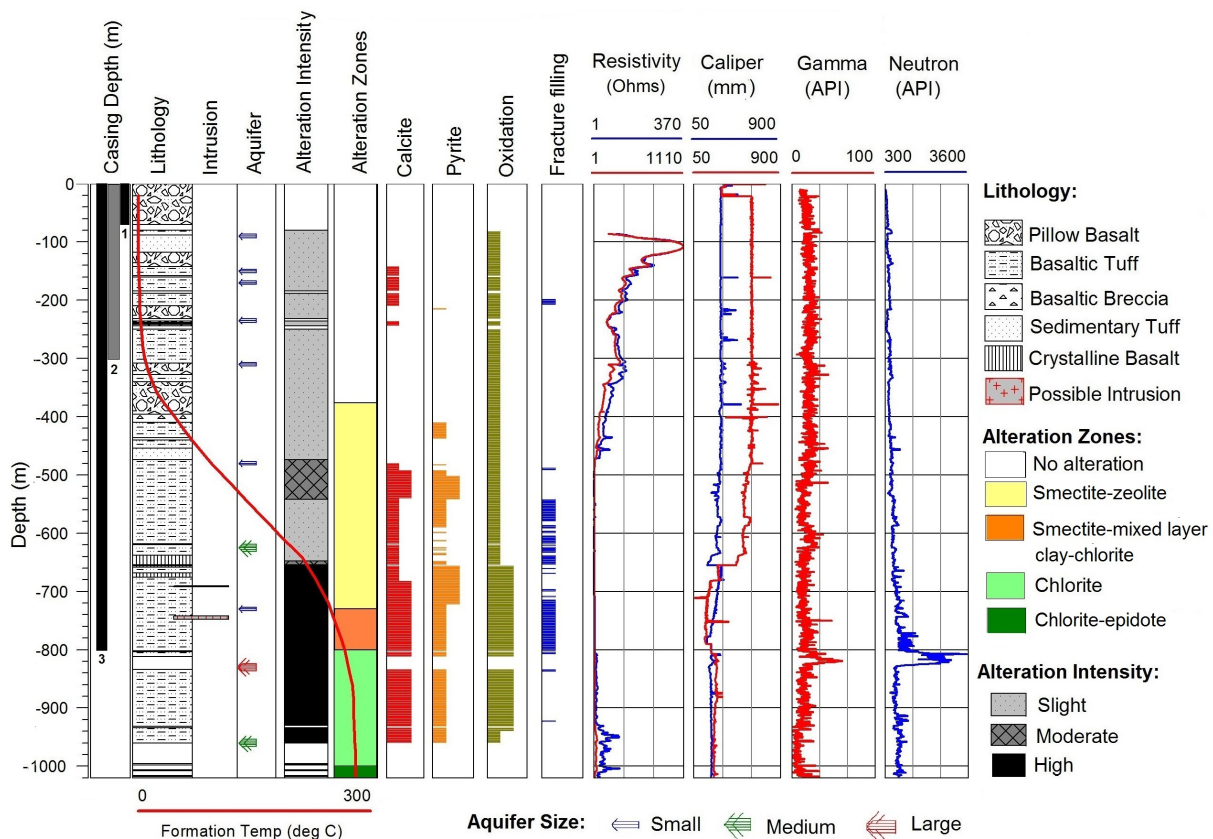
FIGURE 4: Stratigraphy, alteration, aquifers and geophysical logging data

Stratigraphy:

Hyloclastite I (0-88 m): this formation consists of two units, *pillow basalts* from 0-70 m and *basaltic tuff* from 80-88 m; olivine and glass are the first to show signs of oxidation as they are coated with limonite along fractures and as fillings in the vesicles.

Lava I (88-118 m): *Crystalline Basalt;* this crystalline basalt unit is the largest lava sequence in well HE-6 with a width of approximately 30 m. Limonite and siderite coats some of the crystal surfaces with their orange-brown dirty radiating crystals. Opaline silica deposited on some crystal surfaces.

Hyloclastite II (118-436 m): alternating *pillow basalts* and *basaltic tuff units* and a single unit of *basalt breccia*. The intensity of oxidation in the pillow basalt increases at 118 m. Calcite first appeared at 142 m filling vesicles; Pyrite occurs at 410 m and zeolites (scolecite, analzyme, thomsonite) occur at 412 m onward.

Lava II (436-440 m): *Crystalline Basalt*

Hyloclastite III (440-454 m): Mostly of basaltic tuff

Lava III (454-474 m): *Crystalline Basalt*

Hyloclastite IV (474-1018 m): The intensity of pyrite increases from 656 m in the sedimentary tuff and oxidation is intense giving the yellowish-red dirty appearance. Quartz, wairakite and phrenite start to occur within the basaltic tuffs at 804 m depth. The clays are reddish-yellowish-whitish-green coloured, mostly mixed layered clays from 676 m to approximately 800 m. Chlorite green colour is noticeable in the samples after 812 m. *Two possible **intrusions*** were encountered in the basaltic tuffs from 690-692 m and 742-748 m respectively.

Alteration Zones:

No alteration (0-376 m)
Smectite-zeolite (376-730 m)
Smectite-mixed layer clay-chlorite (730-800 m)
Chlorite (800-1000 m)
Chlorite-epidote (1000-1018 m)

relatively fresh with no fillings in the vesicles. However, the olivine and glass are the first to show signs of oxidation as they are coated with limonite along fractures and as fillings in the vesicles.

80-88 m - basaltic tuff: The unit consists of approximately 90% glass with the rest composed of basaltic rock with phenocrysts of olivine, pyroxene and plagioclase. Similar to the pillow basalt unit, vesicles are common. This unit is relatively fresh from 80 m for approximately 8 meters down to a section with few of the vesicles filled with opaline silica, limonite and siderite.

Lava I (88-118 m): This crystalline basalt unit is the largest lava unit in well HE-6 with a thickness of approximately 30 m.

88-118 m – Crystalline basalt: This unit is greyish to dark in colour, fine grained with consolidated microcrystalline phenocrysts of olivine, plagioclase, pyroxene and magnetite. The rock and crystals are relatively fresh with traces of creamy white opaline silica deposited on some crystal surfaces. Limonite and siderite coat some of the crystal surfaces with their orange-brown radiating crystals.

Hyaloclastite II (118-436 m): This formation is the single largest hyaloclastite formation in well HE-6 with a thickness of approximately 318 m, comprising several alternating pillow basalts and basaltic tuff units and a single unit of basalt breccia. There was a circulation loss at 158-160 m, 184-188 m, 232-236 m and at 244-250 m.

118-142 m - pillow basalt: This unit consists of approximately 10% glass, partially crystalline basalt and basaltic rock compositions with phenocrysts of olivine, pyroxene and plagioclase. The glass and basaltic phenocrysts have many vesicles. The intensity of oxidation in the pillow basalt increases at 118 m as more crystals are coated with both siderite and limonite. Opaline silica and chalcedony occur on the crystal surfaces or as fillings in some vesicles.

142-210 m - basaltic tuff: This unit is grey to black in colour and consists of approximately 90% glass with the rest composed of basaltic rock with phenocrysts of pyroxene and plagioclase. The glass alters to brown coarse grain palagonite. Calcite first appears at 142 m, filling vesicles of the basaltic tuffs, and the glass has been partially palagonized. Chalcedony and opaline silica are deposited on the crystals and in some of the vesicles.

210-244 m - pillow basalt: greyish-black coarse grained and highly porous crystals (vesicular) of partially crystallized basaltic glass, olivine, plagioclase and pyroxene. From 212-214, no glass was seen in the cuttings. Amorphous silica is deposited on the surfaces of the cuttings. Siderite and limonite fill fractures in the cuttings as well. Siderite occurs as small spherules with a radiating pattern and is yellowish to reddish brown in colour. There was a circulation loss from 244-250 m.

