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BOREHOLE GEOLOGY AND HYDROTHERMAL MINERALISATION OF WELL HE-27, HELLISHEIDI GEOTHERMAL FIELD, SW-ICELAND

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ABSTRACT

Well HE-27 is an exploration/production well located on Skardsmýrarfjall Mountain situated in the Hellisheidi high-temperature field, within the Hengill volcanic system, SW Iceland. This directional well is 2116 m deep. An analysis of the uppermost 1000 m is presented in this report. Following binocular and petrographic analysis, it can be concluded that the lithology of well HE-27 comprises six hyaloclastite units based on textural differences. The rock types within these hyaloclastite units are: pillow basalts, basaltic breccia, basaltic tuff and a minor basaltic scoria layer between 854 and 860 m. Fine- to medium-grained tholeiitic basalt intrusions occur below 700 m and a sandstone sediment formation is present between 262-278 m. Eight feeder and water receiving zones were identified using temperature, neutron and caliper logs. These zones show an association with circulation loss zones and lithological boundaries in the upper levels and occur in relation to intrusive bodies at depth. Resistivity and gamma logs generally designate the alteration zones. Alteration within HE-27 is primarily influenced by the temperature of the formation which is approximately 248°C, based on the occurrence of epidote and wollastonite as alteration products. Five alteration zones were identified: unaltered, smectite/zeolite, mixed-layer clay, chlorite and chlorite/epidote zone, established by the alteration mineralogy in the uppermost 1000 m of the well. Based on the sequencing of the alteration minerals, there is a strong indication that the area experienced a heating period. Nonetheless, the occurrence of calcite as a late deposit in the sequences implies subsequent cooling. Fluid inclusion analyses, however, indicate a renewed heating episode after the period of cooling. A brief comparison of hydrothermal minerals in well HE-27 and OW-908A in Olkaria, Kenya emphasise the role of temperature in all hydrothermal systems.

1. INTRODUCTION

The Hengill central volcano, located in southwest Iceland, is home to the Hellisheidi high-temperature field, which is situated in the southern sector of the 110 km^2 Hengill low-resistivity anomaly. Hellisheidi comprises four potential geothermal fields, Skardsmýrarfjall, Hverahlíd, Gráuhnúkar and Reykjafell. Well HE-27 is located on Mt. Skardsmýrarfjall which is the northernmost geothermal area of Hellisheidi. The comparison of alteration and formation temperatures indicates minor cooling on the western side of Skardsmýrarfjall, as well as a cooling front from the east between Skardsmýrarfjall and Hverahlíd (Helgadóttir et al., 2010).

Well HE-27 is one of the numerous production wells that have been drilled in Hengill since 1985 in an effort to meet the increasing demand for electricity and hot water for space heating in the industrial and domestic sectors of Iceland. HE-27 is a directional well drilled to 2116 m measured depth; cuttings were sampled at 2 m intervals for analysis, identification of subsurface formations, the location of aquifers or feed zones and the determination of the alteration temperature. Binocular analysis of well cuttings, petrographic analysis of thin sections, X-ray diffractometer tests, fluid inclusion analysis and interpretation of geophysical logs are the analytical methods used to study the upper 1000 m of the well. The samples in this high-temperature field are characterised by several vesicle fillings which can be used to interpret the thermal history of the area.

The study presented here is carried out as a requirement of the fulfilment of the United Nations University Geothermal Training Programme six month course, in the year 2010.

2. REGIONAL GEOLOGICAL SETTING AND TECTONICS

2.1 Geology

Iceland is located at latitude 65° N in the North Atlantic Ocean, just south of the Arctic Circle, straddling the Mid-Atlantic Ridge between the Eurasian and North-American plates. These continental plates are drifting apart approximately 1 cm per year in each direction and the island is the largest land mass on the ridge.

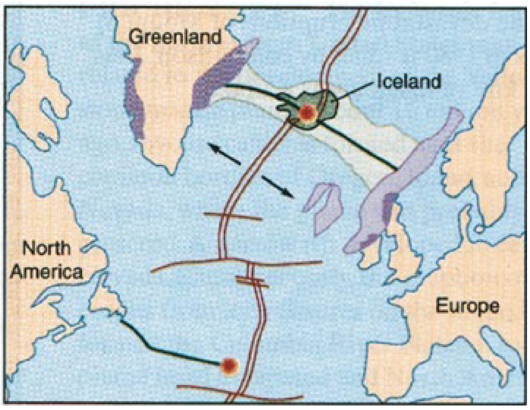

FIGURE 1: Location of Iceland; the orange circle indicates the location of the Iceland plume

Furthermore, the island is located above a mantle plume which is believed to have caused the formation of Iceland itself through the emission of volcanic material, created by rifting and crustal accretion along the NE-SW axial rift system. Therefore, the geology is such that the oldest rocks are found to the east and the west and the youngest in the central rifting zone (Figure 1; Saemundsson, 1979; Hardarson et al., 1997). As a result of its unique location, Iceland is one of the most active and productive subaerial regions on Earth, with an eruption frequency of more than 20 events per century and a magma output rate of about 8 km^3 per century in historic times, i.e. over the last 1100 years (Thórdarson and Höskuldsson, 2008).

The surface geology (Figure 2) is entirely made up of volcanic rocks with basalts being 80-85% of the volcanic pile, and acid and intermediate rocks 10%. The amount of sediments of volcanic origin is 5-10% in a typical Tertiary lava pile, but may locally be higher in Quaternary rocks (Saemundsson, 1979). The basalts themselves can be classified into three main types: the compound flows of olivine tholeiite, simple tholeiite (with little or no olivine) and porphyritic basalt with plagioclase and pyroxene. Olivine tholeiite morphologically produces pahoehoe lava fields, while olivine poor tholeiite gives rise to aa lava fields (Saemundsson, 1979). Various investigations have shown that the type of effusion is dependent upon the volcanic zone from which it emerges. The zones, volcanic rift zones and volcanic flank zones are categorised according to the degree of mantle melting. The former are associated with tholeiitic basalts and their derivatives whereas the latter give rise to transitional-alkaline through to alkaline rocks.

2.2 Tectonic setting

Iceland is the only subaerial part of the Mid-Atlantic Ridge and, while the North Atlantic is spreading symmetrically away from the mid-ocean ridge (Hjartarson, 2009), the ridge itself migrates over the plume and repeatedly shifts its spreading axis through rift jumping (Saemundsson, 1979; Hardarson et al., 1997). The shift of the plate boundary can be explained as a response to a gradual westward drift of the boundary away from the central plume upwelling under Iceland. Being at the intersection of two major structures, namely the Mid-Atlantic Ridge and the Greenland – Iceland – Faeroes Ridge (Figure 3), the island is home to approximately 60% of the world's fissure eruptions.

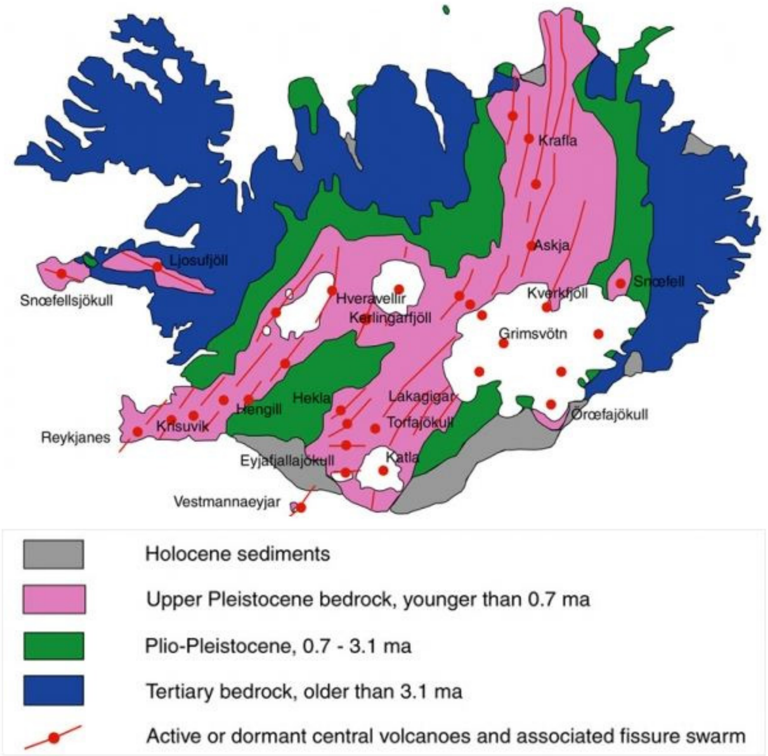

FIGURE 2: Surface geology of Iceland (modified from Jóhannesson and Saemundsson, 1998); the ages of the rocks increase perpendicular to the rift axis (Pálmason and Saemundsson, 1974)

The Mid-Atlantic Ridge is a constructive plate boundary while the Greenland-Iceland-Faeroes Ridge is thought to be the trail of the Icelandic mantle plume, which has been active from the time of the opening of the North Atlantic Ocean 60 million years ago to the present (e.g. Sigmundsson and Saemundsson, 2008). The mantle plume is seen to be located below Central East Iceland (Figures 1 and 3), close to the volcanic rift zone which crosses Iceland from southwest to northeast and is divided into two parallel branches in South Iceland (Pálmason and Saemundsson, 1974). In South and North Iceland, the Mid-Atlantic Ridge has been displaced to the east by transform faults (Figure 3) which are defined as fracture zones. The southern fracture zone is called the South Iceland Seismic Zone (SISZ) while the northern one is called the Tjörnes Fracture Zone TFZ, shown in Figure 3.

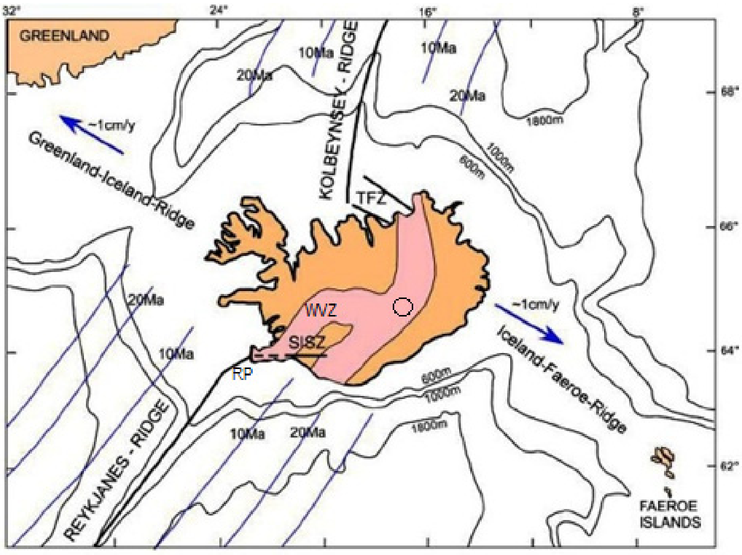

FIGURE 3: Tectonic systems in Iceland (modified from Marosvolgyi, 2009); the location of the Iceland mantle plume is outlined by a black circle

The volcanic rift zone is, thus, a zone of active rifting and volcanism and is characterised by well developed extensional structures such as tension fractures, normal faults and grabens with dykes and normal faults occurring at deeper levels (Gudmundsson, 1998). In their paper, Dauteuil et al. (2005) pointed out that the active zone has at least doubled in width since 200,000 years BP, and forms a unique wide area ranging from 200 km in the south to 120 km in the north. According to Gudmundsson (1992) the swarms may be up to 5-10 km wide and 40-80 km long and make up most of the volcanic rift zone. The volcanic rift zone is about one-third of the area of Iceland.

2.3 Hengill volcano

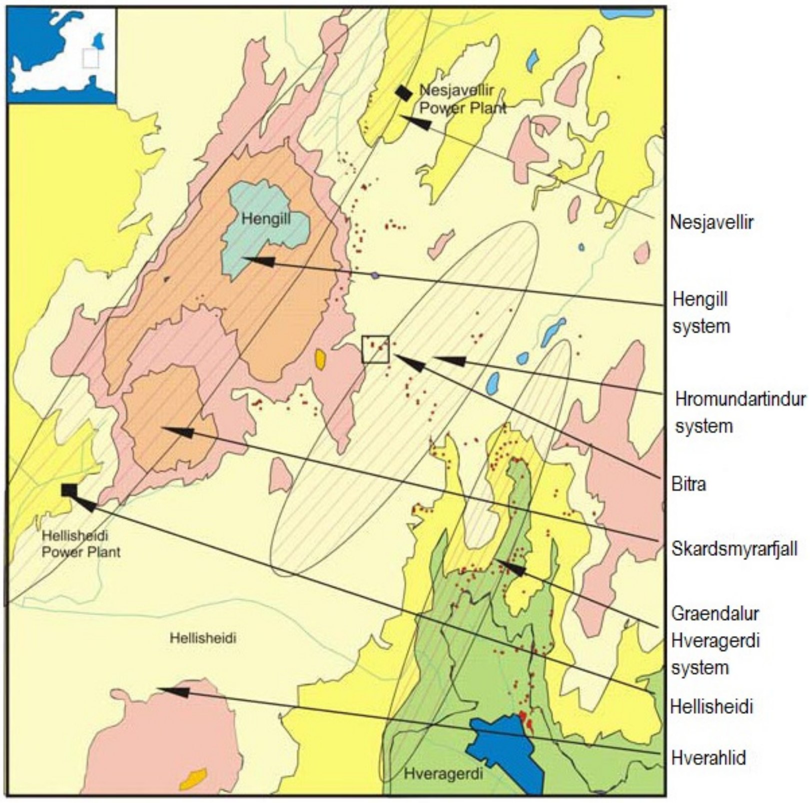

FIGURE 4: The greater Hengill volcanic complex (Pendon, 2006)

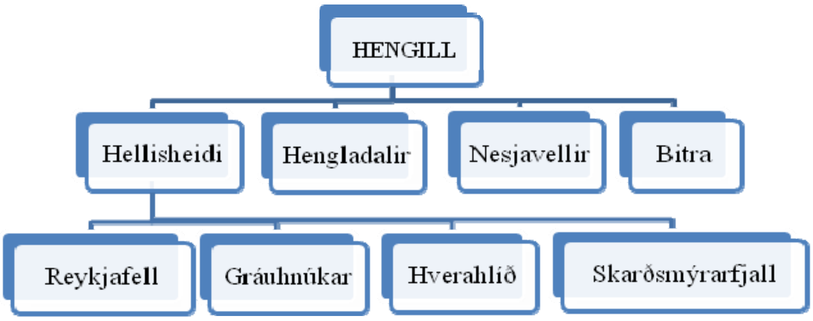

FIGURE 5: Classification of the Hengill geothermal fields

The Hengill central volcano is situated approximately 30 km from the capital of Reykjavik and is a component of a 40-60 km long fissure/fault swarm, part of the Western volcanic zone. The last eruption is considered to have occurred 2000 years ago and geothermal activity shows that the system is still active. An age of about 0.4 million years has been proposed for the Hengill central volcano which means the geothermal system is slightly younger (Franzson et al., 2010).

