GEOLOGICAL MAPPING IN GEOTHERMAL EXPLORATION WITH SPECIAL REFERENCE TO TEPHROCHRONOLOGY AND PALEOMAGNETIC TECHNIQUES

Roberto Maciel Flores[+],
UNU Geothermal Training Programme,
National Energy Authority,
Grensásvegur 9, 108 Reykjavík,
Iceland

[+] Permanent address:
Comisión Federal de Electricidad,
Oficina de Exploraciones Geotermicas,
Alejandro Volta 655,
Colonia Electricistas,
Morelia Mich.,
Mexico

Nr. 4

ABSTRACT

A description is given of methods applied in geological field mapping and a ground magnetic survey carried out in the low temperature area of Brautartunga, where a fluxgate magnetometer was used to map the paleomagnetic stratigraphy. Field work and interpretation of a geophysical survey in the same area using a proton magnetometer is presented as an example of how this technique is used in geothermal exploration to find the relation between hot springs and faults or dykes in places where the basement is partly covered or completely hidden by soil.

A small tephrochronological study carried out in the Snaefellsnes peninsula is described . Three acidic tephra layers were identified all being from the Snaefellsjökull central volcano. Their extent and grainsize were studied and the thickness plotted on isopach maps.

Results of previous studies of three central volcanoes in Iceland visited by the author are summarized to compare two deeply dissected central volcanoes and their extinct high temperature areas with the Krafla active central volcano,where a geothermal power station has been built inside the Krafla caldera. The Krafla volcano has been volcanically active since 1975. The two extinct central volcanoes, Hafnarfjall-Skardsheidi and Reykjadalur, are located in the western part of Iceland. They are deeply eroded making it possible to look at the geology of the high temperature areas at depth and their internal structures.

A brief account is given of the different techniques used in volcanic surveillance in Iceland. It serves to illustrate how active volcanoes and their associated geothermal fields can be monitored.

LIST OF CONTENTS

Page

ABSTRACT

5 VOLCANIC SURVEILLANCE

LIST OF FIGURES

LIST OF TABLE

1 INTRODUCTION

1.1 Scope of work

The author of this report was awarded a UNU Fellowship to attend the 1981 UNU Geothermal Training Programme held at the National Energy Authority in Iceland. The Training Programme is supported by the United Nations University, and the government of Iceland. The training started with an introductory lecture course lasting for about 4 weeks on various aspects of geothermal science such as geothermal energy around the world, geology, geophysics, geohydrology, geochemistry, borehole geology, borehole geophysics, reservoir engineering, utilization and economics.

In the author's view this lecture course provides a general background knowledge on most aspects of geothermal energy, and during the two weeks geothermal UNU field excursion (29 June-9 July) an opportunity was realized to visit most of the geothermal fields in Iceland (low- and high-temperature) which included lectures and seminars on each respective area.

After the four weeks of lectures the author received a specialized training in geology. This included theoretical and practical aspects of tephrachrology (3 weeks), a study of tephra layers in Snaefellsnes region (2 weeks), field mapping in Lundarreykjadalur, which in addition to stratigraphical mapping involved an exercise in using fluxgate magnetometer and proton magnetometer (2 weeks), field work in active and extinct high-temperature areas (2 weeks) and volcanic surveillance (3 days).

1.2 Geological setting and geothermal fields of Iceland

Iceland is the biggest landmass astride the Mid-Atlantic Ridge and consists nearly entirly of volcanic rocks with small portion of intercalated sediments also of volcanic origin. The recent volcanicm is mainly limited to an axial rift zone which stretches (in branches) across the island from southwest to northeast.

The oldest dated rocks in Iceland are found in the extreme northwest and east and their age is 14-16 m.y. (Moorbath et al., 1968). These rocks were originally erupted in an axial rift zone in the central part of the country but drifted away from it as time passed. Thus the rocks in general become younger towards the active rift zone.

The rocks are predominantly basaltic, of the tholeiite suite, but about 10% of the rocks are intermediate to acidic in composition. Sediments make up only about 5% on the average of the strata. In Fig. 1 a geological map of Iceland is presented, compiled by K. Saemundsson of NEA (1979).

The active rift zone consists of en echelon fault and fissure swarms with a central volcano sitting in the centre of each swarm (Saemundsson, 1980). The swarms are inclined at a samll angle to the trend of the zone. These same structures have been observed in the older part of the island. The ruins of central volcanoes have been identified with dyke and fault swarm running through the centres. The present fault and fissure swarms have been interpreted as surface manifestations of the dyke and fault swarms of the older and dissected areas. Presently about 80 central volcanoes have been identified and about 20 of these have been active in postglacial time (last 10.000 years) or may be regarded as dormant.

A small volcanic zone stretches from WNW to ESE along the Snaefellsnes peninsula but it can not be classified as a rifting zone. The composition of the volcanics related to this zone is transitional to alkalic in contrast with the tholeiites of the main rift zone.

Most of the extrusives in Iceland are regular lava flows but during glaciations hyaloclastites and pillow lavas are formed instead. Submarine eruptions produce similar products.

The predominant types of volcanoes in the volcanic zones are the following:

a) Lava shields are formed in one but long monogenetic eruption. Subglacial eruptions may eventually lead to table mountain formations which are predominantly built up of pillow lava, hyaloclastite with a lava cap at the top.

b) Monogenetic eruptive fissures are characterized by eruption of fluid basaltic magma, which may flow over distances of tens or even over a hundred kilometers. The length of one individual fissure may reach 120 km.

c) Central volcanoes (both of strato type and shield volcanoes). The eruption of intermediate to acid magma is restricted to these volcanoes. Usually a caldera subsidence is developed in the central part of the volcano, often with a diameter of 10 km. High temperature fields develop in the central part of the volcanoes. These volcanoes have a life-span of 0.5 - 1.5 m.y. and in deeply dissected volcanoes major intrusions are exposed.

In general the volcanic activity in historical times (i.e. last 1100 years) has been more or less continuous and there is a volcanic eruption on the average every fifth year.

As mentioned above the Mid-Atlantic Ridge which is the plate boundary between the North-American plate and the Eurasian plate, runs across Iceland. Heat flow is anomally high along the plate boundary. The geothermal gradient in Iceland varies from 50° to 160°C/km and is highest close to the axial rift zone, but still higher gradients are reported from high temperature areas inside the active central volcanoes (Saemundsson and Fridleifsson, 1980) The geothermal fields in Iceland are divided into two main categories or groups:

a) High temperature fields, which are always inside an active or dormant central volcano. Presently about 20 such fields have been identified. A geothermal field falling into this group has to have a subsurface temperature of 200°C or more in the topmost 1000 m. They are predominantly characterized by large areas of hot ground, fumaroles, mud pools and high alteration state of the rocks (propylitization).

There are supposed to be two types of hydrothermal systems within high temperature fields in Iceland:

i) Vapor dominated systems
ii) Water dominated systems

b) Low temperature fields. There are several hundred geothermal fields which fall into this group. The geothermal fields in this group have a subsurface temperature of less than 150°C in the topmost 1000 m. These occur as warm and hot springs and in some cases as geysers. The water discharge is variable but the maximum discharge is 180 l/s of 100°C hot water (Deildartunga hot spring). The hydrothermal alteration is usually insignificant. The aquifers are believed to be along faults and dykes or stratification boundaries. The low temperature fields are now extensively exploited. About 80% of houses in Iceland are space heated by hot water, and it is used also in green houses, fish hatching etc.

The high temperature areas are all inside the active volcanic zone but the low temperature areas are confined to areas outside the zone.

2 FIELD MAPPING IN LUNDARREYKJADALUR

2.1 Introduction

The low temperature field in the Borgarfjordur region is the largest of its kind in Iceland. The natural discharge in this area is between 300 and 400 l/s of boiling water. It is situated in the Tertiary region where the aquifers are believed to be mainly confined to narrow zones along dykes or faults.

Brautartunga, the area of study, is demarcated by lines of latitude 64°30′30" and 64°33′30" North and lines of longitude 21°25′ and 21°15′45" West (Fig. 1).

The aim of this field work was to do an exercise in using portable fluxgate magnetometer and proton magnetometer, mapping of the main geological features, like faults and dykes and the identification of zeolite zones.

The research was necessary to be able to decide where to drill for hot water.

2.1.1 Description of methods

In this area 18 sections were measured along steep slopes or gullies where a continuous succession of lava flows could be observed without much difficulties.

The thickenss of the units was established both from altimeter readings on steep escarpements and from direct measurement at the outcrop, considering the dip of lava flows. It is difficult to assess the error limits of the thickness measurements. Check measurements indicate that the thicknesses assigned tend to be a little low rather than too high. An error of as much as 5 percent may thus be present.

Mapping involved also the definition of lava types, measurement of the thickness of individual lava flows and sedimentary beds, measurements of the paleomagnetic polarity of each individual flow, measurements of the tectonic dip and mapping of dykes and faults.

2.2 Paleomagnetism

All rocks exhibit magnetic properties, one of which is the tendency of a rock specimen to behave as a magnet, with a north and south pole and a magnetic axis. This magnetization is a naturally occuring property and is referred to as fossil magnetism or natural remanent magnetism (Irving, 1964).

The great majority of observations of paleomagnetism have been published since 1930 and particularly since 1950. In the early part of the twentieth century Bernard Brunhes, a French physicist, accidentally discovered that the remanent magnetism of a lava flow in Central France was exactly opposite to the magnetic field; that is, the magnetic poles were reversed, but the phenomenon remained unexplained. Later (1926) a Japanese geophysicist, M. Matuyama found that many of the Pleistocene volcano rocks were reversely magnetized. In this way the longer epochs of constant polarity have been named after the several geophysicists who made contributions to this study; for example the Brunhes normal epoch extended from the present to about 0.7 m.y. while the Matuyama reversed epoch extends from 0.7 to 2.3 m.y.

Types of magnetization (Lapedes, 1978)

i) Primary magnetization

There are three types of remanent magnetization in igneous or sedimentary rocks.

a) Thermoremanent magnetization (T.R.M.)

The permanence of natural remanent magnetization of igneous rocks is due

During the present field work a portable fluxgate magnetometer (Segulmaelir FT 74-5,Fjarskiptataekni) was used to measure the magnetic polarity of each lava flow. Three to four samples were taken from each individual layer to identify the magnetic polarity. Effort was made to avoid proximity to dykes because the dykes may have reheated the lava flows and changed their original polarity.

2.2.3 Rock types

The rocks in the Lundarreykjadalur valley, are predominantly basaltic flows, but subglacial rocks such as basaltic hyaloclastites and pillow lavas occur close to this area. Sedimentary rocks constitute less than 5% of the total thickness.

In this report the author uses the classification of Walker (1959) for olivine-basalt and tholeiite. The classification is based on the field appearance of the flows. Following are their main characteristics.

Tholeiite

- Very fine grained
- Weathered crust grey to pale brown. Spheroidal weathering uncommon
- Amygdales empty or bearing celadonite, chalcedony, quartz and chlorophaite, usually without zeolites
- Pipe-amygdales and basalt pegmatites rare
- Well-developed flow structure within the body of the rock

Olivine-basalt

- Much coarses in grain
- Weathered crust brown to black. Spheroidal weathering common
- Amygdales bear zeolite
- Olivine sometimes visible in the hand-specimen
- Pipe-amygdales and basalt-pegmatites common
- Flow structure confined to a streaking out of amygdale

The basalts with more than about 10 per cent phenocrysts (mainly plagioclase) are referred to in this paper as porphyritic basalts.

to the magnetization of different minerals[+] during cooling in the geomagnetic field from the temperatures at which they solidify, which are usually above the Curie points[++] of their magnetic minerals.

b) Depositional remanent magnetization (D.R.M.)

