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GEOPHYSICAL RECONNAISSANCE STUDY OF THE HENGILL HIGH-TEMPERATURE GEOTHERMAL AREA, SW-ICELAND

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ABSTRACT

The active central volcano Hengill,within the Neovolcanic zone in SW-Iceland, is intersected by a major NE-SW trending fissure swarm. There is an abundance of hot springs and fumaroles within a high-temperature geothermal area, which is characterized by a low-resistivity anomaly some 120 km^2 in areal extent. The central hottest part is about 50 km^2 and is indicated by a resistivity-high at depth. An aeromagnetic map shows negative anomalies, caused by hydrothermal alteration. The best geothermal prospects are presumably within the fissure swarm outside the central region.

INTRODUCTION

A high-temperature geothermal area is within the active central volcano Hengill in SW-Iceland some 40-50 km east of Reykjavik and is near the most densely populated part of the country. Therefore, the geothermal field is economically important for an extension of the Reykjavik district heating service, industrial use and electricity production in the coming years. Deep drillholes have only been drilled in two sub-fields, Nesjavellir and Hveragerdi near the boundaries of the geothermal area. An intensified exploration has been initiated in order to investigate the geothermal area. This paper describes the geophysical work in this area, completed or in progress, and plans for further research are discussed briefly.

GEOLOGICAL SETTING AND TECTONICS

An active zone of rifting and recent volcanism, the Neovolcanic zone, crosses Iceland from southwest, where it connects with the Reykjanes ridge, towards the north-east. In south Iceland it has a NE-SW direction, devides into two separate parallel zones, and is characterized by several fissure and fault swarms, most of them passing through a central volcano (Saemundsson, 1978). Nearly all high-temperature areas in Iceland are situated within a central volcano or a fissure swarm, indicating that the main heatsource must be magmatic intrusions in the upper crust. The Hengill high--temperature area is located in the western brance of the Neovolcanic zone. The geology of the Hengill area is characterized by a central volcano named Hengill, which rises about 500 m above the surroundings. The volcano is intersected by a fissure swarm which has the structure of nested grabens. The majority of faults, fissures and hyaloclastite ridges strike NE-SW, but some linear transverse structures are also found near the center of Hengill having a NW-SE direction (Saemundsson, 1967). There are numerous fumaroles in the area, distributed on a narrow zone from Nesjavellir to Hengill. From there a clear lineation of fumaroles extends to SE and forms a continuation of the NE-SW fissures dissecting Hengill. Single fumaroles are in other places for example in the fissure swarm SW of Hengill near Hveradalir (Figure 1).

DC-RESISTIVITY SOUNDINGS

In the last four decades the Schlumberger sounding method has been the most powerful geophysical tool in geothermal exploration in Iceland. The measurements are usually made with current arms up to 1500 m. The interpretation for layered earth is made by use of an inversion computer program. Information about deeper layers are obtained by dipole-dipole equatorial soundings with spacings between potential and current electrodes up to 7 km. About 70 Schlumberger- and dipole-soundings were made in the Hengill geothermal area, which is associated with a highly conductive layer at depth, with a resistivity lower than 20 Ωm. The depth to this layer varies from 100 m beneath Hengill to a few hundred meters near the boundaries of the geothermal area. Outside the area the resistivity is higher. Figure 2 shows location of Schlumberger soundings and a resistivity low at 200 m depth beneath sea-level (b.s.l.). All surface geothermal manifestations are within the boundaries of this anomaly. At 400 m depth b.s.l. the low resistivity area becomes more elongated along the main fissure swarm (Figure 3). The low-resistivity anomaly in the upper few hundred meters is interpreted as caused by high porosity and high temperature (up to 300°C) and also by ionic conduction in thermally altered rocks. The results of the dipole-dipole soundings are in good agreement to the Schlumberger data and show that the low-resistivity boundaries at 400 m depth b.s.l. extend to a greater depth. Seven dipole-dipole soundings, made near the center of the volcano, show, however, increasing resistivity with depth beneath 500-700 m b.s.l. The central high-resistivity anomaly is very likely caused by lower permeability due to dense intrusions and by temperatures higher than 300-400°C. According to Quist and Marshall (1968) the conductivity of solutions has a maximum values around 300-400°C. This inner high-resistivity anomaly is shown in Figure 3 and outlines presumably the center and the hottest part of the geothermal system.

AEROMAGNETIC MEASUREMENTS

An aeromagnetic map, flown at an altitude of 800m above sea-level (Figure 4) shows a positive anomaly, trending NE-SW along the young volcanic zone. To the northwest and southeast the magnetic intensity is lower. Within the central positive anomaly there are two distinct negative anomalies. One is nearly circular and is situated on the main fissure swarm south of Hveradalir. The other has a NW-SE direction and dissects Hengill near its center and coincides with tectonic features having this direction. The negative magnetic anomalies are probably caused by hydrothermal alteration of magnetic minerals, due to high temperature of the water and presence of volcanic gases.

MAGNETOTELLURIC MEASUREMENTS

In a MT-survey across the volcanic zone in SW-Iceland, an anomalous conductive layer was found at about 8 km depth beneath the Hengill area. The depth to it increases to 20 km in about 30 km distance from the rifting axis. Similar results were found in NE-Iceland by Beblo and Bjornsson (1980). They interpreted the conductive layer to consist of partially molten basalt at the crust-mantle interface. The measurements did not reveal an upper crustal magma chamber beneath Hengill.

SEISMICS

A microearthquake study is presently being made in cooperation with the University of Iceland. The first results, reported by Foulger and Einarsson (1980), show high seismic activity and some correlation of epicenters with the distribution of fumaroles. Recordings at 20-30 stations simultaneously will be done in 1981, in order to achieve information about active fissures, faults and possible magma chambers.

FUTURE PLANS

A gravity survey of the whole of the Hengill area in a grid with spacing of 1 km is planned for 1982. Further exploration drillholes will be drilled in Nesjavellir this year and drillholes are planned in the central region and in the fissure swarm to the south in 1982-1984. An extensive investigation of water chemistry is proposed.

DISCUSSION

The low-resistivity anomaly at 400 m depth shows a geothermal area of about 120 km^2. The hottest central part is about 50 km^2 and is characterized by a resistivity high. A fissure swarm dissects the volcano. A comparison of these results with the structure of the Krafla geothermal area in N-Iceland indicates that the best geothermal prospects are most likely within the fissure swarm outside the central region. The permeability is probably lower and there is less ground water flow to transport heat from depth in the central topographic high than in the fissure swarm. Further difficulties could occur due to too high temperature and presence of volcanic gases. This has to be tested with deep drillholes. Like in many other high-temperature areas in Iceland the Hengill area is characterized by a transverse tectonic features perpendicular to the main fissure swarm. The existance of the central volcanoes in general and hence the geothermal high--temperature areas is probably somehow correlated to an intersection of two major lineations in the crust.

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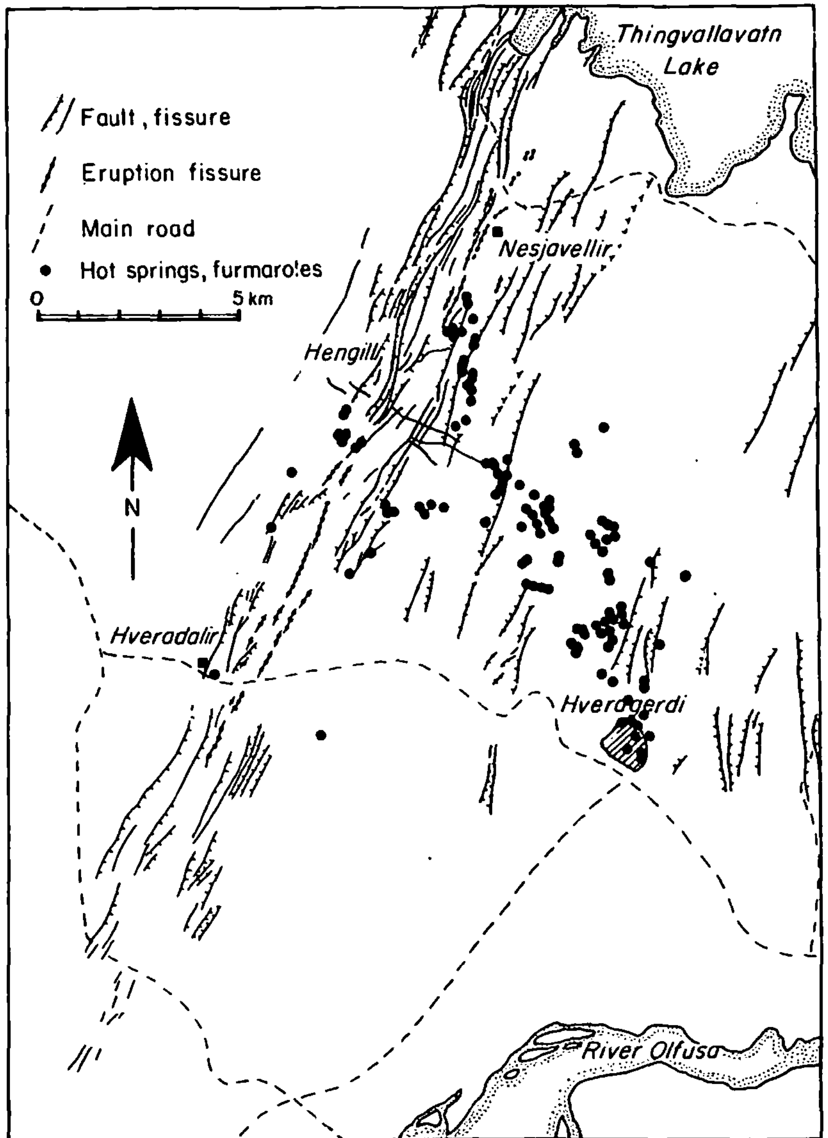

Figur 1. Map of the Hengill area showing the fissure swarm and the distribution of hot springs and fumaroles. From Saemundsson (1967)

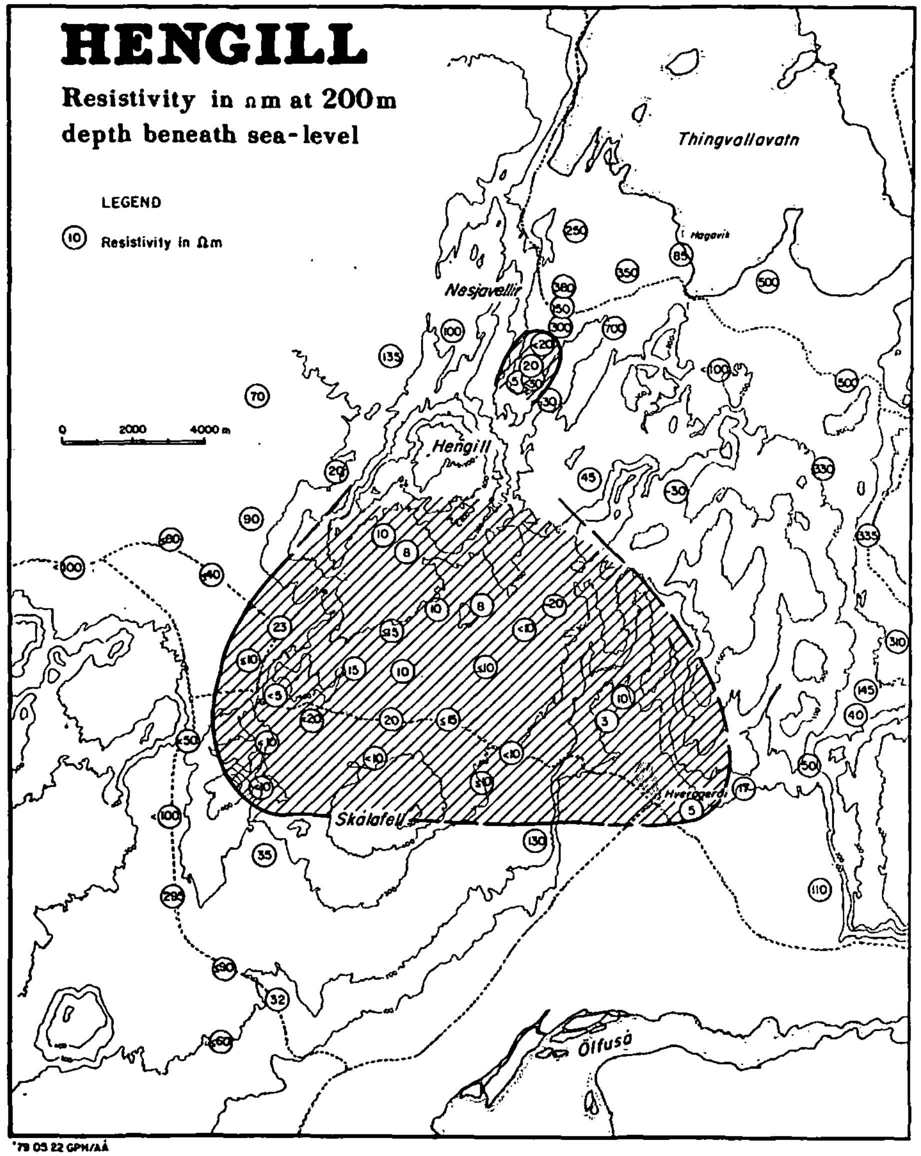

Figure 2. The numbers denote resistivities in Ωm, obtained from Sclumberger-soundings. Nearly all geothermal surface manifestations are within the shaded area, which indicates resistivities lower than 20 Ω m.

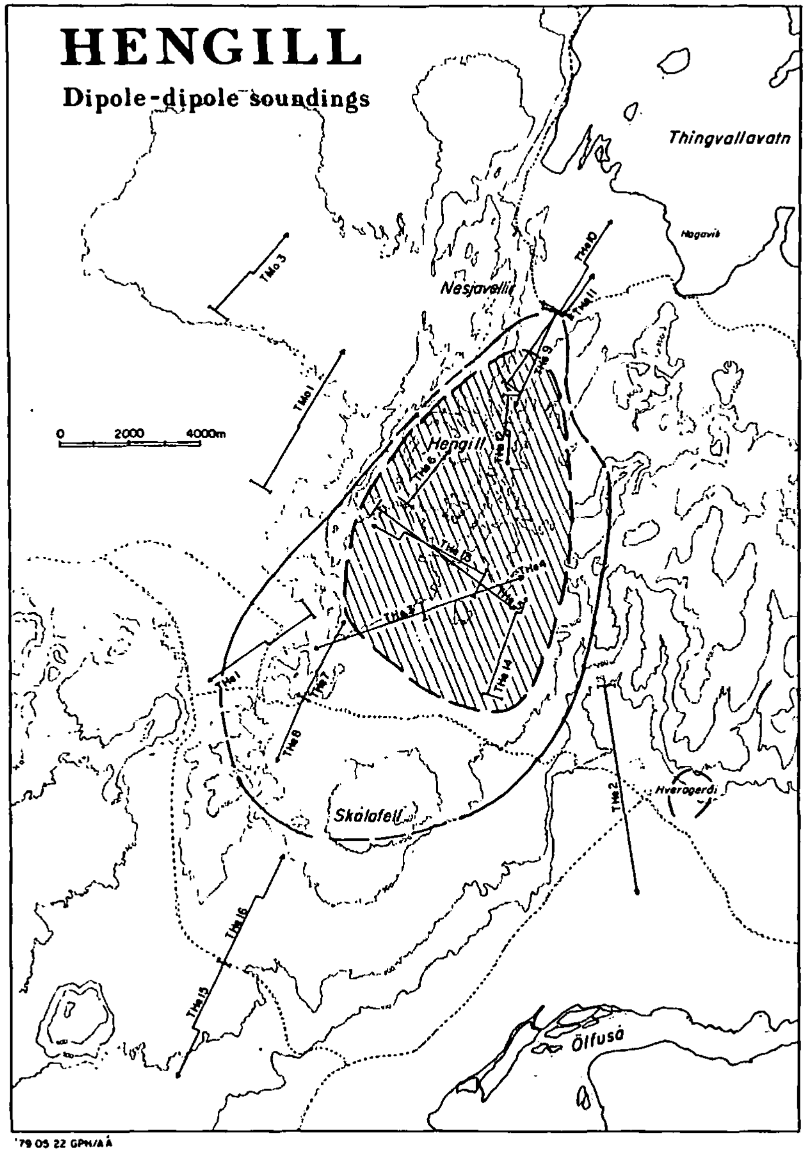

Figure 3. The outer contour-line shows a low-resistivity area (≤20 Ωm) at 400 m depth b.s.l. Within the inner shaded area the resistivity increases with depth, below 500-700 m b.s.l. The arrows show locations of dipole-dipole-equatorial soundings.

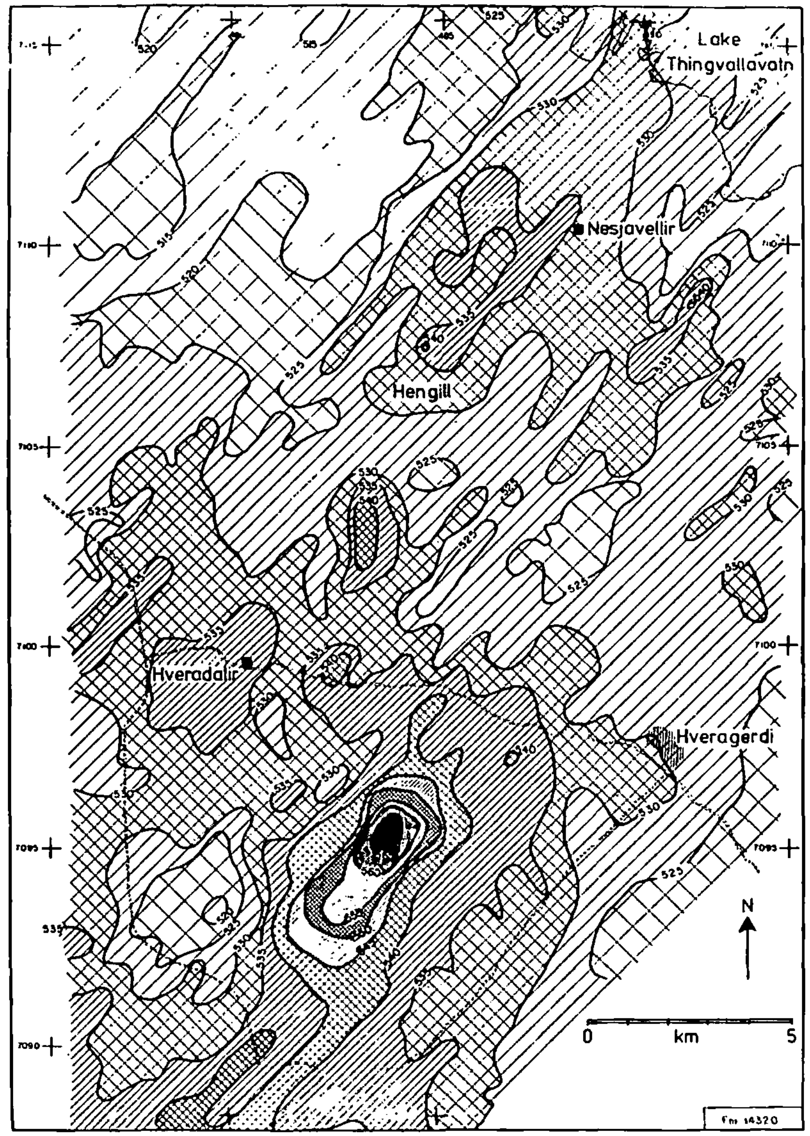

Figure 4. Aeromagnetic map of the Hengill area flown at an altitude of 800 m. Contour lines are drawn with 500 gamma intervals. A negative magnetic anomaly dissects Hengill from NW to SE. Flown by Th. Sigurgeirsson.

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COMBINED GENERATION OF HEAT AND ELECTRICITY FROM A GEOTHERMAL BRINE AT SVARTSENGI IN S.W. ICELAND

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ABSTRACT

At Svartsengi in Iceland, a geothermal plant has been built to generate 50 MW_t (megawatt thermal) and 2 MW_e (megawatt electricity) for a district heating system. A 75 MW_t and a 6 MW_e addition will be completed in two years. What makes the plant at Svartsengi unique is that it makes better use of thermal energy than is normal for a high-temperature geothermal installation. It also differs from other district heating systems in Iceland because of its use of heat-exchangers to utilize the 240°C brine source. The heat is extracted by flashing the brine in two stages to 60°C and using the flash steam for heat and electricity generation.

INTRODUCTION

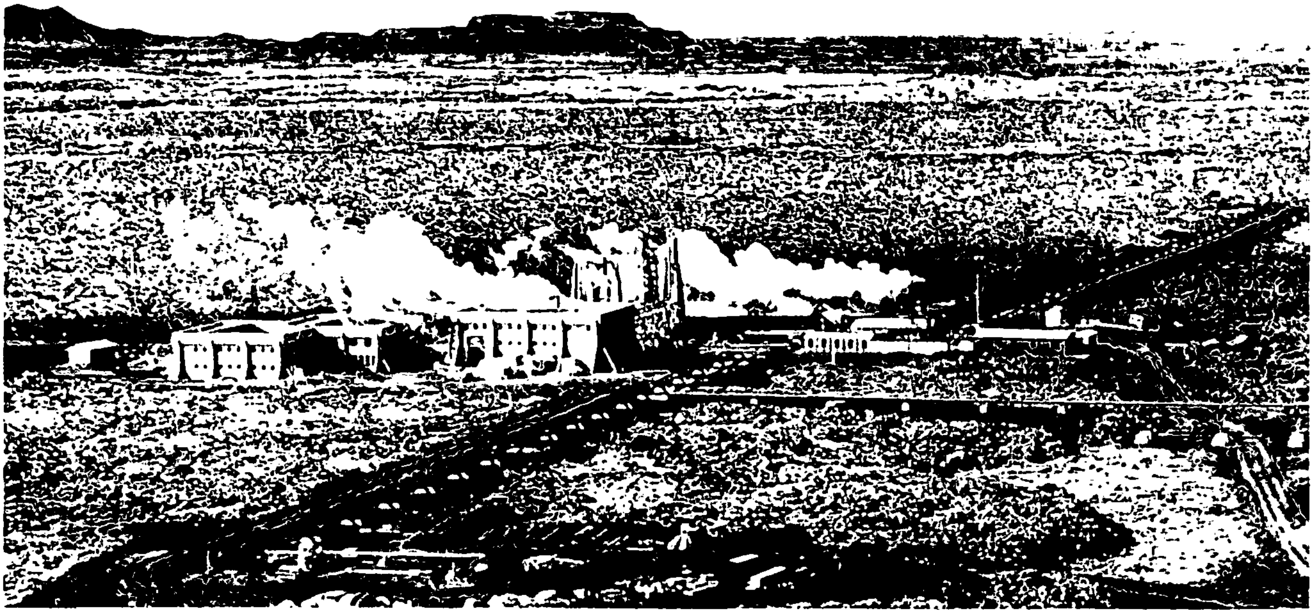
At present (1979) geothermal district heating systems in Iceland have a total nominal peak load of 600 MW_t and supply 2500 GWh of heat energy to 148,000 persons, or to 66 per cent of the population. The project described in this paper is one of many geothermal projects in Iceland that have been undertaken over the past few years for reducing the use of imported oil for space heating purposes. A joint district heating system for the six communities on the Reykjanes peninsula -Keflavik, Njardvik, Sandgerdi, Gerdar, Vogar, Grindavik- (overall pop. some 13,000) and the airport and NATO base at Keflavik (pop. some 7,000) was proposed after two successful wells had been drilled in 1971 at Svartsengi. Heat-exchangers had to be used and a pilot-plant study was carried out in 1974-1975 (ARNORSSON, et al. 1975). The municipalities of the area and the state of Iceland formed a corporation in 1975, The Sudurnes Regional Heating, and in December 1975 contracts were signed with the Icelandic consulting engineering firms Fjarhitun h.f. and Verkfraedistofa Gudmundar & Kristjans for design and supervision of works. The National Energy Authority has carried out the exploration and drilling of the geothermal field and fresh water reservoir, the pilot-plant study and has since 1975 collaborated with the engineering firms.

DISTRICT HEATING

The peak nominal load for the six communities on Reykjanes is at present 38 MW_t. The total amount of energy supplied is 154 GWh per yr for the heating of 1.7×10^6 m^3 of space. In 1990 this is expected to increase to 60 MW_t and 242 GWh per yr. In addition to this, the airport facilities at Keflavik will be heated from Svartsengi in 1981 with a nom. peak load of 55 MW_t. Because the low population density of the six coastal communities and the long distances involved, it was decided to use a single-pipe distribution system but a two pipe system for the airport. The water is pumped to Grindavik at 85°C, but at 125°C to Keflavik where it will be mixed with 54°C return water from the airport. This arrangement reduces the fresh water requirements by about 50%. The hot water supply temp. to the consumer is 80°C in the communities and 94°C at the airport. Tap water requirements are met directly from the system, and this in part dictates the maximum temperature of 80°C. The spent water is discharged to the sewer system at 30-40°C. At present the connection charge (paid only once) for a typical 400 m^3 single family house is $1250, and the monthly hot water bill comes to $54.

PROCESS DESCRIPTION

The power plant at Svartsengi is designed for
heating fresh water for a district heating system
by using geothermal steam. The fresh water is
pumped from shallow wells 4 km away from the pl
The fresh water aquifer is in porous surface la
where a fresh water lens of only 45 m is float
on sea water below (INGIMARSSON, et al. 1978)
Svartsengi geothermal site is a part of a hig
temperature area with a base temperature of
and containing a brine whose salinity is ro
2/3 that of sea water (KJARAN, et al. 1979
chemical composition of the geothermal br
shown in Table 1. Scaling of equipment i
operating problem because of the high si
tent (600 ppm). High fouling rates of h
exchanger surfaces dictated that only f
be used for the the heat-exchange proc
MUNDSSON, 1979). The pilot plant stu
strated, moreover, that the flash ste
used directly (ARNORSSON, et al. 197
fresh water by injection. Subseque
that because of high CO_2 content,
was much easier to accomplish if t
pressure) steam was condensed in
exchanger, rather than being inje
Flow diagram is shown on Fig. 1,
rate and temperature for each fl



Photograph of the geothermal powerplant at Svartsengi.