250-308 m - basaltic tuff: Brown to greyish-black, partially recrystallized glass with pyroxene and plagioclase phenocrysts. Amorphous silica occurs on the crystal surfaces with siderite and limonite filling fractures. Some glass altered to palagonite.

308-328 m - pillow basalt: Greyish-black coarse grained and highly porous crystals (vesicular) of partially crystallized basaltic glass, with olivine, plagioclase and pyroxene. Amorphous silica is deposited on the surfaces of the cuttings. Siderite and limonite fill fractures of the cuttings as well.

328-340 m - basaltic tuff: Brown to greyish-black, partially recrystallized glass with pyroxene and plagioclase phenocrysts. Some of the glass has been altered to palagonite and is porous with a significant amount of vesicles. Amorphous silica is deposited on the surfaces of the cuttings. Siderite and limonite fill fractures.

340-396 m - pillow basalt: Massive and consolidated, partially crystalline basalt with glass, olivine, plagioclase and pyroxene. Very porous but no vesicle fillings; glass is altered to palagonite. Amorphous silica and chalcedony are deposited on the crystal surfaces. Siderite and limonite fill fractures and occur on some crystal surfaces. Scolecite and analcime were identified in the thin-section analysis at 392 m. Zeolites occur together with calcite filling vesicles of mostly glass.

396-410 m - basaltic breccia: The basaltic breccia was intersected at 396 m, with a total thickness of 14 m. This unit consists of approximately 50% glass, with some crystalline to partially crystallized glass, and approximately 50% crystalline basaltic rock comprising phenocrysts of olivine, pyroxene and plagioclase. Some of the glass has been altered to palagonite. There are a number of vesicles, however, most of them have no fillings, while the few that have are filled predominantly with amorphous silica. While scolecite was noted in the overlying pillow basalt unit, the basaltic breccia appeared relatively

fresh and no zeolites were seen. Siderite and limonite occur on the crystal surfaces, giving a yellowish-red colour, while fine clays have formed along some of the vesicles in the glass and palagonite.

410-436 m - basaltic tuff: This unit is brown to black and consists of approximately 90% glass and occasional olivine, pyroxene and plagioclase phenocrysts. This unit is also highly vesicular and some of the pores are filled with amorphous silica. Clay, siderite and limonite occur along fractures. Most of the glass is palagonized and chalcedony appears, together with opaline silica, filling the vesicles. Pyrite occurs at 410 m along fractures and on some crystal surfaces while the zeolites occur at 412 m. Thomsonite, a low temperature zeolite, was identified in the binocular analysis. Fine grained clays and calcite occur along the fractured surfaces of the glass.

Lava II (436-440 m)
436-440 m - crystalline basalt: This unit is similar to the previous unit (Lava I) and is approximately 4 m thick. It is grey to dark in colour, fine grained with consolidated microcrystalline phenocrysts of olivine, plagioclase, pyroxene and magnetite. The rock and crystals are relatively fresh, and a little creamy white opaline silica is deposited on some crystal surfaces. Limonite and siderite coat some of the crystal surfaces.

Hyaloclastite III (440-454 m)
440-454 m - basaltic tuff: This unit consists mostly of glass (90%) and some phenocrysts of olivine, pyroxene and plagioclase. Most of the glass is partially palagonized and is highly vesicular. Some of the pores are filled with opaline silica and chalcedony. Clays, siderite and limonite occur along fractures and as fillings in the vesicles. Pyrite is also seen along the crystal fractures. Zeolite crystals are present in the cuttings but were impossible to distinguish through the binocular microscope.

Lava III (454-474 m)
454-474 m - crystalline basalt: This unit is similar to the previous two crystalline basalt units and is approximately 20 m thick. It is grey to dark in colour, fine grained with consolidated microcrystalline phenocrysts of olivine, plagioclase, pyroxene and magnetite. There was some recrystallized glass mixed within the cuttings. The rock is relatively fresh, with a little creamy white opaline silica deposited on crystal surfaces. Limonite and siderite coat some of the crystal surfaces.

Hyaloclastite IV (474-1018 m): There are no cuttings from the following intervals: 618-620, 654-656, 802-804, 812-834, 932-934, 960-996, 998-1006 and 1008-1016.

474-638 m - basaltic tuff: This unit consists predominantly of altered glass and partially crystallized glass, mixed with a few plagioclase phenocrysts. The vesicles have been filled by opaline silica, clays and zeolites. The zeolites were difficult to distinguish under the binocular microscope. Calcite and zeolites fill vesicle spaces and calcite also occurs as vein infills. Yellow and white clays occur along fractures and vesicle fillings. Limonite and siderite coat some of the crystal surfaces and are quite intense in this section.

638-658 m - sedimentary tuff: This unit is comprised of cuttings grey to black in colour, fine grained, laminated mudstones and sandstones mixed with altered glass, palagonites and phenocrysts of plagioclase. The unit has a significant amount of calcite and zeolites filling the vesicles, and clays, together with the calcite, fill veins and fractures. The intensity of clay alteration is high in this unit. Olivine and glass have been completely altered to smectite and mixed layer clays. Pyrite crystals occur in the sedimentary tuff units at 648 m. The intensity of pyrite increases from 656 m and oxidation is intense, resulting in a yellowish-red dirty appearance.

658-668 m - basaltic tuff: This unit also consists predominantly of clay altered glass with partially crystallized glass mixed in with a few plagioclase phenocrysts. Olivine and glass have been completely altered to smectite and mixed layered clays. The vesicles have been filled by opaline silica, clays and zeolites. The zeolites are difficult to distinguish under the binocular microscope. Calcite and zeolites fill

vesicle spaces and calcite also occurs as vein fillings. Yellow and white clays occur along fractures and as vesicle fillings. Limonite and siderite coat some of the crystal surfaces and are quite intense in this section. Sulphide (mostly pyrite) is disseminated throughout.

668-676 m - sedimentary tuff: This unit is similar to the unit encountered at 638 m, grey to black in colour, fine grained, laminated mudstones and sandstones mixed with clay altered glass, palagonites and phenocrysts of plagioclase. Calcite and zeolites fill the vesicles, and the clays, together with calcite, fill veins and fractures. The intensity of the clay alteration is also high in this unit. Olivine and glass have been completely altered to smectite and mixed layered clay. Pyrite crystals occur in the sedimentary tuff units at 648 m. Sulphide (mostly pyrite) is disseminated throughout.

676-1018 m - basaltic tuff: This is the thickest single unit intersected in the hyaloclastite formation; it occurs for approximately 442 m, assuming all the units that were lost due to circulation losses were part of this same formation. Basaltic tuff consists of approximately 90% glass with the rest being composed of basaltic rock with phenocrysts of olivine, pyroxene and plagioclase. Similar to the other units, there are numerous vesicles in both the glass and basalt units which were encountered at 676-802, 804-812, 834-932, 934-960, 996-998, 1006-1008 and 1016-1018 m. The intensity of clay alteration is much higher at 658 m in the basaltic tuffs with some glass and olivine completely altered to clay minerals. Quartz, wairakite and prehnite start to occur within the basaltic tuffs at 804 m. The clays are reddish-yellowish-whitish-green coloured, mostly mixed layer clay from 676 to approximately 800 m. Chlorite texture and green colour is noticeable in the samples after 800 m.

4.1.1 Intrusions

Two possible intrusions were encountered in the basaltic tuff unit (Hyaloclastite IV) at 690-692 m and 742-748 m. Both intrusions are composed of microcrystalline olivine, plagioclase, pyroxenes and magnetite phenocrysts. The intensity of clay alteration in the basaltic tuff is quite intense as most of the glass and olivine have been partially to completely altered. Pyrite and calcite occur as fracture fillings, while some calcite and zeolites occur as fillings in the vesicles. A spike occurs in the gamma log at 742 m (Figure 4) which may indicate that the intrusion is of intermediate chemical composition. There is no spike at 690 m indicating that this intrusion is of basaltic composition.