In sediments the magnetization may arise through the orientation of the detrital iron oxide grain by the geomagnetic field during deposition in water.

c) Chemical remanent magnetization (C.R.M.)

This form of magnetization may arise during chemical changes, as in the growth of grains of hematite in the red sandstone, during or perhaps slightly later than deposition.

ii) Secondary magnetization

However, the original direction of magnetization in a rock sample is often mastered by secondary or unstable components of magnetization which grow in the rock after its formation.

a) Viscous remanent magnetization (V.R.M.) is the most common secondary component of this form of magnetization. In most rocks there are portions of the magnetic minerals which are very easily realigned so that, if the direction of the Earth's magnetic field changes, these components realign themselves with the changing field.

[+] The magnetic properties of rocks result from iron oxide minerals present to a few percent in many rocks. Two groups are especially important; the magnetite (Fe_3O_4) - ulvospinel (Fe_2TiO_4) solid solution series and the hematite (Fe_2O_3) - ilmenite ($FeTiO_3$) solid solution series. Intergrowths between magnetic minerals occur frequently.

[++] French scientist Pierre Curie in 1895 discovered that the magnetic substances loose their magnetism on heating above a certain temperature appropriately named the Curie point. No mineral has a Curie point higher than 800°C, for example magnetite has a Curie point of 578°C.

This process leads to the masking of the original direction of magnetization in the rock.

b) Isothermal remanent magnetization (I.R.M.)

This type of secondary magnetization is a very important way in which the original direction of magnetization is altered when the outcrop from which the specimen is obtained is struck by lightning.

2.2.1 Magneto-geological mapping in Iceland

Einarsson and Sigurgeirsson (1955) measured the polarity of basaltic lava flows by the means of an ordinary field compass.

Specimens of a convenient size were removed from the rock and its magnetic polarity tested by moving it close to the compass. In most cases the permanent magnetization of basalts was strong enough to give an easily detected deflection of the magnetic needle. If the north magnetic pole of the specimen points downwards when it is placed in the original position, the specimen is normally magnetized; if it points upwards it is inversely magnetized (Einarsson and Sigurgeirsson, 1955).

This method was later used to map the magnetic stratigraphy of the basalt lava pile and the magnetic zonation of the pile was established (Sigurgeirsson and Einarsson, 1955). Later these magnetic zones were correlated with the magnetic stripes on the ocean floor (Piper, 1971). In the early sixties fluxgate magnetometers were designed , being far more sensitive than the compass.

In the last two decades the K-Ar-dating method has been applied to the basaltic lava flows to date the various magnetic events. The combination of K-Ar dating, the use of the portable fluxgate magnetometer in the field and the analyses in the laboratory of samples taken in the field, made it possible to determine the paleomagnetic polarity in different areas with accuracy and extend the geomagnetic polarity time scale to 6.5 m.y. (McDougall et al., 1977).

There is no clearcut boundary between the two types and some flows show characteristics of both and are thus classified as transitional basalts.

Very coarse grained and olivine rich lava flows which usually comprise many flow units are probably flows from ancient lava shields.

2.3 Geological mapping

2.3.1 The lava succession

As mentioned earlier 18 sections were measured. The total thickness of the investigated lava pile is close to 600 m. A representative section across the pile is presented in Fig. 2. The tholeiite flows are most common in the lower part of the section but olivine basalt and pophyritic flows are more common in the central part but the tholeiites become again more common in the uppermost part.

A simplified geological map is shown in Fig. 3.

Usually very thin sedimentary beds intercalate the lava flows; They are usually only few cm thick and bright red in colour. In the uppermost part of the section at least two distinct conglomerate beds are present which can easily be traced for long distances and these aided in the correlation of the lava sequence across faults.

The paleopolarity of the flows is shown in Fig. 2. The lowest part of the section is reversely magnetized proceeding into a zone of anomalous polarity but the upper part consists of normally magnetized flows.

By comparing these results to the work of Piper (1971) (Fig.4) and McDougall et al (1977) (Fig.5), the polarity change observed in the measured section is believed to represent the Gilbert/Gauss boundary which is close to 3.3 m.y. old.

The strike of the lava flows is close to NE-SW and dip about 8° towards SE.

2.3.2 Dykes

During field work all dykes were grouped with respect to composition. The polarity was measured with a portable fluxgate magnetometer. The knowledge of the polarity of each dyke helps to interprete the ground magnetic data.

In all 17 dykes were found in the research area (Fig. 6). The thickness and direction of each dyke was measured in the field. The width of the dykes varied from 0.5 to 5.0 m. The predominant trend of the dykes is NE-SW i.e. roughly parallel to the strike of the lava flows. The dykes are subvertical and usually cut the lava flows at approximately right angles.

All are basaltic in composition and constitute only a small fraction of the total exposed rock volume.

2.3.3 Faults

A number of faults were identified in the field (Fig. 6). These are normal faults usually with a throw of 10-25 m but the extremes are 5 to 80 metres.

Most of the faults, which in general strike NE-SW, have a throw to the west and the plane of these faults tend to be at right angles to the stratification (i.e. dipping 82° NW). The faults with the opposite throw (i.e. on the east side) are close to vertical.

As seen later there is a conspicuous connection between a particular fault and a hot spring.

2.3.4 Alteration state of the succession

Walker (1960) designated the uppermost zeolite zone as the chabazite-thomsonite zone, which passes down into the analcite zone at a depth of about 600 m from the top of the lava succession. The analcite zones passes successively at greater depths into the mesolite-scolecite and further down into the laumotite zone.

The thermal metamorphism is part of the regional alteration which takes place below the water table by reaction of water with the basaltic flows to produce secondary minerals in response to the rather high geothermal gradient observed in Iceland, which are commonly in the order of 50° to 160°C per km (Jóhannesson, 1977).

In Lundarreykjadalur valley, the topmost zeolite-free zone is absent, has probably been removed by glacial erosion. The only zeolites identified were chabazite and thomsonite suggesting that the lavas belong to the chabazite-thomsonite zone.

According to McDougall et al. (1977) lavas within the chabazite-thomosonite zone have been metamorphozed at a temperature of less than 140°C. The zonal pattern in Lundarreykjadalur valley is probably similar to the one presented by McDougall et al. (1977) (see Fig. 7).

2.4 Hot springs

In the research area there are a few hot and warm springs which will be described in this section. The springs are classfied as warm springs if the temperature is in the range of 10-50°C, but as hot springs if 50°C to 100°C. The location of the springs is shown in Fig. 6.

2.4.1 Brautartunga hot spring

The main hot spring in the area is the Brautartunga hot spring. The maximum temperature is 97°C and the discharge is about 2 l/s. The water springs from a gravel terrace.

According to local information a small hot spring was found west of the community house but it was later buried by gravel.

The hot water of the Brautartunga spring is presently used for space heating of the community house, farm house and for the swimming pool but formerly also for greenhouses.

2.4.2 Snartarstadir hot spring

On the south bank of river Grímsá about 1400 m north-east of Snartarstadir farm is a hot spring. The temperature is 43-51°C and the water seeps through a postglacial gravel terrace. The hot springs are seen in a 30 m long stretch along the river bank.

Higher up on the slope on the south side of the valley are a few warm springs which seem to be aligned roughly N-S:

a. Lowest on the slope and nearest to the river is a quaking bog about 2-3 m in diameter. The maximum temperature is 22°C but discharge is negligible. It has been dug out and is about 2 m deep.

b. A quaking bog about 6-7 m in diameter. Maximum temperature is 18°C but discharge is negligible.

c. A quaking bog about 10 m diameter with two eyes of ascent, one with a temperature of 12°C and the other of 18°C.

d. A quaking bog about 3-4 in diameter. Maximum temperature was recorded 10,5°C.

e. A small pool with a temperature of 13°C and some discharge.

f. A quaking bog with two eyes of ascent. The temperatures measured were 14°C and 13°C respectively.

g. A warm spring with temperature 9-12°C. It is about 6-8 m in diam. and is used for drinking water for the farm Holl.

In the gully north of the farm Gullberastadir is a warm spring with a temperature of 11°C.

The hot springs in the Lundarreykjadalur valley can easily be related to faults which can be traced across the valley.

At Brautartunga the ground magnetic survey shows clearly the relation between a fault and the hot springs (see later) and by the field observations the hot and warm springs of Snartarstadir could be related to a normal fault with a throw of about 20 m (see Fig. 6).

A few chemical analyses of hot water from this area are available (Table I). The base temperature is about 130-140°C in Brautartunga but about 90°C at Snartarstadir.

Place	Date	Measurements of temp. °C	Silica temp.
Brautartunga	1944	91.0	135
- " -	15/11/44	86.0	116
- " -	1969	92.0	(64)
- " -	19/04/79	86.5	132
- " -	30/6/81	96.5	
Snartarstadir	30/06/81	43-51	92
	- " -	22	
	- " -	18	
	- " -	12 - 18	
	- " -	10.5	
	- " -	13	
	- " -	13 - 14	
	- " -	9 - 12	
Gullberastadir	- " -	11	21

Table 1. Temperature of thermal waters in Lundarreykjadalur valley. Partly from Gunnlaugsson, 1980).

2.5 Ground magnetic survey

Ground map surveys can be successfully used to reveal the finer structural details of geothermal areas, especially were the bedrock is covered by sediments.

For this reason this method was used to try to find the connection between the hot springs and faults or dykes as the bedrock near the hot springs in Brautartunga is covered by unconsolidated postglacial sediments.

The geology of the Brautartunga area was mapped prior to the ground magnetic survey, but due to the sediments, the relation between the hot springs and faults or dykes was not clear.

The instrument used was a portable proton magnetometer model Geometrics. For the location and topographic control we used an enlarged aerial photograph in the scale 1:5000 and a map in the scale 1:50.000.

The magnetic survey covered an area of 0.25 km^2, in all 25 profiles. Of these 21 were 400 m long but 4 were 300 m long. The interval between measuring points in each profile was 5 m.

The distance between the 21 first profiles was 20 m but between the last four profiles was either 50 or 100 m (Fig. 8).

The geomagnetic contour map is shown in Fig. 8 . From the shape and characteristics of each anomaly it is possible to estblish a fairly good model for the basement underlying the sediments. The interpretation is shown in Fig. 9 . There are two dykes just east of the Brautartunga farm. The trend of these dykes is close to NE-SW. The eastermost dyke has a reverse polarity but the other has a normal polarity.

Furthermore, there is evidence from the magnetic survey that the normally magnetized dyke transects the reversely magnetized dyke (Fig. 9).

At least two faults can be identified from the geomagnetic map. One is a N-S trending fault about 100 m west of the farm Brautartunga. The other is a NE-SW trending fault which seems to run just where the hot spring is at the surface. With reference to the geological mapping these faults are probably normal faults. It seems quite obvious, that the hot spring is related to the latter one.

2.6 Siting of drillhole

The surface investigations including geological mapping, geophysical survey and chemical analysis of the hot water in the Brautartunga area. has led to the following conclusion:

a. According to the geological and geophysical survey it seems likely that the hot springs and the aquifer at depth, are connected by a subvertical fault (see Fig. 9).

b. The temperature of the Brautartunga hot spring has varied from 86° to 96.5°C, probably due to changes in the ground water level i.e. infiltration of cold groundwater. The chemical analyses suggest a temperature of 130-140°C at depth.

c. The most feasible location of a drillhole is about 50 m NW of the Brautartunga hot springs (see Fig. 10). The aim is to intersect the north-westerly dipping fault (82°) at 300-400 m depth.

