GEOLOGICAL MAPPING AND BOREHOLE GEOLOGY IN GEOTHERMAL EXPLORATION.

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ABSTRACT

Geothermal exploration is carried out by a team of scientists and engineers of different disciplines such as geology, geochemistry, geophysics, drilling technology, geothermal utilization, etc. This report deals with the contribution of geological mapping and borehole geology to the understanding of the geothermal systems.

The first section deals briefly with a general procedure of geological exploration and the mapping of the TrΦlladyngja area is described as an application of the method. During the field mapping, a special attempt is made to reconstruct the histories of tectonic and volcanic activity and to combine them into a single relative timescale for the area. A correlation is made of the tectonic activity and the geothermal surface manifestations.

The second section of the report describes the tasks of a borehole geologist in geothermal exploration. A study of drill cuttings from the well HS-17 in Grafarvogur, Reykjavik, constitutes a practical example.

In the last part of the report, some conclusions are drawn on geothermal exploration and a brief discussion made on a tentative programme of geological exploration in the home country of the author.

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1 INTRODUCTION

1.1 Purpose and scope of the study

The exploitation of geothermal energy within an area is based on the possibility of extraction of the natural heat of the earth's crust in the area. As for any other resources, the feasibility of the exploitation supposes enough abundance and accessibility to be competitive with other sources of energy. A prospective area will therefore be the one which can deliver a high amount of thermal energy at relatively shallow depth, or in other words, a zone of abnormally high heat flux relatively near the surface of the earth.

This natural heat is produced either by a magma cooling at depth or by decay reactions of radioactive elements contained in the crust such as uranium, thorium and potassium. Its transfer to the surface can be by conduction, convection or diffusion whereas the most effective and fast mode is the convection. For this type of transfer,the heat is conveyed to the surface by circulating groundwater. Therefore, the main objective of geothermal exploration prior to drilling is to locate the areas of relatively high thermal gradient and within these areas the zones of high permeability that can serve to the hot water as paths to the surface or to shallow depths.

Carrying out the difficult tasks of geothermal exploration needs a team of scientists and engineers of different disciplines such as geology, geochemistry, geophysics, drilling technology, etc. The purpose of this report is to stress the contribution of geology, both surface mapping and borehole logging, in geothermal exploration.

Such a wide topic, as well as the training of the author itself, has been chosen because his home country, Burundi, is still at an early stage of geothermal development, and it is thus important that he should be allowed to gather all the information on the methods used to carry out the tasks expected from both exploration and borehole geologists in geothermal exploration.

The training of the author was scheduled as follows: five weeks of introductory lecture course, one week of specialized literature reading, three weeks of cuttings analysis, one week of X-ray diffraction techniques, four weeks of excursions, two weeks of geological mapping and one week of training on the drill site. The rest of the time was used on the preparation of this report. The purpose of the excursions was to have a direct look on geological setting of geothermal fields in Iceland, on the way the exploration has been adapted to different field conditions and on different possibilities of geothermal utilization.

1.2 Geological setting of geothermal fields

Economic aspects of geothermal utilization have led to the definition of geothermal fields as areas of anomalous concentrations of heat in rock and water at depths shallow enough to be economically extractible. As the transfer of heat to the surface is mostly by convection of steam or water, a second condition for good geothermal system will be a permeability high enough to allow the extraction of large quantities of water or steam for a long time.

Experience from geothermal exploration has shown that these two conditions are usually realized in areas of young tectonism and volcanism, primarily along active plate boundaries. Therefore, the best geothermal fields are localized in areas of spreading ridges, subduction zones or intraplate melting anomalies.

Spreading ridges are zones where a new crust is created by intensive igneous intrusion and extrusion, and accordingly, they are favourable locus for discharge of hydrothermal fluids. The probability to find a major hydrothermal convection system at a spreading zone has been found to be a direct function of spreading rate (Lister, 1979). In fact, intracontinental rifts are also the sites of young volcanism and geothermal fields. But their low rate of extension result in a lower probability of finding major geothermal areas than along fast-spreading oceanic ridges. Geothermal areas found in Iceland are an example in association with oceanic spreading ridge while those in Ethiopia, Kenya and Uganda are related to an intra-continental rift, the East African Rift.

Subduction zones are belts along which two tectonic plates move toward each other, resulting in the consumption of the earth's crust by the thrusting of one plate beneath the other. Melting of the down thrust crust produces pods of magma that rise into the upper plate and act as heat sources for overlying hydrothermal convection systems. An example of such fields are those resulted from the subduction of oceanic crust beneath the continental crust along the west coast of South America.

Recent volcanism and associated geothermal fields are also found along intraplate melting anomalies as has been observed in Hawaii and Yellowstone areas in the United States.

The most common geothermal fields in the world are found along orogenic belts or within intracratonic basins where the heat flow is significantly greater than the worldwide normal value of 1.5 hfu (1.5 heat flow units = 1.5 E-6 cal/cm^2 sec). This high regional heat flow may originate from decay reactions of radioactive elements contained in the rocks such as uranium, thorium, potassium. Another possible origin of this heat is the exothermal processes that accompany the consolidation and dehydration(during diagenesis) of rocks with increase of external load or of depth (Shvetsov, 1975) and the heat produced by friction along shear planes of orogenic belts. This heat is transmitted to the uppermost layers by convective hydrothermal cells controlled by tectonic features. An example of geothermal fields associated with orogenic belts are those found in the areas just north and south of the Caucasus Mountains in the USSR, whereas the fields within the Pannonian Basin of Hungary constitute an example of association with intracratonic basins.

At last, low enthalpy fields have been observed in regions of normal heat flow where geothermal gradients are 20 to 40°C/Km. The Paris Basin is an example of this type. Some of those areas of normal heat flow constitute large, porous aquifers that contain water at pressures in excess of hydrostatic. The best known fields of that type are those in the northern part of the Gulf of Mexico Basin.

1.3 Outline of the geology of Iceland

Due to the location of Iceland on the boundary of the American and Eurasian tectonic plates, its geology is dominated by extensional features which result in intensive tectonism and active volcanism. Both central volcano and fissure swarm eruption types are observed. The geological map shows a SW-NE zone of active volcanism flanked successively by Quaternary and Tertiary volcanic formations as the distance increases from the neovolcanic zones (Fig. 1). In its southern part, the neovolcanic zone is shifted in two branches separated by Quaternary formations (Saemundsson, 1979).

The exposed volcanic pile is predominantly built of basalts that form 80-85 % of the total outcrops and acidic including intermediate rocks that constitute about 10 %. The amount of volcanic sediments is in the order of 5-10 % in a typical Tertiary pile but much higher in Quaternary rocks due to subglacial eruptions during the ice ages.

