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SYSTEMATIC EXPLORATION OF THE KRÍSUVÍK HIGH-TEMPERATURE AREA, REYKJANES-PENINSULA, ICELAND.

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ABSTRACT

The exploration programme consisted of two phases, (1) a surface exploration by means of geophysical, geochemical, and geological methods and (2) drilling of slim 800-1000 m deep, exploratory wells.

Surface thermal manifestations are scattered over an area of 10-20 km^2. Schlumberger soundings indicate a continous hydrothermal reservoir under this area at depths of less than 1000 m. The size of the area is about 40 km^2 as determined by the 30 Ωm isoline at 600 m depth. Drilling of slim wells proved difficult for many technical reasons and not as inexpensive as expected.

All the five deep wells in the area display inverse thermal gradients. Possible explanations are: (1) narrow upflow zone or zones and horizontal hot-water movement at shallow depth and (2) weakening of an intrusive complex heat-source without much decrease in the flow rate of water through the system. Maximum temperature in each well ranges from 180°C to 260°C. There is a positive relation between maximum temperatures and the depth to this maximum. This variation can be explained by mixing of rather saline hot deep-water with fresh warm water in the upflow zones. The distribution of the hydrothermal mineral assemblages does not fit well with the present-day underground temperature distribution. Average porosity in selected samples of core is 11%. The rocks at depths of 3-5 km appear to be relatively porous as inferred from MT measurements. All the exploratory wells penetrated several permeable aquifers. For better understanding of the inverse gradients, a 2000 m deep exploratory well is needed.

INTRODUCTION

In 1970 a paper appeared at the United Nations Symposium on the Development and Utilization of Geothermal Resources, held in Pisa, which described a programme for systematic exploration of the high-temperature areas in Iceland. (Björnsson, 1970). The Krísuvík high-temperature area, southwest Iceland, which this article is concerned with, was the first area whose systematic study was founded on this programme. The exploration of the Krísuvík area revaeled several minor snags in the exploration programme, in particular the drilling of the slim exploratory wells.

The systematic exploration involves mostly two phases, namely (1) surface exploration by geological, geophysical, and geochemical methods, and (2) drilling of slim 800-1000 meters deep exploratory wells. A third phase brigdes exploration and exploitation and involves the drilling of "test wells" which are designed as production wells. The surface exploration was initiated in 1970. At the beginning of 1973 the two phases had been almost completed. One or two 2000 meters deep exploratory wells are still required for a better evaluation of the production characteristics of the area. This requirement may be an exception in systematic exploration of the high-temperature areas in Iceland and is to be related with the inverse thermal gradients observed in every 800-1000 meters deep well. The first phase of the systematic exploration is far the least costly but the last phase most costly. Drilling of one 800-1000 meters deep exploratory well costs some 100.000 U.S. dollars. A 2000 meters deep production well will cost about twice as much. The total cost of the surface exploration is comparable with the drilling of one to two exploratory wells.

Before the systematic exploration was initiated in 1970 considerable research and drilling had been carried out in the geothermal area, including drilling of two, rather closely spaced, 1200 meters deep wells (H.1 and H. 2). Although fragmentary, this previous research provided some valuable information and influenced therefore the details in the set-up of the exploration programme.

2. GEOLOGICAL FEATURES IN THE REGION OF THE HYDROTHERMAL FIELD

The Reykjanes Peninsula on which the Krísuvík high-temperature area is located is characterised by extensive post-glacial lava fields and steep-sided mountains and ridges of pillow lavas, pillow breccias, and hyaloclastites, which protrude through the lava fields. The latter mentioned formations are of upper Quaternary age, in all likelyhood from the last glaciation, formed during volcanic eruptions in melt-water chambers within the ice-sheet. All the rocks are of basaltic composition. The whole of the Reykjanes Peninsula has been mapped by Jónsson (1975).

In the west the Reykjanes Peninsula is low-lying and here the lava fields dominate but towards the east there is a general rise in the topography with simultaneous increase in the proportion of hyaloclasties and pillow lavas at the surface. Numerous volcanic eruptions have occurred in historical times (last 1100 years) on the peninsula, including several in the vicinity of Krisuvík high-temperature area. Minor post-glacial volcanic edifices on fissures are particularly abundant within and in the vicinity of the Krísuvík area.

The post-glacial volcanic activity on the Reykjanes Peninsula consists of fissure eruptions and central eruptions. The latter have led to the formation of rather flat cones or shields. The central eruptions are more voluminous than the fissure eruptions. It seems likely that during prolonged eruptions that the activity tends strongly to become concentrated in a single vent although the magma may initially have protruded through a fissure. This would explain the volume differences of central edifices on one hand and fissure edifices on the other hand.

Four high-temperature areas exist in the Reykjanes Peninsula proper. They all appear to lie on the plate boundary where an en-echelon shaped fault swarm crosses this boundary at an angle of some 30-40° (Klein et al., 1973). The faults which are mostly of the normal tensional type, but also slip-strike, appear to have formed as a result of secondary stress forces set up in the crust by the moving plates (Björnsson, Sv., 1975).

Comparable regional geological setting exists in all the high-temperature areas in Iceland. High-temperature areas outside the Reykjanes Peninsula are located within central volcanic complexes which are characterised by differentiated rocks, intense extrusive magmatic activity as well as cauldron subsidence at a mature stage. These geological features seem to be lacking in all the four high-temperature areas on the Reykjanes Peninsula. As deduced from Tertiary volcanic complexes, which have been exposed by erosion, intrusive activity is also particularly intense within the central complexes.

Sea water percolates into the bedrock under the whole western part of the Reykjanes Peninsula due to the extremely high permeability of the bedrock, particularly the lava flows. This is strongly reflected in the salinity of the geothermal water in the two westernmost high-temperature areas, Reykjanes and Svartsengi. The thermal water in the Krísuvík area is also high in dissolved solids in comparison with the other drilled high-temperature areas in Iceland outside the Reykjanes-Peninsula and it is believed that this high dissolved solids content is due to slight mixing of sea water with fresh water within the Krísuvík hydrothermal system.

Topographically the Krisuvík high-temperature area is characterised by two major southwest striking hyaloclastite ridges (fig. 1). Inbetween the two ridges is a valley covered with lava flows. Most of the surface thermal activity is to be found within the hyaloclastite ridges and on their "outer" sides by the boundaries on the hydrothermal area but in the lava covered valley between the two ridges, thermal activity at the surface is practically non-existing.

Detailed mapping of the easternmost ridge has revealed that it is composed of four farmations and has possibly been built up during four eruptive episodes (Imsland, 1973). It seems likely that the westernmost ridge consists also of more than one formation. It appears therefore that volcanic activity has been most intense on relatively narrow zones represented by the two hyaloclastite ridges.

The faults of the fault swarm that dissects the Krísuvík area are much more dense in the hyaloclastite ridges than in the lava covered valley inbetween. This has been explained by longer exposure of the glacial formations to the tectonic activity than the young lavas. Despite this, it is also possible that the tectonic activity is truly more intense in the ridges where the volcanic activity has been most intense.

As discussed in a later section the inverse thermal gradients observed in every deep well in the area raises major difficulties in defining the production characteristics of the hydrothermal reservoir. It is possible that the hot water rises to the surface under the hyaloclastite ridges in a zone of presumed maximum tectonic activity. No drilling has been performed on these ridges.

Several explosion craters (maars) occur within the Krísuvík field but not in the surrounding area. Some magma has been brought to the surface during the formation of some of these craters. It is considered that this activity results from the crystallisation of shallow magma batches.