The greater Hengill volcanic complex is related to three distinct volcanic systems. The oldest system is the eroded Pleistocene Hveragerdi – Grensdalur (Graendalur) volcanic system. Hrómundartindur is the smallest of the three systems in which the latest recorded eruption dates back to 10,000 years ago (Foulger and Toomey, 1989). These systems are extinct in terms of volcanic activity but still active seismically and host geothermal reservoirs. Hengill is the youngest and most active of the three volcanic and geothermal complexes. There are four geothermal fields in Hengill, one of which is Hellisheidi, located southwest of the volcano and home to Skardsmýrarfjall Mountain (Figures 4 and 5), the area of interest in this study.

2.4 Geological geophysical and geochemical aspects the of Skardsmýrarfjall volcano

Skardsmýrarfjall Mountain is part of the Hellisheidi geothermal field (Figures 4 and 5) and consists primarily of hyaloclastites and Postglacial lava flows with intrusions of basaltic and intermediate composition. The hyaloclastite formations have further been classified into seven different formations based on their texture, crystallinity and compositional variation (Gebrehiwot, 2010). Despite having fissures trending NNE-SSW, the volcano exhibits few surface manifestations. Studies in the area have further shown that permeability is mainly associated with stratigraphic boundaries for the upper aquifers, whilst the deeper ones are associated with intrusive boundaries. The comparison of alteration and formation temperatures seems to indicate minor cooling on the western side of Skardsmýrarfjall as well as a cooling front from the east between Skardsmýrarfjall and Hverahlíd (Figure 4, Helgadóttir et al., 2010).

At shallow depths the Hengill area shows low resistivity in a WNW–ESE elongated form (Björnsson et al., 1986; Árnason et al., 2010). At this depth the resistivity is low and is mainly controlled by a clay cap. The underlying higher resistivity is interpreted as reflecting a transition in dominant alteration minerals from low-temperature clays (smectite and mixed-layer clays) to the formation of chlorite and less conductive alteration assemblages (Árnason et. al., 2000). The NW-SE oriented, low-resistivity anomaly at 3-9 km under and to the southeast of Mt. Hengill is found where intense seismic activity associated with transform tectonics occurs. Since no attenuation of S-waves is observed under the Hengill area, the deep conductors are believed to reflect hot, solidified intrusions that are heat sources for the geothermal system above (Árnason et al., 2010).

According to geochemical and isotopic studies, the Hellisheidi system is younger than the Nesjavellir system to the northeast (Figure 4). At Nesjavellir, the fluid is richer in ^{18}O and chemically more mature than at Hellisheidi, due to more intense water-rock interaction. The calculated quartz geothermometer for both fields indicates a conservative temperature range, between 189 and 255° C, but the Na-K geothermometer indicates higher temperatures between 210 and 290°C (Mutonga, 2007).

3. SAMPLING METHODS AND ANALYSIS

Sampling of well cuttings is carried out at the rig concurrently with drilling operations. The cutting samples, collected at 2 m intervals are analysed on site in an effort to obtain information concerning the alteration temperatures. This information is crucial, particularly to the driller, as it acts as a guide and enables avoidance of situations such as an imminent eruption, collapse of formations, sticking of the drill strings and rapid wearing of bits. The samples are analysed further in the lab to provide ample information regarding the history and future of the geothermal system of interest.

Besides visual examination, the cuttings were first subjected to binocular microscope analysis. The primary objective of this examination was to identify lithology and, if altered, grade it accordingly. Intrusives, veins, vein fillings and many alteration minerals were identified at this stage although some require further and more detailed analysis. Use of acid to identify carbonates is only applicable at this phase but one must be careful as drilling cement is rich in calcium carbonate. It is imperative to wash the samples prior to analysis to remove dust particles and, most importantly, to enhance the visibility of certain obscured features.

Petrographic microscopic analysis was carried out to fine-tune, or clarify, results obtained from the binocular analysis. The sample was cut and polished to approximately 2μm and glued on glass using epoxy. This analysis gives specific information particularly on mineral types and sequences in vein and vesicle fillings. Uncertainties encountered during binocular investigation concerning rock type, texture, porosity etc. can be clarified during this analysis. In order to further understand the alteration zones in a geothermal field, X-ray diffractometer (XRD) tests were carried out. This analysis is of key importance towards classifying the various types of clay which, in turn, provide information on the alteration temperatures. The analysis of twenty samples from well HE-27 and their interpretations are part of this study and the results are seen in Appendices I and II.

The history and predictions of the behaviour of a geothermal system are crucial for planning in all geothermal fields. Although there are other methods that can be used for these predictions, fluid inclusion analysis is fairly simple as it is specific. It involves the analysis of trapped fluids in mineral vesicles, thereby giving information on the temperatures at which the minerals formed.

The data obtained from all the above analyses were compiled and plotted using LogPlot 2007 (RockWare Inc., 2007). The software allows one to display all geological and geotechnical data as a graphic model for easy reference and comparison purposes.

4. BOREHOLE GEOLOGY

4.1 Drilling at Skardsmýrarfjall

A total of fourteen wells are located at Skardsmýrarfjall Mountain; well HE-27 was the third to be drilled at this location after HE-23 and HE-24. It was drilled for further studies of the Hellisheidi geothermal system and more so to explore the eastern part of Mt. Skardsmýrarfjall.

4.2 Drilling well HE-27

Well HE-27, as highlighted in Figure 6, is located at coordinates 64°03'00''North and 21°19'55''West at an elevation of 532 m above sea level. The primary objective of the well was steam production for the Hellisheidi power station. The 6 MWe well was drilled to 2116 m in a N70°E direction dipping at an angle 30°, targeting the NE-SE trending structures in the Skardsmýrarfjall formation (Figure 6).

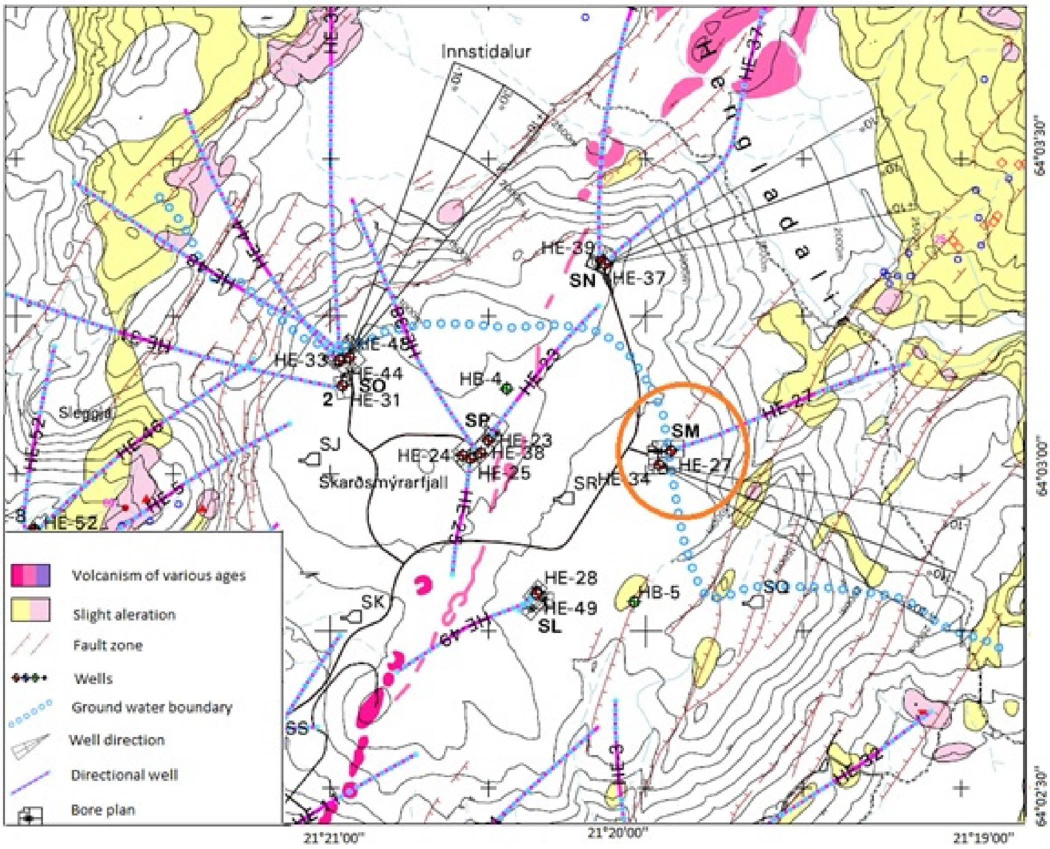

FIGURE 6: Location and direction of HE-27, extrapolated from depth to the surface according to gyro measurements (ISOR map, unpublished)

The drilling process of HE-27 was initially carried out without much difficulty but with several severe circulation losses, both at shallow levels and at depth, which forced the crew to carry out blind drilling. Geysir, a top drive rig, was used for the entire drilling operation of the well. Drilling was carried out in four phases based on the casing type and bit size (Table 1). Drilling of HE-27 was carried out over a total of 49 days.

TABLE 1: Drilling data

Rig	**Drilling phase**	**Drilling bit**	**Drilled depth**	**Casing size**	**Casing shoe**
Geysir	Pre-drilling	26½"	88 m	22½"	84.14 m
Geysir	Phase one	21"	313 m	18⅝"	312.18 m
Geysir	Phase two	17½"	753 m	13⅜"	750.42 m
Geysir	Phase three	12¼"	2116 m	No casing	

As shown in Figure 7, the pre-drilling phase started on the 30th of September 2006 and ended on the 2nd of October four days later, due to well collapse and sticking of the drill string. Circulation loss occurred from 72 to 80 m so drilling to casing depth was carried out blindly. A 26" bit was used for this phase; the 22½ surface casing was set with 34 m^3 of cement.

Several days after the cement had set, phase one of the drilling was initiated on the 12th of October using a 21" bit. Phase one was uneventful and reached a casing depth of 314 m. Geophysical and gyro loggings were under-taken and the 18⅝" anchor casing was set with 27 tons of cement.

Phase two commenced on the 18th of October, drilling a total of 440 m using a 17½" bit. To control the development of azimuth and inclination, drilling was regularly stopped for gyro measurements. At 700 m the well started

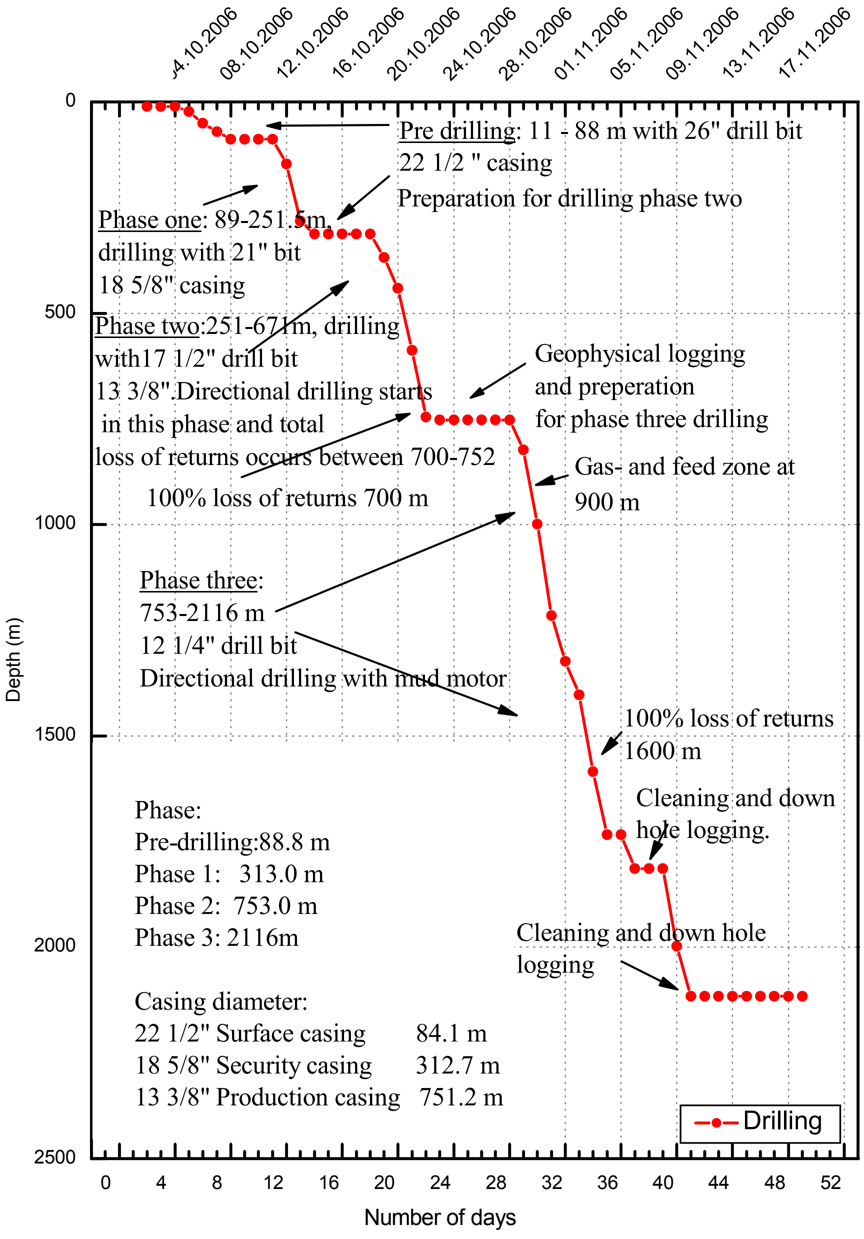

FIGURE 7: Drilling program for well HE-27
(ISOR data, unpublished)

losing water at a rate of approximately 50 l/s with a total circulation loss down to 754 m. Once again drilling was carried out blindly to a casing depth of 753 m. Geophysical logging was carried out again and, subsequently, the 13⅜" production casing was set with 94.6 m^3 of cement to the casing shoe at 750 m.

