



BOREHOLE GEOLOGY AND HYDROTHERMAL MINERAL ALTERATION OF WELL HN-3, HELLISHEIDI GEOTHERMAL FIELD, SW-ICELAND

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ABSTRACT

Well HN-3 is located in the Hellisheidi high-temperature field, within the Hengill geothermal area in SW-Iceland. It is a directional reinjection well reaching a total depth of 1864 m. The uppermost 1272 m are presented in this report. A binocular microscope, a petrographic microscope, XRD-analysis, a fluid inclusion analysis and geophysical logs were used to gather data. The lithology of well HN-3 comprises basaltic hyaloclastite formations, tuffaceous sediment and fine- to medium-grained olivine tholeiitic and tholeiitic basalt. The rock types within the hyaloclastite units are: partially crystallized basalts, basaltic breccia, and basaltic tuff. There are also basaltic intrusions at depths of 1026-1032 m and 1234-1244 m. Four alteration zones were identified in this well: a zone of no alteration at 0-92 m depth, a zeolite-smectite zone (<200°C) from 92 to 972 m, a mixed-layer clay zone (200-230°C) from 972 to 1236 m and a chlorite zone (>230°C) from 1236 to 1272 m. Seven very small possible aquifers were identified above the production part of the well and six small aquifers at 440, 550, 660, 725, 1180 and 1245 m depth by using temperature, neutron-neutron, gamma, resistivity and calliper logs. Fluid inclusions studied show a wide range of homogenization temperatures which range from 180 to 330°C. Fluid inclusion temperatures, at 1200-1270 m, show higher values than the present formation temperature, indicating that the well is cooling at that depth.

1. INTRODUCTION

1.1 Geology of Iceland

Iceland is the third largest island in the Atlantic Ocean, with an area of 103,106 km² and an average altitude of around 500 m (Jakobsson, 1979; Saemundsson, 1979). Geographically, it is located in the northern part of the Mid-Atlantic Ridge; the rift zone along the constructive boundary between the American and the Eurasian tectonic plates which move apart at an average rate of 2 cm per year. Iceland resides on a mantle plume and a hot spot in the rift zone and has formed in frequent volcanic eruptions from Miocene time to present (Figure 1). This explains why this part of the ridge rises above sea level and forms an island. The presently active zone of rifting and volcanism crosses Iceland from southwest northeast with volcanic eruptions, witnessed every few years. Most of the eruptions are effusions of

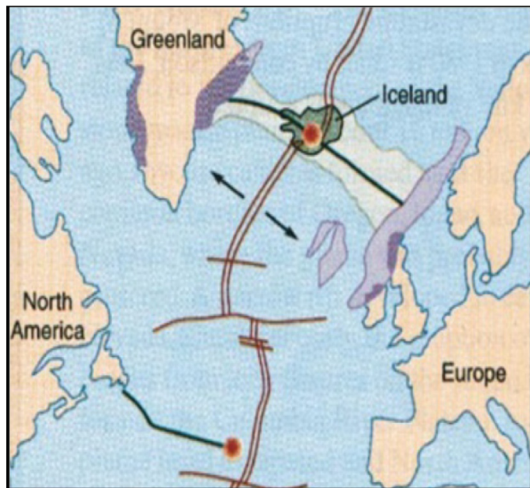


FIGURE 1: Location of Iceland; the orange circle indicates the location of the Icelandic plume (taken from Njue, 2010)

basaltic lava from fissures or shield volcanoes. Each fissure represents a small amount of crustal extension to perhaps about one metre on average (Saemundsson, 1979). In geological terms, Iceland is a young country, little eroded and dominated by volcanic rocks. About 10% of the country is covered with glaciers, some of which are underlain by active volcanoes (Figure 2). Crustal spreading has resulted in older rocks being found to the east and west of the island, while the youngest rocks are found in the centre (Saemundsson, 1979).

Iceland is composed mainly of igneous rock; sedimentary rock accounts for only 8-10% of Iceland's volume, and there is no true metamorphic rock on the island (Saemundsson, 1979; Jakobsson et al., 2008). A total of 25 types of igneous rock have been found in Iceland, the most common of which are tholeiite, olivine tholeiite, gabbro and rhyolite. The main rock-forming minerals in the tholeiite and olivine tholeiite are plagioclase, augite, olivine, and magnetite. Icelandic igneous rock forms three rock series: the tholeiitic series, the transitional series and the alkali series (Jakobsson et al., 2008).

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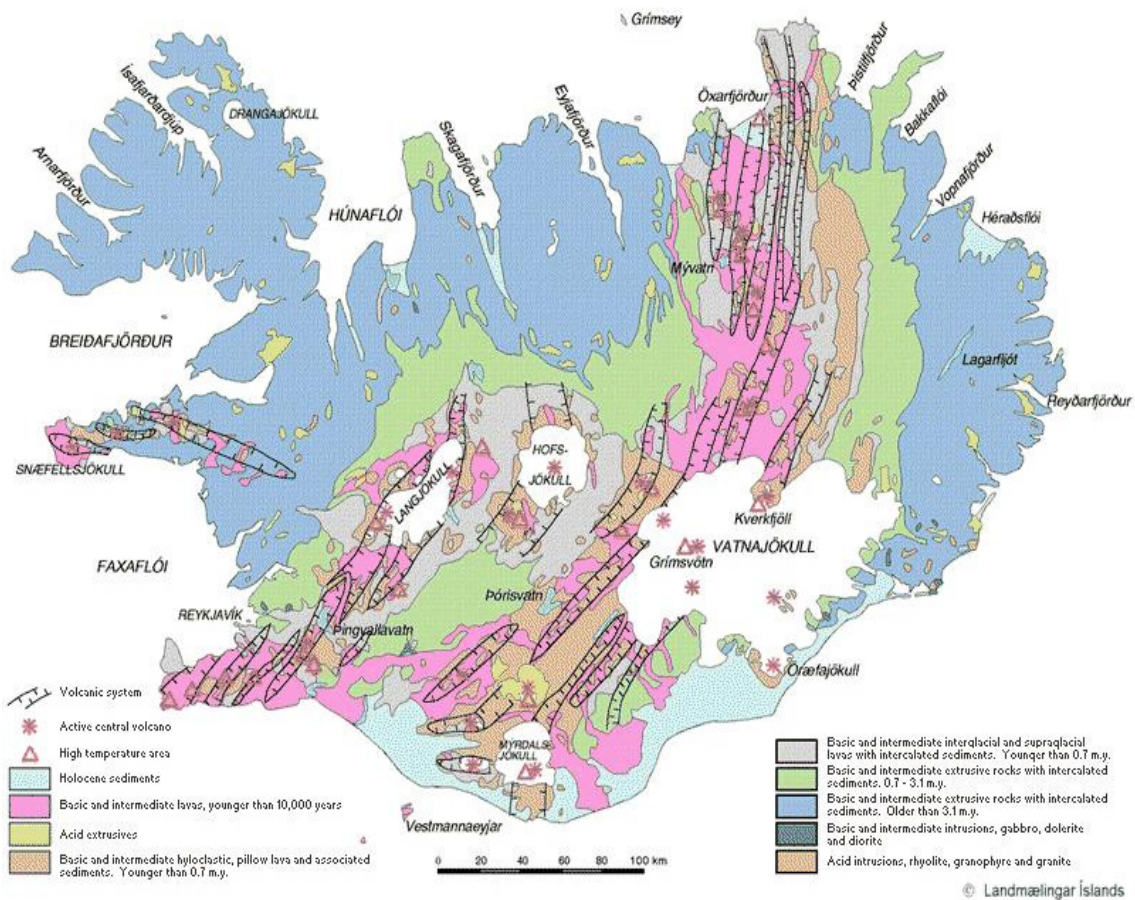


FIGURE 2: Geological map of Iceland showing oldest Tertiary rocks, older Plio-Pleistocene eruptives, Holocene rocks plus other young formations; volcanic systems follow the oceanic ridge (Jóhannesson and Saemundsson, 1999)

The tholeiitic rock series is confined to volcanic systems in the rift zones which delineate the crust of the Mid-Atlantic Ridge. The alkali and transitional series are confined to the volcanic system in the flank zones (Jakobsson et al., 2008). Geologically, Iceland can be divided into three formations that, although predominantly made of basalt, are distinguished by very different land forms: the Tertiary, Pleistocene, and Holocene. Tertiary flood basalt occupies the northwest, northern and eastern parts of the country (Figure 2). The successions of flat lying basalt, which generally dip towards the respective centres of origin, bear witness to rather flat and uneventful landscape in the Tertiary. Towards the coasts, the basalt plateau has been deeply dissected by the Pleistocene glaciers, and subsequent isostatic uplift resulted in the mountainous landscape characterizing these regions. The monotonous basalt successions are interspersed with extinct central volcanoes but, in general, the individual layers of basalt represent large flows that reached far outside the volcanic zone of origin, as evidenced by the observation that the mean difference in age between two adjacent flows varies between regions from 4,000 up to 10,000 years (e.g. Watkins and Walker, 1977; McDougall et al., 1977, 1984; Saemundsson et al., 1980), that of the entire Holocene. In the Pleistocene, alternating glacial and interglacial periods resulted in the formation of mountains made up of palagonitic tuffs occasionally capped by subaerial lavas, on the one hand, and thick valley-filling basaltic flows on the other. The palagonitic mountains now exposed at the surface, derive especially from the last two glaciations (last ~0.2 m.y.). One effect of the Pleistocene was to transform Iceland from the gentle landforms of the Tertiary to the mountainous and rugged landscape of the present: Glacial erosion gouged out deep valleys and fjords towards the coasts while subglacial volcanoes created palagonite hills and ridges along the volcanic zones.

Volcanic products in the Holocene are predominantly lava flows, although special circumstances (such as silicic composition, high groundwater table, or overlying glacier ice) may locally lead to explosive volcanism. Certain distinctive volcanic formations appear to be confined to the uppermost Pleistocene and early Holocene, tuyas (table mountains) and lava shields of primitive composition. These formations represent, in all probability, rapid and extensive magma production associated with the isostatic rebound at the end of glaciation (Sigvaldason et al., 1992) and, since then, the rate of volcanism appears to have decreased exponentially, the present productivity being but 1/30th of that of 10,000 years ago (Sigvaldason et al., 1992). The stratigraphy of the rocks with regard to age is also divided into four groups: Tertiary (Mio-Pliocene), older than 3.1 m.y.; Plio-Pleistocene, 3.1-0.7 m.y.; Upper-Pleistocene, 700,000-11,000 years; and Postglacial, 11,000 years and younger (Fridleifsson, 1983).

1.2 Geothermal fields

Iceland is rich in geothermal resources, as it is situated in an active zone of rifting and volcanism. Many volcanoes are located within the active volcanic zone stretching through the country from the southwest to the northeast, and at least 30 of them have erupted since the country was settled. In the volcanic zone there are at least 20 high-temperature areas containing steam fields with underground temperatures reaching 200°C within 1,000 m depth. High-temperature areas are confined to the active volcanic zones (Figure 3). They are mostly located on high ground and geologically they belong to very young and permeable rocks. As a result of the topography and high bedrock permeability, the groundwater table in the high-temperature areas is generally deep, and surface manifestations are

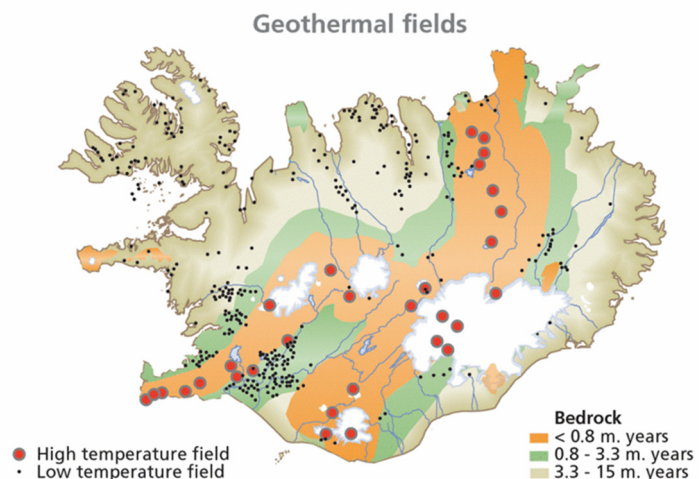


FIGURE 3: High- and low-temperature fields and bedrock age in Iceland (Orkustofnun database)

largely steam vents. The system's heat source is generally shallow magma intrusions (Björnsson et al., 2010).

About 250 separate low-temperature areas, with temperatures not exceeding 150°C in the uppermost 1,000 m, are found mostly in the areas flanking the active zone (Figure 3). The largest of these systems are located in southwest Iceland on the flanks of the western volcanic zone, but smaller systems can be found throughout the country. On the surface, low-temperature activity is manifested in hot or boiling springs, while no surface manifestations are observed on top of some such systems (Björnsson et al., 2010). To date, over 600 hot springs (temperature over 20°C) have been located (Figure 3).

1.3 The Hengill high-temperature geothermal area

The Hengill high-temperature geothermal area is located within the volcanic rift zone in SW-Iceland, 40 km east of Reykjavík. The geology is characterized by the active Hengill central volcano, and its fracture zone which has a NW-SW direction (Saemundsson, 1967; Franzson et al., 2005). This is an

extensive high-temperature area containing several economically promising geothermal prospects.

An extensive regional reconnaissance study was made of the whole Hengill area in the late seventies and early eighties. A short review and preliminary analysis of some of the data was presented by Björnsson and Hersir (1981). Hellisheidi is one of the geothermal areas within the Hengill system, located in the southern part.

The Hellisheidi power plant, situated in the southwest part of the Hengill area, was commissioned in 2006. It utilizes the geothermal activity in the southern part of the SW-NE fracture zone which goes through the Hengill area. The power plant produces electricity (303 MWe) and hot water for space heating using a water-steam mixture from 34 wells in the area (Gunnarsson, 2011). The Gráuhnúkar area was planned as the reinjection zone of the power plant. The high formation temperature in the Gráuhnúkar area is comparable to the hottest parts of the Hellisheidi field (Figure 4) making it promising for production (Gunnarsson, 2011).