3. THE HYDROTHERMAL RESERVOIR AS DEDUCED FROM SURFACE EXPLORATION

3.1 Introduction

The surface thermal manifestations in the Krísuvík area consist of hot ground where the bedrock has been intensily altered by acid surface leaching. In this hot ground steam vents and mud pools are common and locally also solfataras. Alkaline water springs presenting the flashed water fraction of the "deep water" do not exist in the area. Secondarily heated carbonate water appears in warm springs in a few localities. This type of surface thermal activity is to be related with high underground temperatures and low ground-water table. The low ground-water table is believed to reflect the high permeability of the bedrock but not so much the local topography.

Typically the hot ground forms isolated spots few tens of meters across and there may be hundreds of meters between such spots. The greater part of the thermal manifestations are located in two parts of the area, Trölladyngja, near its northwestern boundary and in Sveifluháls and south of lake Kleifarvatn by the southeast boundary of the area.

3.2 Resistivity Survey

About 70 electrical DC depth soundings with the Schlumberger configuration have been taken in order to estimate bedrock resistivity at different depths in the geothermal field. The current arm of the depth profiles was in most cases 900 or 1500 meters.

The resistivity of the surface layers above the groundwater table is usually very high, especially in the postglacial lava fields were it is in the range of 10.000 Ωm. By contrast the resistivity is low, or about 10 Ωm in areas where the surface layer has been strongly altered by acid surface leaching. The resistivity decreases rapidly with depth and within a circular area of about 70 km^2, which encloses all surface thermal activity, it is in the range of 5-50 Ωm at few hundred meters depth (fig. 2). The resistivity map is still somewhat incomplete because only a few profiles have been taken on the hyaloclastite ridges. Surveying there is difficult due to rugged topography. However, dipole-dipole profiling measurements with 1500 meter center spacing along the easternmost ridge, Sveifluháls, showed apparent resistivity of 5 to 8 Ωm under the whole length of the ridge (depth corresponding to about 1000 meters).

There is a reasonably good correlation between measured temperatures in wells and bedrock resistivity by the wells. Therefore, the variation in the bedrock resistivity in the area is considered to reflect variations in underground temperatures and not so much rock porosity and water salinity.

Two separate lows occur within the resistivity anomaly. They are identified by resistivity of less than 10 Ωm at about 600 meters depth. The central portion of both lows coincides with the crest of the two hyaloclastite ridges and both are elongated in the direction of the axes of these ridges. Few deep (over 1500 meters) Schlumberger and dipole-dipole soundings close to the easternmost ridge, Sveifluháls, suggest that the areal extent of the $\leq 10\,\Omega$m low becomes more restricted with depth but still underlies the crest of the ridge. It is possible that the two resistivity lows represent two major upflow zones of thermal water. This has not been demonstrated by drilling.

Bedrock resistivity to depths of some 10 km has been estimated in one locality by means of magneto-telluric measurements (fig. 3). The MT-station was located in the valley between the two hyaloclastite ridges and recorded both the magnetic and electrical fields in the period range from 6 to about 10.000 seconds. The Cagniard estimates for the apparent resistivity were calculated. For the shallower depths the MT-results fit well with the Schlumberger and dipole-dipole soundings. The surface resistivity was found to be about 20.000Ωm. At 30 meters it decreased to 40Ωm, presumably at the ground-water table, and decreased further to some 4Ωm between 150 and 200 meters depth but increased again to 20Ωm and remains so to 2.5-3 km depth. This is somewhat higher than the resistivity at similar depth under the easternmost hyaloclastite ridge. The low resistivity layer at about 200 meters depth may well reflect the zone of maximum underground temperatures, which is recorded in the wells.

Below 3 km depth the resistivity increases rapidly to about 300Ωm. There is some evidence for a thin low resistivity layer on top of the 300Ωm layer, but its true resistivity and thickness is somewhat uncertain.

The top of the deep high resistivity layer coincides with the upper boundary of seismic layer 3 (Pálmason, 1971) (fig. 3). This seismic layer has a P-velocity of about 6.5 km/sec and is believed to represent a metamorphic facies of amphibolite grade or rock formations which are mostly of intrusive origin (Pálmason, 1971; Walker, 1974). Microearthquakes, which identify the plate boundary (Klein et al., 1973) are particularly frequent near the top of layer 3 where the low resistivity layer occurs as suggested by the MT measurements (fig. 3).

The high bedrock resistivity of layer 3 under the Krísuvík high-temperature area is very anomalous in comparison with measured resistivity in this layer under a few other high-temperaute areas and other localities within the active volcanic belts where measurements have been taken (Björnsson, 1975). Here, the resistivity of layer 3 is about 10 Ωm. By contrast the resistivity of layer 3 under the Tertiary flood basalts in Iceland is about 300 Ωm or comparable with that of the Krísuvík area.

The high resistivity of seismic layer 3 under the Krísuvík area may be explained by relatively low temperature of this layer because intrusive activity has ceased some time ago. It is also possible that low overall porosity is responsible for the anomalously high resistivity which would be caused by unusually abundant intrusions.

Airborne magnetic survey did not reveal any magnetic low over the hydrothermal area, a feature which is so distinct for most of the explored high-temperature areas in Iceland. The basaltic rock in Iceland is strongly, although variably, magnetic and during hydrothermal alteration of this rock by the hydrogen sulphide bearing and reducing high-temperature waters the original magnetite of the rock is destroyed, resulting in lowering of the magnetism of the rock. The lack of a negative magnetic anomaly in the Krísuvík area is believed to reflect the lack of any large upflow zone now and in the past where a large body of rock has lost a sufficient part of its original magnetite through alteration to produce a magnetic low.

3.3 Geochemical exploration

In estimation of underground temperatures by geochemical methods, the silica and alkali geothermometers could not be applied because the "deep water" does not reach the surface in springs. It has long been "believed" in Iceland that the hydrogen content of gas emerging in steam vents reflected underground temperature conditions, the higher the hydrogen, the higer the underground temperature. Indeed, a positive relationship between measured temperatures in wells and the hydrogen content of the gas has been observed at Námafjall (Gudmundsson et al., 1971). By applying this empirical relationship underground temperatures were estimated from the hydrogen content of steam vents in the Krísuvík area.

It seems possible that the hydrogen content of the deep geothermal water below the zone of flashing is controlled by temperature dependent equilibrium with water, sulphate and sulphide:

$$2H^{+} + 4H_2 + SO_4^{--} = 4H_2O + H_2S$$

For constant distribution of sulphur atoms between sulphate and sulphide, the variation in the hydrogen concentration depends only on changes in the equilibrium constant with temperature. By applying thermodynamic data of Helgeson (1969) it was found that the calculated variation compared reasonably with the empirical one at Námafjall.

The hydrogen geothermometer gave underground temperature of 235-260°C where well 5 was drilled but the maximum measured temperature in that well was only 183°C. The same figures for wells 1 and 2 were 220-245°C and 222°C and for well 6 they were 215-240°C and 262°C. Although a maximum temerature of 262°C was recorded in well 6, the temperature was only 218°C at the bottom of the well. Underground temperature estimates in the Krísuvík area by the hydrogen geothermometer were, therefore, not successful, although it cannot be said that they were a complete failure. There may be many reasons for the discrepancy, which is particularly large in the case of well 5. Here the gas in steam vents rising from the "deep water" moves through a body of water near the surface which is well below the boiling point. The more water soluble gases, carbon dioxide and hydrogen sulphide, may have dissolved in this water leading to a relative enrichment of hydrogen in the gas that escapes to the surface. The salinity at Krísuvík is also somewhat different from that at Námafjall, which may influence the distribution of sulphur atoms between sulphate and sulphide, so the empirical relation at Námafjall is not well applicable to the more saline waters in Krísuvík. Much more study is required before the hydrogen geothermometer can be considered adequately quantitative.