3 TEPHROCHRONOLOGY OF THE SNAEFELLSJOKULL CENTRAL VOLCANO

3.1 Methods and usefulness in geothermal exploration

3.1.1 General aspect

Studies of tephra layers as a tool for various research began in Iceland and other countries around 1930. In Iceland these studies led to a doctoral thesis: Tefrokronologiska studier pa Island (Thorarinsson, 1944). He introduced the terms tephra and tephrochronology, and uses the word tephra as a collective term for all airborne pyroclasts, and the word tephrochronology as the dating method based on identification, correlation and dating of tephra layers. Thorarinsson (1944) also points out the importance of tephrochronology as a tool in volcanological, pollenanalytical, glaciological, geomorphological and archaeological research.

According to Thorarinsson (1979) tephrochronology has been applied in Iceland in different fields such as:

I Studies of fluvial and wind erosion
II Dating of glacier oscillations
III Dating of ice cores from glaciers
IV Studies of periglacial phenomena, especially frost crack polygons
V Archaeological studies, especially dating of farm ruins
VI Pollenanalytical studies of vegetation changes
VII Establishing a tephrochronological connection between Iceland and other countries
VIII Cryopedological research
IX Establishing the volcanic history of active central volcanoes and to estimate volcanic hazards (i.e.intensity of volcanic activity (see chapter 5)
X Study of eruptive characteristics of volcanoes
XI Dating of lava flows

3.1.2 Field methods

Where the tephra layers have not been stripped off by wind or water, but covered by subsequent soil formation, they appear in more or less distinct horizons in the soil profiles.

To identify tephra layers in the field it is necessary to note their colour (which usually reflects the chemical composition), thickness and grain size along with physical characteristics of the grains and stratigrafic relation.

3.1.3 Laboratory work

The samples of tephra layers in the laboratory can be identified by the determination of their mineralogical, chemical and physical characteristics the microprobe analysis of the glasses being the most useful method. Various granulometric parameters can reveal the type of explosive activity (Shaw et al., 1974).

Tephra layers can often be dated by various means, such as by C^{14}-dating of peat soil immediately above and beneath them or trees or shrubs embedded in them (Steinthorsson, 1967). They can also be dated by extrapolation from other exactly dated tephra layers which show the rate of sedimentation, by dendrochronology, pollenanalysis and with the aid of archaeology (Thorarinsson, 1979).

3.1.4 The preparation of samples

To appreciate the grain size characteristics, the tephra samples should be washed to separate organic and soil material. In this process, beaker or plastic glass should be used to put the sample in and finally cold water to wash the mixture. After the washing and decanting several times the sample is dried. After drying, the different tests can be done.

There are some problems, however, which have to be considerd in the interpretation of tephra layers. Eaton (1964) listed several common causes of modification of tephra deposits:

1) Erratic directional variation of high altitude winds.
2) Redistribution of particles by erosion and masswasting.
3) Fluid redistribution of tephra (transportation and reworking in oceans and lakes).
4) Stratigraphic complexities that prevent corration between various sample points.
5) Pronounced physical and/or chemical alteration.

Several other, more subtle variations also lead to (often unrecognized) effects which modify tephra deposition. Local thunderstorms commonly develop near an eruption. Rain from these storms may scavenge particles from the atmospheric tephra (cloud), thus producing anomalous size distribution. A certain fraction of all eruptions must occur coincidentally with periods of general rain as a weather disturbance passes through (Know et al., 1964).

These problems demonstrate the need for more thorough, systematic studies of tephra in terms of mineralogical, textural, chemical and size characteristics in addition to geographical distribution and geometrical configuration.

3.2 An exercise in tracing the phreatric activity of the Snaefellsjokull volcano in postglacial time

On the westernmost tip of the Snaefellsnes peninsula is a 1446 m high stratovolcano which has been active in postglacial times. Numerous basaltic to trachytic lava flows have erupted during that time, either from the summit crater or from small craters on the lower flanks or on the surrounding lowlands. In the Snaefellsnes peninsula three light coloured acidic tephra layers have been identified in postglacial soil sections (Steinthorsson, 1967, Sigurdsson, 1967). The aim of this work was to establish the phreatic activity of the Snaefellsjokull volcano and to trace the tephra layers.

A total of 95 soil profiles were studied in the region and their location is shown on Fig. 11. The thickness of individual tephra and soil layers were measured and the characteristics of the tephra layers studied. In Fig. 11 are shown some representative profiles. As mentioned above three light coloured tephra layers have been identified in the region and they are confirmed by this study. The layers are referred to as Sn-1, Sn-2 and Sn-3, counted from top to bottom. They all seem to have a common origin; Snaefellsjokull volcano. Isopach maps of the distribution of the three layers, respectively are presented in Figs. 12-14. In some cases the tephra layers have been transported by water or been severly affected by wind. These cases were omitted.

Sn-1 is the uppermost layer. It is light-coloured acidic tephra, generally at 7-37 cm depth below the present surface. Its maximum thickness is 21 cm close to Olafsvik northeast of the volcano. The axis of maximum thickness stretches northeast from the volcano (Fig. 13). The maximum grain size is about 6 cm close to the volcano and the average size is close to 1-2 cm. The layer consists mainly of frothy acidic pumice but pitchstone fragments are common and reach the same size as the pumice. Sn-2 is a light-coloured acid tephra and is usually found at 14-120 cm depth. Its maximum thickness is 23 cm northeast of the volcano and the axis of maximum thickness stretches northeast from the volcano (Fig. 13). The grain size is 1-2 cm close to the volcano but decreases to less than 0.2 cm further away. Most of the fragments are made of frothy pumice but glassy fragments are also common. Sn-3 is the lowest layer found. It is usually close to the bottom of the soil cover which is underlain by gravel, morainic material or basement rocks. The maximum thickness of this layer is about 30.5 cm northeast of the volcano but its distribution pattern is strange, as can be seen from Fig. 14. The westernmost part of the layer seems to be missing for some reason or another. This layer consists of frothy acidic pumice like the others and the grain size is 1-2 cm closest to the volcano but drops to 0.2 cm furthest away.

A few basaltic tephra layers were found but they could usually be traced to cinder cones in the close neighbourhood of the sections that they were found in. For example a basaltic tephra layer from the Berserkjahraunskulur craters is found in many sections west of the craters. The age of the upper two light coloured layers has been dated by the means of C^{14}-method (Steinthorsson, 1967). The layer Sn-1 turned out to be 1750±150 years old but Sn-2 about 3960±100 years. The bottom layer must be 8000-10000 years old or so because it is close to the base of the soil cover which is assumed to have started to form about 10000 years ago.

It is clear from this research that three major phreatic eruptions have occured in the Snaefellsjokull volcano in the last 10000 years and the relative dating of the lava flows should be relatively easy by using the present knowledge of the age relations in the area. It would be useful to date the layer Sn-3.

4 GEOLOGY OF HIGH TEMPERATURE FIELDS IN ICELAND, ACTIVE AND EXTINCT

4.1 An active high temperature field: Krafla

4.1.1 General geology

The structure of the active rift zone in northern Iceland is dominated by large fault and fissure swarms (Fig. 15). Most of the swarms pass through different central volcanoes where fissure eruptions, differentiated rocks and high temperature geothermal fields are concentrated. Two of the central volcanoes in this region have developed calderas; Krafla and Askja. The Krafla volcano and fissure swarm is reviewed by Bjornsson et al. (1977), and the following description is based on that paper but the main places of interest were visited. The Krafla fissure swarm is about 100 km long and has a maximum width of 10 km. The Krafla central volcano forms a low broad shield about 25 km in diameter and in the central part of the volcano is a caldera. It measures about 10 km east-west and about 8 km north-south (Fig. 16). It formed during the last interglacial period and has since been filled almost to the rim with volcanic material. The caldera's collapse almost certainly followed the eruption of a sheet of dacitic welded tuff, which is exposed around the caldera and is an airfall tuff, blown mainly towards the north-east from a source near the centre of the caldera. A shield volcano with a diameter of at least 20 km existed before the caldera formation, and its remnants enclose the caldera on the east and west sides, exposing lavas and breccias dipping outward at low angles.

A high temperature geothermal field lies within the caldera (Fig. 16). Drilling has revealed temperatures> 340°C at 2 km depth. Basaltic intrusions become increasingly common below 1.2 km depth and granophyre intrusions are found in some wells near 2 km depth.

About three km south of the Krafla calder is another high temperature geothermal field, the Namafjall field and drilling to a depth of 1.8 km has revealed temperatures of over 290°C. The latter area is in the same fissure swarm as runs through the Krafla volcano.

The composition of the eruptives of the Krafla volcano ranges from olivine tholeiites to rhyolites. The basalts of the fissure swarm through Namafjall and Krafla area are usually quartz normative tholeiites.

Basaltic fissure volcanism is concentrated mainly within the Krafla caldera and near Namafjall. In postglacial time about 20 eruptions have occurred in the Krafla caldera and its nearest surroundings and about 15 in the Namafjall area. Most of the fissure eruptions are basaltic, but andesite and dacite flows have also been erupted in both areas. Voluminous silicic eruptions have not occurred in post-glacial time but four subglacial silicic eruptions within and around the Krafla caldera have produced large domes or ridges during the last glacial period (Hlidarfjall and Jorundur). The Krafla caldera contains explosion craters that have ejected small quantities of rhyolitic pumice. The most recent such crater was formed in 1724 at the beginning of the 1724-29 volcanic episode (Myvatn fires). Tephrochronological studies indicate that postglacial volcanism (i.e.less than 10000 years) in the Namafjall and Krafla areas occurred in two main periods; (1) in early postglacial time and (2) during the last 3000 years. The period of repose may have lasted more than 4000 yrs. The Krafla and Namafjall part of the fissure swarm have erupted 6-7 times each during the latter period of activity. Volcanism near Krafla has been concentrated along the central part of the fissure swarm during this period; earlier it was dispersed over the entire caldera. The largest volume of basaltic lava produced in a single eruption in the area comes from the Þrengslaborgir-Lúdentsborgir crater row on the Namafjall part of the swarm about 2000 yrs ago. Its total volume has been estimated as 2-3 km^3 and the areal extent is about 220 km^2. Few flows of the swarm, however, exceed 0,2 km^3.

4.1.2 Historic volcanic activity in Krafla

Only two major volcanic events have occurrred in the Krafla fissure swarm in historical time. The first in the years 1724-1729. It started with the formation of the explosion crater Viti which ejected both silicic pumice and basaltic scoria along with mud and rock debris. Ithas been suggested that the formation of the Víti explosion crater may have been caused by magma injection into the hydrothermal system. In the following years a repeated activity is recorded, many earthquakes and opening of ground fissures. New steam and mud craters were formed in the Krafla and Namafjall geothermal fields, but little or no volcanic activity was noticed. In 1727 a fissure eruption started in Leirhnukur near the centre of the Krafla caldera. This eruption continued with lapses until late 1729. The lava flows covered about 35 km^2 but the fissure is about 11 km long and discontinuous. A small eruption was reported on this fissure in 1729.