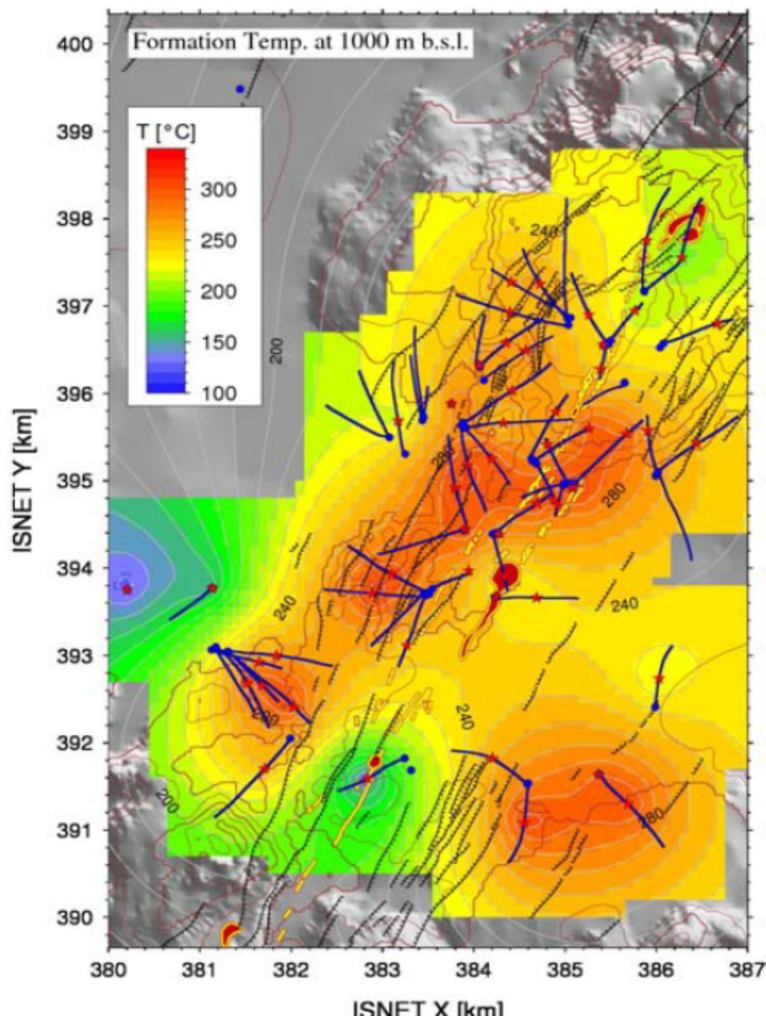


FIGURE 4: Formation temperature at a depth of 1000 m below sea level in the Hellisheidi field; well heads are depicted as blue points, directionally drilled wells as blue lines, and the data points (the intersection of the wells to 1000 m b.s.l.) are depicted as red stars; formation temperature between data points is calculated using the surface interpolation scheme of the GMT software package (Wessel and Smith, 1998)

1.4 Geological and tectonic setting of the Hengill area

The Hengill area is located on the eastern border of the Reykjanes Peninsula, SW-Iceland (Figure 5). It is set at the continuation of the Mid-Atlantic Ridge (Reykjanes Ridge) in Iceland, at the triple junction of the Reykjanes Peninsula volcanic zone, the Western volcanic zone and the South Iceland seismic zone. The Hengill volcanic system has a 100 km long NE-SW axis and is 3-16 km wide, and is almost entirely built up of volcanic rocks. Subglacially formed hyaloclastites together with pillow basalts constitute the main rock types in the area. Second in extent are Pleistocene and Postglacial lava flows (Saemundsson, 1967).

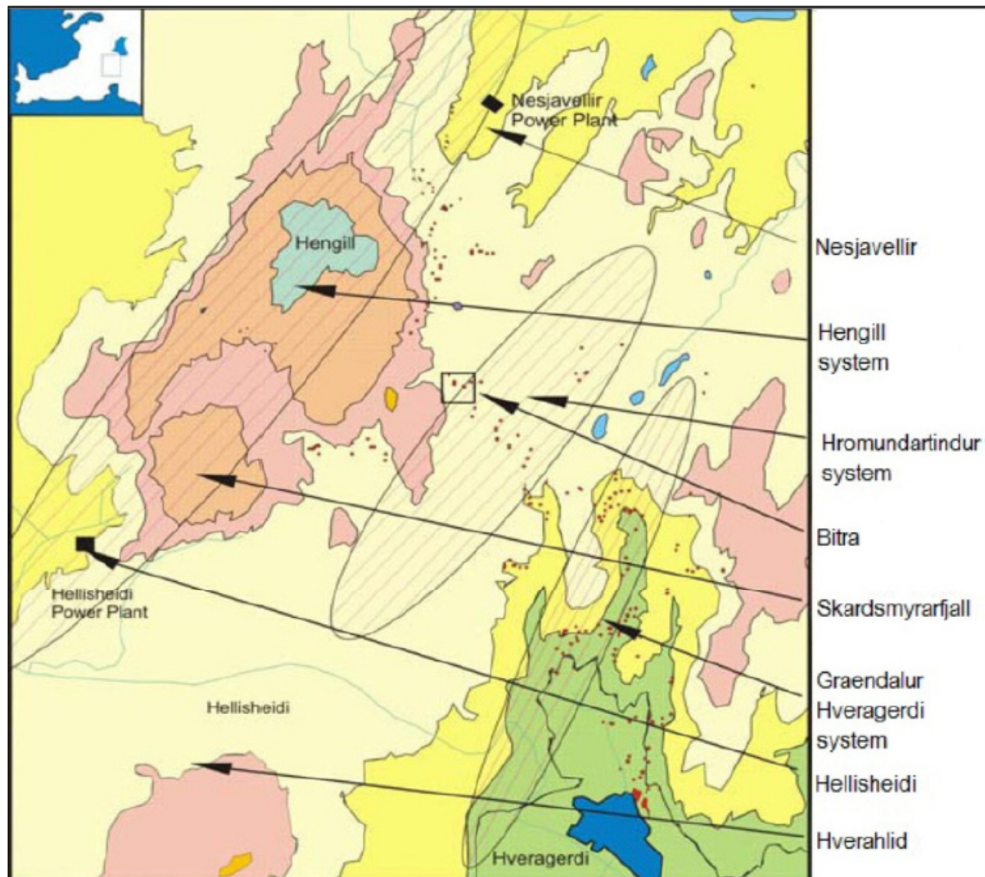


FIGURE 5: Location map of the Hengill volcanic complex (from Pendon, 2006)

The Hengill volcanic area is commonly separated into three volcanic systems; the youngest and most active one is the Hengill system itself, centred in the west and located within the axial rift zone. The second youngest is the Hrómundartindur system, further to the east and much smaller in distribution. The third one, the Hveragerdi system, furthest to the east, is now extinct and was considerably eroded in Pleistocene time (Figure 5). The oldest rocks, about 0.8 my old from the Matuyama epoch, are located in the lowlands southeast of the town of Hveragerdi, and the youngest are the Holocene lava flows from the fissure swarm cutting the Hengill volcano in the west. The Hrómundartindur system also includes an early Holocene lava flow from the Tjarnarhnúkur volcanic cone, extending into western Ölkelduháls (Saemundsson, 1967). An age of about 0.4 million years is proposed for the Hengill central volcano which puts a lower age limit on the geothermal system (Franzson et al., 2005). Tectonically, Hengill is the easternmost of a series of four closely spaced basaltic fissure systems that cut diagonally across the Reykjanes Peninsula. It is traversed by a graben about 10 km broad which runs NE-SW parallel to the hyaloclastite ridges. This graben is part of a greater structure which accompanies the Reykjanes-Langiökull volcanic zone. The western part of the Hengill area is split up by numerous sub parallel normal faults. These constitute a 5 km broad inner graben of intense faulting and fissure volcanism. During Postglacial time, six fissure eruptions occurred within this graben, four south of and two north

of the mountain (Franzson et al., 2010). Faults and major fractures strike mostly NE-SW and are most conspicuous in the east and west, marking the boundaries of the fault and fissure zones of the volcano. Postglacial volcanism includes three fissure eruptions of ages 9000, 5000 and 2000 years. The two younger NE-SW volcanic fissures are believed to provide an inflow of colder water towards the reservoir of the main geothermal system (Franzson et al., 2005).

1.5 Geophysical studies in the Hengill area

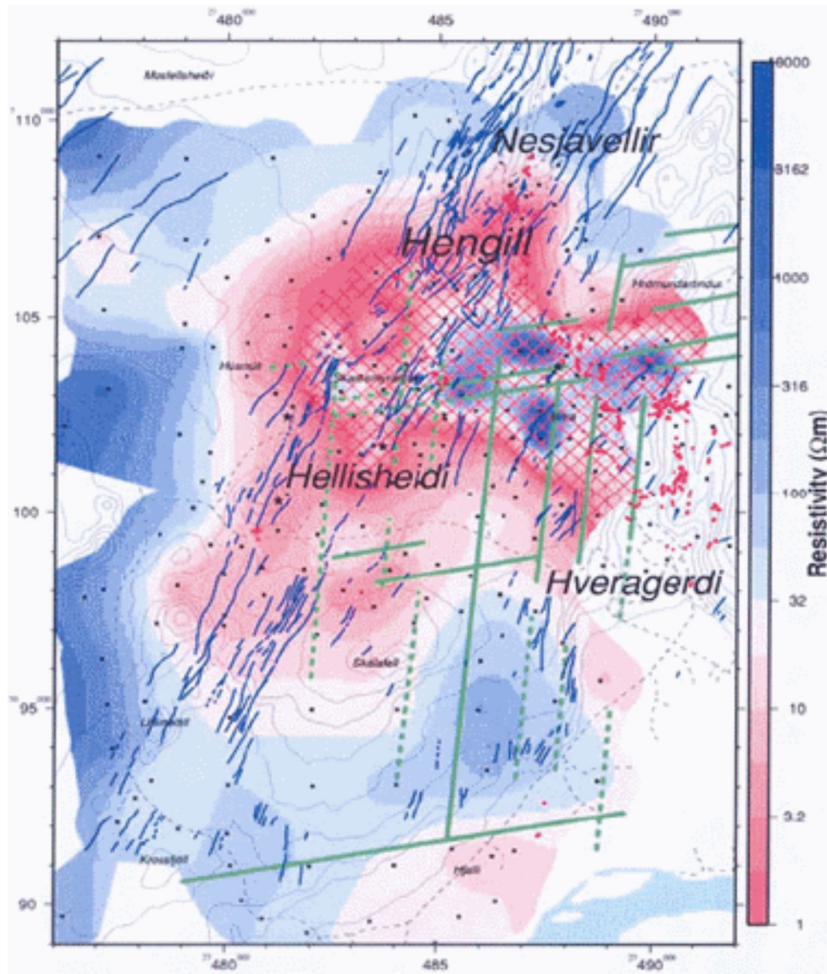


FIGURE 6: Hengill area, resistivity at 100 m b.s.l. according to a recent TEM survey, faults (blue/irregular lines) and interpreted earthquake fractures from recent seismic events (green/straight lines) (Árnason and Magnússon, 2001)

Extensive resistivity surveys, such as Schlumberger and TEM resistivity measurements (Björnsson et al., 1986; Árnason and Magnússon, 2001), have been conducted in the Hengill region, including the Hellisheidi geothermal field. Aeromagnetic and gravity surveys have, furthermore, been done in the Hengill geothermal area, including Hellisheidi (Björnsson et al., 1986). The resistivity surveys delineated a 110 km² low-resistivity area at 200 m b.s.l., and the magnetic survey showed a negative and transverse magnetic anomaly coherent with the most thermally active grounds. Recent transient electromagnetic soundings (TEM) have led to a revision of the resistivity map (Figure 6). The electrical resistivity also indicates roughly the extent of the high-temperature fields. All surface manifestations, like fumaroles and altered ground, are within this area (Hersir et al., 1990). The main resistivity feature is a high-resistivity zone (50-500 Ωm) which is observed beneath the low-resistivity layer. A correlation between resistivity, rock temperature and alteration at Nesjavellir (northeast of Hellisheidi) shows high-resistivity values close to the surface which can be attributed to fresh unaltered rocks. The low-resistivity values (1-5 Ωm) are connected with the smectite-zeolite alteration belt at temperatures between 50 and 200°C. Below the low resistivity is the high-resistivity core or layer, mentioned above, and associated with a high-temperature alteration zone. This high-resistivity core is related to the chlorite-epidote zone, located under a chlorite zone, indicating temperatures of more than 240°C (Árnason et al., 2000).

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2. BOREHOLE GEOLOGY

Gráuhnúkar area was planned for the reinjection of excess water from the Hellisheidi power plant. It is located southwest of the power plant, north of Gráuhnúkar peaks (Figures 7 and 8). Gráuhnúkar peaks form the northeast segment of a hyaloclastite ridge. On the surface the formation consists mainly of fine-grained, layered tuff, but fractured interglacial lava can be found in places on top of the tuff. It is believed that the formation was erupted from a relatively short volcanic fissure early during the last glacial period during a short period of warmer climate, possibly some 70,000 years ago (Saemundsson et al., 1990). This is supported by the fact that lava flows are found at only 400 m above sea-level, showing that the eruptive product melted their way up through the glacier at that altitude.

The rocks of Gráuhnúkar belong mostly to the olivine tholeiite and olivine basalt series. Postglacial lavas surrounding the Gráuhnúkar formation date between 5500 and 2000 years (Saemundsson, 1995). Geothermal alteration is not seen at the surface at Gráuhnúkar but some is found to the east, for example at Hveradalir.

