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BOREHOLE GEOLOGY AND HYDROTHERMAL ALTERATION OF WELL HE-24, HELLISHEIDI GEOTHERMAL FIELD, SW-ICELAND

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ABSTRACT

Well HE-24 is located in the Hellisheidi high-temperature field, Hengill geothermal area in SW-Iceland. It is a vertical well reaching a total depth of 2587 m. The uppermost 1200 m are analysed here. The well was drilled in August 2006 targeting a 5000 years old NE-SW trending volcanic fissure. The lithology of well HE-24 comprises basaltic hyaloclastite formations and dyke intrusions. Five alteration zones were identified in the well: a zone of no alteration at 12-244 m, a smectite-zeolite zone at 244-616 m, a mixed layer clay zone at 616-672 m, a chlorite zone at 672-1074 m, a chlorite-epidote zone at 1074-1172 m and an epidote-actinolite zone below 1172 m depth. These zones depict a trend from a cold groundwater system down to 244 m and passing through a cap rock and into a high-temperature system. Based on mineral sequence deposition and a comparison between hydrothermal alteration and formation temperatures, the geothermal system appears to have cooled in the upper 700-800 m of the well. Around twenty-three feed points were found in the well and categorized into weak, moderate and large aquifers. Seven of these are located in the production part and are mostly associated with intrusions while sixteen aquifers are located above the production part, and are mostly associated with stratification boundaries. The correlation of hydrothermal alteration minerals of wells HE-23, HE-24 and HE-25 at Skardsmýrarfjall shows that the alteration mineral temperature has a similar increase in gradient with depth in the three wells, although it is only in well HE-24 that the epidote-actinolite zone is found above 1200 m depth.

1. INTRODUCTION

Iceland lies astride the Mid-Atlantic Ridge which is a divergent plate boundary separating the American and the European plates. An active zone of volcanism and tectonism across Iceland strikes roughly NE-SW. On the surface the rock sequence in the rift zone consists of interglacial lavas and sub-glacial hyaloclastites with an age of less than 0.7 million years (Björnsson et al., 1986). Geothermal areas in Iceland are divided into high- and low-temperature areas. The former are located within the rift zones. One of the largest high-temperature geothermal areas in Iceland is found within the Hengill central volcano about 30 km east of Reykjavik, the capital city of Iceland. Hellisheidi field is a part of the Hengill geothermal area (Franzson et al., 2005).

This report deals with the borehole geology and hydrothermal alteration in the upper 1200 m of well HE-24 in the Hellisheidi field. The data of the geology and hydrothermal alteration from the well were collected from cutting analyses using a binocular microscope (Olympus SZ12 with a magnification of 7x to 90x), thin section petrography, X-ray diffraction analysis, fluid inclusion geothermometry, and geophysical logs. A total of approximately 600 samples of drill cuttings starting at a depth of 12 m down to 1200 m were collected at 2 m intervals and analysed, to identify sub-surface formations, locate aquifers or feed zones and determine the temperature of the reservoir. For more detailed analysis, petrographic thin section analysis was used to recognize minerals based on optical properties, textural relationships and the intensity of rock alteration. This method was also used to study the replacement of the primary minerals and to determine the depositional sequence in veins and vesicles. Additional methods included X-ray diffraction (XRD) and fluid inclusion analysis. XRD analysis was used to identify the types of clay and the study of fluid inclusions was used to determine the temperature of the reservoir fluids as well as their apparent salinities.

This report is the result of a study done during the six-month course at the UNU Geothermal Training Programme, Iceland, in the year 2007.

2. GEOLOGICAL OUTLINE

2.1 Regional geology

Iceland is located where the astenospheric flow under the NE-Atlantic plate boundary interacts and mixes with a deep-seated mantle plume. The buoyancy of the Icelandic plume leads to a dynamic uplift of the Iceland plateau, and high volcanic productivity over the plume produces a relatively thick crust. The Greenland-Faeroe ridge represents the Icelandic plume track through the history of the NE-Atlantic.

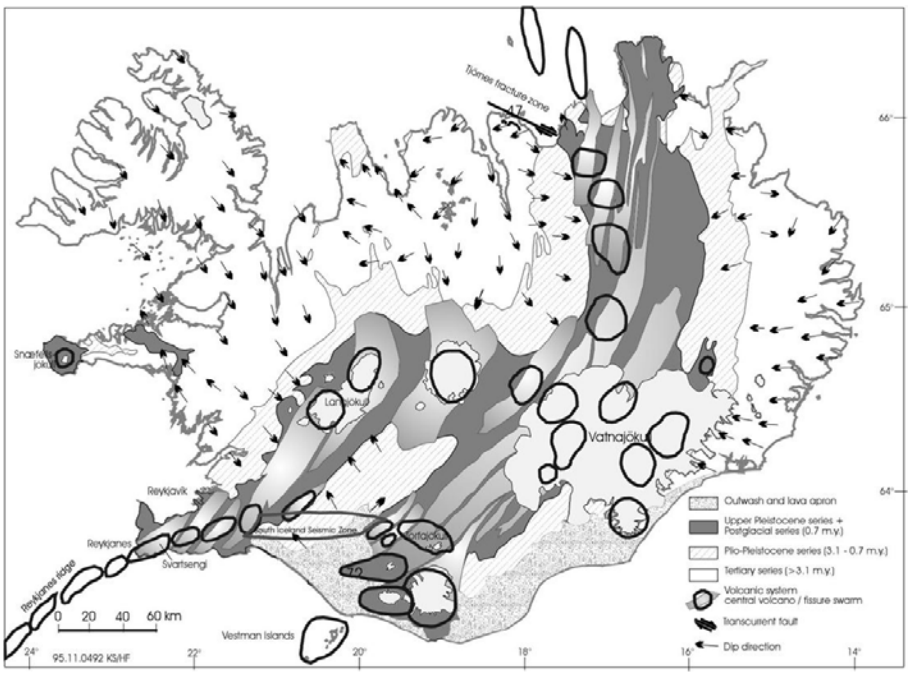

FIGURE 1: A simplified geological map of Iceland showing the volcanic zones, fissure swarms and central volcanoes

The two main volcanic belts are connected in central Iceland by the Hofsjökull volcanic system (Figure 1). The northern end of the eastern main volcanic belt is connected to the crestal zone of the Mid Atlantic Ridge by active transcurrent faults. In southern Iceland this volcanic belt is propagating south; at the same time dilation is dying out in the northern half of the western main volcanic belts.

The main volcanic belts in Iceland are

displaced eastward relative to the crest zone of the Mid Atlantic Ridge because the lithosphere tends to break up above the mantle plume and the plume has been moving east relative to the plate boundary. Some 13-8 Ma ago the active volcanic belts in Iceland lay E-W along the Snaefellsnes peninsula in western Iceland and then curved northwards to the Húnaflói region in northern Iceland. At the end of that period it shifted from Snaefellsnes to its present position between Reykjanes and Langjökull. One to two million years later the volcanic belt in northern Iceland shifted from Húnaflói to its present position. Since then, this eastern belt has been propagating towards the southwest and simultaneously the dilation in the northern end of the western belt has been dying out (Saemundsson, 1979).

Vertical sections of the volcanic sequence in Iceland expose up to 1500 m thick pile of volcanic rocks below which lies at least another 2-5 km thick sequence of extrusives. The exposed volcanic pile is built predominantly of basalts (80-85%), while acidic rocks, including intermediate rocks, constitute about 10%. The amount of sediments of volcanic origin is in the order of 5-10% in a typical Tertiary lava pile but much higher in Quaternary rocks (Saemundsson, 1979).

The dyke or fissure swarms are characterized by extensional tectonic features such as open fissures, graben structures and crater rows at the surface, with dykes and normal faults at deeper levels. The active period of the volcanic systems has been found to vary from 300,000 years to over 1 million years (Saemundsson, 1979).

Based on the geological setting and temperature data from drillholes, the geothermal areas in Iceland are classified as high or low in temperature. The high-temperature areas are located within the active volcanic belts in the country, whereas most of the low-temperature areas occur in Quaternary and Tertiary formations. Twenty high-temperature areas have been confirmed in Iceland. Most of the high-temperature areas are located on high ground. The rocks are geologically very young and permeable. However, a few high-temperature areas, such as Geysir, Hveravellir and the Hveragerdi field east of the Hengill area, are less elevated and there the geothermal groundwater table reaches the surface and silica sinters have been deposited around the hot springs (Arnórsson, 1995).

2.2 The Hengill high-temperature area

The Hengill geothermal area is situated in the western rift zone of Iceland and belongs to the Hengill volcanic system. The volcanic system includes an approximately 60-100 km long NNE-SSW trending fissure swarm with normal faults, fissures, frequent magma intrusions and a central volcano. Three well fields have been developed within the greater Hengill area: 1) Nesjavellir where a 120 MW power plant is currently in operation, 2) Hellisheidi where a 90 MW power plant is currently in operation, and 3) Hveragerdi, where the geothermal resource is utilized by the local community (Björnsson et al., 2003; Franzson et al., 2005). Two other fields, Ölkelduháls and Hverahlíd, are under consideration.

The geothermal fields within and around the Hengill volcano have been studied extensively from as early as 1947. Permeability in the reservoir is believed to relate largely to intrusive boundaries and major faults. Of particular interest are two NNE-SSW basaltic dykes from 2000 and 5000 years old fissure eruptions, which are believed to provide strong geothermal flow channels in the system from a proposed upflow zone in the central part of the Hengill volcano to the north and south (Franzson et al., 2005).

Hengill central volcano is mainly built up of sub-glacial hyaloclastite formations. Interglacial sub-aerial lavas erupting within the volcano flowed down the volcano flanks and accumulated in the surrounding lowlands. The age of the volcano has been assessed at around 400,000 years (Franzson et al., 2005).

2.2.1 Geology and structures of Hellisheidi field

The Hellisheidi high-temperature field is located in the southern part of the Hengill central volcano (Figure 2). This is a part of the 110 km^2 Hengill low-resistivity anomaly (Árnason and Magnússon, 2001). Exploration started in 1985 with a well drilled at Kolvidarhóll, followed by a well at Ölkelduháls in 1995 (Franzson et al., 2005). Fault and major fractures strike mostly NNE-SSW and are conspicuous in the east and west, marking the boundaries of the volcano's fault and fissure zone. Post glacial volcanism includes three fissure eruptions of 9000, 5000 and 2000 years in age. The fissures can be traced further to the north, through Nesjavellir field and into Lake Thingvallavatn (Saemundsson, 1995).