4.2 Alteration mineralogy

4.2.1 Alteration of the primary mineral assemblages

The primary minerals of the basaltic hyaloclastite formations, basaltic lavas (crystalline unit) and intrusions in well HE-6 are predominantly glass, olivine, plagioclase, pyroxene and opaque minerals, mostly magnetite and ilmenite. Table 1 shows the primary minerals in well HE-6 and some of their most common alteration products.

TABLE 1: Primary minerals of the rocks studied in well HE-6 and the common alteration products

Sensitivity to alteration	Primary rock minerals	Alteration mineral products; *Not identified in this study*
Most	Glass	clay, calcite
	Olivine	clay, calcite, smectite, chlorite
	Plagioclase	clay, calcite, albite, wairakite
	Pyroxene	clay, calcite, chlorite, smectite, epidote (including actinolite*)
Least	Opaque (mostly magnetite and ilmenite)	limonite, siderite, clay, pyrite, (including sphene*)

The primary minerals are relatively fresh from the surface to approximately 114 m. Figure 5A shows fresh basaltic phenocrysts of olivine, plagioclase, pyroxenes and glass in the basaltic tuff unit at 34 m. As depth in the well increases, temperature and fluid chemistry changes and the primary minerals become unstable under the new physiochemical conditions. The primary minerals alter to secondary mineral products by their interaction with hot water (and steam which helps the dissolution of the primary minerals, replacement and the precipitation of new minerals).
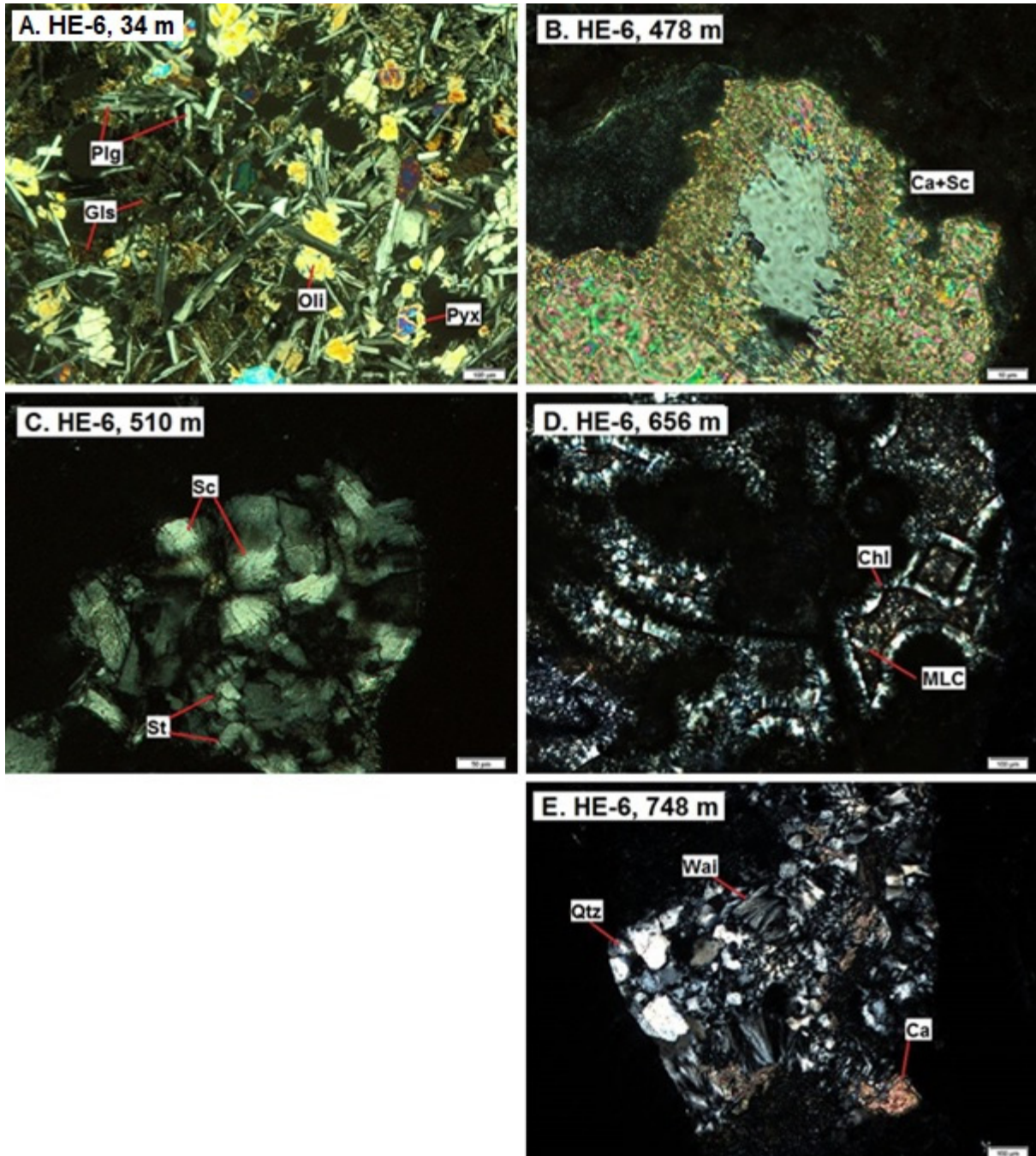
FIGURE 5: Photomicrographs of some of the samples studied, taken under transmitted light (crossed polars): A) Glass (Gls), olivine (Oli) and plagioclase (Plg) crystals in the basaltic tuff unit; B) Calcite (Ca) and scolecite (Sc) filling vesicle in the sedimentary tuff unit; C) Sc and stilbite (St) occur as fillings in the vesicles of the glass; D) Chlorite (Chl) replacing mixed layer clay (MLC); E) Quartz (Qtz) and wairakite (Wai) are formed and calcite (Ca) replaces the plagioclase

Glass is the first to alter and is replaced by calcite at 142 m, mostly along the fractures. *Olivine* is also early to alter to clays and calcite along fractures; this was seen at 158 m in the petrographic study. *Plagioclase* alters to clays and calcite at 394 m, noticeable from the increase in clay alteration intensity. At deeper levels (greater than 750 m depth), the plagioclase alters to albite which was seen in the petrographic study. Albite alteration of plagioclase starts at temperatures above 220°C and is known as albitization. *Pyroxene* alters to clays and calcite along fractures at relatively shallow levels in the geothermal system (Franzson, 2014). At deeper levels, chlorite replaces the pyroxenes more easily; this was seen at 808 m and, at greater depths, the pyroxene alters to epidote and actinolite. Epidote and actinolite are high temperature minerals which occur in high temperature geothermal systems in Iceland at temperatures ranging between 250 to greater than 300°C (Franzson, 2014). Opaque minerals usually alter within oxidation zones, characterized by the presence of limonite, pyrite and sometimes sphene**.**

4.2.2 Hydrothermal alteration

The hydrothermal minerals identified in well HE-6 are shown in Table 2 and are presented in their mineral groups and order of occurrence. Figure 6 shows the depth range at which the alteration minerals were formed and the alteration zones based on clay analyses.