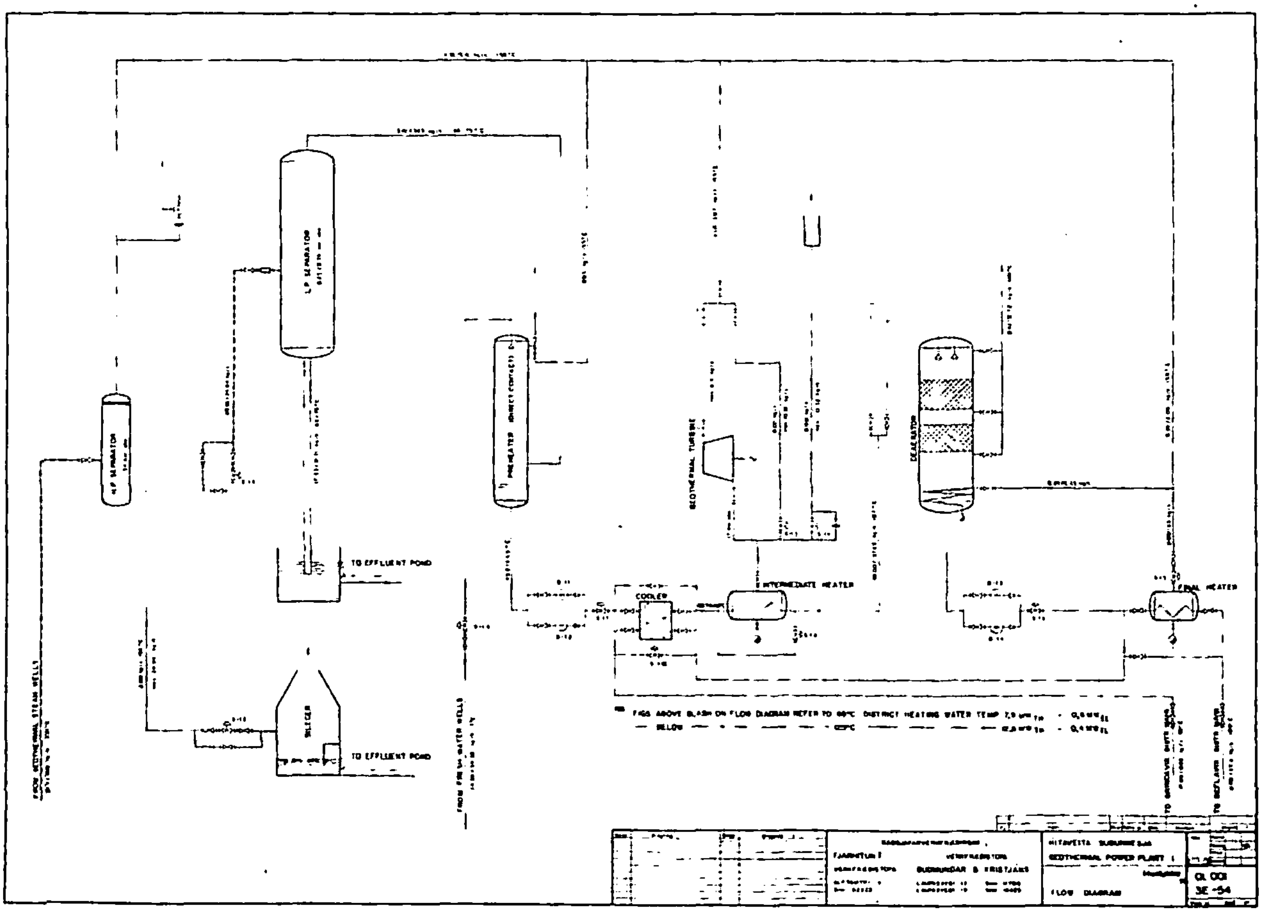


Fig. 1 Flow diagram of the geothermal powerplant at Svartsengi.

Table 1. Chemical composition of formation water at Svartsengi. Concentrations in mg/kg.

PROPERTY	WELL 3 (402 m deep)	WELL 4 (1670 m deep)
Date	19.04.78	18.04.78
Temperature (°C)	235	240
SiO_2	447	437
Na^+	6959	6837
K^+	1140	1060
Ca^{++}	1021	1036
Mg^{++}	0.74	1.08
SO_4^{--}	36.1	31.6
Cl^-	12440	12593
F^-	0.10	0.11
H_2S (total)	4.03	6.82
CO_2 (total)	183	360
TDS	22244	21400

Table 2. Composition of high-pressure and low-pressure flash steam from well H-4. Concentrations in mg/kg.

PROPERTY	H.P.	L.P.
Date	19.07.78	19.07.78
Pressure, bar, abs.	4.6	1.0
pH/°C	5.35/28	5.37/30
CO_2 (total)	2011	61.3
H_2S (total)	27.3	2.3
Carry over (brine)	50	

The decision to run the H.P. separator above the amorpous silica saturation temperature of 140°C, and to use thermal deaeration at atmospheric pressure, determines the temperatures and flow within the system. The flow is balanced to use all of the H.P. and L.P. steam generated, based on a reservoir temperature of 240°C. Each geothermal well is designed to produce 60 kg/s, an output which is split between two units as shown on Fig. 1, and the power plant consists of a total of four parallel units (4 x 12.5 MW_t). The geothermal fluid is piped in two-phase flow from the wells to a flash plant located by the power-house. Two centrifugal steam separators in series produce the H.P. (5.4 bar abs) and L.P. (0.25-0.39 bar abs) steam. The composition of the flash steam is shown in Table 2. Water level in the H.P. separator is controlled and the spent brine discharged from the barometric leg of the L.P. separator to surface disposal. The H.P. steam is used for generation of electricity in a back-pressure turbine before being condensed in a plate heat-exchanger. The L.P. steam is piped to a direct-contact condenser where it preheats the fresh water from 5°C to 65°C and removes 90% of the dissolved gases from the fresh water. This water is pumped to a second heat-exchanger (intercooler) and on to the turbine condenser mentioned above. There, the water is heated to 105°C before the atmospheric deaeration. At this point, the hot water is potable and is either cooled to 85°C in the intercooler for direct use in the town of Grindavik, or heated further by H.P. steam in a plate heat-exchanger to 125°C for pumping to Keflavik. The degree of instrumentation allows the plant at Svartsengi to be run by one operator per shift.

EQUIPMENT AND MATERIALS SELECTION

The equipment for the plant is mostly of standard manufacture, selected with the service conditions in mind. The flash plant and deaerating equipment is specially designed. Mild steel is used for the geothermal brine, steam pipes and equipment. Stainless steel and reinforced epoxy pipes are used for the heated fresh water before deaeration and for the steam condensate. The plate heat-exchangers are from titanium and stainless steel. Degradation of materials has not been noted except for corrosion of mild steel from the H.P. condensate, and cracking of the epoxy pipes. Scaling in the H.P. separator (water phase) has been 0-2 mm per year, and 1-3 mm per year in the L.P. separator. Silica scaling from stagnant water in pipe branches is a problem at present, which requires design modifications.

ELECTRICITY GENERATION

Co-generation of electricity and heat is common in district heating plants burning fossil fuels. A plant of that type might generate 0.5 MW_e per MW_t. At Svartsengi, the electricity produced is only 0.03 MW_e per MW_t (BJÖRNSSON, 1978), but if all available steam were to be used for electricity production only, the output would be appr. 7 MW_e, compared to the present output of 1.2 MW_e and 50 MW_t. The turbines selected (2 x 1 MW) are conventional double row Curtis type with few modifications made for the geothermal service (steam rate 32 kg/kWh). Operating experience has been satisfactory, except for corrosion in the labyrinth seal and of the governor valve spindle in the gland.

CHANGES IN DESIGN

The plant has been run for one and a half year, and in this period operating problems have been few. The deep wells (H-4, H-5) 1500-1700 m have to be cleaned of calcite deposits in the 9 5/8" casing at a depth of 350-400 m (ARNORSSON, 1978). However a shallow well (H-3) which is only 400 m deep, has been operated for two years without signs of plugging, because boiling takes place outside the well in the reservoir. The wells to be drilled for the future extensions will be provided with a 13 3/8" casing to allow for more area forscaling. The second power plant scheduled for completion in 1981, a different arrangement of L.P. separator, pre-heater and deaerator has been selected. Successful tests were carried out in a unit where these three pieces of equipment were combined in one column. The L.P. steam goes to a heater/deaerator which is operated under vacuum, and the water heated further in a plate heat-exchanger using back-pressure steam from the turbine and H.P. steam. At present the geothermal brine is being discharged to a pond where a silica sludge is deposited, and the water percolates into the surface lavas. Eventually, reinjection will be tested, but different alternatives are now under study. One interesting aspect of that study is the possibility of producing silica of high purity. The commercial value and methods of recovery are now being evaluated.

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EXPLORATION AND EXPLOITATION OF LOW-TEMPERATURE GEOTHERMAL FIELDS FOR DISTRICT HEATING IN AKUREYRI, NORTH ICELAND

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ABSTRACT

The Tertiary basalt formations in the vicinity of the town of Akureyri in central northern Iceland are characterized by low permeability, except for thin interlayer and narrow fissures along dykes. Although several thermal springs are found in the area, decades of geothermal exploration brought no success.
A renewed geothermal exploration was started in 1975. Geological and geophysical reconnaisance survey revealed three geothermal prospects. Drilling of the Eyjafjordur area 12 km south of Akureyri, was successful and some 150 l/s of 80-96°C hot water is now produced from 6 wells. The water is low in dissolved solids and is used direct for cooking and space heating. About 85% of the town is now supplied with 70-80°C hot water from a district heating service. The total cost savings for the people of Akureyri, by using geothermal instead of imported oil, are some $ 5 million per year. Research and drilling is continuing for further utilization of thermal water.

INTRODUCTION

Akureyri is the largest town in northern Iceland with about 12,000 inhabitants. From 1933 to 1965 several boreholes were drilled in the surrounding geothermal areas but without success. The drilling cost was high and the permeability of the aquifers too low to produce sufficient water for district heating. From 1970 there has been great progress in geothermal exploration and exploitation in Iceland. New techniques for surveying geothermal areas and analyzing the data were developed. New drilling rigs were acquired, that could drill to a depth of 3500 m, and developments in downhole pumping technology were being made, which allowed to place pumps at a depth of 300-400 m. So, at the beginning of the first oil crisis in 1973, it was possible to reconsider older ideas and plans for utilization. It was now justifiable to spend more money on search for hot water by reconnaisance survey and drilling, and it became feasible to pipe hot water much longer distances than before. This paper describes a geological, geophysical and chemical research which started in 1975, and drilling for hot water near Akureyri. Further a description is given of a district heatins system which now is supplying the town with hot water for spaceheating

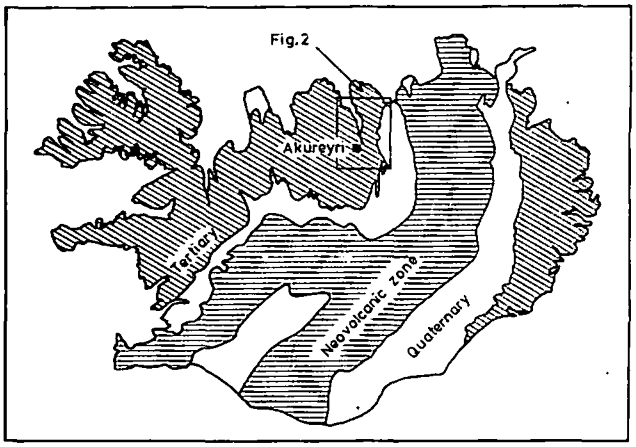

Figure 1. Geology of Iceland and location of the town Akureyri.

GEOTHERMAL FIELDS IN ICELAND

The Mid-Atlantic Ridge lies across Iceland and is characterized by a zone of recent volcanism and tectonic activity. On both sides of the Ridge there are regions of Quaternary flood basalts and subglacially erupted formations. In the east, west and north of Iceland there are Tertiary subaerial flood basalt areas (Figure 1). Within the active central zone, there are over 20 geothermal fields characterized by boiling water and fumaroles. In the Quaternary and Tertiary areas there are about 600 thermal springs in about 250 localities. They are characterized by hot water with a temperature varying from a few degrees above mean annual temperature up to 100°C. The flow rate varies from nil up to maximum of 180 l/s from a single spring. The water in the low-temperature areas originates as meteoric rain water falling in the highlands near the center of the country. It percolates deep into the bedrock and flows laterally in the upper part of the crust, driven by the hydrostatic head along faults, dikes and permeable layers (Einarsson, 1942). The water can flow tens of km before it ascends to the surface, often along faults and dikes, in valleys on the lowlands. On its way the water draws heat from the surrounding rock, which is supplied by the high regional heat flow. The temperature of the water depends on the depth and length of the main flow channels. The permeability is usually rather low in the Tertiary lava pile but is much higher in the Quaternary areas.

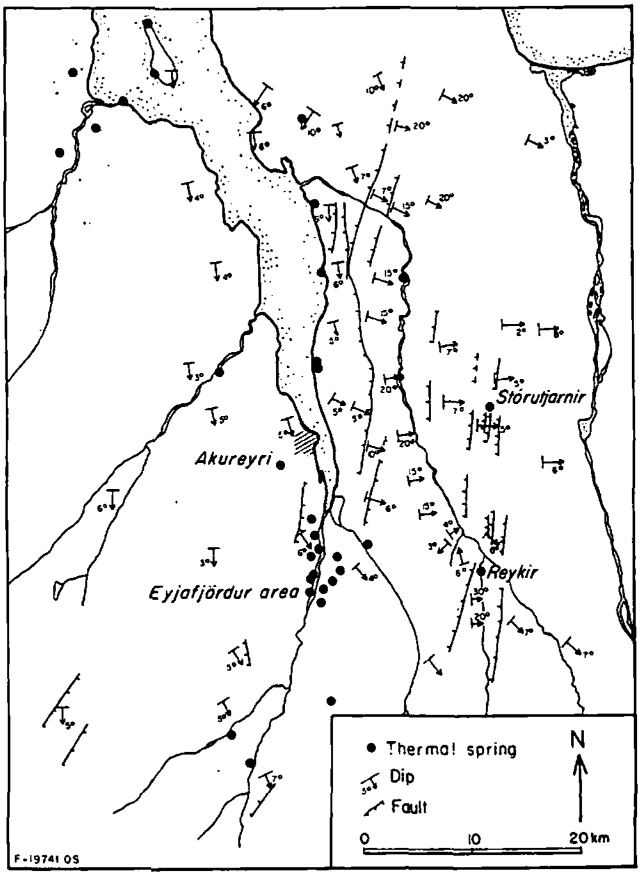

Figure 2. A generalized geology and tectonic map of the Akureyri area. Mapped by K. Saemundsson.

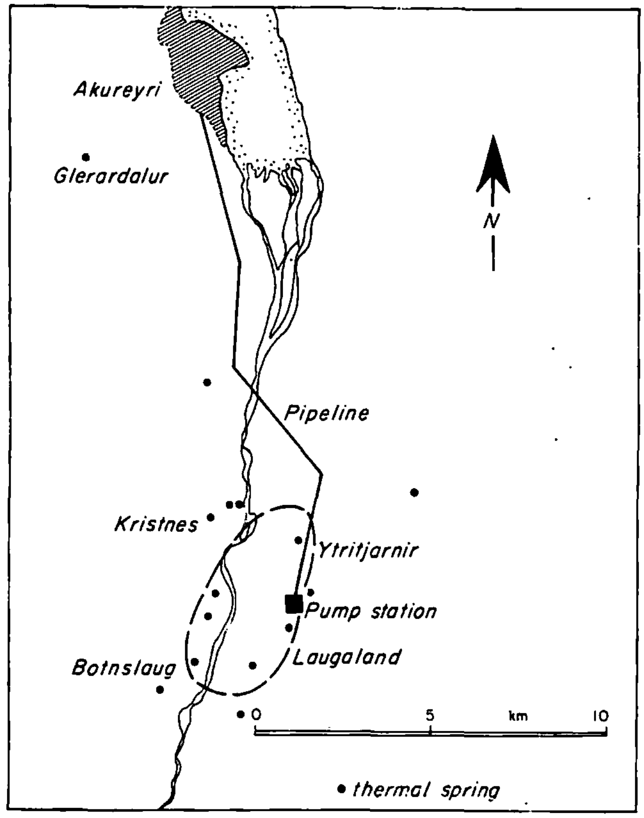

Figure 3. The Eyjafjordur geothermal area south of Akureyri is characterized by a low--resistivity anomaly (broken line). the present production fields are at Ytritjarnir, Laugaland and Botnslaug.

ECONOMIC IMPORTANCE OF GEOTHERMAL ENERGY FOR ICELAND

About one third of the net energy consumption in Iceland in 1979 came from low-temperature geothermal fields. Another third was from hydro power stations and the rest from imported oil. About 45% of all energy consumption was in space heating. In 1979 there were 24 public district heating services in operation in towns and in over 100 rural localities hot water is used for space heating, swimming pools, greenhouses etc. In 1974 about 50% of the population enjoyed geothermal district heating. In 1977 this figure had reached 60% and is about 70% at present. When the already initiated district heating services have been completed this number will have increased to about 80%. It is now economically feasible to pipe hot water over much longer distances and spend more money on geothermal research and drilling than before. The longest pipeline presently being build in Iceland is 60 km long. Much more effort has been put into reconnaisance study and search for geothermal fields, also where the temperature and permeability is low and probability for successful production of water is low. This is especially true for the Tertiary regions where productive geothermal fields are only found in few areas.

GEOLOGICAL SETTING

The Tertiary strata near Akureyri are built up of subaerial basalt flows usually a few meters thick, with thin interlayers of sediments and volcanic scoria. The age ranges from 8 to 10 m.y. Extensive secondary mineralization (zeolites) has filled pores and cracks and lowered the permeability (Saemundsson et al., 1980). The dip of the lava pile is 5-7° towards south and southeast towards the highlands and is perpendicular to the main flow of the geothermal water from south to north (Figure 2). Numerous, generally north-south trending dikes intersect the strata. The total volume of dikes is around 6%. All major faults have the same direction (Bjornsson et al., 1979). The main direction of the dikes, i.e. parallel to the hydrostatic gradient, and their close association with the thermal springs indicates that some of the dikes must be the main flow channels of the geothermal water.

GEOPHYSICAL EXPLORATION

The geology of a large part of northern Iceland was mapped, and a resistivity survey was carried out. About 70 DC-Schlumberger soundings with a current arm of 1500 m were performed, and then extended with several dipole-dipole equatorial soundings with a spacing up to 6000 m.

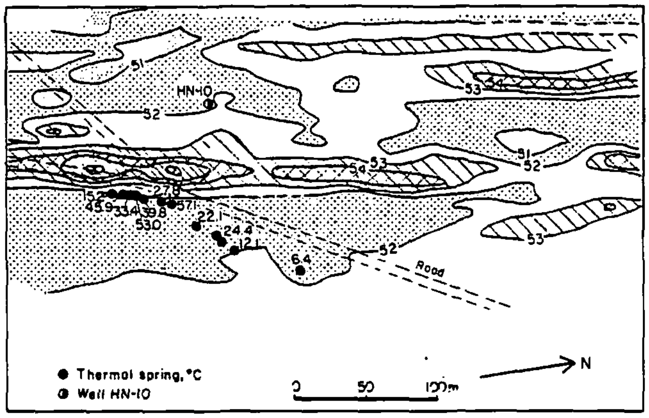

Figure 4. Ground magnetic map of the Botnslaug area. The numbers are magnetic intensity in kilogamma. The drillhole HN-10 intersects the positive dyke at 800 m depth.

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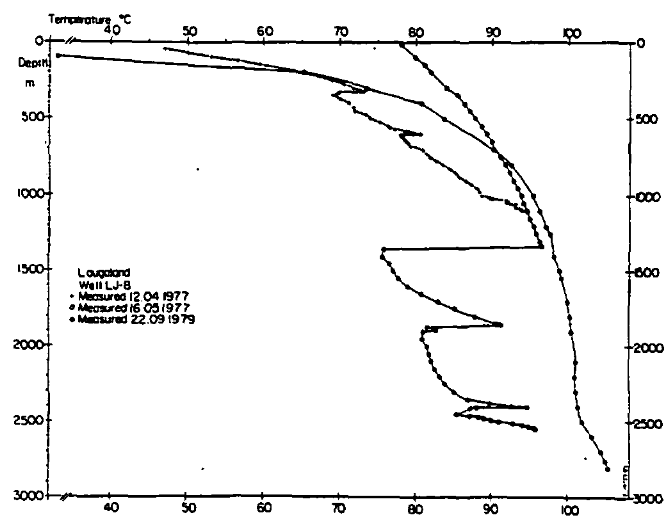

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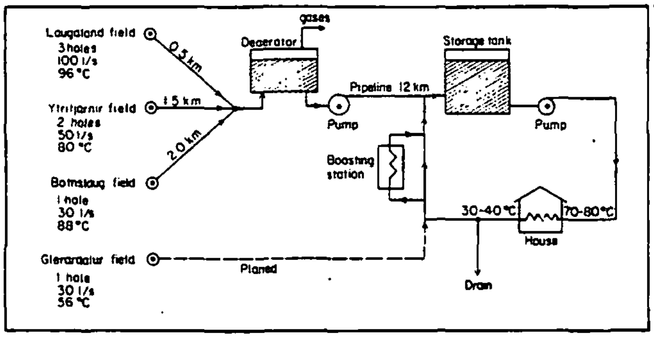

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WARD

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EXPLORATION AND EXPLOITATION OF LOW-TEMPERATURE GEOTHERMAL FIELDS FOR DISTRICT HEATING IN AKUREYRI, NORTH ICELAND

Axel Bjornsson

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ABSTRACT

The Tertiary basalt formations in the vicinity of the town of Akureyri in central northern Iceland are characterized by low permeability, except for thin interlayer and narrow fissures along dykes. Although several thermal springs are found in the area, decades of geothermal exploration brought no success.

A renewed geothermal exploration was started in 1975. Geological and geophysical reconnaisance survey revealed three geothermal prospects. Drilling of the Eyjafjordur area 12 km south of Akureyri, was successful and some 150 l/s of 80-96°C hot water is now produced from 6 wells.

The water is low in dissolved solids and is used direct for cooking and space heating. About 85% of the town is now supplied with 70-80°C hot water from a district heating service. The total cost savings for the people of Akureyri, by using geothermal instead of imported oil, are some $ 5 million per year. Research and drilling is continuing for further utilization of thermal water.

INTRODUCTION

Akureyri is the largest town in northern Iceland with about 12,000 inhabitants. From 1933 to 1965 several boreholes were drilled in the surrounding geothermal areas but without success. The drilling cost was high and the permeability of the aquifers too low to produce sufficient water for district heating. From 1970 there has been great progress in geothermal exploration and exploitation in Iceland. New techniques for surveying geothermal areas and analyzing the data were developed. New drilling rigs were acquired, that could drill to a depth of 3500 m, and developments in downhole pumping technology were being made, which allowed to place pumps at a depth of 300-400 m. So, at the beginning of the first oil crisis in 1973, it was possible to reconsider older ideas and plans for utilization. It was now justifiable to spend more money on search for hot water by reconnaisance survey and drilling, and it became feasible to pipe hot water much longer distances than before. This paper describes a geological, geophysical and chemical research which started in 1975, and drilling for hot water near Akureyri. Further a description is given of a district heatins system which now is supplying the town with hot water for spaceheating

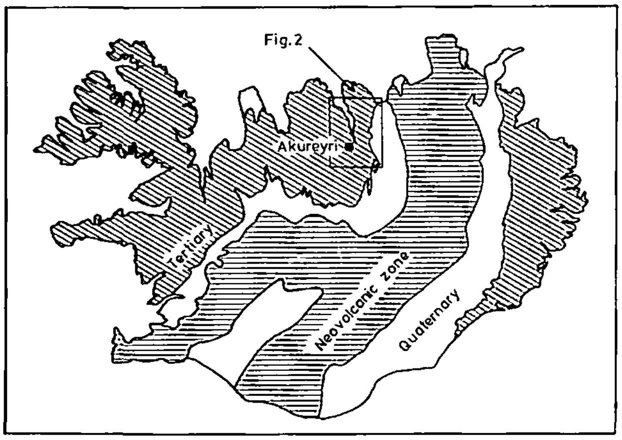


Figure 1. Geology of Iceland and location of the town Akureyri.

GEOTHERMAL FIELDS IN ICELAND

The Mid-Atlantic Ridge lies across Iceland and is characterized by a zone of recent volcanism and tectonic activity. On both sides of the Ridge there are regions of Quaternary flood basalts and subglacially erupted formations. In the east, west and north of Iceland there are Tertiary subaerial flood basalt areas (Figure 1). Within the active central zone, there are over 20 geothermal fields characterized by boiling water and fumaroles. In the Quaternary and Tertiary areas there are about 600 thermal springs in about 250 localities. They are characterized by hot water with a temperature varying from a few degrees above mean annual temperature up to 100°C. The flow rate varies from nil up to maximum of 180 l/s from a single spring. The water in the low-temperature areas originates as meteoric rain water falling in the highlands near the center of the country. It percolates deep into the bedrock and flows laterally in the upper part of the crust, driven by the hydrostatic head along faults, dikes and permeable layers (Einarsson, 1942). The water can flow tens of km before it ascends to the surface, often along faults and dikes, in valleys on the lowlands. On its way the water draws heat from the surrounding rock, which is supplied by the high regional heat flow. The temperature of the water depends on the depth and length of the main flow channels. The permeability is usually rather low in the Tertiary lava pile but is much higher in the Quaternary areas.

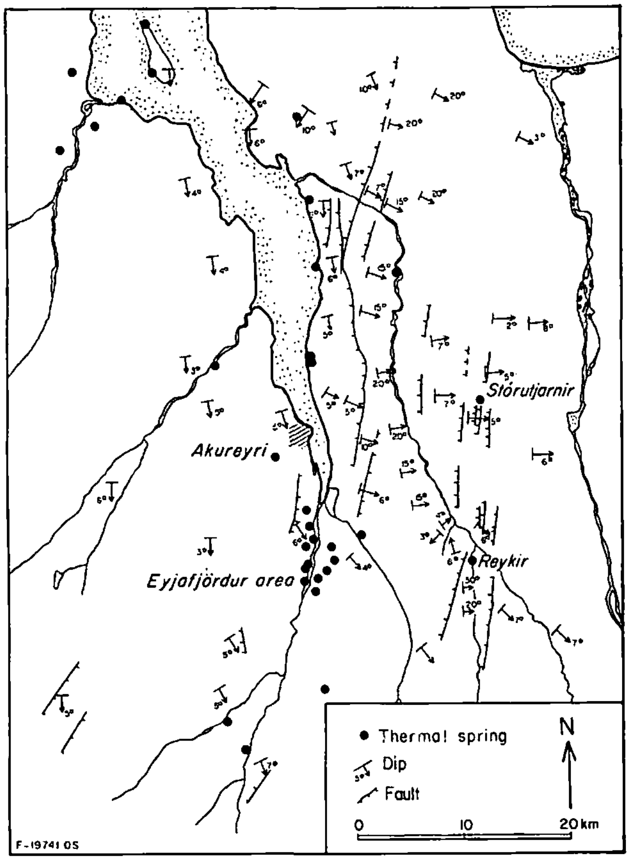


Figure 2. A generalized geology and tectonic map of the Akureyri area. Mapped by K. Saemundsson.