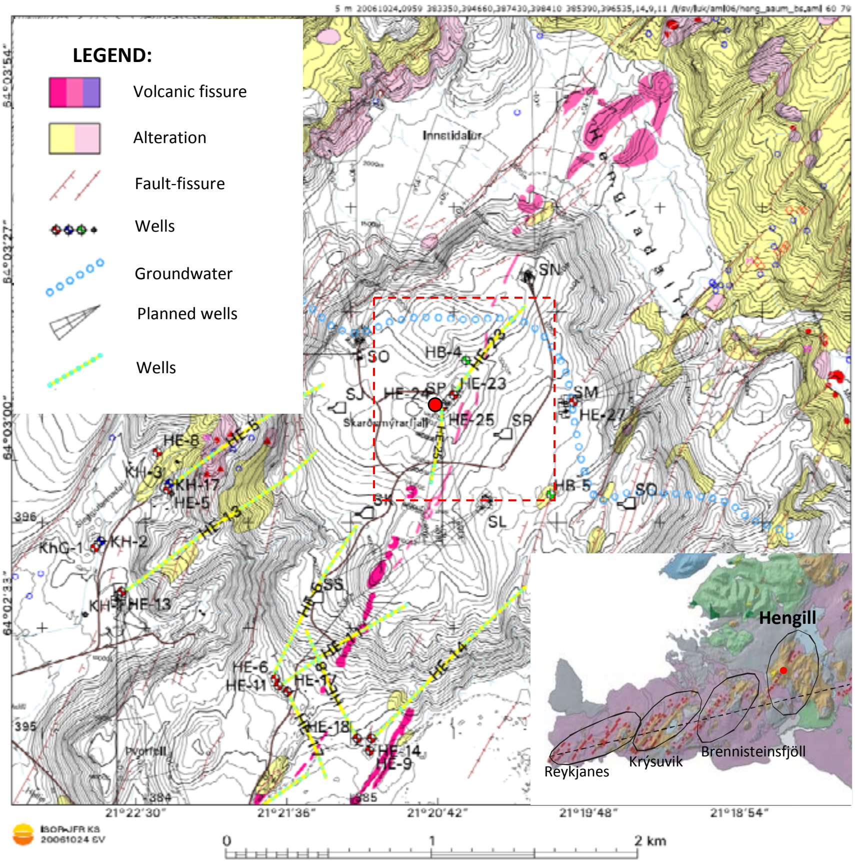

FIGURE 2: Volcanic activity, alteration, fissure system and location of well HE-24 and surrounding wells at Skardsmýrarfjall

The rock sequence of the Hellisheidi field predominantly consists of hyaloclastites and lava series (Figure 3). Hyaloclastites are formations of relatively limited horizontal extent which makes them of limited use as marker horizons. Lava series are seen to bank up against the volcano in the western part, as lavas are a feature of valley infillings. Unusually large faults delineate the western margin of the fissure swarm and indicate the termination of volcanic activity west of the Hengill area. This may, furthermore, imply that the high-temperature reservoir deepens sharply west of the faults (Franzson et al., 2005).

2.2.2 Geophysics

Aeromagnetic, gravity and DC-resistivity surveys were carried out between 1975 and 1986. These delineated a 110 km^2 low resistivity area at 200 m b.s.l. a negative and transverse magnetic anomaly coherent with the most thermally active grounds (Björnsson et al., 1986). The resistivity map was revised between 1986 and 2000, by applying the central loop transient electromagnetic sounding method (TEM) (Figure 4). These data imply that despite being widespread, the resistivity anomaly is complex and affected by processes such as faulting, shearing and spreading (Árnason and Magnússon, 2001).

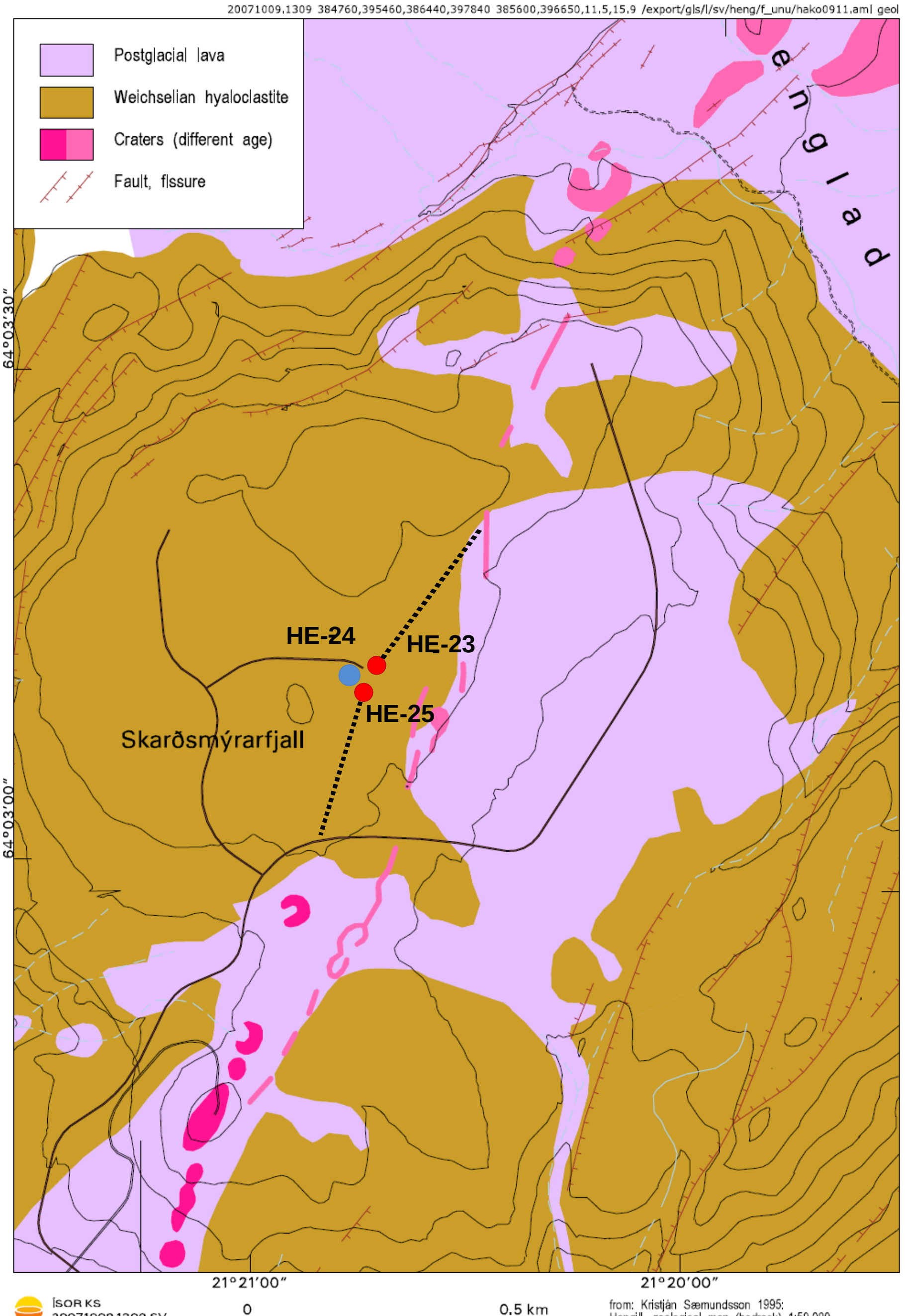

FIGURE 3: Geology of the Skardsmýrarfjall and the location of well HE-24 (slightly modified from Saemundsson, 1995)

3. BOREHOLE GEOLOGY

Skardsmýrarfjall mountain is part of the Hellisheidi geothermal field. Geologically Skardsmýrarfjall consists of hyaloclastites and postglacial lava (Figure 3). The mountain is succeeded by 5000 years old volcanic fissures trending in a NNE-SSW direction. Three wells are located in the middle part of the area, namely well HE-23, a directional well with a northeasterly direction, drilled in 2006 to a total depth of 1968 m, followed by well HE-24, which is vertical and reached a depth at 2587 m. The other directional well is well HE-25, which was drilled in December 2006 to a depth of 2155 m.

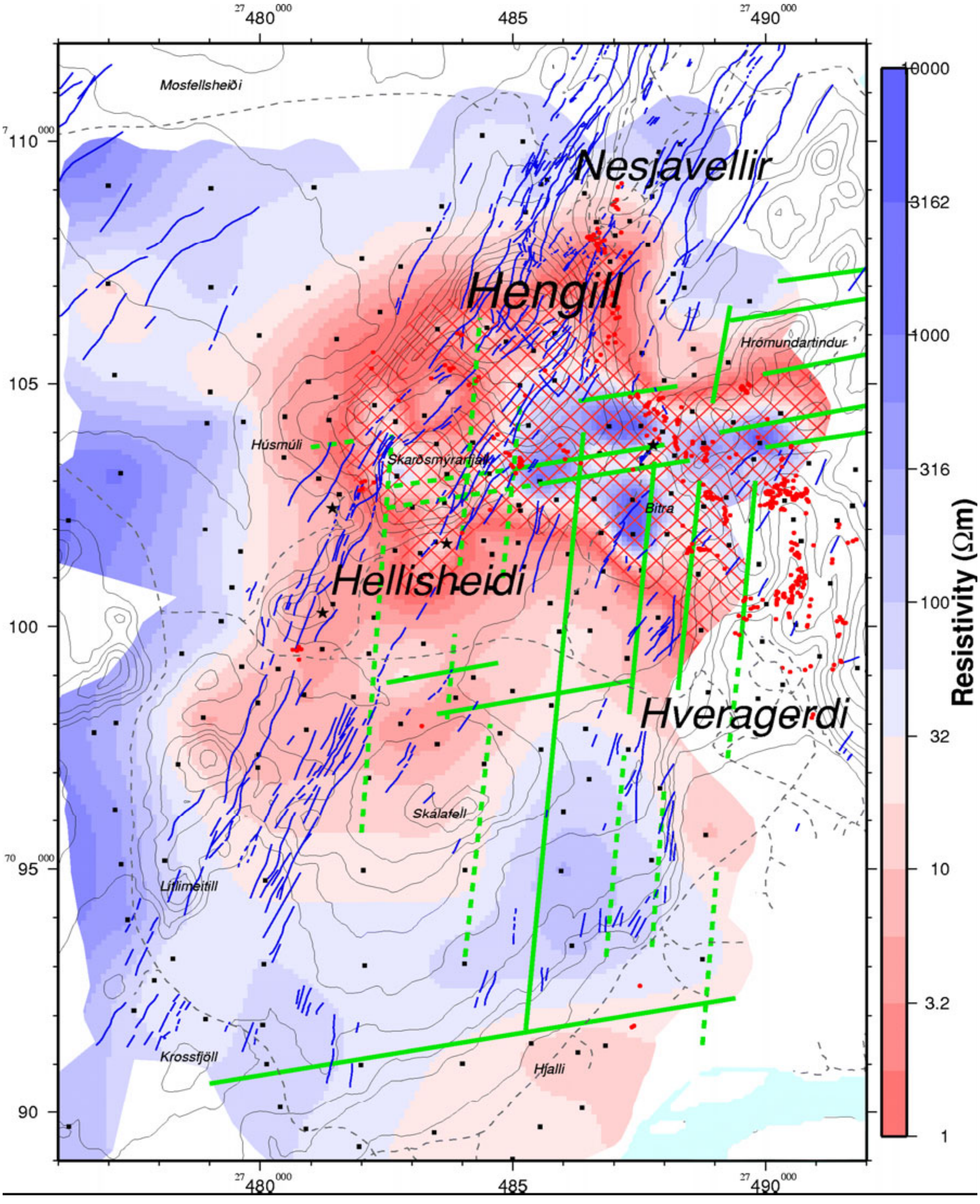

FIGURE 4: Resistivity in the Hengill area at 100 m b.s.l. according to a recent TEM survey (Árnason and Magnússon, 2001)

3.1 Drilling of well HE-24

Well HE-24 is a vertical well with a total depth of 2587 m. The design of the well is shown in Figure 5. The well pad is located at coordinates X = 385418,13 E, Y= 396557,04 N and Y= 572,5 m a.s.l. and on the same drillpad as directional wells HE-23 and HE-25. The drilling target of this well was the NNE - SSW fractures associated with the 5000 years old volcanic fissures that dissect the Hellisheidi geothermal field.