TABLE 2: Alteration minerals identified in well HE-6, structural formula, formation temperature range and depth of first occurrence

Mineral groups/ minerals	Chemical formula (Gribble and Hall, 1992) *not specified*	Formation temperatures (Saemundsson and Gunnlaugsson, 2014) *estimate numbers*	Depth of first occurrence
Oxide:			
Limonite	$FeOOH.nH_2O$	from low temp $< 40°C$ *	2 m
Carbonates:			
Siderite	$FeCO_3$	from low temp $< 40°C$ *	80 m
Calcite	$CaCO_3$	$< 40°C - 280°C$ *	142 m
Sulphides:			
Pyrite	FeS_2	$< 40°C - > 200°C$ *	216 m
Zeolites:			
Scolecite	$Ca[Al_2Si_3O_{10}].3H_2O$	$80 - 120°C$	392 m
Thomsonite	$NaCa_2(Al,Si)_5$	$50 - 120°C$	410 m
Analcime	$NaAlSi_2O_6.H2O$	$50 - 170°C$	392 m
Stilbite	$Na_2CaK_2[Al2Si_7O_{12}].4H_2O$	$80 - < 150°C$	510 m
Laumontite	$Ca[Al_2Si_4O12].4H_2O$	$120 - 180°C$	656 m
Wairakite	$CaAlSi_2O_6.H_2O$	$> 200°C$ *	748 m
Clays:			
Smectite	$(1/2Ca,Na)(Al,Mg,Fe)_4(Si,Al)_8O_{20}(OH)_4.nH_2O$	$50 - 170°C$	378 m
Mixed layer clay	(smectite and chlorite complex)*	$200 - 230°C$	748 m
Chlorite	$(Mg,Al,Fe)_{12}(Si,Al)_8O_{20}(OH)_{16}$	$230 - 290°C$	808 m
Others minerals:			
Opaline silica	SiO_2	$< 100°C$	80 m
Chalcedony	SiO_2	$70 - 200°C$	118 m
Quartz	SiO_2	$180 - > 300°C$	748 m
Prehnite	$CaAl[AlSi_3O_{10}](OH)_2$	$240 - 250°C$	912 m
Epidote	$Ca_2Fe^{3+} Al_2[Si_3O_{12}](OH)$	$250 - > 300°C$	1000 m

Oxides and carbonate:
Limonite forms by the oxidation of ferrous minerals, such as magnetite, which is found in all the basalts and occurs immediately at shallow levels to depths of more than 600 m in the well. It is dark or yellowish brown, sometimes reddish and opaque with a slight metallic lustre and forms thin layers with a globular surface coating.

Siderite was identified at 80 m and occurs together with limonite down to more than 600 m. Similar to limonite, the crystals are tabular, however, have spherules and show a radiating pattern. Siderite usually pseudomorphs after limonite and is yellowish brown or reddish brown in appearance.

Calcite is the most common carbonate precipitate in well HE-6 and forms at all temperature ranges, however, it is not stable at temperatures above 300°C (Table 2). Calcite is generally white but may be coloured by impurities and was first identified at 142 m. It has a vitreous lustre, conchoidal fractures and occurs in various forms from granular to orthorhombic. Platy calcite was recognized at depths greater than 800 m in the well.

Sulphides:
Pyrite is the most common sulphide mineral in well HE-6 and was first identified at 216 m. It forms yellowish cubic crystals which have a strong metallic lustre. Pyrite is abundant at 510 m to 540 m and from 656 m to the bottom of the stratigraphic profile (i.e. 1018 m) which may indicate good permeable zones.

Zeolites:
Scolecite was identified at 392 m together with calcite, analcime and clay, filling vesicles of the pillow basalts. Scolecite is a colourless or white, fibrous zeolite with a vitreous, or slightly silky lustre, and forms between 80°C and 120°C (Table 2). It is four-sided, densely packed, and forms aggregates where the crystal fibres radiate from a single point, growing thicker, and may separate towards an obtuse point at the end. The fractures are uneven and thick, having faint longitudinal marks. Figure 5B shows the typical fibrous structure of scolecite surrounded by calcite, filling the vesicles in the pillow basalt unit at 478 m.

Thomsonite is a radiating zeolite, identified at 410 m, and forms at temperatures ranging between 70°C and 110°C. The crystals are prismatic, acicular, elongated, bladed, slightly flattened, with a slanted end. They form dense masses of radiating clusters, with mammillary structures whose surface are finely bristled. Its colour is mainly white and is translucent with a vitreous or pearly lustre and has uneven fractures.

Analcime is another low temperature zeolite identified at 392 m and forms between 50 and 170°C. It is also known as analcite and forms colourless or white, many sided (trapezohedral) crystals with vitreous lustres and conchoidal fractures.

Stilbite is a low temperature zeolite mineral identified at 510 m. It forms between 80°C to less than 150°C. It is commonly observed as thick tabular crystals with pointed terminations. The crystals often grow wider towards the end, forming shelf-like aggregates which develop ends like a "bow-tie".

Stilbite is generally milky white, clear and translucent. In the petrographic study, stilbite occurred mostly with scolecite. Figure 5C shows scolecite and stilbite filling vesicles in the sedimentary tuff formation at 510 m.

Laumontite was identified at 656 m and forms at temperatures ranging from 120 to 230°C. It is a radiating or tabular zeolite and creates thin elongated fibres, or prisms, with longitudinal marks and a square end. A cross-section of the crystal is square and it is normally white, but may be pale pink or reddish brown.

Wairakite is a zeolite mineral with an analcime structure but contains a calcium ion (Gribble and Hall, 1992). It is the calcium rich variant related to analcime that forms between 200 and 300°C, i.e. at higher temperatures than the other zeolites. It was first identified at 748 m together with quartz.

Clays:
Smectite is the most common clay mineral group in well HE-6 and occurs at 378 m to depths of more than 900 m. Smectite is a low temperature clay formed from the alteration of glass and the ferromagnesian primary minerals and occurs at temperatures ranging from 50 to 170°C. Smectite precipitates in pores and occurs as linings in vesicles at shallower depths but occurs with mixed layer clay at deeper levels.

Mixed layer clay was identified at 748 m in well HE-6 and forms at temperature ranges between 200 and 230°C. The mixed layer clay is a product of reactions involving clay minerals end member in which the different kinds of clay layers alternate with each other. In well HE-6, the mixed layer clay is a complex of smectite and chlorite clay minerals. It appears coarse grained around the fine grained smectite clay and is identified by the brightly coloured radiating crystals under transmitted light (crossed polars) in the petrographic study.

Chlorite is generally green, radiating, flaky or fibrous aggregates and was first identified at 808 m. It is a common secondary mineral formed from the alteration of mafic primary minerals or smectite, filling the vesicles and occurs at temperatures between 230 and 290°C. It replaces the mixed layer clay completely at higher temperatures. Chlorite is sometimes colourless in thin-section with no pleochroism and has a low order maximum interference colours (rarely above grey). Figure 5D shows chlorite replacing mixed layer clay at 656 m in the well.

Silica and other minerals:
Opaline silica forms at temperatures of less than 100°C and was first identified in the well at 80 m. It is almost amorphous, and is softer and less dense than all other silica minerals. The most common variety is pale or milky white, opaque (sometimes translucent) with a vitreous or greasy lustre.

Chalcedony is another silica mineral that occurs at much higher temperatures than opaline silica, at 70°C to 200°C. It is often regarded as amorphous and forms tiny thread-like crystals identified by their greyish blue colour and slight vitreous or greasy lustre. Chalcedony was first identified at 118 m by the binocular analysis.

Quartz appears at temperatures above 180°C and was first identified at 748 m together with wairakite. It is white-milky white or grey and is translucent or opaque and has a vitreous lustre and uneven or conchoidal fractures. Quartz forms hexagonal prismatic crystals which end in slanted crystal faces meeting at a point, i.e. a six sided pyramid. Figure 5E shows the occurrence of quartz, wairakite and calcite in thin-section analysis. The quartz and wairakite crystals have low order interference colours and are distinguished by their various optical properties.

Prehnite was first identified at 912 m and forms at temperatures ranging between 240 and 250°C. It forms small spherical clusters of crystals with vitreous lustre, sometimes pale green, white or grey in colour. The crystal clusters are radiating, and this is visible when broken up. The fracture is irregular and the streak is white.