Among the basalts, three main lava types can be distinguished: compound flows of olivine tholeiites, simple flows of olivine poor tholeiites (little or no olivine) and flows of plagioclase and/or pyroxene porphyritic basalts. A gradation in the occurrence of these three types has been observed in the field.

The acidic rocks are mainly lavas and intrusions (60-70 %), but a considerable part (30-40 %) consists of pyroclastic material deposited as agglomerate in vent regions, as ash flow tuff sheets or as airfall tuff beds carried downwind from the source for long distances. Some of the ash flow sheets form marker horizons of considerable help in geological mapping of volcanic piles.

Alkalic rocks are limited to branches of the neovolcanic zones that are termed flank zones and are superimposed on the tholeiite rocks that build up the lava pile. Studies carried out on the dip of the lava pile have shown a general trend towards the central part of Iceland. The dips increase gradually from near zero at the highest exposed levels of the pile to about 5-10 degrees at sea level.

On the basis of climatic evidence from inter-lava sediments or volcanic breccias and from palaeomagnetic reversal patterns, the volcanic pile of Iceland has been divided into four groups which are from the youngest to the oldest:

-Postglacial: last 9000 to 13000 years

-Upper Pleistocene: back to 0.7 million years (m.y.), corresponding to the pre sent normal geomagnetic epoch, Brunhes.

-Plio-Pleistocene: 0.7 - 3.1 m.y., including the Matuyama epoch and the Gauss epoch upwards of the Mammoth event.

-Tertiary: rocks older than 3.1 m.y. and younger than 16 m.y.

The tectonic features are dominated by an intensive fracturing of the axial rift zones into fissure swarms up to 20 Km broad and sometimes over 100 Km long. Their trend is slightly variable, but rather uniform within each branch of the axial rift zones. The eastern branch of the neovolcanic zone south of Iceland has poorly developed extensional features, indicating that crustal spreading along the volcanic axis is there negligible and the extra load of the volcanic edifices is compensated by isostatic subsidence on a more regional scale (Saemundsson, 1979).

The connecting segment between the Reykjanes Peninsula and the Eastern volcanic branch has been referred to as the South-Iceland seismic zone. It has a slightly more east-westerly trend than the branch of the neovolcanic zone that lies along the Reykjanes Peninsula (Saemundsson, 1979).

2 GEOLOGICAL MAPPING

2.1 General procedure

One of the most important steps in the preliminary geothermal study of an area is the geological exploration which has to map any geological and tectonical features that can provide information on the geothermal possibilities of the area. In a new area, where no geological map has yet been made, the first step will be a regional mapping, whereas in well mapped areas one can start directly with detailed mapping of selected prospective fields. Both in regional and prospect investigations, air photographs should be carefully studied and a preliminary map made before direct inspection in the field.

During the regional mapping, the different geological formations are identified, their limits outlined and their relative ages established. For a first approximation of possible aquifers and aquicludes that might be present underground, an attempt is made to estimate the primary permeability of each formation and to deduce secondary permeability from regional structures such as open fault zones, dykes, fractures and fissures.

The estimation of primary permeability is made on the basis of factors that are different in clastic rocks as compared to lava formations. In clastic rocks, the primary permeability results from intergranular porosity inherited from deposition of sediments or eruption of pyroclastics. This permeability decreases with time due to compaction and/or cementation.

In lava formations, the primary permeability originates from cooling cracks and joints, scoriaceous upper part of aa flows and interbedded ash, blocks and lapilli beds. Feeder dykes disrupt rocks and create permeable channels both along the contact zones with the host rocks and along the cooling cracks within the dyke itself. These types of permeability may be relatively short-lived due to mineral deposition by self-sealing and absence of rejuvenation.

The surface manifestations, hot springs or altered areas, are mapped with their measured temperature and samples are taken for chemical analysis and petrographic determination. The geochemical interpretation of the results leads to the estimation of possible underground temperature by the use of geothermometers.

In the laboratory, a petrographic study of thin sections is done to check the field identification of the rocks before the establishment of the final version of the map. The regional geological map obtained can then be used in combination with the geochemical results for the definition of selected small prospects that must be inspected in the next step of detailed investigations.

For detailed mapping of selected prospects, the procedure used is the same as for regional mapping except that the scale is larger. As the secondary permeability constitutes by far the most important path for the upflow of geothermal fluids, a special emphasisis is put on the mapping in detail of every tectonic feature such as faults, dykes, fractures and fissures. Hot springs and areas of surface alteration are exactly marked on the map and their alignments correlated with the tectonical features observed on the surface or inferred from regional and local structures. Indeed, a surface manifestation may be observed fairly far from the upflow zone due to permeability patterns in the uppermost stratigraphic units that result in a lateral flow. Therefore, all the hot springs may not be obviously correlated with observed local tectonic activity.

The results of detailed mapping are confined on a large scale geological map (1/25,000 or 1/20,000) which shows the different geologic formations of the area, tectonic activity in detail, surface manifestations and their measured temperature as well as an estimation of the yield of hot springs.

Combined with the results from geophysical surveys of the prospect area, such a map will be useful for the elaboration of a preliminary geological model that is used for an accurate siting of exploratory wells in the next step of geothermal exploration.

2.2 Mapping of the Trolladyngja area

2.2.1 General geology of the area

The mapped area is located near the middle of the Krisuvik fissure swarm, one of the five fissure swarms that form the active volcanic zone of the Reykjanes Peninsula (Fig. 2). These fissure swarms are the surface expressions of dyke swarms related to central volcanoes that constitute the continuation of the Mid-Atlantic Ridge across Iceland. They form en echelon arrays that may be dextral or sinistral, the controlling factor being the direction of maximum tensional stress which is parallel to the direction of spreading (Jakobsson et al., 1978).

The dominant structures of the fissure swarms are volcanic fissures, open non-eruptive fissures and faults. When traced along their trend, the faults and fissures appear as short sinuous and sometimes dendritic en echelon segments.

The active volcanic zone where plate growth is taking place are the sites of episodic rifting where an interplay has been evidenced between magmatic processes in a central volcano and rifting in the associated fissure swarm (BjΦrnsson et al, 1977). During the interval between two successive events, tensional stress accumulates which is released during the rifting episodes. The central volcanoes appear to play a very active role during such events by allowing the ascent of magma that collects at shallow depth in their roots. From time to time, during the rifting episode, magma is injected from the shallow reservoir into the fissure swarm, the controlling factor being the magma pressure that must reach a certain level to initiate a jerk of rifting. If the ascent of magma into the magma chamber continues beyond the widening capacity of a fissure swarm, an eruption may eventually result (Saemundsson, 1978).