This well was drilled by the Icelandic drilling company, Jardboranir Ltd. The drilling started on August 26 and finished on October 13 2006. Two drill rigs, Sleipnir and Ódinn, drilled the well. The drilling progress is shown in Figure 6. The drilling was carried out in four stages as shown in Figure 5. The pre-drilling stage was done using Sleipnir drill rig to 96 m with a 26" drill bit and cemented using a 22½" casing. The first stage was 21" drilling from 96 to 351 m and cased with an 18⅝" casing. After that, a larger drill rig, Ódinn, continued the drilling. The next stage was drilling to 711 m with a 17½" drill bit and a mud motor. The well was cemented using production casing, sized 13⅜". The final stage, or the production stage, was drilled to 2587 m using a 12¼" drill bit. A 9⅝" slotted liner was hung at 675 m to the bottom. In the first and second main drilling stages, mud was used. In the production part, the well was drilled using aerated water.

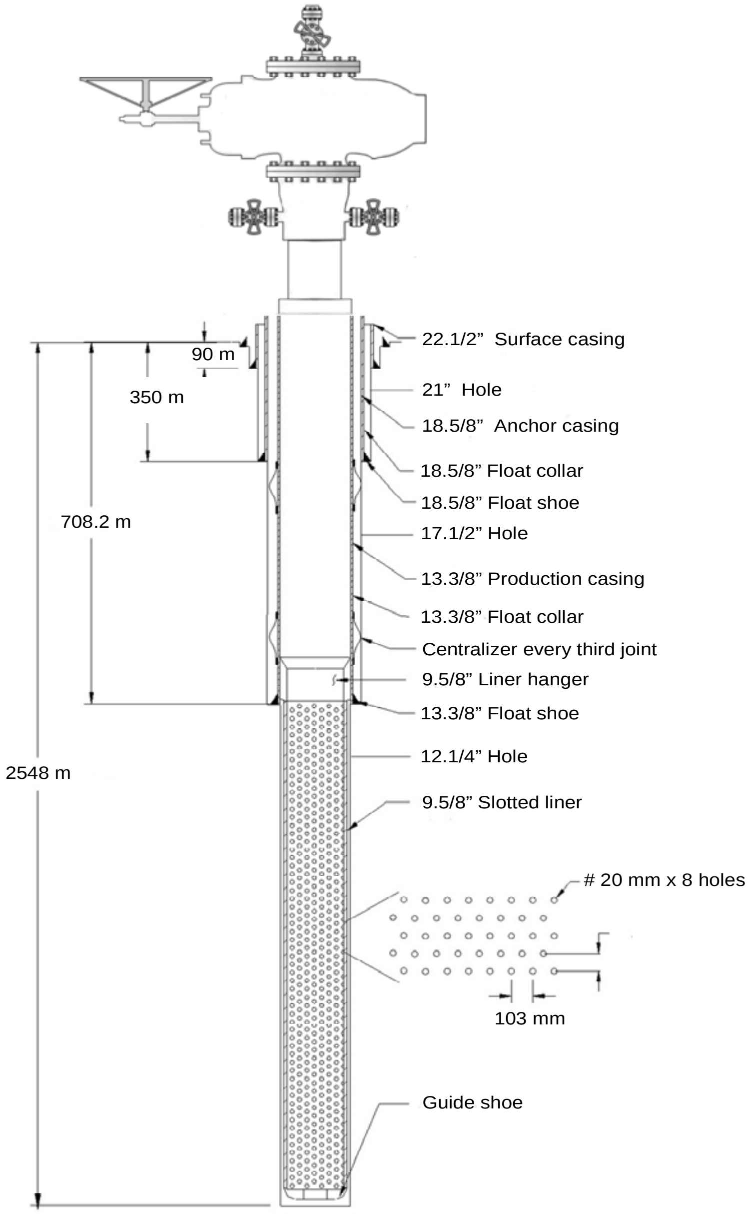

FIGURE 5: Well design of well HE-24 (Mortensen et al., 2006a and 2006b)

3.2 Stratigraphy

The stratigraphy of well HE-24 down to 1200 m depth is analysed here and shown in Figure 7. It is divided into a number of rock types, mainly depending on the crystallinity of the rock. Thus, basaltic tuff is composed of volcanic glass, basaltic breccia is a mixture of partially crystallized basalt and volcanic glass, glassy basalt, which often is interpreted as pillow basalt, is mostly made up of partially crystallized rock with minor amounts of volcanic glass, and finally crystallized (fine- to coarse-grained) basalt, which forms either sub-aerial lavas or intrusions. The volcanic products are, in most cases, very porous, while intrusions are dense. The porphyritic or aphyric character of the rock is very useful in separating one volcanic formation from another.

The description of the rock formation from well HE-24 is mostly based on the binocular microscope and aided by petrographic thin section analysis. The stratigraphy of the well consists predominantly of alternating sequences of sub-glacial hyaloclastites and basaltic intrusive rocks. The stratigraphy of the uppermost 1200 m of well HE-24 is divided into the following volcanic formations:

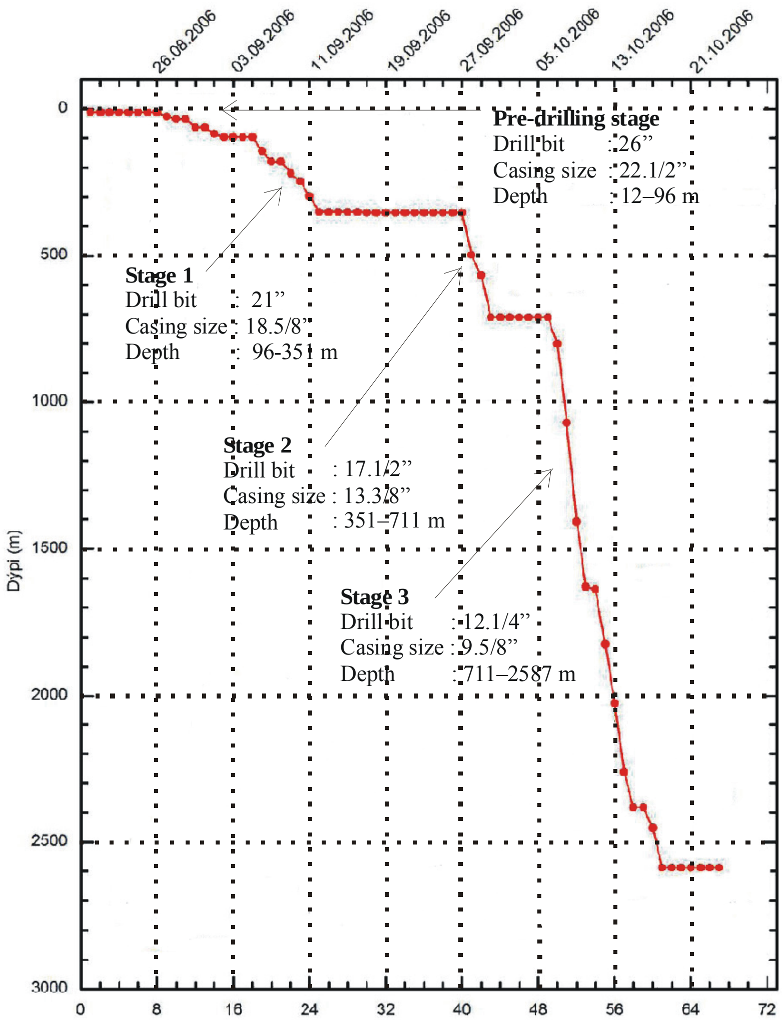

FIGURE 6: Drilling progress of well HE-24 (Mortensen et al., 2006a and 2006b)

Hyaloclastite I (12-350 m)
This volcanic hyaloclastite formation is divided into three main units: a) Cutting samples are scarce down to 102 m, but basaltic tuff dominates where samples were taken. b) Pillow basalt dominates the succession below the tuff down to 236 m with the exception of a medium- to coarse-grained basalt at 190-196 m and a breccia layer at 196-218 m. c) Medium- to coarse-grained basalt is found between 240 and 286 m. d) Basaltic breccia is found in samples from 286 to 350 m.

The rock is plagioclase porphyritic, vesicular and very fresh. The voids are empty except for minor amounts of limonite and spherical siderite. The unit is believed to belong to the Skardsmýrarfjall hyaloclastite formation that erupted during the last glacial (15,000 - 115,000 years).

Hyaloclastite II (350-478 m)
This hyaloclastite formation consists of three sub units. Basaltic tuff forms the upper part (354-392 m), the middle part is a basaltic breccia (392-446 m) and the lower part is pillow basalt (446-478 m). The rock sequence is entirely fresh. The basaltic tuff is black in colour, very vesicular and aphyric. The tuff appears to become more vesicular with depth. Occasional calcite and pyrite were identified but they were never common. The basaltic breccia is oxidized mainly in the upper part. The pores in the rock are, to some extent, filled by limonite and siderite. The rock shows some alteration to clay or zeolite. The glassy basalt (pillow basalt) is fresh, grey-black in colour and more vesicular than the glass. The rock is differentiated from the upper formation by its aphyric texture.

Hyaloclastite III (478-773 m)
This formation includes a thick layer (295 m) of basaltic tuff in the upper part, basaltic breccia in the middle part and basaltic tuff and basaltic breccia at the base.

The basaltic tuff (478-516 m depth) is plagioclase porphyritic which distinguishes it from the overlying formation. It shows increasing alteration with depth obtaining greyish to green colour. Voids are predominantly filled by calcite, clay and pyrite. Pyrite and calcite are moderately abundant and increase with depth. Low-temperature hydrothermal minerals such as chalcedony, limonite, siderite and zeolites are present. A thin section at 510 m depth shows that the rock consists of mostly altered glass to clay, calcite and quartz with some fragments of crystallized basalt.

The unit of basaltic breccias at 516-666 m depth is slightly altered at the top, moderately altered in the middle, and highly altered in the lower part. The rock sequence is vesicular and fractures are filled by calcite, pyrite and oxidation. The oxidation increases mostly in the middle part. Some zeolites such as thomsonite and scolesite appear from 520-526 m and 568 m. Three thin sections are available at 560, 616 and 668 m depth. These samples are basaltic breccia with pore and vein fillings including thomsonite, mesolite, and scolesite.