The second volcanic event started in December 1975. It followed an increased seismic activity earlier that year culminating with an earthquake swarm and a small eruption. There have been 6 volcanic eruptions until now. The activity has been characterized by inflation and deflation periods. These periods have been described by Bjornsson et al. (1979). The inflation periods are characterized by:

1. Continuous and nearly constant uplift of the Krafla region. The maximum uplift is near the centre of the caldera, 7-10 mm/day decreasing outwards 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.

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

The inflation is interrupted by a sudden subsidence/deflation which usually lasts few hours or days. It is characterized by:

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 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 during some deflation.

The inflation is interpreted as being caused by magma transported from below and accumulating in a magma chamber at a shallow depth within the Krafla caldera. The deflation on the other hand is caused by release of pressure in the magma chamber by widening of the rifts and flow of magma into the fault swarm north of the caldera. This flow is accompanied by volcanic tremors, faulting and thermal activity in the fault swarm. The movement of the magma has been mapped with the help of migrating earthquakes (see Fig. 17). In this model the earthquakes in the caldera are caused by inelastic deformation of the crust above the magma chamber.

4.1.3 Geothermal field

In the Krafla central volcano and the associated fissure swarm are two reported high temperature geothermal fields (Stefansson, 1981).

One is the Krafla field proper which is inside the caldera (Fig. 16). The other is the Namafjall field which is in the fissure swarm some 5 km south of the caldera. During the present volcanic activity a third field has been formed north of the caldera (Fig. 16), the Gjastykki field.

The eastern part of the Krafla geothermal field was mapped by use of the Schlumberger D.C. resistivity method. The resistivity measurements show a good correlation between surface alteration and a low resistivity region in the depth range 0-800 m (Fig. 18) but at greater depth the resistivity measurements show somewhat a more complicated picture. Inside the low resistivity region the values seem to increase beneath 800 m. So far 17 wells have been drilled in the field, the first one in the year 1974. One of the wells is used to monitor the ground water level.

From information obtained by drilling, a geological cross-section of the drilling area has been constructed (Fig. 19). The sub-surface rocks can be split into three main lithological units:
(1) Hyaloclastite formation 0-700 m (2) lava formation (mainly altered basalt lavas) in 700-1200 m and (3) intrusive formation (dolerite, granophyre etc.) down to at least 2200 m.

The Krafla geothermal system consists of two separate geothermal zones: The upper zone between 200-1100 m depth is a water dominated system with a temperature range of 195-215°C and the lower zone below 1600 m with temperatures between 300 and 340°C. The latter is considered to be vapour dominated and it has a high content of gas (like CO_2) and a high enthalphy.

There seems to be a barrier between the two zones at about 1100 m depth but in part of the area (near Hveragil) this barrier seems to be absent where there is an upflow channel from the lower zone to the surface.

Building of a geothermal power plant in Krafla started simultaneous to the production drilling in early 1975 and was completed in 1977. The power plant is designed for 60 MW (two turbines each 30 MW). So far only one of the turbines has been installed but due to production difficulies only about 12 MW are presently being generated.

4.2 Reykjadalur central volcano

Reykjadalur central volcano was visited and all the main units were inspected. It is located in the northeast part of the Snaefellsnes peninsula between Borgarfjordur and Dalir districts.

The diameter of the central volcano ranges from 25 to 30 km. The centre itself is represented by a caldera approximately 10 km in diameter, which is best exposed in the Reykjadalur valley. The flanks of the volcano extend, on average, between 10-13 km out from the caldera rim (Jóhannesson 1975).

The extrusive rocks of the volcano can be divided into the following four series:

1. The Reykjadalur thick layered series form the base of the volcano and also part of the cone itself. These series consist mainly of thick tholeiite flows which dip outwards closest to the volcano. The base of the series is about 6.0 m.y. old and the total thickness is 250-300 m.
2. The main phase of differentiated extrusives lies conformably on top of the preceding series. These series comprise most of the intermediate and acid extrusives of the Reykjadalur volcano. The rocks enclose both lava flows and pyroclastic material like ash and unwelded ignimbrites. The thickness varies from 10 to 200 m and two thirds of the rocks are acidic in composition.
3. Unconformably on top of the differentiated rocks is the Reykjadalur thin layered series which is at least 600 m thick. The series

consist of numerous thin tholeiite flows generally 2-4 m thick.

4. Caldera filling and the final differentiated extrusives.

At some time during the formation of the thin-layered series a caldera was formed and this caldera was gradually filled by volcanic and sedimentary rocks, many of which are unconsolidated and later intruded by cone sheets. The following are the main units of the caldera filling (Fig. 20).:

1. As the floor of the caldera subsided screes (talus-breccias) were formed on the sloped of the caldera escarpment.
2. Rhyolite flows from the early stages of the caldera occur on the lower slopes of the Reykjadalur valley.
3. Basaltic pillow lavas, hyloclastites and two-tiered jointed flows intercalate with the breccias of unit (1).
4. A 25 m thick tuffaceous sediment with leaf impressions occurs in Kalfagil in Reykjadalur.
5. A 200 m unit of basaltic subaerial lava flows overlies unit 4.
6. A unit of basic-intermediate lava flows overlies unit (4) in Kalfagil in Reykjadalur.
7. Units (5) and (6) are covered by a sedimentary bed, which has been found in many places inside the caldera and also outside the caldera rim , near Lambahnukur, indicating that no subsidence occurred after the bed was laid down. Plant fossils are abundant east of Kalfagil.
8. A unit of miscellaneous volcanic rocks overlies the sedimentary bed. The lower part of the unit consists of irregularly columnar jointed lava flows and the upper part consists of hyaloclastites and pillow lava.

The hyaloclastites and pillow lavas of unit 8 in the caldera filling seem to be formed under subglacial conditions and when traced across the caldera rim they overlie a tillite horizon. This horizon has been dated 4.3 - 4.4 m.y.

Intrusive rocks

A few large intrusive bodies occur at the present erosional surface with the exception of the Baula complex. In the Reykjadalur central volcano the intrusives are mainly dykes and cone sheets (Fig. 21). Jóhannesson (1975) divided the intrusives into 3 groups.

1. Intermediate and acid dykes. These dykes cut the thick-layered series and are presumably feeders to the main phase of differentiated rocks. These have been extensively penetrated by later cone sheets.
2. Basic to acid cone sheets. These cone sheets tend to follow the trend of the caldera rim, and also tend to form a circular inclined funnel-like structure around the volcano. An individual cone sheet does not cover more than a few degrees of an arc. The sheets dip decreases outwards from the centre of the caldera from 45° to 10°. The intensity distribution of the sheets depends on the host rocks. They intrude more easily ignimbrites and other acidic rocks along with talus-breccias of the caldera than the basaltic lava flows.
3. Late rhyolite cone sheets and intrusives. The main difference between rhyolite and basic cone sheets is that each rhyolite cone sheet can be traced for many kilometers. The sheets have a dip of about 45° towards the centre of the caldera.

Baula and Litla Baula intrusions are about 10 km south of the southern caldera rim. These are the youngest intrusions in this area. The Baula complex is according to the K-Ar dating about 3.4 m.y.old (Jóhannesson, 1975). It consists of columnar-jointed rhyolite (Fig. 22). At the base of Mt. Baula rhyolite dykes cut the intrusion and form concentrically inclined sheets. These usually dip about 45° towards Baula. This intrusion seems to have uplifted the overlying lava flows. This intrusion consists of rhyolite and has intruded at shallow level.

Litla Baula is a volcanic vent filled with rhyolite breccias and with a rhyolite plug in the centre. Traces of the eruptives which seem to be mainly tephra are to be found just north of the vent.

During the life-span of the central volcano (around of 1.5 m.y.if the Baula complex is excluded), a great variety of volcanic rocks were produced, but only a few of them had the characteristics to be able to act as a reservoir rocks. The rocks units which are more likely to have high porosity are mainly hyaloclastites, conglomerates, sedimentary layers, ignimbrites, pillow lavas and talus-breccias and some intrusives such as Baula, which is usually columnar jointed. In the Reykjadalur volcano the most intense hydrothermal alteration has taken place around the caldera rim where the cone sheets are most numerous.

The alteration displays a zonal distribution (Fig. 23). Laumontite forms a zone 2-5 km wide outside the caldera rim but inside it is a zone dominated by epidote. The most intense alteration is in the rocks originally unconsolidated like talus-breccias and hyaloclastites. It seems reasonable to conclude that the paleohydrothermal system in Reykjadalur was confined to these rocks and that the heat was donated by the numerous cone sheets intruding these rocks.

4.3 Hafnarfjall-Skarðsheiði central volcano

4.3.1 Geological succession

The author had the opportunity to visit for four days the Hafnarfjall-Skarðsheiði central volcano, which is located similar to Reykjadalur central volcano in the deeply eroded Tertiary region of western Iceland, and was able there to observe on surface many of the deep seated geological features of a central volcano.

The geology of the central volcano and its surroundings has been mapped by Franzson (1978) and the data here is mainly from his work.

The setting of the central volcano is somewhat anomalous as is summarize below.

The Borgarfjörður regional anticline is a major structural feature in the Tertiary succession of western Iceland. For most of its length, from Hredavatn in the northeast towards Borgarnes in the southwest it is characterized by a gentle curvature, but at its southern end west of Hafnarfjall it splits up into a number of synclines and anticlines (Fig. 25).

The rocks at the crest of the anticline at Borgarnes have been dated with the K-Ar method to be 13.2±2 m.y. (Moorbath et al.,1968) and 9.4±0.7 m.y. (Aronson and Saemundsson, 1975). The succession of the anticlinal crest dips gently and becomes younger towards a synclinal axis on the Snaefellsnes peninsula to the northwest. It is believed that the synclinal axis represent the location of an axial rift zone which became extinct some 6 m.y. ago and that this succession was derived from that rift zone.

Just east of but paralel to the Borgarnes anticline runs a major unconformity. To the north it is called the Hredavatn unconformity, but the Hofn unconformity to the south, named after a locality near the farm of Hofn where it is best exposed.

The succession to the east and above the unconformity has been shown to get progressively younger towards the Reykjanes-Langjökull axial rift zone or from 6-7 m.y. to Recent. This has led to the conclusion that the unconformity represents the division between successions derived from two different axial rift zones, i.e. the one on the Snaefellsnes peninsula which became extinct some 6-7 m.y. ago and another rift zone which became active about 6-7 m.y.ago and which is called the Reykjanes-Langjökull rift zone.

As mentioned earlier a central volcano within an axial rift zone is the locus of maximum intrusive and extrusive activity within each swarm. The position of the Hafnarfjall-Skarðsheiði central volcano at this crustal boundary brings out the differential strain response of these two different crusts, i.e. the relative rigidity and impermeability of the older crust to the west and the young ductile crust of the developing axial rift zone to the east as indicated in Fig. 24.

This may explain some of the features found in the central volcano, such as the flexuring of the basalt succession east of the unconformity, the certain alignment of the larger intrusives paralel to and east of the unformity, the relatively sudden disappearance of the dyke-swarms near to the unconformity etc.

Somewhat anomalous for the Hafnarfjall-Skarðsheiði central volacno (as well as the Reykjadalur one) is the long life span of about 1.5 m.y. (i.e. from about 5.5-4.0 m.y.) but according to Saemundsson (1979) the normal life span of central volcanoes in Iceland is 0.5-1.0 m.y.

The volcanic activity of this central volcano can be divided into 4 main phases (Fig. 25) with respect to age, eruptive centres, composition and structural characteristics.