In the Reykjanes Peninsula, eruption fissures appear to dominate the axial area of each swarm whereas the lava shield craters, where present, lie on the periphery of the swarm. Most of the swarms are cut by a shallow graben of 60-80 m of maximum subsidence, with an actually occurring vertical displacement in the order of several mm per year.

Basaltic lava flows of Holocene age (Postglacial) are dominant in the Reykjanes Peninsula (Kjartansson, 1960; Jonsson, 1978) whereas vast basaltic lavas probably from the last interglacial period have been observed on the western part of the peninsula. The axis of the peninsula is morphologically dominated by numerous hyaloclastite ridges that are considered to have formed by subglacial eruptions during the last glaciation.

2.2.2 Field mapping

The aim of this mapping was to get an information on the risk of volcanic activity in the area and to identify the geological and tectonic features that can enhance understanding of the geothermal reservoir and its hydraulic links with the surface manifestations observed in the area. For a good understanding of the heat transfer conditions, an attempt has been made to reconstruct the histories of the volcanic and the tectonic activity and to combine them to establish a single relative timescale for the area.

2.2.2.1 History of the volcanic activity

The mapping of the area led to the identification of seven different lava flows (Fig. 3) distinguished on the basis of petrography, morphology and crater of origin. Several hyaloclastite ridges have been observed with a NE-SW general trend whereas two areas have been found covered by recent alluvium. A brief description of each lava flow is given hereafter in the probable order of extrusion. The order numbers increase from the oldest to the youngest. A brief summary on the formation of hyaloclastites is made at the end of this part.

Lava flow 1 This flow seems to have come from the east of the area and is characterized by a porphyritic texture. The petrographic inspection in thin section revealed it to be a holocrystalline olivine tholeiite within which plagioclase laths form more than 60% of the groundmass that contains in addition olivine, pyroxenes and magnetite.

Lava flow 2 Characterized by a relatively high amount of plagioclase phenocrysts in a groundmass of microlites of plagioclase, pyroxene, magnetite and a few olivine, this lava belongs to the tholeiite type. The craters of origin are on the north-eastern margins of the Fiflavallafjall hyaloclastite hill and form alignments that suggest an eruption of fissure type. The flow is to a large extent covered by the younger lavas.

Lava flows 3a, 3b Two flows 3a and 3b having similar petrographic composition but craters of origin apart have been given a same relative age in the volcanic history of the area. The lava 3a has erupted from the area NE of Trølladyngja and flowed to the north and to the west whereas the lava 3b has erupted through an irregular fissure whose segments are observed at the east of the Fiflavallafjall hyaloclastite hill. The lava 3b has flowed to the north and to the south along the hyaloclastite hill that limits it to the west. As no contact exists between these flows, I have distinguished them apart but concluded that they derive from the same erupting episode on grounds of petrographic similarity. The rock is an aphyric hypocrystalline tholeiite with over 70% of glass. The crystallized minerals are olivine, pyroxene, plagioclase laths and iron ore.

Lava flows 4a, 4b The two flows have been erupted from a group of craters west of the Sog area and seem to be of the same age. Their separation in two flows is based on the occurrence of sparse olivine phenocrysts in 4b that have not been observed in 4a. From observations made further to the south, these lavas have been found to be older than lava flow 7 that flanks 4a flow to the west (S. Einarsson, personal communication). Flow 5 is suggested to be younger than 4, but if there has been any contact, it is covered now by alluvium. And lava flow 4a is younger than 3 as has been deduced from the observation of their contact in the field. The flow 4a is hypocrystalline tholeiite whereas 4b is olivine porphyritic basalt.

Lava flow 5 Erupted from Eldborg crater, it has flowed to the north covering the lava flows 2, 3 and 4a that are thus younger. Field observations revealed it older than lava flow 7. The petrographic study in thin section showed it to be hypocrystalline tholeiite with a few phenocrysts of plagioclase and olivine.

Lava flow 6 Its crater is located just west of the Mavahlidar hyaloclastite hill and it has flowed to the west and to the north. The only information provided by the field mapping is that the lava is younger than 3 but nothing can show its exact age position between flow 3 eruption and the present day time. The petrographic inspection revealed it to be olivine-plagioclase porphyritic basalt with a hypocrystalline groundmass of glass, plagioclase laths, pyroxene, olivine and iron ore that form above 80% of the lava.

Lava flow 7 The crater of this flow is located to the south of the mapped area and the lava has been flowing to the north. Its age is younger than 5 but any age relation can be established with lava flow 6 as they do not have any contact. Absolute age determination has revealed it to be younger than 900 years (Jonsson, 1983). A tholeiite composition was obtained from the petrographic study with very few plagioclase phenocrysts in a groundmass of glass, olivine, pyroxene, plagioclase and iron ore that form more than 90% of the lava.

Formation of hyaloclastites The hyaloclastites observed are believed to have formed during subglacial eruptions. Three main types of hyaloclastites may result from such an eruption: pillow lava, pillow breccia or well bedded fine grained tuffs.

The pillow lava represents the effusive initial phase of a subglacial (or subaqueous) eruption that tends to form a pile around the orifice. The pillow lava is enveloped in a layer of irregularly stratified glassy debris formed together with it due to direct contact of water from ice melting and moving lava. This debris is transported by currents and deposited between the ice walls and the pillow lava itself. This type of formation is usually found on the slopes of the hyaloclastite hills.

The pillow breccias are also found on the slopes and are accumulated by gravitational slumping on the slopes of the pillow lava piles. The high content of glass in pillow breccias seems to claim for either the participation of molten material or the formerly described irregularly stratified debris in the slumping procedure.

As the eruption lasts, the ice cap keeps on melting. When the subglacial (subaquous) eruption has built up a mountain nearly up to the water level (or the water level has been lowered by an escape of glacial melt water), the eruptive mechanism evidently changes over to explosive phreatic activity producing mainly well bedded fine grained hyaloclastites commonly termed tuff (Saemundsson, 1967).

From this brief summary on hyaloclastite formation, it appears that a hyaloclastite mountain can be looked at as a closed box that may contain hyaloclastites from different erupting episodes that took place consecutively during a glaciation.

All the lava flows described above are of Postglacial age and therefore younger than the hyaloclastites. In addition, several eruptive fissures have been mapped over hyaloclastites, showing a recent volcanic activity that may be contemporaneous with some of the lava flows mapped. An accurate relative dating of these eruptions is impossible due to the usually local extension of the erupted material.