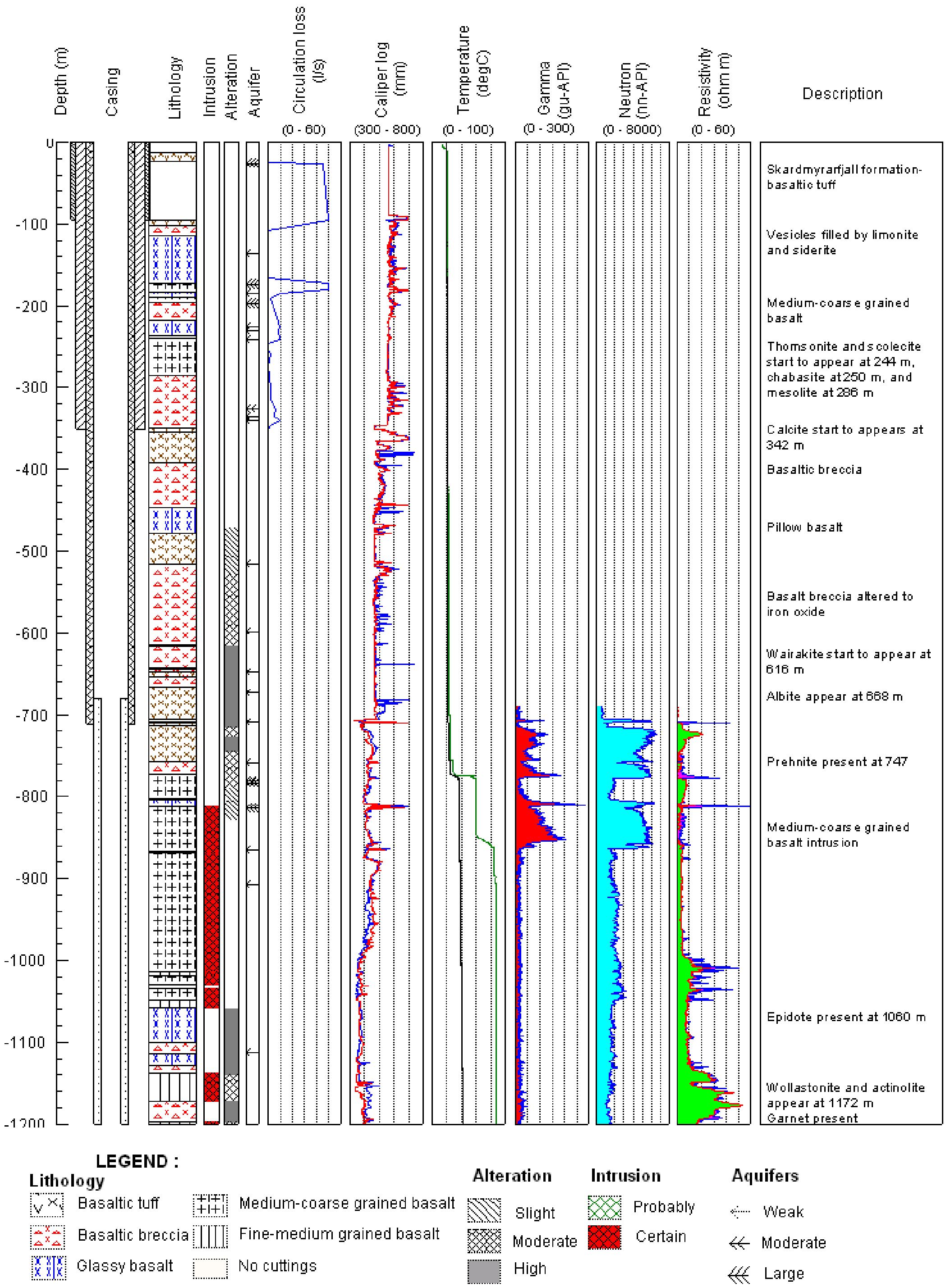

FIGURE 7: Simplified stratigraphic section and geophysical logs down to 1200 m depth in well HE-24

The third sub unit is basaltic tuff (666-757 m depth) underlain by a thin layer of basaltic breccia (757-773 m depth). This sequence is highly altered in the upper part but moderately to highly altered in the lower part. Calcite and pyrite are very common in this sub-unit but decrease to moderate occurrence at the bottom part. Clay alteration is intense in this rock. Some zeolites, such as scolecite, were identified at 688-698 m depth. Quartz becomes very common especially in the pores of the rock. Veins and vesicles are filled by various minerals such as calcite, quartz, wairakite, clay and sphene.

Basalt and intrusion (773-1058 m)
This unit predominantly comprises three sub-units of medium- to coarse-grained basalt (773-802 m, 812-1014 m, 1018-1048 m), a thin layer of glassy basalt (804-812 m depth), and fine- to medium-grained basalt (1014-1018 m, 1048-1058 m depth). This unit is moderately altered at the top but becomes slightly unaltered at the base. The glassy basalt in the upper part is predominantly green in colour caused by hydrothermal alteration. Calcite and pyrite are very common minerals here and oxidation is present.

The medium- to coarse-grained basalt at the base is connected by thin fine- to medium-grained basalt in the middle and bottom parts, and is relatively fresh. Vesicles are rare, but some of them as well as the veins are filled by clay, quartz, wairakite and calcite.

Hyaloclastite IV and intrusion (1058-1200 m)
This rock consists of alternating layers of glassy basalt and basaltic breccia (1058-1138) and inter-layered fine- to medium-grained basaltic intrusions. The rock is porphyritic, and moderately to highly altered. Epidote starts to appear at 1074 m depth, and is mostly found in pores and veins in association with quartz. Some are also found with prehnite together with wollastonite, which was first identified at 1172 m depth from petrographic thin section and at 1176 m with the binocular microscope. Actinolite also appears first at 1172 m as an alteration of pyroxene. Epidote, wollastonite and actinolite are present down to the 1200 m depth. Garnet starts to appear with anhedral-subhedral shape at 1184 m.

In the middle and at the bottom part of the unit there is a fine-medium grained basaltic intrusion with an interlayer of highly altered basaltic breccia. The fine-medium grained basalt is greyish-yellow to pale green, and is porphyritic in texture. In this part, quartz, epidote and wollastonite were identified in small-moderate veins at depths down to 1190 m.

3.3 Intrusive rocks

Intrusive rocks are mostly characterized by relatively low alteration compared to the surrounding rock, of a compact nature, and sometimes marked by oxidation near their margin. Intrusions usually show relatively high peak values in neutron-neutron and resistivity logs (Franzson et al., 2005).

Intrusive unit 1 (812-1058 m)
This intrusion is a coarse- to fine-grained aphyric basalt. The rock is fresh and is marked by high peaks of neutron-neutron and resistivity in the lower part. The intrusion may be divided into two units based on a fine-grained contact. The apparent thickness is 146 m. True thickness is believed to be only a fraction of that, the reason being that the well is drilled along the dyke. It is of interest to note in Figure 7 that oxidation, which is an indication of the dyke's contact thermal effect, starts more than 40 m above the first encounter of the intrusion, indicating that the well is already very near to the dyke far above. This is important when considering the relation of the many feed points to the dyke that are below 700 m. It is possible that this intrusion represents the dyke feeder to the 5000 year old eruption, which is exposed a few tens of metres east of the well. Chemical analysis of selected cuttings from the margins of the dyke may confirm that.

Intrusive unit 2 (1138-1172 m)
This is fine- to medium-grained basalt with an apparent thickness of approximately 34 m. The rock is greyish-pale green in colour and plagioclase porphyritic. The rock is moderately altered. The intrusion is probably a different intrusion from the one encountered above. The geophysical logs show moderate peaks of neutron-neutron and high resistivity values.

Intrusive unit 3 (1198-1200 m)
This is a 2 m thick fine- to medium-grained basalt dyke intrusion. The rock is moderately altered and characterized by being plagioclase porphyritic; it may be the same intrusion as the one above. It shows high-resistivity peaks.

4. HYDROTHERMAL ALTERATION

The factors which can affect the formation of hydrothermal minerals are temperature, pressure, parent rock types, reservoir permeability, fluid composition and duration of activity. The hydrothermal alteration in the geothermal system is a product of water-rock interaction. The parent rock influences hydrothermal alteration mainly through the control of permeability by texture and porosity. Studies on alteration in geothermal fields have clearly recognized the important control of permeability on hydrothermal mineral deposition (Browne, 1978). The hydrothermal minerals can be useful as geothermometers (Figure 8), in determining depth of production casing, in estimating fluid pH and other chemical parameters, as well as predicting scaling and corrosion tendencies of fluids, measuring permeability and possible cold-water influx, and as a guide to hydrology (Reyes, 1990).

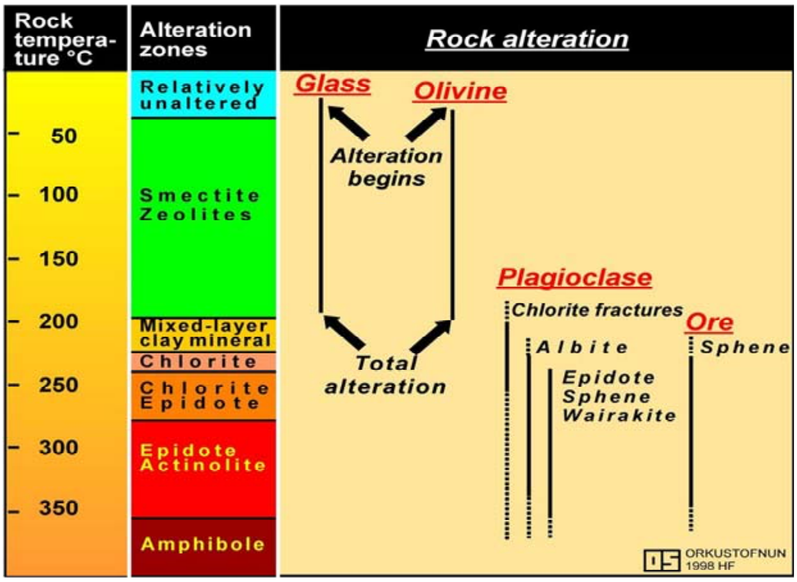

FIGURE 8: Mineral alteration temperature diagram (Franzson, 2007)

4.1 Primary rock minerals

Most rocks in geothermal areas contain some primary minerals which are unstable in a geothermal environment. These have a tendency to be replaced by new minerals that are stable, or at least metastable, under the new conditions. The replacement hydrothermal minerals record the interactions between the wall rocks and the hydrothermal fluids, while space-fill minerals reflect the processes that affected the circulating fluids (Browne, 1978).

The primary minerals in the rocks penetrated by well HE-24 (Table 1) are characterized by an abundance of glass, olivine, plagioclase, pyroxene and opaques. The replacement of the minerals can best be studied by petrographic thin-section analysis. The primary minerals first began to alter intensively in this well below 478 m depth.

TABLE 1: Primary rock minerals and their products as found in well HE-24

Relative susceptibility	Primary rock minerals	Alteration mineral products
Most susceptible	Glass	clay, calcite, quartz
↓	Olivine	clay, calcite, sphene
	Plagioclase	clay, albite, calcite, quartz, wairakite, epidote
	Pyroxene	clay, actinolite, sphene
Least susceptible	Opaque	sphene, sulphides (pyrite)

Glass: The most common replacement minerals of glass are clay minerals and calcite. It started to alter to clay at 160 m depth, to calcite at 334 m, and to quartz at 510 m.

Olivine is the most unstable mineral. It was easily identified in medium- to coarse-grained basalt below 190 m and in intrusions below 826 m. The mineral usually starts to alter along fractures and is usually replaced by clay. Some of the olivine occurs as part of the matrix in microcrystalline rocks.

Plagioclase is common in the microcrystalline ground-mass and as phenocryst in basalt. This is seen throughout the well where crystallization has taken place, and is obvious where the rock is plagioclase porphyritic. The subhedral-euhedral shape of this mineral is well recognized from 160 m. Plagioclase starts to alter commonly into clay and sphene below 510 m, to wairakite at 616 m, to albite at 668 m depth and at 1172 m it was replaced by epidote.

Pyroxene was observed as phenocrysts and in the ground-mass in this well. It began altering to clay at 668 m depth and to actinolite at 1172 m.

Opaque minerals mostly alter to sphene. They were first observed to alter to pyrite and sphene in a petrographic thin section at 510 m depth.

4.2 Distribution of hydrothermal minerals

The most common hydrothermal alteration minerals in well HE-24 are quartz, calcite, pyrite, low-temperature minerals such as the zeolite group, smectite and chlorite (Figure 9). Calcite is of special importance as it is deposited closest to the present time in the geothermal system. The temperature of calcite deposition is relatively difficult to determine; experience has taught us that this mineral disappears at temperatures above 290°C (Franzson, 2000; Kristmannsdóttir, 1979).