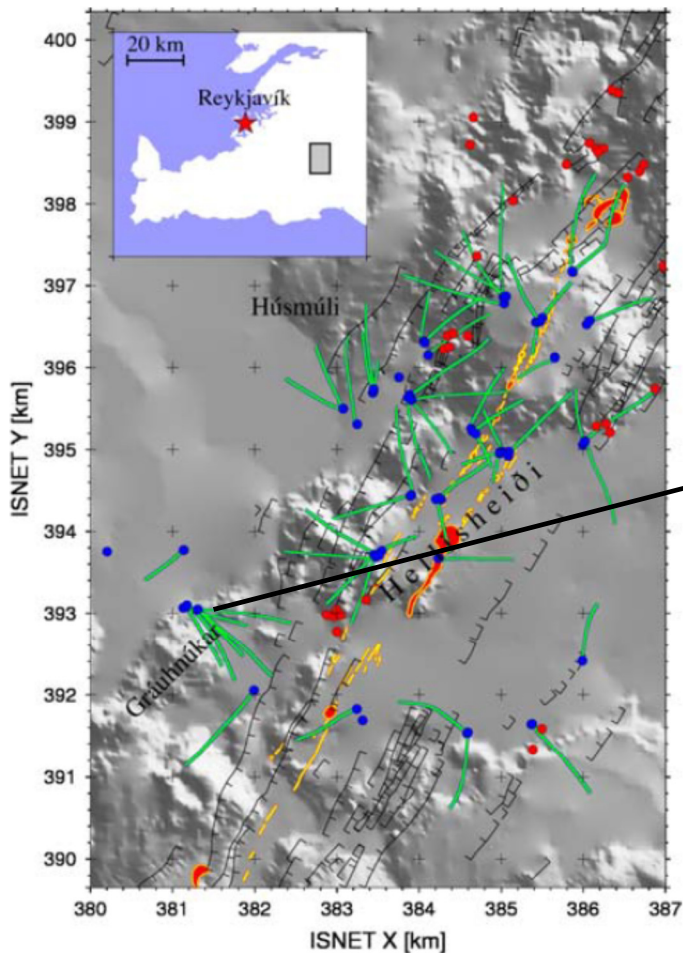


FIGURE 7: Map of the Hellisheidi geothermal field; the inset shows the location of the field in SW-Iceland; well heads are depicted with blue dots, tracks of directionally drilled wells with green lines, hot springs and fumaroles with red dots, faults with combed lines, and volcanic fissures and craters with yellow/red areas

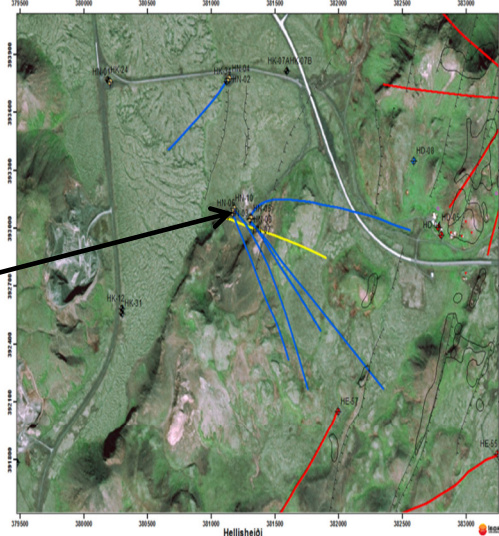


FIGURE 8: Aerial photo of well HN-3, marked with a yellow line; blue lines are other reinjection wells; red lines are production or exploration wells

The reinjection wells at Gráuhnúkar are all directional and trend towards the east (Figures 7 and 8). They are located west of the extensive NE-SW faults that mark the western boundary of the Hengill volcanic system but these faults are quite permeable. The reinjection wells cut the faults and thereby extend into the Hengill geothermal system. E-W trending fissures, probably related to the South Iceland seismic zone, which cut the NE-SW faults, may increase the permeability of the reinjection area (Hardarson et al., 2007). As the reinjection area was developed, it was discovered that the temperatures in the wells ranged between 270 and 305°C. Consequently, Reykjavik Energy has been looking for a new injection field so that the present area can be harnessed for production and is currently also using the Húsmúli area (Figure 7) for reinjection.

2.1 Drilling of well HN-3

Well HN-3 was drilled directionally to the east-southeast intersecting fractures and faults on the western margin of the Hengill volcanic system in order to reach the high-temperature system (Thórarinnsson et al., 2006). The well's coordinates (ISN93) are as follows:

X=381127.55
 Y=393064.11
 Z=263 m a.s.l.

The well is of the wider kind with an 18⁵/₈" safety casing and a 13³/₈" production casing and a 12¹/₄" production section. The drilling of the well started February the 8th, 2006 and was finished March the 16th, 2006. The drilling progress is shown in Figure 9.

The first section of the well was drilled with a 21" drillbit with water. Drilling proceeded slowly since the rock formations were both hard and fractured. Circulation losses were noted at 102, 116, 134, and 140 and 143 m. Mica was put in the well to prevent further loss. The first section of the well was finished at 148 m. Temperature and calliper logs were made before an 18⁵/₈" casing was put in and

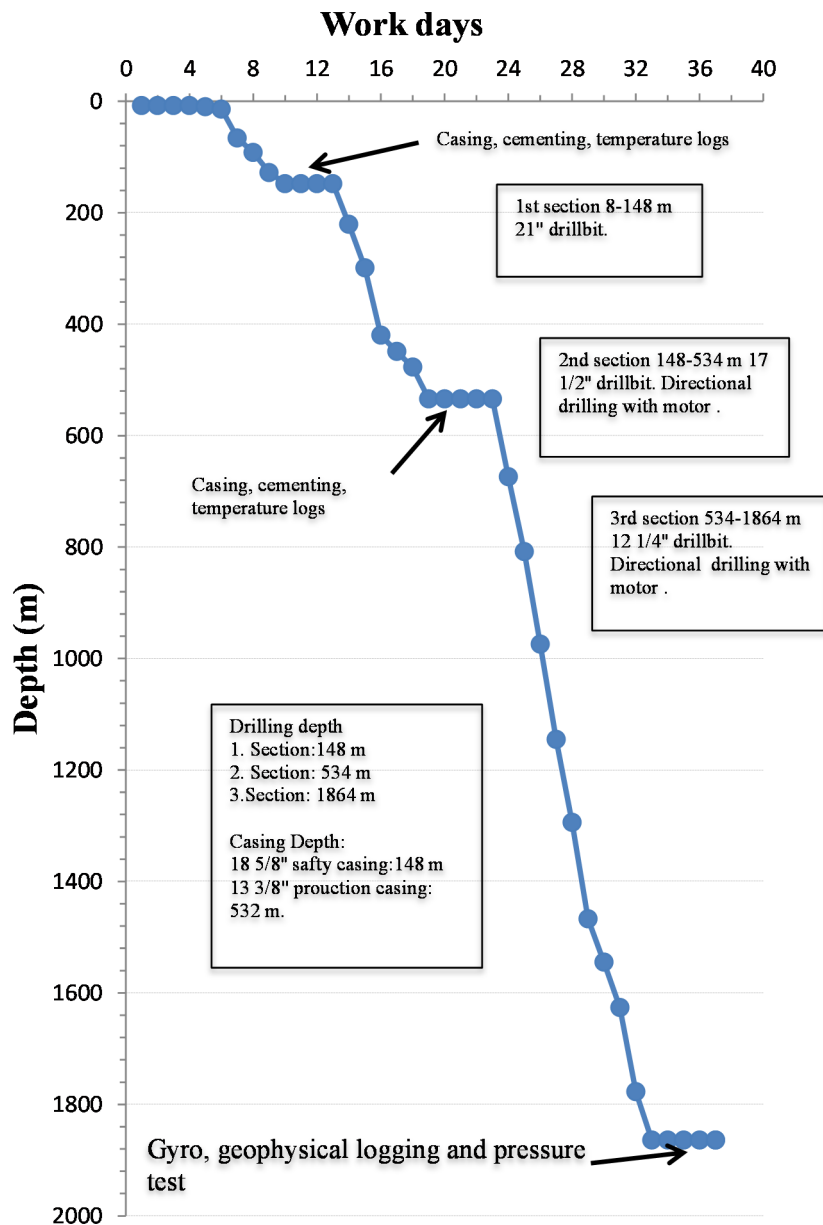


FIGURE 9: Drilling progress of well HN-3 and logging carried out during the drilling

cemented. A total of 32 m³ was used for cementing the casing.

The second section of the well was drilled with a 17½" drillbit with a motor and MWD equipment to direct the well to 110° with a 30° inclination. The initiation of the inclining of the well proved difficult, perhaps because of the hardness of the rock. At 537 m drilling was stopped when the inclination had reached 27°. Circulation loss of 20 l/s was measured at that depth. Temperature and geophysical logging was carried out before the well was cased with a 13⅜" casing. The first cementing included 74.5 m³ but a 5 l/s circulation loss was, however, still measured which meant that the cement had sunk below the first casing at 150 m depth. Cement bond logs confirmed that. Then, 20 m³ of cement were pumped down between the casings and still further 4 m³ to finish the job.

The third section was drilled with a 12¼" drillbit, slotted liner excluded. According to the drilling schedules it was planned to drill to 1500 m measured depth. Hardly any circulation losses had occurred at that depth and it was decided to continue with the pumping of water down the well for 3 hours. The circulation loss did not increase and it was decided to drill further, to 1800-2000 m. No circulation losses occurred until at 1689 m when 26 l/s were lost. At last total circulation loss occurred at 1830 m. The drilling of the third section was then stopped at 1864 m. Geophysical logs were conducted along with temperature logs and gyro measurements (for direction and inclination of the well).

Following logging, the well was circulated with water and air to clear out cuttings from the aquifers. The temperature in the well was measured again and proved to be around 40°C at the bottom. A short estimation of the injection coefficient gave the result of 4.4 (l/s)/bar. The drilling of well HN-3 took 37 work days (Figure 9).

2.2 Analytical methods

Binocular microscope

Rock cuttings were sampled at an interval of 2 m except where there were no cuttings due to circulation losses. The samples were mounted on a binocular microscope to analyse rock type, grain size, oxidation, alteration minerals, veins, vein fillings and primary minerals.

Petrographic microscope

The petrographic analyses were based on 20 thin sections from different depths of the uppermost 1272 m of well HN-3. These were used to confirm rock type, texture, porosity, alteration minerals and to identify additional minerals not seen in the binocular microscope. The petrographic analyses were also very helpful in identifying the alteration of primary minerals.

XRD analysis

The XRD analysis is a method used to examine crystallized materials based upon the scattering of X-rays according to the crystal type of the material. The X-ray diffractometer analyses were carried out to distinguish between various clay minerals. In this study, the prepared samples represent the different rocks and the degree of alteration intensity. Such analyses are useful in determining the alteration zones along with other index minerals found in binocular and petrographic analysis. The results of the analyses of clay mineral samples from well HN-3 are listed in Appendix I.

Interpretation of geophysical logs

Neutron logs: The physical basis of measurements of neutron logs is the scattering and capture of neutrons; the main purpose is to estimate the formation porosity (Ásmundsson, 2011). This is e.g. useful in the identification of intrusions, which are usually compact compared to the intruded rock formations.

Calliper log measures the diameter of the well as a function of depth using a logging probe with arms placed symmetrically around it (Steingrímsson, 2011). Indications of wider areas in the well can point

to the locations of soft formations (e.g. tuff) or fractured areas where the walls of the well are easily broken.

Resistivity logs: The specific resistivity of the reservoir rock is the result of two different contributions, the resistivity of the rock matrix and the formation fluid. An igneous rock matrix is generally a poor electrical conductor at geothermal temperatures (e.g. Mostaghel, 1999). Thus, an igneous rock with appreciable porosity or fluid filled vesicles will show somewhat lower resistivity. The resistivity graph in Figure 10 shows lower resistivity in the aquifer zones.

Gamma log measures the natural gamma radiation, used to determine the clay content in rocks. Investigations in Iceland show, however, that the gamma ray activity in volcanic rocks is related to the SiO₂ content of the rock and can, therefore, be used to identify rocks of evolved compositions (e.g. Mostaghel, 1999).

2.3 Stratigraphy of well HN-3

The stratigraphic column of well HN-3 consists primarily of alternating sequences of hyaloclastite units and layers of lavas with basaltic intrusive bodies. The units are further divided into various lithologies: tuffaceous sediment, the hyaloclastite lithology of basaltic breccia, basaltic tuff, and partially crystallized basalt and lava flows with fine- to medium-grained texture. Overall, three kinds of formations were recognized: hyaloclastite units, lava flows and tuffaceous sediment along with two intrusions. The hyaloclastites and pillow lavas have a glassy texture. The outer rim of the pillow lava has a black glassy surface known as tachylite (Saemundsson and Gunnlaugsson, 2002). However, the interior parts of the pillow lavas have crystallized textures of tholeiites and olivine tholeiites, the latter being sometimes porphyritic. Volcanic breccia is characterized by abundant lithic fragments. The lava flows are tholeiitic and olivine tholeiitic in composition. Tholeiitic basalt is fine to medium grained. It shows crystals of plagioclase, clino-pyroxene and magnetite, with minor olivine (only as phenocrysts). Olivine tholeiite is fine to medium grained with crystals of plagioclase, clino-pyroxene, magnetite and some olivine. Intrusive rocks in this well are fine- to medium-grained basalts. They are relatively unaltered compared to the surrounding rocks. The intrusive bodies are found at 1024-1032 and 1234-1244 m depth. The lithological descriptions are based mostly on binocular microscopic analysis with additional information from petrographic analysis of thin sections. Figure 10 shows the simplified lithology of the well along with geophysical logs. The descriptions and characteristics of the lithological units are as follows:

0-14 m. No cuttings.

14-16 m. Fine- to medium-grained basalt

Solid, evenly grained, fine-grained tholeiite lava, oxidized and vesicular at the lower boundary.

16-26 m. Tuffaceous sediments

Mixture of grey, fine- to medium-grained glass which is less oxidized than the above formation.

26-60 m. Hyaloclastite unit (basaltic tuff)

Grey to brownish fine- to medium-grained rock, slightly oxidized with vesicles. Olivine and plagioclase porphyritic.

60-130 m. Hyaloclastite unit (partially crystallized basalt)

Light grey to dark brownish, vesicular fine- to medium-grained basalt, moderately oxidized. Olivine and plagioclase porphyritic. Limonite and siderite were found at 96-102 m and small traces of carbonate at 106 m, most likely aragonite.