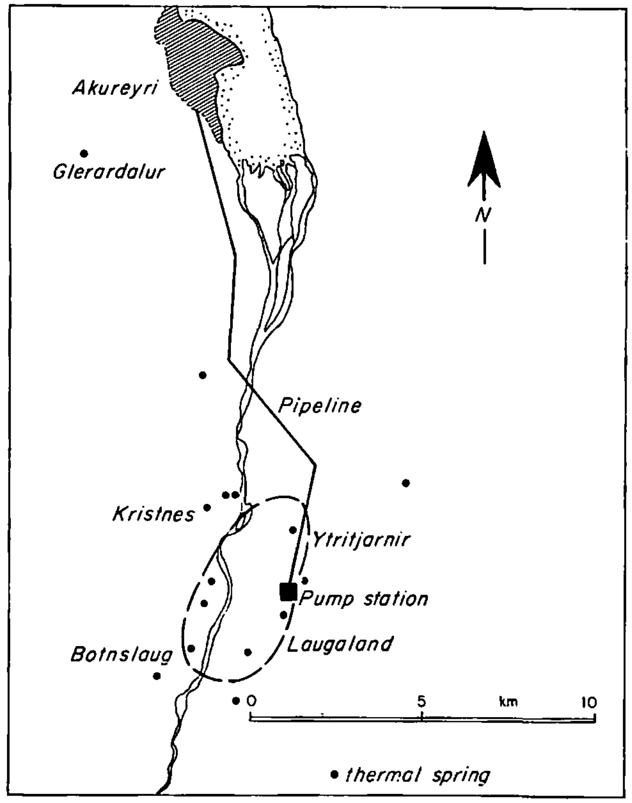


Figure 3. The Eyjafjordur geothermal area south of Akureyri is characterized by a low-resistivity anomaly (broken line). the present production fields are at Ytritjarnir, Laugaland and Botnslaug.

ECONOMIC IMPORTANCE OF GEOTHERMAL ENERGY FOR ICELAND

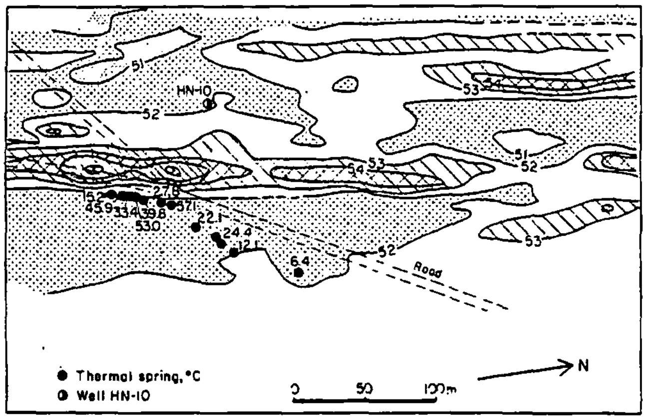
About one third of the net energy consumption in Iceland in 1979 came from low-temperature geothermal fields. Another third was from hydro power stations and the rest from imported oil. About 45% of all energy consumption was in space heating. In 1979 there were 24 public district heating services in operation in towns and in over 100 rural localities hot water is used for space heating, swimming pools, greenhouses etc. In 1974 about 50% of the population enjoyed geothermal district heating. In 1977 this figure had reached 60% and is about 70% at present. When the already initiated district heating services have been completed this number will have increased to about 80%. It is now economically feasible to pipe hot water over much longer distances and spend more money on geothermal research and drilling than before. The longest pipeline presently being build in Iceland is 60 km long. Much more effort has been put into reconnaisance study and search for geothermal fields, also where the temperature and permeability is low and probability for successful production of water is low. This is especially true for the Tertiary regions where productive geothermal fields are only found in few areas.

GEOLOGICAL SETTING

The Tertiary strata near Akureyri are built up of subaerial basalt flows usually a few meters thick, with thin interlayers of sediments and volcanic scoria. The age ranges from 8 to 10 m.y. Extensive secondary mineralization (zeolites) has filled pores and cracks and lowered the permeability (Saemundsson et al., 1980). The dip of the lava pile is 5-7° towards south and southeast towards the highlands and is perpendicular to the main flow of the geothermal water from south to north (Figure 2). Numerous, generally north-south trending dikes intersect the strata. The total volume of dikes is around 6%. All major faults have the same direction (Bjornsson et al., 1979). The main direction of the dikes, i.e. parallel to the hydrostatic gradient, and their close association with the thermal springs indicates that some of the dikes must be the main flow channels of the geothermal water.

GEOPHYSICAL EXPLORATION

The geology of a large part of northern Iceland was mapped, and a resistivity survey was carried out. About 70 DC-Schlumberger soundings with a current arm of 1500 m were performed, and then extended with several dipole-dipole equatorial soundings with a spacing up to 6000 m.



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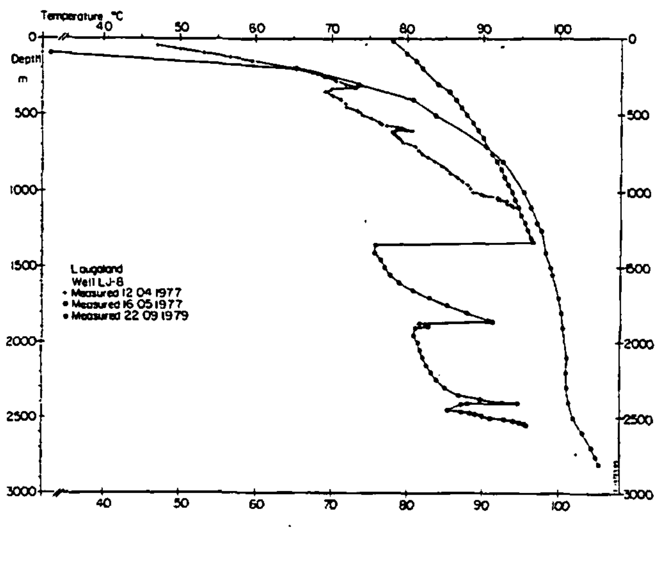

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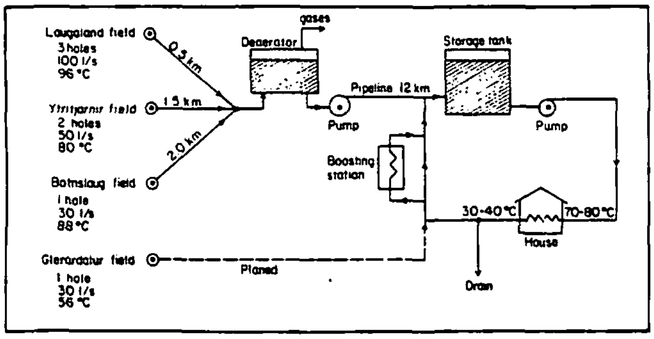

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to S.H. Ward with best wishes — Axel Björnsson

Rifting of the Plate Boundary in North Iceland 1975–1978

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A rifting episode started in 1975 on the accreting plate boundary in North Iceland after 100 years of quiescence. Horizontal extension of some 3 m has been observed in the Krafla caldera and the associated 80 km long fissure swarm. The rifting occurs periodically in short active pulses at a few months intervals. Between these active pulses, continuous inflation of 7–10 mm/day of the caldera is caused by 5 m^3/s inflow of magma into a magma chamber at 3 km depth. The active pulses are caused by a sudden east-west expansion of the fissure swarm and a contraction of zones outside the fissure swarm. Rapid flow of magma out of the magma chamber and into the fissures toward north or south is indicated. These pulses are accompanied by earthquake swarms and vertical ground movements of up to 2 m and sometimes also volcanic eruptions and formation of new fumaroles. The magma chamber below the Krafla caldera thus acts as a trigger for the plate movement along the constructive plate boundary in North Iceland.

INTRODUCTION

The plate boundary between the European and the American plates follows the Mid-Atlantic Ridge and crosses Iceland from south-west to north-east. In Iceland, the boundary is characterized by zones of recent volcanism, graben structures, and seismic activity, and is generally named the Neovolcanic Zone. The Neovolcanic Zone in north Iceland has a north-south direction and is characterized by several fault and fissure swarms, each passing through a central volcano (Figure 1). The tectonic and volcanic activity of the Neovolcanic Zone is restricted to these central volcanoes and the associated fissure swarms, and occurs episodically rather than continuously, with a period of 100–150 years. During each active period, which probably lasts 5–20 years, only one central volcano and fissure swarm is active [*Björnsson et al.*, 1977].

The Krafla fissure swarm, which is presently active, extends from the Tjörnes Fracture Zone in the Axarfjördur bay in the north and some 100 km to the south. Its width is approximately 5 km, but varies considerably along the swarm. It passes through the Krafla caldera, which formed during the last interglacial period, but has since been filled to the rim with eruptive material (Figure 2). In postglacial time, about 35 eruptions have taken place within this fissure swarm, most of them either within the Krafla caldera or in the Namafjall area, about 10 km south of Krafla [*Björnsson et al.*, 1977].

A geothermal field with temperatures exceeding 340°C at 2 km depth exists within the Krafla caldera, and another geothermal field located in the Namafjall area has a temperature exceeding 290°C at 1.8 km depth. The economic importance of these geothermal fields is responsible for a more intense research of this area than would have been possible otherwise.

There are historical records of only one period of volcanic and tectonic activity within the Krafla fissure swarm, the Myvatn Fires of 1724–1729. A very small eruption in 1746 may be regarded as the last observed pulse in the Myvatn Fires episode. Another period of volcanic and tectonic activity in North Iceland occurred in 1874–1875, but in another fissure swarm, the Askja fissure swarm. The Theistareykir fissure swarm was active in 1618, but no volcanic activity was reported [*Thoroddsen*, 1925]. After the volcanic and tectonic activity in the Askja fissure swarm in 1874–1875, there has been little tectonic activity within the neovolcanic zone in North Iceland until the Krafla fissure swarm became active in 1975. A period of volcanic activity in Askja 1921–1926 was not associated with any observed tectonic activity, and neither was the Askja eruption of 1961, except for some vertical ground movements within the Askja caldera.

Geodetic measurements intended to observe tectonic movements in the neovolcanic zone in North Iceland were initiated in 1938 [*Niemczyk*, 1943]. Remeasurement of the 1938 network in 1965 showed no significant ground deformation. More precise remeasurements in 1971 and 1975 showed significant expansion of the area around Krafla during the 1971–1975 interval after an indicated contraction of the same area during the 1965–1971 time interval [*Gerke*, 1969, 1974, 1977; *Schleusener and Torge*, 1971; *Spickernagel*, 1966; *Torge and Drewes*, 1977]. These observations may indicate that the present tectonic event had started before the summer of 1975 as an inflation of the Krafla area. Increased seismic activity of the Krafla region in early 1975 may also be interpreted as an indication of abnormal tectonic activity.

NARRATIVE OF EVENTS

Contemporary description of the Myvatn Fires episode in 1724–1729 shows clearly that the volcanic and seismic activity was largely confined to short periods of high activity interrupted by much longer periods of quiescence. Each active period lasted for only a few days, while the quiet periods lasted for several months. Each of the pulses of activity was characterized by strong earthquakes and either volcanic activity or noticeable changes in the geothermal activity. Some of the pulses were associated with changes in the level of Lake Myvatn, indicating large scale vertical ground movement [*Thoroddsen*, 1925].

The present volcano-tectonic episode in the Myvatn-Krafla area is also characterized by a similar pulsation. It is possible to divide the time since the active episode started in December 1975 into periods of two kinds, inflation periods and subsidence events. Figure 3 shows the elevation changes with time of bench mark FM-5596 near the center of the Krafla caldera from 1975 to early 1978. Since the initial subsidence event, in December 1975, the rate of uplift has been relatively constant,

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around 7 mm/day at this point, but it has been interrupted by seven sudden subsidence events lasting 2 hours to several days each. The inflation periods last for 1–7 months and are characterized by the following:

1. Continuous and nearly constant uplift of the Krafla region. The maximum uplift is near the center of the caldera, 7–10 mm/day, decreasing outward to less than 1 mm/day at a distance of 10 km from the apex of uplift.

2. Gradually increasing seismic activity within the caldera after the land elevation has reached a certain critical level. Decreasing or no seismic activity within the fissure swarm outside the caldera (P. Einarsson, personal communication, 1978).

3. Gradual widening of fissures near the center of uplift, up to 1 mm/day.

The duration of the subsidence events or active pulses is much shorter than the inflation periods. Some are so small that they are not noticed, except on measuring equipment, while other pulses correspond exactly to those described in the eighteenth century episode. These pulses of activity have the following common characteristics according to the available observations:

1. Subsidence of the Krafla region. The maximum subsidence, near the center of the caldera, has been from 3 to about 250 cm, but decreasing outward.

2. Continuous seismic tremor (volcanic tremor) which usually starts at the same time as the subsidence and lasts for a few hours.

3. Earthquake swarm in the fissure zone outside the Krafla caldera.

4. New fissures and east-west widening of the fissure swarm at the same place as the earthquake swarm. Widening of 2 m has been measured during a subsidence event.

5. Subsidence of the active part of the fissure swarm, sometimes exceeding 1 m, and uplift of both flanks of the swarm amounting to tens of centimeters.

6. Development of new geothermal areas or increased activity in old ones. Increased pressure in drillholes.

7. Outpouring of basaltic lava, mostly within the caldera, has been observed in three of the active pulses.

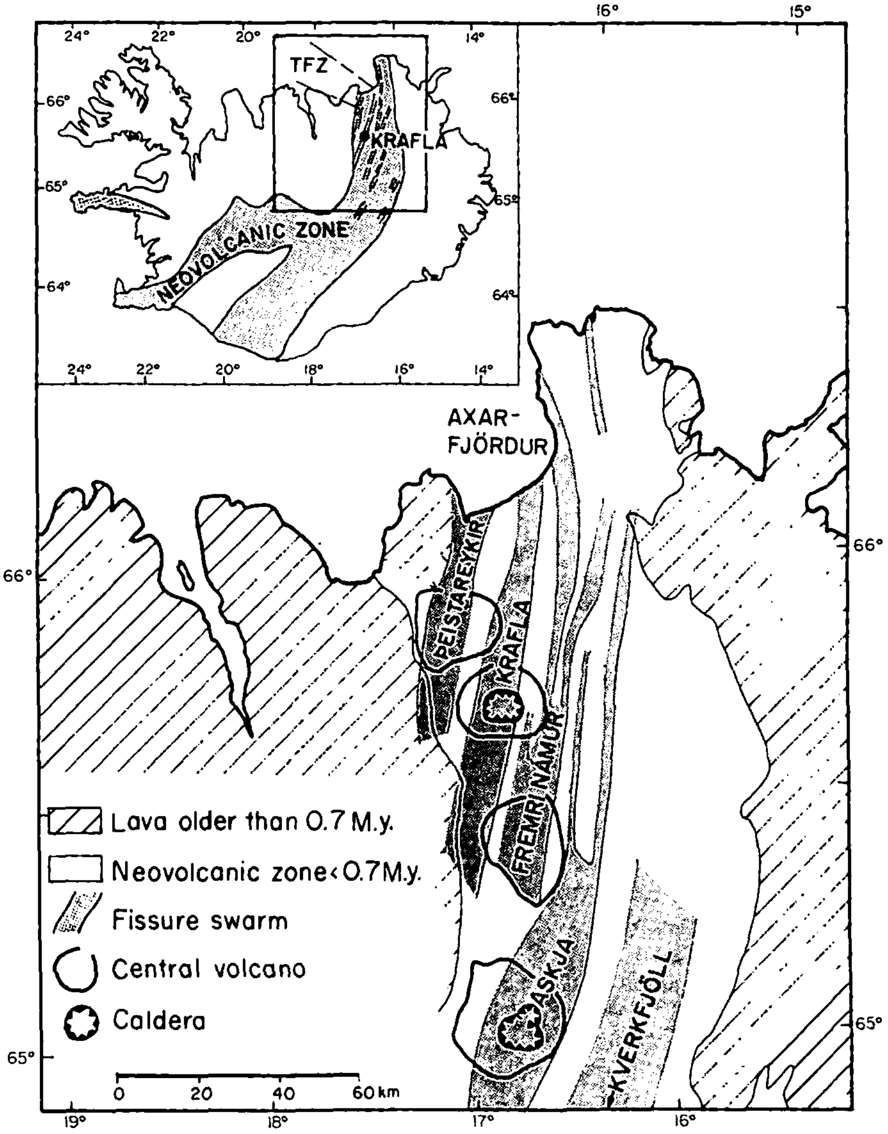

Fig. 1. The spreading zone in North Iceland. Central volcanoes and associated fissure swarms are named after the high-temperature geothermal fields in the central volcanoes. Two of them, Askja and Krafla, contain calderas. Mapped by Kristjan Saemundsson. The Tjörnes Fracture Zone (TFZ) is shown in the inset.

Figure 4 shows the areas of maximum ground deformation and rifting within the fissure swarm outside the caldera during different subsidence events.

The first pulse of high activity started on December 20, 1975, and lasted for several weeks. The maximum subsidence, near the center of the caldera, was some 2.5 m and an intense earthquake swarm was observed some 40–60 km north of Krafla where large scale ground movements occurred [*Björnsson, 1976*; *Björnsson et al.*, 1977; *Tryggvason*, 1976].

The second period of activity started about September 29, 1976. It lasted for some 5 days and the maximum subsidence was about 15 cm. The most noticeable feature of this pulse was the complete cessation of the seismic activity within the caldera, but this activity had been increasing gradually during the four preceeding months [*Björnsson et al.*, 1977].

The third active period started on October 31, 1976, at about 2 a.m. and lasted for less than 48 hours. The maximum subsidence was about 50 cm and intense volcanic tremor ac-

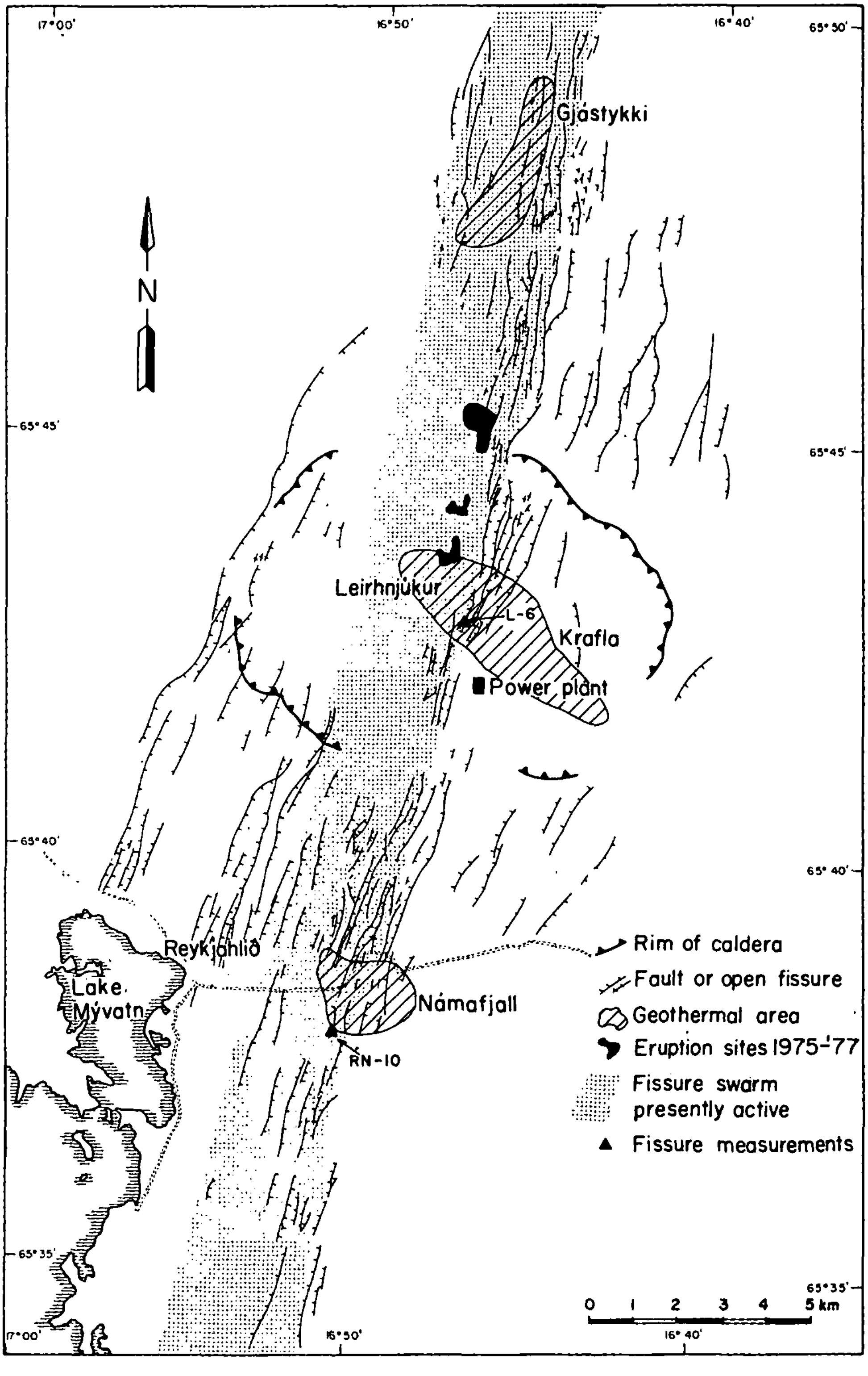

Fig. 2. Outline geological map of the Krafla caldera and the associated fissure swarm. Mapped by Kristjan Saemundsson.

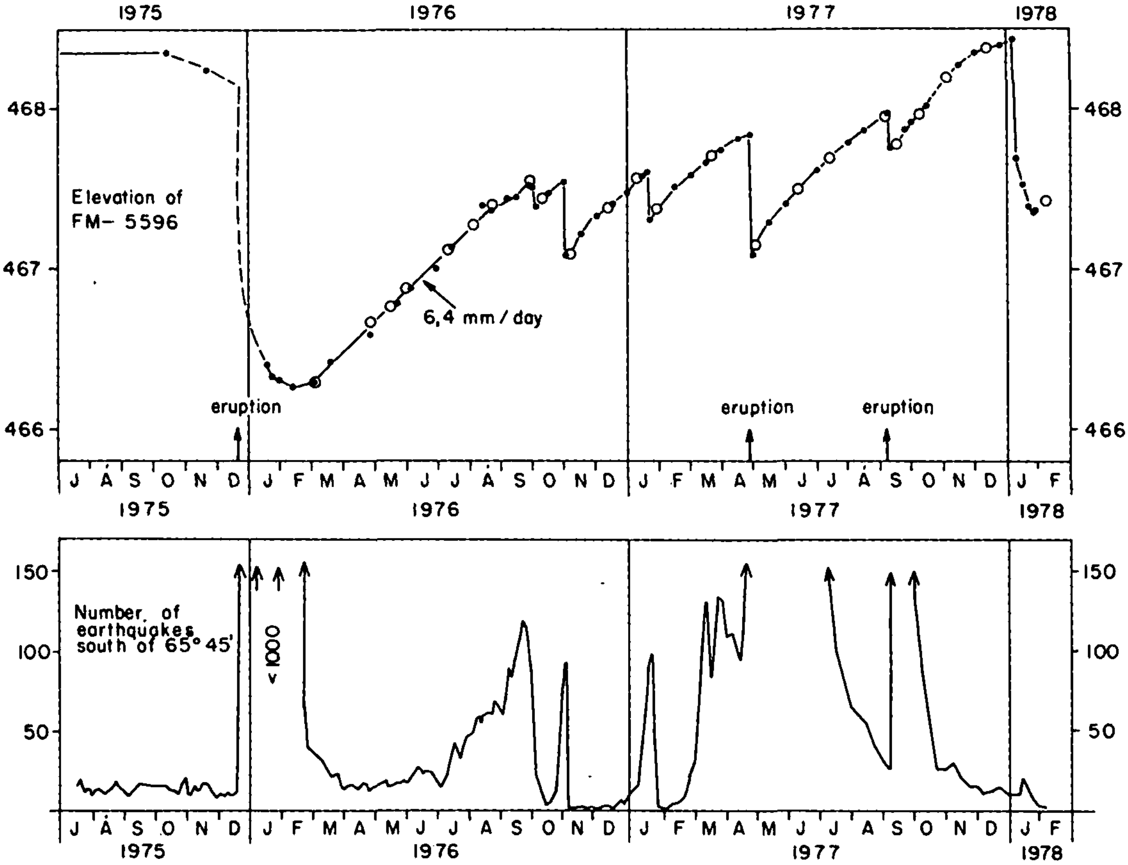

Fig. 3. Changes in elevation of bench mark FM-5596 within the Krafla caldera from 1975 to early 1978 (see Figure 5 for location). Levelling data (open circles) are supplemented with tilt data at the Krafla power house (dots). The rate of uplift is relatively constant, 5-7 mm/day, interrupted by eight sudden subsidence events. The subsidence event of November 2, 1977, was too small to be shown on this graph. The lower part shows running 5-day averages of number of earthquakes within and south of the Krafla caldera. After April 27, 1977, most of the earthquakes occurred south of the caldera. The earthquake information was kindly supplied by P. Einarsson.

companied the subsidence. Widening of old fissures and new fumaroles were observed in the fissure swarm 10-15 km north of the caldera.

The fourth active period started on January 20, 1977, shortly after midnight and lasted for only about 20 hours. The maximum subsidence was about 30 cm and widening of fissures, and new fumaroles were observed about 10 km north of the caldera.

The fifth subsidence event started on April 27, 1977, at about 1 p.m. with very intense volcanic tremor. A small lava, covering only 0.01 km², was erupted about 5 p.m. near the north rim of the caldera. The floor of the caldera subsided irregularly but the maximum subsidence was more than one meter. Widening of fissures to the south of the caldera was observed. An east-west widening of the fissure swarm of 2 m was obtained the next day by measuring the opening of individual cracks in frozen ground on a profile across the Namafjall area. A remeasurement of a geodimeter line along the same profile showed a widening of 2.0 m in excellent agreement with the direct measurements of fissures. The widening of fissures spread out from the caldera and reached the Namafjall area in about 5 hours. This indicates a velocity of the order of 0.5 m/s for the horizontal movement of activity. During this active pulse the central part of the active fissure swarm subsided about one meter but the flanks to the east and west were uplifted some tens of centimeters.

The sixth subsidence event started with volcanic tremor around 4 p.m. on September 8, 1977. The course of events was very similar as on April 27. A volcanic eruption started at about 6 p.m. near the north rim of the caldera. The area covered with lava was about 0.8 km² and the volume of lava is estimated as $2 \cdot 10^6$ m³. Another volcanic eruption occurred about 11:45 p.m. in the Namafjall area where some 2500 kg of basaltic pumice erupted through a borehole, 1138 m deep. This location is about 12 km south of the lava eruption, in the fissure swarm where the east-west widening was 1 m.

The seventh subsidence event occurred on November 2, 1977, and was the least noticeable of the observed active periods to date. It lasted for only 2 hours and the total subsidence was 2-3 cm. A small tremor was seen on the local seismometers, but no movement of fissures was observed.

The eighth subsidence event started in the afternoon of January 6, 1978, and lasted for some 3 weeks. The maximum subsidence within the caldera during this event exceeded 1 m, and the accompanying earthquake swarm was strongly felt 20-50 km north of Krafla, where significant subsidence and widening of fissures was observed.