Limonite is a hydrated iron oxide mineral, characterized by a reddish brown, spherical, concentric pattern mostly found in veins and vesicles. This mineral was identified in the first samples taken from the well and is common down to about 350 m depth. Minor amounts of limonite were seen in the pillow basalt around 470 m. Limonite is associated with cold groundwater systems and is usually the earliest mineral to precipitate in that environment.

Siderite is a mineral composed of iron carbonate. The colours are yellow-brown; it has a curved and striated shape, commonly found in veins and vesicles. It began to appear at 24 m depth.

Zeolites are among the most common products of chemical interaction between groundwater and bedrock during diagenesis and low-grade metamorphism. It is formed during alteration and precipitation in vugs and vesicles (Saemundsson and Gunnlaugsson, 2002). They are often classified by shape into three main categories: fibrous/acicular, tabular/prismatic, and granular. Groups of zeolites such as mordenite, heulandite, stilbite, epistilbite, chabazite, thomsonite, scolecite, and mesolite occur at temperatures below 100°C; maximum temperatures vary from 100 to 120°C (Kristmannsdóttir and Tómasson, 1978; Kristmannsdóttir, 1979).

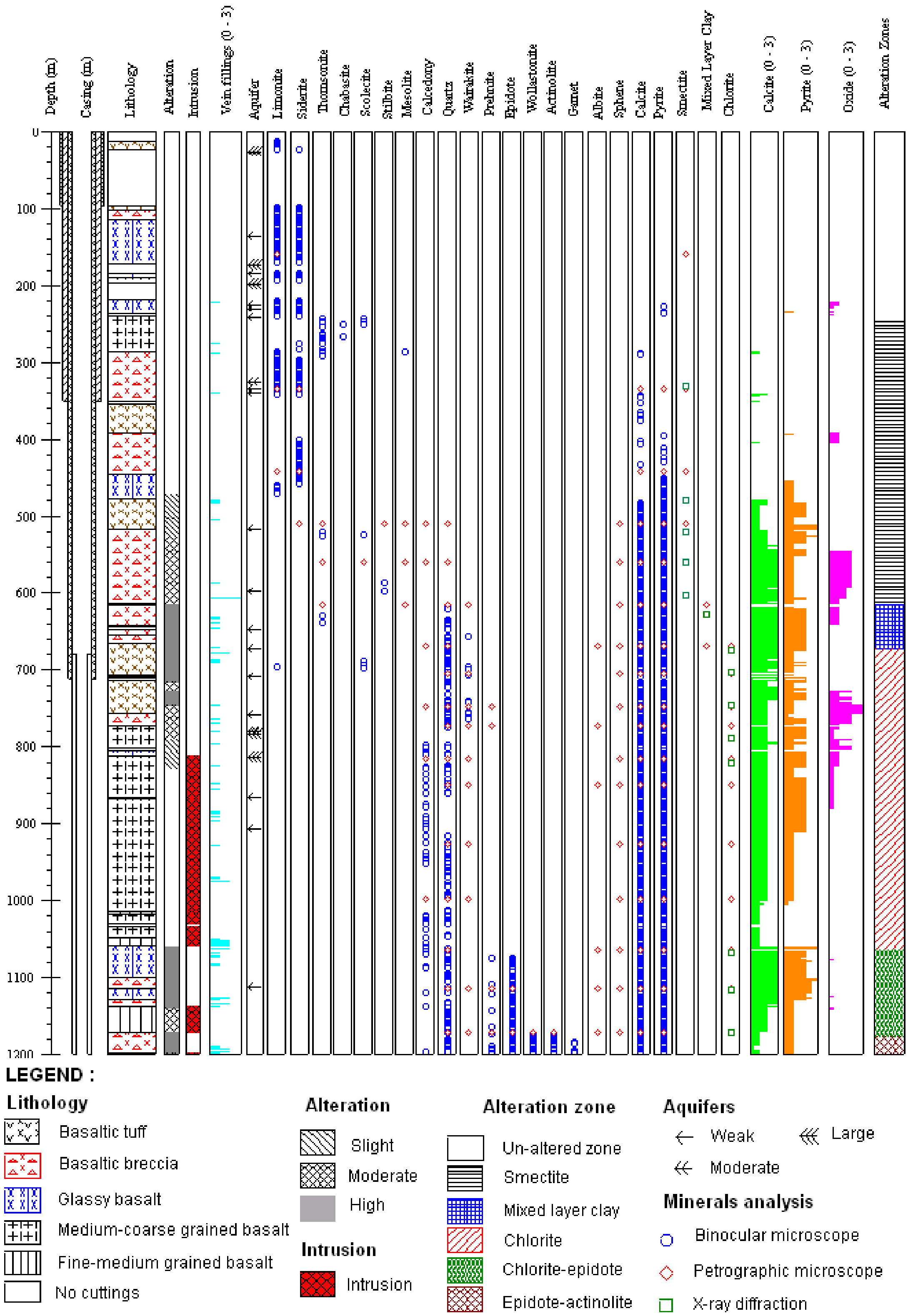

FIGURE 9: Distribution of hydrothermal alteration minerals in the first 1200 m of well HE-24

Scolecite/Mesolite. Scolecite occurs as acicular and fibrous aggregations. In the well, this mineral was identified at 286 m and down to 510 m depth. This mineral is associated with stilbite/heulandite and quartz. Scolecite and mesolite are believed to form at temperatures above about 80°C up to just above 100°C.

Stilbite was recognized in the binocular microscope and petrographical thin section at a depth of 510, 586 and 598 m. It is characterized by radial and fan-like aggregates.

Thomsonite is usually colourless-white, radial aggregates. Sometimes it appears together with other low-temperature zeolites in the vesicles and veins. In this well, it began appearing at 244 m depth.

Chabasite was observed as colourless-white, near cubical, with uneven fractures. It was identified in vesicles at 250 and 266 m depth. It is stable from about 30 to about 80°C (Kristmannsdóttir, 1979).

Wairakite was identified from 616 m depth and is found sporadically down to 1200 m depth. Under petrographic microscope, wairakite is characterized by a colourless to low grey colour under cross-polarized light and has conspicuous twinning. Wairakite appears first at about 200°C and has been recorded at temperatures as high as 300°C (Kristmannsdóttir, 1979).

Chalcedony was recognized in the binocular from 798 m and in thin section at 510 m. It is characterized by colourless-blue colour under the binocular. In thin section it is seen as light white in colour as a thin layered lining in vesicles, often replaced by quartz. This mineral began to precipitate as chalcedony at temperatures above 120°C (Kristmannsdóttir, 1979).

Quartz is a common alteration mineral found in this well from 510 m depth. It is mostly observed as a filling in veins and vesicles. This mineral appears at temperatures above 180°C (Saemundsson and Gunnlaugsson, 2002).

Albite was recognized from thin section as an alteration of plagioclase first seen at 668 m depth and is seen in most rocks below that except in the intrusions, which are relatively fresh. This alteration is known as albitization. It becomes more pervasive toward the centre of the zone of alteration (Thomson and Thomson, 1996). Albite alteration of plagioclase starts at temperatures above about 220°C.

Prehnite. In this well, prehnite is mostly found together with epidote precipitating in veins and vesicles, occasionally replacing olivine. In binocular microscope it sometimes is seen as being radial and colourless-white. Under petrographic thin section it is marked by high relief, strong interference colours and a bow-tie texture. The lower temperature stability of prehnite is not precise. It appears to be able to form at temperatures above 200°, even up to about 250°C, and is present in the geothermal systems to above 300°C (Browne, 1978; Saemundsson and Gunnlaugsson, 2002). Prehnite was found in thin sections at 747 m depth and then below about 1114 m.

Epidote is a colourless-pale green mineral, with a weak to moderate green colour of pleochroism. It often is found in pores and veins and may replace the matrix and plagioclase. Epidote is often associated with quartz. The lower temperature stability of epidote is 240-250°C and it remains stable to well above 300°C (Browne, 1978). It was first encountered at 1074 m depth and continuously from there to the bottom of well.

Wollastonite is hairy in shape and white in colour. Wollastonite is found together with epidote and sometimes with quartz. In this well it was found below 1172 m. It is usually found at temperatures above 270°C (Kristmannsdóttir, 1979).

Actinolite is a high-temperature amphibole mineral. In this well the mineral is greenish in colour, replacing mostly pyroxene. It is characterized by being fibrous and having a vitreous lustre. It was first found below 1172 m. Actinolite forms at temperatures around 280°C (Kristmannsdóttir, 1979).

Garnet was identified in a few samples below 1184 m, in highly altered rock to moderate alteration close to the intrusion body. Garnet forms at and above 280°C (Saemundsson and Gunnlaugsson, 2002).

Sphene is predominantly found as a replacement of opaque minerals. This mineral was identified from petrographic thin section at 510 m depth. Sphene is characterized by red-dark brown coloured crystals.

Calcite. **V**arious carbonate minerals have been identified in the geothermal reservoirs in Iceland. In well HE-24 sporadic small radial carbonate is found from near the surface to about 460 m depth. It is here tentatively identified as siderite. This type is predominantly found in relatively cold groundwater systems. Calcite was first identified sporadically below 288 m. It becomes very common below 482 m and is still common at 1200 m depth. A quantity assessment of calcite is shown to the right in Figure 9. Calcite is dissolved or precipitated by groundwater depending on several factors such as temperature, pH, and dissolved ion concentrations (Browne, 1978). In the thin section, calcite was identified by the very high to very low birefringence. Calcite has a wide temperature stability range up to about 300°C.

Pyrite occurs sporadically below 228 m and more or less continuously below 396 m depth. It is easily identifiable in a binocular microscope. Pyrite was assessed quantitatively in the cutting analysis and is presented in a histogram in Figure 9. This mineral is used sometimes as a permeability indicator. The temperature range of pyrite is around 120 to 260°C (Reyes, 2000).

Clay minerals are mainly formed by geothermal alteration from the surface to the deepest parts of geothermal systems (Saemundsson and Gunnlaugsson, 2002). These minerals have been widely used in geothermal exploration as temperature indicators (Kristmannsdóttir, 1979). These minerals have been studied in three different ways: In binocular analysis their increasing crystallinity was looked at with temperature in vein and vug fillings. Petrographically we studied the change in optical characteristics going from smectite to chlorite. Both these methods are rough guides to the identification of the clays as one tends to focus on the mineral deposition as opposed to the whole rock alteration. The third method is XRD-analysis, which is based on shaking the cutting sample in water, allowing the small clay particles to break from the rock and be suspended in the water. These clay particles, which are derived from the whole rock, are then run in the XRD giving an overall picture of the clays present in the rock.

Smectite. The rocks are very fresh down to around 472 m depth. Very minor clays can be seen petrographically, and then as a transformation from palagonite. The low clay content in the rock resulted in smectite being below analytical resolution in the XRD until a sample was found at 472 m. On XRD, smectite has peaks around 12.48-13.72 Å for untreated, 15.54-16.75 Å for glycolated and 9.75-9.94 Å for heated samples. Petrographical deposition of smectite is first seen as a thin lining in vesicles. Smectites form at temperatures below 200°C (Kristmannsdóttir, 1979).

Mixed layer clays identified in this well are mostly inter-layered smectite-chlorite. In the thin section, it has a vibrant brownish green colour and high pleocroism under plane polarized light. In the X-ray diffractometer analysis, the mixed layer clays are identified at 628 m. This type of clay has peaks at 14.12Å for untreated, expanding to 16.6 Å on glycolated and collapse to 9.78-14.12 Å on heating to 550°C.