4. RESULTS OF EXPLORATORY DRILLING

4.1 Outline of Results of Exploratory Drilling

The location of the first three exploratory wells (H.5, H.6, and H.7) was mostly based on the results of the resistivity survey and the interpretation of the hydrogen content in steam vents. The results of previous drillings were also taken into account. The principal aim of the exploratory drillings was to reveal underground temperatures in different parts of the hydrothermal area and "calibrate" the results of the resistivity survey but it was uncertain to what extent the low resistivity was due to temperature, porosity, and salinity of the water. The first well was located just east of the larger $\leqslant 10 \Omega m$ resistivity low where the hydrogen content of gas in the steam vents suggested the highest underground temperature. The second well was located in the westernmost resistivity low and the third well in the valley between the two hyaloclastite ridges, mostly to reveal to what extent water salinity and rock porosity could be responsible for the low bedrock resistivity in relatively large areas where there were no manifestations of surface activity.

After the discovery of the inverse thermal gradients in wells 5 and 6, which had also been observed in two 1200 meters deep wells located some 1000 meters from well 5, the first two models for the hydrothermal system developed. One of them took into consideration the overall circular shape of the resistivity anomaly. It was postulated that a major deep upflow zone existed under the center of the resistivity anomaly and that the hot water would migrate upwards under a small angle from this upflow zone towards the surface. The model of a major intrusion with cone sheets extending from its apex was born in mind in the development of this model. Yet, no geological structures on the surface, such as a caldera formation were observable, but it was believed that it might not be visible, even if it existed, because it was covered by young volcanics. This model explained the inverse thermal gradients in all the deep wells, but wells 5 and 6 were located near the east and west boundary of the resistivity anomaly respectively. Unfortunately well 7 was not sufficiently deep (540 meters) to give a decisive answer in this respect. The other model assumed that more than one upflow zone existed and that all the wells so far had penetrated a mushroom shaped body of hot rock outside the top of these upflow zones.

Bearing in mind these two explanations of the inverse thermal gradients the forth well (H.8) was located in the western part of the eastern $\leqslant 10\Omega m$ low (see figure 2). If separate upflow zones existed it was to be expected that this hole would penetrate the maximum underground temperature but if the "cone sheet model" proved correct, increasing temperatures to the bottom of the hole were expected. As with other deep wells the maximum temperature was penetrated, now at a depth of about 450 meters. Therefore, the "cone sheet model" was rejected. Later the idea of cooling from below of the hydrothermal system has developed.

4.2 Design of Exploratory Drillholes

In designing the exploratory wells a compromise was made between an estimated cost of drilling and the information that could be obtained from the well depending on its design. It was decided to drill slim holes of 3 3/8 inch diameter and cement the anchor casing to a depth of 200-250 meters. A liner to the bottom of the hole was omitted and as a result is was decided not to let the wells "blow" since a non-lined wet steam hole is likely to collapse during discharge. This excluded direct information on the productivity of individual wells but was expected to yield the same information on underground temperatures, geological sections, hydrothermal alteration, and location of aquifers, as indeed was the case. It is considered that sampling from a non-blowing well yields in many respects less reliable samples than from blowing wells due to imperfect sampling technique.

Experience showed that many technical problems rise in drilling slim wells and they did not prove to be as inexpensive compared with wider wells as expected. Firstly, full load on the drillbit is obtained at much greater depth in case of the slim wells which reduces the drillspead at the beginning. Secondly, it was found to be difficult to clog aquifers temporarily during drilling, to recover the cuttings with the circulation water to the surface. Usually aquifers below the anchor casing are clogged successfully during drilling by pumping wood chips down with the circulation water. The failure to do this with the slim wells can be related to the small width of the drillrods and small drillbit. The result was costly cementing of aquifers which were sufficiently permeable for total loss of circulation fluid.

4.3 Hydrothermal alteration

The general pattern of the hydrothermal alteration in wells 5, 6, and 8 in the Krísuvík high-temperature area studied by Gíslason (1973), is similar to that of other drilled high-temperature areas in Iceland (fig. 4). This pattern involves zonal distribution of many of the most distinguished hydrothermal minerals. At the shallowest depths various zeolites form along with smectite. Chlorite replaces the smectite at depths where the zeolites generally dissappear. At the boundary of the smectite and chlorite zones a mixed layer mineral occurs, which appears to be built up of regular stacking of smectite and chlorite. Epidote usually appears at a shallow level in the chlorite layer.

The alteration in the Krísuvík area is anomalous in many respects to the most commonly occuring pattern. The "typical" zeolite zone is lacking although several zeolites, (wairakite, analcime, heulandite, and mordenite) have been reported at different depths in one or more of the wells. The chlorite zone appears at a depth of about 100 meters in well 5, which is unusually shallow. Epidote has only been reported in well 8 at depths of more than about 800 meters.

Other hydrothermal minerals that have been reported from Krísuvík include quartz and calcite, which occur at all depths below about 100-200 meters. Pyrite is also widely dispersed. Hematite occurs locally and siderite and kaolinite have been found down to depths of about 100 meters in well 5. Heulandite occurs above 400 meters depth in wells 6 and 8, wairakite in

2 samples from well 5 at 430 and 700 meters depth. Analcime is widely dispersed in well 6 but has only been identified at a depths of 550 and 850 meters in well 8. Mordenite occurs below 300 meters in well 8.

The different facies of basaltic rock show highly different susceptability to the hydrothermal alteration. Thus, the hyaloclastite tuffs, which are mostly glassy, are altered most easily, but the massive middle parts of basalt lavas are least prone to alteration. The primary rock is as a rule only partially decomposed. Only occasionally is the hyaloclastite tuff so heavily altered that no relicts of glass nor primary crystalline fragments are visible. The glass is most susceptable to alteration being mostly altered into smectite and/or chlorite. Olivine and magnetite are also altered very easily, although primary magnetite is observed at various depths in well 5. Pyroxene is more resistant and usually plagioclase has suffered little alteration.

It seems likely that the hydrothermal mineral assemblage does not represent an equilibrium assemblage. Basaltic rock that is subjected to alteration by hydrothermal fluids may go through increasing alteration stages represented by the depth zoning. The higher the temperature and the more prolonged the geothermal activity, the further will the alteration proceed. The crude zonal arrangement that is observed with depth and therefore underground temperatures is therefore considered to reflect different stages in the approach to a stable mineral assemblage, which is mostly aided by the temperature conditions and the rate of through-flowing water.

The poor relation between the depth zonal arrangement of hydrothermal minerals and the extisting underground temperatures in well 5 in comparison with the most common pattern of high-temperature areas, suggests that the alteration we observe today is a fossil one, bearing evidence of higher underground temperatures in the past. The same applies to well 8 but not for well 6, whose alteration fits reasonably with the most common depth-zonal pattern. The relation between hydrothermal alteration and present-day underground temperatures suggests that different parts of the area pass through their maximum underground temperature at different times in the evolution of the hydrothermal field as a whole.

The unusually high level of the chlorite layer (100 meters depth) in well 5 and the lack of the topmost zeolite layer could be explained by relative high pressure at the surface of the bedrock caused by an overlying ice-sheet at the time of formation of the presently observed hydrothermal minerals. The ice-sheet of the last glaciation retreated from this area 10-12.000 years ago and its thickness in Krísuvík was of the order of 300 meters. The pressure caused by the overlying ice-sheet was therefore about 30 atmosphere pressure corresponding to a temperature of about 235^{o}C for saturated steam. Chlorite appears first at depths in wells at Námafjall, Nesjavellir, and Reykjanes which correspond to temperatures of 240-265^{o}C for saturated steam.

4.4 Thermal Fluid Compositions

Water samples from the exploratory wells were collected at different depths by sinking special sampling apparatus into the well. When the sample is released from the sampling apparatus and transfered to the sample bottle, some of the volatiles may escape, and the results presented in table 1 are not considered reliable with respect to these compounds (CO_2 and H_2S). A few samples were also collected from carbonate springs presenting secondarily heated water.