Epidote was not seen in this study but was identified at 1000 m in the preliminary stratigraphic study (Jónsson et al., 2002b). It is a product of hydrothermal alteration of the feldspars and pyroxenes and forms at 250°C to temperatures greater than 300°C. Epidote is prismatic and fibrous with striations and is usually yellow-green or greenish black in colour.

4.2.3 Alteration zones

The first and last appearance of the hydrothermal minerals studied under both the binocular and petrographic microscopes were combined with the clay analyses from the XRD to create a time relationship of the minerals relating to the formation temperatures in well HE-6. Figure 6 shows the

sequence of mineral deposition in the well. Five alteration mineral zones have been identified in well HE-6:

i. Unaltered zone (0-376 m): This zone, although relatively unaltered, consists of oxides (limonite), carbonates (siderite and calcite) and silica minerals (opaline silica and chalcedony). The presence of these minerals suggests a carbonaceous cold groundwater system warming up.

ii. Smectite and zeolite zone (376 – 730 m): This zone is well defined and is quite extensive in comparison to the other zones. It comprises, all the low temperature zeolites (scolecite, analcime, thomsonite, stilbite) and smectite clays. Calcite, chalcedony and opaline silica occur, and are only accessory minerals that do not define the alteration mineral zonation boundaries. The temperature range for the occurrence of this mineral assemblage is a little under 40 to just over 170°C.

iii. Smectite, mixed layer clay and chlorite zone (730 – 800 m): This zone is a transitional zone, predominantly consisting of smectite with some occurrence of mixed layer clay and chlorite. The temperature range for the occurrence of these minerals ranges from a little under 180 to 290°C.

iv. Chlorite zone (800 – 1000 m): This zone is also well defined, and is the second largest zone studied. The temperatures defining the zone range from a little under 230 to 290°C.

iv. Chlorite and epidote (1000 – 1018 m): This zone is predominantly of chlorite and epidote with temperatures ranging from 250 to over 300°C. Although epidote was not identified in this study, it was referred to in an ISOR report (Jónsson et al., 2002b). This zone is quite common in the high temperature geothermal systems in Iceland (Franzson, 2014). Epidote has been found in nearby wells, in well HE-3 at 792 m (Getaneh, 2001) and in well HE-9 at 660 m (Eshaghpour, 2003).
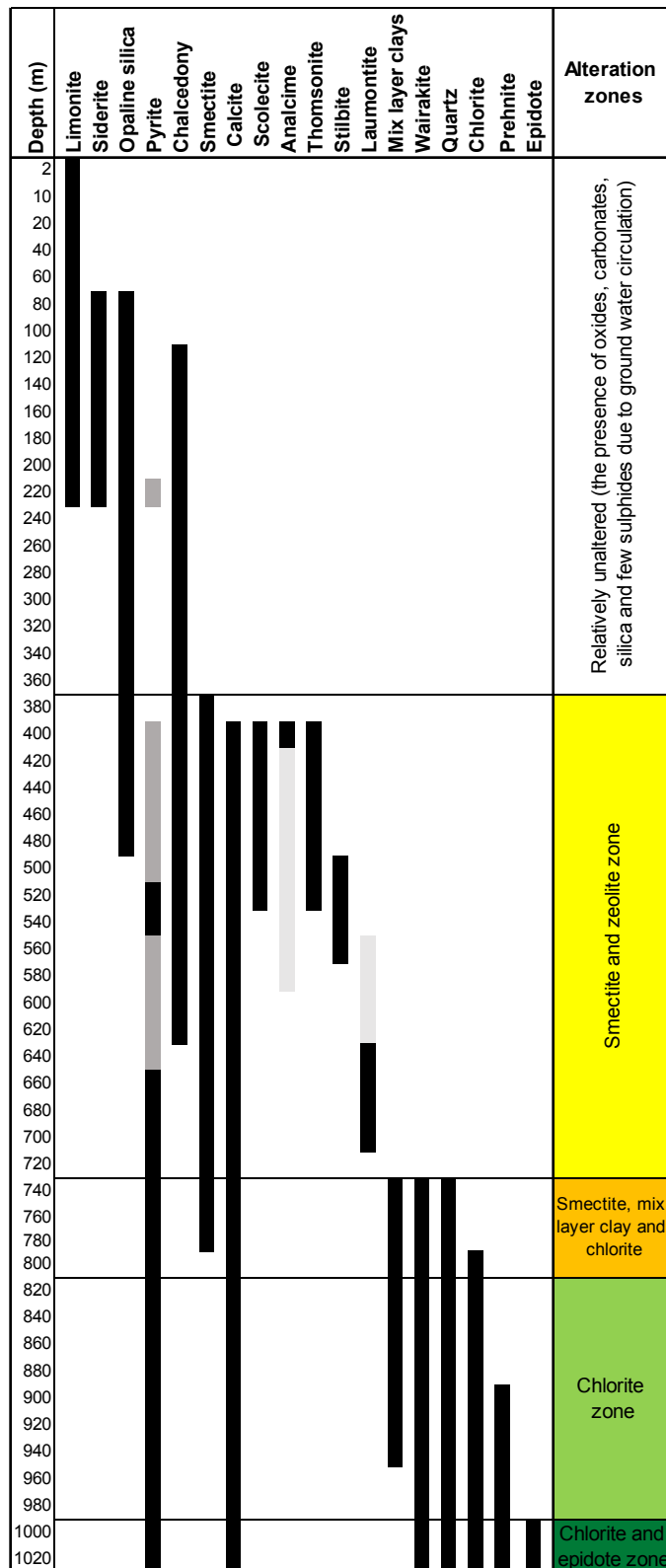
FIGURE 6: Sequence of mineral deposition in well HE-6: The black shades indicate strong occurrence, the dark grey shade indicate weak occurrence, while the light grey shade indicates a probable occurrence

4.3 Clay mineralogy from the XRD analysis

The clay minerals are the predominant hydrothermal alteration products in this borehole. Table 3 shows the summary of the results from the XRD analyses. Figure 7 shows examples of some of the interpretations from the XRD clay measurements. The following clay minerals were detected.

TABLE 3: XRD results for the clay analyses

No.	Depth (m)	Untreated (Å)	Glycolated (Å)	Heated (Å)	Clay type
1	162	nd	Nd	nd	nd
2	262	nd	Nd	nd	nd
3	318	nd	Nd	nd	nd
4	336	nd	Nd	nd	nd
5	376	13	17	nd	Smectite
6	404	16.9	17	10.01	Smectite
7	450	15	17	10	Smectite
8	472	15	17	nd	Smectite
9	480	14	17	10	Smectite
10	500	nd	Nd	nd	nd
11	516	13	17	13	Smectite
12	552	14	17	16-Oct	Smectite
13	590	13	17	16-Oct	Smectite
14	600	14.16	14.31	10.01	Smectite
15	630	12.9	16.9	9.8	Smectite
16	650	13.59	13.9	10.01	Smectite
17	660	12.9	16.9	12/9.8	Smectite
18	690	12.9	16.9	12/9.8	Smectite
19	700	13.59	13.9	10.01	Smectite
20	720	14	17	13	Smectite
21	730	13.9	10.01	7.21	Chlorite+Smectite
22	740	14.47/13.07	10.01	7.21	Chlorite+Smectite
23	752	30/14	31/17	14	Smectite+MLC
24	782	31/14	31/17	14	Smectite+MLC
25	788	17.67/15.05	10.01	7.27	Chlorite+Smectite
26	794	14.29/12.97	10.01	7.18	Chlorite+Smectite
27	800	14.59/13.04	10.01	7.18	Chlorite+Smectite
28	808	31/14	31/17	14	Chlorite+MLC
29	810	14.59/13.04	10.01	7.18	Chlorite+MLC
30	846	14.5	14.5	14.5	Chlorite
31	850	15.12	15.12	7.33	Chlorite
32	866	14.5	14.5	14.5	Chlorite
33	880	14.97	14.97	7.28	Chlorite
34	892	31/14	32/15.5	12	Chlorite+MLC
35	900	14.97	14.97	7.28	Chlorite
36	910	14.47	14.47	7.23	Chlorite
37	934	31/14	32/15.5	12	Chlorite+MLC
38	940	14.47	14.47	7.23	Chlorite
39	954	14.3	14.3	14.5	Chlorite unstable
40	960	14.47	10.2	7.23	Chlorite