Chlorite. Petrographically, chlorite is light greenish in colour in plane polarized light, has a fibrous cleavage and no pleochroism. It was also seen as an alteration of the primary rock constituents It was observed especially from XRD analysis starting from 674 m depth down to 1200 m. This mineral was identified by peaks at 14.04-14.32 Å. Some chlorite has peaks 7.07-7.13 Å for untreated, 7.07-7.13 Å for glycolated and collapses after heating to 550°C. Chlorite is characterized in the microscope by a fine- to coarse-grained radial texture in voids. It is a common hydrothermal alteration, particularly in prophylitic alteration. The occurrence of chlorite in geothermal areas generally indicates temperatures exceeding 230°C (Kristmannsdóttir, 1979).

4.3 Vein and vesicles fillings

The rocks encountered in the well are generally porous with a number of veins. Porosity can be classified into several types such as intergranular, joint and vesicular or vug type. Vesicular is common in Iceland where basaltic rocks predominate (Browne, 1984). These open voids become gradually filled with increasing alteration where e.g. limonite, siderite, and low-temperature zeolite are found in the upper part and mostly clay, calcite, quartz, wairakite and epidote at higher alteration. Hydrothermal mineral deposition is mostly found in vesicles and veins. In the well, voids are abundant in the hyaloclastites. A summary of the vein fillings in this well are shown in Table 2.

TABLE 2: The distribution of vein fillings in well HE-24

No.	Depth	Lithology	Degree of alteration	Vein fillings	Number of veins
1	~160-442	Glassy basalt, basalt breccias	No alteration	-	-
2	~478-516	Basaltic tuff	Slight	cc, qtz, cly	14
3	~560-668	Basaltic breccia, basaltic tuff	Moderate - high	cc, qtz, cly	31
4	~704-773	Basaltic tuff, basaltic breccias	Moderate - high	cc, qtz, cly, wai, chl, sph	39
5	~814-826	Medium- to coarse-grained basalt	Slight	cc, qtz, cly, wai, sph	18
6	~850-998	Medium- to coarse-grained basalt	No alteration	cc, qtz, cly, wai, sph	35
7	~1064 -1110	Glassy basalt	High	cc, cly	19
8	~1114-1172	Fine- to medium-grained basalt	Moderate - high	cc, qtz, cly, wai, epi, sph	49

Explanation: cc = calcite, qtz = quartz, cly = clay, wai = wairakite, sph = sphene, epi = epidote

4.4 Alteration mineral zonation

In geothermal areas, the study of altered basaltic rocks shows that the sequence of mineral assemblages relates to increased temperature and depth. The most common alteration minerals are the clay minerals. Other hydrothermal minerals present are silica, feldspar, calc-silicates, zeolites, carbonates, iron oxide, iron sulphides, sulphate, and sulphides (Browne, 1978). Below the hydrothermal alteration has been divided into temperature-dependent zones as practiced in Iceland:

Unaltered zone (0-244 m). The formations down to 244 m depth contain no alteration that is related to hydrothermal activity. XRD analysis shows hardly any indication of smectite and the only mineral precipitations are limonite and siderite, both of which relate more to cold groundwater conditions. The Skardsmýrarfjall formation belongs to this zone.

Smectite-zeolite zone (244-616 m). The upper boundary of this zone coincides with the first occurrence of zeolites (including thomsonite, chabazite, scolesite) at about 244 m depth. The XRD signature of smectite is still weak and remains so until below 446 m where it becomes stronger. Experience and data have confirmed that smectite forms below 200°C (Kristmannsdóttir, 1979).

Mixed layer clay zone (616-672 m). The upper boundary is set by the first analysis of mixed layered clay at 620 m depth and the lower boundary is determined by the first appearance of chlorite at 672 m. The temperature assessment of this zone is 200-230°C (Browne, 1978, Kristmannsdóttir, 1979). Petrographic evidence shows the mixed layered clays have high colours and are very pleochroic.

Chlorite zone (672-1074 m). The upper boundary of this zone is marked by the first appearance of chlorite in the XRD analysis at 672 m and the lower boundary is marked by the first appearance of epidote at 1074 m depth. Chlorite is identified petrographically as low-colour and non-pleochroic radial clays. With XRD-analysis chlorite is identified with peaks appearing at 14 and 7 Å, although the chlorite is considered unstable as the 7 Å peak collapses upon heating. Chlorite has been estimated as forming at a minimum temperature of 230°C (Browne, 1978; Franzson, 1987).

Chlorite-epidote zone (1074-1172 m). The upper boundary of this zone is characterized by the appearance of epidote. Other minerals in the zone include quartz, wollastonite and sometimes prehnite. The upper boundary of the zone is believed to conform to 240-250°C (Kristmannsdóttir, 1979).

Epidote-actinolite zone (1172 - >1200 m). The upper boundary of this zone is marked by the first appearance of actinolite. Actinolite was first identified from a petrographic thin section at 1172 m depth and with the binocular microscope at 1176 m. This mineral is mainly found as an alteration of pyroxene. Actinolite appears to form at a minimum of about 280°C (Kristmannsdóttir, 1979).

4.5 Mineral deposition sequence

The mineral sequences deposited from the geothermal system into vesicles and veins were studied petrographically. The results of that study are shown in Table 3.

The depositional minerals were found mostly in vesicles and veins. The alteration mineral assemblages change from low-temperature minerals such as zeolites to moderate- to high-temperature minerals with increasing depth, such as quartz, wairakite, prehnite, wollastonite and actinolite.

Clay and calcite are the most common minerals participating in the mineral sequence in this well. The fine-grained clay is mostly found as thin linings in the walls of vesicles and veins, associated or deposited after chalcedony, which is also found near the boundary of veins and vesicles. Coarse-grained clay is found especially as fillings in the veins or vesicles.

TABLE 3: Mineral depositional sequence of well HE-24

DEPTH	**Degree of alteration**	**Early**							**→ Later**
160	No alt	smec							
334	No alt	smec							
442	No alt	smec							
510	Slight	smec		zeo					cc
560	Moderate		fgc				cgc		cc
616	High	cc		zeo					cc
668	High	chal	fgc		qtz		cgc		
		chal	fgc						cc
704	High	chal	fgc	zeo			cgc	qtz	cc
747	Moderate				qtz	wai			cc
773	Moderate				qtz	wai			
816	Slight	chal			qtz				cc
850	No alt		fgc		qtz				cc
926	No alt		fgc		qtz				cc
998	No alt		fgc		qtz				
1064	High		fgc				cgc		cc
1114	High				qtz				cc
1172	High		preh		woll	act	wai		

Explanation: No alt = unaltered, moderate = moderate alteration, slight = slight alteration, high = high alteration, smec = smectite, zeo = zeolite, fgc = fine-grained clay, cgc = coarse-grained clay, cc = calcite, qtz = quartz, wai = wairakite, preh = prehnite, woll = wollastonite, act = actinolite.

The mineral depositional sequences observed with petrographic thin section analysis are shown in Table 3. In the upper part only fine-grained smectite is seen. At 510 m depth smectite is succeeded by zeolite and later by calcite. Further down, zeolites become unstable, and calcite and quartz replace them. Also, chalcedony, fine-grained clay and coarse-grained clay precipitate and are then superimposed by calcite. Fine-grained clay is succeeded by quartz and calcite below 850 m depth. In the mineral sequence, the deposition of calcite is mostly present at the end of the sequence. The presence of calcite at the end of the sequence may imply that the system is cooling. At the lower part, the precipitation of prehnite is succeeded by wollastonite, actinolite and wairakite.

4.6 Fluid inclusions

Fluid inclusions form during the growth of crystals, which are termed primary inclusions. Similarly it is common for crystals to be deformed after mineral precipitation is complete. Where deformation features, e.g. cracks, develop, they will be filled with the fluid present during or after deformation. Fluid may be trapped between the deformed surfaces by either subsequent precipitation or by dissolution-reprecipitation processes, leading to the formation of secondary fluid inclusions (Goldstein and Reynolds, 1994). Analysis of fluid inclusions, thus, makes it possible to elucidate temperature and fluid changes during and subsequent to crystal growth.

In well HE-24, quartz and calcite crystals were collected but only the calcite crystals were found to contain measurable fluid inclusions. A fluid inclusion study was conducted to assess whether the geothermal system is heating or cooling. The fluid inclusion study in well HE-24 was done with samples collected at a depth around 664-710 m. The homogenization temperature (Th) was identified from primary inclusions with a total number of 44 inclusions. The range of homogenization temperatures were found to range from 195 to 285°C (Figure 10). The fluid inclusions indicate that the calcite grew under a wide range of temperatures. The highest Th measured is up to 285°C indicating that the crystal deposited while the geothermal system was hotter than the present state, possibly boiling. As earlier mentioned, the homogenization temperatures in the calcite crystals exhibit a wide range in temperatures and indicate cooling of the system down to the current state at around 195°C at this depth in the reservoir.

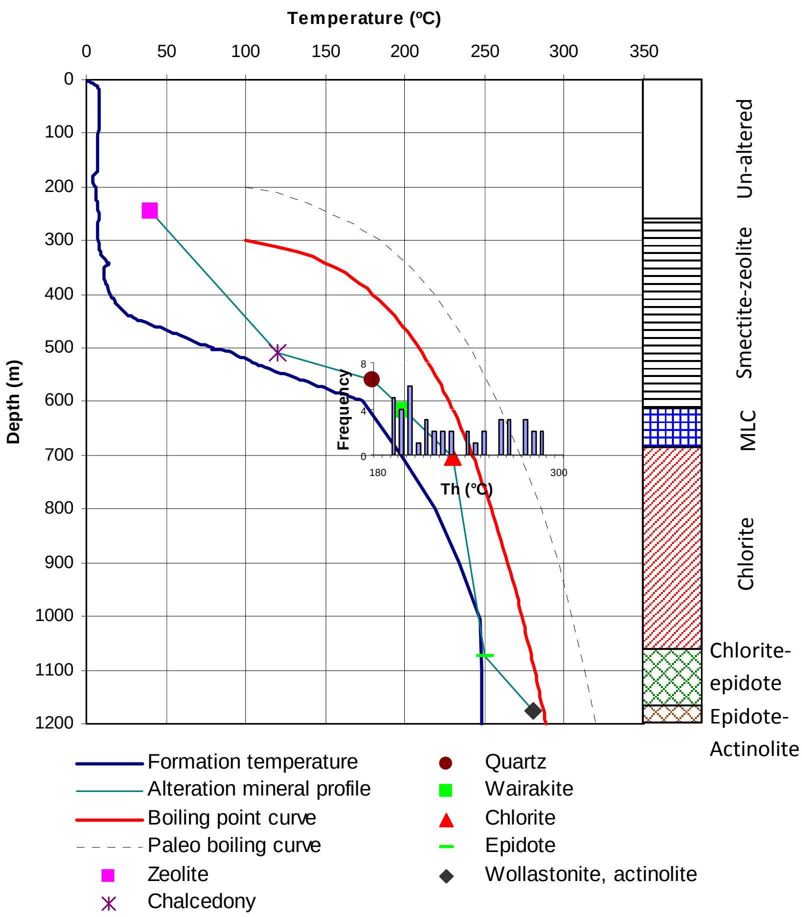

FIGURE 10: Profile of formation and alteration mineral temperature and fluid inclusions of

5. AQUIFERS

Aquifers or feed points in the wells are located using data such as circulation losses, temperature logs, hydrothermal alteration, well completion test and other data from drilling (Franzson et al., 2005; Browne, 1978). Data from drilling such as the rate of penetration and pump pressure may also be important. A high rate of penetration during drilling may indicate aquifers, whereas a drop in pipe pressure during drilling, especially while drilling during total loss of circulation, may also indicate opening of an aquifer. According to Reyes (1990), who studied permeability in the Philippine geothermal systems, the source of permeability in rock formations can be faults, intrusions and lithological contacts, joints, clast matrices or fragment contacts. Permeability, as a physical parameter, depends on the structure, shape, and occurrence of cavities or fractures and how well connected they are (Browne, 1978).