OBSERVATIONS, TECHNIQUES, AND DATA

The credibility of the observed ground deformation and the interpretations based thereon are entirely dependent on the nature and extent of the measurements. Therefore it is certainly in order to give some details of the observational techniques and procedures along with the presentation of data. Only those observations which are pertinent in analyzing the tectonic processes are described.

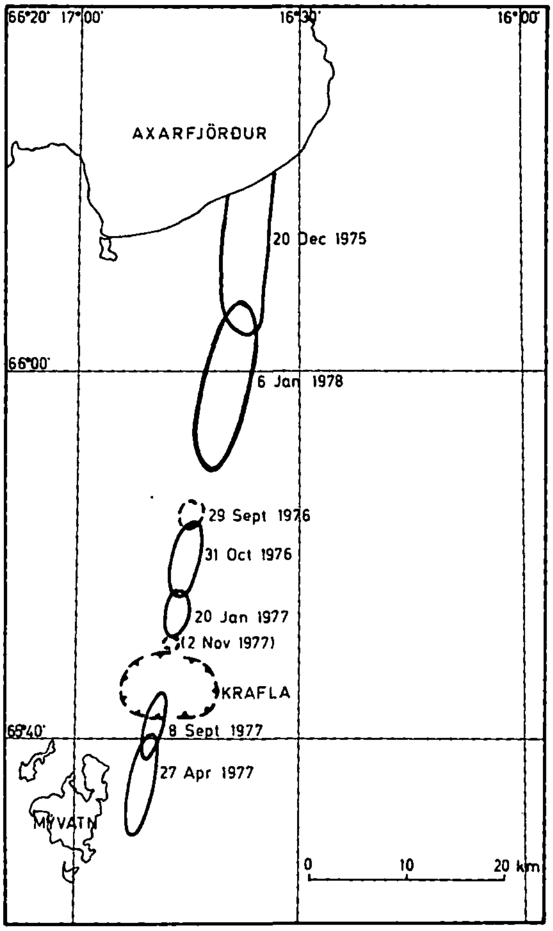

Fig. 4. Areas of maximum ground deformation and rifting outside the Krafla caldera during different subsidence events. No rifting was observed during the events of September 29, 1976, and November 2, 1977.

Levelling

Levelling was carried out in 1974 and earlier along the road from Myvatn to Krafla and around the geothermal field at Krafla. This levelling network was extended in 1976 and includes presently about 70 bench marks. Levelling has been performed at intervals of 1–2 months since March 1976 in a large part of this network and at longer intervals in the remainder. Zeiss Ni2 level and wooden measuring rods, compared with Wild invar rods, are used. The standard error of the levelling has been determined as approximately $1.5L^{1/2}$ mm, where L is the length of the foreward and backward measured levelling line in kilometers.

The area which has been rising and subsiding in the Krafla region has remained nearly constant during the last two years. The inflation bulge and the deflation bowl is nearly circular, although east-west elongation is indicated. Maximum vertical ground movement is observed near the center of the caldera. The half-width of the bowl or bulge is about 3 km and only very minor movement is measured at 10 km distance from the center. Some details of the inflation-deflation pattern are shown in Figure 5.

The elevation changes within and around the Krafla caldera have been cyclic up and down movements (Figure 3), while elevation changes on the fissure swarm to the south and north of the caldera are more or less transient phenomena. During subsidence events rapid and permanent changes in land elevation are observed, accompanied by east-west widening and earthquake swarms. Figure 6 shows land elevation changes on an east-west profile across the fissure swarm near Namafjall during the subsidence event of April 27, 1977. A 1 km wide segment in the middle of the fissure swarm subsided about 80 cm, while the flanks to the east and west of the active zone were uplifted about 30 cm. Similar elevation changes have occurred in other subsidence events where these movements are indicated by vertical displacements of faults, changes of ground water level, and changes in shore-lines of lakes.

Measured gravity changes support these observations, as can be seen in Figure 6. Similar gravity changes were observed in Gjastykki, north of the caldera, during the subsidence event of January 20, 1977.

Tilt Measurements

A water tube tiltmeter was installed on a semi-permanent basis in the Krafla power house on August 19, 1976. Three measuring pots are connected. The north-south arm of this tiltmeter is 68.95 m long and the east-west arm 19.50 m. The reading accuracy is better than 1 microradian, but temperature variations and corresponding thermal expansions of the building cause tilt error of the same magnitude. These can be partly corrected for. Readings are usually made once a day, but during periods of rapid ground movements, more frequent readings are made. During the period January–August 1976, optical levelling was frequently performed between markers on the four corners of the power house with an accuracy of better than 1 mm.

The power house is located about 1.3 km from the apex of the inflation-deflation bowl, and the north-south arm of the tiltmeter is directed almost exactly toward the apex. Hence it is ideally located and oriented to monitor tilt changes caused by elevation changes in the Krafla caldera. An excellent correlation has been found between these two parameters, tilt of the power house and ground elevation inside the Krafla caldera. The tilt can thus be used to monitor daily variations in elevation and volume changes during inflation and deflation events.

Figure 3 shows the land elevation as derived from levelling and tilt observations in the power house. Uplift or subsidence of the apex of the inflation-deflation area is approximately 3.4 mm for each microradian of tilt at the power house.

Additional dry-tilt stations have been established in the Myvatn-Krafla area, consisting of 5–6 bench marks each. Most of these tilt stations are so constructed that the bench marks lie on a circle of 25 m radius and, during measurements, an optical level, Wild N3, is placed exactly in the center of this circle and invar rods are placed on the markers. The relative elevation of the bench marks is established with an accuracy of approximately 0.1 mm, allowing to determine ground tilt of less than 5 microradians. The principal source of error of these tilt measurements is, however, internal deformation of the area covered by each tilt station. A total of 12 such tilt stations are under observation about once every month in the summer, but these measurements are rarely made in the winter due to the snow cover. The arrows in Figure 7 show observed tilt variations during the inflation period April 29 to September 8, 1977 [*Tryggvason,* 1978]. This correlates well with the levelling measurements. Continuously recording electronic tiltmeters were installed at two locations in late 1977.

Gravity Survey

A gravity survey was carried out in the Krafla area in August 1975 prior to the first Leirhnjúkur eruption. The purpose was to monitor gravity variations that might be caused by

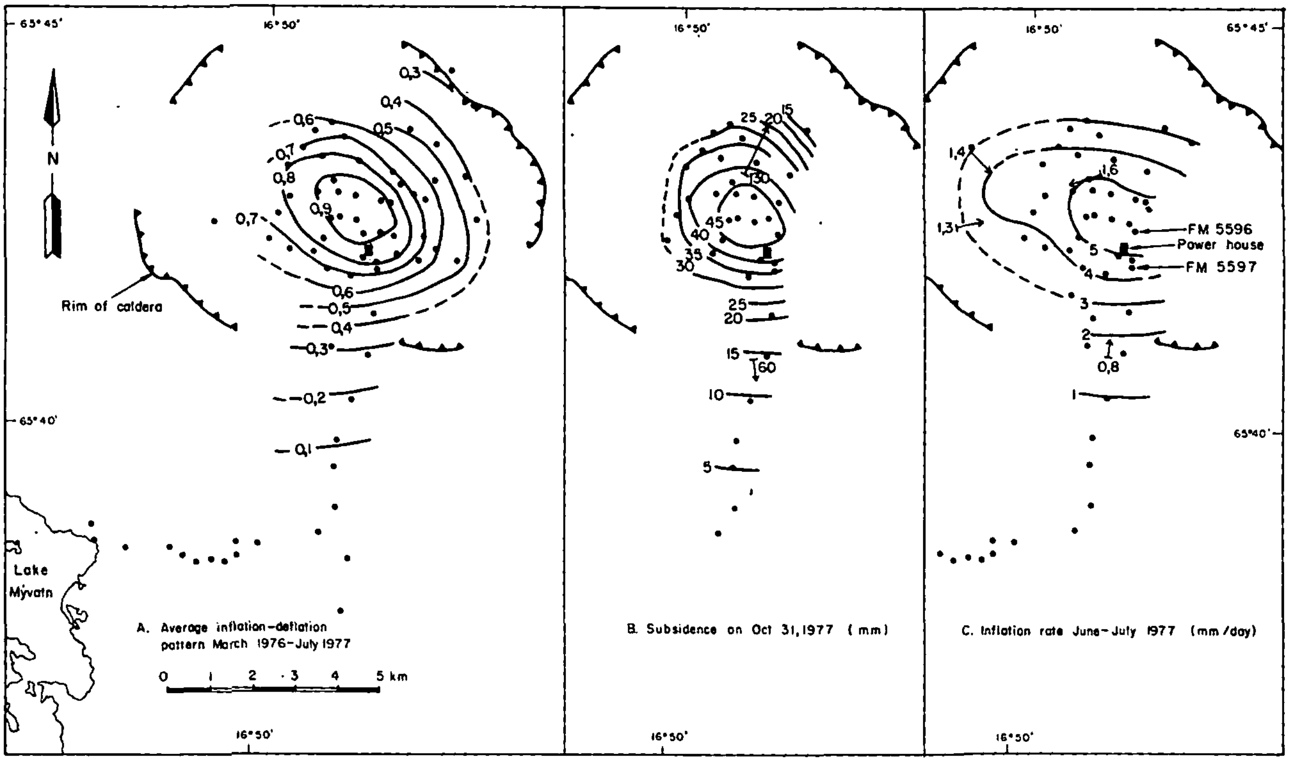

Fig. 5. Pattern of uplift and subsidence in the Krafla area. (*a*) Average vertical ground movement throughout the period March 1976 to July 1977 in fractions of maximum movements. (*b*) Total subsidence in centimeters during the subsidence event of October 31 to November 1, 1976. (*c*) Rate of uplifting in millimeters per day during a period of 1 month, June–July 1977. Arrows show tilt changes in microradians at four stations.

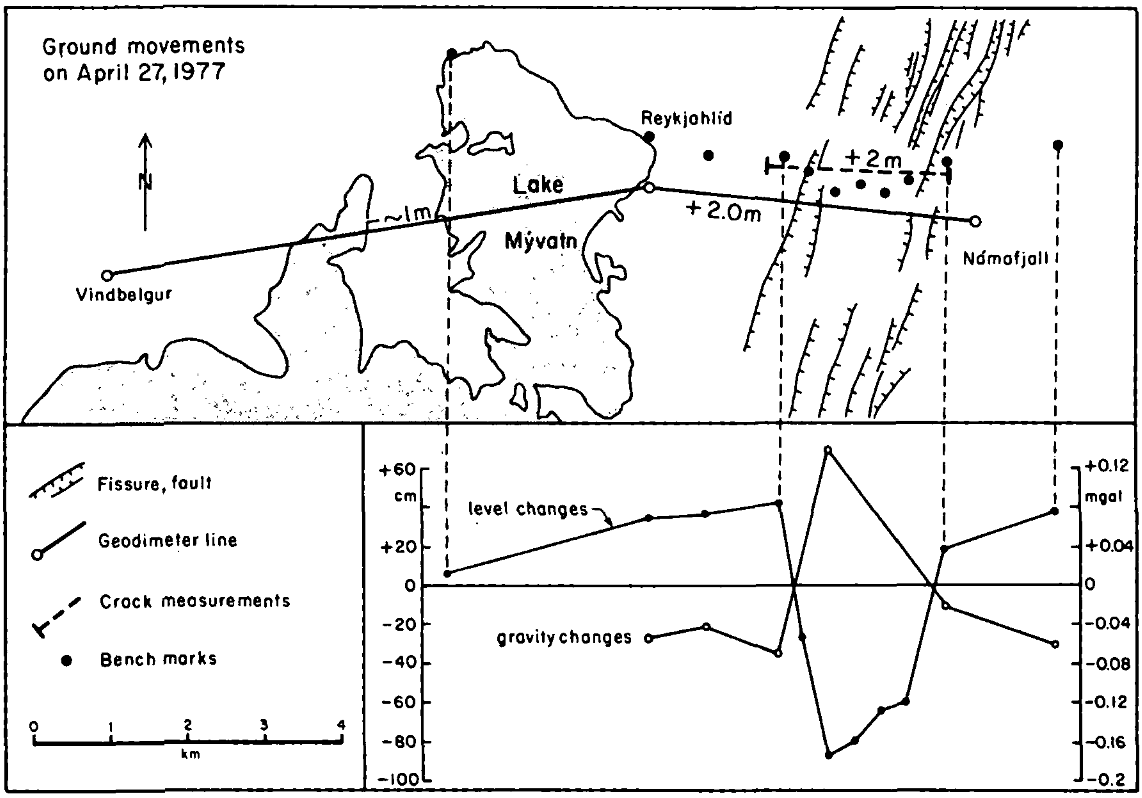

Fig. 6. Horizontal and vertical ground movements across the Krafla fissure swarm near Namafjall during the subsidence event of Arpil 27, 1977. Expansion is measured on individual cracks and by geodimeter. Elevation and gravity changes are shown in the lower part of the figure.

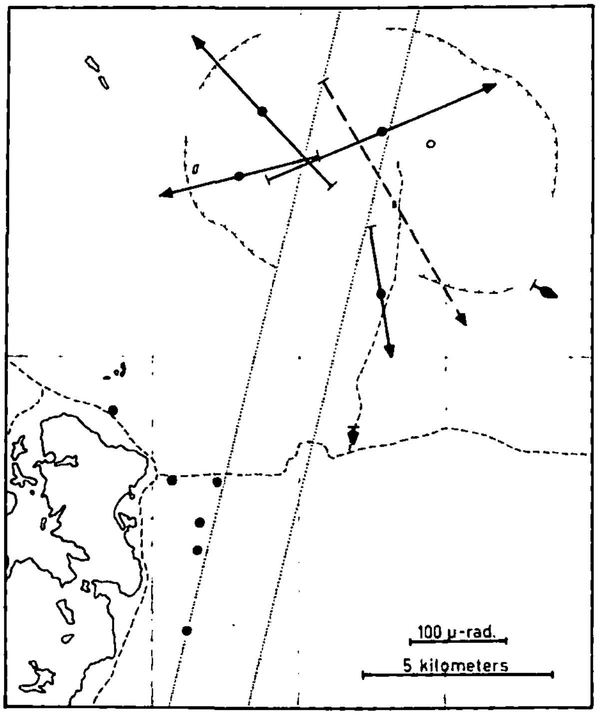

Fig. 7. Observed ground tilt during the inflation period April 29 to September 8, 1977. Arrows show direction and magnitude of observed tilt, and if no arrows are drawn, the tilt was less than the error of the observations. Solid dots are dry-tilt stations. The direction of tilt at the power house (dashed arrow) is based on an electronic tiltmeter which was operated for a few weeks during this period. Dotted lines outline the most active part of the Krafla fissure swarm. For location see Figures 1 and 2.

groundwater changes due to removal of fluid from drillholes in the goethermal fields. The network included some 30 stations and the instrument used was a LaCoste-Romberg gravity meter, G-10. This network was reoccupied in March 1976 and June 1976 using an old Worden gravity meter, W-68. This instrument, however, turned out to be rather unreliable, and one could only conclude that a significant gravity change had occurred (a few tenths of mgal) in the Krafla caldera since August 1975. In September 1976 a survey was carried out on an extended net using a new LaCoste-Romberg instrument, G-445, and the survey has since then been repeated at intervals of 1–2 months with the same instrument. The network has gradually been extended to more than 100 stations, although not all are included in each survey. Gravity changes were originally referred to a base station at Reykjahlid by Lake Myvatn, but since April 1977 corrections have been made for changes in that station relative to another base station at Husavik. The accuracy of the values measured by the G-meters is 0.01 to 0.02 mgals after corrections have been made for tidal effects and instrumental drift.

The main purpose of the gravity survey since 1976 has been to monitor elevation changes over a more extensive area than is covered with the level surveys. However, the conversion of gravity changes into elevation changes requires knowledge of changes in mass distribution. As this knowledge is not present, two models have been considered; (1) a free air model, leading to approximately 3 cm elevation change for 0.01 mgal gravity change, and (2) a Bouguer model assuming a density of 2.5 g/cm^3, leading to approximately 5 cm elevation change for 0.01 mgal gravity change.

Repeated gravity measurements have been made at several stations that are regularly included in the levelling survey. Figure 8 shows the results from one of these stations, FM-5597, located near the Krafla power house (for location see Figure 5). A comparison is made between measured gravity changes and gravity changes that correspond to measured level changes using the two models. The comparison indicates that the Bouguer model leads to a fairly close agreement with the level changes, although notice should be taken of the additional assumption that no ground movement took place between the levelling survey in 1974 and the gravity survey in 1975. Results from other stations support this conclusion. Since the Bouguer model corresponds to an assumption of no change of density, a subsidiary result of the gravity survey has been to lend support to the theory that the inflation and

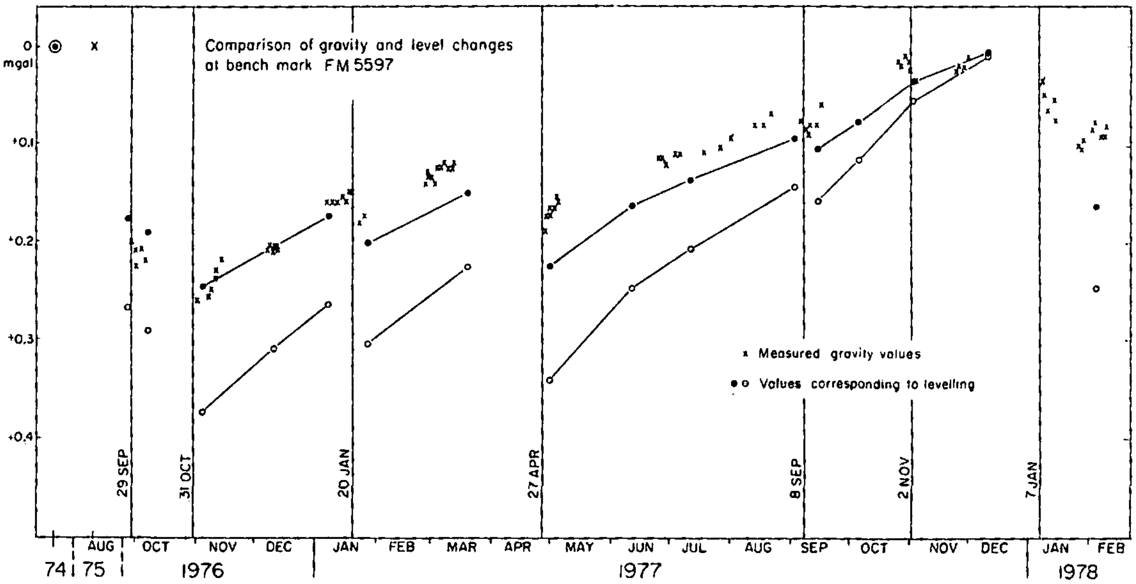

Fig. 8. Comparison of gravity and level changes with time at benchmark FM-5597. For location see Figure 5. Crosses show measured gravity values. Solid circles show gravity calculated from measured level values assuming the Bouguer model (0.01 mgal ≃ 3 cm). Open circles show gravity values corresponding to measured level values using the free air model (0.01 mgal ≃ 5 cm).

deflation of the Krafla caldera is accompanied by inflow and outflow of magma.

Taking the Bouguer model for granted, there remain some noteworthy discrepancies between the levelling and gravity observations. In particular, we point out the increasing gap between measured gravity values and those calculated from levelling after each deflation event during the winter and spring of 1977, and the sharp increase in measured gravity immediately after some of the deflation events (Figure 8). A possible explanation of both these effects is sinking of the groundwater level. In the first case we may have gradual sinking during the winter months when the ground is covered by snow and in the second case relatively rapid local sinking following an even more abrupt rising during the deflation event. Changes in groundwater pressure to this effect have been observed in drillholes. Expansion of the crustal rocks above the magma chamber, due to, e.g., increased steam pressure, would, however, affect the gravity values in the same way as groundwater sinking. Results from other stations are not detailed enough to provide further insight into these effects.

Movements of Fissures

Widening of numerous preexisting fissures was observed during the earthquake swarm of December 1975 to February 1976. Due to this, simple fixtures were installed on a number of fissures in the Krafla fissure swarm. This allowed measurements of changes in the width of these fissures by a micrometer with an accuracy of better than 0.1 mm. At some other fissures or swarms of fissures, nails were driven into the lava on opposite sides and their separation measured with a steel tape. Measurements are now made at about 30 sites both within the caldera and in the fissure swarm to the north and south. These are made daily on some of the fissures and on others at several days intervals. Additional measurements have been made after each period of rapid ground movement. New cracks in frozen ground or packed snow could be measured rather accurately. The good agreement between the aggregate widening of fissures and Geodimeter measurements show that these fissure measurements are quite accurate in determining the total widening of the fissure swarm.

Some of the fissures within the caldera show opening during inflation and closing during subsidence, correlating well with the tilt of the power-house as shown in Figure 9. The fissures in the fissure swarm outside the caldera show no or very little movement except during subsidence events. Then most of the fissures in the center of the active area widen rapidly up to several centimeters. Other fissures on the flanks of the fissure swarm close at the same time. These deformations are mostly permanent.

Continuous recording extensiometers have now been installed at five locations, but no results are available as yet.

Distance Measurements

An extensive program of Geodimeter measurements was initiated in early 1977 to determine the horizontal component of ground deformation in the Krafla fissure swarm. Preliminary analyses of measurements on single east-west lines across the active area before and after subsidence events have shown expansion of more than 3 m across the Namafjall area. There the extension seems to be confined to the central part of the fissure swarm, which is about 1 km wide. On both the flanks to the east and west of the center region the measurements have

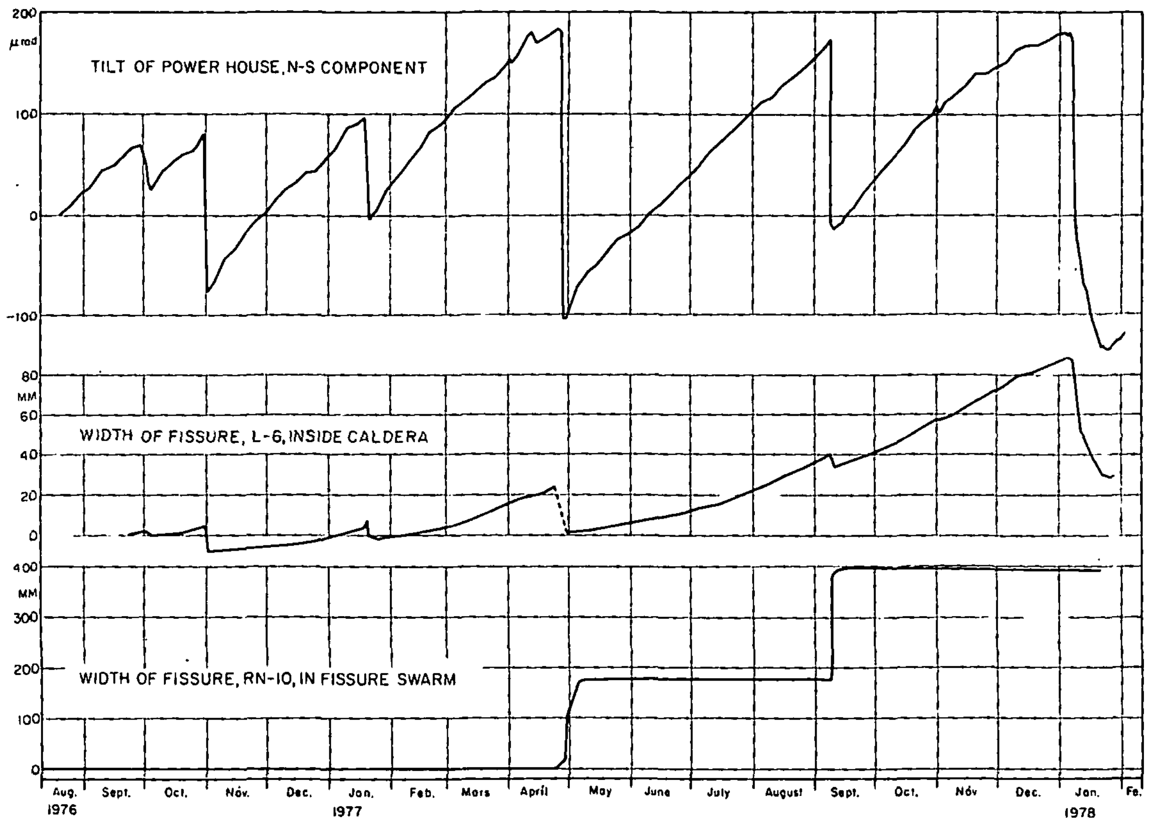

Fig. 9. North-south component of tilt at the Krafla power house from August 1976 to February 1978 and variation in width of fissures, near Leirhnjukur inside the caldera (L-6), and in the fissure swarm by Namafjall about 10 km south of the center of the caldera (RN-10). See Figure 2 for location.

shown compression of the order of some tens of centimeters. Figure 6 shows the expansion which occurred on April 27, 1977, in the Namafjall region south of the Krafla caldera. The expansion measured with the Geodimeter is nearly the same as the aggregate widening of individual cracks and also in good agreement with measurements made on fixtures on the preexisting fissures. Frequent Geodimeter measurements on a line from the center of the caldera to Namafjall have shown periodic horizontal variations which correlate well with the level variations.

Some Other Observations

A very dense seismic network has been in operation in the Krafla-Axarfjördur region during the present tectonic episode in North Iceland.

Several other measurements have been made in the area either in order to investigate the high-temperature field or to monitor the tectonic movements presently taking place. They include dc-resistivity soundings, monitoring of ground temperature and groundwater level, self-potential measurements, and magnetotelluric measurements. Temperature, pressure, and chemical composition of borehole discharge is monitored regularly. Gases emitted from fumaroles are analyzed, and close watch is kept on changes of the geothermal fields.

Some of these observations are important for analyzing the horizontal and vertical ground movements. During all major subsidence events an increased pressure or uplift of groundwater level has been observed in the deep drillholes both in the Krafla and the Namafjall geothermal fields. At several locations in Gjastykki, the fissure swarm to the north of the caldera, new geothermal fields have been formed and increased activity has been observed in old ones, especially at Namafjall.

Interpretation and Discussion

The inflation of the Krafla area between subsidence events is interpreted as being caused by inflow of magma from below into a magma chamber at shallow depth beneath the center of the caldera, analog to conditions at Kilauea, Hawaii [*Decker*, 1966; *Fiske and Kinoshita*, 1969; *Kinoshita et al.*, 1974].

Model calculations using the simple model of a spherical chamber within an elastic half-space (the Mogi model, [*Mogi*, 1958]) show good agreement between the calculated and observed ground deformation. Best agreement is obtained if the center of the spherical magma chamber is at a depth of 3 km (Figure 10). The volume increase of the inflation bulge is close to 5 m^3/s, and this value is taken as the average rate of inflow of magma from below into the Krafla magma chamber. The observations do not show where this magma flow originates.