2.2.2.2 History of tectonic activity

The tectonic features of the area are characterized by the occurrence of faults, fractures and fissures that have a general NE-SW trend. The northwestern corner of the map seems to be outside the fissure swarm as no tectonic activity has been observed there.

The oldest lavas (1 and 2) are more affected by faults and fractures than the younger ones. The lava flow 2 is dissected by faults with up to 5-10 m of vertical displacement. Lavas 3a and 3b are not faulted but open fissures of up to 0.2 m width have been observed. No tectonic activity has been found in the younger flows indicating that the area has suffered no or little tectonic activity during the last eruption phases. Another possible explanation is that the younger lavas are of similar age and the volcanic activity has just followed the tectonic activity as the last phase of a rifting episode.

The estimation of the relative age of faults and fractures affecting hyaloclastites is more difficult. In fact, the hyaloclastites should be more severely faulted than the lavas since they have suffered rifting episodes for much longer time than the oldest lavas. But this emphasis of tectonic activity is not observed in the field due more likely to the fact that faults and fractures will in a relatively short time be covered by talus and other sediments within the hyaloclastites. And it is also likely that the retreating ice cap immediately wiped out all such features in the hyaloclastites.

2.2.3 Geothermal activity and hydrothermal alteration

As described above, the fissure swarms observed in the neovolcanic zones are the surface expression of shallow intrusions and dyke swarms associated with central volcanoes. Three main factors control the degree of alteration: the temperature, the rock type and the water circulation. The rock types have varying susceptibility to alteration depending on the degree of crystallisation and mineralogical composition. Glassy breccias and tuffs are the most easily altered rocks, whereas holocrystalline dense basalts are much more resistent. A gradation in the alteration state is observed in the field. The glass becomes first hydrated and oxidized and the colour of the rock turns to brownish to yellow or orange, leading to a typical alteration termed palagonitization in specialized literature. Where the rock is heavily altered, the glass is completely replaced by clay minerals. Olivine is the first mineral to alter, first to iddingsitic aggregates and later it is completely replaced by clay minerals. Pyroxene is also altered to clay minerals, but this alteration is rarely completed.The plagioclase is nearly unaffected by alteration (Kristmannsdottir et al., 1974). Such surface alteration was observed around the Eldborg crater and the Sog craters zone.

Hot springs were also found associated with the propylitized areas with a maximum outlet temperature of 83°C. The highest temperature measured within an altered zone was 97°C on the Lambafell hyaloclastite.

The geothermal activity and associated hydrothermal alteration seem to be controlled by a major fault that runs through the Sog crater zone, the Eldborg crater and the Lambafell hyaloclastite after which a lateral shift of the fault to the east is observed but still keeping the same general trend as to the south.

The X-Ray Diffraction analysis of two samples of clays from the propylitized surface on TrΦlladyngja and Lambafell hyaloclastite mountains respectively showed that TrΦlladyngja clays were of vermiculite-illite mixed layer group whereas Lambafell clays were of serpentine-kaolin group. But for both samples, no peak was observed after heating up to 550-600°C, which seems exceptional for those types of clays.

3 BOREHOLE GEOLOGY

3.1 Contribution to geothermal system modelling

After the geothermal reservoir has been broadly outlined by geological and geophysical exploration on the surface, the next step is the study of the subsurface structure especially as it affects the permeability. For this purpose, exploratory wells will be drilled. The decision where to drill must be based on locating adequate secondary permeability preferably as close as possible to upflow zones. And even where the primary permeability is present, the best places to drill will be where secondary permeability may also be inferred.

The main information expected from a well are the location and yield of aquifers as well as their physical parameters such as temperature and pressure that control their dynamics. Location of aquifers and evaluation of their yield and geological affinities are tasks of a borehole geologist whereas a borehole geophysicist will be responsible for physical properties measurements and interpretation. This part of the report deals with the tasks of a borehole geologist as a contribution to the understanding of a geothermal system. Three sections are developed: drill site duties, laboratory work and geological modelling.

3.1.1 Drill site duty

Simultaneously with drilling, an analysis of the cuttings is done and a preliminary stratigraphic section constructed with beside it a graph of mean penetration rate against depth. All information on the proceeding of the drilling are recorded. These include the penetration rate, the load pressure of the drillstring, the dimensions of the hole, drillpipes and casings, the quantity of circulation water, the location of aquifers and probably the temperature of the water in the aquifers, diameter and inclination measurements.

From cuttings analysis, the exact location of aquifers is done on evidences such as the petrographic type, the degree of alteration and/or the density of fracturing of the rock. A gain or loss of circulation recorded during drilling may be in that way exactly correlated with the stratigraphic section. This data may be very helpful as to determine the optimal depth of the end of a casing that must be in hard solid rock and to cement all unwanted aquifers (e.g.cold aquifers). Therefore, the drilling crew expects advice from the geologist in matters depending on geology and stratigraphy.

The identification of hydrothermal mineral assemblages in cuttings or cores will provide information on the range of temperature and permeability that can be expected downhole.

After drilling, the drill site geologist in some cases performs with the project hydrogeologist the completion tests of the wells consisting of a permeability assesment by pumping with compressed air, stimulation with injection packer, multiple step injection tests, etc. The purpose of these tests is to evaluate the maximum flow rate of the well, to estimate the drawdown during production and the bulk permeability of the field. The effects of the tests on other wells in the surroundings are recorded contemporaneously.

3.1.2 Laboratory work

As the sample boxes are marked by the drillers with the time of collection and the corresponding depth reached, the first task of the drill site geologist is to correct the sample depth using the penetration rate, the mean settling velocity of the cuttings, the depth reached and the time of collection. A petrographic study of thin sections prepared from cuttings is carried out and the results used to make the final version of the stratigraphic section of the well. The dimensions of the hole and casings are marked on the same drawing as well as the load pressure of the drillstring, location and yield of the aquifers, penetration rate and the geophysical loggings. The identification of alteration minerals is also completed by the use of both thin sections inspection and X-Ray diffraction techniques.

Simultaneously with this work, a geochemical study may be carried out of fluid and host rock compositions, their interactions and the stability ranges of the alteration minerals observed during the petrographic study.

3.1.3 Geological modelling

The combination of all the information provided by the stratigraphic sections of all the exploratory wells and the different parameters recorded beside them with the results of geochemical and geophysical interpretation are used to refine the working geological model derived from the surface regional and prospect surveys. A new geological model is constructed taking into account subsurface conditions of temperature and permeability. This model can then be used in the construction of a general model of the geothermal area.

3.2 Well HS-17 in Grafarvogur

3.2.1 Location, objective and available data

The well HS-17 is a shallow (282 m) thermal gradient well drilled in the Reykjavik low temperature area. As the objective of a gradient well is the study of the conductive heat flow, the siting of such a well is done in such a way that no aquifer is encountered. Therefore, the only available data from the well are information on the penetration rate, differential measurements of temperature and drill cuttings.