In well HE-24, twenty three aquifers (feed points) were identified using various methods such as circulation loss data, temperature logs, intensity of alteration and other geological data (Figure 11). These feed points were divided into three relative sizes of aquifers: five large, two moderate and sixteen small aquifers. Seven aquifers are located in the production part and sixteen aquifers are located above the production part. In this well, the aquifers are mainly located at stratigraphic boundaries above the production part, while they appear mainly to be related to intrusions in the production part.

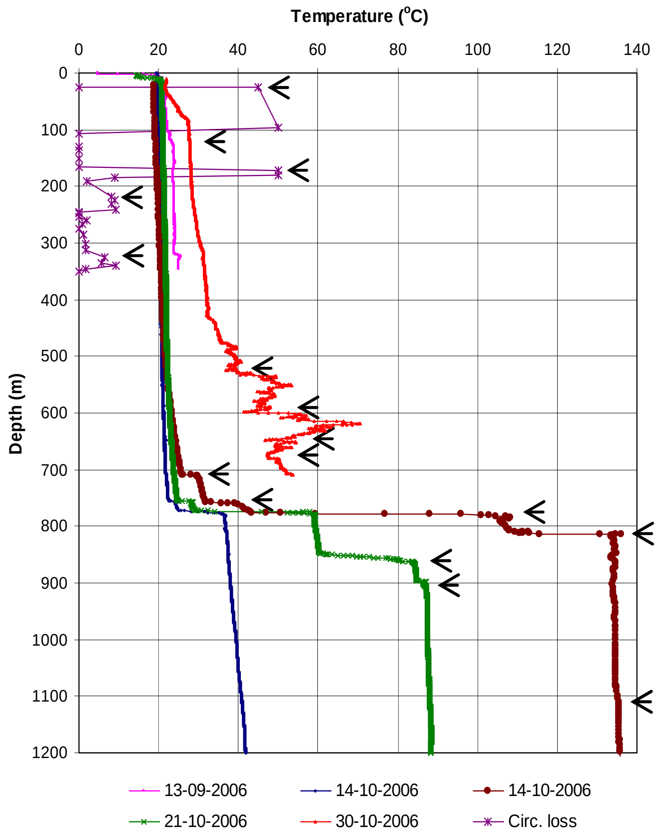

FIGURE 11: Loss of circulation and temperature log showing the locationof aquifers/ feed points in well HE-24

Aquifers above the production part. Around sixteen feed points were identified above the production part. Above 340 m, the feed points are associated with circulation losses with a range around 6, 9, 45 and 50 l/s, or total circulation loss. Circulation losses range from small, moderate to total circulation losses where a highly permeable zone is encountered. Some aquifers can also be confirmed by comparison of temperature logs. Three feed points at depths 26-96, 136 and 326 m were associated with circulation losses and were identified in the temperature logs.

In this part, from 340 down to 709 m depth the feed points are mainly marked by peaks/deflections in the temperature profile. The sizes of these deflections correspond to the relative size of the feed points. Deflections of the temperature profile at this depth are in the range of small peaks and small sized feed points and are mainly associated with lithological boundaries.

Aquifers in the production part. Here, the aquifers were identified from the measured temperature profile. The deflection in temperature shows small to large peaks and indicates that the relative size of the aquifers ranges from small to large. Around seven aquifers were found and divided into four small, one medium and two large aquifers. The feed points are predominantly at or near to the intrusion. No or slight alteration was observed at the feed points which are located within the intrusion, while the alteration is moderate to high at feed points above and below the intrusion. The distribution of aquifers correlated to the geology of the well is shown in Table 4.

TABLE 4: The location of aquifers (feed points) above 1200 m depth well HE-24

No.	Depth (m)	Circ. loss (l/s)	Temp. profil	Alteration intensity	Geological structure	Relative size
1.	26-96	45 – TLC	S	NA	Strat (?)	L
2.	136		S	NA	Unk	S
3.	174-182	50 – TLC		NA	Strat (?)	L
4.	185	9		NA	Strat (?)	S
5.	198-219	TLC		NA	Strat (?)	L
6.	225	9.1		NA	Unk	S
7.	230	8.2		NA	Unk	S
8.	241	9.16		NA	Strat	S
9.	326	6.4	S	NA	Unk	M
10.	335	5.56		NA	Unk	S
11.	340	9.16		NA	Unk	S
12.	516		S	SA	Strat	S
13.	598		S	MA	Unk	S
14.	648		S	HA	Strat	S
15.	673		S	HA	Strat/Frac	S
16.	709		S	HA	Unk	S
17.	758-778		S	MA	Strat/Int/Frac	S
18.	780		M	MA	Int	M
19.	783		L	MA	Int	L
20.	814		L	SA	Int	L
21.	865		S	NA	Int	S
22.	907		S	NA	Int	S
23.	1113		S	HA	Unk	S

Explanation : Circ.=Circulation, Temp.=Temperature, TLC = Total loss circulation, S=Small, M=Moderate, L=Large, NA=No alteration, SA=Slight alteration, MA=Moderate alteration, HA=Highly alteration, Strat=Stratification boundary, Int=Intrusion, Frac=Fracture, Unk=Unknown relation

6. DISCUSSION

Generally, the stratigraphy of the upper 1200 m of well HE-24 consists of hyaloclastites and basaltic intrusions. The volcanic sequence is divided into five units based on the textural differences and the intensity of alteration. The distinction of the volcanic sequences in the well is predominantly based on whether they are porphyritic or aphyric. Four hyaloclastite units (separate sub-glacial eruptions) and one basalt unit and intrusion were identified in the well.

In well HE-24, the geological and hydrothermal alteration study shows that the degree and the intensity of rock alteration and the distribution of mineral alteration increase with depth. Below about 478 m depth, the degree of alteration increases rapidly both in temperature dependent minerals and alteration intensity.

Temperature has been defined in two ways, in well HE-24. Hydrothermal alteration mineral temperature was assessed according to the first appearance of the hydrothermal alteration minerals. Formation temperature was determined by calculations based on the temperature measurements during the heating-up period. The alteration temperature curve of well HE-24 shows a progressive temperature increase with depth. The alteration mineral assemblage shows a trend, where low-temperature minerals like zeolites form in the upper part of the well and are gradually replaced by moderate-temperature minerals like chalcedony, quartz and wairakite which in turn give way to higher-temperature mineral assemblages like chlorite, prehnite, epidote, wollastonite, actinolite and garnet in the lower part of well HE-24. The correlation of hydrothermal alteration, formation and boiling point temperature curve is shown in Figure 10.

The mineral deposition sequence shows that low-temperature zeolites form in the early stages of the sequence. In the middle of the sequence, moderate-temperature minerals such as quartz and wairakite are deposited in the veins and vesicles. In the lower part of the well high-temperature minerals such as wollastonite and actinolite precipitate. The clay minerals are sensitive to changes in temperature. In the well, clays were found to become more crystalline with depth. In petrographic thin sections, the clay minerals deposited in vesicles consisted of two types: either fine-grained clay which usually is found as a thin layer lining the voids and vesicles, and coarse-grained clay usually as a chlorite and mixed layer clay, similarly deposited in the veins and vesicles. In well HE-24, calcite is predominantly deposited in the last stage of the mineral sequences in the well. Data from the Hellisheidi geothermal field indicate that the last deposition of calcite was associated with cooling in the later stages of the geothermal system (Franzson, 2000).

The aquifers in the production part mostly relate to a basaltic intrusion. Feed points below 758 m are believed to be associated with a vertical dyke, possibly the feeder to the 5000 years old eruption. The rock, where the feed points appear, is heavily oxidized and is interpreted as a contact aureole adjacent to the dyke. Below this, the aquifer is located within the fresh intrusion body.

Besides well HE-24, two other wells were drilled in Skardsmýrarfjall mountain, namely HE-23 and HE-25. (Figure 12). HE-23 is a directional well drilled to a total depth of 1968 m in a NE direction. Well HE-25 is also directional with a SW direction and reached a total depth of 2155 m. The low-temperature zeolite minerals in the three wells are found at a similar shallow depth. The minerals are used to determine the upper boundary of the smectite zone. Quartz is largely found at the same depth in the three wells and is located in the lowermost part of the smectite zone. On the other hand, wairakite appears fairly low, that is to say below the mixed

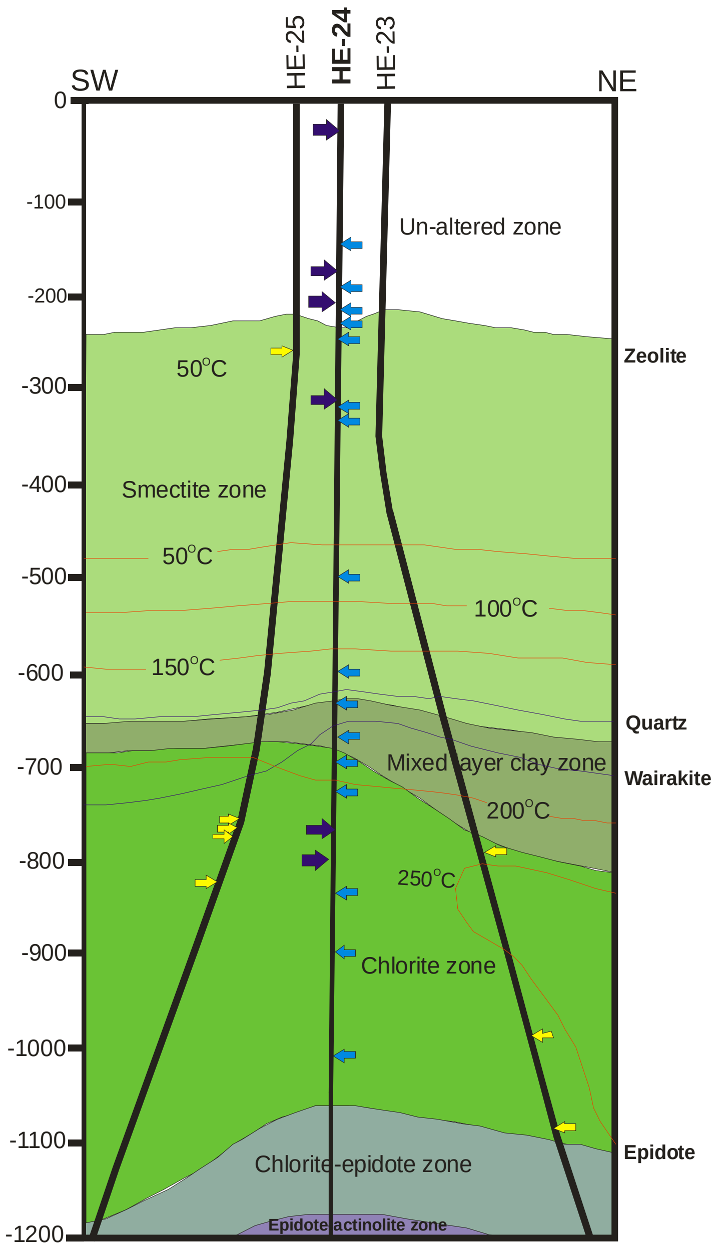

FIGURE 12: Correlation of the hydrothermal alteration of the three wells at Skardsmýrarfjall

layer clay zone in well HE-25. The first appearance of this mineral is within the mixed layer clay zone of wells HE-23 and 24. The mixed layer clay and chlorite zones were determined by XRD analysis in wells HE-23 and 25, but combined with a study of petrographic thin sections in well HE-24. The mixed layer clay zone becomes thinner towards the southwest, while the chlorite zone is fairly thick in the wells at Skardsmýrarfjall. The thick chlorite zone suggests that the alteration gradient decreases at around 750-1100 m depth in this part of the reservoir. In the lowermost part, the first appearance of the high-temperature minerals, epidote and actinolite, marks a noticeable increase in alteration and was used to determine the boundary of the chlorite-epidote and epidote-actinolite alteration zones. Actinolite appears shallower in well HE-24 than in the other two wells. The correlation of the alteration minerals is shown in Figure 12.