Phase three began on October 27th with air drilling using a12¼" drill bit; no circulation losses were encountered down to a depth of 881 m. An attempt to decrease the pumping rate of water from 45 l/s to 30 l/s did not facilitate the return of cuttings. The pumping rate of water was reduced to 43 l/s and the pumping rate of air increased from 30 m^3/min to approximately 46 m^3/min with no success. Polymer pills and a further increase of pumped air did not ensure the return of cuttings and blind drilling continued.

Finally, at a depth of 1267-1275 m, cuttings returned to the surface after continued efforts to increase air pumping. Total loss was again encountered at 1323 m and the torque of the bit was noted to be unusually high which forced the crew to pull out the string to inspect the bit. Fortunately, everything was in order and this was an opportunity to perform temperature and gyro measurements. Gyro measurements revealed that the inclination had increased from 2° up to 36° at 1270 m. Drilling continued using 45 l/s of water with no success in recovering cuttings except in a few incidences where partial returns were encountered. On November 4th at a depth of 1800 m, the torque of the bit became unusually high and once more temperature and gyro measurements were executed. Azimuth and inclination showed that everything was in order and the well was proceeding in the planned direction and angle. The well was completed on the 9th of November at a depth of 2116 m; geophysical loggings were then performed.

Complications due to the final temperature measurement kept the Geysir rig an extra week on the well pad giving a total number of 49 days for the entire process. A liner was not installed in the well.

4.3 Stratigraphy of HE-27

The entire lithology of well HE-27 (shown in Figure 8) can be summarised as being primarily hyaloclastite with a few basaltic lava flows and one minor sedimentary formation. The classification of the different lithologies into hyaloclastite formations is, in this case, based on the texture of the rock being aphyric or porphyritic. The classification of the hyaloclastites in the Skardsmýrarfjall formation was previously achieved using not only texture but also composition (whether olivine tholeiitic or tholeiitic) and crystallinity. The following description of the rock formation is based on binocular observations aided by petrographic thin sections and XRD analysis.

Skardsmýrarfjall formation (12-262 m)
12-262 m: *Glassy basalt*. The lithostratigraphic column is largely made up of dark grey to brownish grey fine-grained, feldspar and olivine phyric rock, with glass forming part of the assemblage. Quartz veins and vesicles are also physical characteristics of the pillow basalt. The low-temperature zeolites identified include chabasite, stilbite, and analcime although thomsonite is also evident in some samples. Amorphous silica and chalcedony and sometimes opal are seen to be deposited on some surfaces and within a number of vesicles. The basalt is fairly oxidised with limonite and siderite forming the common iron compounds. Calcite is the major alteration mineral and the rock is seen to have a porosity of up to 50%. X-ray diffraction results, as well as thin sections, indicate that this lithology is within the smectite clay zone. A circulation loss zone is present between 72-78 m. The glassy basalt between 250-262 m is dark and grades to an almost completely glassy variety at depth. This may represent the outer margins of the pillows as the porphyritic interior is typically bound by glassy margins.

Sandstone (262-278 m)
Brown sandstone which grades from a fine- to coarse-grained variety interrupts the volcanic sequence.

A siliceous cement is what appears to bind the rock which also contains some clays and, to a lesser amount, magnetite. A calcite lining is apparent around some of the particles in thin section and the porosity is estimated to be approximately 30%. The deposition of sediments in between hyaloclastite formations implies a period of decreased volcanic activity.

Hyaloclastite formation II (278-450 m)

278-450 m: *Porphyritic tuff*. A large proportion of the samples from between 312 and 326 m were composed of drilling cement as compared to the geological samples but specks of basaltic tuff can still be identified. Dark brown tuff is, hence, the only rock unit in this large lithologic column although the colour changes to green towards the bottom. The tuff is rich in feldspar phenocrysts which increased significantly near 450 m. The brown sample is rich in calcite which is present as crystals and

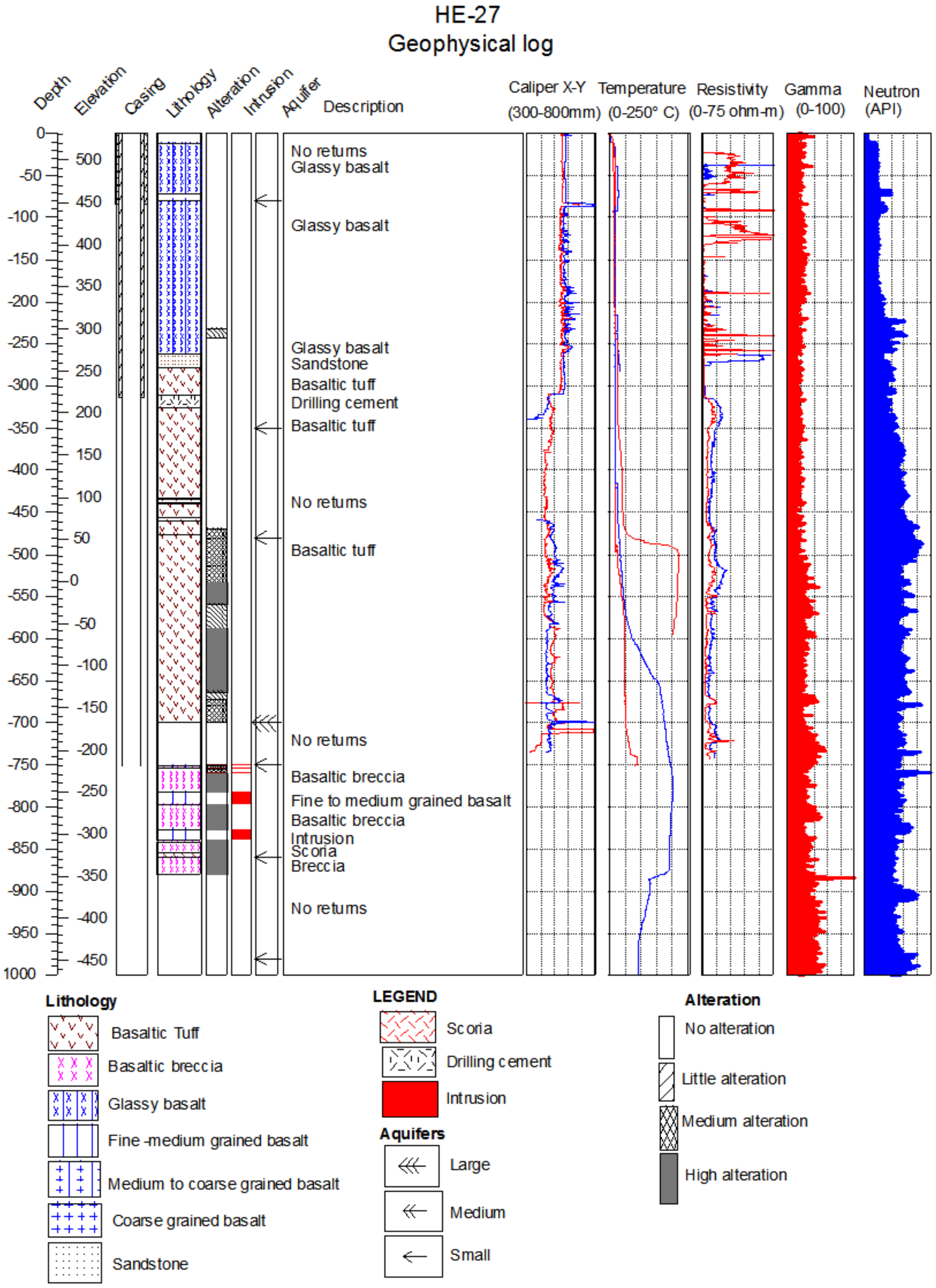

FIGURE 8: Geophysical logs of well HE-27

sometimes occurring in vesicles on top of zeolites such as mesolite. Palagonisation outlines are obvious along the vesicles and form sequences with the fine grained clay. Smectite clays are evident in thin section and appear to be overlain by calcite. The tuff in the upper zones is fairly porous as many of the vesicles are void. However, this scenario changes with the increase in calcite deposition.

Hyaloclastite formation III (450-472 m)
450-472 m: *Aphyric tuff.* Although the basaltic tuff greatly resembles the one above in colour, the feldspar phenocrysts are notably absent. A minor loss zone occurs between 456-460 m. Calcite heulandite and analcime are present within this formation.

Hyaloclastite formation IV (472-562 m)
472-500 m: *Porphyritic basaltic tuff.* The samples are greenish grey, the glassy fragments being green while the few crystalline ones are grey. The tuff is altered and contains significant amounts of calcite and amorphous silica. Calcite in thin section appears to be layered and in most sequences succeeding smectite and mixed-layer clays.

500-562 m: *Porphyritic highly altered tuff.* Although there are a few crystalline fragments, the rock is still mostly glassy and is therefore classified as tuff. The brown and greenish white mix exhibits high alteration and large amounts of calcite and pyrite. The vesicles are filled with calcite, quartz and mixed-layer clays.

Hyaloclastite formation V (562-700 m)
562-700 m: *Aphyric tuff rich basaltic breccia.* Binocular analyses indicate that the samples are a mixture of brown tuff and dark brown fine- to medium-grained basalt. The degree of alteration is much less as is the amount of calcite and pyrite. The majority of the vesicles are filled with calcite, quartz, prehnite and mixed-layer clays. The sequences show quartz being the last to be formed after calcite and sometimes after clays. Minor loss zones occur at 600-604 and 608-612 m. Prehnite and traces of epidote are visible both in the cuttings and in thin sections. At deeper levels (654 m), the alteration sequence shows prehnite depositing after quartz. The breccia is slightly porphyritic to non-porphyritic unlike the variety in the upper formation.

700-752 m *loss zone*: This major loss zone is an indication of permeability.

752-754 m: *Olivine tholeiite basaltic intrusion.* The sample is fresh, black in colour, fine- to medium-grained and contains olivine and pyroxene crystals.

Hyaloclastite formation VI (754-854 m)
754-782 m: *Porphyritic basaltic breccias.* Binocular analysis indicates that the rock is mostly crystalline with minor tuff fragments. The crystalline fragments are of two types, a moderately altered one, and a much fresher one which contains olivine and pyroxene evident in thin section. This indicates that the sample contains an olivine tholeiite intrusion. Quartz and prehnite form most of the vein and vesicle fillings and are the last to be formed in the sequences. Epidote is seen to occur with sphene, also identified in the calcite rich breccia.

782-798 m: *Fine- to medium-grained basalt.* The crystalline fragments are grey but appear to be discoloured to a greenish red tinge. Slight oxidation is evident and so are specks of epidote.

798-826 m: *Basaltic breccia.* This breccia formation is very similar to the one encountered at 754 m. The minerals identified in the upper lithology are similar and the olivine tholeiite intrusive is also present.

826-840 m: *Fine- to medium-grained basalt.* The intrusion at this location resembles the one at 782 m except it is much thicker and some fresh glass is evident in thin section. Epidote, sphene and wairakite are the key hydrothermal minerals in the samples.

840-854 m: *Aphyric basaltic breccia*. The formation is whitish green in colour mixed with dark and fresh olivine tholeiite intrusion. Alteration is high and calcite is in abundance. Epidote and quartz occur in the groundmass and oxidation increases towards the bottom.

Hyaloclastite formation VII (854-880 m)
854-860: *Scoria.* Alteration and oxidation are very high in this vesicular formation. The rock is fine grained although some fragments are poorly crystallized. Olivine tholeiite intrusion specks are evident in the formation; epidote, wairakite, prehnite quartz and clays are also perceptible.

860-880: *Tuff rich porphyritic breccia.* The cuttings consist mainly of poorly crystallized fragments, altered glass and fresh intrusive fragments. Calcite, clays, prehnite, epidote and sphene are the alteration minerals identified.

4.4 Intrusions

An intrusion is an emplacement of magma into pre-existing rock. It is for this reason that the crystals in intrusive rocks (at depth) are large since the subterranean magma cools slowly giving time for crystal growth. If cooled at shallow depths, the rocks tend to be more finely crystalline or even aphanitic. The intrusions also show oxidation where heating effects have been prominent along their boundaries (Gebrehiwot, 2010). Furthermore, the rock is more discernible as it appears less altered than the surrounding lithology.

As mentioned earlier, intrusions form a source of permeability at depth in most Icelandic geothermal systems; Hellisheidi is not excluded. The composition of intrusions in the Skardsmýrarfjall formation is mainly basaltic although intermediate composition intrusions occur at depth with textures varying from fine over medium to coarse grained. In the upper 1000 m, well HE-27 is characterised by multiple intrusions below 700 m. These are at 750-752, 754-756, 758-760, 782-798 and 826-840 m. The properties of the intrusive rock include a dark grey colour, no vesicles and they appear shiny and generally fresher than other cuttings. Thin section analyses show that the intrusive rock is plagioclase porphyritic and has, in addition, phenocrysts of olivine and pyroxene.

4.5 Interpretation of the geophysical logs

Caliper: These logs are used to identify fractures and possible water-producing or receiving zones and to correct other geophysical logs for changes in the borehole diameter (Senior et al., 2005). Furthermore, caliper logs can be used to avert trouble by identifying cavings within the well column. As can be noted in Figure 8, there are two peaks that almost coincide with the circulation loss zones at 72-80 m and at 700-752 m. These can be interpreted as possible water receiving zones.