3.2.2 Stratigraphy

The samples of cuttings are collected in 100 ml boxes at every second meter of penetration. The sample boxes are marked with the time of collection and the depth reached.

This dept is directly read on the geolograph when the latter is available (big drillrig in Iceland), or calculated from the length of the drillstring already sunk into the hole.

At 100 m interval, bigger samples are taken in 1-3 l boxes for chemical analysis and mineral separation, as the normal size samples are too small for that purpose. When drilling in new areas, the larger samples are taken at randomly chosen depths whereas in well-known areas, the sample depths are chosen from the stratigraphic sections of previous drillholes.

The stratigraphic section constructed from the binocular inspection of drillcuttings and thin sections study of selected samples is shown in Fig. 4 with beside it a drawing of the penetration rate. The latter does not show any evident correlation with the stratigraphy due most likely to the fact that the penetration rate is not accurate enough as it is the calculated mean for drilling a depth equal to the length of one drillpipe. Seven rock formations have been distinguished on the basis of petrography and alteration. Each formation will be briefly described below.

Clastic sediments (0-12 m): This layer isformed by badly sorted detrital sediments where the coarse fragments are finely to intermediately crystalline basalts and the finer grained matrix is of tuffaceous composition. The proportion of the tuff seems to increase with depth and is at a maximum between 8 and 10 m where the tuff constitutes more than 90% of the cuttings. The lower part of the layer is slightly oxidized. The sediments are fresh and alteration minerals rare.

Tholeiite formation I (12-108 m):This formation is characterized by fresh finely crystalline tholeiite lavas. Thirteen different lava flows have been identified by using the stratification usually observed within a single lava flow, i.e. the upper part is often glassy to very fine grained, highly vesicular and sometimes red due to oxidation, whereas the lower part is coarse, darker and more dense. Within each lava flow, numerous almost completely filled vesicles were observed in the upper part whereas a few, usually empty vesicles characterize the lower part. The degree of fracturing is very low as deduced from the amount of fragments consisting of single alteration minerals. The amygdale minerals observed were chabazite, scolecite and clays.

Basalt-rich breccia (108-114 m): This layer is composed of almost equal proportions of glass and finely crystalline tholeiite. The porosity is characterized by numerous vesicles, mostly empty, but the degree of fracturing is very low. The rock is relatively fresh.

Detrital tuff (114-118 m): The petrographic inspection of cuttings shows a few finely to intermediately crystallized basaltic fragments embedded in a groundmass of tuff which forms over 90% of the whole rock. The tuff is slightly oxidized but no alteration minerals have been obseved. This layer is indicative of an erosional episode during an interglacial period.

Tholeiite formation II (118-158 m): This series is formed by six lava flows of glassy to crystalline tholeiite. Vesicles are few and half-filled by secondary minerals such as chabazite, scolecite/mesolite, iron ores and clays. The degree of fracturing is relatively low and the primary constituents of the rock are fresh to weakly altered.

Altered olivine tholeiite series (158-264 m): The upper boundary of this series is found at 158 m whereas the lower boundary is arbitrarily put at 264 m where the alteration state of the primary constituents of the rock decreases considerably. Eleven lava flows have been distinguished formed by finely crystalline olivine tholeiite. The petrographic texture seems not to be homogeneous as some fragments in the cuttings from 188 m have been identified as olivine-plagioclase porphyritic. The flows display numerous vesicles half-filled by secondary minerals. The degree of fracturing is high as well as the alteration of the rock as deduced from the alteration of olivine and plagioclase. Chabazite, mesolite/scolecite, stilbite/-

heulandite and clays have been observed in cuttings and thin sections. Three thin sediment layers intersect this basaltic profile.

Slightly altered olivine tholeiite series (264-282 m): The upper boundary of this layer has been arbitrarily fixed at 264 m on the basis of the alteration state of the rock whereas the lower boundary is tentatively put at the bottom of the well. Three lava flows were identified. The cuttings are composed of finely to intermediately crystalline olivine tholeiite fragments which show numerous vesicles in the upper part of the flows. These vesicles are half-filled by zeolites and clay minerals that have been identified by the study of thin sections as chabazite, mesolite, stilbite/heulandite and clays. The degree of fracturing of the rock is low and the alteration state of the primary constituents relatively weak.

3.2.3 Alteration

In order to confirm and complete the identification of secondary minerals in the well, several XRD analysis were carried out. For that purpose, amygdale minerals were handpicked from cuttings and clay samples prepared from selected depths. The amygdale samples were taken at depths where a particularly high amount of zeolites was observed (166 m, 180 m, 214 m, 236 m), whereas the clay samples were prepared from cuttings from basalt-rich breccia, detrital tuff and olivine tholeiite layers where an alteration leading to clay formation could be expected from the relatively higher permeability. A brief description of the preparation of a clay sample is given in Appendix I.

As shown in Table 1,the results of XRD analysis confirm and sometimes complete the petrographic study showing the following secondary minerals: chabazite, mesolite/scolecite, calcite and clays. The 236 m amygdale sample showed some peaks that have not been well identified and belong probably to the zeolites epistilbite and wairakite/analcite. The XRD identification of clay samples showed that all of them were smectites.

In the alteration mineral assemblage identified, laumontite is lacking within the zeolites and only smectite clays are present, showing that the well intersects the uppermost alteration zone of a low temperature area as defined by Kristmansdottir and Tomasson (1976).

TABLE 1. Results of XRD analysis

Depth (m)	Rock type	Sample type	Alteration minerals
112	basaltic breccia	clays	smectites
116	tuff	clays	smectites
166	alt.ol.thol.	amygdale	chabasite, scolecite/ mesolite, clays.
170	alt.ol.thol.	clays	smectites
180	alt.ol.thol.	amygdale	chabasite, scolecite, clays.
214	alt.ol.thol.	amygdale	calcite, mesolite, wairakite/analcite, clays and ?
236	alt.ol.thol.	amygdale	mesolite, analcite/ wairakite?, epistilbite ? and clays.
240	alt.ol.thol.	clays	smectites
272	weak.alt.ol.thol.	clays	smectites
278	weak.alt.ol.thol.	clays	smectites

4 CONCLUSIONS AND DISCUSSION

Geothermal fields are found in areas of young tectonism and active volcanism where the heat source formed by cooling intrusions is localized at shallow crustal depths and the secondary permeability is high enough to allow the convection of hydrothermal fluids. Hence, when prospecting for geothermal resources, a special attention must be put on spreading oceanic or intracontinental ridges, subduction zones and intraplate melting anomalies, orogenic belts and intracratonic basins. A geological mapping of every tectonic feature, surface manifestations and volcanic activity will be very useful in the understanding of the geothermal system. The subsurface geology obtained from exploratory wells will complete the information needed for the construction of a geological model of the geothermal system.