The formation temperature shows that the temperature increases rapidly from 50 to 200°C at 480-650 m depth in the three wells. At 660 m down to 760 m depth, the temperature appears to be higher towards the NE in well HE-23, where the temperature even reaches the boiling point 250°C at a depth of 790 m. In all three wells, the temperature is very close to 250°C at 1000-1300 m depth, but below 1300 m depth the temperature decreases. In well HE-24, the temperature is reversed again at 1750 m, and at a depth of 2250 m a temperature of 250°C is reached.

The three wells were drilled through the Skardsmýrarfjall formation. Drilling has confirmed the presence of a relatively cold groundwater system in the upper approximately 250 m of the formation. The formation is characterized by fresh hyaloclastites and pillow lavas that are very porous and permeable. A correlation of alteration mineral temperatures and measured temperature studies indicates that at 250-700 m depth, through the cap rock zone of the geothermal reservoir, the system has cooled significantly. On the other hand, at approximately 750-1200 m, especially in well HE-24, depth cooling appears to be minor in the reservoir. Indications of cooling were also found in the lower part of the three wells, characterized by decreasing temperature especially around 1600 m. This is probably caused by cold groundwater influx, either along fractures or faults, or the high permeability of the reservoir. Discharge tests show that the temperature continues to decrease in the lower parts of the wells. This is probably caused by cold water influx from beneath, which influences the reservoir in the upper part causing cooling. This defines a reservoir of fairly limited vertical extent which appears to be associated with a fissure eruption 5000 years in age. The temperature profile is shown in Figure 13.

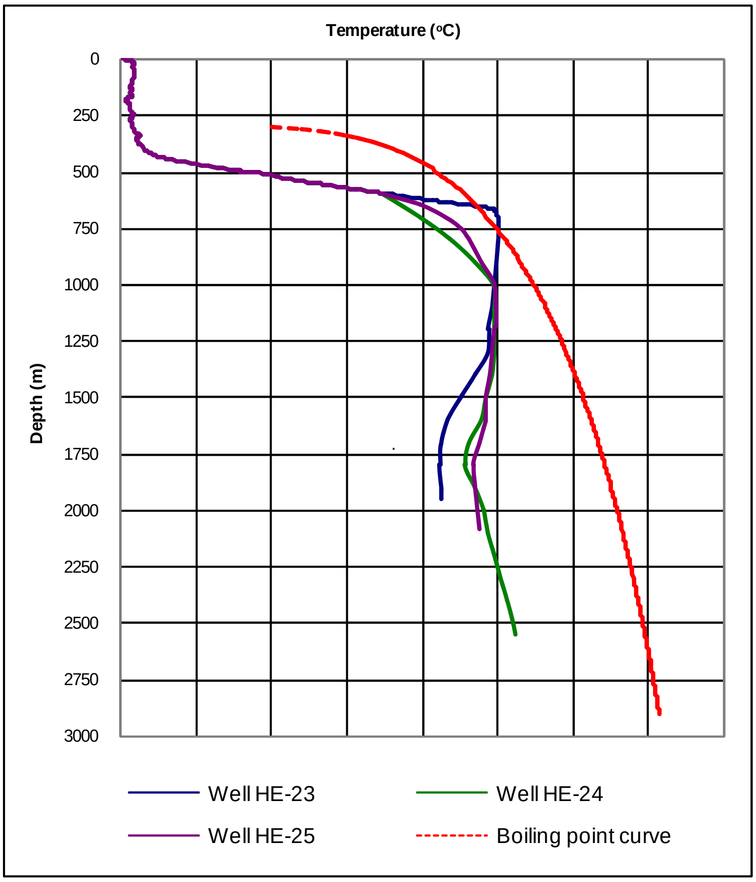

FIGURE 13: Comparison of the measured temperature of the three wells at Skardsmýrarfjall

7. CONCLUSIONS

The following conclusions can be deduced:

- Stratigraphy of the first 1200 m of well HE-24 consists of sub-glacial hyaloclastite formations and basaltic intrusion.
- According to the distribution of alteration minerals, an un-altered zone and five alteration zones were identified: Un-altered zone, smectite-zeolite zone (<200°C), mixed layer zone clay (200-230°C), chlorite zone (230-250°C), chlorite-epidote zone (>250°C) and epidote-actinolite zone (>280°C).
- According to circulation loss data, temperature logs, and the intensity of alteration, three relative aquifers sizes were distinguished giving five large aquifers, two moderate ones and sixteen small aquifers. Seven aquifers are located in the production part and sixteen aquifers are located above the production part.
- By studying the hydrothermal alteration minerals, it was found that the temperature rises rapidly at about 472-672 m depth with the appearance of zeolites, quartz, wairakite and mixed layered clays, while the temperature gradient is lower in the fairly thick chlorite zone at 672-1072 m depth. Below 1072 m depth the alteration increases again with the deposition of epidote, wollastonite, actinolite and garnet.
- Comparison of alteration mineral temperature and measured temperature shows that the geothermal system has cooled in the upper 700-800 m of the well, while cooling appears to be minor at 750- 1200 m depth. In the lowermost parts of the wells, especially below 1600 m, the reservoir appears to be cooling again.
- Correlation of hydrothermal alteration minerals shows that the temperature gradient, according to the alteration minerals, is similar in all three wells, HE-23, HE-24 and HE-25, at Skardsmýrarfjall.

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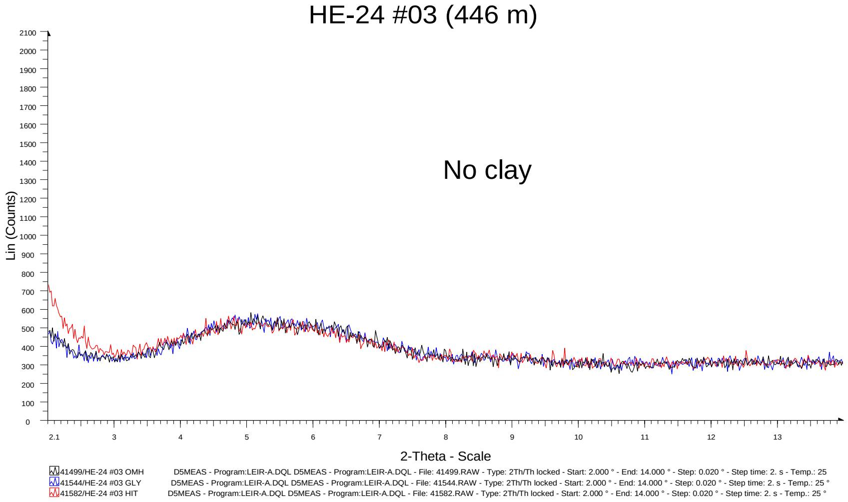
APPENDIX I: Results of the XRD analysis of clay minerals

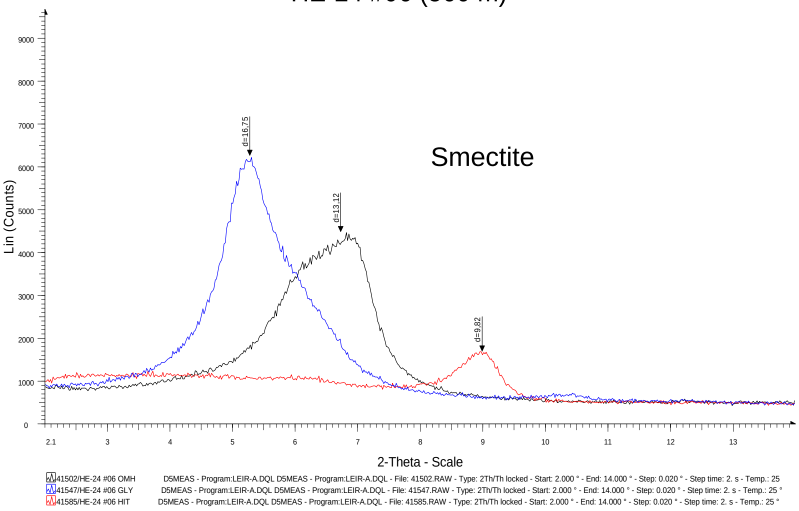
Depth	Untreated (Å)	Glycolated (Å)	Heated (Å)	Type of clay
330	13.68		9.95	Smectite (?)
402	13.68			No clay
446				No clay
480	13.72	16.64	9.94	Smectite
520	13.04	15.54	9.9	Smectite
560	13.12	16.75	9.82	Smectite
602	12.48	16.42	9.75	Smectite
628	14.12	16.6	14.12 (9.78)	Smectite-unstable chlorite
	7.17	7.17		
674	14.22	14.22	14.22	Unstable chlorite
	7.09	7.09		
702	14.32	14.32	14.32	Unstable chlorite
	7.13	7.13		
745	14.4	14.4	14.4	Unstable chlorite
	7.13	7.13		
788	14.04	14.04	14.04	Unstable chlorite
	7.08	7.08		
820	14.03	14.03	14.03	Unstable chlorite
	7.08	7.08		
1068	14.04	14.04	14.04	Unstable chlorite
	7.07	7.07		
1116	14.18	14.18	14.08	Unstable chlorite
	7.08	7.08		
1172	14.08	14.08	14.08	Unstable chlorite
	7.07	7.07		

APPENDIX II: Characteristic XRD patterns for the clay minerals of well HE-24

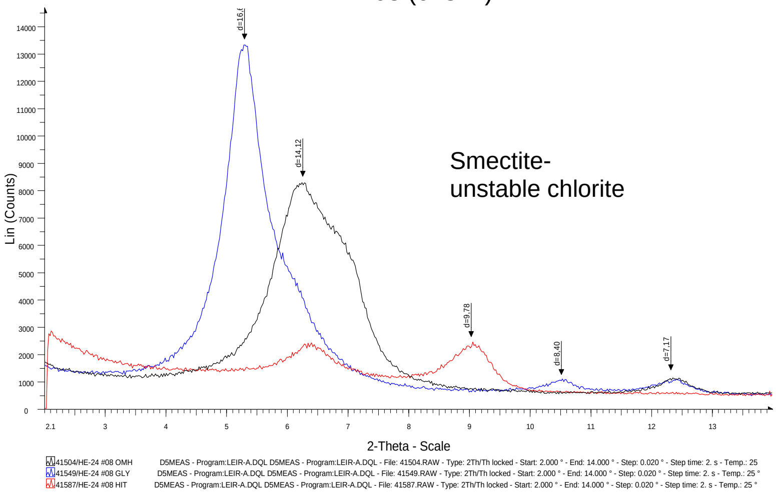
HE-24 #03 (446 m)



HE-24 #06 (560 m)



HE-24 #08 (628 m)



HE-24 #11 (745 m)