Temperature: Temperature logs are used to identify possible aquifers. The effect of water flowing into or out of the well from the formation causes a sudden increase or decrease in the temperature and it may affect the circulating water pressure. The temperature logs indicate a large aquifer at 480 and 700 m and small ones at approximately 350, 750 and 1000 m. In Figure 9, three small cold aquifers, or water receiving zones, can be identified at about 80, 860 and 980 m, indicated by the decline in temperature.

Resistivity: The resistivity and conductivity of objects are inversely proportional. 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 8 shows lower resistivity in the aquifer zones.

Gamma: Natural gamma radiation is 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 SiO_2 content of the rock and can, therefore, be used to identify rocks of evolved compositions (e.g. Mostaghel, 1999). The gamma results show low values in the fresh rocks and a general increase in the alteration zone.

Neutron: Neutron logs respond to the fundamental formation property of hydrogen richness. If all of the formation's hydrogen is contained in the form of liquids, and if these liquids completely occupy the total pore volume, hydrogen richness is an index to porosity. Hence, a neutron log can be used to determine the porosity (Mostaghel, 1999). The neutron logs show an increase around the aquifer located at 450 m and, as such, the pores in the rocks are probably saturated.

4.6 Aquifers

The temperature logs (Figure 9) are the most common measurements run in any given well because they suit diverse purposes. However, they are mainly used to locate aquifers (e.g. Danielsen, 2010). Furthermore, aquifers can be located using circulation loss zones, geophysical logs as well as hydrothermal alteration. Drilling data, such as the rate of penetration or the pumping rate, may also be used. A high rate of penetration during drilling may be an indicator of aquifers, whereas a drop in pipe pressure, especially a total loss of circulation in the zones, may indicate the same (Koestono, 2007).

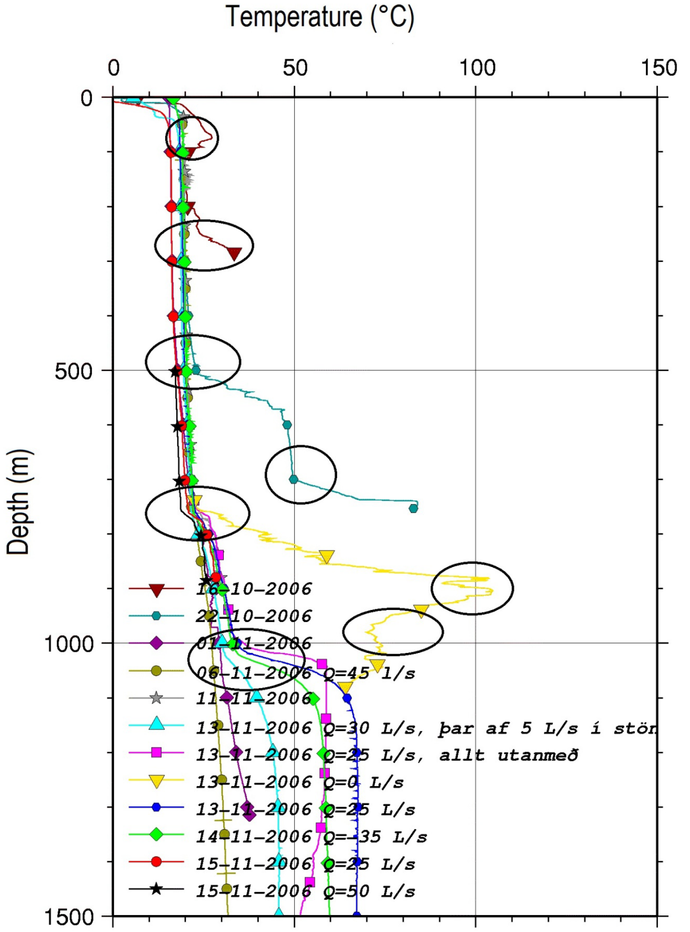

FIGURE 9: Temperature logs of well HE-27 (ISOR Data)

High permeability zones

80 m: A cold feeder point seen to be within a circulation loss zone indicates a temperature decline of approximately 10°C. This permeable zone is within the cold groundwater system and is cased off to avoid cooling of the well.

350-370 m: A temperature rise of almost 15°C is noted in two temperature profiles carried out during drilling (Figure 9). However, this aquifer is cased off because its temperature is not high enough and would ultimately cool the well.

470-480 m: The boundary between the tuff and altered breccia is very clear in the sample cuttings and is evidently a high permeability conduit. This aquifer is the largest in well HE-27; the fluid temperature rises above 200°C.

700 m: At this depth, a significant circulation loss was encountered; an aquifer is inferred to be associated with this loss. This is one of the main aquifers in the uppermost 1000 m of the well.

750 m: This aquifer is located within the loss zone and temperatures measured while pumping 45 l/s of water showed temperature up to almost 110°C.

860 m: This aquifer is associated with a lithological boundary as well as an intrusion.

980 m: A major loss of circulation was encountered at this depth and the temperature drop can be associated with a water receiving zone.

1000 m: The temperature logs indicate a temperature increase from 1000 m; this aquifer is located within a circulation loss zone.

5. HYDROTHERMAL ALTERATION

It was the interest in epigenetic ore deposits that led to the recognition that hot fluids are important in their deposition and that wall rocks are almost always highly altered. Hydrothermal alteration is simply the process by which the mineralogy, chemistry and texture of a rock changes due to interaction with hot ground fluids (gases included). The physical changes in the rock are in some incidences very conspicuous, especially where there is colour change (due to bleaching or otherwise), deposition of minerals, or development and obliteration of porosity and leaching. The changes in mineralogy and chemistry are far more intricate as re-crystallization is involved, resulting in chemical systems of extreme complexity. Most geothermal systems display unique hydrothermal alteration characteristics based on several factors. The composition of the rock and the composition of the solution are the two principals involved in the process. Other important factors that come into play include the temperature and pressure of the geothermal system, the structures within the system, composition of magma (as this ultimately influences the types of gases) and the age of the system/the duration of the water rock interaction.

In all geothermal fields, the proper understanding of hydrothermal alteration is crucial as it is this information that gives the general picture of the geothermal system, its history and possibly its future. Moreover, hydrothermal minerals can be useful as geothermometers and thus assisting in determining the depth of the production casing whilst drilling. Further still, these minerals are also used 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 the hydrology (Reyes, 1990).

In Iceland, for example, hydrothermal alteration is bracketed by either the replacement of primary components in the rocks by alteration minerals or by the precipitation of alteration minerals into voids in the rock (Gebrehiwot, 2010). Detailed studies have contributed to the identification of various hydrothermal alteration zones and these are closely linked to the formation temperature (Figure 10). These zones are laid out using several temperature dependent minerals likely to deposit when at equilibrium with the surroundings. For example, secondary quartz in a typical high-temperature field starts to deposit at a minimum temperature of 180°C and exists at a maximum of over 300°C.

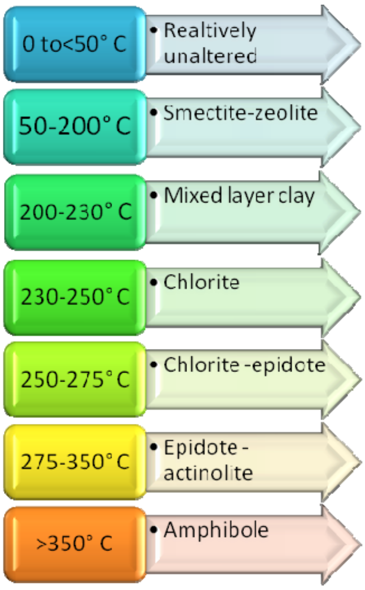

FIGURE 10: Alteration zones in high temperature geothermal fields of Iceland (modified from Franzson, 2010)

5.1 Primary rock minerals in HE-27

Basaltic rocks are the dominant rock types in Iceland and such is the case in well HE-27 at Skardsmýrarfjall. Thus, the primary minerals that form basaltic rocks, as indicated in Figure 11, are olivine, plagioclase, pyroxene and opaque minerals such as magnetite. These minerals are, however, not stable in geothermal environments and alter to new minerals which are likely to be stable in the geothermal environment.

The order of their alteration depends on the Bowens reaction series so that the first mineral to be formed is also the first to be altered. Volcanic glass, which is a constituent of the basaltic rocks, cannot be classified as a mineral but is relevant in this discussion as its replacement products are quite relevant as hydrothermal minerals. The primary mineralogy of the basaltic host rocks in the volcanic zones of Iceland is relatively homogeneous (Árnason et al., 2000).

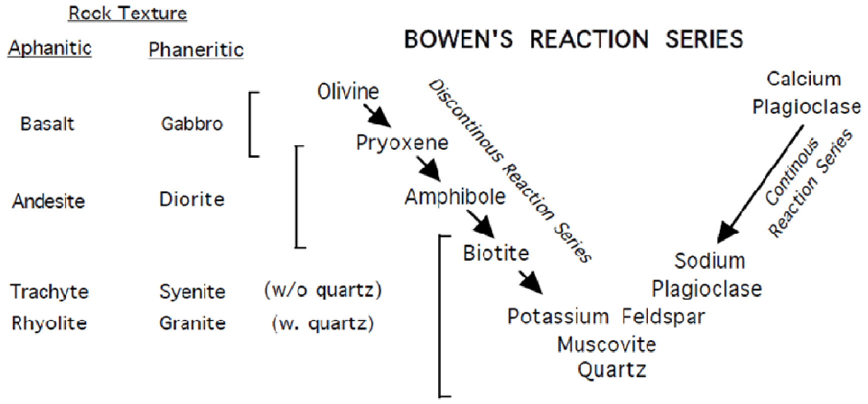

FIGURE 11: Bowen´s reaction series (Thomas, 2010)

Glass: Glass (amorphous quenched liquid) is the first primary constituent to be altered to clays and is replaced, mainly by calcite and zeolites. Altered glass is abundant in the tuff and breccia formations and ranges from fresh to slightly altered in the upper pillow lavas of HE-27. The glass appears dark with a vitreous lustre exhibiting good conchoidal fracturing, although some samples exhibit brownish oxidation. In petrographic analysis, it is transparent when fresh and brownish when altered. Palagonitisation is evident in the upper samples and forms sequences with fine-grained smectite clay layers.

Olivine: In well HE-27 olivine is evident in the pillow lavas as a green crystal with a glassy lustre. In thin section, it has very poor cleavage or a lack of it, very clear fracturing (often with alteration to a reddish coloured mineral called iddingsite along the fracture), and fair to high birefringence makes it distinguishable, especially from the pyroxenes. Olivine is present in the intrusions between 725-754 m and between 826-840 m. The crystal shape, generally subhedral, helps one to identify the pre-existence of olivine even when it has been replaced by another mineral.

Plagioclase: This is the most common mineral in well HE-27. Plagioclase phenocrysts have been used as the main distinction between the hyaloclastite formations. It appears almost like quartz except it is milky and has a characteristic brownish alteration. A substantial amount of plagioclase phenocrysts have liquid inclusions. The most common type of plagioclase, identified in thin section based on the extinction angle, is labradorite where it occurs as phenocrysts as well as in the groundmass of many samples. Albite is also evident as an alteration product, especially at deeper levels of the well.

Pyroxene: Pyroxenes are relatively easy to identify in thin section. They are present in the pillow lavas, some breccias, as well as in the olivine tholeiite intrusions. The mineral appears dark and shiny with a metallic lustre. In thin section, pyroxenes are identifiable by their good cleavage in comparison to olivine.

Opaques: Opaque minerals appear as dark minerals which show a high resistance to alteration and are common in fresh intrusive rocks. Magnetite is identifiable in well HE-27 and is found in association with sphene below 700 m. These minerals can easily be confused with pyrite during petrographic analysis and consequently further confirmation using a binocular microscope is required.

5.2 Hydrothermal minerals of well HE-27

Hydrothermal minerals are those that occur as a result of hydrothermal alteration either by deposition or by alteration of the primary minerals. Numerous hydrothermal minerals are encountered in the well although the most common are calcite, pyrite, quartz and clays. The probable cause of the abundance of calcite and pyrite is the H_2S and CO_2 content commonly associated with geothermal systems. The occurrence of alteration minerals with depth in well HE-27 is shown in Figure 12.

Limonite: This hydrated iron oxide is present in the well to a depth of 242 m and re-emerges at approximately 760 m in abundance within the scoria formation found at this depth. It is formed due to the interaction of rocks with cold groundwater.

Siderite is an iron carbonate with curved and striated faces and occurs with limonite except at 760 m where it is nonexistent.

Zeolites: These are microporous crystalline solids with well-defined structures. Generally they contain silicon, aluminium and oxygen in their framework and cations, water and/or other molecules within their pores (Bell, 2001). There are over 40 naturally occurring zeolites which are often classified according to shape into three main categories: fibrous/acicular, tabular/prismatic, and granular (Koestono, 2007). Moreover, these minerals form at temperatures of up to 120° C and typically occur in veins and vesicles although they may also be deposited on the surface of rocks. Table 2 summarises the various zeolites present in well HE-27.