Thermal waters from wells in the Krísuvík high-temperature area display a large variation in their dissolved solids content (table 1) By contrast, other drilled high-temperature areas in Iceland have a very homogeneous hot-water chemistry. It should, however, be pointed out that wells in Krísuvík are spread over a much larger area than wells in other high-temperature areas so the comparison may not be entirely relalistic. Yet, a concentration range of an element like chlorine of 50-1100 ppm must be considered unusually high for a single hydrothermal system.

In those exploratory wells (H.7 and H.8), which were located some distance away from any surface thermal manifestations, a 200-300 meters layer of fresh and relatively cold groundwater was found to overly the weakly saline water of the hydrothermal reservoir (see table 1). On the other hand, the salinity of thermal water is about the same at all depths in those wells (H.5 and H.6), which are located close to surface thermal activity.

There is a distinct relationship between the salinity (chlorine content) of the thermal water and its temperature as estimated from its unionized silica concentration assuming equilibrium with quartz or chalcedony (fig. 5). Equilibrium with quartz or chalcedony is selected on the basis of assumptions given by Arnórsson (1975). There is generally a good agreement between these equilibrium temperatures and the measured temperatures at the sampling depth in the wells (fig. 5). There is, however, considerable discrepancy between measured temperatures and silica temperatures for some of the lower temperature waters, which are from the carbonate springs. This is explained by conductive cooling of the ascending water without sufficient precipitation of silica to retain equilibrium. Discrepancy between measured temperatures and silica temperatures for 2 points corresponding to water at about 120°C is considered to be due to cooling by flashing without much silica precipitation.

The salinity-temperature relationship presented in figure 5 can be explained by mixing of weakly saline, hot-water with fresh water. It is considered that the salinity of the hot-water is of marine origin. The fresh water may have become heated to some extent before the mixing took place. In figure 5 the topmost line indicates chlorine-temperature relationships in water mixtures derived from dilution of weakly saline hot-water at 260°C with cold water (5°C), where the hot-water contains 1100 ppm of chlorine but the cold water 10 ppm. The lower line shows, on the other hand, dilution with water at 150°C containing 100 ppm of chlorine. The chlorine contents of the 260° and 150°C waters correspond with those from wells 6 and 5 respectively. If the 260°C water is the hottest participating in the mixing, one would expect that the distibution of the points of figure 5 would fall approximately inbetween the two lines. This is not the case, which may be explained by assuming that the original hot-water had a temperature of more than 260^{o}C, may-be 280^{o}C for chlorine content of 1100 ppm. A higher chlorine concentration would require a still higher temperature of the original hot-water.

Individual hydrothermal systems have been indentified by constancy in the Cl/B ratio of their thermal waters (Arnórsson, 1970). Arnórsson (1970) summarised those factors expected to influence the chlorine and boron concentrations in thermal waters. Waters from wells in the Krísuvík area show considerable scatter in their Cl/B ratio indicating non-constancy in some of the variables that influence the chlorine and boron concentrations in the waters (fig. 6).

The content of boron is not highest in those waters where the chlorine content is highest, which indicates that the variation in the Cl/B ratio cannot be caused by mixing of fresh water with saline water deriving its salinity from sea water.

Relatively constant Cl/B ratio is obtained for samples from different depths in the same well despite variations in the concentrations of the two elements. Such variations are explained by dilution of the thermal water with fresh groundwater. It is interesting that such dilution has occured near the bottom of wells 5 and 6, below the temperature maximum (see section 4.5). On the other hand dilution occurs in the uppermost part of well 8 corresponding to the 200 meters layer of cold groundwater.

There is an apparent tendency for the Cl/B ratio of the thermal waters to decrease from east (well 5) to west (well 6) across the geothermal area. The reason for this trend is not understood. It seems, however, possible that boron becomes more easily leached from the rock with progressive alteration so the mentioned trend would reflect decreasing alteration from east to west.

If the explanation given above on the temperature-salinity relationship is correct, the hydrology of the Krísuvík reservoir will be quite different from the generally accepted picture of high-temperature hydrothermal reservoirs in Iceland. These reservoirs are believed to be capped with any impermeable layer originally formed by precipitation of minerals from the water at the interface of cold groundwater and the ascending thermal water. The difference between Krísuvík and the most common situation could be that Krísuvík presents a cooling, mature hydrothermal system. In such case the capping layer would be opened through tectonic movements, allowing cold water to seep into the reservoir, and not sealed off again because of decreasing geothermal activity.

4.5 Inverse Thermal Gradients and Models for Water Movement

Every well in the Krísuvík area, which penetrates to depths greater than about 800 meters, displays inverse thermal gradient (fig. 7). These wells are spread over the larger part of the area so the inverse gradient cannot be a localized phenomonon. The maximum temperatures are recorded at depths of 200 to 500 meters and there is a distinct tendency for the lowest maxima to occur at the shallowest depths and highest maxima to occur at the greatest depths. Below the maxima the temperatures are remarkably constant to the bottom of each well. The temperature difference between the maximum and the straight line below bears a positive relation with the temperature maximum. Thus, in well 6, which has the highest recorded temperature of 262°C, the constant temperature below the maximum is about 218°C or 44°C lower. By constrast the maximum temperature in well 5 is only 183°C and the constant temperature below the maxium is 153°C or 30°C lower.

Relatively cold water "floats" on top of the hot-water in wells 7 and 8. These wells are located some distance away from any surface thermal activity on the boundary of the lava fields inbetween the two prominent hyaloclastite ridges in the area (see figs. 1 and 2). Thermal gradients are extremely high near the bottom of the cold water lense or as much as 90°C per 100 meters. This cold layer and the high thermal gradient at its bottom are taken to indicate insignificant vertical flow of water in this part of the hydrothermal field.

Approximately constant temperatures in the hydrothermal areas in Iceland have been interpreted by substantial vertical movement of the thermal water. The extremely small variation in temperatures with depth below the maxima in all the deep wells in the Krísuvík area remind inevitably of temperature distribution considered to be caused by vertical movement of water.

After drilling of the four exploratory wells two hypotheses have been developed to explain the observed temperature gradients. They are: (1) separate upflow zones not yet localized by wells and a mushroom shaped body of hot-water and rock at the top of these zones and (2) gradual cooling from above and below of the hydrothermal reservoir by relatively fresh water which is replacing an originally more saline water. If the former hypothesis is correct, the constant temperature distribution would be explained by "sinking" of flashed water which moves outwards away from the upflow zones as the flashing takes place. This hypothesis would also explain the temperature maxima in relation to their depth. The hot-water chemistry fits both hypotheses equally well.

The second hypothesis implies that the heat source of the hydrothermal system is dying, but at the same time the flow rate of water through the roots of the system has not decreased significantly relative to the cooling, so the temperature of the ascending water becomes lower. The cooling would also tend to diminish the ascending flow and explain the beforementioned lense of cold water on top of part of the hydrothermal reservoir.

The second hypotheses gives a better explanation of the distribution of the hydrothermal minerals in relation to the present-day underground temperature conditions. The high bedrock resistivity of seismic layer 3 at depths of more than some 3 kilometers, as indicated by the MT measurements, is also in favour of the second hypothesis. This high resistivity fits well with a dying or dead heat source as it is comparable with the resistivity of seismic layer 3 under the Tertiary flood basalts.

4.6 Permeability and Porosity

The only information that is available on permeability has been inferred from loss of circulation fluid during drilling. In Iceland water is always used as drilling fluid and aquifers struck during drilling in the high-temperature areas may always be located by loss of the circulation fluid. When the cold drilling fluid is replacing an originally hot-water, movement of this water into the aquifer is caused by the higher pressure of the water column in the well, which is determined by its higher density. Experience has shown that there is a crude relationship between the total loss of circulation fluid in one well and its mass flow.