The present study has been conducted in thermal areas in Reykjavik and the Reykjanes Peninsula where a thermal gradient above the average has already been demonstrated and surface thermal manifestations recorded by previous research. The main target of the exploration in such cases is to identify the factors that control the permeability in the area. Two types of permeability were found in both the Grafarvogur and the TrΦlladyngja thermal areas, namely horizontal permeability and vertical permeability.

The horizontal permeability is mainly a primary one resulting from intergranular porosity in sediments and pyroclastics, cooling cracks and joints in lavas, scoriaceous upper part of aa flows and interbedded ash, blocks and lapilli beds. This type of permeability decreases with time due to factors such as compaction and cementation in sediments and pyroclastics and mineral deposition in vesicles of lava flows. The determination of the age of the different formations mapped and the estimation in drill cuttings of the amount of amygdales filling the pores are very important in assessments of the nature of the permeability.

The vertical permeability which is by far the most important path for the upflow of geothermal fluids is related to tectonic features such as faults, contact zones between dykes and host rocks, fractures and fissures. When mapping for geothermal exploration, these features will require a special attention of the geologist and an attempt will be made to correlate them with the surface manifestations observed in the area in order to understand the upflow of hydrothermal fluids. During the binocular inspection and the petrographic study of rock cuttings, the vertical permeability is estimated on the evidence of degree of fracturing as deduced from the amount of fragments consisting of single alteration minerals.

At last, a correlation on cross-sections of several stratigraphic sections of exploratory wells with data from surface mapping may give complementary information on permeability due to buried faults, dykes and fractures.

As the home country of the author intends to start a geothermal exploration in the foreseeable future, a brief discussion is made below on a tentative programme of geological exploration of the geothermal resources of Burundi.

The geothermal manifestations in Burundi are hot springs or warm water pools where the highest temperature recorded is 68°C at Ruhwa spring (Mc Nitt 1969, Deelstra et al. 1972, Edeline et al. 1981, Armannsson et al. 1983). No record exists of surface steam or mud pools. These hot springs are associated with the western branch of the East African Rift that runs from Turkey to Mozambique and includes the Dead Sea, the Red Sea, the Gulf of Aden and crosses East Africa from Ethiopia to Mozambique. The Burundian coast of Lake Tanganyika and the Rusizi Valley belong to that rift.

The formation of the rift is of Tertiary age whereas the geological formations crossed by the rift are Proterozoic. The rift is still active as numerous active volcanoes have been recorded further to the north, e.g. in Zaire and Rwanda (Fairhead et al. 1982, Williams 1982, Barberi et al. 1982). The graben of the western Rift Valley is filled by young sediments that can form a good geothermal reservoir whereas the Precambrian rocks are considered very poor in this respect.

The most promising gethermal fields of Burundi are expected to be in the northwest on the edge of the Tshibinde volcanic region south of Lake Kivu. The highest temperature recorded in Burundi was found in sediments covering the Rift Valley in that area. Therefore, the geothermal exploration of Burundi s hould be concentrated on this region and should cover the whole Tshibinde volcanic zone that extends to Zaire and Rwanda. An anomalously high geothermal gradient may be expected there as a result of dyke swarms and intrusions associated with central volcanoes recorded further to the north and cooling at relatively shallow depths. Therefore, the exploration of the area should start with a detailed geological mapping of the Tshibinde volcanic zone where geological formations, folds, faults, fractures and fissures would be mapped in detail. An attempt should also be made to establish a relative timescale for tectonic and volcanic activities and to correlate them with the geothermal manifestations. The data would then be confined in a geological model of the geothermal system that shows possible heat sources, reservoir rock, cap rock, possible upflow zones and different aquifers.

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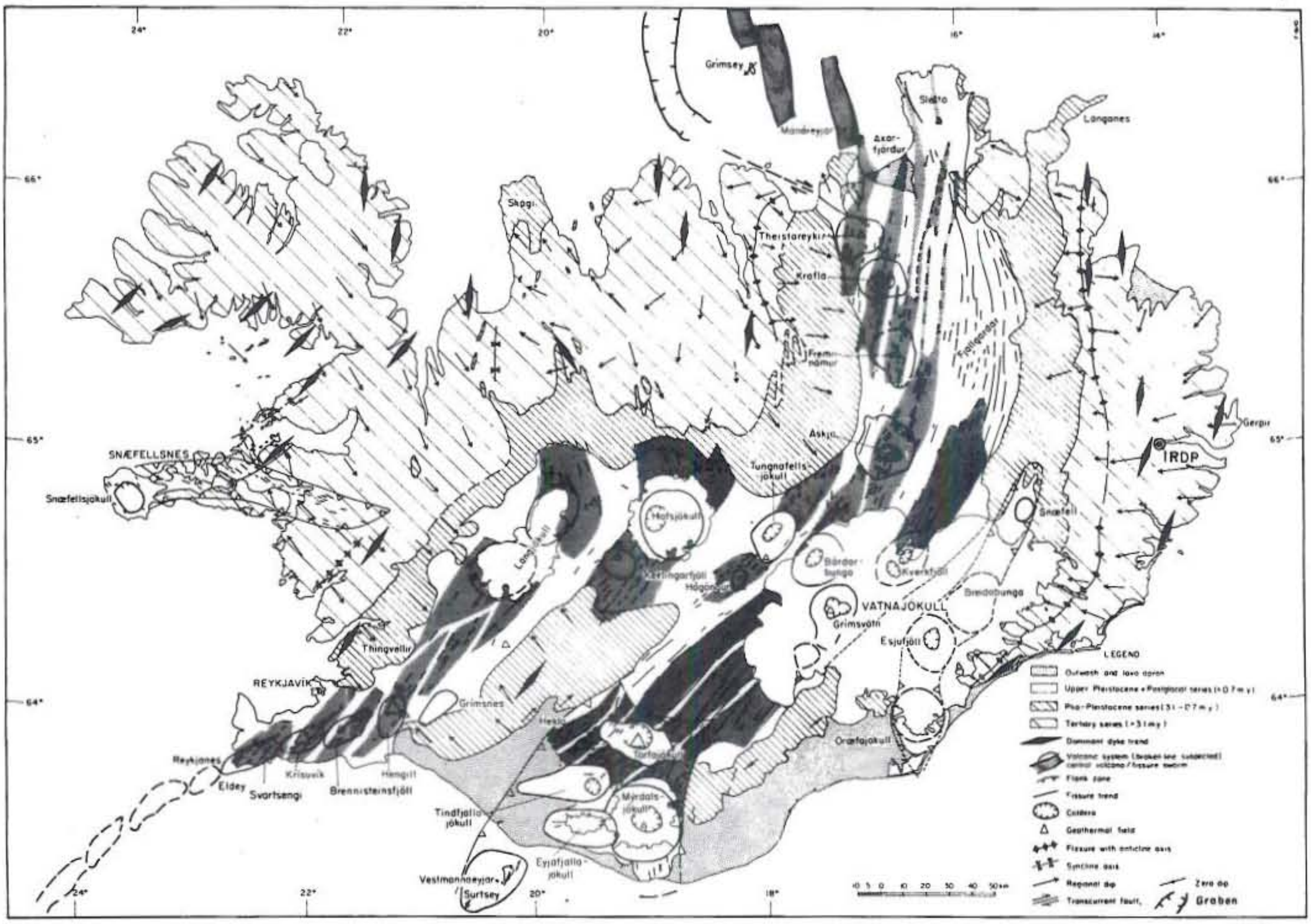