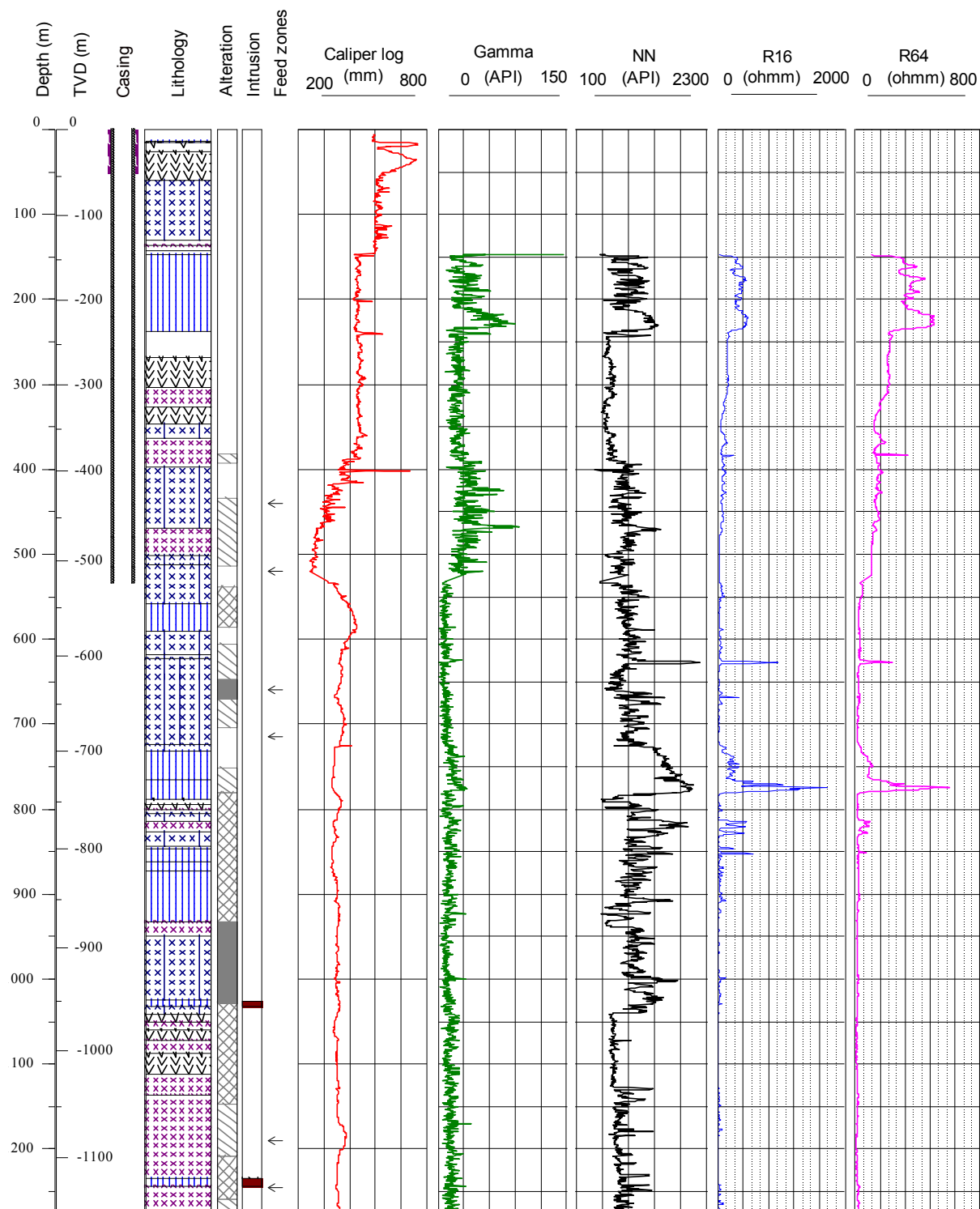


FIGURE 10: Simplified lithology and geophysical logs of well HN-3

130-136 m. No cuttings

136-142 m. Hyaloclastite unit (basaltic breccia)

Light grey to light brownish, fine- to medium-grained with vesicles. Slightly oxidized.

142-148 m. No cuttings

148-238 m. Fine- to medium-grained basalt

Fine- to medium-grained olivine-tholeiite, slightly porous, mixture of grey to brownish grains. Olivine porphyritic and a few plagioclase phenocrysts were found. Aragonite, siderite and limonite are common at this depth. Fine-grained basalt at 206-214 m.

*238-268 m. No cuttings**268-304 m. Hyaloclastite unit (basaltic tuff)*

Dark grey to black, medium-grained glass with vesicles.

304-326 m. Hyaloclastite unit (basaltic breccia)

Dark grey to light grey, glassy and vesicular with limonite and calcite.

326-346 m. Hyaloclastite unit (basaltic tuff)

Well compacted, light grey to brownish colour. Some deposition minerals were found like chabazite and/or analcime. The rock is relatively fresh. Common alteration minerals are calcite, limonite and fine-grained smectite.

346-364 m. Hyaloclastite unit (partially crystallized basalt)

Light grey to brownish, vesicular, pillow basalt. Aragonite, calcite, dogtooth-calcite, limonite and smectite are common alteration minerals. Zeolites are also common.

364-398 m. Hyaloclastite unit (basaltic breccia)

Light grey to brownish, fine- to medium-grained basalt combined with glass. Basalt has become less crystallized. Zeolites are more common, mainly chabazite.

398-470 m. Hyaloclastite unit (partially crystallized basalt)

Light grey to brownish, vesicular pillow basalt. The majority of the rock is crystallized or slightly crystallized tholeiite, non-porphyritic. The pillow basalt has a rather solid crystallization which is usually fresh (>60%). Pyrite, chalcedony, fine-grained clay and smectite are common alteration minerals together with some zeolite minerals.

470-502 m. Hyaloclastite unit (basaltic breccia)

Light grey to brownish in colour, moderately altered. The most common alteration minerals are smectite, thomsonite, scolecite, mesolite and calcite.

502-512 m. Hyaloclastite unit (partially crystallized basalt)

Light grey to dark grey, plagioclase porphyritic basalt with rare olivine phenocrysts. Alteration minerals are calcite, smectites and zeolites (mainly chabazite).

512-558 m. Hyaloclastite unit (partially crystallized basalt)

This layer is variably crystallized with olivine and plagioclase phenocrysts, most likely olivine-tholeiite. Dark grey to brownish, partial crystallization, but that which can be seen is solid and plagioclase porphyritic with rare olivine. Above, it is still relatively fresh, but with depth alteration increases a little. The most common alteration minerals are calcite, smectite, thomsonite, chabazite, and scolecite/mesolite. Small traces of pyrite are sometimes seen.

558-590 m. Fine- to medium-grained basalt

Light grey to dark grey, vesicular, olivine-tholeiite, with rare plagioclase phenocrysts. Deposition minerals are very common, mainly calcite, scolecite/mesolite, phillipsite, chabazite, thomsonite, heulandite, smectite and limonite.

590-618 m. Hyaloclastite unit (partially crystallized basalt)

Dark grey to reddish, vesicular basalt, rare olivine and plagioclase phenocrysts. Thomsonite, mesolite/scolecite, stilbite, calcite with a few pyrite crystals are common alteration minerals.

618-622 m. Hyaloclastite unit (partially crystallized basalt)

Dark grey to reddish, fine-grained basalt, though there are still a few grains in between that are medium-grained and pyroxene porphyritic. Few deposition minerals, trace of calcite, stilbite, scolecite and thomsonite. Moderately oxidized.

622-724 m. Hyaloclastite unit (partially crystallized basalt)

Dark grey basalt, plagioclase and olivine porphyritic and with rare pyroxene phenocrysts. A lot of deposition minerals are present, mainly calcite, thomsonite, chabazite, stilbite, scolecite/mesolite and smectite.

724-732 m. Hyaloclastite unit (partially crystallized basalt)

Dark grey to reddish, evenly-grained basalt, likely tholeiite. Fine-grained smectite and calcite are the most common alteration minerals at this depth.

732-766 m. Fine- to medium-grained basalt

Dark grey, crystallized rock, likely olivine-tholeiite. Very rarely pyroxene phenocrysts can be seen. A lot of deposition minerals are at the top, mainly zeolites like scolecite/mesolite, thomsonite and stilbite, also including calcite and fine-grained clay.

766-788 m. Fine- to medium-grained basalt

This layer is light to dark grey in colour, moderate pores, made up of crystallized olivine-tholeiite basalt. Common minerals at this depth are calcite, fine-grained clay, stilbite, phillipsite and scolecite.

788-794 m. Tuffaceous sediment

The tuff is dark grey to reddish in colour because of oxidation. Calcite and phillipsite are the main alteration minerals.

794-800 m. Hyaloclastite unit (basaltic tuff)

Light to dark grey, vesicular tuff. Vesicles are mostly filled with dark green clay. The most common alteration minerals are smectite, calcite, heulandite, scolecite and stilbite.

800-806 m. Hyaloclastite unit (basaltic breccia)

Light grey pillow breccia, a mixture of tuff, crystallized basalt and more crystallized olivine-tholeiite. Smectite and scolecite are common alteration minerals.

806-814 m. Hyaloclastite unit (partially crystallized basalt)

This unit is light to dark grey, pillow basalt with calcite, scolecite and thomsonite.

814-826 m. Hyaloclastite unit (basaltic breccia)

This layer is light to dark grey breccia with light grey tuff and crystallized basalt. The most common alteration minerals are stilbite, scolecite, fine-grained clay and calcite. Included are small traces of chalcedony.

826-844 m. Hyaloclastite unit (partially crystallized basalt)

Light grey to black, vesicular rock. It is a mixture of partially crystallized basalt and fully crystallized basalt. Most common alteration minerals are smectite, scolecite, chabazite and heulandite.

844-846 m. No cuttings

846-862 m. Fine- to medium-grained basalt

This unit is light to dark grey, fine-grained basalt, most likely tholeiite. Calcite, analcime, scolecite and heulandite are common alteration minerals.

862-872 m. *Fine- to medium-grained basalt*

This unit is light to dark grey, vesicular, olivine tholeiite basalt. Common alteration minerals are chabazite, scolecite and thomsonite along with calcite.

872-932 m. *Fine- to medium-grained basalt*

Lava flow, light grey to brownish, fine- and evenly-grained tholeiite. Olivine and plagioclase are quite altered. Main alteration minerals are calcite, heulandite, stilbite, chalcedony, chabazite, smectite, fine-grained clay and a trace of pyrite.

932-950 m. *Hyaloclastite unit (basaltic breccia)*

Light grey to brownish and slightly oxidized. Heulandite, stilbite, yugawaralite and calcite are common minerals.

950-1024 m. *Hyaloclastite unit (partially crystallized basalt)*

This is light to dark grey, vesicular, variedly crystallized olivine-tholeiite and crystallized basalt. Oxidation increases towards the bottom. A lot of deposition minerals were found, mainly stilbite, scolecite/mesolite, chalcedony, yugawaralite, calcite, MLC and a trace of pyrite. MLC was confirmed at 1008 m through XRD analysis (see Appendix I).

1024-1032 m. *Basaltic intrusions*

Medium-grained fresh basalt, olivine and plagioclase porphyritic.

1032-1042 m. *Hyaloclastite unit (partially crystallized basalt)*

This unit is dark grey to brownish. Fine- to medium-grained crystallization slightly oxidized. Stilbite, scolecite, chalcedony, yugawaralite, stilbite and calcite are common alteration minerals as well as a trace of pyrite.

1042-1050 m. *Hyaloclastite unit (basaltic tuff)*

This is dark grey vesicular tuff. Phillipsite, scolecite, chlorite and quartz are common.

1050-1060 m. *Hyaloclastite unit (basaltic breccia)*

This is light to dark grey tuff rich breccia, slightly altered and porous. Common alteration minerals are chalcedony, heulandite and quartz.

1060-1072 m. *Hyaloclastite unit (basaltic tuff)*

Light grey to dark green tuff mixed with breccia. Slightly porous with coarse-grained clay, fine-grained clay, MLC (which was confirmed at 1064 m by XRD analysis), mesolite/scolecite, heulandite, prehnite and traces of pyrite.

1072-1088 m. *Hyaloclastite unit (basaltic breccia)*

Light grey to brownish tuff rich breccia, mostly dark green, sometimes lighter. Vesicles in the tuff are to a large extent filled with calcite, heulandite and quartz with a trace of pyrite.

1088-1112 m. *Hyaloclastite unit (basaltic tuff)*

This is grey-green tuff, quite vesicular. Vesicles mostly filled with deposition minerals. Prehnite, heulandite, MLC and quartz with pyrite are quite common. MLC was detected by XRD analysis at 1104 m.

1112-1136 m. *Hyaloclastite unit (basaltic breccia)*

This unit is dark grey to brownish, glassy and highly oxidized rock. Laumontite, phillipsite, heulandite and stilbite are common alteration minerals.

1136-1234 m. Hyaloclastite unit (basaltic breccia)

Tuff rich breccia, light grey to brownish, slightly oxidized. Common hydrothermal minerals are MLC, heulandite, yugawaralite and laumontite. At the depth of 1188-1234 m there is light green, quite vesicular tuff. Vesicles are usually filled. Alteration minerals are mainly scolecite, chalcedony, stilbite, calcite and pyrite. MLC was identified through XRD at 1140, 1166 and 1198 m.

1234-1244 m. Basaltic intrusions

Fine- to medium-grained, fresh basalt.

1244-1272 m. Hyaloclastite unit (basaltic breccia)

Light grey to brownish tuff rich breccia. Quite a lot of deposition minerals are here, including scolecite, chalcedony laumontite, MLC at 1262 m, coarse-grained clay, quartz, calcite, stilbite, heulandite and pyrite.

2.5 Intrusive rocks

An intrusion is an emplacement of magma into pre-existing rock. As such, they usually show relatively coarser textures compared to the host rock since the subterranean magma cools slowly giving time for crystal growth (Lugaizi, 2011). Intrusive rocks are characterized by their massive, compact nature and appear fresh in comparison to the surrounding lithology, and are sometimes marked by oxidation near their margins. Two intrusions were identified in well HN-3 at different depths. They are not prominent in the neutron-neutron and resistivity logs (Figure 10) which should show the compactness of the intrusions.

Intrusive unit I (1024-1032 m)

This is fine- to medium-grained basalt with an apparent thickness of approximately 8 m. The rock is dark grey in colour, slightly altered and plagioclase porphyritic with accessory micro-crystals of olivine.

Intrusive unit II (1234-1244 m)

This is a 10 m thick fine- to medium-grained basaltic intrusion. The rock is light grey in colour, fresh and characterized by plagioclase phenocrysts.

3. HYDROTHERMAL ALTERATION

Hydrothermal alteration refers to the change in mineralogy, texture, porosity and permeability of rocks due to its interaction with hydrothermal fluids. Primary minerals are replaced by secondary minerals as a result of change in the prevailing conditions to which the rock is subjected. These changes are primarily variations in temperature, pressure, or chemical conditions or any combination of these. The fluids carry metals in solution, either from a nearby igneous source, or from leaching out of nearby rocks. Hydrothermal fluids cause hydrothermal alteration of rocks by passing hot fluids through the rocks and changing their composition by adding or removing or redistributing components. Temperatures can range from weakly elevated to boiling. Fluid composition is extremely variable. They may contain various types of gases, salts (brines), water, and metals. The metals are carried as different complexes, thought to involve mainly sulphur and chlorine (Lagat, 2009).

In a geothermal system, hydrothermal alteration can reveal the history, and possibly the future, of the system involved. In addition, hydrothermal minerals are useful geothermometers and are used, for example, to assist in determining the depth of the production casing during drilling. Hydrothermal minerals are also used to determine the alteration temperatures of wells and, when compared to formation temperatures, can predict whether the rock formations in a specific area are cooling or heating. Hydrothermal minerals can also be used in estimating fluid pH and other chemical parameters, as well

as predicting scaling and corrosion tendencies of fluids, measuring permeability, possible cold-water influx and as a guide to the hydrology (Reyes, 1990). When the rock heats up, especially in active volcanic areas, it will release ions into the solution. This hydrothermal solution, after undergoing further heating, will rise up as a mineral rich fluid. The mineral rich fluid will tend to crystallize out of the solution into open cavities and/or open structures depending on temperature, rock type, pressure, permeability, fluid composition and duration of activity. For example, epidote starts to appear at about 240°C (Table 1). Figure 11 shows the mineral alteration zones in Iceland and the alteration of primary minerals vs. temperature.