The number of aquifers penetrated by each well is highly variable (fig. 8). Considering only aquifers below 250 meters depth, since shallower aquifers will be sealed by the anchor casing, well 5 struck 1 aquifer at 460 meters depth with a circulation loss of more than 18 liters/sec, whereas well 8 penetrated a total of 11 aquifers in the depth interval of 440-860 meters. The total loss of circulating fluid in this well was 120 liters/sec. The present drillhole data are not sufficient to conclude that one part of the hydrothermal field may be more productive than the rest. In this respect, however, it must be concluded that the results for well 6 are positive and very positive for well 8. What experience is available from deeper drilling in other high-temperature areas in Iceland indicates that the number and size of aquifers will not be less below 1000 meters depth than it is at shallower depths. As indicated by figure 8 it is also apparent that far the greater part of the aquifers are located in basalt lavas, an observation that conforms with results in other high-temperature areas.

Porosity was measured in selected samples of core from different depths in wells 5, 6, and 8 (table 2). These measurements take only to primary pores, most of which, if not all, are of vesicular origin. It is very difficult to estimate that fraction of the total porosity, which is of secondary origin, caused by jointing and tectonic movement. No attempt has been made here to do such an estimate. The figures for the porosity given in table 2 should therefore be regarded as minimum ones.

The porosity is highest on average in cores from well 5 and lowest in cores from well 8, or quite the reverse of the permeability as deduced from the loss of circulating fluid during drilling. This probably demonstrates the fact that it is the perviousness of the rock that determines its permeability and there may be no relation between vesicular porosity and permeability.

The average porosity for all the wells is 11%. On the basis of this average, it is evident that each cubic kilometer of bedrock contains at least o.11 cubic kilometers of water. The size of the hydrothermal area as fixed by the $30\Omega m$ isoline at 600 meters depth, is about 40 km^2. Each section of 1 kilometers thickness of the hydrothermal reservoir contains therefore at least 4.4 km^3 of water. This amount of water corresponds to a flow rate of 1400 liters/sec over a period of 100 years. This figure must not be taken as any estimate of the productivity of the hydrothermal reservoir but should give an indication of its size.

5. PRODUCTION CHARACTERISTICS

The productions characteristics of the Krísuvík high-temperature area are still quite uncertain due to the inverse-thermal gradients observed in every deep well. The two hypotheses put foreward in the previous section imply very contrasting production characterists and influence for that reason the potential utilization of the area. If the hypothesis of separate upflow zones is correct, one could expect that deep production drilling could recover water of at least 260°C and the area could be exploited for electric production or some industrial purposes requiring high pressure steam. If, on the other hand, the hypothesis of a dying heat source is correct, insignificant amount of water at temperatues above about 230°C is available for exploitation and the area would not be feasible for any large scale steam production. Large quantities of water at 150-230°C are, however, available in the uppermost 1-2 kilometers of the system and the area could be feasable for large scale exploitation of water for space heating. The geothermal water could in such case be either mixed directly with cold water to obtain the appropriate temperature or the heat could be transferred into cold water by means of some heat exchange. If the temperature is assumed to be 200°C, the quantity of water in 1 km thick layer of the reservoir issufficient for a power plant of about 500 MW thermal for the mentioned period of 100 years. For comparison it may be mentioned that Municipal Heating District Service in Reykjavík supplies today water equivatent to 340 MW thermal for the 80.000 inhabitants of the city. These figures do not imply that the quanity of water represents the quantity available for exploitation but they are meant to give an indication of the size of the hydrothermal reservoir,

In order to reveal better the production characteristics of the area and test the two models a 2000 meters deep well is required. This well should be located inside one of the resistivity lows (see figure 2). In choosing between the two areas and selecting the drilling site, it will be necessary to make a compromise between the geographical situation influencing the drilling cost and interests in either of the areas with respect to potential exploitation. Penetration of the maximum underground temperature by a deep well would be in favour of the "dying heat source" hypothesis whereas increasing temperatures with depth or very high temperatures at the bottom of the hole would be in favour of the "separate upflow zone" hypothesis.

CAPTION TO FIGURES

(1) Geological map of the Krísuvík high-temperature field and the surrounding area. Mapped by J. Jónsson, National Energy Authority, Reykjavík.

(2) Electrical resistivity at 600 meters depth as determined by depth soundings with the Schlumberger configuration. Location of exploratory wells is also shown.

(3) Comparison between bedrock resistivity, P-velocity, and distribution of microearthquakes to depths of 10 km beneath the Krísuvík area.

(4) Distribution of "zonal" hydrothermal minerals from several geothermal areas in Iceland. Data from Gíslason (1973), Sigvaldason (1963), and Tómasson and Kristmannsdóttir (1972).

(5) Relation between temperature and salinity of water from the Krísuvík high-temperature area.

(6) Cl/B relationships in water from various depths in deep wells in the Krísuvík high-temperature area.

(7) Temperatures in wells in the Krísuvík high-temperature area. The location of the wells is given in figure 2.

(8) Geological sections showing aquifers in exploratory wells H.5, H.6, and H.8. The location of the wells is shown in figure 2.

(9) Scematic sections showing possible models for water movement characterisising the hydrothermal reservoir.

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TABLE 1

Chemical composition of thermal waters from the Krísuvík high-temperature area. Samples from wells were collected at different depths by sinking special samling apparatus into the non-discharging well. Concentrations in ppm.

Location	pH/°C	SiO_2	B	Na^+	K^+	Ca^{++}	Mg^{++}	CO_2 [x] (total)	SO_4^{--}	H_2S [xx] (total)	Cl^-	F^-	diss. solids	Sampling depth m	Temp. °C at sampl. depth
spring	6.48/23	133		42.0	3.5	32.4	32.5	14.5	29.0	<0.1	42.3	0.10	552		
spring	6.98/23	77		34.3	2.8	52.4	21.2	14.1	69.5	<0.1	21.2	0.10	426		
well 3	7.02/20	142			10.8	24.6	1.4	316.0	31.3	<0.1	34.4	0.10	610	200	105
well 3	6.91/20	154		148.0	12.8	19.1	1.3	310.0	31.9	<0.1	70.4	0.15	654	300	115
well 5	8.35/20	222	1.25	206.0	21.5	9.4	0.24	96.0	157.0	3.3	102.0	0.34	856	200	173
well 5	8.80/20	226	1.23	205.0	12.9	9.1	0.24	55.0	175.3	0.3	122.0		850	350	154
well 5	8.80/20	220	1.18	210.0	13.8	10.6	2.22	63.0	178.2	4.4	118.0	0.45	861	470	151
well 5	8.60/20	210	1.15	200.0	12.9	10.0	0.23	72.0	141.6	0.2	151.5		822	650	151
well 5	8.85/20	164	0.59	233.0	16.7	16.5	0.51	63.0	324.7	1.3	52.0	0.60	896	800	151
well 6	7.85/20	205	0.65	680.0	40.4	90.8	0.50	55.7	103.1	<0.1	1234.0	0.20	2563	200	183
well 6	8.35/20	514	0.79	700.0	119.0	42.4	0.38	62.2	49.6	6.6	1094.0	0.50	2605	500	258
well 6	7.30/20	304	0.39	596.0	64.0	40.0	0.44	59.5	40.1	1.7	914.0	0.30	2020	800	218
well 7	8.00/20	50		30.6	1.7	18.4	10.4	110.0	7.9	<0.1	16.1	0.20	208	325	30
well 7	7.15/20	178	0.14	160.0	8.2	15.3	1.4	120.0	75.1	<0.1	163.2	0.40	692	475	139
well 8	8.90/18	210	0.24	140.0	8.3	5.5	0.30	66.9	90.0	<0.1	96.8	0.70	600	240	129
well 8	6.82/18	332	0.52	227.0	21.5	11.5	0.13	121.0	240.0	<0.1	246.4	0.70	1000	450	192
well 8	6.83/18	332	0.50	230.0	20.8	12.5	0.14	117.5	106.3	<0.1	245.6	0.80	1000	700	184
well 8	6.90/18	298	0.53	226.0	21.3	17.0	0.23	121.0	104.0	<0.1	244.0	0.80	1030	920	170
well 14	9.23/23	490	1.77	465.0	57.2	10.4	0.04	49.8	92.4	9.9	759.0	0.30	1876	flashed to 100°C	