Fig. 1. Geological map of Iceland

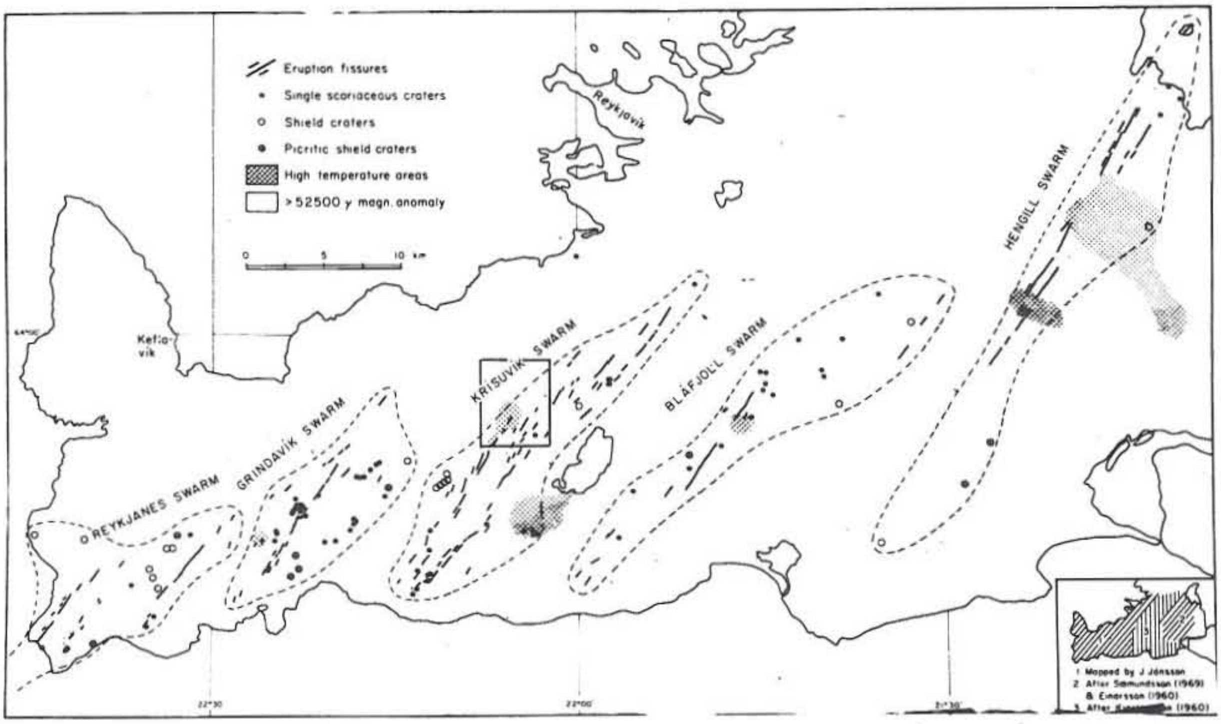

Fig. 2. The active volcanic zone of the Reykjanes Peninsula, SW-Iceland (after Jakobsson et al., 1978).