Smectite, the most common low temperature clay mineral in Iceland geothermal systems (Kristmannsdóttir, 1985) occurs from 376 to 800 m in well HE-6. However, smectite exists as a single clay from only 376 to 730 m, detected from 15 samples. Smectite occurred predominantly from 730 to 800 m with mixed layer clay and chlorite. Smectite and chlorite coexisted in two samples representing
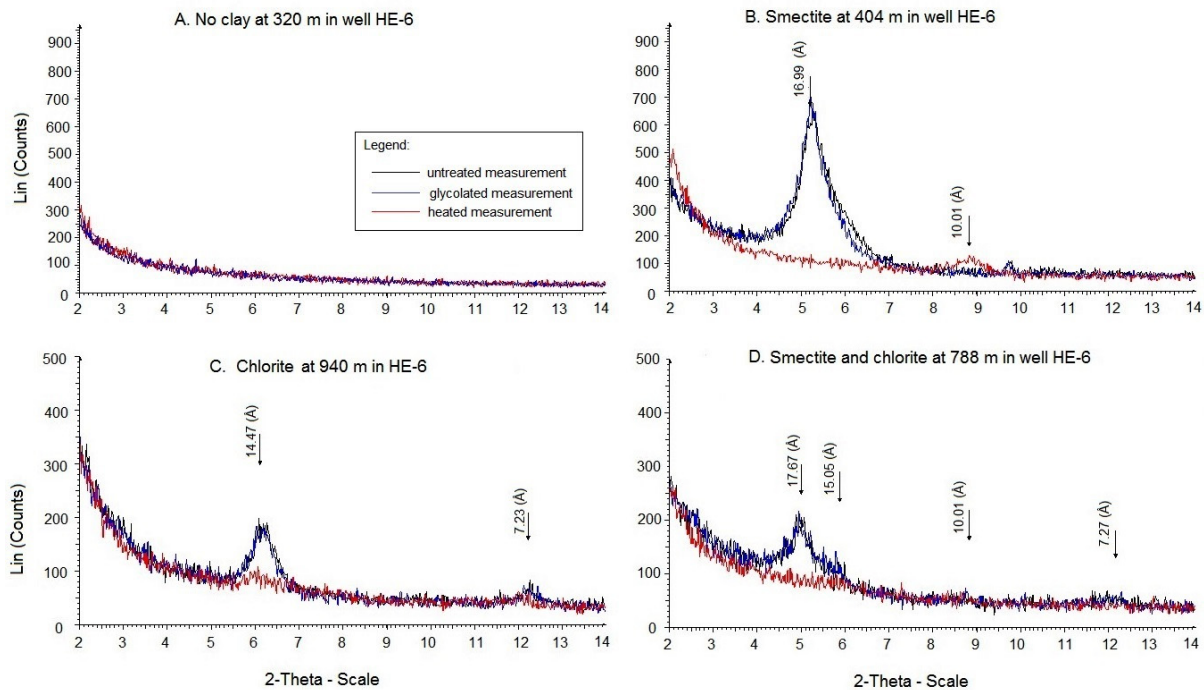
FIGURE 7: The XRD measurements for the clay analyses in well HE-6: A) No clay was detected at 320 m; B) Smectite clay was detected at 404 m; C) Smectite and chlorite were detected at 788 m; D) Chlorite was detected at 940 m

730 to 752 m. This was followed by the coexistence of, again, smectite and mixed layer clay in two samples representing 752 to 788 m. From 788 to 800 m, the smectite and chlorite coexist again, identified in 3 samples. Chlorite occurred predominantly from 800 to 960 m in 13 samples, although a few samples between these depths contained chlorite and mixed layer clay.

The X-ray diffraction patterns of the clays may vary depending on the humidity, exposure to certain organic molecules, heat treatment and exchangeable cations (Wilson, 1987). The X-ray diffraction patterns which show no peaks in all the measurements are typical for samples lacking clays (Figure 7A). The smectite clay usually has an untreated basal reflection between 13 and 15 Ångstroms (Å). When saturated with ethylene glycol, the reflection of smectite will swell to about 17 Å and, when heated to 400°C, the reflection will collapse to about 10 Å. Figure 7B shows the typical X-ray diffraction pattern for smectite. Chlorites have their peak at approximately 14 to 14.5 Å, depending on the individual species (Figure 7C). Peak positions are unchanged by ion saturation, treatment with ethylene glycol, or heating.

Clay minerals are composed of only two types of structural units (octahedral and tetrahedral sheets) and, therefore, different types of clay minerals can coexist, giving rise to mixed layer clays. The mixed layer clays sometimes result in much higher Ångstrom values (see examples in Table 3; mixed layer clay and smectite or mixed layer clay and chlorite). The peaks appear quite distinct in samples containing two or more clays and are identifiable by their individual diffraction patterns as shown in Figure 7D, a sample containing smectite and chlorite clays.

4.4 Calcite microthermometry from fluid inclusion

Platy calcite, representing the hydrothermal alteration at 814 m, 840 and 906 m, was analysed to deduce the homogenization temperatures in the well at the time of formation. Figure 8 is a photomicrograph of the fluid inclusion containing gas bubbles analysed at 840 m. The summary of the depth, number of

FIGURE 8: Secondary fluid inclusion from fracture fillings of platy calcite in the mixed layer clay-chlorite zone at 840 m

crystals and analyses of each crystal are shown in Table 4 and the statistical analyses presented in Figure 9.

The homogenization temperature range for platy calcite at 814 m (Platy calcite 1 in Figure 9) is from 240°C to 270°C. The homogenization temperature range for platy calcite at 840 m (Platy calcite 2) gave a narrow range from 265°C to 270°C. Because these two samples occur not far apart in the well, the statistical analyses are shown together in Figure 9A. The platy calcite from 906 m has a homogenization temperature ranging from 260°C to 275°C, shown in Figure 9B.
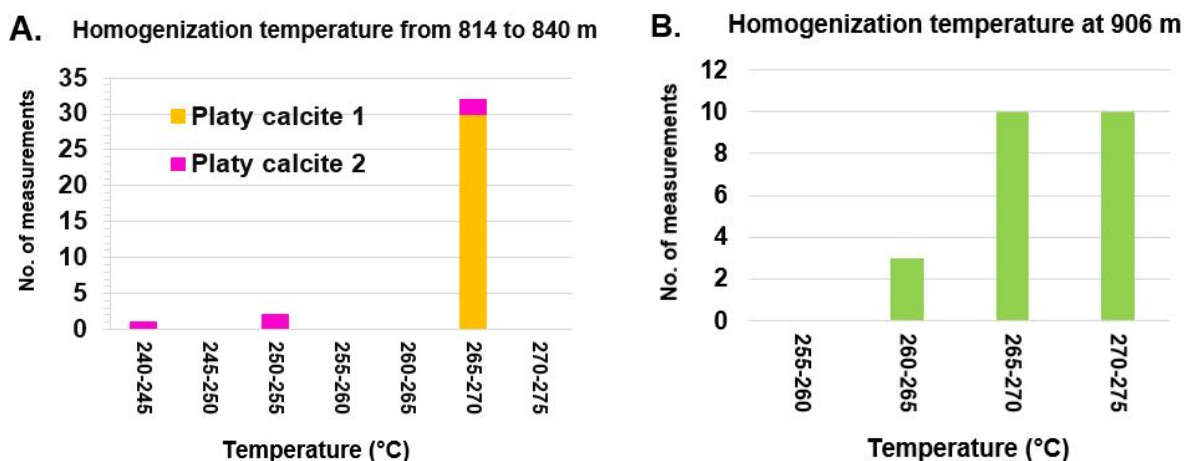
FIGURE 9: The homogenization temperatures in the geothermal system

TABLE 4: Fluid inclusion measurements from well HE-6

Depth (m)	Crystals (No.)	No. Measurements
814	1	5
840	1	30
906	1	23
Total:	3	58

The homogenization temperature in an alteration mineral represents a restricted past temperature in the geothermal system (Franzson et al., 2014). The platy calcite formations are secondary inclusions which occurred mostly in fracture fills within the chlorite zone of this geothermal system.