TABLE 1: Some temperature dependent index minerals in high-temperature areas in Iceland (e.g. Kristmannsdóttir, 1979, Franzson, 1998)

Minerals	Min. temperature (°C)	Max. temperature (°C)
Zeolites	40	120
*Laumontite	120	180
*Wairakite	200	
Smectite		<200
**MLC	200	230
Chlorite	230	>300
Calcite	50-100	280-300
Quartz	180	>300
Prehnite	240	>300
Epidote	230-250	>300
Wollastonite	270	>300
Actinolite	280	>300

*Belong to the zeolite group; **Mixed-layer clay

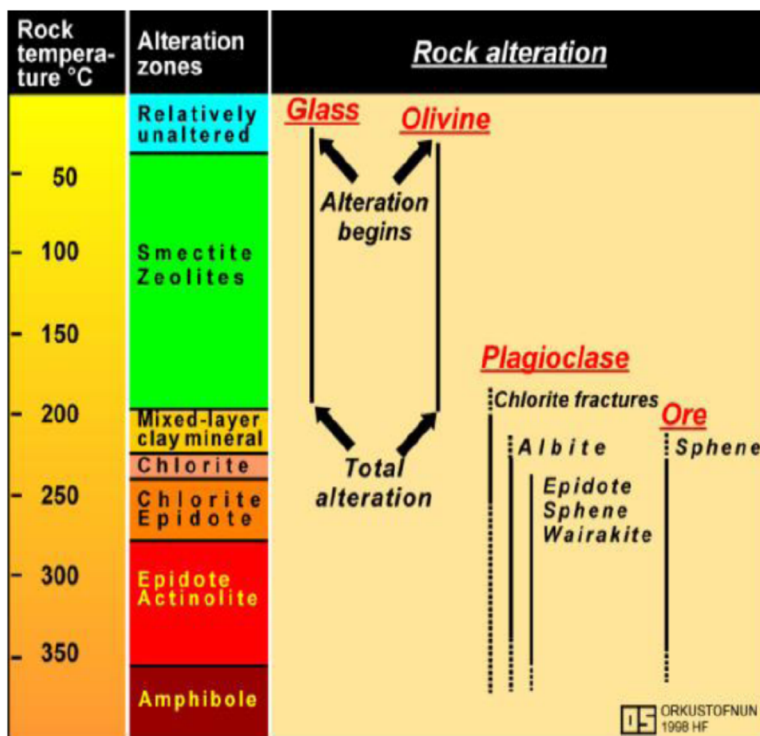


FIGURE 11: Mineral alteration vs. temperature diagram (Franzson, 1998)

3.1 Primary rock minerals and alteration products

Primary minerals, crystallized from magma, governed by the physico-chemical conditions under which the magma solidifies, become unstable in a geothermal environment where there is high permeability, elevated temperature and intense fluid activity (Gebrehiwot, 2010).

The primary constituents found in well HN-3 are: Plagioclase, olivine, pyroxene and opaque minerals (magnetite-ilmenite) (Table 2). The interaction between the wall rocks and the hydrothermal fluids is the main factor that leads to the replacement of primary minerals with hydrothermal minerals,

while the process of fluid circulation affects mineral deposition in veins and cavities (Browne, 1978). Glass alters mostly to clays, calcite, zeolites and quartz, olivine alters to clays, calcite or oxides, and pyroxene is relatively resistant to hydrothermal alteration but mainly alters to clays and, at higher temperatures, to titanite and actinolite. Plagioclase is also relatively resistant to alteration but starts to alter to clay minerals and, in accordance with increasing temperature, to albite, calcite and chlorite.

TABLE 2: Primary mineral alteration in well HN-3
(the arrow shows decreasing susceptibility to alteration)

Fresh/unaltered	Alteration products
Glass	Clay, zeolite, calcite, quartz Clay, calcite, quartz Clay, calcite, quartz, wairakite, chlorite Clay
Olivine	
Plagioclase	
Pyroxene	
Opaque	

The main primary minerals in Iceland in basaltic rocks are (excluding glass): olivine, plagioclase, pyroxene and opaques. As seen glass is the most unstable constituent of basaltic rocks (Gebrehiwot, 2010).

Volcanic glass: Amorphous quenched magma showing a highly vitreous lustre and good conchoidal fractures. It is the first constituent to be altered and replaced by alteration minerals. The replacement products of volcanic glass are clays, zeolites (e.g. mordenite, laumontite), quartz and calcite (Browne, 1978). At 624-720 m and at 1064 m and below, glass is completely altered in well HN-3.

Olivine: Compositional series of minerals (forsterite-fayalite) based on the proportion of magnesium to iron. Olivine is green (colourless in thin sections) but transparent or translucent with a vitreous lustre and the streak is white. Fracture is conchoidal, and cleavage indistinct. It is one of the primary minerals that forms basaltic rocks (e.g. olivine tholeiite) and is very susceptible to alteration. Iddingsite is formed by the alteration or weathering of olivine. It is common in slightly altered basalt. Olivine is distinguished in thin section by its high birefringence, distinctive irregular fracture pattern, and alteration products, usually clay, forming along fractures. It is completely replaced by clay and sometimes calcite and oxides. Olivine is completely altered at 624-720 and 1064-1270 m depth (Table 3).

Plagioclase: It is the most abundant mineral occurring in most igneous rocks. In crystalline rocks it is readily identified by its low relief and conspicuous polysynthetic twinning. It shows moderate alteration in well HN-3, increasing significantly at 1064 m (Table 3).

Pyroxene: It is a compositional series of minerals. It is black or dark green in colour and forms prismatic crystals with a vitreous lustre and perfect cleavage. It is translucent or transparent; the streak is green or grey, and fractures are uneven. The dominant pyroxenes that occur in Icelandic basalts are the clinopyroxenes occurring mostly as phenocrysts and in the ground mass (Gebrehiwot, 2010). In the upper part of the well it was difficult to identify the alteration of pyroxene. Pyroxene often shows signs of rapid cooling in hyaloclastite formations where it seems to be partially crystallized. In those cases it is often difficult to see if alteration has really occurred.

Opaque: Those minerals are less altered compared to others. Magnetite/ilmenite occurs in small amounts in basaltic rocks and appears opaque in transmitted light. It is mildly altered in the well, in most cases only showing some signs of oxidation (Table 3).

TABLE 3: Distribution of primary mineral alteration; numbers refer to the intensity of alteration; 4 - very strongly altered; 3 - strongly altered; 2 - moderately altered; 1- weakly altered; 0 - unaltered

Depth (m)	Distribution of primary minerals alteration in well HN-3				
	Glass	Olivine	Plagioclase	Pyroxene	Opakes
238	2	2	1	1	1
310	2	2	1	1	1
340	3	3	2	2	1
404	3	2	1	1	1
440	3	3	2	1	1
508	3	2	1	1	1
586	2	2	1	1	1
624	4	4	2	1	1
720	4	4	2	1	1
728	3	3	2	1	1
740	3	3	2	1	1
780	3	3	1	1	1
820	2	2	0	0	0
850	2	2	1	0	0
880	1	1	0	0	0
1020	2	2	1	1	1
1064	4	4	3	1	1
1190	4	4	2	2	1
1270	4	4	3	1	1

3.2 Distribution of hydrothermal minerals

Hydrothermal minerals are formed as a result of the alteration of primary minerals. The hydrothermal alteration depends on many factors such as temperature, permeability, rock type fluid chemistry, pressure and rock structure (Kristmannsdóttir, 1978; Reyes, 1990; Browne, 1978 and 1993; Malapitan, 1995). The most common minerals in well HN-3 are limonite, siderite, pyrite, calcite, clays, zeolites, chalcedony, prehnite and quartz. The distribution of alteration minerals with depth in well HN-3 is shown in Figure 12, and the different minerals are described below:

Calcite: It is the most abundant alteration mineral in the well, appearing from 30 m and seen all the way to the bottom of the well. It is very common in the pores of hyaloclastites and lava flows but less in the more compact intrusions. Calcite either occurs as a replacement of primary minerals or as a direct depositional mineral. Hydrochloric acid was used to identify it in the cuttings. Calcite crystals have clear cleavage and high interference colour in thin section and can be distinguished from plagioclase by these properties.

In thin section, platy and dog tooth-like calcites were observed at 340 m. It also occurs as vesicle fillings. Calcite can be linked to boiling, dilution and condensation of carbon dioxide in the geothermal system. It can also form during the heating of cooler peripheral fluids (Simmons and Christenson, 1994).

Pyrite: It is most abundant in highly altered rock of extinct central volcanoes, where it forms shiny coatings on the surface of fissures, or clusters of glittering golden granules in the rock. The mineral forms yellowish cubic crystals and is seen at the depth of 28-30, 36-42, 142-152 and 866-868 m in well HN-3. Pyrite in a geothermal system can be used as an indication of permeability.

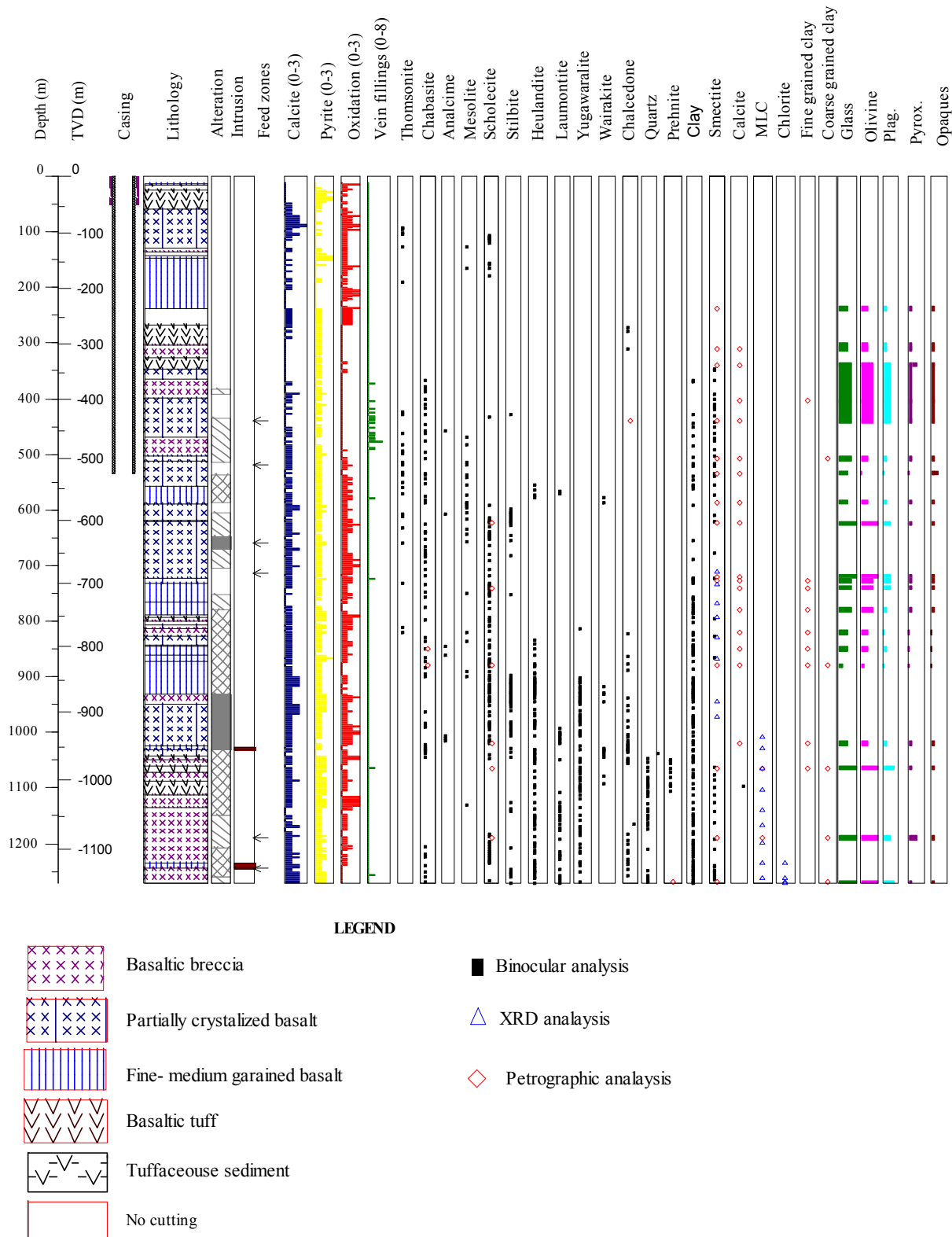


FIGURE 12: Distribution of hydrothermal alteration minerals, location of aquifers and alteration intensity

Limonite: It is characterized by a yellowish brown colour. It often forms spherical precipitation in vesicles. It is hydrated iron oxide and is formed due to the interaction of rocks with cold groundwater. It is present in the well at 96 m down to 582 m.

Siderite: Shows a radiating pattern when broken. It also occurs as tabular crystals, which may be curved. It is yellowish or yellowish-brown, occurs with limonite and belongs to the group of carbonates. In the cuttings, it appears at 190-218 m depth.

Zeolites: These are a category of hydrous secondary minerals composed of sodium calcium aluminium silicates. Zeolites are generally white or colourless, but sometimes yellow by minor impurities. There are different types of zeolites. Zeolites are formed during alteration and precipitation in vesicles. As water percolates down into a rock pile, it heats up, depending on the depth of penetration. In nature, zeolites are temperature dependent which can help in determining the temperature at a given depth in geothermal systems. In well HN-3 zeolites appear at 92 m and can be found all the way down to 1270 m. The different types of zeolites in well HN-3 are described below:

Thomsonite: It is an acicular radiating zeolite. Crystals are elongated, slightly flattened, somewhat fibrous, with a slanted end. They form dense masses of radiating clusters. Its colour is mainly milky white and brownish. Thomsonite was observed in the binocular microscope starting from 92 m depth and crystallizes at around 30-40°C.

Scolecite: It is a typical fibrous zeolite with flattened radiating crystals from a common centre (Figure 13). Scolecite is colourless or white with a vitreous or slightly silky lustre. It is found with various other zeolites that are relatively low in silica, such as mesolite, chabazite, analcime and thomsonite. It started to form at a depth of 164 m in the well. Scolecite precipitates between ~80 and 120°C. In thin section scolecite was observed at 620, 740 and 880 m depth.