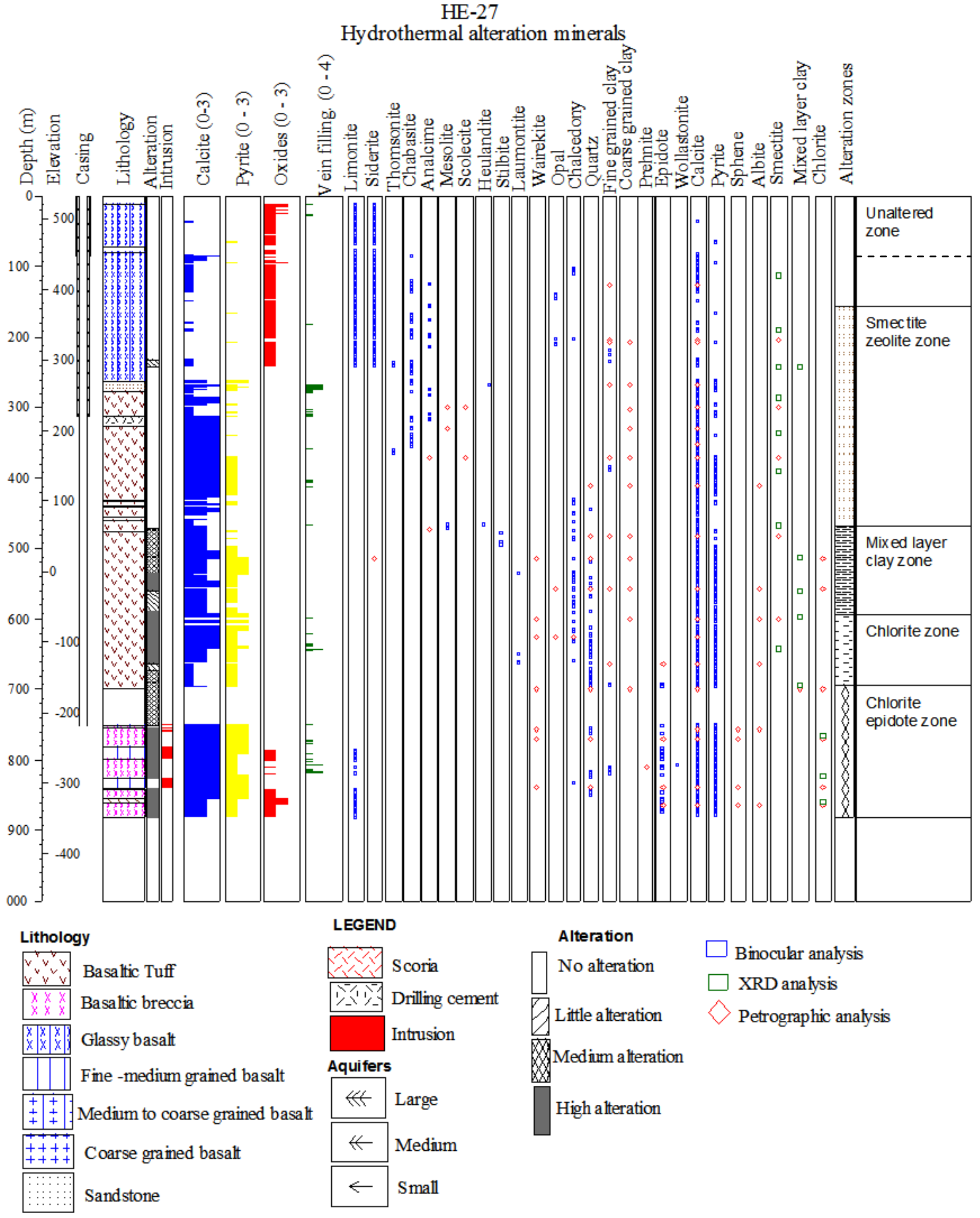

FIGURE 12: Alteration minerals and alteration zones of well HE-27

TABLE 2: Zeolites in well HE-27

Zeolite	Characteristics	Occurrence in HE-27
Thomsonite	Appears colourless, white, to off white and appears as radial aggregates and forms temperatures of approximately 50°C.	200-364 m
Chabasite	It is stable from about 30 to about 80°C, and would fall in the tabular category due to its cubic nature. Although it can be found exhibiting many colours, it is white to clear in this case.	Occurs intermittently between 96 and 354 m
Analcime	Analcime occurs as clear trapezohedron shaped crystals.	First sited at 124 m
Mesolite/scolecite	These fall in the fibrous category of zeolites and the two appear similar. However, mesolite appears to be lighter in colour and would be found higher up in the well as it forms at lower temperatures.	Observed in thin section at 300 m
Heulandite	The mineral is pearly white and exhibits perfect cleavage with wedge shaped edges.	Sited at 466 m
Stilbite	Stilbite in HE-27 exists in a white to off white colour and appears to have a bow tie structure.	Exists at 478-490 m
Laumontite	The laumontite crystals are angular in shape and white in colour and are quite rare. Formation temperature is approximately 150°C.	Present between 648 and 662 m
Wairakite	Wairakite is easily identified in thin section as it appears transparent with very low birefringence and characteristic cross hatched twinning.	Present between 514 and 880 m

Chalcedony/opal: These silica compounds are inferred in thin sections mainly by the characteristic smooth, almost perfectly circular deposition. They appear to be replaced/altered to quartz and from sequences with clays. Chalcedony, however, appears bluish under the binocular microscope, covering the inside of vesicles evenly.

Quartz: Secondary quartz forms at temperatures of approximately 180°C and is identified in cuttings by its typical euhedral shape. In thin section the high refractive index is used to distinguish quartz from the zeolites. Quartz deposition is seen to occur in veins and in vesicles, both evidenced in binocular and petrographic analysis. Evidently quartz appears to be deposited after clay.

Fine-grained clay: These are identified under the petrographic microscope as fine layers mostly composed of smectite and mixed-layer clays.

Coarse-grained clays: These are formed at higher temperatures than the fine-grained clay. The edges of these clays appear fibrous and more defined. In cases where the fine and coarse clays occur together, the latter overlay the former because they are formed later.

Prehnite: Appears as a highly birefringent mineral but is distinguished from epidote by its bow tie structure. The formation temperature of prehnite is nearly 200°C and can be stable up to 250°C. In HE-27, it was first identified using the petrographic microscope at 556 m.
Epidote: The most distinctive feature of epidote is its yellowish/green colour. This mineral was first sited at 652 m where the crystals are not very obvious but at 752 m the prismatic shape becomes evident. In thin section, the mineral has a strong pleochroism of green, yellow and brown.

Wollastonite: Is rare in this well and is identified by its hairy structure in the cutting analysis at 806

m. The temperature of its occurrence is about 270°C.

Calcite: Calcite formation 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). This mineral is abundant and occurs in almost all lithologies. In binocular analysis calcite is identified using dilute hydrochloric acid. Furthermore, calcite crystals have obvious cleavage and can be distinguished from the plagioclase in this way. In thin section calcite occurs as radial, euhedral and platey. The mineral is aggressive as a replacement mineral, occurs in many sequences and as vein and vesicle fillings.

Pyrite: Like calcite it is abundant in HE-27 samples. It occurs as vein fillings and its occurrence can be used to infer permeability in a well.

Sphene: The mineral is identified in a thin section at 756 m by its characteristic brown colour under crossed polars and anhedral crystal shape. Sphene is an alteration product of opaque minerals (Gebrehiwot, 2010).

Albite: In HE-27, albite occurs as an alteration product of plagioclase although it is also perceived to precipitate in some vesicles. The first occurrence of albite is noted in thin section at 482 m where it replaces plagioclase, destroying the characteristic twinning habit.

Clays: These are water-rich phyllosilicates that form by hydrous alteration of primary silicate minerals and require the presence of water in liquid and/or vapour form. The composition, structure and morphology of clay minerals depend on a number of environmental parameters – temperature, fluid composition/amount, pH, etc. Clay crystals are finely crystalline or meta-colloidal and occur in flake-like or dense aggregates of varying types (Ahmed, 2008). In geothermal systems, different types of clays are seen to occur from the surface, where there is slight or no alteration, to the deeper zones where intense alteration is at play.

Three main clay alteration zones have been distinguished in Icelandic geothermal fields. The first zone is mainly characterised by a gradual transformation of trioctahedric Fe/Mg to smectite. An interlayer (second zone) of mixed-layer clay occurs between the first layer and the transformation of Fe-Mg rich saponite to chlorite in zone three (Marosvolgyi, 2009). These three dominant zones fall within a temperature range of 200-240°C. In this case study, clays were analysed using the binocular microscope, petrographic microscope and XRD analysis and the three clay zones listed below were identified.

- *Smectite zone,* as identified using the XRD, lies between 112 and 468 m where the rocks exhibit slight alteration. The smectite group of clays can be classified as swelling clays because they demonstrate high peak values in untreated and glycolated samples but designate lower values when heated. This indicates that the structure collapses with less moisture. The peaks values of the untreated samples show values of 13.05-15.47 Å, but 13.64-17.53 Å for glycolated samples and values of 7.76-10.14 Å for heated samples. Smectite falls in the lower range of the temperature zone at approximately 200°C.
- *Mixed-layer clay* group is clearly identified under the petrographic microscope where it occurs as fine- and coarse-grained clay layers. The clay is strongly pleochroic exhibiting brown orange and green colours. When analysed with the XRD, mixed-layer clays show peaks that are characteristic of both smectite and chlorite. Thus the collapse character of the smectite is evident when heated plus the two peaks of the chlorite at 14 and 7 Å are also present. The presence of these clays in HE-27 occurs between 468 and 594 m.
- *Chlorite* (unstable) appears green and non-pleochroic under plane polarized light and very dark green to brown under crossed polars. The surfaces of the clay also appear very smooth compared to other clays under the petrographic microscope and can be fine- or coarse-grained. In XRD analysis the chlorite has two curves, between 14.57 and 14.75 Å when the sample is

untreated, glycolated and heated. The other curve at 7.2 Å appears when the sample is untreated and glycolated but disappears completely when heated, thus the unstable chlorite connotation. The presence of chlorite indicates temperatures of above 230°C as chlorite has a crystallisation temperature of over 230°C (Franzson, 1998).

5.3 Veins and vesicle fillings in well HE-27

Veins are micro fractures that are filled up either by fluid or by the deposition of secondary minerals. Vesicles, on the other hand, are pore spaces. Despite the difference in structure, both are very important, not only as sources of permeability, but also as ´sample holders'. Hydrothermal alteration minerals, known to be very important geothermometers, deposit within these structures and must be carefully studied, using both binocular and petrographic analysis. Basically, porosity and permeability are two of the primary factors that control the movement and storage of fluids in rocks and lead to the deposition of minerals either in veins or vesicles (Gebrehiwot, 2010).

In well HE-27 vesicles in the upper part of the well are filled with zeolites, calcite, quartz and clay while in the deeper section vesicles contain clays, wairakite, prehnite and epidote. Veins and vesicles are also important in the study especially for the interpretation of mineral deposition sequences in these structures. The veins that could be identified under the binocular microscope were quartz, pyrite and the very dominant calcite veins. Petrographic analysis aids in the determination of additional vein fillings by clays, wairakite and epidote. Moreover, cross-cutting relationships between several veins are also apparent.

5.4 Alteration mineral zones

Progressive alteration of basaltic rocks in high-temperature geothermal areas leads to the formation of mineral alteration zones as expressed by their characteristic minerals. The temperature range in low-temperature geothermal areas is well within the uppermost alteration zone of the high-temperature geothermal areas (Kristmannsdóttir, 1975). As mentioned earlier, the formation of alteration minerals in any environment mainly depends on the compositions of the primary rocks and the nature of the circulating fluids. Given that the primary minerals in Icelandic rocks are homogeneous, and the fluids can be categorised into two homogenous groups, the role of temperature becomes very important. This correspondence of different stable and dominant alteration minerals with different temperatures is used extensively in geothermal exploration and drilling (Árnason et al., 2000).

The study of HE-27 down to 1000 m revealed five zones based on the alteration mineral of high abundance at each particular zone. The boundary between one zone and another was defined by the first appearance of the successive dominant alteration mineral. Reference to Figure 10 is important as each zone can be associated with the formation temperature at depth. Binocular examination, petrographic scrutiny and XRD analysis were used jointly to determine the following:

The unaltered zone (0-156 m): This zone consists mainly of pillow lavas which show some oxidation alteration that is not related to geothermal activity. Limonite and siderite, the main alteration products, are related to cold groundwater circulation zones. XRD analysis shows no smectite presence above 112 m.

Smectite-zeolite zone (96-468 m): There are some uncertainties in the placement of the upper boundary of the smectite-zeolite zone. The first occurrence of smectite by XRD-analysis is at about 112 m but smectite may occur above that depth. A strong smectite signature occurs below 112 m. The first occurrence of zeolite is the presence of chabasite first noticed at 96 m, and this depth is tentatively established as the upper boundary of the smectite-zeolite zone. Amorphous silica is also present in some samples in this interval accompanying limonite and siderite alteration.

Mixed-layer clay zone (468-594 m): In the temperature range from 220 to about 240-250°C, the low-temperature zeolites disappear and smectite is transformed into chlorite in a transition zone, the so-called mixed-layer clay zone, where smectite and chlorite coexist in a mixture (Árnason et al., 2000). The clays are fine and coarse in texture and pleochroic. Hydrothermal minerals present in this zone include some zeolites, wairakite, calcite, quartz and albite.

Chlorite zone (594-694 m): In petrographic analysis the first occurrence of chlorite was identified at 514 m but the XRD analyses indicate that the boundary is lower in the succession. The clay is mainly deposited on the surface of cuttings and within vesicles. In thin section, the coarse chlorite clay is more dominant towards the boundary of the chlorite-epidote zone.

Chlorite-epidote zone (694-1000 m): Epidote is first recognised in the cuttings at 694 m and this marks the upper boundary of this zone. The XRD analyses indicate that unstable chlorite is the dominant chlorite clay present. The co-relationship between lithology and the type of chlorite clay which forms (whether stable or unstable) is still not understood.

5.5 Mineral deposition sequences of HE-27

A mineral deposition sequence/zoning is a powerful tool in the geothermal sciences as it gives a historical account while at the same time (all parameters held constant) it predicts the future of geothermal systems. A sequence in the mineralogical sense is a stepwise or gradual progression of mineral deposition over time. Polymorphs of the same mineral may be deposited or different minerals may succeed each other depending on the prevailing conditions. Many of these minerals are formed either by replacement or deposition, and are temperature dependent (Gebrehiwot, 2010).

Sequences, which typically occur within veins and vesicles, may have clearly defined boundaries and may have cross-cutting relationships. The sequences are such that the earliest mineral to deposit is the one located on the outermost boundary of the vein or vesicle and the one that succeeds it forms later and inward. Moreover, within veins that exhibit cross-cutting relationships, the vein that cuts through another is younger in age. In HE-27, various mineral sequences were identified, clays and calcite showing significant dominance in many of them. Table 3 summarises the sequences as observed using petrographic analysis. The deposition sequence is in chronological order. The earliest to form is marked 1 and so on in Table 3.