A comparison can be made between the alteration mineral temperatures, the formation temperatures and the homogenization temperature of the platy calcite to deduce the past and present temperature conditions of the geothermal system. Figure 10 shows temperature versus depth with the interpreted temperatures of the alteration minerals, homogenization temperature of platy calcite and the formation temperature taken from the drilling data. The boiling point curve is added for comparison as that depicts the maximum temperature attainable in a water dominated system (Franzson et al., 2014).

An approximate equilibrium exists between the alteration temperatures and the formation temperatures at shallow levels of the well, less than 500 m depth (Figure 10). This is seen by the appearance of analcime, scolecite and thomsonite, the low temperature zeolites. As the depth in the well increases, the alteration temperature curve shifts slightly lower beneath the formation temperature curve which indicates that the geothermal system in the upper 1000 m of well HE-6 has heated up. This is possibly related to the heat emitted from the 2000 and 5000 year old fissure eruptions to the east of well HE-6 (Figure 3). Lack of cuttings below 1018 m, due to circulation losses, makes it impossible to compare

deeper depths. However, it can be seen from Figure 10 that the formation temperature has decreased below about 1500 m depth, indicating cooling in the geothermal system in that area.

Fluid inclusion temperatures fill in the gap of the alteration temperatures and gives the history of the temperature variation during the lifetime of the geothermal system (Franzson et al., 2014). In this study, the homogenization temperatures of the platy calcite conform relatively well with the present formation temperature of the geothermal system.

4.5 Aquifers (feeder zones)

The locations of the aquifers, identified from temperature profiles logged during drilling, is shown in Figure 11. The geophysical logs and the circulation losses helped with the determination of these aquifers. The size of the aquifers is based on the change in the temperature gradient of each curve. The aquifers are grouped as large, medium and small. The aquifers and their locations are shown in Figure 11, together with the alteration mineral curve, the formation temperature curve (this was taken after the drilling process was completed) and the boiling point curve for comparison. The identified aquifers are listed below:

FIGURE 10: Temperature versus depth showing the formation and alteration temperature curves in well HE-6

i. Small aquifers: Nine small aquifers were observed at 90, 150, 170, 310, 480, 730, 1280, 1500 and 1625 m depths. Only 3 of these small aquifers are in the production zone of the geothermal reservoir, the rest occur above the production zone and have been cased off.

ii. Medium aquifers: Four medium aquifers were identified at 235, 625, 960 and 1025 m. Only 2 of these aquifers are in the production zone of the geothermal reservoir, the two higher up have been cased off.

iii. Large aquifer: Only one large aquifer, located at 830 m, is in the production zone of the geothermal reservoir.

The aquifers at 830 and 960 m are further indicated by loss zones. Apart from high intensity alteration, sheared or fractured rocks, the abundance of some minerals such as calcite, quartz and sulphide (mostly pyrite) occur commonly adjacent to these aquifers as vein fillings or well-formed crystals.

FIGURE 11: Temperature versus depth showing the formation temperature curve, alteration
temperature curves and the location of the aquifers

5. DISCUSSION

Identifying and understanding the controls on permeability is the key to a productive geothermal
reservoir in high temperature systems. The study of permeability can be achieved by studying the
structures, mineralogy and rock types (Reyes, 1990). Assessment of the geothermal system and the
potential for productivity is also largely dependent on the study of the hydrothermal alteration minerals.
The formation of these hydrothermal alteration minerals depends mainly on the permeability,
temperature, fluid composition, host rock composition and the age span of the hydrothermal activity in
the geothermal system. These issues were addressed by using the various analytical studies undertaken
and presented in this report.

5.1 Permeability

The permeability in well HE-6 is related to lithological contacts, intrusive contacts, fractures and major
faults. The aquifers gave good indications of permeability of the reservoir rocks. There were 14 aquifers
identified in the well. The temperature logs measured during the three drilling phases were used to locate
these aquifers in addition to circulation losses and geophysical loggings (Jónsson et al., 2002ab; Richter
et al., 2002). However, not all the aquifers may have been identified because there is the possibility that
some smaller aquifers may have been overlooked, for example, due to very small gradient changes in
the temperature logs or they might have been clogged by mud during drilling. One main factor that can

affect the interpretation of the temperature logs is the location of the pivot point in wells and the use of mud while drilling (Helgadóttir et al., 2010).

In a high temperature geothermal system, heat is transported from the depths to the surface mainly by two processes, heat conduction and convection (Hjartarson, 2014; Figure 12). Aquifers are very important in promoting conductivity and heat transfer in a high temperature geothermal system. Therefore, permeability is crucial for good conductivity and convection in the system.

In Iceland, the permeability of the reservoir rocks is controlled predominantly by stratigraphic boundaries in the upper part of the wells, while faults and fractures,

FIGURE 12: Conceptual model of a high temperature geothermal system (Hjartarson, 2014)

along intrusive boundaries control permeability in the deeper parts of the well (Franzson et al., 2001). The Hellisheidi field is dominated by northeast-southwest striking major faults and fractures intersected by easterly striking features which may affect the permeability (Hardarson et al., 2007). The alteration mineralogy of the reservoir rocks can also be affected by the permeability in the geothermal reservoir. Therefore, by studying the alteration minerals, we can deduce and confirm this speculation locally in well HE-6.

5.2 Alteration minerals

In geothermal environments, primary minerals alter and secondary minerals are formed. The mineral alteration temperature diagram in Figure 13 shows the stages of alteration in high temperature geothermal systems in Iceland. The factors that cause this hydrothermal process are the permeability, temperature, pressure, fluid chemistry, initial composition of the rock and the duration of the hydrothermal activity. These factors are independent, but the effects of one or more of the factors can exert a dominant influence on the location and extent of the hydrothermal alteration (Lagat, 2007). Some of these factors affecting the hydrothermal minerals encountered in well HE-6 are examined.
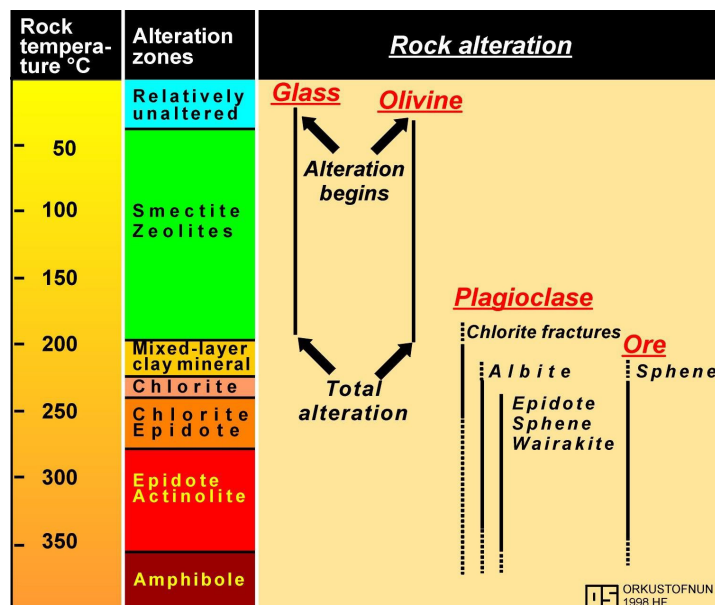
FIGURE 13: Mineral alteration temperature diagram (from Franzson, 2014)

High rock permeability and porosity should result in extensive alteration and low permeability and porosity should result in limited alteration (Franzson, 2014). However, in this study, the porosity of the rock was not a good measure of

permeability. Even though most volcanic rocks are porous they may not be permeable as often seen in altered hyaloclastite formations. Tuff and breccia are highly porous or permeable to begin with. However, they alter easily and the alteration products tend to plug the permeability. Therefore, these formations often form cap rocks. Of the hyaloclastites, the pillow lavas frequently remain quite permeable (Franzson et al., 2001).