At the beginning of each inflation period the crust behaves in an elastic manner and no significant earthquakes are observed within the Krafla caldera. After the uplift has reached a certain critical level, the deformation of the floor of the caldera is no longer purely elastic and earthquakes start occurring. The fissures near the center of the apex of uplift widen in good correlation with the uplift.

The subsidence of the Krafla area is caused by flow of magma out of the magma chamber. This outflow is primarily horizontal and follows the fissure swarm toward north or south, to the region where rifting occurs during each subsidence event (Figures 4 and 11). The widening of the rift system which allows the flow of magma into the fissures is presumably caused by regional east-west tension which has been built up gradually for one or two centuries due to east-west drift on the constructive plate boundary. The sudden widening of the fissure swarm is accompanied by elastic contraction of wide zones on both sides of the fissures (Figure 6). The pulsation of the rifting episode may be due to limited supply of magma at the beginning of the present episode, so the part of the fissure system which was ready to move could not all be filled with magma in one pulse. Additional supply of magma was needed and this has been entering the magma chamber at the rate of 5 m^3/s since the beginning of the rifting episode in December 1975. Outflow of magma from the magma chamber into the fissure swarm is blocked after each

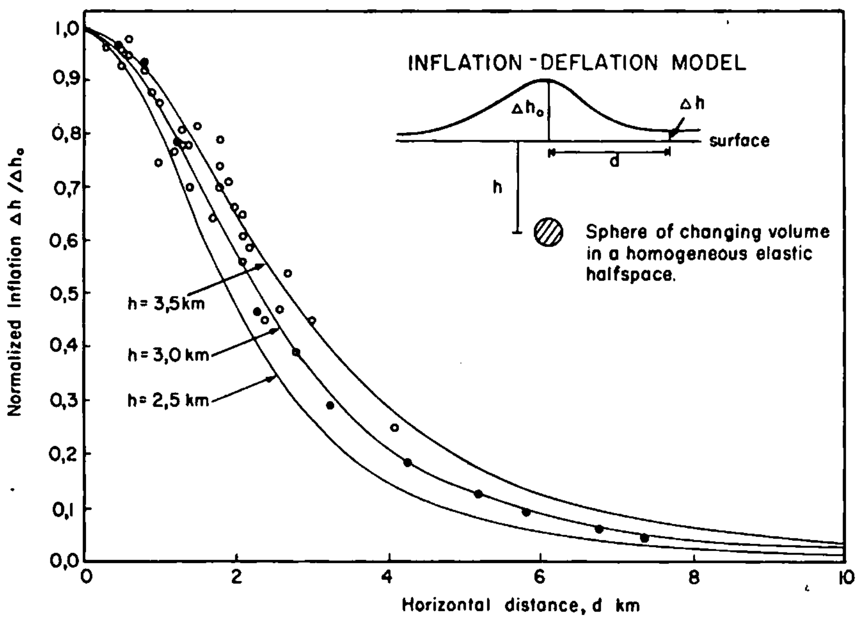

Fig. 10. Inflation-deflation model of the Mogi-type consisting of sphere of changing volume in homogeneous elastic half space, compared with elevation data from the Krafla area. Open circles are data from inside the caldera. Solid circles are measurements on a line from the apex of uplift to the south. The apex of uplift is near the center of the Krafla caldera at 65°42.7'N, 16°47.8'W (see Figure 5).

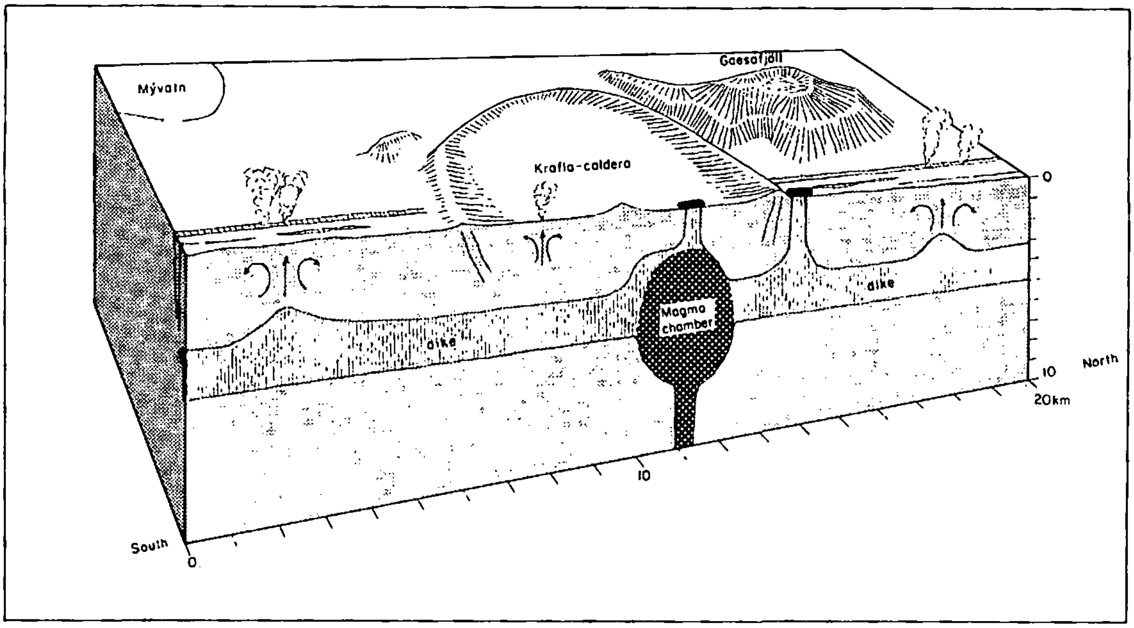

Fig. 11. Block diagram showing schematically the magma chamber below the Krafla caldera and the dike that has been formed in the present tectonic episode. This dike extends tens of kilometers farther north than indicated. Approximate location of new and intensified steam fields and lava eruptions in 1975 and 1977 are also shown.

pulse of activity and this flow is not resumed until a certain critical pressure has been built up in the magma chamber.

Continuation of the present processes must at some time in the future lead to a condition where no more widening of the fissure system is possible. If the inflow of magma from below into the Krafla magma chamber continues after this condition is reached, the expected result is intermittent or continuous lava eruption as long as the inflow continues.

According to this interpretation, the flow of magma in the crust is a secondary phenomena and the regional plate movement the primary factor. On the other hand, it seems clear that when rifting occurs it starts where the mechanical strength of the crust is lowest. In the Krafla area this place is within the caldera where the magma chamber is located. Thus we can look at the cladera and the magma chamber as a trigger for the episodical drift movements on the plate boundary in northern Iceland.

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WARD

Rifting of the Plate Boundary in North Iceland 1975–1978

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A rifting episode started in 1975 on the accreting plate boundary in North Iceland after 100 years of quiescence. Horizontal extension of some 3 m has been observed in the Krafla caldera and the associated 80 km long fissure swarm. The rifting occurs periodically in short active pulses at a few months intervals. Between these active pulses, continuous inflation of 7–10 mm/day of the caldera is caused by 5 m^3/s inflow of magma into a magma chamber at 3 km depth. The active pulses are caused by a sudden east-west expansion of the fissure swarm and a contraction of zones outside the fissure swarm. Rapid flow of magma out of the magma chamber and into the fissures toward north or south is indicated. These pulses are accompanied by earthquake swarms and vertical ground movements of up to 2 m and sometimes also volcanic eruptions and formation of new fumaroles. The magma chamber below the Krafla caldera thus acts as a trigger for the plate movement along the constructive plate boundary in North Iceland.

INTRODUCTION

The plate boundary between the European and the American plates follows the Mid-Atlantic Ridge and crosses Iceland from south-west to north-east. In Iceland, the boundary is characterized by zones of recent volcanism, graben structures, and seismic activity, and is generally named the Neovolcanic Zone. The Neovolcanic Zone in north Iceland has a north-south direction and is characterized by several fault and fissure swarms, each passing through a central volcano (Figure 1). The tectonic and volcanic activity of the Neovolcanic Zone is restricted to these central volcanoes and the associated fissure swarms, and occurs episodically rather than continuously, with a period of 100–150 years. During each active period, which probably lasts 5–20 years, only one central volcano and fissure swarm is active [*Björnsson et al.*, 1977].

The Krafla fissure swarm, which is presently active, extends from the Tjörnes Fracture Zone in the Axarfjördur bay in the north and some 100 km to the south. Its width is approximately 5 km, but varies considerably along the swarm. It passes through the Krafla caldera, which formed during the last interglacial period, but has since been filled to the rim with eruptive material (Figure 2). In postglacial time, about 35 eruptions have taken place within this fissure swarm, most of them either within the Krafla caldera or in the Namafjall area, about 10 km south of Krafla [*Björnsson et al.*, 1977].

A geothermal field with temperatures exceeding 340°C at 2 km depth exists within the Krafla caldera, and another geothermal field located in the Namafjall area has a temperature exceeding 290°C at 1.8 km depth. The economic importance of these geothermal fields is responsible for a more intense research of this area than would have been possible otherwise.

There are historical records of only one period of volcanic and tectonic activity within the Krafla fissure swarm, the Myvatn Fires of 1724–1729. A very small eruption in 1746 may be regarded as the last observed pulse in the Myvatn Fires episode. Another period of volcanic and tectonic activity in North Iceland occurred in 1874–1875, but in another fissure swarm, the Askja fissure swarm. The Theistareykir fissure swarm was active in 1618, but no volcanic activity was reported [*Thoroddsen*, 1925]. After the volcanic and tectonic activity in the Askja fissure swarm in 1874–1875, there has been little tectonic activity within the neovolcanic zone in North Iceland until the Krafla fissure swarm became active in 1975. A period of volcanic activity in Askja 1921–1926 was not associated with any observed tectonic activity, and neither was the Askja eruption of 1961, except for some vertical ground movements within the Askja caldera.

Geodetic measurements intended to observe tectonic movements in the neovolcanic zone in North Iceland were initiated in 1938 [*Niemczyk*, 1943]. Remeasurement of the 1938 network in 1965 showed no significant ground deformation. More precise remeasurements in 1971 and 1975 showed significant expansion of the area around Krafla during the 1971–1975 interval after an indicated contraction of the same area during the 1965–1971 time interval [*Gerke*, 1969, 1974, 1977; *Schleusener and Torge*, 1971; *Spickernagel*, 1966; *Torge and Drewes*, 1977]. These observations may indicate that the present tectonic event had started before the summer of 1975 as an inflation of the Krafla area. Increased seismic activity of the Krafla region in early 1975 may also be interpreted as an indication of abnormal tectonic activity.

NARRATIVE OF EVENTS

Contemporary description of the Myvatn Fires episode in 1724–1729 shows clearly that the volcanic and seismic activity was largely confined to short periods of high activity interrupted by much longer periods of quiescence. Each active period lasted for only a few days, while the quiet periods lasted for several months. Each of the pulses of activity was characterized by strong earthquakes and either volcanic activity or noticeable changes in the geothermal activity. Some of the pulses were associated with changes in the level of Lake Myvatn, indicating large scale vertical ground movement [*Thoroddsen*, 1925].

The present volcano-tectonic episode in the Myvatn-Krafla area is also characterized by a similar pulsation. It is possible to divide the time since the active episode started in December 1975 into periods of two kinds, inflation periods and subsidence events. Figure 3 shows the elevation changes with time of bench mark FM-5596 near the center of the Krafla caldera from 1975 to early 1978. Since the initial subsidence event, in December 1975, the rate of uplift has been relatively constant,

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around 7 mm/day at this point, but it has been interrupted by seven sudden subsidence events lasting 2 hours to several days each. The inflation periods last for 1–7 months and are characterized by the following:

1. Continuous and nearly constant uplift of the Krafla region. The maximum uplift is near the center of the caldera, 7–10 mm/day, decreasing outward to less than 1 mm/day at a distance of 10 km from the apex of uplift.

2. Gradually increasing seismic activity within the caldera after the land elevation has reached a certain critical level. Decreasing or no seismic activity within the fissure swarm outside the caldera (P. Einarsson, personal communication, 1978).

3. Gradual widening of fissures near the center of uplift, up to 1 mm/day.

The duration of the subsidence events or active pulses is much shorter than the inflation periods. Some are so small that they are not noticed, except on measuring equipment, while other pulses correspond exactly to those described in the eighteenth century episode. These pulses of activity have the following common characteristics according to the available observations:

1. Subsidence of the Krafla region. The maximum subsidence, near the center of the caldera, has been from 3 to about 250 cm, but decreasing outward.

2. Continuous seismic tremor (volcanic tremor) which usually starts at the same time as the subsidence and lasts for a few hours.

3. Earthquake swarm in the fissure zone outside the Krafla caldera.

4. New fissures and east-west widening of the fissure swarm at the same place as the earthquake swarm. Widening of 2 m has been measured during a subsidence event.

5. Subsidence of the active part of the fissure swarm, sometimes exceeding 1 m, and uplift of both flanks of the swarm amounting to tens of centimeters.

6. Development of new geothermal areas or increased activity in old ones. Increased pressure in drillholes.

7. Outpouring of basaltic lava, mostly within the caldera, has been observed in three of the active pulses.

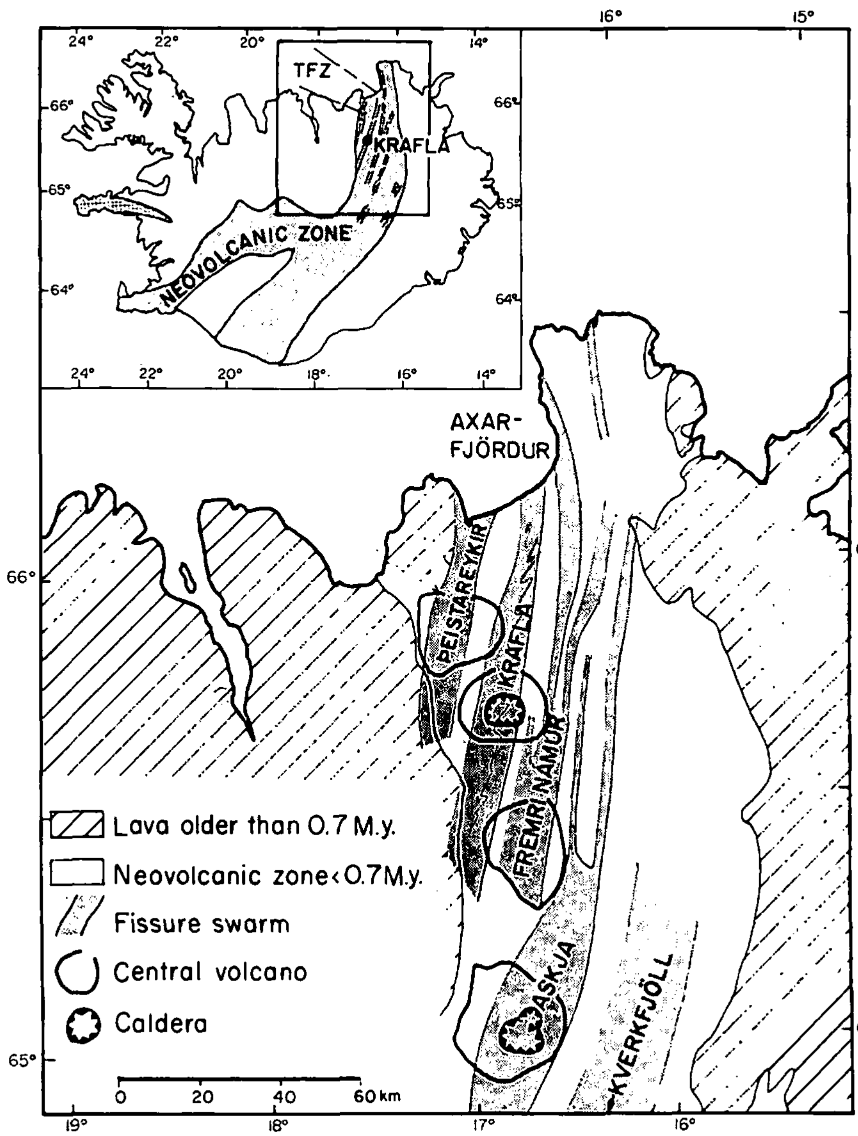

Fig. 1. The spreading zone in North Iceland. Central volcanoes and associated fissure swarms are named after the high-temperature geothermal fields in the central volcanoes. Two of them, Askja and Krafla, contain calderas. Mapped by Kristjan Saemundsson. The Tjörnes Fracture Zone (TFZ) is shown in the inset.

Figure 4 shows the areas of maximum ground deformation and rifting within the fissure swarm outside the caldera during different subsidence events.

The first pulse of high activity started on December 20, 1975, and lasted for several weeks. The maximum subsidence, near the center of the caldera, was some 2.5 m and an intense earthquake swarm was observed some 40–60 km north of Krafla where large scale ground movements occurred [*Björnsson*, 1976; *Björnsson et al.*, 1977; *Tryggvason*, 1976].

The second period of activity started about September 29, 1976. It lasted for some 5 days and the maximum subsidence was about 15 cm. The most noticeable feature of this pulse was the complete cessation of the seismic activity within the caldera, but this activity had been increasing gradually during the four preceeding months [*Björnsson et al.*, 1977].

The third active period started on October 31, 1976, at about 2 a.m. and lasted for less than 48 hours. The maximum subsidence was about 50 cm and intense volcanic tremor ac-

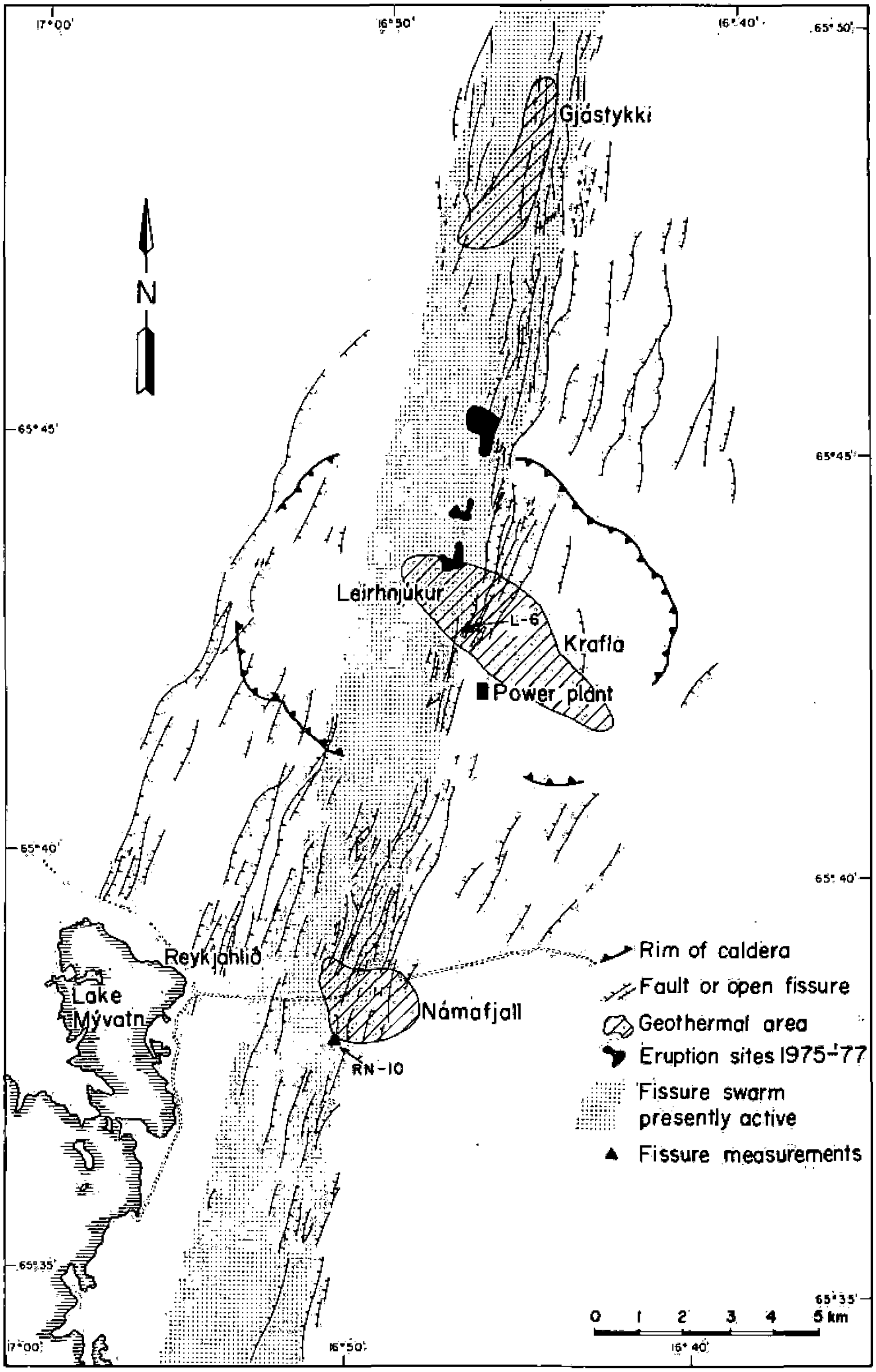

Fig. 2. Outline geological map of the Krafla caldera and the associated fissure swarm. Mapped by Kristjan Saemundsson.

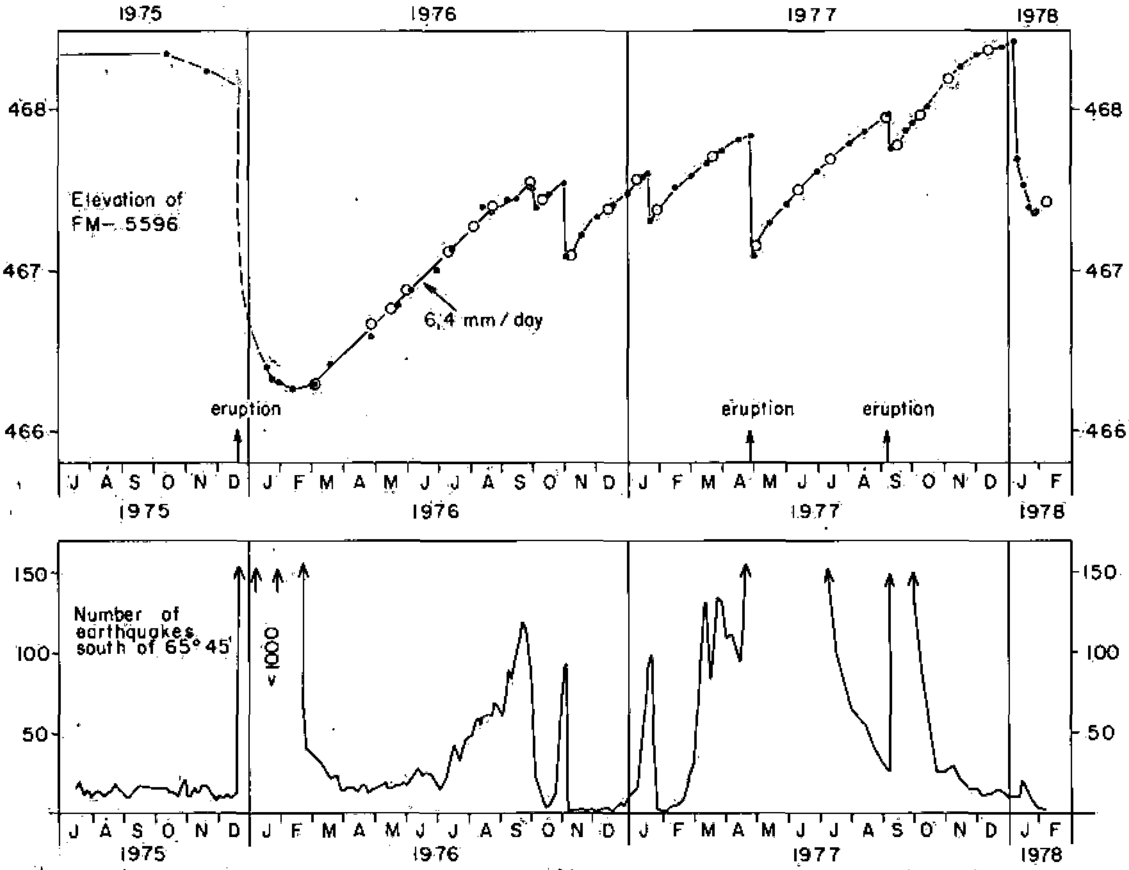

Fig. 3. Changes in elevation of bench mark FM-5596 within the Krafla caldera from 1975 to early 1978 (see Figure 5 for location). Levelling data (open circles) are supplemented with tilt data at the Krafla power house (dots). The rate of uplift is relatively constant, 5–7 mm/day, interrupted by eight sudden subsidence events. The subsidence event of November 2, 1977, was too small to be shown on this graph. The lower part shows running 5-day averages of number of earthquakes within and south of the Krafla caldera. After April 27, 1977, most of the earthquakes occurred south of the caldera. The earthquake information was kindly supplied by P. Einarsson.

companied the subsidence. Widening of old fissures and new fumaroles were observed in the fissure swarm 10–15 km north of the caldera.

The fourth active period started on January 20, 1977, shortly after midnight and lasted for only about 20 hours. The maximum subsidence was about 30 cm and widening of fissures, and new fumaroles were observed about 10 km north of the caldera.

The fifth subsidence event started on April 27, 1977, at about 1 p.m. with very intense volcanic tremor. A small lava, covering only 0.01 km², was erupted about 5 p.m. near the north rim of the caldera. The floor of the caldera subsided irregularly but the maximum subsidence was more than one meter. Widening of fissures to the south of the caldera was observed. An east-west widening of the fissure swarm of 2 m was obtained the next day by measuring the opening of individual cracks in frozen ground on a profile across the Namafjall area. A remeasurement of a geodimeter line along the same profile showed a widening of 2.0 m in excellent agreement with the direct measurements of fissures. The widening of fissures spread out from the caldera and reached the Namafjall area in about 5 hours. This indicates a velocity of the order of 0.5 m/s for the horizontal movement of activity. During this active pulse the central part of the active fissure swarm subsided about one meter but the flanks to the east and west were uplifted some tens of centimeters.

The sixth subsidence event started with volcanic tremor around 4 p.m. on September 8, 1977. The course of events was very similar as on April 27. A volcanic eruption started at about 6 p.m. near the north rim of the caldera. The area covered with lava was about 0.8 km² and the volume of lava is estimated as $2 \cdot 10^6$ m³. Another volcanic eruption occurred about 11:45 p.m. in the Namafjall area where some 2500 kg of basaltic pumice erupted through a borehole, 1138 m deep. This location is about 12 km south of the lava eruption, in the fissure swarm where the east-west widening was 1 m.

The seventh subsidence event occurred on November 2, 1977, and was the least noticeable of the observed active periods to date. It lasted for only 2 hours and the total subsidence was 2–3 cm. A small tremor was seen on the local seismometers, but no movement of fissures was observed.

The eighth subsidence event started in the afternoon of January 6, 1978, and lasted for some 3 weeks. The maximum subsidence within the caldera during this event exceeded 1 m, and the accompanying earthquake swarm was strongly felt 20–50 km north of Krafla, where significant subsidence and widening of fissures was observed.