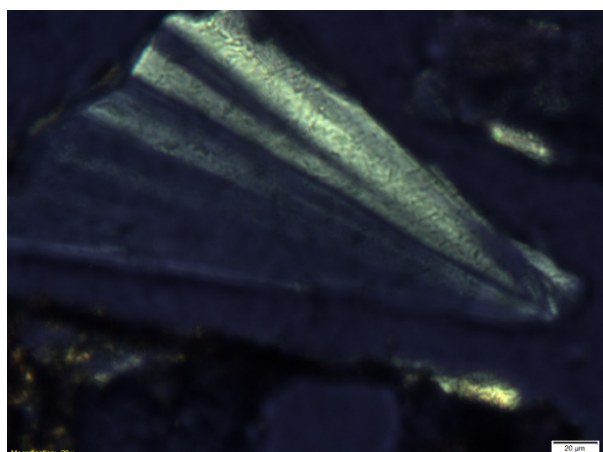


FIGURE 13: Radiating crystal of scolecite at 620 m depth in well HN-3

Mesolite: It forms small scale fibrous radiating clusters. Mesolite is usually translucent to white in colour. It occurs with other zeolite minerals especially with scolecite. It was first found at a depth of 128 m.

Stilbite: It is white and transparent to translucent, either colourless or white. It occurs associated with heulandite in this well, first found at a depth of 600 m. Stilbite is an indication of the temperature range 90-120°C (Ahmed, 2008).

Chabazite: It is white or brownish in colour. In the well, its first appearance was at 374 m depth together with thomsonite and other zeolite minerals. It is stable from about 30 to about 80°C (Kristmannsdóttir, 1979). In thin sections chabazite was observed at 850 and 880 m depth.

Analcime: This mineral forms white or colourless crystals and forms many sided (trapezohedral) crystal faces. It was first found at the depth of 606 m. Analcime occurs along with chabazite and thomsonite and has a formation temperature of about 50-70°C (Ahmed, 2008).

Heulandite: It is transparent or translucent and forms tabular crystals with a whitish-colour appearance. This mineral was recognized in the binocular microscope, first found at the depth of 556 m. It has a crystallization temperature of 90-120°C (Franzson, 1998).

Yugawaralite: It forms colourless or white platy crystals with oblique ends. It was first found at a depth of 906 m, where it occurred with calcite, stilbite and heulandite.

Chalcedony: It is a form of silica, with a cryptocrystalline texture (Kristmannsdóttir, 1979). It occurs in thin section as a thin lining inside vesicles. It is colourless, grey to greyish blue and semi-transparent to translucent with a waxy lustre and occurs within the depth interval of 720 to about 966 m, below that depth it was replaced by quartz. In thin sections, it can be seen as a vesicle filling, followed by fine-grained clay at 404 m depth.

Laumontite: In well HN-3, it started to appear at 992 m. It has a prismatic and fibrous structure and is mainly found as cavity fillings. It was found in partially crystallized basalt, and indicates temperatures of 120-200°C.

Clay minerals: The clay minerals are the most voluminous alteration minerals. Their formation involves the chemical reactions and physical movement of hydrothermal fluids and, in many cases, there is a zonal arrangement of the clay minerals around the source of alteration, depending on the parent rock and the nature of the hydrothermal solutions (Gebrehiwot, 2010). The composition, structure and morphology of clay minerals depend on a number of environmental parameters – temperature, fluid composition/amount, pH, etc. (Njue, 2010). Clay minerals are alteration products of basaltic glass, olivine, plagioclase, and partly pyroxene, as well as vesicle fillings. Smectite, mixed-layer clays and chlorite comprise the different types of clays in the well.

Smectite: It is a low-temperature clay formed by the alteration of glass and ferromagnesian primary minerals such as olivine and commonly replaces them completely (Lugaizi, 2011). Smectite has a brownish colour and is found in patches or as thin coatings in altered rocks. In well HN-3 it was first found at 310 m depth. Under the microscope, smectite is fine grained with a dark-brownish colour. It is deposited as a thin lining layer in vesicles and as a mineral replacement. In XRD, smectite clay was identified between 712-972 m. Smectite forms at temperatures below 200°C (Franzson, 1998).

Mixed-layer clays: Mixed-layer clay (MLC) minerals are intermediate products of pure end-members of clays (Sradon, 1999). The reactions that produce mixed-layer clays are progressive high-temperature reactions altering smectite into chlorite via mixed-layer smectite/chlorite (Lugaizi, 2011). This type of clay was identified, in thin sections, at 1064 m (Figure 14) and 1190 m depth. Mixed-layer clays are usually green or bluish green in plane polarized light. XRD analysis detected mixed-layer clays between 1008 and 1198 m. At 1198 m depth, the sample shows peaks at about 12.72 Å for untreated samples, 13.40 Å for glycolated ones, and 9.74 Å for heated samples.

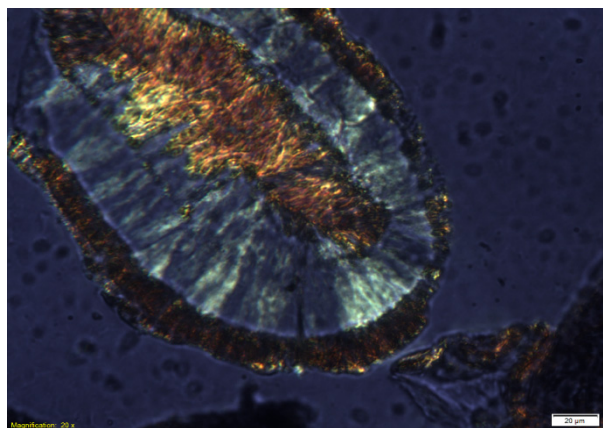


FIGURE 14: MLC found at 1064 m depth in well HN-3

Chlorites: This mineral is generally green, forming radiating, flaky or fibrous microscopic crystals or anhedral aggregates with a stacked structure (Gebrehiwot, 2010). In thin sections, it is coarse grained and has a light green colour in plane polarized light and grey in bipolarized light. Its texture shows tiny green needles, and sometimes radial forms. XRD analysis showed typical chlorite at 1236 and 1262 m depth. At 1262 m, the sample shows peaks at about 7.11 Å for an untreated sample and 7.11 Å for a glycolated sample. Chlorite has a crystallization temperature above 230°C (Franzson, 1998).

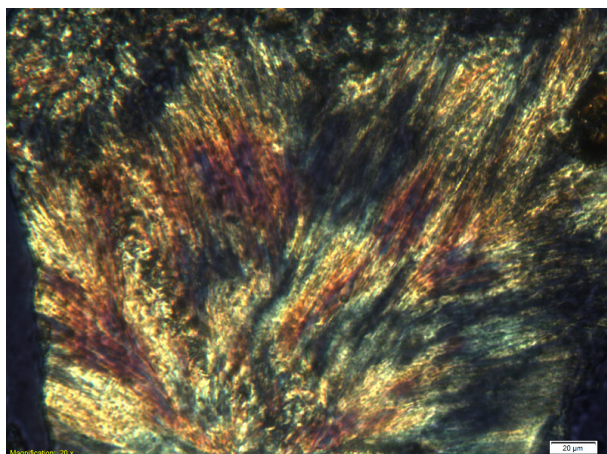


FIGURE 15: Prehnite found at 1270 m depth in well HN-3

Quartz: It is a white, transparent mineral. Quartz was first seen at a depth of about 1048 m and continued to the bottom of the well. Quartz has a temperature stability of $>180^{\circ}\text{C}$. Quartz is stable in a geothermal system and can be used as an indicator mineral for production casing if it shows a clear prismatic shape.

Prehnite: It occurs as rounded to nearly spherical in binocular microscope and shows colours of second order in bipolarized, transmitted light and is transparent in plane polarized light. It crystallizes at temperatures above 240°C . The first appearance of prehnite was observed at 1070 and 1272 m depth in cuttings and thin section (Figure 15).

3.3 Alteration mineral zonation

Many hydrothermal minerals in Iceland are highly temperature dependent and definite mineralogical changes are seen to take place with increasing temperature (Kristmannsdóttir and Tómasson, 1978; Malapitan, 1995). In the Icelandic geothermal setting, low-temperature zeolites and amorphous silica form below 100°C , chalcedony below 180°C , quartz above 180°C , wairakite above 200°C , epidote above $230\text{--}250^{\circ}\text{C}$ and garnet and amphibole above 280°C (Kristmannsdóttir, 1979; Saemundsson and Gunnlaugsson, 2002; Franzson, 1998). The clay minerals crystallize below 200°C for smectite, at $200\text{--}230^{\circ}\text{C}$ for mixed-layer clays, and above 230°C for chlorite (Kristmannsdóttir, 1977).

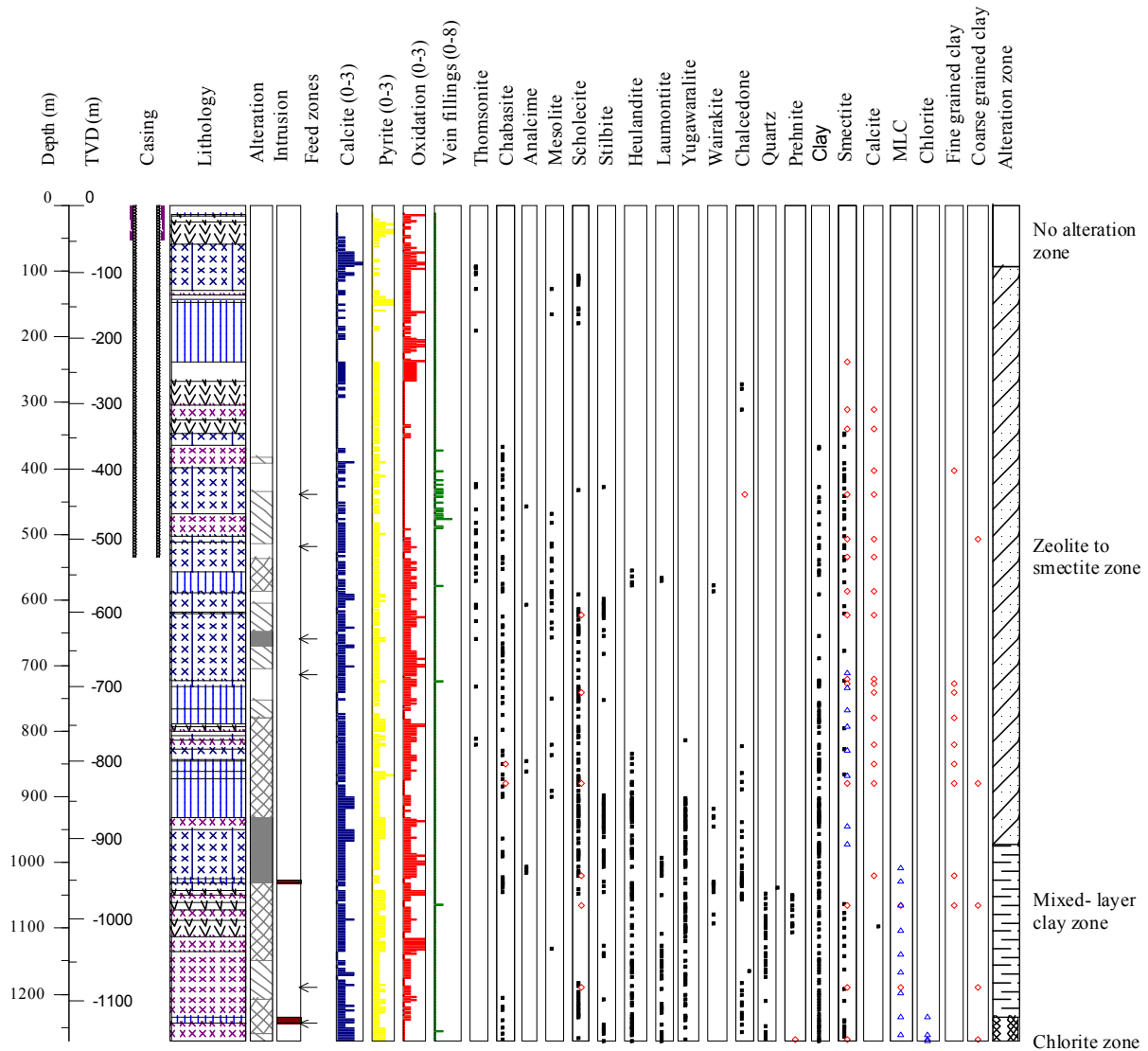
Based on binocular, petrographic and XRD analysis, four mineral alteration zones were distinguished in HN-3. The zones were revealed by considering the alteration mineral temperatures and by the abundance of certain minerals (Figure 16). The alteration zones are as follows:

Unaltered zone (0-92 m): This zone is composed of fresh rocks with no signature of alteration or development of any secondary minerals.

Zeolite-smectite zone (92-972 m): Low-temperature alteration zone which is characterized by the presence of zeolites and low-temperature clays (smectite). The upper boundary of this zone is represented by the first appearance of zeolites at 92 m. The zeolites present as secondary minerals in this zone are commonly scolecite/mesolite, chabazite, stilbite and thomsonite. Smectite indicates temperature less than 200°C .

Mixed-layer clay (MLC) zone (972-1236 m): It is the transition products of reactions involving clay mineral end-members in which the different kinds of clay layers alternate with each other. The characteristic temperature range of this alteration zone in Iceland is about $200\text{--}230^{\circ}\text{C}$.

Chlorite zone (1236-1272 m): It is a high-temperature clay mineral which signifies a temperature range of $230\text{--}250^{\circ}\text{C}$ and its first appearance marks the top boundary of this zone. Chlorite indicates temperatures above 230°C (Browne, 1978; Franzson, 1987).



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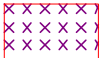

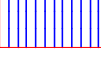

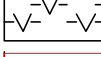




-  Basaltic breccia
-  Partially crystallized basalt
-  Fine- medium garained basalt
-  Basaltic tuff
-  Tuffaceous sediment
-  No cutting
-  Binocular analysis
-  XRD analysis
-  Petrographic analysis

FIGURE 16: Alteration minerals and alteration zones of well HN-3

3.4 Mineral deposition sequence

Minerals are formed at their characteristic physico-chemical conditions during the history of geothermal systems. Many of these minerals formed either by replacement or deposition, are temperature

dependent, hence by studying their depositional sequence in veins or vesicles, one can explore the parent thermal history and relative time scale of alteration minerals within a system (Gebrehiwot, 2010).