A rise in temperature causes the primary minerals in the host rocks to alter to secondary minerals, resulting in the precipitation of new minerals. Some of the minerals identified are zeolites, clays, quartz, and prehnite. These minerals are temperature dependant and, therefore, act as geothermometers in identifying temperature changes in the geothermal system. In this study, 4 alteration zones where identified in well HE-6. These are the smectite-zeolite zone, the smectite-mixed layer clay-chlorite zone, the chlorite zone and the chlorite-epidote zone.

A comparison between alteration minerals, the fluid inclusion temperature range and the formation temperatures in Figure 10 gives a clue about the history of the geothermal system in this particular area. Three geothermal phases may exist within this geothermal system; the first one is related to the present conditions, the second one relates to a cooling episode caused by an inflow of the surrounding groundwater system, and a third one may be related to the geothermal up-flow reheating the system to its present conditions. This heating up probably relates to the renewed geothermal outflow along the 2000 and 5000 years old volcanic fissures (Franzson et al., 2005).

A comparison of the formation temperatures and hydrothermal alteration from several wells, including well HE-6, in a conceptual model of the Hellisheidi and Hverahlíd reservoirs, proposed three up-flow zones beneath Gráuhnúkar, Reykjafell and Hverahlíd (Helgadóttir et al., 2010). Well HE-6 is north of Reykjafell and south of Skardsmýrarfjall (Figure 3). Cross-sections suggest minor cooling west of Skardsmýrarfjall (Figure 14) while towards the southeast of Skardsmýrarfjall, it could be in equilibrium but there is heating up taking place beneath Gráuhnúkar in the southwest sector and in Hverahlíd (not shown in Figure 3). Figure 14 shows an example of the conceptual model and cross-section view taken from Franzson et al. (2005).

Since the Hellisheidi high temperature field is surrounded by cold ground water systems (Franzson, 2014), any change in the permeability may affect the conditions in the geothermal reservoir. For
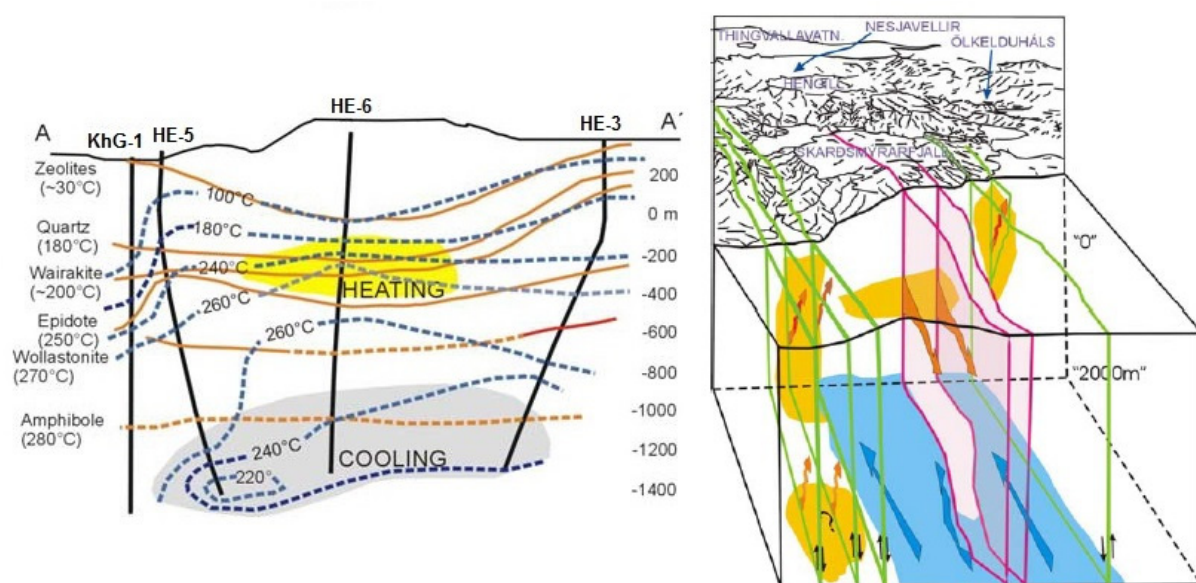

FIGURE 14: The conceptual model of the Hellisheidi high temperature geothermal field showing the cold water inflow into the geothermal reservoir beneath Skardsmýrarfjall and the cross-section of four wells including well HE-6 (Franzson et al., 2005)

example, if the pressure of the geothermal system is higher than the surrounding cold groundwater system, any permeability change will result in an outflow and heating up will occur. High pressure in the surrounding groundwater system will cause an intrusion of cold water into the geothermal system resulting in cooling in that area. This heating and cooling applies to high permeability zones and was observed in well HE-6.

6. CONCLUSIONS

Borehole geological logging is an integral part of subsurface geothermal exploration which adds to the understanding of the natural conditions in the geothermal reservoir. It involves a combination of studies to produce a conceptual model and history of the geothermal system. These studies include:

i. Lithological logging, to identify the host rocks, alteration minerals and structures.

ii. Petrographic studies to define host rock minerals, alteration minerals, structures and sequence of mineral deposition.

iii. XRD analyses to define the clay zones and confirm the alteration mineralogical zones.

iv. Fluid inclusion measurements to deduce the homogenization temperatures of the alteration minerals and salinity in the geothermal reservoir.

This study of well HE-6 is an exploratory or introductory experience for the author, being new to the study of geothermal systems. All the results presented in this report are the findings from this study, except where otherwise quoted. Several conclusions can be made from the findings of this work:

The stratigraphy of well HE-6 is comprised predominantly of four distinct hyaloclastite formations, three lava sequences and two possible intrusions, one being of intermediate composition. The hyaloclastite formations, made up of pillow basalts, breccias and tuffs, are porous in comparison to the basaltic lavas and intrusions. The rocks were not a good measure of permeability in this study because the cuttings were too small for a detailed analyses on the porosity. However, in studies undertaken for Icelandic rocks, the hyaloclastite units are not always permeable because hydrothermal alteration can plug permeability in easily altered rocks such as breccias and tuffs. Pillow basalt is the most permeable unit of the hyaloclastites because it consists of a smaller amount of glass, which is most sensitive to alteration.

The 14 aquifers identified in this study were along lithological contacts, intrusive boundaries, and along faults and fractures which are indicative of high permeability zones. The mineral structures of some hydrothermal alteration minerals such as quartz, calcite and pyrite can depict good permeable zones. These minerals occurred mostly as vein fillings or well-formed crystals adjacent to the aquifers.

The mineralogical studies recognized four distinct alteration zones: smectite-zeolite zone, smectite-mixed layer clay-chlorite zone, chlorite zone and chlorite-epidote zone. The chlorite-epidote alteration zone is the highest mineral alteration temperature zone encountered for the upper 1018 m of the borehole. The temperatures of the alteration minerals formed a hypothesis regarding the history of the geothermal system suggesting three geothermal phases: the first one is related to present conditions, the second one relates to a cooling episode caused by an inflow of the surrounding groundwater system, and a third one may be related to volcanic activity causing geothermal up-flow and reheating the system to its present condition.

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