Observations, Techniques, and Data

The credibility of the observed ground deformation and the interpretations based thereon are entirely dependent on the nature and extent of the measurements. Therefore it is certainly in order to give some details of the observational techniques and procedures along with the presentation of data. Only those observations which are pertinent in analyzing the tectonic processes are described.

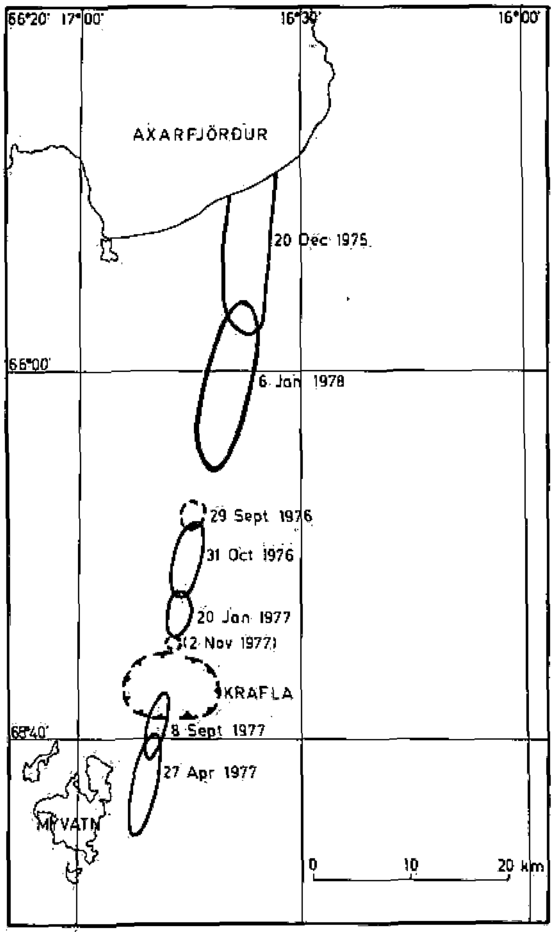

Fig. 4. Areas of maximum ground deformation and rifting outside the Krafla caldera during different subsidence events. No rifting was observed during the events of September 29, 1976; and November 2, 1977.

Levelling

Levelling was carried out in 1974 and earlier along the road from Myvatn to Krafla and around the geothermal field at Krafla. This levelling network was extended in 1976 and includes presently about 70 bench marks. Levelling has been performed at intervals of 1–2 months since March 1976 in a large part of this network and at longer intervals in the remainder. Zeiss Ni2 level and wooden measuring rods, compared with Wild invar rods, are used. The standard error of the levelling has been determined as approximately $1.5L^{1/2}$ mm, where L is the length of the foreward and backward measured levelling line in kilometers.

The area which has been rising and subsiding in the Krafla region has remained nearly constant during the last two years. The inflation bulge and the deflation bowl is nearly circular, although east-west elongation is indicated. Maximum vertical ground movement is observed near the center of the caldera. The half-width of the bowl or bulge is about 3 km and only very minor movement is measured at 10 km distance from the center. Some details of the inflation-deflation pattern are shown in Figure 5.

The elevation changes within and around the Krafla caldera have been cyclic up and down movements (Figure 3), while elevation changes on the fissure swarm to the south and north of the caldera are more or less transient phenomena. During subsidence events rapid and permanent changes in land elevation are observed, accompanied by east-west widening and earthquake swarms. Figure 6 shows land elevation changes on an east-west profile across the fissure swarm near Namafjall during the subsidence event of April 27, 1977. A 1 km wide segment in the middle of the fissure swarm subsided about 80 cm, while the flanks to the east and west of the active zone were uplifted about 30 cm. Similar elevation changes have occurred in other subsidence events where these movements are indicated by vertical displacements of faults, changes of ground water level, and changes in shore-lines of lakes.

Measured gravity changes support these observations, as can be seen in Figure 6. Similar gravity changes were observed in Gjastykki, north of the caldera, during the subsidence event of January 20, 1977.

Tilt Measurements

A water tube tiltmeter was installed on a semi-permanent basis in the Krafla power house on August 19, 1976. Three measuring pots are connected. The north-south arm of this tiltmeter is 68.95 m long and the east-west arm 19.50 m. The reading accuracy is better than 1 microradian, but temperature variations and corresponding thermal expansions of the building cause tilt error of the same magnitude. These can be partly corrected for. Readings are usually made once a day, but during periods of rapid ground movements, more frequent readings are made. During the period January–August 1976, optical levelling was frequently performed between markers on the four corners of the power house with an accuracy of better than 1 mm.

The power house is located about 1.3 km from the apex of the inflation-deflation bowl, and the north-south arm of the tiltmeter is directed almost exactly toward the apex. Hence it is ideally located and oriented to monitor tilt changes caused by elevation changes in the Krafla caldera. An excellent correlation has been found between these two parameters, tilt of the power house and ground elevation inside the Krafla caldera. The tilt can thus be used to monitor daily variations in elevation and volume changes during inflation and deflation events.

Figure 3 shows the land elevation as derived from levelling and tilt observations in the power house. Uplift or subsidence of the apex of the inflation-deflation area is approximately 3.4 mm for each microradian of tilt at the power house.

Additional dry-tilt stations have been established in the Myvatn-Krafla area, consisting of 5–6 bench marks each. Most of these tilt stations are so constructed that the bench marks lie on a circle of 25 m radius and, during measurements, an optical level, Wild N3, is placed exactly in the center of this circle and invar rods are placed on the markers. The relative elevation of the bench marks is established with an accuracy of approximately 0.1 mm, allowing to determine ground tilt of less than 5 microradians. The principal source of error of these tilt measurements is, however, internal deformation of the area covered by each tilt station. A total of 12 such tilt stations are under observation about once every month in the summer, but these measurements are rarely made in the winter due to the snow cover. The arrows in Figure 7 show observed tilt variations during the inflation period April 29 to September 8, 1977 [*Tryggvason*, 1978]. This correlates well with the levelling measurements. Continuously recording electronic tiltmeters were installed at two locations in late 1977.

Gravity Survey

A gravity survey was carried out in the Krafla area in August 1975 prior to the first Leirhnjúkur eruption. The purpose was to monitor gravity variations that might be caused by

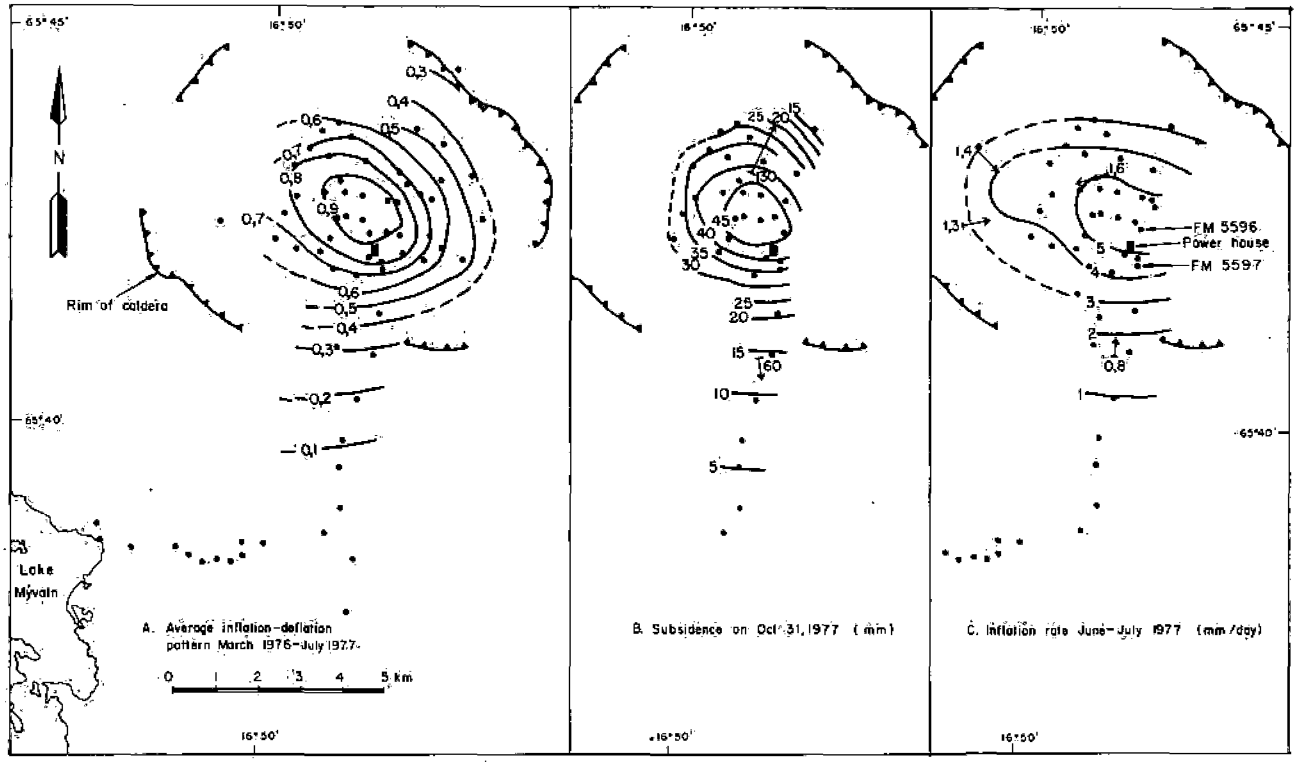

Fig. 5. Pattern of uplift and subsidence in the Krafla area. (*a*) Average vertical ground movement throughout the period March 1976 to July 1977 in fractions of maximum movements. (*b*) Total subsidence in centimeters during the subsidence event of October 31 to November 1, 1976. (*c*) Rate of uplifting in millimeters per day during a period of 1 month, June–July 1977. Arrows show tilt changes in microradians at four stations.

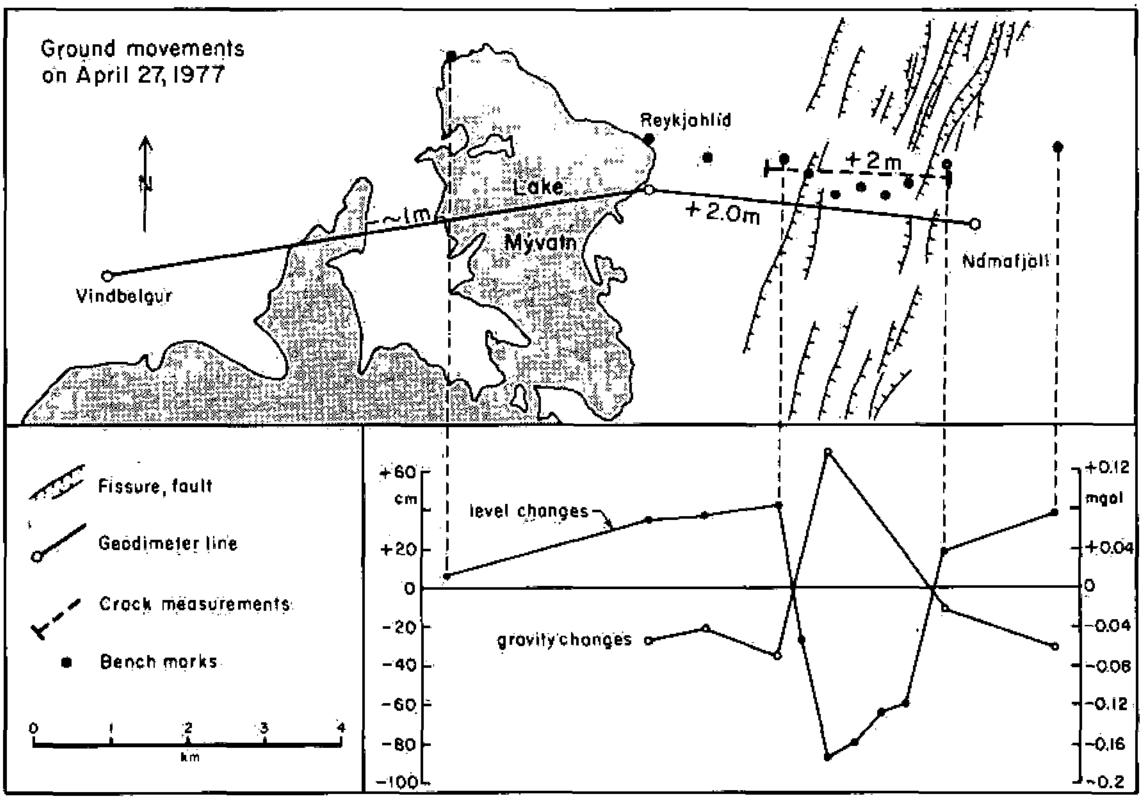

Fig. 6. Horizontal and vertical ground movements across the Krafla fissure swarm near Namafjall during the subsidence event of Arpil 27, 1977. Expansion is measured on individual cracks and by geodimeter. Elevation and gravity changes are shown in the lower part of the figure.

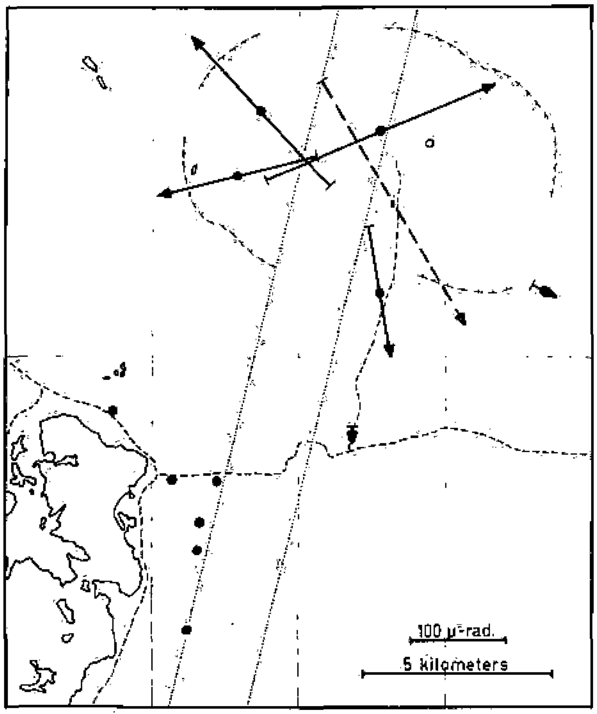

Fig. 7. Observed ground tilt during the inflation period April 29 to September 8, 1977. Arrows show direction and magnitude of observed tilt, and if no arrows are drawn, the tilt was less than the error of the observations. Solid dots are dry-tilt stations. The direction of tilt at the power house (dashed arrow) is based on an electronic tiltmeter which was operated for a few weeks during this period. Dotted lines outline the most active part of the Krafla fissure swarm. For location see Figures 1 and 2.

groundwater changes due to removal of fluid from drillholes in the goethermal fields. The network included some 30 stations and the instrument used was a LaCoste-Romberg gravity meter, G-10. This network was reoccupied in March 1976 and June 1976 using an old Worden gravity meter, W-68. This instrument, however, turned out to be rather unreliable, and one could only conclude that a significant gravity change had occurred (a few tenths of mgal) in the Krafla caldera since August 1975. In September 1976 a survey was carried out on an extended net using a new LaCoste-Romberg instrument, G-445, and the survey has since then been repeated at intervals of 1–2 months with the same instrument. The network has gradually been extended to more than 100 stations, although not all are included in each survey. Gravity changes were originally referred to a base station at Reykjahlid by Lake Myvatn, but since April 1977 corrections have been made for changes in that station relative to another base station at Husavik. The accuracy of the values measured by the G-meters is 0.01 to 0.02 mgals after corrections have been made for tidal effects and instrumental drift.

The main purpose of the gravity survey since 1976 has been to monitor elevation changes over a more extensive area than is covered with the level surveys. However, the conversion of gravity changes into elevation changes requires knowledge of changes in mass distribution. As this knowledge is not present, two models have been considered; (1) a free air model, leading to approximately 3 cm elevation change for 0.01 mgal gravity change, and (2) a Bouguer model assuming a density of 2.5 g/cm^3, leading to approximately 5 cm elevation change for 0.01 mgal gravity change.

Repeated gravity measurements have been made at several stations that are regularly included in the levelling survey. Figure 8 shows the results from one of these stations, FM-5597, located near the Krafla power house (for location see Figure 5). A comparison is made between measured gravity changes and gravity changes that correspond to measured level changes using the two models. The comparison indicates that the Bouguer model leads to a fairly close agreement with the level changes, although notice should be taken of the additional assumption that no ground movement took place between the levelling survey in 1974 and the gravity survey in 1975. Results from other stations support this conclusion. Since the Bouguer model corresponds to an assumption of no change of density, a subsidiary result of the gravity survey has been to lend support to the theory that the inflation and

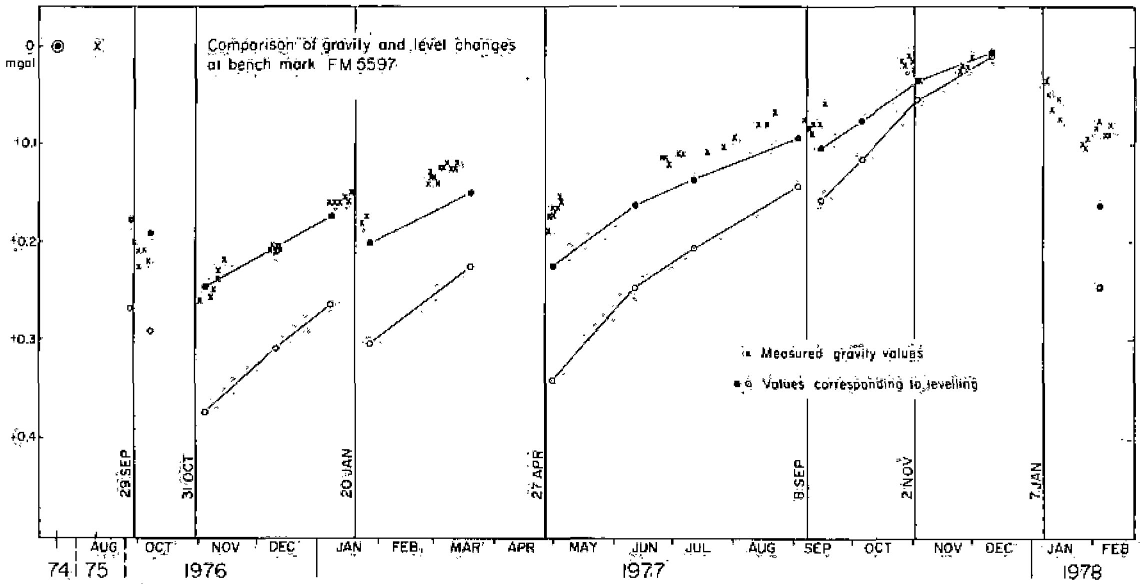

Fig. 8. Comparison of gravity and level changes with time at benchmark FM-5597. For location see Figure 5. Crosses show measured gravity values. Solid circles show gravity calculated from measured level values assuming the Bouguer model (0.01 mgal ≈ 3 cm). Open circles show gravity values corresponding to measured level values using the free air model (0.01 mgal ≈ 5 cm).

deflation of the Krafla caldera is accompanied by inflow and outflow of magma.

Taking the Bouguer model for granted, there remain some noteworthy discrepancies between the levelling and gravity observations. In particular, we point out the increasing gap between measured gravity values and those calculated from levelling after each deflation event during the winter and spring of 1977, and the sharp increase in measured gravity immediately after some of the deflation events (Figure 8). A possible explanation of both these effects is sinking of the groundwater level. In the first case we may have gradual sinking during the winter months when the ground is covered by snow and in the second case relatively rapid local sinking following an even more abrupt rising during the deflation event. Changes in groundwater pressure to this effect have been observed in drillholes. Expansion of the crustal rocks above the magma chamber, due to, e.g., increased steam pressure, would, however, affect the gravity values in the same way as groundwater sinking. Results from other stations are not detailed enough to provide further insight into these effects.

Movements of Fissures

Widening of numerous preexisting fissures was observed during the earthquake swarm of December 1975 to February 1976. Due to this, simple fixtures were installed on a number of fissures in the Krafla fissure swarm. This allowed measurements of changes in the width of these fissures by a micrometer with an accuracy of better than 0.1 mm. At some other fissures or swarms of fissures, nails were driven into the lava on opposite sides and their separation measured with a steel tape. Measurements are now made at about 30 sites both within the caldera and in the fissure swarm to the north and south. These are made daily on some of the fissures and on others at several days intervals. Additional measurements have been made after each period of rapid ground movement. New cracks in frozen ground or packed snow could be measured rather accurately. The good agreement between the aggregate widening of fissures and Geodimeter measurements show that these fissure measurements are quite accurate in determining the total widening of the fissure swarm.

Some of the fissures within the caldera show opening during inflation and closing during subsidence, correlating well with the tilt of the power-house as shown in Figure 9. The fissures in the fissure swarm outside the caldera show no or very little movement except during subsidence events. Then most of the fissures in the center of the active area widen rapidly up to several centimeters. Other fissures on the flanks of the fissure swarm close at the same time. These deformations are mostly permanent.

Continuous recording extensiometers have now been installed at five locations, but no results are available as yet.

Distance Measurements

An extensive program of Geodimeter measurements was initiated in early 1977 to determine the horizontal component of ground deformation in the Krafla fissure swarm. Preliminary analyses of measurements on single east-west lines across the active area before and after subsidence events have shown expansion of more than 3 m across the Namafjall area. There the extension seems to be confined to the central part of the fissure swarm, which is about 1 km wide. On both the flanks to the east and west of the center region the measurements have

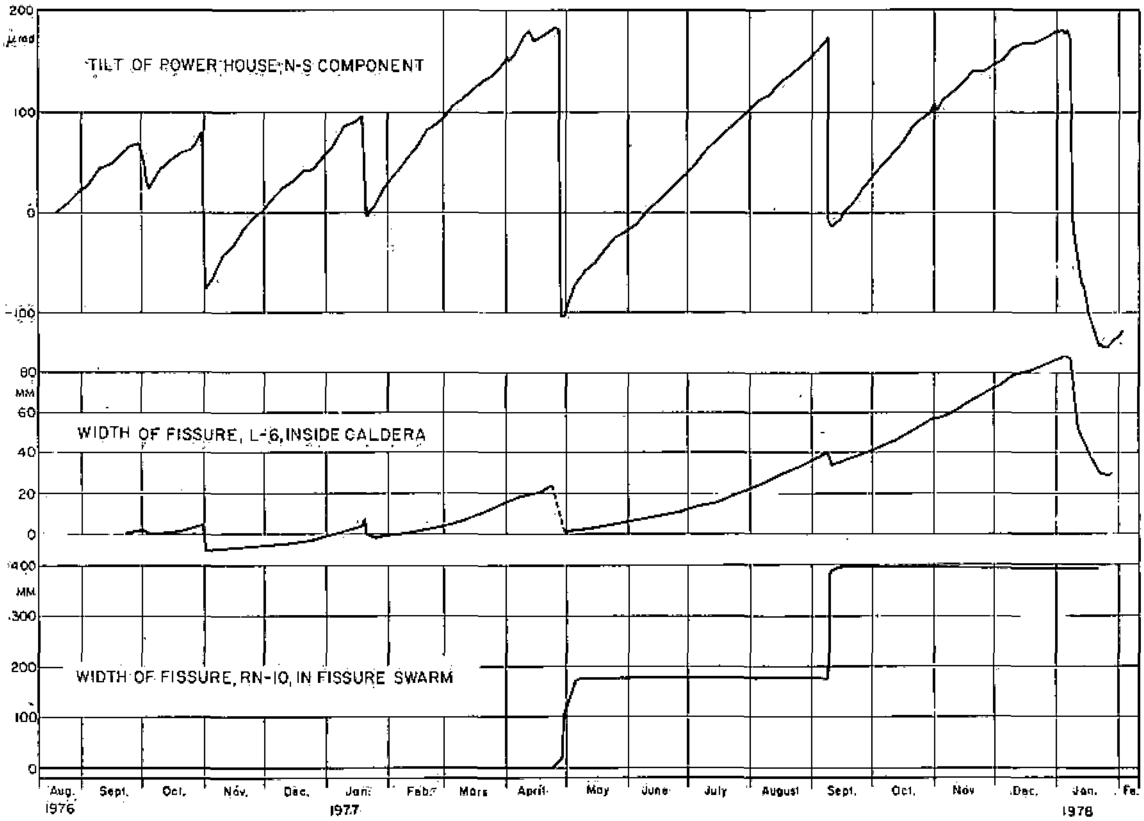

Fig. 9. North-south component of tilt at the Krafla power house from August 1976 to February 1978 and variation in width of fissures, near Leirhnjukur inside the caldera (L-6), and in the fissure swarm by Namafjall about 10 km south of the center of the caldera (RN-10). See Figure 2 for location.

shown compression of the order of some tens of centimeters. Figure 6 shows the expansion which occurred on April 27, 1977, in the Namafjall region south of the Krafla caldera. The expansion measured with the Geodimeter is nearly the same as the aggregate widening of individual cracks and also in good agreement with measurements made on fixtures on the preexisting fissures. Frequent Geodimeter measurements on a line from the center of the caldera to Namafjall have shown periodic horizontal variations which correlate well with the level variations.

Some Other Observations

A very dense seismic network has been in operation in the Krafla-Axarfjördur region during the present tectonic episode in North Iceland.

Several other measurements have been made in the area either in order to investigate the high-temperature field or to monitor the tectonic movements presently taking place. They include dc-resistivity soundings, monitoring of ground temperature and groundwater level, self-potential measurements, and magnetotelluric measurements. Temperature, pressure, and chemical composition of borehole discharge is monitored regularly. Gases emitted from fumaroles are analyzed, and close watch is kept on changes of the geothermal fields.

Some of these observations are important for analyzing the horizontal and vertical ground movements. During all major subsidence events an increased pressure or uplift of groundwater level has been observed in the deep drillholes both in the Krafla and the Namafjall geothermal fields. At several locations in Gjastykki, the fissure swarm to the north of the caldera, new geothermal fields have been formed and increased activity has been observed in old ones, especially at Namafjall.

INTERPRETATION AND DISCUSSION

The inflation of the Krafla area between subsidence events is interpreted as being caused by inflow of magma from below into a magma chamber at shallow depth beneath the center of the caldera, analog to conditions at Kilauea, Hawaii [*Decker*, 1966; *Fiske and Kinoshita*, 1969; *Kinoshita et al.*, 1974].

Model calculations using the simple model of a spherical chamber within an elastic half-space (the Mogi model, [*Mogi*, 1958]) show good agreement between the calculated and observed ground deformation. Best agreement is obtained if the center of the spherical magma chamber is at a depth of 3 km (Figure 10). The volume increase of the inflation bulge is close to 5 m^3/s, and this value is taken as the average rate of inflow of magma from below into the Krafla magma chamber. The observations do not show where this magma flow originates.

At the beginning of each inflation period the crust behaves in an elastic manner and no significant earthquakes are observed within the Krafla caldera. After the uplift has reached a certain critical level, the deformation of the floor of the caldera is no longer purely elastic and earthquakes start occurring. The fissures near the center of the apex of uplift widen in good correlation with the uplift.