The alteration mineral sequence in well HN-3 can be seen in Table 4. In the thin sections, mineral sequences can be seen from 310 down to 1020 m depth. The depositional minerals were mostly found in vesicles. The alteration mineral assemblages change from low-temperature minerals to moderate- to high-temperature minerals with increasing depth. Clay and calcite are the most common minerals in the mineral sequence in this well. The fine-grained clay is mostly found as thin linings in the walls of vesicles. Coarse-grained clay is found, especially as fillings in veins or vesicles (Table 4).

TABLE 4: Sequence of alteration mineral deposition in well HN-3

Depth	Lithology	Alteration mineral sequence
310	Fine- to medium-grained basalt	Smectite > calcite
340	Basaltic tuff	Smectite > coarse-grained clay
404	Partially crystallized basalt	Fine-grained clay > chalcedony
440	Partially crystallized basalt	Smectite > chalcedony
508	Partially crystallized basalt	Calcite > smectite > coarse-grained clay
534	Partially crystallized basalt	Calcite > smectite
586	Fine- to medium-grained basalt	Smectite > calcite
624	Partially crystallized basalt	Smectite > scolecite/mesolite>calcite
728	Partially crystallized basalt	Smectite > calcite
740	Fine- to medium-grained basalt	Scolecite/mesolite > calcite
880	Fine- to medium-grained basalt	Smectite>coarse-grained clay
1020	Partially crystallized basalt	Scolecite/mesolite>calcite

4. AQUIFERS

Aquifers are typically water-saturated regions in the subsurface which produce an economically feasible quantity of water to a well. The movement of sub-surface water is controlled by the type of rock formations, the characteristics of its permeability and porosity, the temperature and pressure of the sub-surface environment, natural recharge, and the hydraulic gradient. The presence of structural formations such as faults, fractures, and joints, lithological contacts, clasts and fragmented matrixes, and paleosols are positive indications of geothermal feed zones (Reyes, 2000). The aim of deep geothermal drilling is to penetrate into a permeable high-temperature zone. The cause of permeability in rock formations can be faults, joints, intrusions and lithological contacts. Pillow basalts have higher effective permeability than any other rock type encountered by drilling in Icelandic geothermal areas (Fridleifsson, 1978).

Loss of circulation may vary from minor or weak loss, due to tight characteristics of the formation, to total circulation loss that is characterized by highly permeable formations. In the geothermal settings of Iceland, the highest permeable zones intercepted are associated with dykes and faults (Arnórsson, 1995; Pendon, 2006). Some of the aquifers encountered in geothermalwells can be directly related to lithological boundaries and margins of intrusions, indicating the presence of fracture permeability in the geothermal reservoir. The rock, which is rich in vein/vesicle networks and has a high concentration of hydrothermal minerals, is an indication of the presence of strong subsurface hydrological circulation (Mgejwa, 2011).

In well HN-3, aquifers were identified by geophysical logs (temperature, calliper and neutron logs), increased alteration minerals (e.g. calcite and pyrite), alteration intensity and circulation loss. Seven

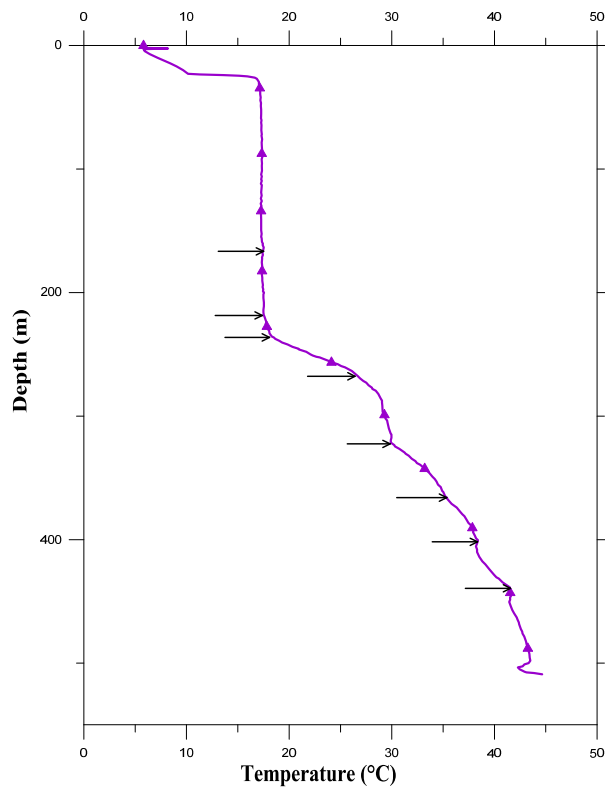


FIGURE 17: Very small aquifers identified from temperature logs at a depth of 165, 220, 240, 270, 325, 370 and 400 m from the second stage of drilling of well HN-3; the aquifer at 440 m can also be seen in this figure

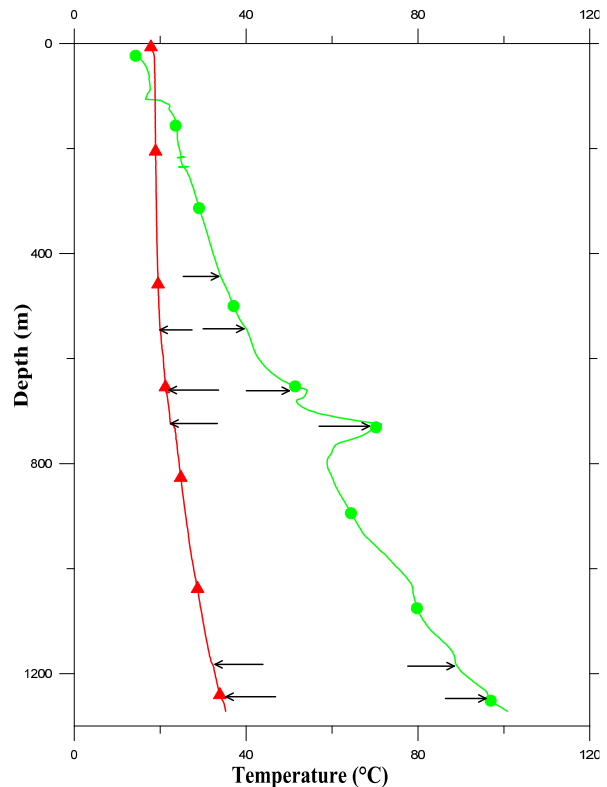


FIGURE 18: Small aquifers identified from temperature logs from the third phase of drilling well HN-3

small possible aquifers were identified at a depth above the production casing of the well in the upper hyaloclastite units (Figure 17). These are all minor. In addition, there are six small aquifers at 440, 550, 660, 725, 1180 and 1245 m (Figure 18). The lowermost aquifer is located by an intrusion.

Aquifer 1: This aquifer is located at 440 m within a unit of partially crystallized basalt. It was indicated by a high peak in the calliper log and a low peak in the neutron-neutron log (Figures 10 and 18).

Aquifer 2: It is located at 550 m in slightly altered, partially crystallized basalt. At this depth, a low peak in the gamma log and low resistivity log was identified (Figure 10). No circulation losses were observed.

Aquifer 3: This aquifer is found at 660 m in partially crystallized basalt. No circulation losses were observed but it was associated with highly altered rock, a low peak in the neutron-neutron log, low resistivity log and a low peak in the gamma log as well as in the temperature log (Figures 10 and 18).

Aquifer 4: It is found in partially altered and crystallized basalt at 725 m. At this depth there is a low peak in the neutron-neutron log, low resistivity log and a low peak in the gamma log (Figure 10). No circulation losses were observed at this depth.

Aquifer 5: It is located at a depth of 1180 m with moderately altered hyaloclastite basaltic breccia. It was indicated by a high peak in the calliper log (Figure 10).

Aquifer 6: This aquifer is located at the boundary of an intrusion and basaltic breccia at 1245 m with a lot of depositional minerals. It was identified by low resistivity and a low peak in the gamma log (Figure 10). No circulation losses were observed.

5. FLUID INCLUSIONS

Fluid inclusions are used to predict whether a geothermal system is cooling or heating and consequently may allow us to interpret the history of the system (Mgejwa, 2011). When hydrothermal minerals deposit, or recrystallize in a fluid environment, tiny growth irregularities trap small amounts of the depositing fluid within the solid crystal. The sealing off of these irregularities yields primary inclusions or, by later crystallization, perhaps in response to micro fractures, secondary inclusions (Browne, 1978). Secondary inclusions typically cross-cut earlier generations (Shepherd et al., 1985). If the fluid trapped is a single phase, this will separate into a vapour (bubble) and a liquid at lower temperatures (two-phase boundary). It has been shown that heating of an inclusion on a heating stage will restore this fluid to single phase at its homogenization temperature. This is taken to be its entrapment temperature which is used to determine the thermal history of a reservoir (Browne, 1978; Shepherd et al., 1985).

A total of 36 fluid inclusion measurements were carried out in this well in both quartz and calcite within the depth range of 1200-1272 m. It was not possible to conclude whether they were of primary or secondary origin. Of these, 21 inclusions are from quartz samples while the remaining 15 are from calcite. Both show a wide range of homogenization temperature ranging from 180 to 330°C and 190 to 300°C for quartz and calcite, respectively (Table 5 and Figure 19). Of the 21 quartz fluid inclusion measurements, more than half show a homogenization temperature above 275°C which is higher than the present formation temperature at the above depth range. Out of 15 measurements in calcite, 11 show more than 245°C, 6 of which have a homogenization temperature of 275°C and above (Figure 19).

TABLE 5: Fluid inclusion homogenization temperatures

Depth (m)	Homogenization temp. (Th) ranges (°C)	No. of fluid inclusions	Type of crystals
1200-1272	180-185	2	Quartz
	190-200	3	Calcite
	205-210	2	Quartz
	210-215	2	Quartz and calcite
	215-240	4	Quartz
	245-275	5	Calcite
	275-285	4	3 calcite and quartz
	285-295	4	Quartz
	300-305	7	3 calcite and 4 quartz
	305-330	3	Quartz

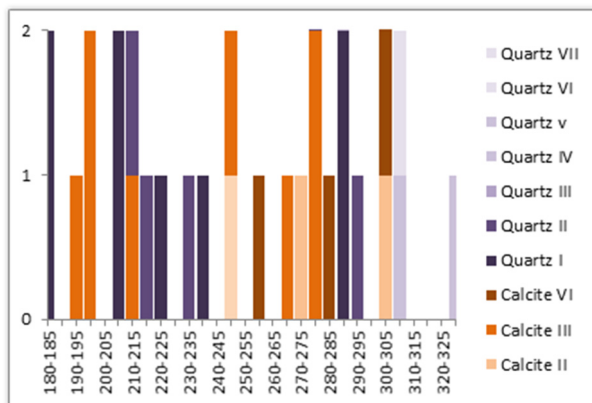


FIGURE 19: Distribution of fluid inclusion homogenization temperatures at 1200-1272 m in well HN-3

6. DISCUSSION

Generally, the stratigraphy of the upper 1272 m of well HN-3 consists mainly of hyaloclastites (basaltic breccia, basaltic tuff and partially crystallized basalt), with some fine- to medium-grained basalt, tuffaceous sediment and basaltic intrusions. Intrusive rocks observed below 1064 m are identified by their massive, compact nature and appear fresh in comparison to the surrounding lithology.

In well HN-3, the deposition of hydrothermal alteration minerals occurs in vesicles and as

replacements of the primary minerals, showing a progressive variation with temperature depicted by temperature-dependent minerals. The minerals undergo chemical reactions with respect to changes in the physico-chemical conditions which resulted in the formation of diverse secondary mineral assemblages, these changes in mineral assemblages being from zeolites and smectite up to quartz, chlorite and prehnite. The alteration temperature can be defined by the first appearance of the temperature dependent alteration minerals (Figure 20). Based on these assessments, three alteration zones have been identified below an unaltered zone. These zones are a zeolite-smectite zone, a mixed-layer clay zone and a chlorite zone.

The sequence of mineral deposition within the well generally shows the hydrothermal system to have evolved from low- to high-temperature conditions, ranging from zeolite-smectite at low temperatures to chlorite at high temperatures. Quartz and prehnite have also been noted as a part of a high-temperature mineral assemblage in well HN-3, along with chlorite.

The dominant aquifers in the studied part of the well (all small) have in most cases no relation to lithological boundaries, although fractures are in some cases associated with them and an intrusion related aquifer occurs at 1245 m. A strong correlation is observed between the aquifers and intensity of alteration of primary minerals, especially at depths 660 and 714 m, and at the intrusion related aquifer at 1245 m. Petrographic studies have shown the highly susceptible primary minerals to be very strongly altered at these depths, showing aquifers as conduits for hydrothermal fluids, facilitating the alteration process.

The distribution of fluid inclusion homogenization temperatures in most of the quartz and some of the calcite samples shows a temperature above 275°C, which is much higher (reaching up to 305°C for both minerals) than the present formation temperature, indicating fluctuations in temperatures (Figure 20).

7. CONCLUSIONS

The following conclusions can be deduced:

- Stratigraphy of the first 1272 m of well HN-3 consists of basaltic hyaloclastite formations (tuff, breccia and partially crystallized basalt), fine- to medium-grained basalt lava flows (mainly olivine-tholeiite) and basaltic intrusions.
- The alteration mineral distribution of well HN-3 shows a trend with depth where low-temperature minerals, like zeolites, are gradually replaced by moderate-temperature minerals like quartz, which are quickly followed by high-temperature mineral assemblages like chlorite and prehnite.