1. Brekkufjall phase. Characterized by basaltic volcanism followed by voluminous and copious extrusion of differentiated rocks which led to a caldera collapse in Brekkufjall area (about 5 km in diameter).
2. Hafnarfjall phase is characterized by a thick sequence of mainly basaltic composition as well as some rocks of more evolved composition. During this time a major caldera formed in the Hafnarfjall region (see Fig. 27) consisting of three partially superimposed basins. The diameter of the caldera as a whole is about 7x5 km and has a total subsidence about 800-1000 m. The subsidence structure is filled with rock formation divisible mainly on lithological grounds into the following 5 units:
   a) The lowest unit consists of a succession of thin basalt lava flows of maximum thickness is about 200 m.
   b) Overlying the basalt unit lies a hyaloclastite unit, with a maximum thickness of about 300 m.
   c) The lower agglomerate unit consists primarily like the underlying unit, of basaltic hyaloclastite. This unit has a maximum thickness of about 200 m.
   d) The andesite unit is a mixture of lavas and hyaloclastites and has a maximum thickness of about 500 m.
   e) The upper basalt unit has about 100 m maximum thickness and apparently tops the caldera succession.

The pyroclastic units which would be considered to have a relatively high porosity and permeability are believed to have formed due to the interaction between the erupting magma and the water of the caldera lake.

3. Skarðsheiði phase. The succession belonging to this phase consists mostly of basalt and rocks of rhyolite composition. Included in the latter is a very interesting ignimbrite eruption centre.
4. The Heiðarhorn phase. This phase is represented by a succession of basaltic rocks with a subordinate amount of more evolved compositions.

When the central volcano and the associated fissure swarm became extinct about 4 m.y. ago as previously mentioned, another fissure swarm developed to the east of the central volcano, issuing lavas which gradually buried Hafnarfjall-Skarðsheiði central volcano and this succession is called on Fig. 26 the Hvalfjordur lenticular lava unit.

4.3.2 Intrusions

Dykes

A dyke swarm is a deep erosional equivalent of the active fissure swarms of the active rift zones, such as in Krafla. Fig. 26 shows the rose diagram representation of the dyke swarm associated with the Hafnarfjall-Skarðsheiði central volcano.

Cone sheets

Inclined intrusions are common within the core of the central volcano and these are likely to represent cone sheet swarms propagating out from the deeper seated larger intrusives, Fig. 27. These cone sheets seem to have been most actively formed during the caldera formation in the Hafnarfjall area.

Larger intrusives

Fig. 28 shows the locations of surface exposures of large intrusions and also locations of inferred intrusions, which are deduced from data such as converging cone sheets updoming of the roof above the intrusions, and the hydrothermal aureol around these.

4.3.3 Alteration

The rocks within the core of the central volcano show in general a high alteration state, similar to that found in the Reykjadalur central volcano and within the Krafla high temperature area. It is very evident that the heat source of the high temperature areas in Iceland are the high percentage intrusives assembled in the core of these volcanoes.

5 VOLCANIC SURVEILLANCE

5.1 The aim of this surveillance is to estimate the hazard of volcanic activity in thermal areas.

In different places like a geothermal field, situated in an active volcano-tectonic area where lavaflows, earthquakes, ashflows and mud avalanches are frequent and may threaten inhabited places, it seems reasonable to try to reduce the damage to already existing installations and human life and to design future installations with respect to the possible danger.

In order to know what kind of protective measures could be taken in certain place, different studies can be made such as:

a) tephrochronology, could be applied in studies of eruption history and eruption-mechanism of volcanoes (see chapter 3).

b) chemical analyses (if there is an active volcano in the surrounding area) in order to study the amount, composition and temperature of gases to emerge from the volcano.

c) volcanology can help us to know the types of volcanic eruptions and intensity of volcanic cycles.

In Iceland the volcanic prediction is a combination of critical evaluation of historical records, a study of the field evidence mainly, tephrochronology, and the information collected in the monitoring station from tiltmeters and seismometers (see chapter 5.2). A network of seismometers has been established in the last two decades in Iceland. The aim is to monitor the axial rift zones and the fracture zones in northern and southern Iceland. It is hoped that the system may give early warnings of major volcanic and tectonic events (Fig. 29).

In many cases a detection of changes in the level of activity may give indication of abnormal state usually preceding major events.

Recently, samples of ground water, especially hot water, have been sampled regularly and the radon content measured. This method has been used quite successfully in China for earthquakes prediction but so far success has been limited in Iceland (Hauksson and Goddard, 1975).

Sigvaldason (1979) has reviewed the monitoring efforts in Iceland. Three volcanoes are constantly monitored, Katla and Heimaey in southern Iceland and Krafla in northern Iceland.

5.2. Surveillance of the Krafla central volcano

In a place like Krafla where volcanic activity during the last 6 years has been vigorous it is of importance to monitor the volcano in order to predict forthcoming events and thus possibly reducing the risk of important places such as the Reykjahlid village, the Kisildiajn diatomite factory and the Krafla power station being damaged.

For this reason since 1975, three institutions (Orkustofnun, the University of Iceland and the Nordic Volcanological Institute) have been operating tigether a volcanic observatory in the Reynihlid village. In this observatory information from all the instruments installed around Krafla geothermal field are collected in order to monitor the volcanic and tectonic activity of the Krafla central volcano.

The instruments installed around the Krafla field (Tryggvason, 1979), are:

1) Monitoring of fumarole gases installed in different places and with the function of sampling gases to know chemical changes and also the gases emitted from fumaroles are analysed every few weeks.
2) Tiltmeter.
   The tilt measurements are of three types:

a) Water table tiltmeters of two components installed in the Krafla power house.

b) Three electronic continously recording tiltmeters, one in the Krafla power house, the second about one kilometer north of the explosive crater Víti, and the third in Reynihlid by Myvatn.

c) In addition there are twelve spirit level tilt stations in the area.

3) Three seismometers are located within 10 km of Krafla and additional three are within 50 km distance (Brandsdottir and Einarsson, 1978).

All the data from the monitoring stations are collected in the observatory at Reynihlid village, and can be interpreted in order to predict volcanic or tectonic event and subsequent planning to evacuate or to protect the town, the diatomite plant and the power station in the event of an eruption.

As described in 4.1 the activity in Krafla is characterized by inflation and deflation periods. During inflation the tiltmeters record an increased ground elevation towards the centre of the caldera but a rapid decrease during deflation. In Fig.30 is shown a diagram of ground level changes for one station.

The seismicity is low during the inflation period but often increases shortly before deflation starts as well as during the deflation (Fig. 30). Just before deflation the CO_2content of fumaroles increases and so does the pressure in some of the drillholes. This suggests a relation between magma and the hydrothermal system.

ACKNOWLEDGEMENTS

I would first like to thank the organizers of the U.N.U. Geothermal Training Programme in Iceland under the direction of Dr. Ingvar Fridleifsson and Dr. Hjalti Franzson for creating good conditions for my study.

I am much indebted to my supervisor Dr. Haukur Jóhannesson for his close supervision and guidance during the field work in Brautartunga area, Snaefellsnes peninsula and Reykjadalur central volcano, and for his many valuable ideas and for critically reviewing the manuscript.

I would also like to express my sincere thanks to Dr. Hjalti Franzson for revising parts of the manuscript and for his guidance during the field work in Hafnarfjall-Skarðsheiði central volcano, in Krafla central volcano and during the U.N.U. Field Excursion and for his kind cooperation in many ways during the training.

The author is also grateful to Dr. Jon Jonsson for his instructive and fruitful help during the tephrachronology training in the southwestern and southern Iceland and also in the Snaefellsnes peninsula.

Special thanks are due to Ludvik S. Georgsson for his supervision and guidance in the execution and interpretation of the ground magnetic survey carried out in the Brautartunga area and for revising part of the manuscript.

Thanks are also expressed to Dr. Sveinbjorn Bjornsson of the Science Institute of the University of Iceland for his explanation of the monitoring station in Krafla volcano. My thanks are also due to the staff at the National Energy Authority who helped the author during his stay in Iceland, in particular to Ms. Solveig Jonsdottir, secretary of U.N.U., for her kind cooperation in many ways during the training.

I would like to extend my thanks to Auður Ágústsdóttir and Ingunn Sigurðardóttir for expertly drawing the figures, to the library staff who provided reports and information essential for this report and to Mrs. Aðalheiður Jóhannesdóttir who typed the manuscript with infinite patience.

Finally I would like to express my thanks to the inhabitants of the Borgarfjörður region, specially to Mr. Ólafur Jóhannesson and his family for their hospitality.

REFERENCES

Aronson, J.L. and Saemundsson, K., 1975. Relatively old basalts from structurally high areas in central Iceland. *Earth Planet Sci. Lett.*, 28, p. 83-97.

Björnsson, A., Saemundsson, K., Einarsson, P., Tryggvason, E. and Grönvold, K., 1977. Current rifting episodes in North Iceland, *Nature*, 266, p. 318-322.

Björnsson, A., Johnson, G., Sigurdsson, S. and Thorbergsson, G., 1979. Rifting of the Plate Boundary in North Iceland 1975-1978. *Geophys. Res.*, 84, p. 3029-3038.

Brandsdottir, B. and Einarsson, P., 1978. Seismic activity associated with the September 1977 deflation of the Krafla central volcano in NE-Iceland. *J. Volc. Geotherm. Res.* 6, p. 197-212.

Eaton, G.P., 1964. Windborne volcanic ash: A possible index to polar wandering. *J. Geol.*, 72, p. 1-35.

Einarsson, T. and Sigurgeirsson, Th., 1955. Rock magnetism in Iceland. *Nature* 175, p. 892.

Franzson, H., 1978. *Structure and Petrochemistry of the Hafnarfjall Skardsheidi Central Volcano and the Surrounding Basalt Succession, W-Iceland. (Ph. D. Thesis)*, Edinbugh, Scotland, Univ. Edinburgh.

Gunnlaugsson, E., 1980. *Borgarfjördur, efnafraedi jardhitavatns*. National Energy Authority Report OS80020/JHD11.

Hauksson, E. and Goddard, J.G., 1981. Radon earthquake precursor studies in Iceland. *Journal of Geophysical Research* 86, p. 7037-7054.

Irving, E., 1964. *Paleomagnetism and its Application to Geological and Geophysical Problems*. Ed. Wiley.

Johannesson, H., 1975. *Structure and Petrochemistry of the Reykjadalur Central Volcano and the surrounding Areas, Midwest Iceland. (Ph. D. Thesis)* Durham, England, Univ. Durham.

Karlsdottir, R., Johnsen, G., Björnsson, A., Sigurdsson, O. and Hauksson, E., 1978. *Jardhitasvædi vid Kröflu*. Orkustofnun report OS JHD 7847, 37 p.

Know, J.B. and Short, N.M., 1964. A Diagnostic model using ash-fall data to determine eruption characteristics. *Bull Volcanol*. 27, p. 5-24.

Kristmannsdóttir, H., 1978. Alteration of basaltic rocks by hydrothermal activity of 100-300°C. In *International Clay Conference 1978*, 359-367, M.M. Mortland and V.C. Former (Eds.). Elsevier, Amserdam.

Lapedes, D.N. (editor), 1978. *McGraw-Hill Encyclopedia of the Geological Sciences*, McGraw-Hill, New York.

McDougall, I., Saemundsson, K., Johannesson, H., Watkins, N.V., Kristjansson, L., 1977. Extension of the geomagnetic polarity time scale to 6.5 m.y. K-A, dating geological and paleomagnetic study of a 3.500 m lava succession in Western Iceland. *Bull. Geol. Soc. Am*. 88, p. 1-15.