The sequences involving clay (Figure 13a) indicate that the system is evolving from a cooler one to a hotter one. This is because fine-grained clays tend to form at lower temperatures while coarse-grained varieties form at higher temperatures because of their temperature dependent structure. Calcite, on the other hand, clearly indicates cooling of the upper part where it occurs after the clay, and cooling in the lower part of the well where it occurs last after wairakite and other high-temperature minerals

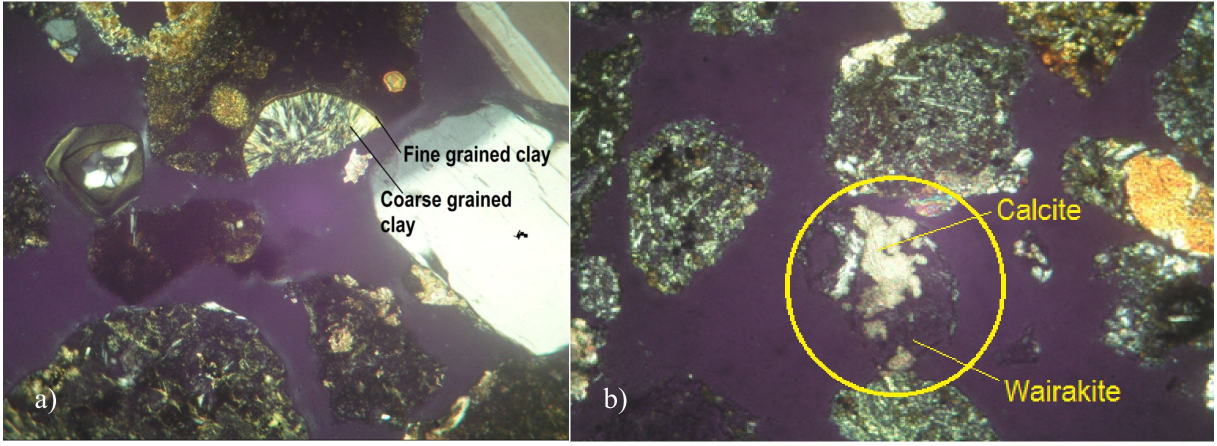

FIGURE 13: a) Fine-grained and coarse-grained clay; b) Wairakite and calcite sequence

(Figure 13b). A study of two wells close to well HE-27, i.e. HE-37 and HE-24, by Gebrehiwot (2010) also indicates that the system is cooling, since calcite comes at a later stage in both wells.

TABLE 3: Alteration sequences in well HE-27

Depth (m)	**Lithology**	**Alteration minerals**											
		fgc	**chal**	**stil**	**scol/mes**	**qtz**	**wai**	**Cgc**	**wo**	**alb**	**preh**	**epid**	**cc**
126	Pillow basalt	1	2										
204	Pillow basalt	1											2
206	Pillow basalt	1	2										
268	Sandstone												
300	Tuff	1			2								3
330	Tuff	2			1								3
	(S2)	1						2					
372	Tuff(S1)				1			2					3
	(S2)	1			2								2
410	Breccia	1						2					3
482	Breccia	1						2					
514	Breccia							1					2
556	Breccia (S1)	1						2					3
	(S2)	1				2							3
600	Breccia							1					2
626	Breccia(S1					2							1
	(S2)							1					2
664	Breccia (S1)									1			2
	(S2)					1					2		
700	Basalt (S1)							1			2		
	(S2)	1				2							
756	Basalt												
770	Breccia					1		2					
838	Basalt (S1)						1	2					
	(S2)					1						2	
864	Basalt (S1)					1						2	
	(S2)									1		2	

fgc = fine grained clay, chal = chalcedony, stil = stilbite, scol/mes = scolesite/mesolite, qtz = quartz, wai = wairakite, cgc =coarse grained clay, wo = wollastonite, alb = albite, preh = prehnite, epid = epidote, act = actinolite, cc = calcite, S 1,2,3 = sequence 1,2,3

5.6 Fluid inclusions

Fluid inclusions are microscopic bubbles of liquid and gas formed during crystallisation or re-crystallisation of minerals. Primary inclusions are formed during mineral crystallisation while secondary inclusions are formed during re-crystallisation. However, both provide vital information with respect to the thermal history of a geothermal system. Fluid inclusion histograms are, thus, plotted with formation temperature curves to establish whether a geothermal system is heating or cooling.

Numerous fluid inclusion studies have been carried out in the Skardsmýrarfjall Mountain exhibiting results that imply heating, cooling and equilibrium within the geothermal system. A study by Gebrehiwot (2010) on 28 primary and secondary inclusions within quartz resulted in two homogenization temperatures, at 230-255°C and 175-185°C. Since the formation temperature was 235°C at the time, the former was associated with a state of equilibrium in the system, whilst the latter indicated heating.

In well HE-27 two depths of varying intervals were sampled for fluid inclusions crystals, these being at 450-500 m and 780-790 m. Unfortunately, only the latter yielded three suitable quartz crystals; hence, homogenization temperature measurements were carried out in a total of 46 secondary fluid inclusions. The homogenization temperature of many of the inclusions is between 240 and 250°C.

The inference is that the quartz crystals re-crystallised within this temperature range, implying that the system attained these temperatures (Figure 14). The formation temperature at 248°C is well within the fluid inclusions homogenization temperature, indicating a state of equilibrium within the area. A comparison of the formation temperature, alteration temperature and fluid inclusion results (Figure 14) confirms cooling due to higher alteration temperatures compared to the formation temperatures. However, the fluid inclusions suggest a state of equilibrium within the system, because many of the inclusions have a homogenization temperature of between 240-250 °C which is within the formation temperature.

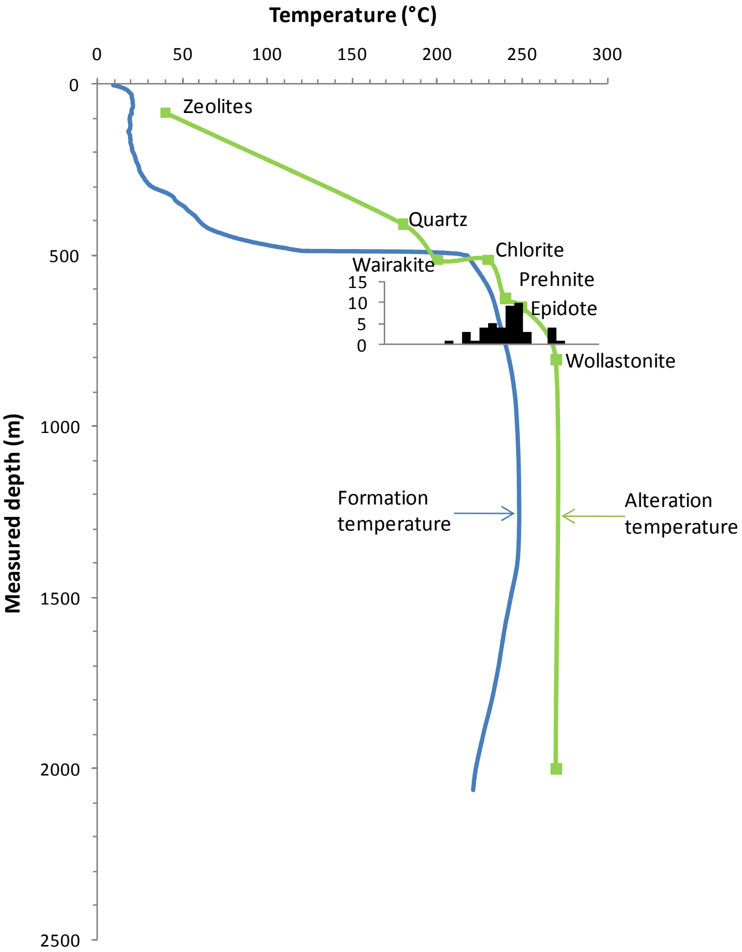

FIGURE 14: Alteration and formation curves of well HE-27; results of the fluid inclusion study are shown in the histogram

5.7 Hydrothermal mineralization well OW-908, Olkaria, Kenya

In an effort to understand better the effects of temperature regarding the formation of hydrothermal minerals, a petrographic analysis of the upper 1000 m of samples from well OW-908 in Olkaria, Kenya, is included in this report. Parameters such as geology, tectonic setting, lithology and geothermal fluids play a significant role in the type of alteration products in addition to temperature. However, Lagat (2004) states that temperature is the most significant factor in hydrothermal alteration because most of the chemical reactions require elevated temperatures and, also, minerals are thermodynamically stable at high temperatures.

Olkaria is a high-temperature geothermal system located within the central sector of the Kenyan Rift Valley and associated with an area of late Quaternary rhyolitic volcanism (e.g. Omenda, 1998). The geothermal system covers approximately 120 km^2 and has been in production since 1981. The area is subdivided into six fields namely Olkaria Central, Northeast, Southwest, Southeast, East and Olkaria Domes where well OW-908 is located. Well OW-908 is a directional production well drilled to a depth of 2988 m and stands at an elevation of 2014 m a.s.l. with a production capacity of 6.5 MWe. The lithology mainly comprises rhyolite, tuff, basalt and trachyte (Figure 15).

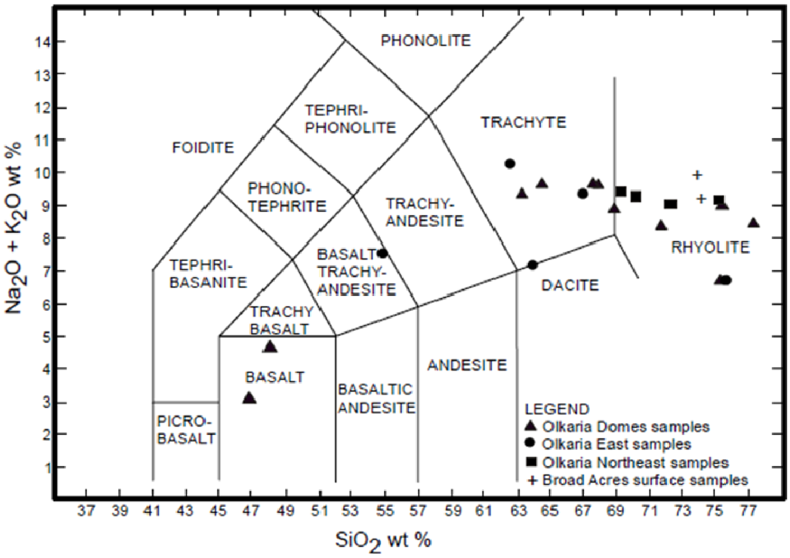

FIGURE 15: Total alkalis silica (TAS) diagram for classification for Olkaria rocks (from Lagat, 2004)

As mentioned earlier, Iceland is located on a mid-oceanic ridge, whereas Olkaria is located within a continental rift zone, both areas being at divergent plate boundaries. However, the geology differs given that lavas erupted from oceanic crust and lavas erupted from continental crust differ in composition.

Magma from oceanic crust is more basic due to the silica, magnesium composition, whereas continental crust magma exhibits some felsic properties due to the silica, aluminium composition. Thus the Icelandic rocks are mostly tholeiitic basalts while those from Olkaria vary from basaltic to rhyolitic rock with an alkaline composition (Figure 15). The difference in location, geological setting, lithology, chemistry of thermal fluids and age of the geothermal systems are significant factors which have to be considered when comparing the alteration minerals of wells HE-27 and OW-908.

The lithology of well OW-908 to 1000 m (Appendix III) constitutes five rock types which are: pyroclastics, rhyolite, trachyte, tuff and basalt. The four alteration zones recognised within Olkaria are zeolite-chlorite zone; illite-chlorite zone; epidote-illite-chlorite zone; and the garnet-biotite-actinolite zone (Lagat, 2004). Petrographic analysis reveals similarities in a number of the alteration minerals when comparing OW-908 and HE-27. The common hydrothermal minerals include various zeolites, calcite, pyrite, albite, chlorite, epidote, prehnite, quartz and sphene. The occurrence of these secondary minerals is largely influenced by the primary rock. Thus, as mentioned earlier, Olkaria and Iceland are located in different kinds of divergent margins which contribute to the difference in geology. Therefore, temperature is the most important factor since it is one of the parameters which influence hydrothermal alteration yet it is a parameter the two different areas have in common.

6. DISCUSSION

The geology of Hellisheidi geothermal system mainly comprises hyaloclastite formations and Mt. Skardsmýrarfjall, which forms a part of the Hellisheidi system, is no exception. The lithology, based on the analysis of well HE-27 samples, consists of formations that can be divided into seven hyaloclastite units, based on their textural differences. The varying amount of plagioclase phenocrysts is the basis of distinction between the various hyaloclastite units. A thin sandstone sediment formation between the first two hyaloclastite units implies a period of reduced volcanic activity. Fine- to medium-grained basaltic intrusions are found below 700 m and they are all tholeiitic basalt in composition.

The geophysical logs are used as a tool to assist in the interpretation of several parameters during and after drilling. The caliper and temperature logs of HE-27 clearly indicate the feed-zones. The resistivity log further indicates high resistivity in the fresh rocks and low resistivity in the alteration zones whilst the neutron log shows higher values where the rocks are saturated with water. Geophysical logs, along with temperature logs, are thus used to determine the position of aquifers in the well. The aquifers in well HE-27 are mainly associated with loss zones because loss zones are associated with permeability, although lithological boundaries and intrusions are also deemed noteworthy particularly with the smaller aquifers. Natural gamma logs show that the rocks of the top 1000 m of well HE-27 are basaltic in composition.

Hydrothermal alteration minerals such as epidote and wollastonite imply that the formation temperature is well over 200°C. Furthermore, the occurrences of fine-grained clays succeeded by coarse-grained clay prove that the system was gradually heated up from lower to higher temperatures. If calcite is ignored as part of the alteration mineral sequence, it would be correct to assume that the system has consistently been heating up with time. However, calcite is associated with cooling if it occurs last in the alteration sequences. On the basis of hydrothermal sequences in HE-27, it can be interpreted that the geothermal system in that area is in the process of cooling.

However, fluid inclusion studies imply that the system is in a state of equilibrium. Results of homogenization temperatures from 46 fluid inclusions give a temperature range of between 240-250°C, namely within the formation temperature of 248°C. Fluid inclusion analysis by Gebrehiwot (2010) in neighbouring wells HE-24 and HE-37 illustrates equilibrium as well as heating, probably due to the opening up of post-glacial fissures.