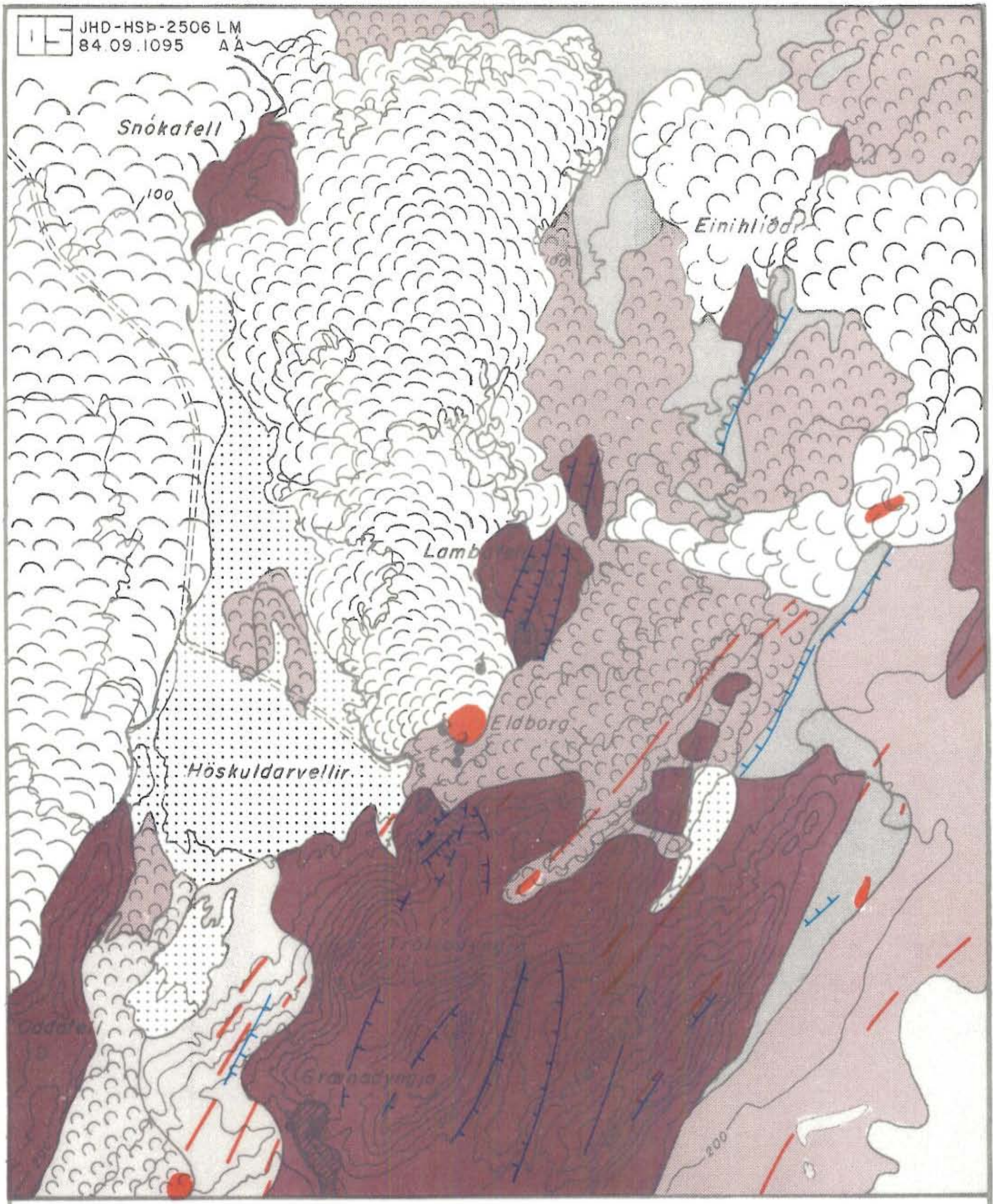

Fig. 3. GEOLOGICAL MAP OF THE TRÖLLADYNGJA AREA, REYKJANES PENINSULA

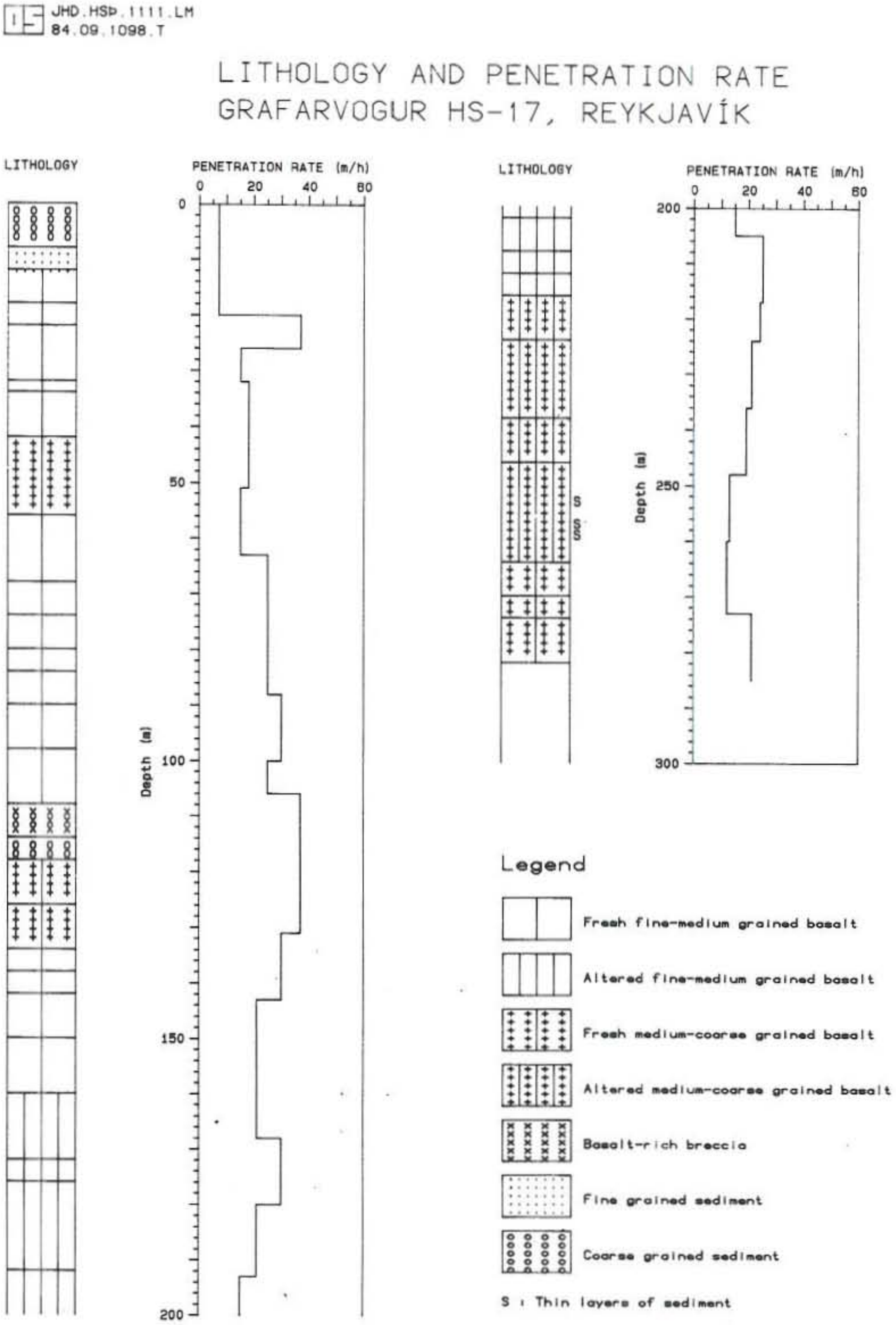

Fig. 4. Stratigraphic section of well HS-17 in Grafarvogur, Reykjavik, Iceland.

APPENDIX I

Preparation of clay samples

Two teaspoons of drillcuttings are placed into a glass tube. All the dust is washed out before the tube is filled up to 3/4 with distilled water and plugged with a rubber stopper. The tube is then put horizontally into a shaker (mixer) and well shocked with sponge. The sample is shaked for approximately 5-6 hours.

The tube is then placed on a table for approximately 2-3 hours to let the larger particles deposite. A few ml are then taken with a pipette from the tube and 3-4 drops placed on a numbered glassplate which is covered entirely with a thin layer of the sample. The sample is afterwards dried on the table before a 24 hours minimum stay in a dessicator containing a solution of $CaCl_2$. The sample is then ready for the first run which is done for values of 2θ comprised between 2 and 17°C.

A duplicate of the sample is placed in a dessicator containing a glycol solution and stored for 48 hours at room temperature before the sample is run for the second time in the same range of 2θ.

After the second run, the sample is placed on an asbestos plate which is put in a preheated oven at 550-600°C and heated for one hour. The sample is then cooled before the last run (same 2θ as above).