[x] $H_2CO_3 + HCO_3^- + CO_3^{--}$

[xx] $H_2S + HS^- + S^{--}$

TABLE 2

Porosity of selected samples of core from exploratory wells. Measurements carried out by Sv. Pálsson.

well 5		well 6		well 8	
porosity/depth m		porosity/depth m		porosity/depth m	
12%	428 m	16%	412 m	5%	377 m
19%	670 m	11%	413 m	10%	736 m
		11%	625 m	8%	930 m
		7%	821 m		


N
ÓBRINNISHÓLAR
KAPELLUHRAUN
UNDIRHLÍÐAR
LANGAHLÍÐ
SNÓKAFELL
SÓLEYJARKRIKI
HRAUNHÓLL
VATNSSKARÐ
FJALLIÐEINA
MÁVAHLÍÐAR
HRÚTAGJÁ
SANDFELLSKLOFI
VATNSHLÍÐARHORN
LAMBAFELL
HÖRÐUVELLIR
HÖSKULDARVELLIR
ELDBORG
HELLUTINDAR
NORÐLINGAHÁLS
LAMBHAGI
TRÖLLADYNGJA
GRÆNADYNGJA
KÖLDUNÁMUR
INNRI-STAPI
FÍFLAVALLAFJALL
FOLALDADALIR
STAPAHRAUN
ODDAFELL
SOG
HRÚTAFELL
HOFMANNAFLÖT
KLEIFARVATN
DRIFFELL
SYÐRI-STAPI
FAGRADALSFJALL
DJÚPAVATN
MIÐDEGISHNÚKUR
HVERINN EINI
GRÆNAVATN
TRAÐARFJÖLL
LAMBATANGI
HVANNAHRAUN
GEITHÖFÐI
BLEIKHÓLL
HRAUNSELSVATNSFELL
HVAMMHOLT
HVAMMAR
KRÓKAMÝRI
ARNARVATN
STÓRA-LAMBAFELL
VIGDÍSARHÁLS
ÍRPUHÓLL
LITLA-LAMBAFELL
SELTÚN
VIGDÍSARVELLIR
AUSTURENGJAHVERIR
HVERADALUR
SLAGA
KRÍSUVÍK
GESTASTAÐAVATN
GRÆNAVATN
SANDFELL
VESTURHÁLS
SVEIFLUHÁLS
STÓRAAUGA
NÓGDI
STÓRI HAMRADALUR
MÆLIFELL
BÆJARFELL
ARNARFELL
GEITAHLÍÐ
LATSFJALL
NÚPSHLÍÐ
STÓRA-ELDBORG
ÖGMUNDARHRAUN
LITLA-ELDBORG
SELALDA
SKRIÐA
KRÍSUVÍKURBJARG
0 1 2 3 km
LEGEND:
Hyaloclastites
Pillow lava
Pyroclastics
Lava scoria
Basalt
Lake
Lava from fissures
Lava shields
Dyngja
Craters and crater rows
Explosion craters
Faults a with displacement, b without displacement
Steam vents
Surface alteration
Road

Fig.2

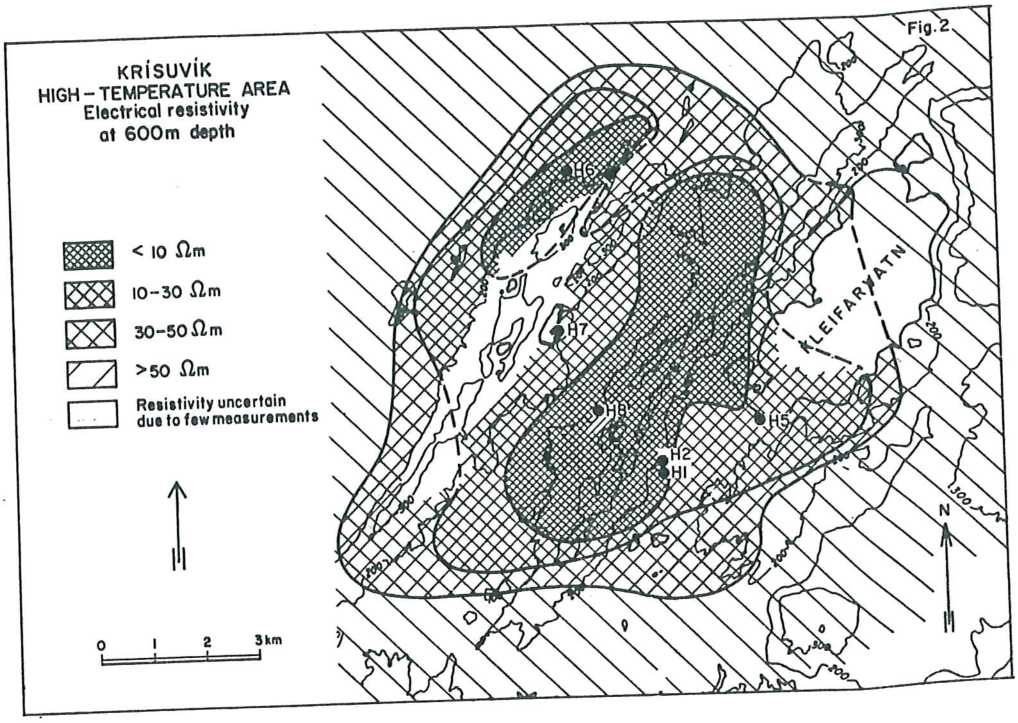
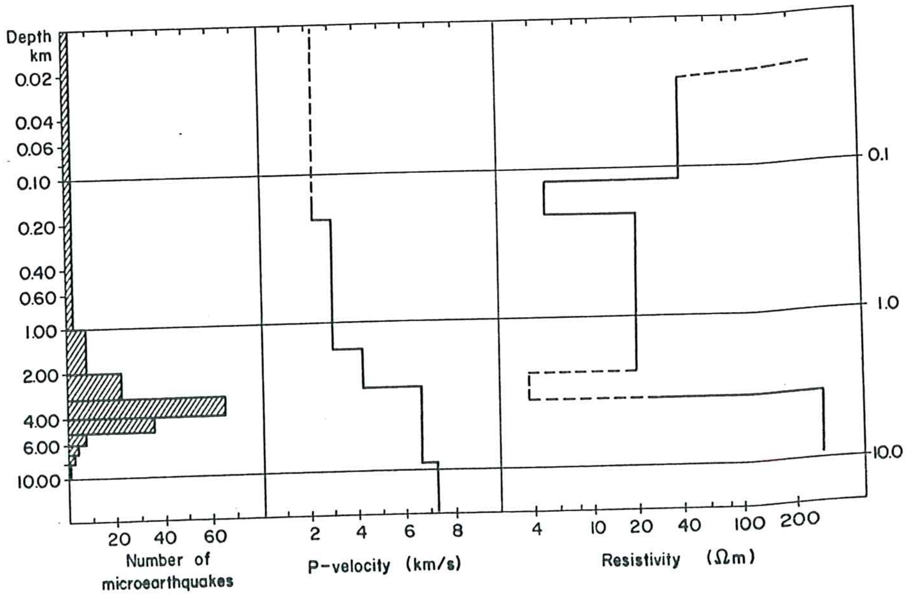