Moorbath, S., Sigurdsson, H. and Goodwin, R., 1968. K-Ar ages of the oldest rocks in Iceland. *Earth Planet. Sci. Lett.*, 4, p. 197-205.

Piper, J.D., 1971. Ground magnetic studies of crustal growth in Iceland. Earth Planet. *Sci. Lett.* 12, p. 199-207.

Saemundsson, K., 1974. Evolution of the Axial Rifting Zone in Northern Iceland and the Tjörnes Fracture Zone. *Geol. Soc. Am. Bull.* 85, 495-504.

Saemundsson, K., 1978. Fissure swarms and central volcanoes of the neovolcanic zones of Iceland. *Geol. J.*, Spec. Issue 10: 415-432.

Saemundsson, K., 1979. Outline of the geology of Iceland. *Jökull* 29, p. 7-28.

Saemundsson, K. and Fridleifsson, I.B., 1980. Jardhiti og jardfrædirannsóknir (Application of geology in geothermal research). *Náttúrufrædingurinn*, 50, 117-188.

Shaw, D.W., Watkins, N.D. and Huang, T.C., 1974. Atmospherically transported volcanic glass in deep-sea sediments: theoretical considerations, *J. Volc. Geotherm. Res.* 79, p. 3087-3094.

Sigvaldason, G.E., 1979. Volcanic prediction in Iceland. Nordic Volcanological Institute, report 79-02, 19 pp.

Sigurdsson, H., 1967. Geology of the Setberg Areas, Snaefellsnes, Western-Iceland. Vísindafélag Íslendinga (Societas Scientiarum Islandica) IV-2, p. 54-123.

Stefánsson, V., 1981. The Krafla Geothermal Field, Northeast Iceland. In: Geothermal Systems: Principles and Case Histories, Chapter 10. Edited by L. Rybach and L.J.P. Muffler 1981, John Wiley & Sons Ltd. 273-293.

Steinthórsson, S., 1967. Tvaer nýjar aldursákvardanir á öskulögum úr Snaefellsjökli, Náttúrufræðingurinn 37, p. 236-238.

Thorarinsson, S., 1944. Tefrokronologiska studier på Island. Geografiska Annaler 26, p. 1-217.

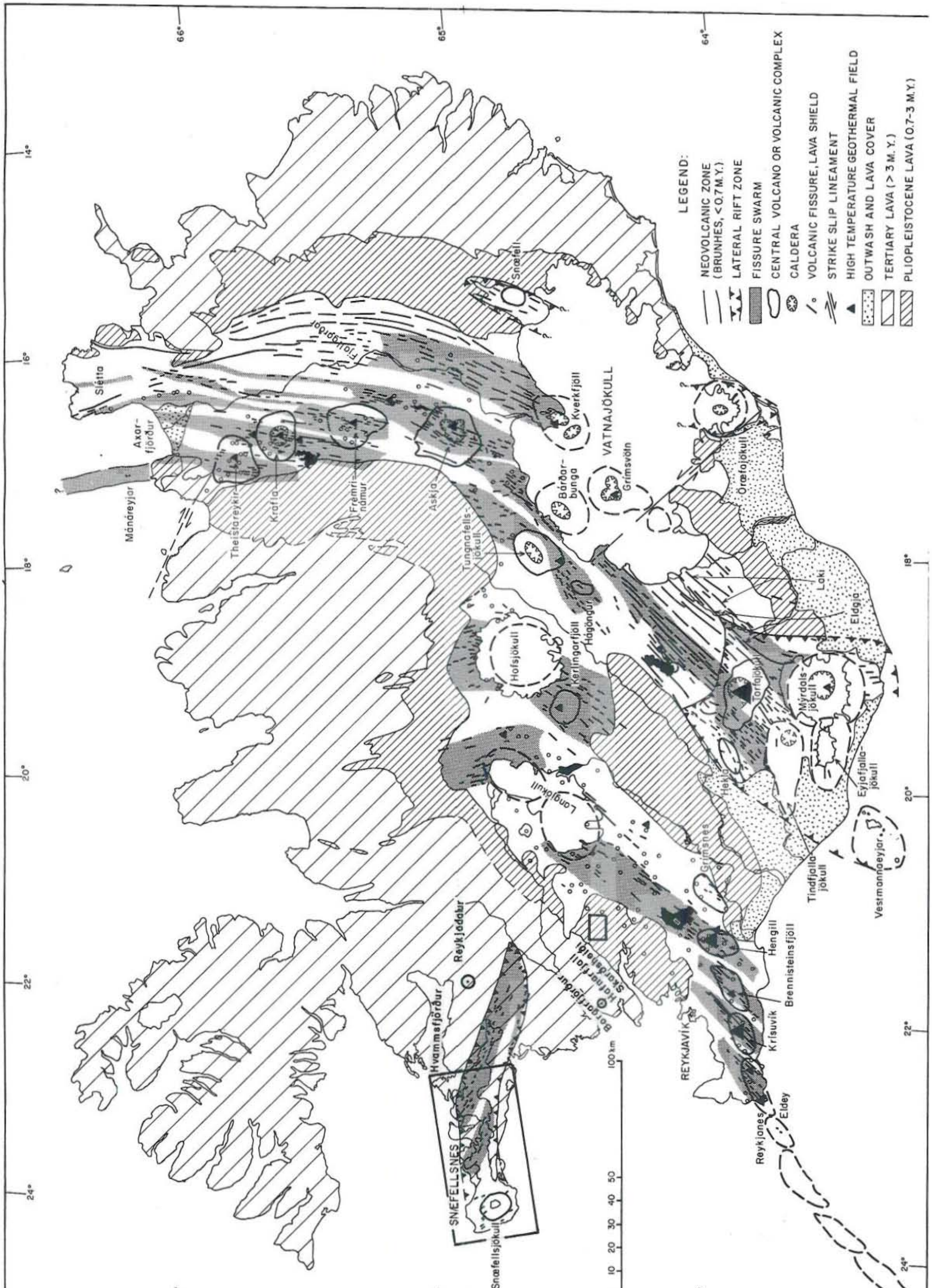
Thorarinsson, S., 1974. Tephrachronology and its application in Iceland. Jökull 29, p. 33-36.

Thorarinsson, S., 197[illegible]. Tephra studies and tephrachronology. A historical review with a special reference to Iceland (a preprint).

Tryggvason, E., 1978. Subsidence events in the Krafla area. Nordic Volcanological Institute, report 1978-14, 65 p.

Walker, G.P.L., 1959. Geology of the Reydarfjördur area, eastern Iceland. Quart. J. Geol. Soc. London, 114, p. 367-393.

Walker, G.P.L., 1960. Zeolite zones and dike distribution in relation to the structure of the basalts in eastern Iceland. J. Geol., 68, p. 515-528.


LEGEND:
NEOVOLCANIC ZONE (BRUNHES, <0.7M.Y.)
LATERAL RIFT ZONE
FISSURE SWARM
CENTRAL VOLCANO OR VOLCANIC COMPLEX
CALDERA
VOLCANIC FISSURE, LAVA SHIELD
STRIKE SLIP LINEAMENT
HIGH TEMPERATURE GEOTHERMAL FIELD
OUTWASH AND LAVA COVER
TERTIARY LAVA (> 3 M.Y.)
PLIOPLEISTOCENE LAVA (0.7-3 M.Y.)
66°
65°
64°
14°
16°
18°
20°
22°
24°
Snæfell
Fjallgarðar
Slétta
Axarfjörður
Mánáreyjar
Theistareykir
Krafla
Fremri-námur
Askja
Kverkfjöll
VATNAJÖKULL
Bárðarbunga
Grímsvötn
Öræfajökull
Tungnafellsjökull
Hágöngur
Hofsjökull
Kerlingarfjöll
Loki
Eldgjá
Torfajökull
Mýrdalsjökull
Eyjafjallajökull
Hekla
Langjökull
Grímsnes
Vestmannaeyjar
Tindfjallajökull
Hengill
Brennisteinsfjöll
Krísuvík
REYKJAVÍK
Reykjanes
Eldey
Reykjadalur
Hvammsfjörður
Borgarfjörður
Hafnarfjall
Skarðsheiði
SNÆFELLSNES
Snæfellsjökull
0 10 20 30 40 50 100km

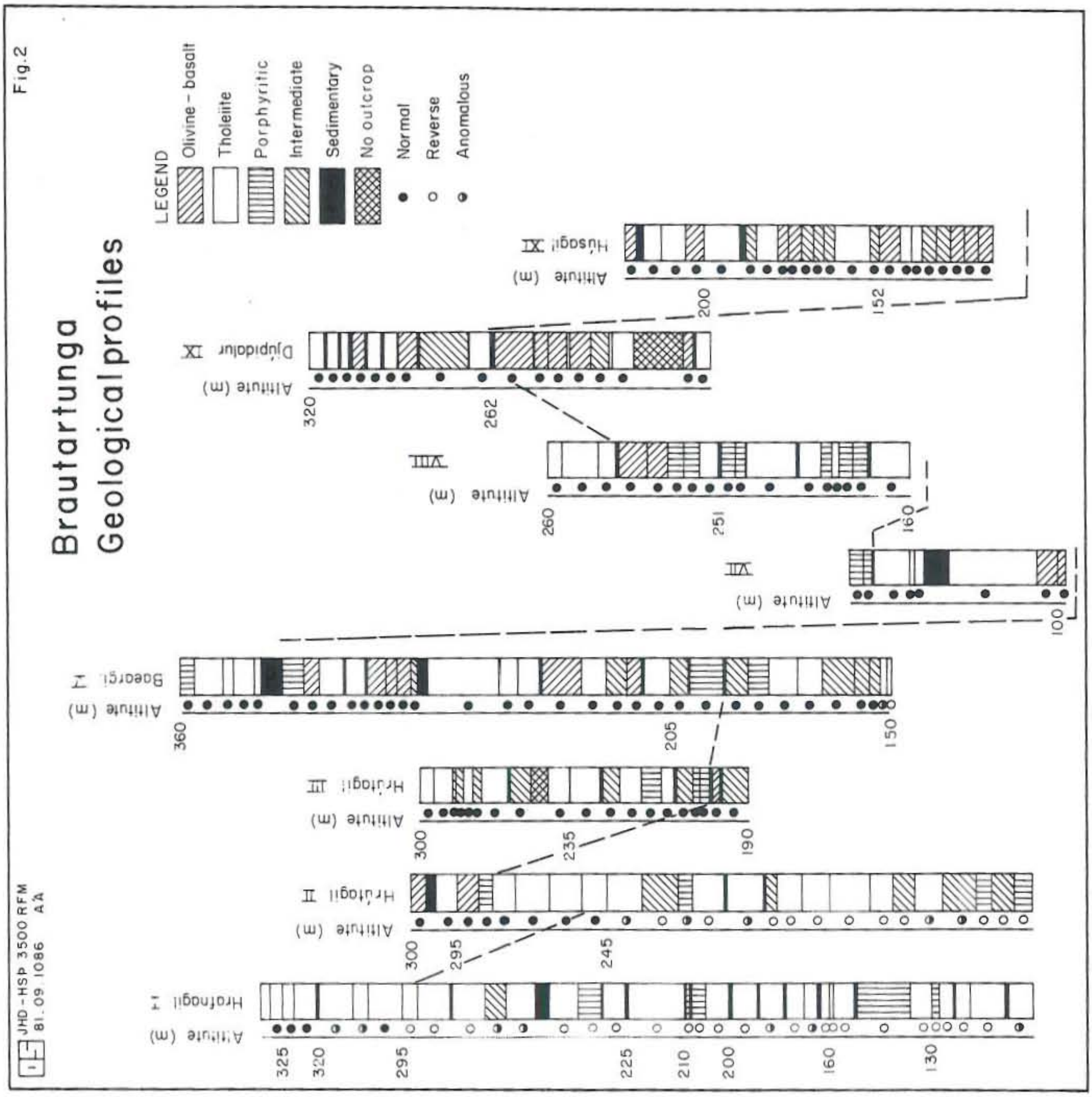


Fig.2
Brautartunga
Geological profiles
LEGEND
Olivine - basalt
Tholeiite
Porphyritic
Intermediate
Sedimentary
No outcrop
Normal
Reverse
Anomalous
JHD-HSP 3500 RFM
81.09.1086 ÁÁ
Hrafnagil I
Altitute (m)
325
320
295
225
210
200
160
130
Hrútagil II
Altitute (m)
300
295
245
Hrútagil III
Altitute (m)
300
235
190
Baergil V
Altitute (m)
360
205
150
VII
Altitute (m)
100
VIII
Altitute (m)
260
251
160
Djúpidalur IX
Altitute (m)
320
262
Húsagil XI
Altitute (m)
200
152