The subsidence of the Krafla area is caused by flow of magma out of the magma chamber. This outflow is primarily horizontal and follows the fissure swarm toward north or south, to the region where rifting occurs during each subsidence event (Figures 4 and 11). The widening of the rift system which allows the flow of magma into the fissures is presumably caused by regional east-west tension which has been built up gradually for one or two centuries due to east-west drift on the constructive plate boundary. The sudden widening of the fissure swarm is accompanied by elastic contraction of wide zones on both sides of the fissures (Figure 6). The pulsation of the rifting episode may be due to limited supply of magma at the beginning of the present episode, so the part of the fissure system which was ready to move could not all be filled with magma in one pulse. Additional supply of magma was needed and this has been entering the magma chamber at the rate of 5 m^3/s since the beginning of the rifting episode in December 1975. Outflow of magma from the magma chamber into the fissure swarm is blocked after each

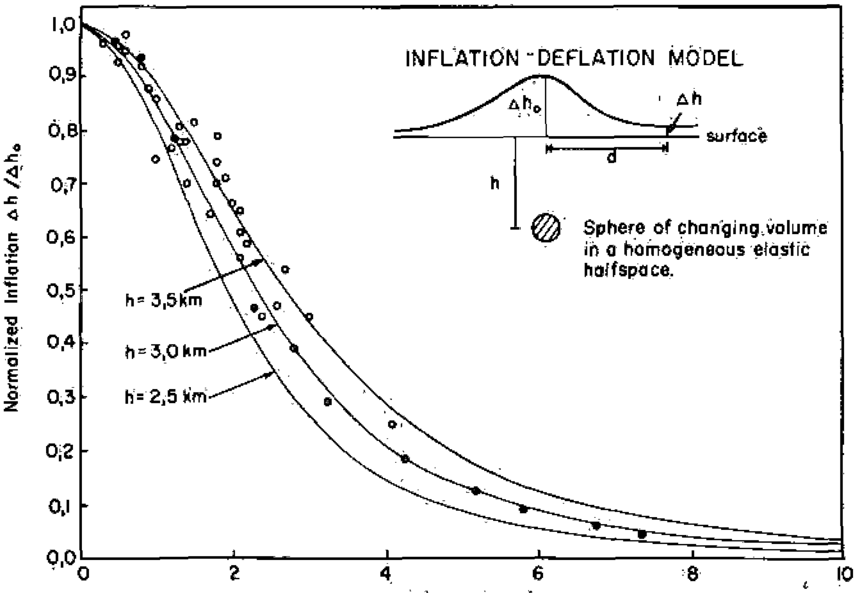

Fig. 10. Inflation-deflation model of the Mogi-type consisting of sphere of changing volume in homogeneous elastic half space, compared with elevation data from the Krafla area. Open circles are data from inside the caldera. Solid circles are measurements on a line from the apex of uplift to the south. The apex of uplift is near the center of the Krafla caldera at 65°42.7'N, 16°47.8'W (see Figure 5).

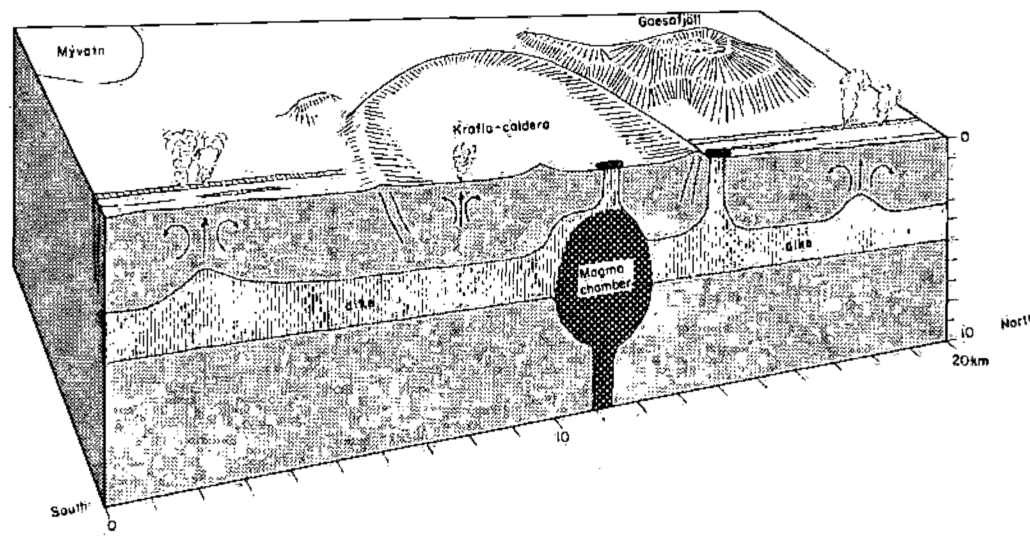

Fig. 11. Block diagram showing schematically the magma chamber below the Krafla caldera and the dike that has been formed in the present tectonic episode. This dike extends tens of kilometers farther north than indicated. Approximate location of new and intensified steam fields and lava eruptions in 1975 and 1977 are also shown.

pulse of activity and this flow is not resumed until a certain critical pressure has been built up in the magma chamber.

Continuation of the present processes must at some time in the future lead to a condition where no more widening of the fissure system is possible. If the inflow of magma from below into the Krafla magma chamber continues after this condition is reached, the expected result is intermittent or continuous lava eruption as long as the inflow continues.

According to this interpretation, the flow of magma in the crust is a secondary phenomena and the regional plate movement the primary factor. On the other hand, it seems clear that when rifting occurs it starts where the mechanical strength of the crust is lowest. In the Krafla area this place is within the caldera where the magma chamber is located. Thus we can look at the cladera and the magma chamber as a trigger for the episodical drift movements on the plate boundary in northern Iceland.

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WARD

A Hydrological Model for the Flow of Thermal Water in Southwestern Iceland with Special Reference to the Reykir and Reykjavik Thermal Areas

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ABSTRACT

Characteristically the temperature of thermal water in Quaternary rocks west of the volcanic zone in southwestern Iceland increases with distance from the volcanic zone which is reciprocal to the trend of the regional heat flow and most likely is caused by a decrease in rock porosity away from the volcanic zone.

A comparison of a regional deuterium/hydrogen (D:H) ratio map of rain and the D:H ratio of thermal water from several springs and wells outside and from the Reykir and Reykjavik thermal fields suggests that the thermal water (100 to 140°C) in the oldest rocks may be precipitated as rain in the interior highlands, whereas that (80°C) in the younger rocks may be precipitated nearby in the volcanic zone.

The thermal areas in question are in Quaternary volcanics characterized by thick successions of low-porosity lavas intercalated by high-porosity subglacial volcanics, which form ridges tens of kilometers long, 1 to 5 km broad, and sometimes hundreds of meters thick within the strata parallel to the volcanic zone. We suggest that these high-porosity volcanics may serve as channels along which water may flow at depth from the highland areas toward the coast.

A regional electrical resistivity survey (to 1500 m depth) supports a picture derived from the geological and hydrological data, wherein close to and within the volcanic zone there may be a large-scale circulation system of local water, but in the older rocks the water may flow long distances parallel to the volcanic zone. Evidence is given for mixing within a thermal system of water derived from the two recharge areas.

INTRODUCTION

Due to Iceland's location on a constructive plate boundary (the Mid-Atlantic Ridge) the regional heat flow is very high there (Palmason, 1973) as compared with most parts of the world. Hydrothermal activity is widespread in the country (Bodvarsson, 1961). The thermal areas are divided into two categories on the basis of the maximum temperature (base temperature) in the uppermost kilometer. The base temperature is thus higher than 200°C in the high-temperature areas, but lower than 150°C in the low-temperature areas.

The high-temperature areas are confined to, or on the margins of, the active zones of rifting and volcanism that run through the country (Palmason and Saemundsson, 1974), and the heat source for each high-temperature area is thought to be a local accumulation of igneous intrusions cooling at a shallow level in the crust. The low-temperature areas are, on the other hand, in Quaternary and Tertiary volcanics, and are thought to draw heat from the regional heat flow.

The present contribution deals with low-temperature hydrothermal activity in early Quaternary rocks west of the active volcanic zone in southwestern Iceland.

GEOLOGICAL FEATURES

The active volcanic zone in southwestern Iceland is flanked symmetrically by Quaternary volcanics which in turn are flanked by Tertiary volcanics. The strata, which dip towards the volcanic zone, reflect continuous volcanic activity and crustal spreading in this part of the country during at least the last 7 million years (Fridleifsson, 1973; Saemundsson and Noll, 1975; Johannesson, 1975).

During the last 3 m.y. there have been over 20 major glaciations in Iceland. The Quaternary stratigraphic succession is, therefore, characterized by sequences of subaerial lava flows intercalated by volcanic hyaloclastites and morainic horizons at intervals corresponding to glaciations.

Subglacial volcanics tend to pile up under the ice around the eruptive orifice. Eruptive fissures are the most common form of volcanoes in Iceland. Individual fissures are commonly several kilometers (and can be tens of kilometers) long. During a major glaciation, fissure eruptions can produce a series of parallel hyaloclastite ridges 1 to 5 km broad and several hundred meters thick along the entire active volcanic zone. In subsequent subaerial eruptions, lava flows will bank up against the hyaloclastites and may eventually bury them (Fig. 1).

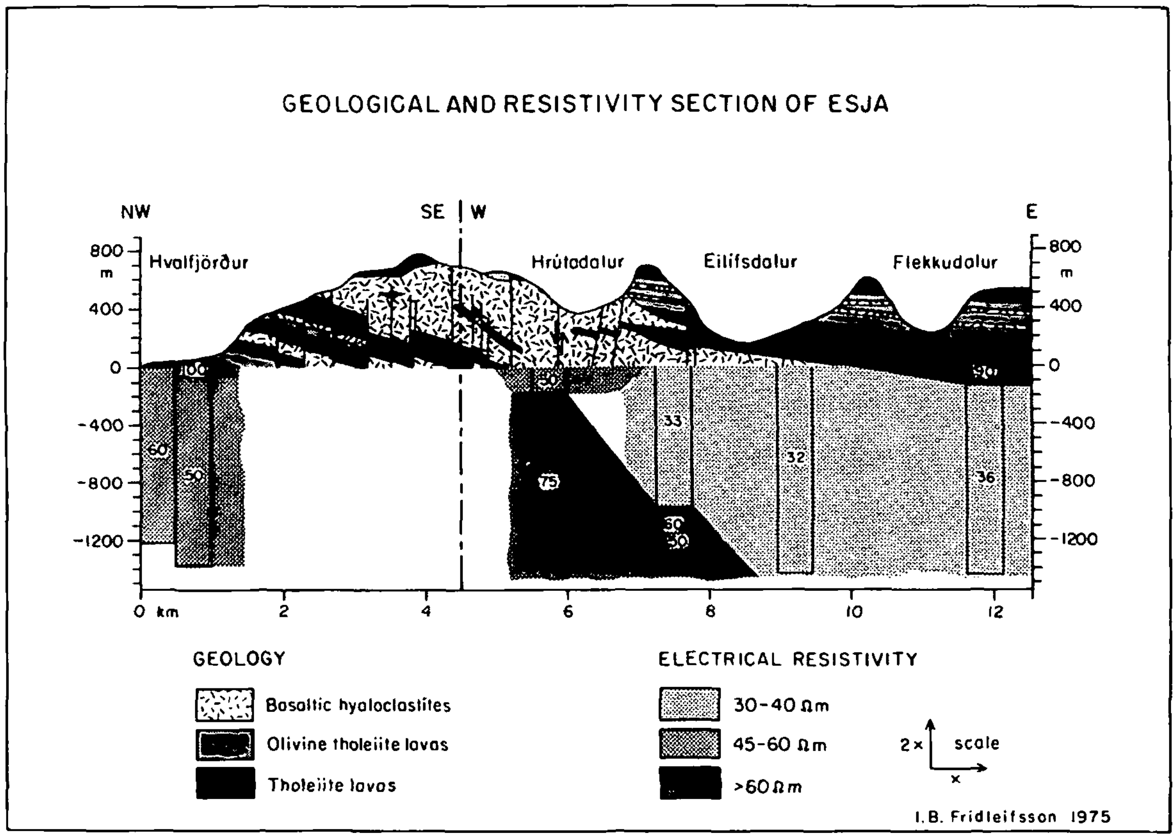

Figure 1. A combined geological (above sea level) and resistivity (below sea level) section of early Quaternary strata. The location of the section is shown in Figure 2. Prominent high-porosity hyaloclastite bodies can be traced to depths of > 1 km by resistivity soundings.

The average porosity of subglacial volcanics is approximately twice that of subaerial lavas (Fridleifsson, 1975). The hyaloclastite ridges can be looked on as high-porosity channels separated by relatively low-porosity lavas in the Quaternary strata. These channels are "thin" and "narrow" where the rate of volcanism has been low during any particular glaciation, but "broad" and "thick" where the extrusion rate has been higher than average.

The Reykir and Reykjavik thermal fields are in Quaternary rocks ranging in age from about 2.8 to 1.8 m.y. There are signs of ten glaciations in the volcanic succession. During this time span there were two central volcanoes active in the region—the Kjalarnes (which is older) and the Stardalur central volcanoes (Fridleifsson, 1973). The rate of volcanic eruption was much higher in the central volcanoes than in other parts of the volcanic zone of the time. This resulted in exceptionally thick accumulations of hyaloclastites in the vicinity of the volcanoes. The volcanoes were further characterized by an abundance of shallow level dykes and sheets; the latter range in thickness from less than a meter to several hundred meters. The emplacement of the intrusions in the strata has probably produced secondary permeability in the strata (Fridleifsson, 1975). The intrusions gave rise to high temperature fields during the life span of the volcanoes (about 0.6 and 0.3 m.y. respectively for the Kjalarnes and Stardalur volcanoes). Now the volcanoes are deeply eroded; the intrusions can be inspected on the surface, and the core regions of the volcanoes are marked by positive gravity anomalies (Fig. 2) which reflect the intensity of intrusions in the strata. The Reykjavik thermal fields are situated on the southern margin of the Kjalarnes central volcano, and the Reykir field is between the central volcanoes, but closer to the southwestern margin of the Stardalur volcano. The low-temperature thermal fields are thus superimposed on the margins of extinct, eroded high-temperature fields.

The ratio of hyaloclastites to subaerial lavas in the strata is variable both within and between the thermal fields; it is lowest in the Seltjarnarnes field, which is in the oldest rocks, but highest in the Reykir field, where in 29 drillholes, 800 to 2043 m deep, the volume percentage of hyaloclastites ranges from 30 to 60%. Table 1 shows the occurrence of aquifers in the different rock types in these holes. Considering that in a 2 km deep hole there may be approximately 1000 m of lavas, 900 m of hyaloclastites, and 100 m of intrusions, but perhaps only 40 to 50 narrow contacts (aggregate thickness of the order of 100 m) between lavas and hyaloclastites, the chances of aquifers occurring in lavas alone or hyaloclastites alone are perhaps tenfold to those of contacts between the formations. It is thus apparent that aquifers are by far most likely to occur at contacts—the higher number of contacts between lavas and hyaloclastites, the higher number of aquifers.

As an extension to the geological investigations, the resistivity of the bedrocks has been studied by numerous direct-current resistivity soundings. A conventional Schlumberger electrode configuration has been used, and the depth of the soundings is 1000 to 1500 m depending on local circumstances.

In an exploration of low temperature thermal activity,

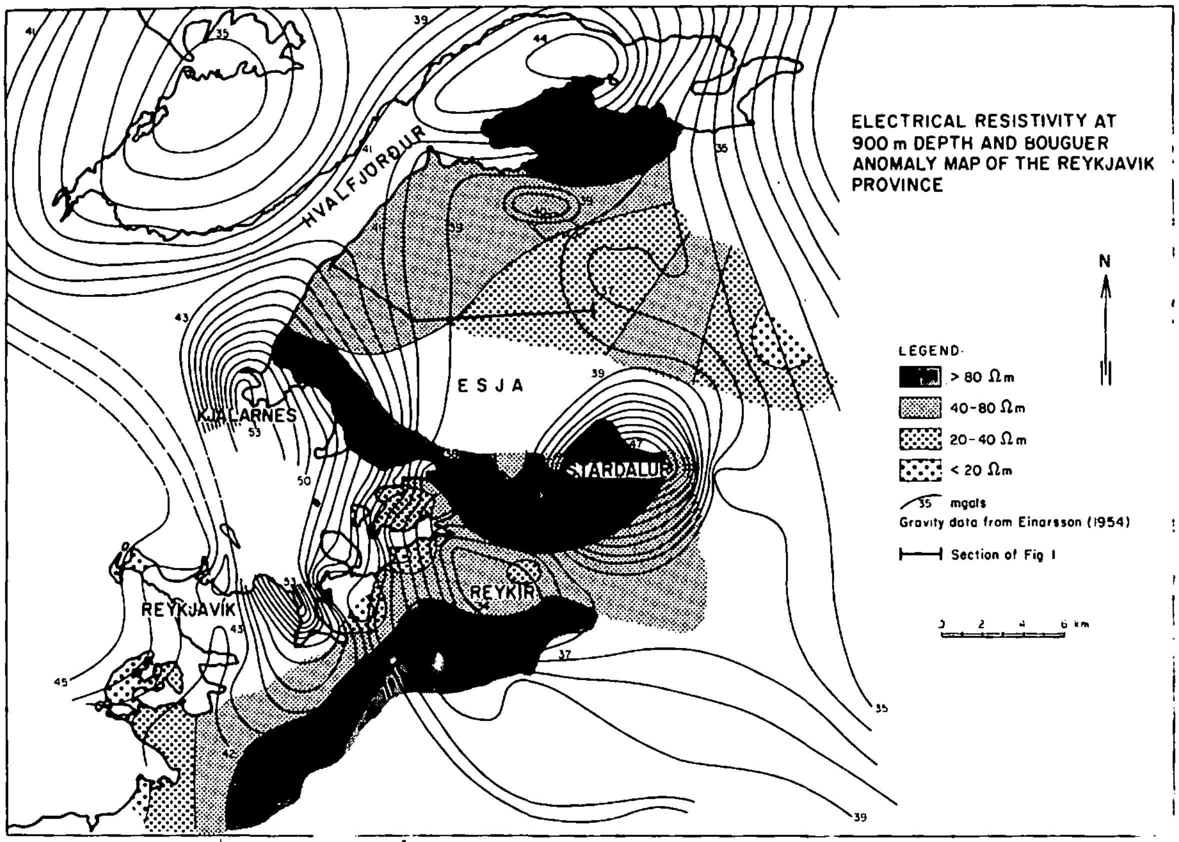

Figure 2. Map of the true resistivity at 900 m below sea level. Prominent high-porosity hyaloclastite bodies can be traced at depths of several km along the strike of the rocks by resistivity soundings. The low-resistivity areas in the upper part of the map are thought to outline two high-porosity "channels" in the strata. Note the coincidence of high resistivity and positive Bouguer anomalies. The exploited low-temperature fields discussed in the text are in low-resistivity (high-porosity) rocks on the outskirts of the central volcanoes, which are marked by positive Bouguer anomalies.

resistivity measurements are largely a structural method. The porosity of the rocks has a very significant influence on the measured resistivity. An informative example of how the resistivity method can be used as a direct extension of surface geological investigations is shown in Figure 1, where the boundaries between high-porosity hyaloclastites and relatively low-porosity tholeiite lavas are traced down to 1200 m below sea level.

About 70 resistivity soundings have been made in the area under discussion. On the basis of these measurements resistivity maps at various depths have been made. The true resistivity at 900 m depth is shown in Figure 2, together with the Bouguer anomalies in the area. The gravity data are from Einarsson (1954). A general northeast-southwest structure can be seen in the low-resistivity areas in Figure 2, which is in agreement with the trend of the hyaloclastite ridges discussed previously.

The higher resistivity values are predominantly found in association with shallow-level intrusions of low porosity. These intrusions are mainly associated with the Kjalarnes and Stardalur central volcanoes as previously mentioned, but also with the Hvalfjordur central volcano. The triple correlation—shallow-level intrusions, positive gravity anomaly, and high resistivity—is rather good.

Table 1. Occurrence of aquifers in the different rock types of 29 drill holes.

Rock type	Aquifers ≦ 2 l/sec	Aquifers 2–20 l/sec	Aquifers >20 l/sec	Total number
Lavas	44	27	2	73
Hyaloclastites*	29	12	4	45
Dolerites		1	1	2
Lavas and hyaloclastites*	53	38	20	111
Lavas and dolerites	13	1	3	17
Hyaloclastites* and dolerites	5	2	1	8

*Included in this group are reworked hyaloclastites and detrital beds.

HYDROTHERMAL SYSTEMS

The thermal gradients (as measured in drillholes deeper than 90 m) to the west of the volcanic zone in southwestern Iceland increase fairly regularly towards the zone from about 70°C/km in Tertiary rocks 100 km west of the zone to about 165°C/km in early Quaternary rocks some 20 km west of the volcanic zone. Assuming thermal conductivity as the only form of heat transport (Palmason, 1973), the thermal gradient should continue increasing towards the volcanic zone axis. But due to water circulation in the Quaternary strata which become increasingly permeable towards the volcanic zone, a trend opposite to that of the regional gradient is found.

The Reykjavik thermal areas lie within or just outside the city boundaries, but the Reykir thermal area is some 15 km northeast of the city center and slightly closer to the active volcanic zone. About 150 drill holes 100 to 600 m deep and 69 drill holes 800 to 2200 m deep have been sunk and about 1400 liters per second (l/sec) of hot water are now pumped from these areas for domestic heating in Reykjavik and its neighborhood.

The surface thermal gradients measured in shallow drill holes in Reykjavik and vicinity are shown in Figure 3. Four areas of thermal maxima are apparent from the isothermal lines in Figure 3, that is, the Alftanes, Seltjarnarnes, Laugarnes, and Ellidaar areas. Only the latter three have been exploited, and hydrological (Thorsteinsson and Eliasson, 1970), thermal, chemical, and isotopic data (Arnason and Tomasson, 1970) indicate that these areas constitute separate hydrothermal systems.

The high surface thermal gradients inside the thermal areas are due to localized transport of water from the thermal systems at depth to the surface. This is best demonstrated in the Laugarnes area, where the highest surface gradients are measured. Prior to exploitation about 10 l/sec of 88°C water issued in free flow from thermal springs in that area, whereas only minor natural thermal activity was found in the other areas in Reykjavik. There is very little or no transport of water from depth in the rocks between the thermal areas, and the depth of the gradient drill holes (at least down to several hundred meters) has little influence on the measured gradients (Fig. 4) outside the thermal areas. The surface gradient of 0°C/km shown southeast of the thermal areas in Figure 3 is due to cold ground water penetrating young volcanic rocks. This cold ground-water zone has been found to reach down to 750 m (measured in a hole 986 m deep) in the volcanic zone 11 km south of the Ellidaar area (Palmason, 1967).

In Figure 4, the estimated temperature of the bedrock is shown for the four geothermal fields (Reykjavik and Reykir). These temperature curves are found from temperature measurements in closed holes in thermal equilibrium and from measurements made on the bottom of holes during drilling. From the bedrock temperature curves and from the concentration of deuterium in the thermal water, it has been concluded (Arnason and Tomasson, 1970) that the Seltjarnarnes and Laugarnes fields each consist of single hydrological systems. However, in the Ellidaar field, the scattered values of deuterium concentration were found to depend on mixing of water from two hydrological systems. The reverse temperature gradient found in the Ellidaar field (Fig. 4) supports the proposed existence of the two systems.

A schematic cross section through the Reykjavik thermal fields perpendicular to the strike of the rocks is shown in Figure 5. The three thermal systems are separated by impermeable barriers (Thorsteinsson and Eliasson, 1970), which we suggest are swarms of dykes and associated faults. The volume percentage of intrusions in drillholes in the Ellidaar area increases towards the hydrological barrier that separates it from the Laugarnes area. It is assumed that the hydrological barriers reach down to layer 3, which probably consists mostly of impermeable intrusions and forms a base to water circulation in the crust.

In the Reykir field there is also evidence for two types of hydrological systems. In the eastern part of the field (nearest to the volcanic zone) a reverse temperature gradient

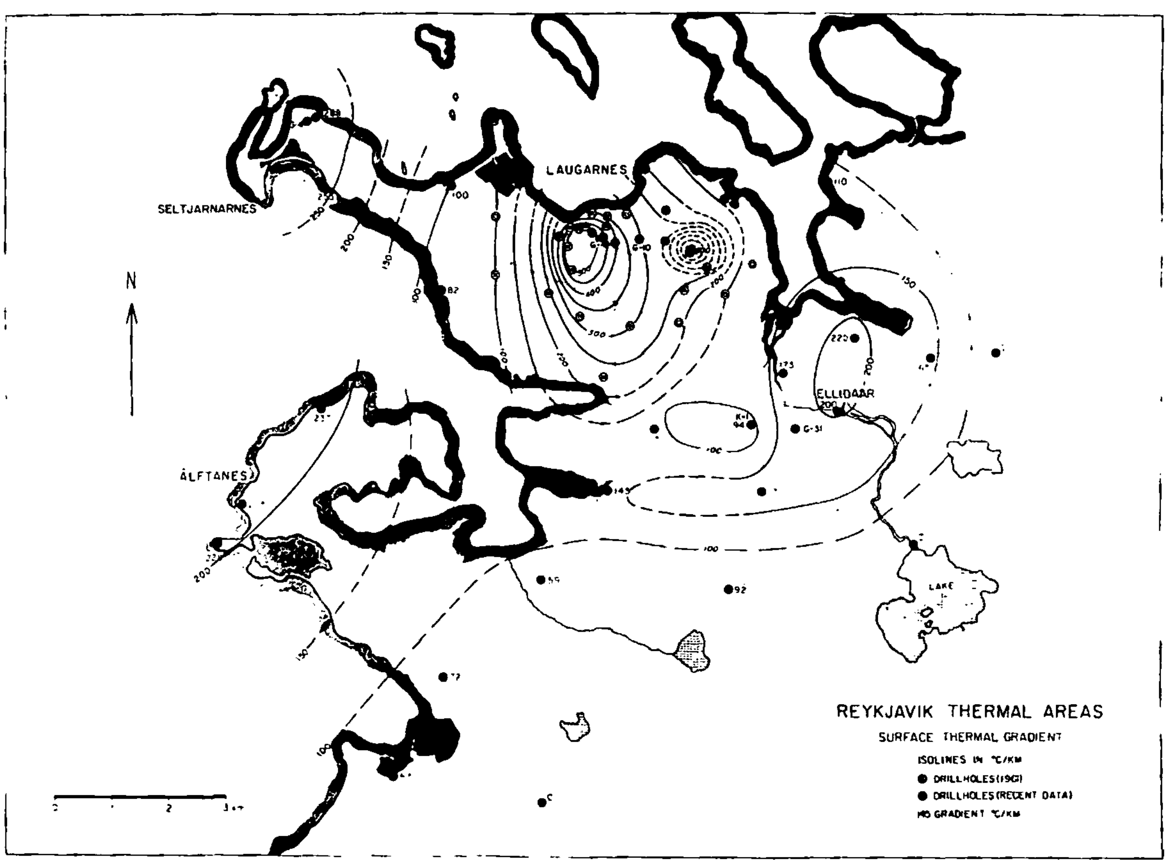

Figure 3. Map of the surface thermal gradients in Reykjavik and vicinity.