KRÍSUVÍK
HIGH-TEMPERATURE AREA
Electrical resistivity
at 600m depth
< 10 Ωm
10-30 Ωm
30-50 Ωm
>50 Ωm
Resistivity uncertain
due to few measurements
0 1 2 3km
KLEIFARVATN
H1
H2
H5
H6
H7
H8
N
200
300

Fig.3


Depth
km
0.02
0.04
0.06
0.10
0.20
0.40
0.60
1.00
2.00
4.00
6.00
10.00
20 40 60
Number of
microearthquakes
2 4 6 8
P-velocity (km/s)
4 10 20 40 100 200
Resistivity (Ωm)
0.1
1.0
10.0

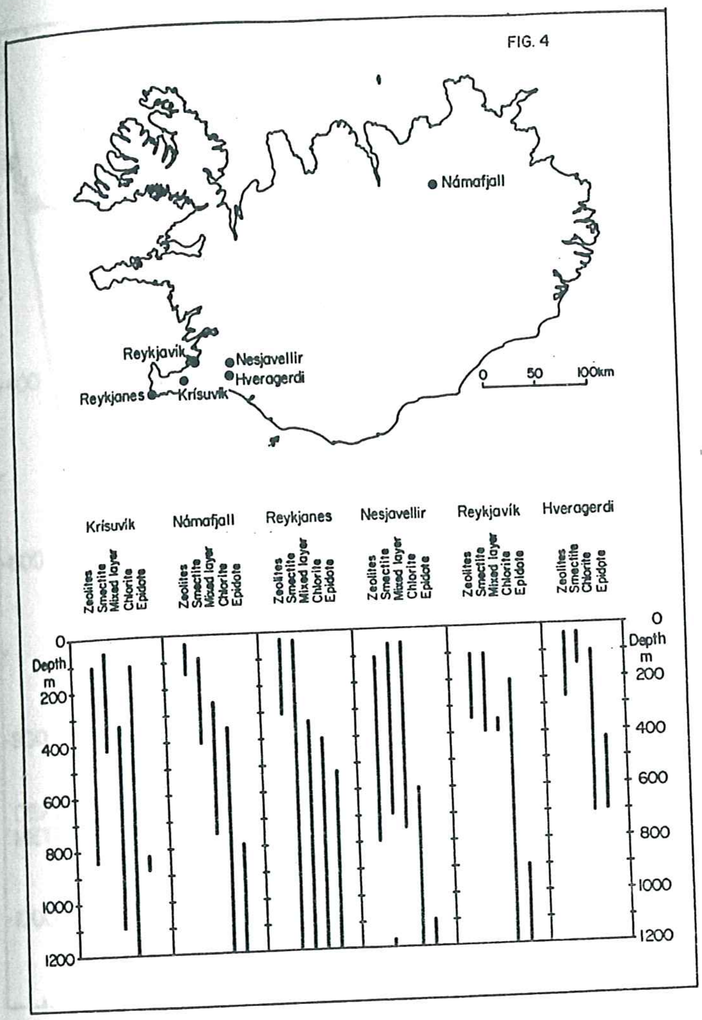


FIG. 4
Námafjall
Reykjavik
Nesjavellir
Hveragerdi
Reykjanes
Krísuvík
0 50 100km
Krísuvík Námafjall Reykjanes Nesjavellir Reykjavík Hveragerdi
Zeolites
Smectite
Mixed layer
Chlorite
Epidote
Depth m
0
200
400
600
800
1000
1200

FIG.5

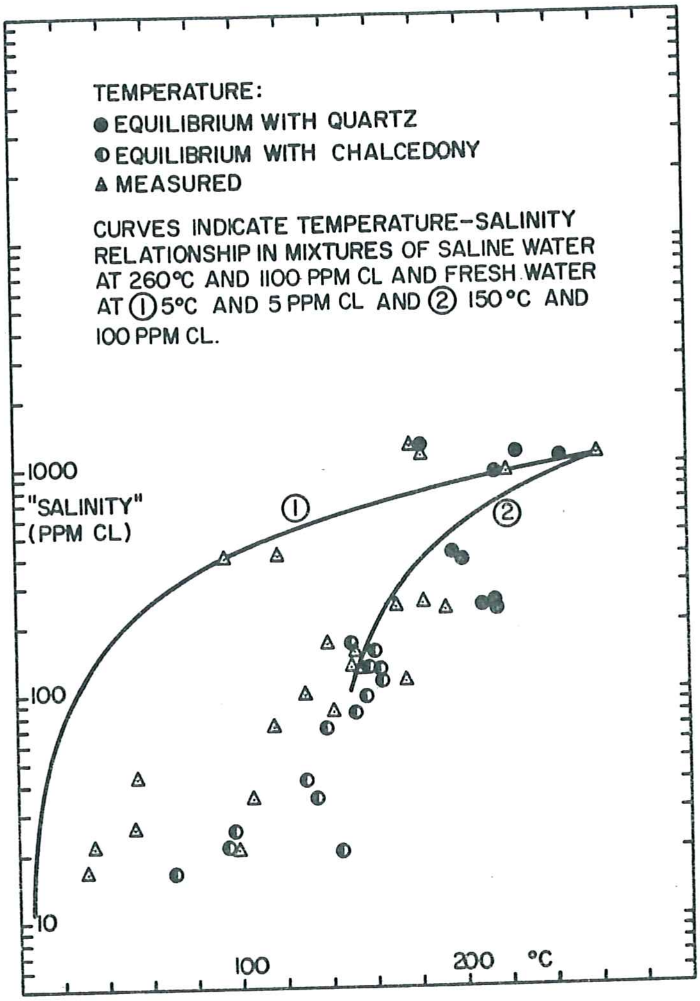

TEMPERATURE:
● EQUILIBRIUM WITH QUARTZ
◐ EQUILIBRIUM WITH CHALCEDONY
▲ MEASURED
CURVES INDICATE TEMPERATURE-SALINITY RELATIONSHIP IN MIXTURES OF SALINE WATER AT 260°C AND 1100 PPM CL AND FRESH WATER AT ① 5°C AND 5 PPM CL AND ② 150°C AND 100 PPM CL.
"SALINITY" (PPM CL)
1000
100
10
①
②
100
200
°C