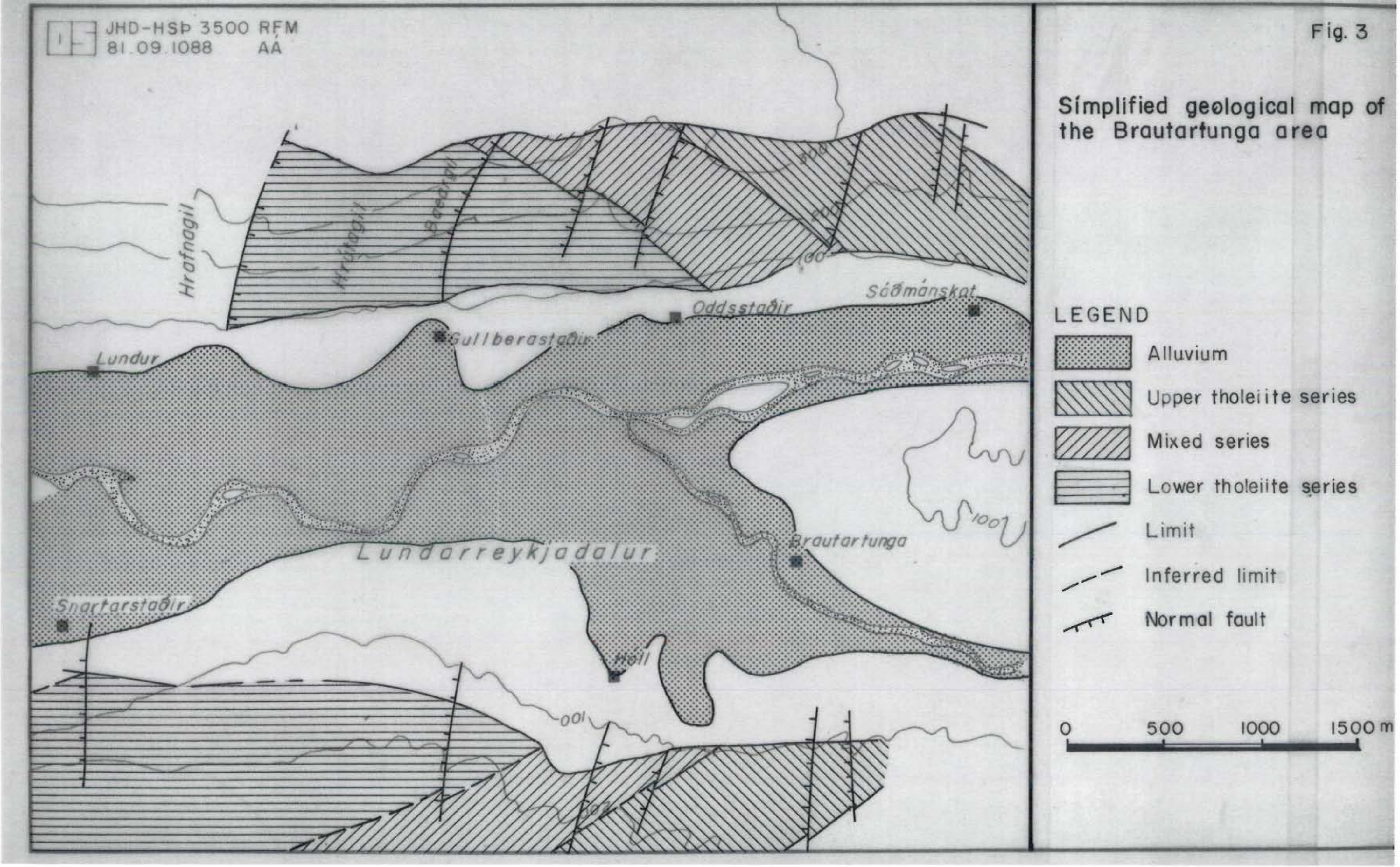


Fig. 3

Simplified geological map of the Brautartunga area

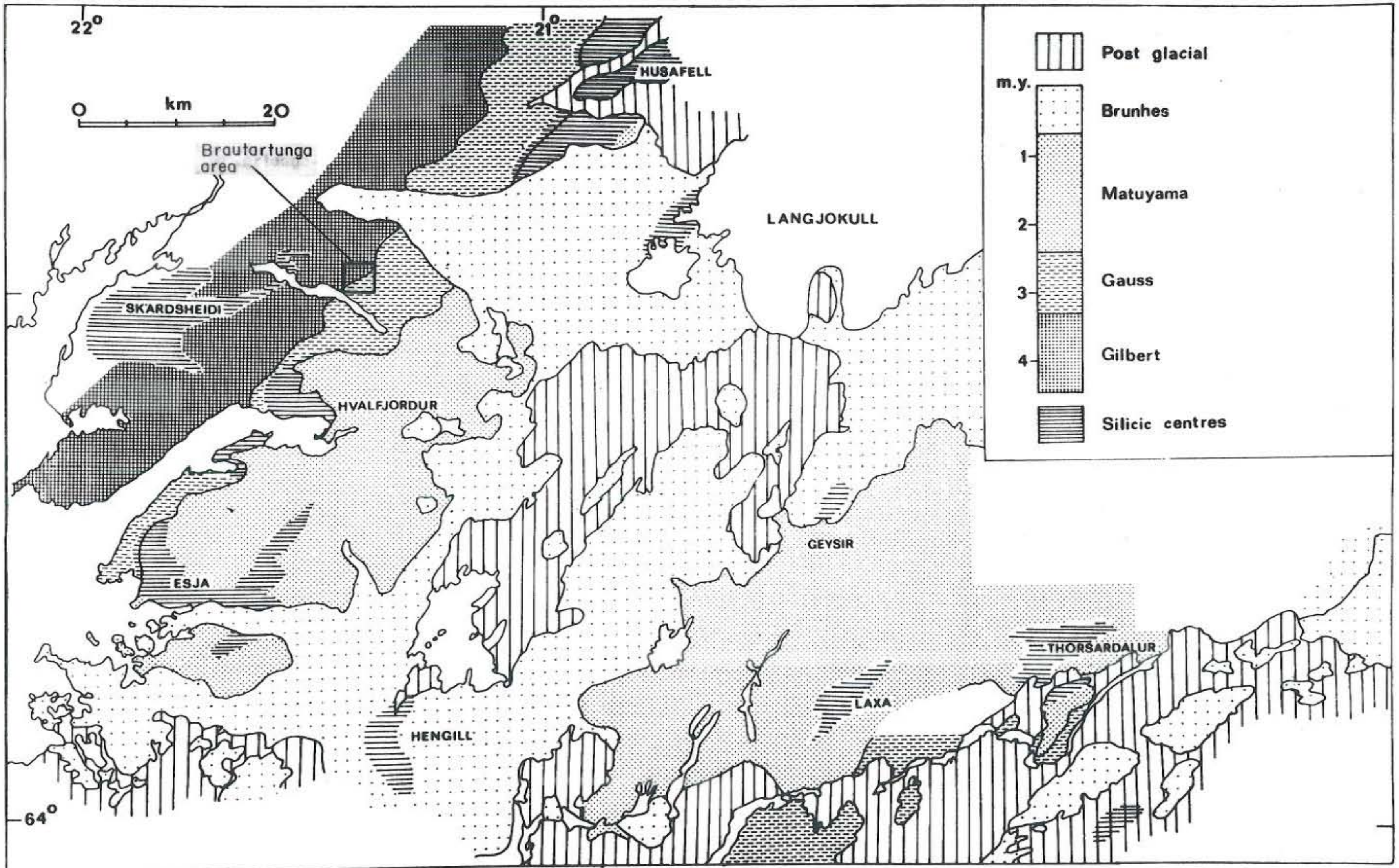


Fig. 4 The distribution of the main magnetic epochs in W-Iceland (Piper J.D.R. 1971)