Comparison of hydrothermal alteration between HE-27 in Iceland and OW-27 in Olkaria emphasises the role of temperature in mineralisation. Despite the differences in rock types, geographic location, fluid chemistry and the age of the geothermal systems, the rocks show similar alteration minerals at almost the same temperatures. This implies that temperature is critical in the hydrothermal alteration process and that alteration minerals can be used to infer this parameter. Furthermore, hydrothermal minerals can designate the evolution of a geothermal system over time.

7. CONCLUSIONS

After a comprehensive study of samples from well HE-27, the following conclusions can be drawn:

- There are six hyaloclastite units in the upper 1000 m of the well primarily comprised of pillow basalts, tuff and breccia which were classified on the basis of texture.
- Basaltic intrusives occur at a depth exceeding 700 m; they are tholeiitic in composition with fine- to medium-grained texture.
- Aquifers in well HE-27 show association with circulation loss zones as well as lithological boundaries in addition to intrusions at depth.
- The upper 1000 m can be classified into five alteration zones based on the hydrothermal alteration mineralogy. The zones include: unaltered zone, smectite/zeolite zone, mixed-layer clay zone, chlorite zone, and chlorite/epidote zone.
- Petrographic analysis of a mineral sequence of low-temperature minerals grading to high-temperature minerals implies a history of heating. Later cooling is revealed by the deposition of calcite on high-temperature minerals.
- Fluid inclusion measurements indicate that the area around well HE-27 is in a state of equilibrium.
- The importance of temperature in geothermal systems is emphasised by the occurrence of similar alteration minerals in different geologic settings, i.e. Iceland and Kenya.

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REFERENCES

Ahmed, M., 2008: Borehole geology and hydrothermal mineralization of well HN-08, Hellisheidi geothermal field, SW Iceland. Report 8 in: *Geothermal Training in Iceland 2007.* UNU–GTP, Iceland 1-29.

Árnason, K., Eysteinsson, H., and Hersir, P., 2010: Joint 1D inversion of TEM and MT data and 3D inversion of MT data in the Hengill area, SW Iceland. *Geothermics, 39-1,* 13-34.

Árnason, K., Karlsdóttir, R., Eysteinsson, H., Flóvenz, Ó.G., and Gudlaugsson, S.Th., 2000: The resistivity structure of high-temperature geothermal systems in Iceland. *Proceedings of the World Geothermal Congress 2000*, *Kyushu-Tohoku, Japan,* 923-928.

Bell, G.R., 2001: *What are zeolites*? British Zeolite Association, website: *www.bza.org.*

Björnsson, A., Hersir, G.P., and Björnsson, G., 1986: The Hengill high-temperature area SW-Iceland: Regional geophysical survey. *Geoth. Res. Council, Transactions, 10,* 205-210.

Danielsen, E.P., 2010: Servicing geothermal wells during completion and follow-up monitoring. *Proceedings of the World Geothermal Congress 2010 Bali, Indonesia*, 6 pp.

Dauteuil, O., Bouffette, J., Tournat, F., Van, Vliet-Lanoë, B., Embry, J.C.,and Quété, Y., 2005: Holocene vertical deformation outside the active rift zone of northern Iceland. *Tectonophysics, 404, 3-4*, 203-216.

Foulger, G.R., and Toomey, D.R., 1989: Structure and evolution of the Hengill-Grensdalur volcanic complex, Iceland: Geology, geophysics and seismic tomography. *J. Geophys. Res., 94-B12*, 17,511-17,522.

Franzson, H., 1998: Reservoir geology of the Nesjavellir high-temperature field in SW-Iceland. *Proceedings of the 19th Annual PNOC-EDC Geothermal Conference, Manila*, 13-20.

Franzson, H., 2010: *Borehole geology.* UNU-GTP, Iceland, unpublished lecture notes.

Franzson, H., Gunnlaugsson, E., Árnason, K., Saemundsson, K., Steingrímsson, B., and Hardarson, B., 2010: The Hengill geothermal system, conceptual model and thermal evolution. *Proceedings of the World Geothermal Congress 2010*, *Bali, Indonesia*, 9 pp.

Gebrehiwot M., K., 2010: *Subsurface geology, hydrothermal alteration and geothermal model of Northern Skardsmýrarfjall, Hellisheidi geothermal field, SW Iceland.* University of Iceland, MSc thesis, UNU-GTP, Iceland, report 5, 65 pp.

Gudmundsson, Á., 1992: Formation and growth of normal faults at the divergent plate boundary in Iceland. *Terra Nova, 4-4,* 464–471.

Gudmundsson, Á., 1998: Magma chambers modelled as cavities explain the formation of rift zone central volcanoes and their eruption and intrusion statistics. *J. Geophys. Res., 103*, B4, 7401-7412.

Hardarson, B.S., Fitton, J.G., and Pringle, M.S., 1997: Rift relocation - a geochemical and geochronological investigation of a paleo-rift in northwest Iceland. *Earth Planet Sci. Lett., 153*, 181-196.

Helgadóttir, H., Snaebjörnsdóttir, S., Níelsson, S., Gunnarsdóttir, S., Matthíasdóttir, T., Hardarson, B., Einarsson, G. and Franzson, H., 2010: Geology and hydrothermal alteration in the reservoir of the Hellisheidi high temperature system, SW-Iceland. *Proceedings of the World Geothermal Congress 2010, Bali, Indonesia,* 10 pp.

Hjartarson, Á., 2009: Central volcanoes as indicators for the spreading rate in Iceland. In: Thórdarson, T., Larsen, G., Rowland,S. K., Self, S., and Höskuldsson, Á. (eds.), *Studies in volcanology: The legacy of George Walker.* Geological Society Publ. House, Bath, UK, 323-330.

Jóhannesson, H. and Saemundsson, K., 1998a: *Geological Map of Iceland. 1:500,000. Bedrock geology* (2nd edition). Icelandic Institute of Natural History, Reykjavík.

Koestono, H., 2007: Borehole geology and hydrothermal alteration of well HE-24, Hellisheidi geothermal field, SW-Iceland. Report 10 in: *Geothermal Training in Iceland, 2007.* UNU-GTP, Iceland 199-224.

Kristmannsdóttir, H., 1975: Hydrothermal alteration of basaltic rocks in Icelandic geothermal areas. *Proceedings of the 2nd U.N. Symposium on the Development and Use of Geothermal Resources, San Francisco,* 441-445.

Lagat, J.K., 2004: *Geology, hydrothermal alteration and fluid inclusion studies of the Olkaria Domes geothermal field, Kenya.* University of Iceland, MSc thesis, UNU-GTP, Iceland, report 2, 71 pp.

Marosvolgyi, K., 2009: *Retrograde alteration of basaltic rocks in the Theistareykir high-temperature geothermal field, North-Iceland.* School for Renewable Energy Science in affiliation with the University of Iceland & the University of Akureyri, MSc thesis, 41 pp.

Mostaghel, B., 1999: Processing and interpretation of geophysical well logs from well KJ-32, Krafla geothermal field, NE-Iceland Report 8 in: *Geothermal Training in Iceland 1999.* UNU-GTP, Iceland 193-220.

Mutonga, M., 2007: The isotopic and chemical characteristics of geothermal fluids in Hengill area, SW-Iceland: Hellisheidi, Hveragerdi and Nesjavellir fields, SW-Iceland. Report 15 in: *Geothermal Training in Iceland 2007*. UNU–GTP, Iceland, 333-370.

Omenda, P.A, 1998: The geology and structural controls of the Olkaria geothermal system, Kenya *Geothermics, 27-1,* 55-74.

Pálmason, G., and Saemundsson, K., 1974: Iceland in relation to the Mid-Atlantic Ridge. *Annual Review Earth Planet. Sci., 2,* 25-50.

Pendon, R. R., 2006: Borehole geology and hydrothermal mineralisation of well HE-22, Ölkelduháls field, Hengill area, SW-Iceland. Report 17 in: *Geothermal Training in Iceland 2006.* UNU-GTP, Iceland, 357-390.

Reyes, A.G., 1990: Petrology of Philippine geothermal systems and the application of alteration mineralogy to their assessment. *J. Volc. Geoth. Res., 43*, 279-309.

RockWare Inc., 2007: *LogPlot program.* Rockware Inc.

Saemundsson, K., 1979: Outline of the geology of Iceland. *Jökull*, *29*, 7–28.

Senior, A.L., Cinotto, J.P., Conger, W.R., Bird, H.P., and Pracht, A.K., 2005: *Interpretation of geophysical logs, aquifer tests, and water levels in wells in and near the North Penn Area 7 Superfund site, Upper Gwynedd Township, Montgomery County, Pennsylvania*. US Geological Survey, scientific investigations report, 2005-5069, 2000-02.

Sigmundsson, F., and Saemundsson, K., 2008: Iceland: A window on North-Atlantic divergent plate tectonics and geologic processes. *Episodes, 31-1*, 92-97.

Simmons, S.F., and Christenson, B.W., 1994: Origin of calcite in a boiling geothermal system*.* *Am. Jour. Sci., 294*, 361-400.

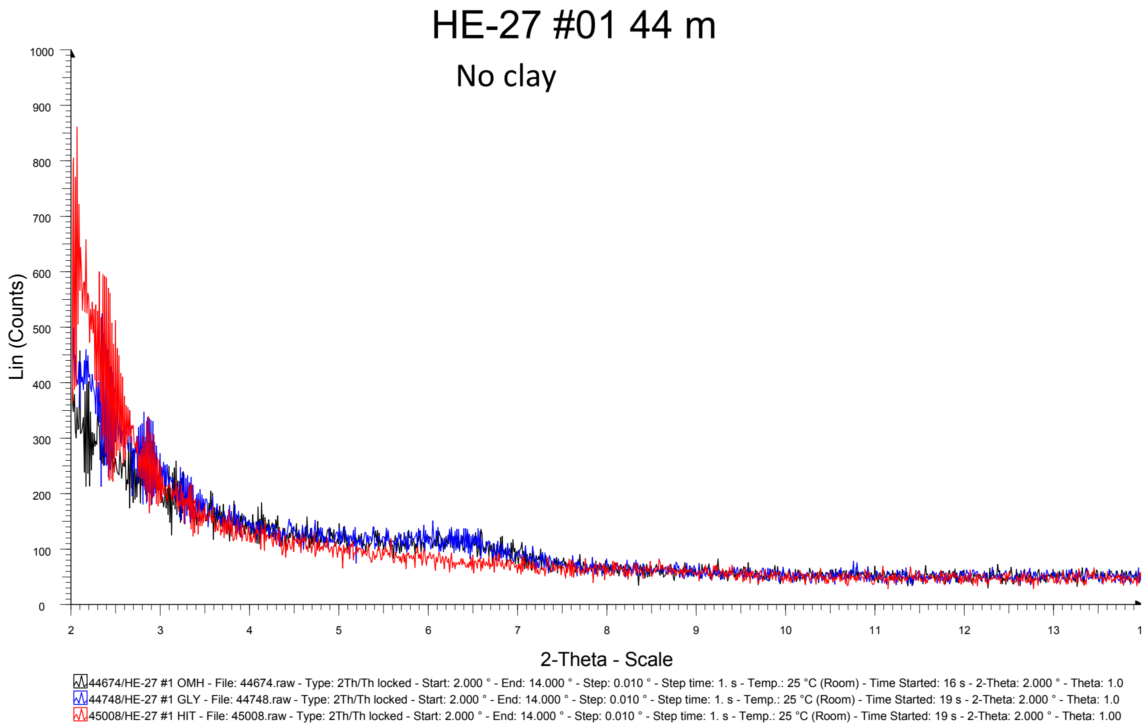
Thomas, J., 2010: *Fairly simple geology exercises for students and their teachers***.** Skidmore College, website, http://www.skidmore.edu.

Thórdarson, T., and Höskuldsson, Á., 2008: Postglacial volcanism in Iceland. *Jökull, 58,* 197-228.