FIG.6

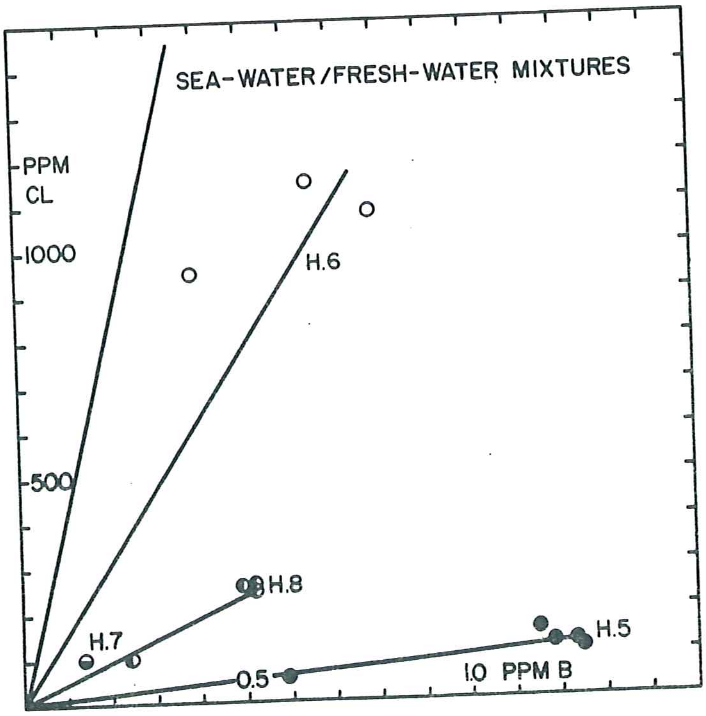

SEA-WATER/FRESH-WATER MIXTURES
PPM CL
1000
500
H.6
H.8
H.7
H.5
0.5
1.0 PPM B

FIG.7

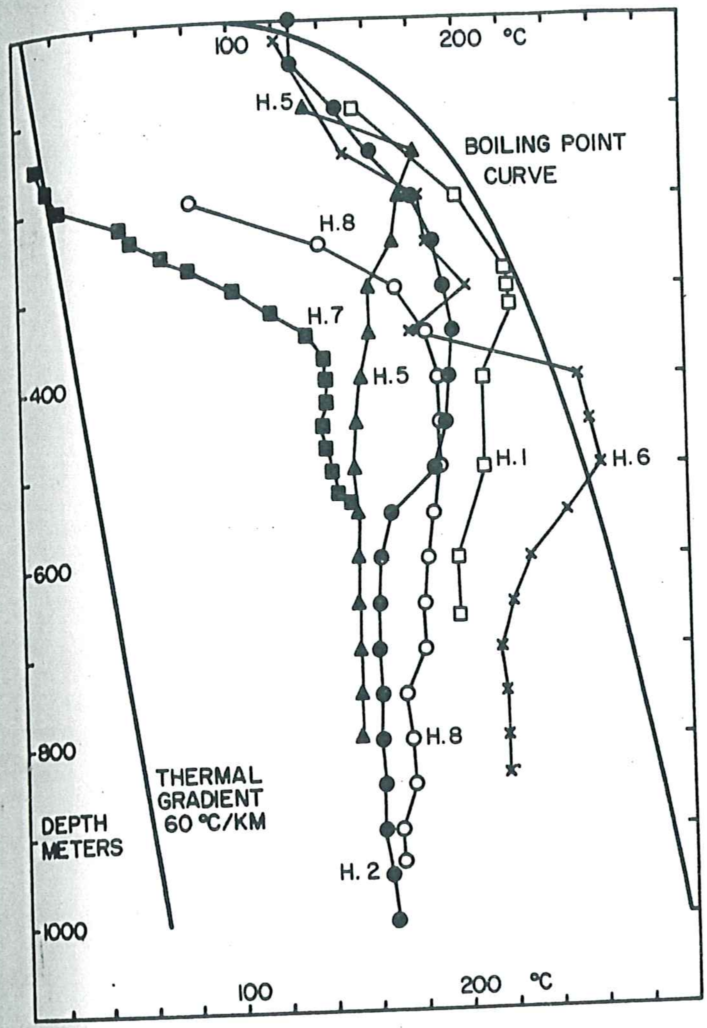

100
200 °C
H.5
BOILING POINT CURVE
H.8
H.7
H.5
H.1
H.6
400
600
800
1000
DEPTH METERS
THERMAL GRADIENT 60 °C/KM
H.8
H. 2
100
200 °C

FIG.8

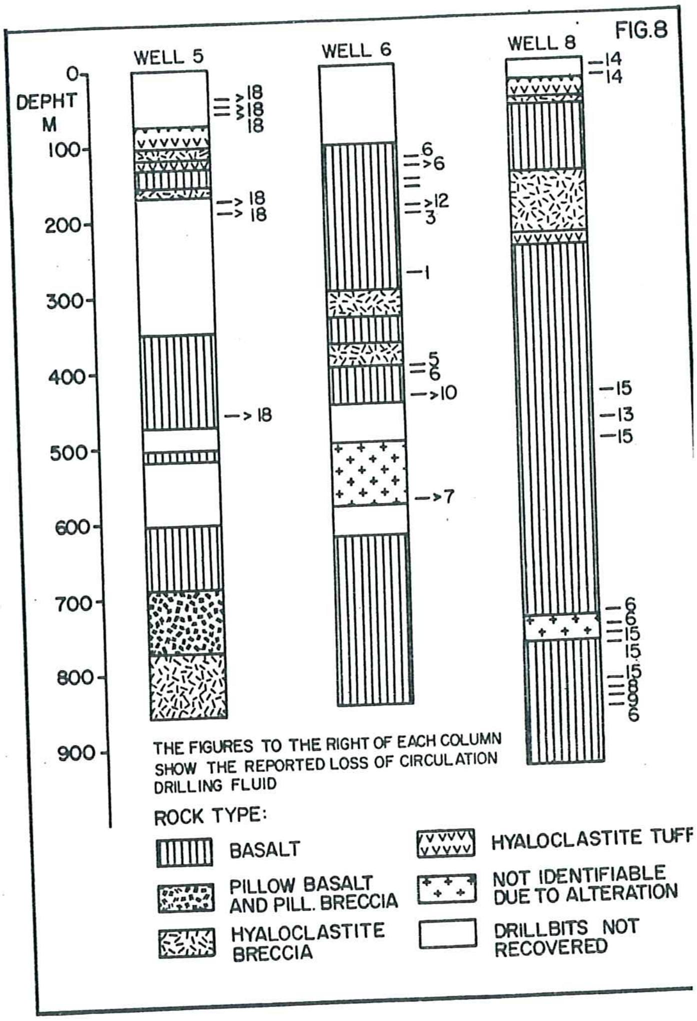
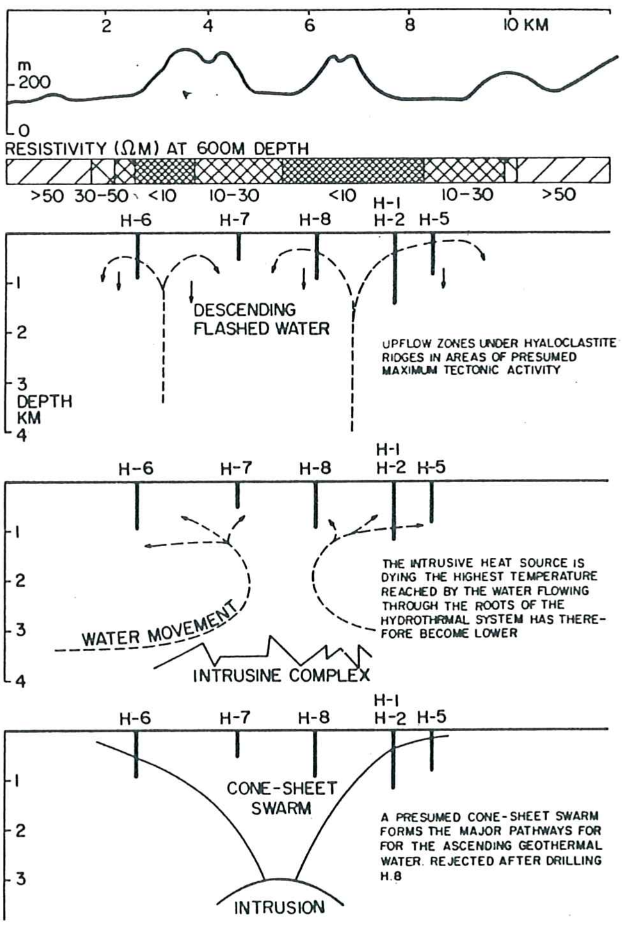

WELL 5
WELL 6
WELL 8
DEPHT M
0
100
200
300
400
500
600
700
800
900
THE FIGURES TO THE RIGHT OF EACH COLUMN SHOW THE REPORTED LOSS OF CIRCULATION DRILLING FLUID
ROCK TYPE:
BASALT
PILLOW BASALT AND PILL. BRECCIA
HYALOCLASTITE BRECCIA
HYALOCLASTITE TUFF
NOT IDENTIFIABLE DUE TO ALTERATION
DRILLBITS NOT RECOVERED

FIG 9



JARDEFNA FR TNR160 J-YM TNR 475 FNR 12795