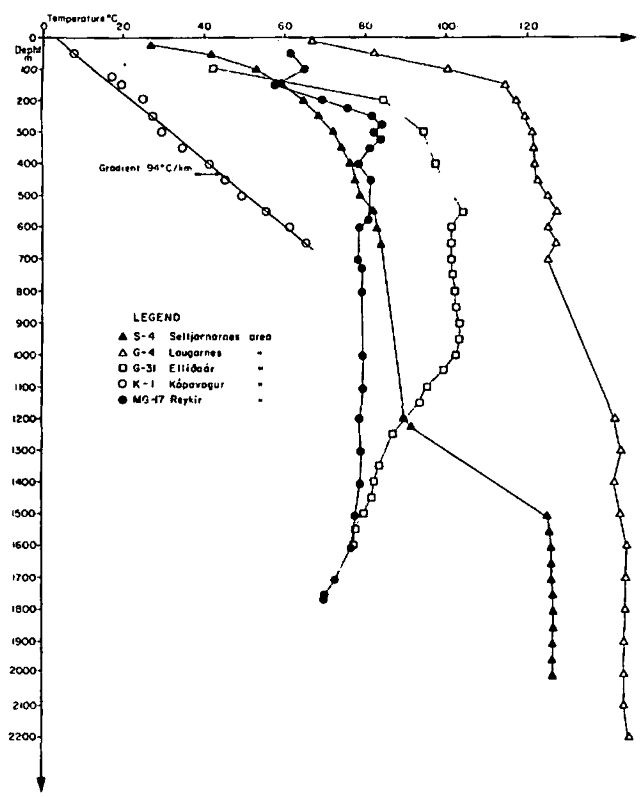

Figure 4. Estimated rock temperature in four thermal fields. The location of the holes, except MG 17 (Reykir), is shown in Figure 3. The drill hole K–1 is between the thermal fields, and shows the regional thermal gradient.

is observed (MG 17 in Fig. 4). The concentration of deuterium in the water from this well is $\delta = -60.0$ ‰, but in a drill hole 1000 m deep (not reaching the deeper, cooler system) situated only 20 m from MG 17, the δ– value is found to be -62.4 ‰ (Arnason, 1975, private commun.). From mixing assumptions it is found most likely that the δ– value of the deeper system is -58 ‰, which is the value for local precipitation.

Outside the thermal fields the thermal gradient is about 100°C/km (see K-1 in Fig. 4). The reverse temperature gradients found in the Ellidaar and Reykir fields can only be accounted for by the circulation of cold water at depth. This cooling effect might be similar to the surface cooling effect observed southeast of Reykjavik (Fig. 3).

HYDROLOGICAL MODEL

The distribution of deuterium concentrations in thermal water in southwestern Iceland has been treated by Arnason and Sigurgeirsson (1967) and by Arnason and Tomasson (1970). A comparison of the deuterium content of thermal water with the distribution of deuterium in the precipitation in Iceland indicates that most of the thermal water originates from precipitation which falls in the interior highlands of the country (Arnason and Sigurgeirsson, 1967), and has been heated by descending to great depth (Einarsson, 1942).

The distribution of the deuterium content in the precipitation in southwestern Iceland (Arnason, et al., 1969) is shown in Figure 6 along with the boundaries between Tertiary, Quaternary, and Recent rocks.

The concentration of deuterium in the thermal water

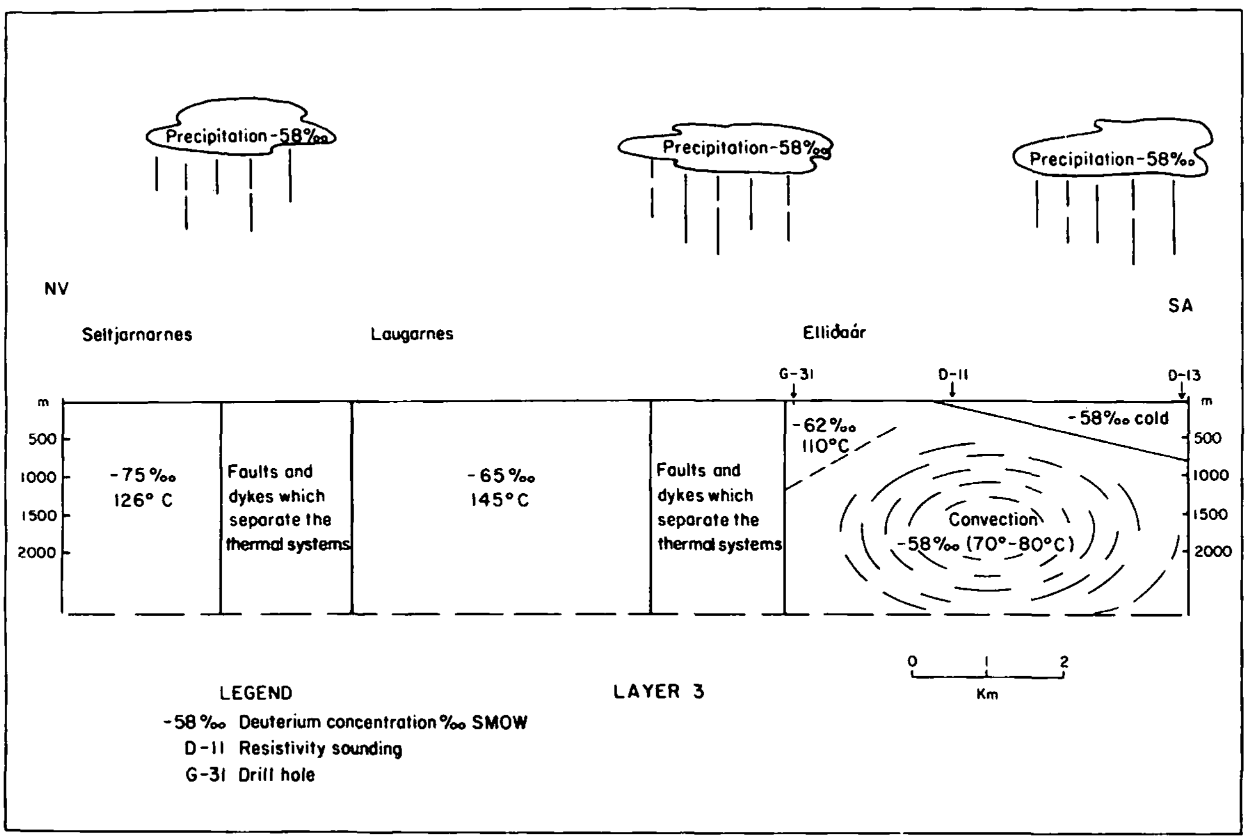

Figure 5. A schematic cross section through the Reykjavik thermal fields showing the isotopic composition and temperature of the thermal water in the fields.

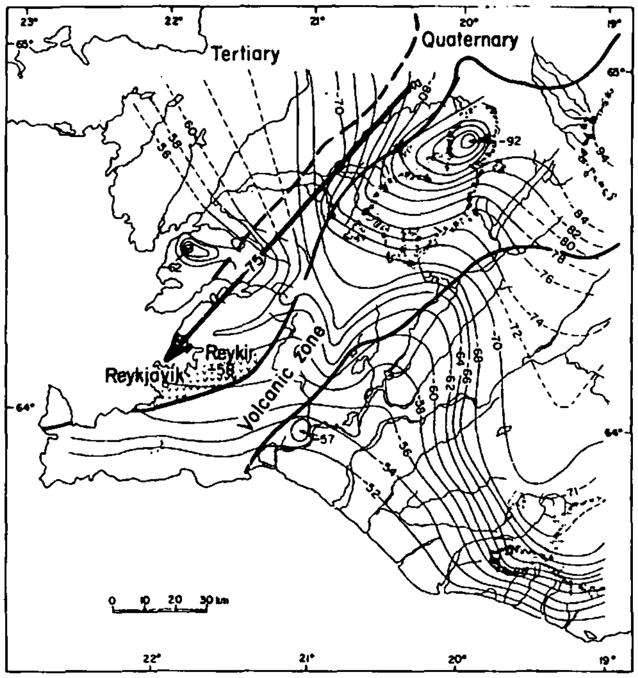

Figure 6. Map of the concentration of deuterium in the precipitation (from Arnason, et al., 1969), and the boundaries of the active volcanic zone, Quaternary and Tertiary rocks in southwestern Iceland. The arrow indicates the direction of flow at depth of low-deuterium water parallel to the volcanic zone, but the shaded area indicates a local connection of −58 ‰ warm water perpendicular to the volcanic zone.

increases towards the volcanic zone. A δ− value −75 ‰ is found in the Seltjarnarnes field, −65 ‰ in Laugarnes field, and higher than −64 ‰ in the Ellidaar and Reykir fields (Arnason and Tomasson, 1970; Arnason, 1975, private commun.).

In the hydrological model proposed here, the thermal water is expected to flow parallel to the volcanic zone along the structural channels discussed previously. A possible path for the −75 ‰ deuterium water is indicated in Figure 6 by an arrow.

Nearest to the volcanic zone (in rocks of relatively high permeability), a certain movement of fresh water, δ = −58 ‰, perpendicular to the zone is allowed for in the model. This high-deuterium water occurs in the thermal fields nearest to the volcanic zone (Ellidaar and Reykir) and is found at greater depth than the "channel water". Thermal water from drill holes penetrating the two water systems should show variations in the δ− values, as is observed in the Ellidaar and Reykir fields.

The thermal properties of the bedrocks, discussed in the previous section, are also in agreement with the hydrological model proposed. The high-deuterium water coming directly from the volcanic zone is much colder than the low-deuterium water, which has flowed some 50 or 100 km underground. The inverse temperature pattern of the rock shown in Figure 4 strongly supports the proposed hydrological model.

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WARD

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A RESISTIVITY SURVEY ON THE PLATE BOUNDARIES IN THE WESTERN REYKJANES PENINSULA, ICELAND

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ABSTRACT

A resistivity survey in the western Reykjanes Peninsula consisted of about 200 Schlumberger geoelectric soundings. It revealed a continous east-west striking low resistivity zone which coincides fairly closely with the Reykjanes seismic zone that has been interpreted as plate boundaries. This low resistivity zone includes at least two but probably all three of the high temperature geothermal areas in the region. Study of correlation between resistivity, salinity and temperature seem to confirm that the low resistivity can only be explained by anomalously high heat flow. This indicates that exploitable geothermal energy may be found in the uppermost 1-2 km along the plate boundaries on the Reykjanes Peninsula even far away from surface manifestations; this may be of great economic significance in the future.

INTRODUCTION

Geophysical investigations on the western part of the Reykjanes Peninsula have been conducted for the past 12 years as a part of the exploration of the three high temperature geothermal areas in region. These geothermal areas are: Svartsengi, Eldvorp and Reykjanes (fig.1).

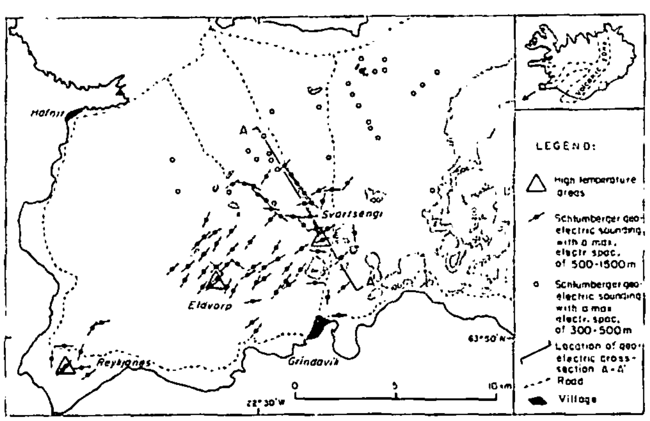

Figure 1. Location of geothermal areas and geoelectric soundings on the western Reykjanes Peninsula.

In Svartsengi a geothermal plant is under construction, with a present capacity of 125 MW_t for a district heating system and 8 MW_e. The base temperature of the field is 240°C and the salinity of the geothermal fluid is approx. 2/3 that of seawater. Of 10 wells drilled in the area, 8 are now productive. The flashed steam is used to heat up fresh water in heat exchangers (Thorhallsson 1979).

At Reykjanes a pilot plant has been in operation for the past three years, producing fishery and commercial salt (NaCl) from the geothermal brine. Commercial production of approx. 8000 tons/year is planned in the immediate future. The base temperature of the area is probably approx. 300°C and the salinity of the brine is similar to that of seawater. Of 8 wells drilled in the area 1 is now productive (Bjornsson et al.1970).

Eldvorp is a small area of fumaroles, but no drilling has yet been done in the area.

It is the purpose of this summary to present an integrated picture of geophysical investigations in the area with the main emphasis on resistivity surveys.

GEOLOGICAL BACKGROUND

The constructive boundaries between the North-American and the Eurasian plate follow the Mid-Atlantic ridge and cross Iceland from south-west to north-east. In Iceland the boundaries are characterized by zones of volcanic and seismic activity. These volcanic zones can be divided into distinct units consisting of a fissure swarm and/or a central volcano. Associated with these are high temperature geothermal areas which are often located near the center of the swarm (Saemundsson 1978).

The Reykjanes Peninsula is the landward extension of the Mid-Atlantic-Ridge. The westernmost part of it is covered by recent basaltic lavas and occasional low hyaloclastite ridges formed in subglacial eruptions. It is an area of freequent volcanic activity, even though it has been dormant for the past few centuries.

The Reykjanes Peninsula is the most active seismic zone in Iceland. The epicenters of the earthquakes in the past ten years group into a narrow zone, 2-4 km wide, along the peninsula striking N70°E. This zone has been interpreted as the plate boundary (Klein et al.]977). Holocene eruption sites on the peninsula have been grouped into 4-5 distinct swarms which are arranged en echelon along the plate boundary; the main direction of faults and eruption sites is N40-50°E(Saemundsson 1978, Jakobsson et al. 1978) The present study deals with the greater part of the westernmost swarms: the Reykjanes-swarm and the Grindavik-swarm. The boundary between them seems though somewhat diffuse.

GEOHYDROLOGICAL CONDITIONS

The geohydrological conditions on the western Reykjanes Peninsula must be considered rather unusual. Because of the very high permeability of the bedrock, the seawater percolates easily through the peninsula. This infiltration of seawater is the reason for the salinity of the geothermal fluid

A lens of fresh water floats on the seawater like ice on water. Despite high infiltration (800 mm/year) and no surface run-off the groundwater-level is only at 1-1.5 m above sea level which means that the thickness of the lens is approx. 50 m. The freshwater lens has been the subject of intense investigations in the past few years, as the geothermal plant in Svartsengi needs 300-400 l/s of fresh water and no water is found at the surface in the vicinity of the plant (Sigurdsson et al. 1978).

THE RESISTIVITY SURVEY

Despite the low relief of the western Reykjanes Peninsula (see fig.1), the lava terrain is extremely rough and only accessible on foot. It is mostly bare lava with litterally no soil and little vegetation.

Approximately 200 Schlumberger geoelectric soundings have been performed in the area. Of this total about 2/3 are shallow soundings specially aimed at measuring the thickness and resistivity of the freshwater lens in the area (Sigurdsson et al.1978). The maximum electrode spacing (AB/2) of these soundings is usually 300-500 m and most of them give only a vague indication of the resistivity of the rocks below the freshwaterlens, if any at all. The remaining third of the soundings has maximum electrode spacing of 700-1500 m and the effective probing depth is of the same order. The location of the soundings is shown on fig.1.

INTERPRETATION

As might be suspected from the geohydrological conditions the resistivity layering in the area is unusually horizontal. Resistivity changes within each layer have been found to be quite subtle, so a one-dimensional interpretation of each sounding gave a reliable picture of the true resistivity. The close connection between the resistivity layers and the geohydrological layers is summarized in the following table:

TABLE 1: The correlation between resistivity layers and geohydrological formations.

Layer no.	Geohydrological formation	Resistivity Ωm	Thickness m
1	"Dry" lava	5000-25000	20-80
2	Freshwater lens	300-3000	40-60
3	Rock penetrated by cold seawater	6-15	
4	Rock penetrated by hot seawater	2-5	

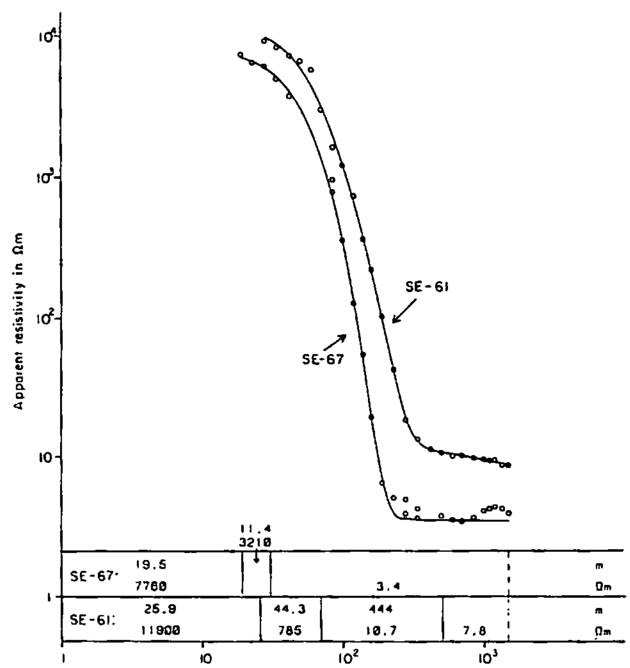

Figure 2. Typical Schlumberger soundings; SE-67 is located within the Svartsengi geothermal field,but SE-61 is unaffected by geothermal activity.

Typical soundings, as shown in fig.2, contain three layers because layer 3 and 4 (table 1) are only found together on the borders of low resistivity areas. In the center of the geothermal areas even layer 2 disappears as the upflow of the geothermal brine disturbs the balance between the freshwater and the seawater.

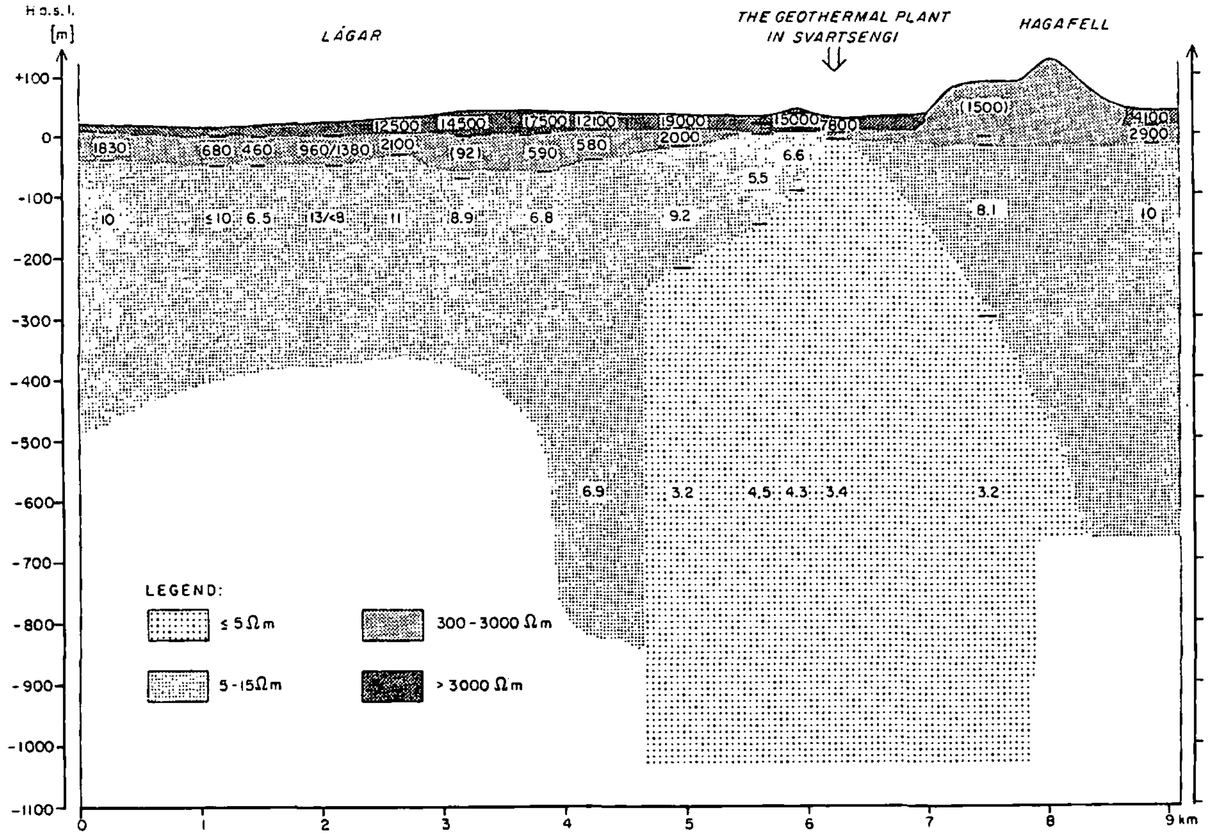

Figure 4. Resistivity cross section A-A'; location is shown on fig.1.

The computer program used for interpretation is the one described by Johansen (]977) and does an inverse interpretation on a starting model, which includes number of layers and some guessed values of each parameter. Minimization is done by an iterative least squares procedure with singular value decomposition.

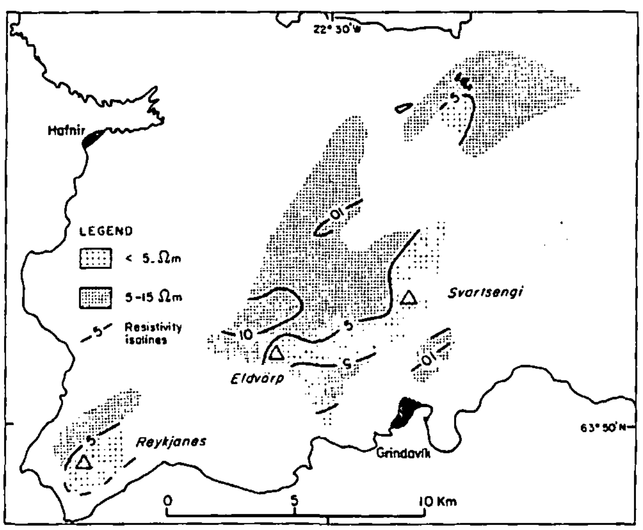

Figure 3. Resistivity at 400 m below sea level.

RESULTS

Figure 3 shows the true resistivity of the bedrock at a depth of 400 m below sea level. The figure shows a east-west striking low resistivity zone($\rho<5\Omega m$) which includes the geothermal areas of Eldvorp and Svartsengi. One sounding indicates that this zone continues further eastwards. Another low resistivity anomaly is seen in connection with the Reykjanes geothermal field (Bjornsson et al.1970). It seems probable that this is a westwards continuation of the west-east striking zone but more data is needed to confirm that. Finally a low resistivity area, which may be a potential geothermal field, emerges on the northern side of the peninsula. No geothermal activity is seen there on the surface.

The NW-SE resistivity cross-section on fig. 4 cuts the low resistivity zone near the Svartsengi geothermal plant and shows the low resistivity anomaly related to the Svartsengi geothermal field. It also reveals the horizontal layering of the area.

DISCUSSION

In fig.5 the main points in the geophysics and the geology of the area are integrated. The seismic zone defines the plate boundaries and in a narrow sense it also defines the volcanic zone in the area. There the

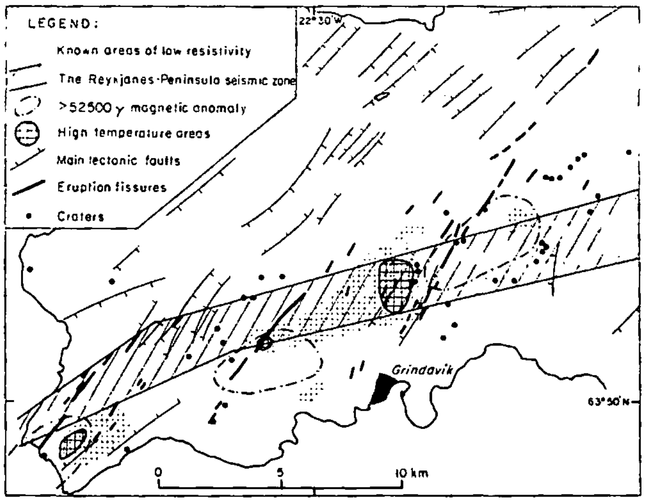

Figure 5. Geophysical and geological features of the western Reykjanes Peninsula;seismic zone based on Klein et al.(1977) and an unpublished map by S.Bjornsson; magnetic anomalies based on Sigurgeirsson(1970); tectonic features and eruptive centers based on Jakobsson et al.(1978).

rocks are most extensively fractured but more important there the heat flow can be expected to be anomalously high.

Resistivity studies based on resistivity measurements on groundwater samples (Karlsdottir 1975), Archie's law and resistivity -temperature relations have shown that salinity and temperature are the only factors besides porosity that affect the resistivity in the area. By making the assumption that porosity is the same inside and outside the geothermal systems (which is supported by geological conditions of the area) these relations have also shown that the resistivity contrast in the Svartsengi (Reykjanes) area (from 3.5Ωm (2.5Ωm) inside the geothermal field to 10Ωm outside) corresponds very well to the measured drop of temperature in drillholes from 240°C within the geothermal system to 40°C outside (at 400 m depth). Similarily it can be shown that the 5Ωm isoline that is used to indicate the low resistivity areas in fig.5 corresponds to a temperature of approx. 160°C near the Svartsengi geothermal area.

The study on temperature-resistivity correlations seems to confirm the hypothesis that the low resistivity zone in the Reykjanes Peninsula must be correlated to anomalously high heat flow. This poses the question whether exploitable geothermal energy may be found everywhere within the low resistivity zone, i.e. whether convective geothermal systems are to be found in the uppermost 1-2 km along the plate boundaries on the Reykjanes Peninsula.

It is worth pointing out that the geothermal areas are in all cases connected with some major geological features fracture zones and/or the youngest major volcanic eruption fissures in the area. Surface geothermal activity might benefit from this. Each geothermal area is also located within or near a >52500 γ magnetic anomaly. These anomalies may indicate centers of intrusive activity, which might be the heat sources of the convective geothermal systems.

So far the resistivity survey indicates a positive answear to the question posed. the real answear though, which is of great economic importance, must wait for the results of an exploratory drillhole located within the low resistivity zone but outside known geothermal areas. The first steps towards that drillhole will hopefully be taken in 1982 with the realisation af extensive plans of geophysidal research covering the whole of the western Reykjanes peninsula.

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