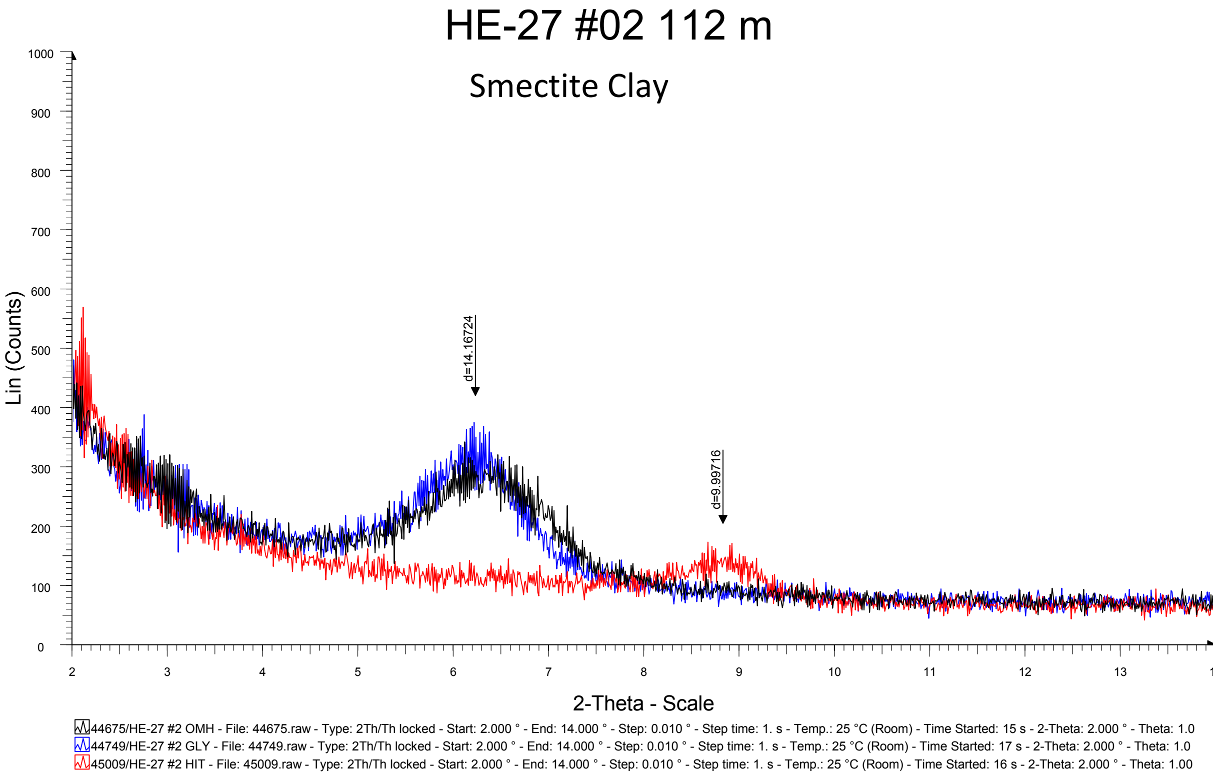
APPENDIX I: XRD clay analysis results

Sample number	Depth (m)	Untreated sample (Å)	Glycolated sample (Å)	Heated sample (Å)	d(002)	Mineral
#01	44	-	-	-	-	-
#02	112	14.17	14.17	9.99		Smectite
#03	190	14.16	14.16	9.99		Smectite
#04	242	15.15	17.5		7.76	Smectite/ mixed-layer clay
#05	286	15.5	17.54			Smectite
#06	336	13.03	13.76	9.9		Smectite
#07	390	13.05	13.97	9.9		Smectite
#08	468	13.64	13.64	10.15		Smectite
#09	512	30.34 / 14.52	30.34 / 14.52	14.52	7.19	Mixed-layer clay
#10	560	15.23 / 13.83	13.83	10.06	7.19	Mixed-layer clay
#11	596	30.91 / 14.55	30.91 / 14.54		7.19	Mixed-layer clay
#12	642	15.21	14.41	9.88		Mixed-layer clay
#13	694	15.52 / 14.04	14.04	10.35	7.27	Mixed-layer clay
#14	764	14.55	14.55	14.55	7.19 HIT=0	Unstable chlorite
#15	822	14.75	14.75	14.75	7.22 HIT = 0	Unstable chlorite
#16	858	14.76	14.76	14.76	7.22	Unstable chlorite
#17	1260	14.57	14.57	14.57	7.19	Unstable chlorite
#18	1402	14.57	14.57	14.57	7.19	Unstable chlorite
#19	1426	14.57	14.57	14.57	7.19	Unstable chlorite
#20	1588	14.57	14.57	14.57	7.19	Unstable chlorite

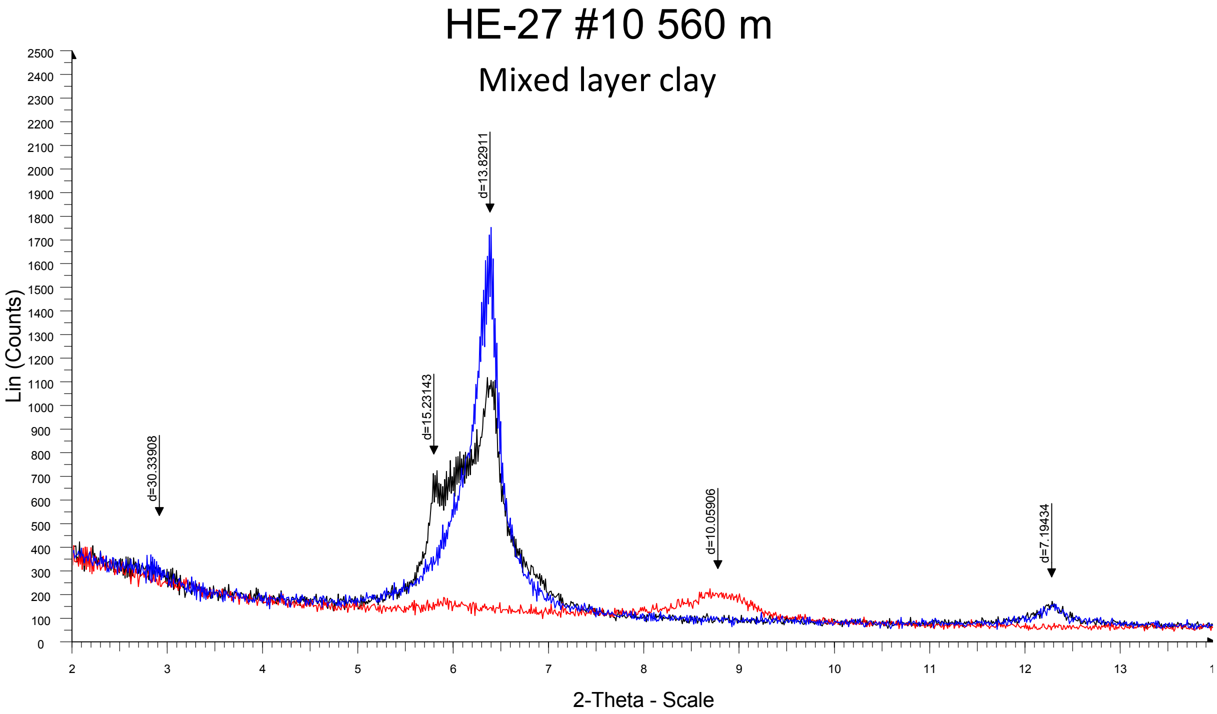
APPENDIX II: Typical XRD patterns for the various clay minerals in well HE-27


HE-27 #01 44 m

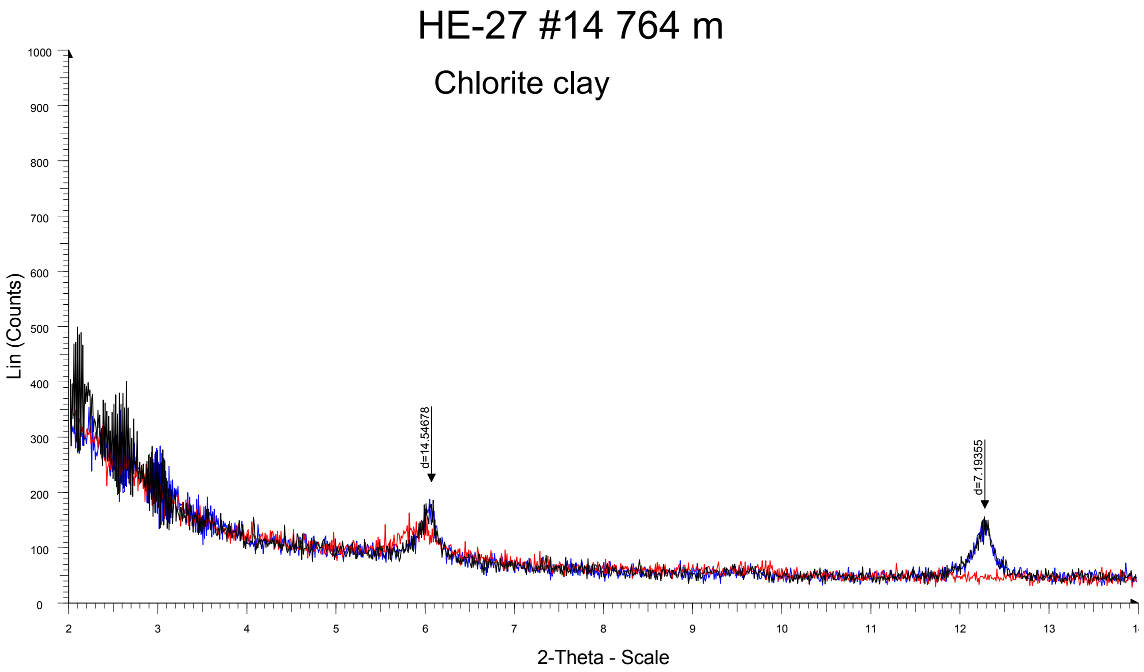
No clay


HE-27 #02 112 m

Smectite Clay



44683/HE-27 #10 OMH - File: 44683.raw - Type: 2Th/Th locked - Start: 2.000 ° - End: 14.000 ° - Step: 0.010 ° - Step time: 1. s - Temp.: 25 °C (Room) - Time Started: 16 s - 2-Theta: 2.000 ° - Theta: 1.
44757/HE-27 #10 GLY - File: 44757.raw - Type: 2Th/Th locked - Start: 2.000 ° - End: 14.000 ° - Step: 0.010 ° - Step time: 1. s - Temp.: 25 °C (Room) - Time Started: 17 s - 2-Theta: 2.000 ° - Theta: 1.
45017/HE-27 #10 HIT - File: 45017.raw - Type: 2Th/Th locked - Start: 2.000 ° - End: 14.000 ° - Step: 0.010 ° - Step time: 1. s - Temp.: 25 °C (Room) - Time Started: 19 s - 2-Theta: 2.000 ° - Theta: 1.0



44687/HE-27 #14 OMH - File: 44687.raw - Type: 2Th/Th locked - Start: 2.000 ° - End: 14.000 ° - Step: 0.010 ° - Step time: 1. s - Temp.: 25 °C (Room) - Time Started: 16 s - 2-Theta: 2.000 ° - Theta: 1.
44761/HE-27 #14 GLY - File: 44761.raw - Type: 2Th/Th locked - Start: 2.000 ° - End: 14.000 ° - Step: 0.010 ° - Step time: 1. s - Temp.: 25 °C (Room) - Time Started: 16 s - 2-Theta: 2.000 ° - Theta: 1.
45021/HE-27 #14 HIT - File: 45021.raw - Type: 2Th/Th locked - Start: 2.000 ° - End: 14.000 ° - Step: 0.010 ° - Step time: 1. s - Temp.: 25 °C (Room) - Time Started: 16 s - 2-Theta: 2.000 ° - Theta: 1.0

APPENDIX III: Olkaria well OW-908 lithology

0-38 Pyroclastics: Pyroclastics consisting mainly of pumice, obsidian and lava (rhyolite) with lithic fragments

38-114 Rhyolite: Light grey quartz and feldspar porphyritic rock. The rock is also porphyritic with amphiboles (arfevedsonite) and flow banding; highly silicic. It's fresh and unaltered.

114-314 No cuttings (total loss)

314-338 Trachyte: Greenish grey to brownish grey fine-grained glassy lava rock. The rock consists of microphenocrysts of euhedral sanidine set in a fine-grained groundmass. The groundmass has tiny feldspar laths exhibiting trachoid flow texture. The rock is altered to light green clays. Rounded chalcedony pellets are deposited in vugs.The iron oxides show very little oxidation.

338-348 No cuttings (total loss)

348-3390 Rhyolite: Light brown highly siliceous spherulitic rock showing local flow banding. It has low phenocryst content and is green to brownish grey intensively altered rock. Abundant clays as a result of alteration are present.

390-404 No cuttings (total loss)

404-572 Rhyolite: Greenish grey to brownish grey fine grained quartz and feldspar porphyritic lava. The greenish rock cuttings are altered into green clays. Reddish-brown minerals result from oxidation of ferromagnesian minerals.

572-576 No cuttings (total loss)

576-618 Rhyolite: Greenish grey to brownish grey fine grained highly siliceous rock. The greenish rock cuttings are altered into green clays. Reddish-brown minerals result from oxidation of ferromagnesian minerals.

618-624 No cuttings (total loss)

624-626 Rhyolite: Brownish grey fine-grained fractured rock. The rock is weak to moderately altered with some cuttings showing flow banding.

626-628 No cuttings (total loss)

628-710 Trachyte: Brownish grey moderately, medium soft, porphyritic, highly altered lava. Large feldspar phenocrysts present. Some cuttings are highly bleached to whitish clays.

710-714 Loss of circulation

714-724 Trachyte: Brownish grey moderately, medium soft, porphyritic, highly altered lava. Large feldspar phenocrysts present. Some cuttings are highly bleached and fractured.

724-730 Loss of circulation

730-738 Trachyte: Extensively bleached, cutting greenish grey, porphyritic lava rock. Primary rock cuttings have large phenocrysts.

738-740 No cuttings (total loss)

740-786 Trachyte: Brownish grey to greenish fine-grained feldspar porphyritic rock. The rock shows a high intensity of alteration with abundant pyrite. The cuttings are extensively bleached.

786-790 No cuttings (total loss)

790-854 Trachyte: Greenish grey mixed cuttings of Trachyte and basaltic. The rock is porphyritic and weakly to highly altered.

854-856 No cuttings (total loss)

856-860 Tuff: Light greenish grey to brownish grey highly altered crystalline rock with some lithic fragments present. Epidote also occurs as hydrothermal alteration mineralogy.

860-864 No cuttings (total loss)

864-868 Tuff: Light grey to whitish tuffaceous rock. The rock is highly bleached into clays. Pyrite is disseminated in the ground mass.

868-874 No cuttings (total loss)

874-87 Tuff: Light grey to whitish tuffaceous rock. The rock is highly bleached into clays. Pyrite is disseminated in the ground mass.

876-884 No cuttings (total loss)

884-886 Tuff: Light grey to whitish tuffaceous rock. The rock is highly bleached into clays. Pyrite is disseminated in the ground mass.

886-902 No cuttings (total loss)

902-904 Tuff: Light greenish grey to brownish grey highly altered crystalline rock with some lithic fragments present.

904-912 No cuttings (total loss)

912-914 Basalt: Light grey fine-grained crystalline lava with whitish tinges. The rock is high in silica and is weakly altered to clays. There are partial returns to total loss returns within the same basaltic zone, indicating a highly permeable zone.

914-922 No cuttings (total loss)

922-924 Basalt: Light grey fine-grained crystalline lava with whitish tinges. The rock is high in silica and is weakly altered to clays. Partial returns to total loss returns within the same basaltic zone indicate a highly permeable zone.

924-932 No cuttings (total loss)

932-934 Basalt: Light grey fine-grained crystalline lava with whitish tinges. The rock is high in silica and is weakly altered to clays. Partial returns to total loss returns within the same basaltic zone indicate a highly permeable zone.

934-940 No cuttings (total loss)

940-980 Basalt: Light grey fine-grained crystalline lava with whitish tinges. The rock is high in silica and is weakly altered to clays. Partial returns to total loss returns within the same basaltic zone indicate a highly permeable zone.

980-1000 Trachyte: Brownish grey porphyritic lava generally showing a low to moderate intensity of alteration. Intermittent fractured zones showing a high intensity of hydrothermal alteration occur.