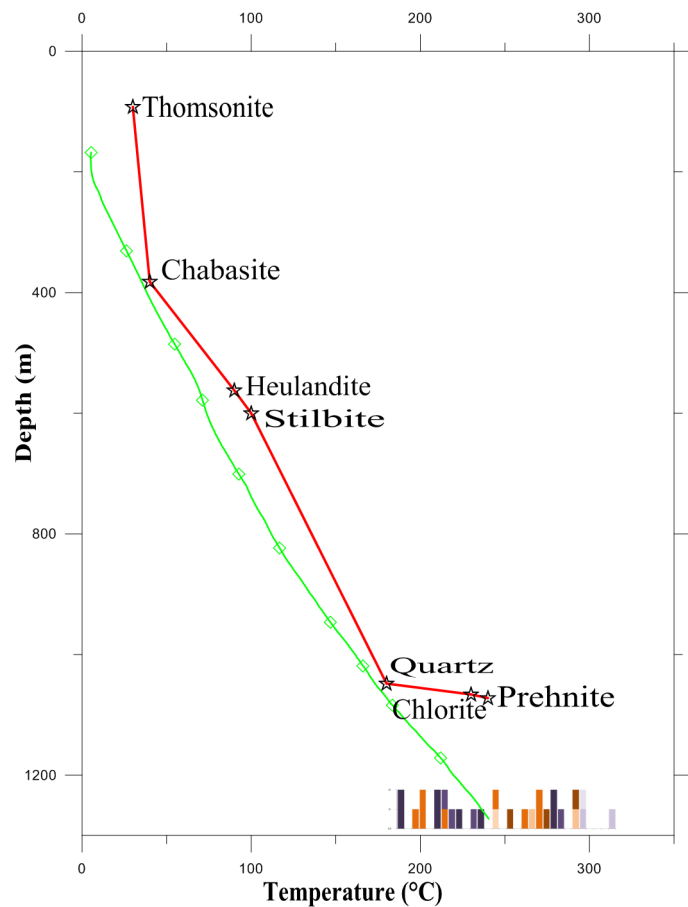


FIGURE 20: Formation (green line) and alteration (red line) temperatures with fluid inclusion homogenization temperatures

- According to the distribution of alteration minerals, four alteration zones were identified and arranged according to key temperature-dependent index minerals: Unaltered zone (< 50°C), a smectite-zeolite zone (< 200°C), a mixed-layer clay zone (200-230°C), and a chlorite zone (230-250°C).
- Most of the aquifers in the studied production part of the well have no obvious relation to lithological boundaries but can, in some cases, be linked to possible fractures by means of increased alteration intensity. The aquifer at 1245 m is associated with an intrusion. Only small aquifers were found.
- Fluid inclusion temperature at 1200-1270 m reaches up to 305°C, which is much higher than the present formation temperature, indicating fluctuations in the system. Comparison of formation temperature and alteration temperature indicates equilibrium in the upper part of the well whereas cooling seems to have occurred below ~1050 m in the well.

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APPENDIX I: XRD clay mineral analysis results from well HN-3

TABLE 1: List of samples with XRD clay mineral analysis results from well HN-3

Sample no.	Depth (m)	D(001) untreated sample	d(001) glycolated sample	d(001) heated sample	d(002)	Mineral
1	712	14.99	17.04	10.2	7.52	Smectite
2	734	14.99	17.23	0	0	Smectite
3	768	14.86	17.17	15	0	Smectite
4	794	14.91	16.33	9.88	0	Smectite
5	830	14.81	16.75	10.02	0	Smectite
6	868	14.6	16.48	9.88	0	Smectite
7	914	14.19	16.78	10.03	7.47	Smectite
8	946	14.65	16.45	9.88	0	Smectite
9	972	14.67	16.42	9.8	0	Smectite
10	1008	13.68	16.45	10	0	MLC
11	1028	14.65	13.58	9.8	0	MLC
12	1064	12.7	15.99	9.8	0	MLC
13	1104	12.63	13.42	9.75	0	MLC
14	1140	14.39	16.42	9.6	7.2	MLC
15	1166	13.07	14.1	9.92	7.10	MLC
16	1198	12.72	13.4	9.74	0	MLC
17	1236	13.99	14.86	9.86	0	Chlorite
18	1262	14.77	13.48	9.88	7.11	Chlorite

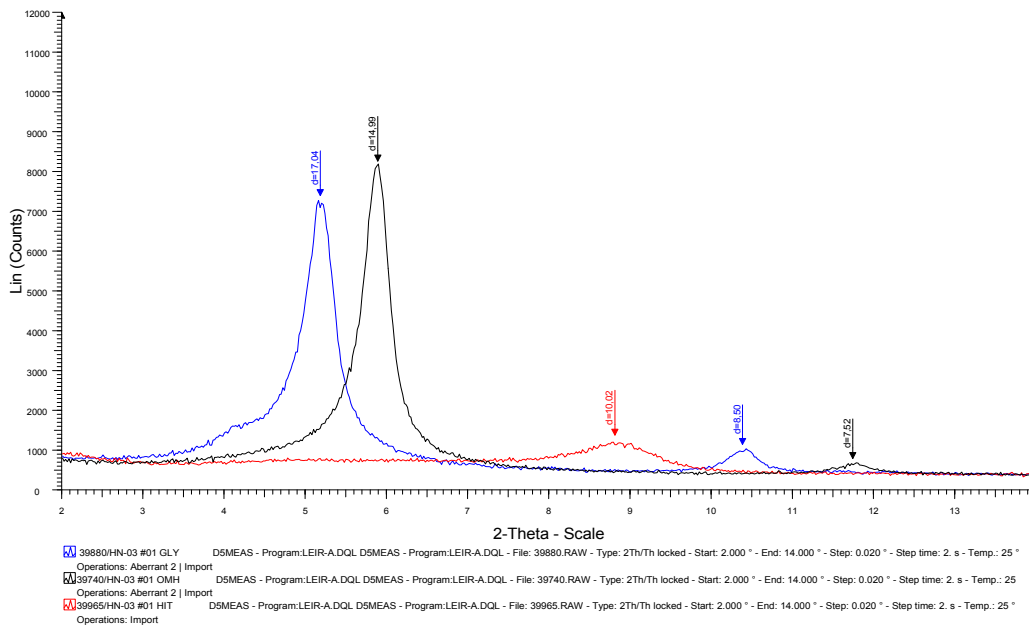


FIGURE 1: HN-03, XRD analysis of a sample 1 from 712 m

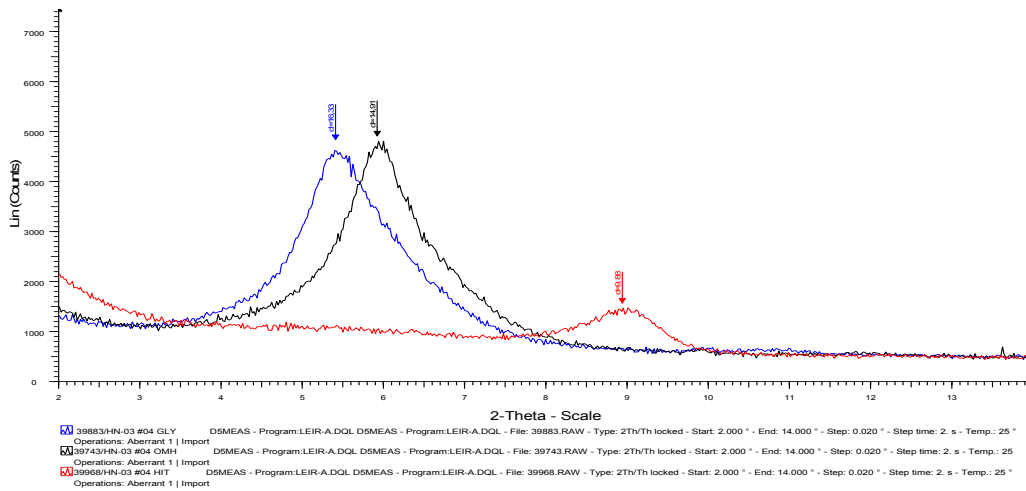


FIGURE 2: HN-03, XRD analysis of a sample 4 from 794 m

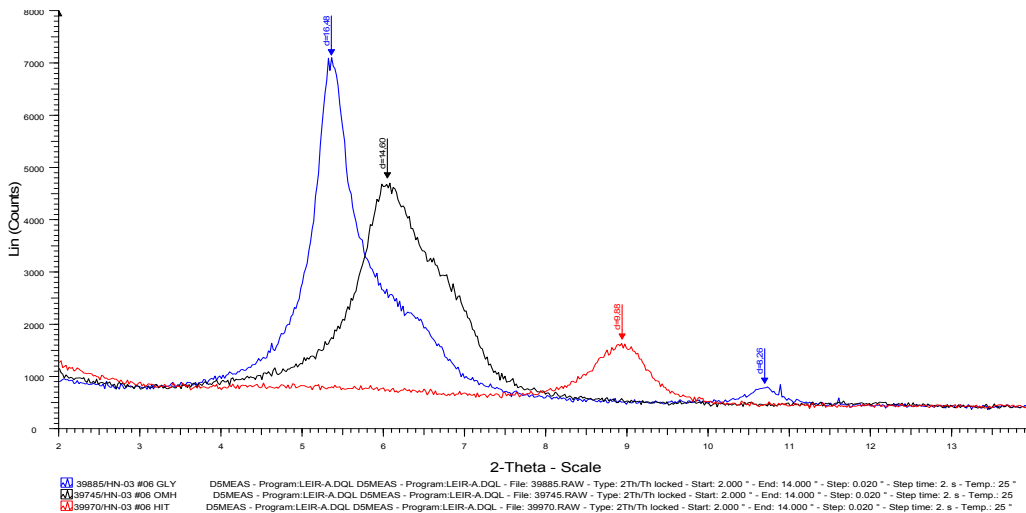


FIGURE 3: HN-03, XRD analysis of a sample 6 from 868 m

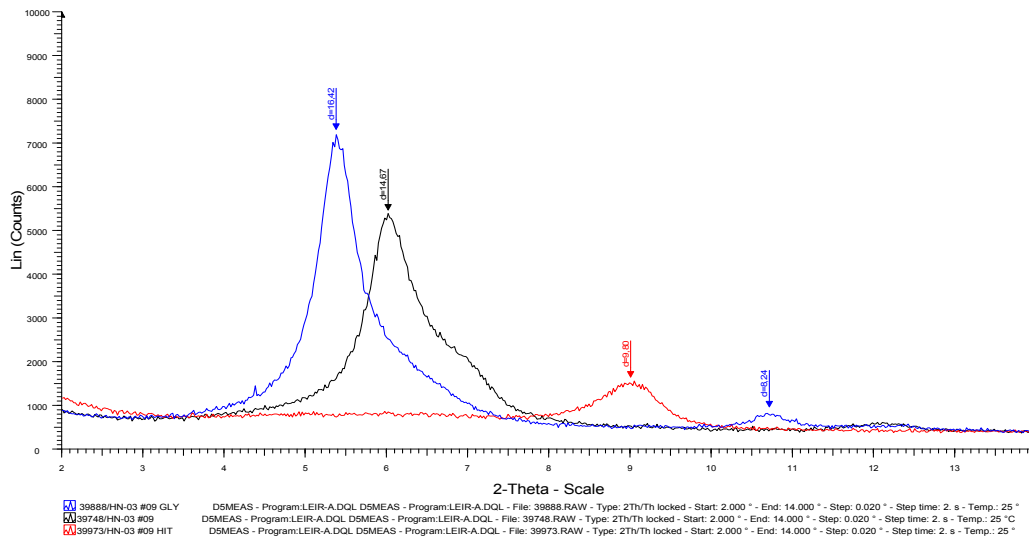


FIGURE 4: HN-03, XRD analysis of a sample 9 from 972 m

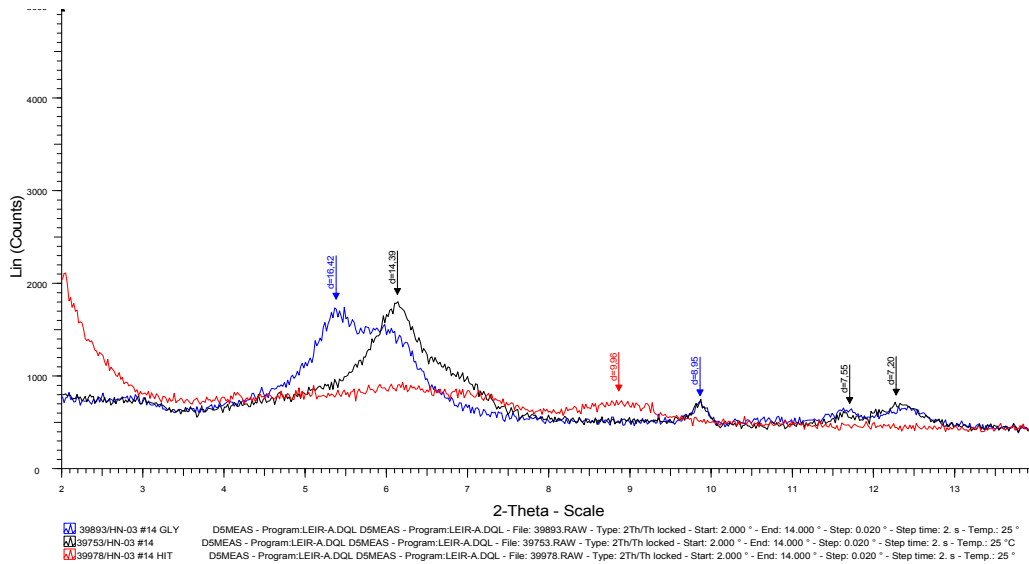


FIGURE 5: HN-03, XRD analysis of a sample 14 from 1140 m
HN-03 #18 (1262 m)

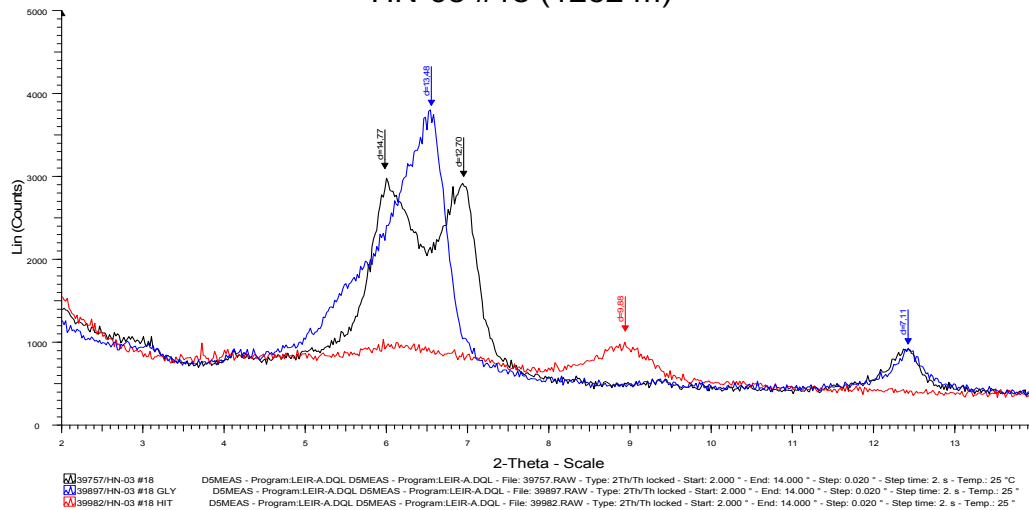


FIGURE 5: HN-03, XRD analysis of a sample 18 from 1262 m