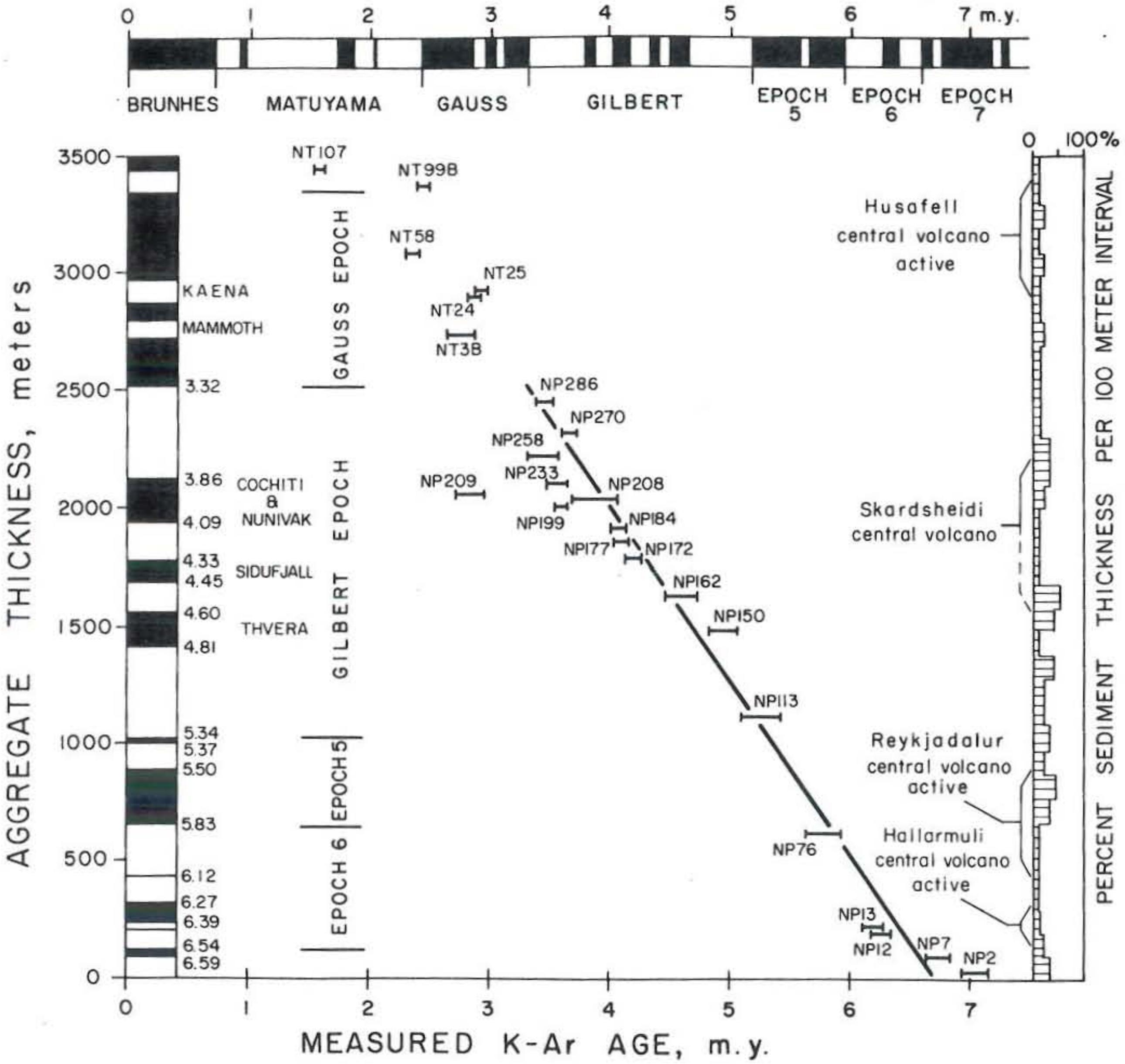


Fig. 5

Magnetic stratigraphy of the upper Borgarfjörður area (McDougall et.al., 1977)

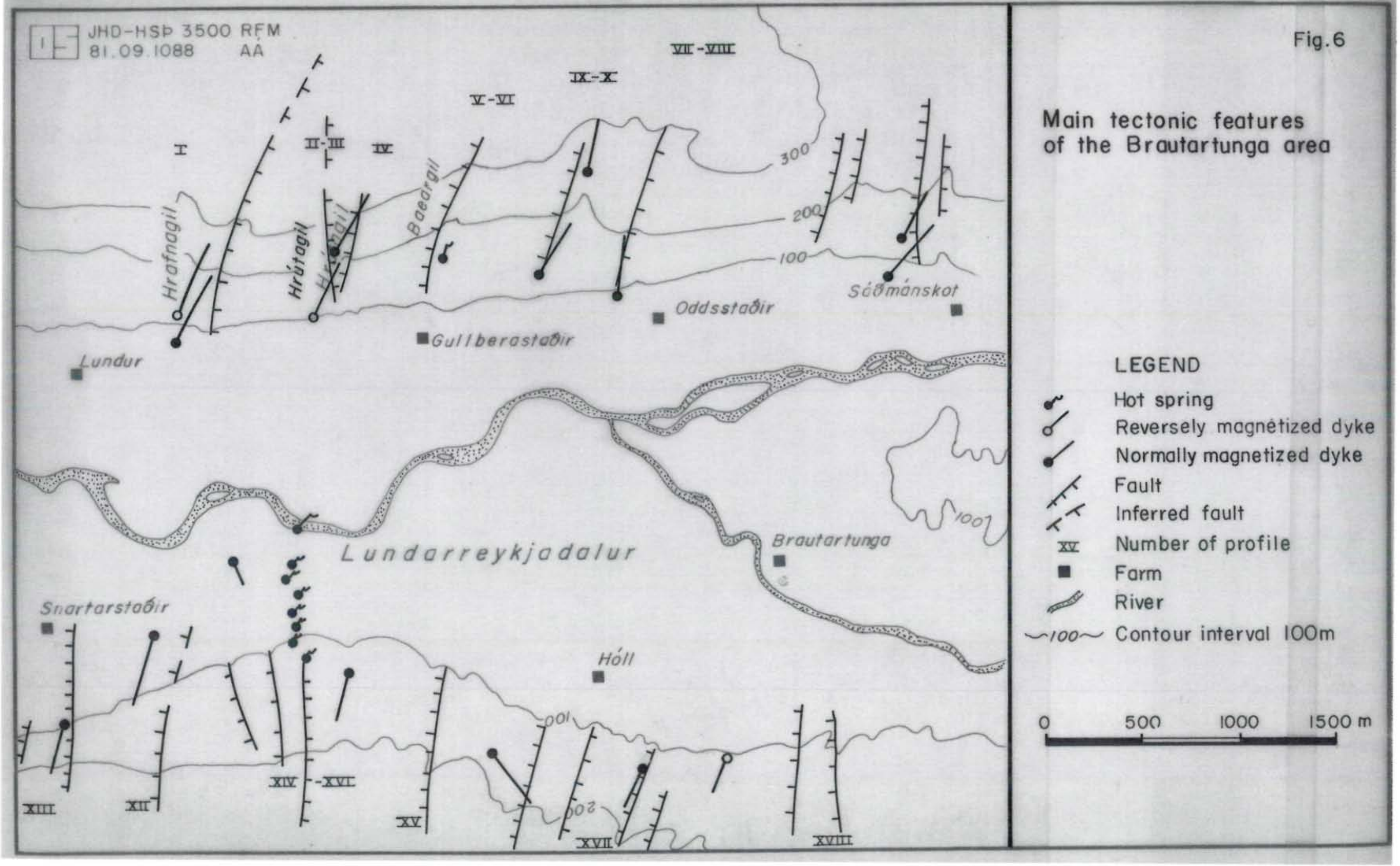


Fig. 6

Main tectonic features of the Brautartunga area

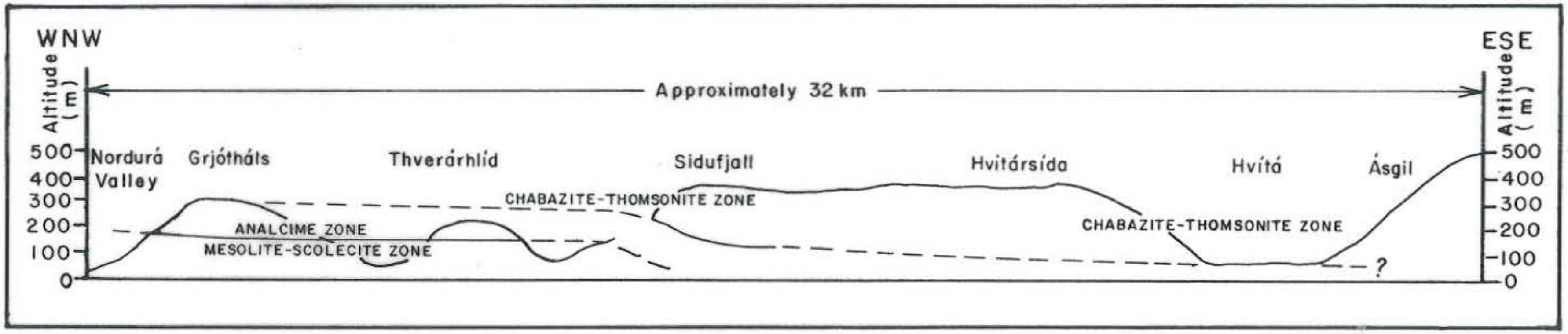
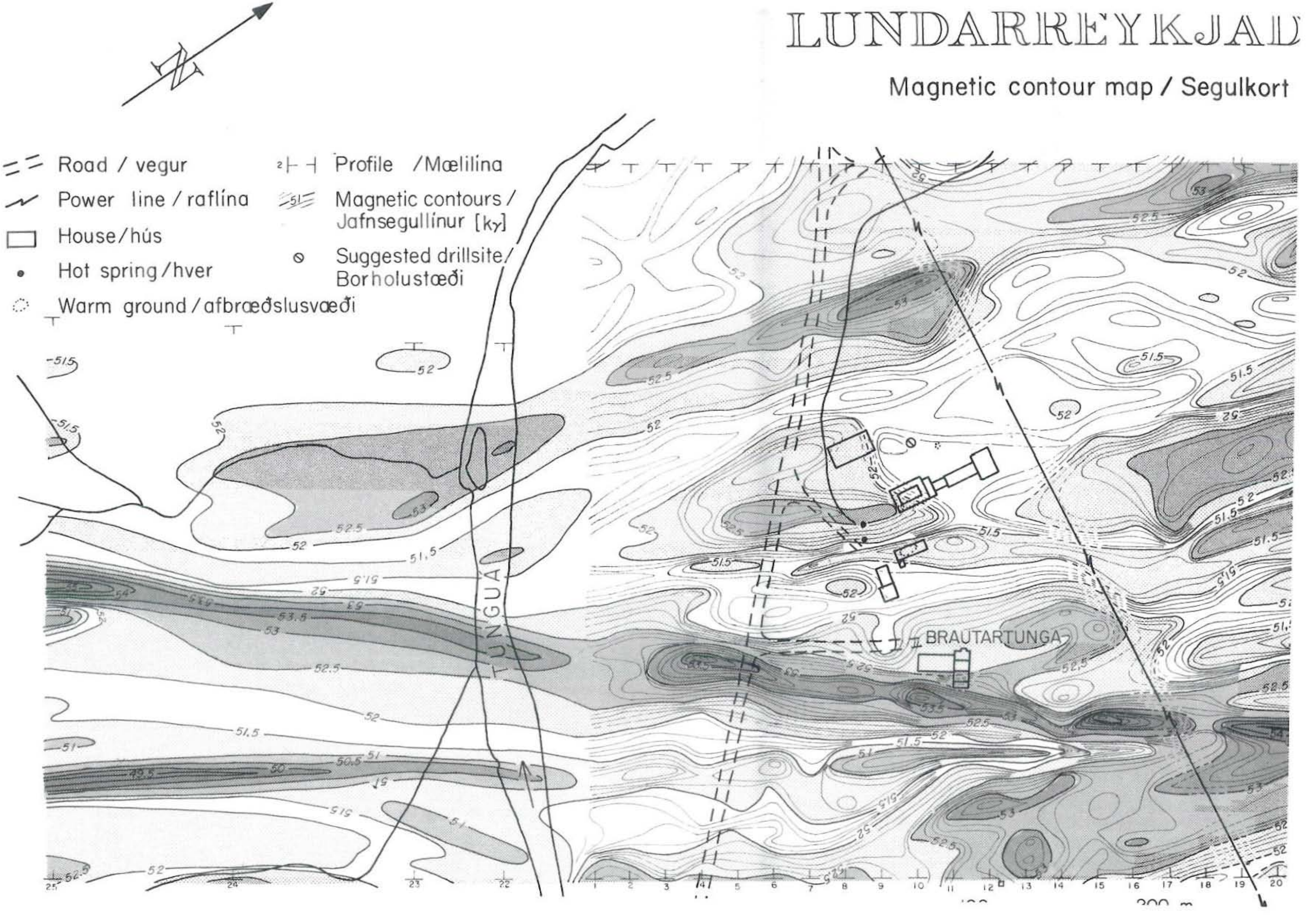


Fig. 7 Diagrammatic sections across the Tertiary lava pile in Borgarfjordur region showing the zeolite zonal distribution (McDougall et al., 1977).

LUNDARREYKJAD

Magnetic contour map / Segulkort



JHD-HSÞ 3507 RFM
81.10.1214 ÁA

Fig. 9

Interpretation of the magnetic anomalies

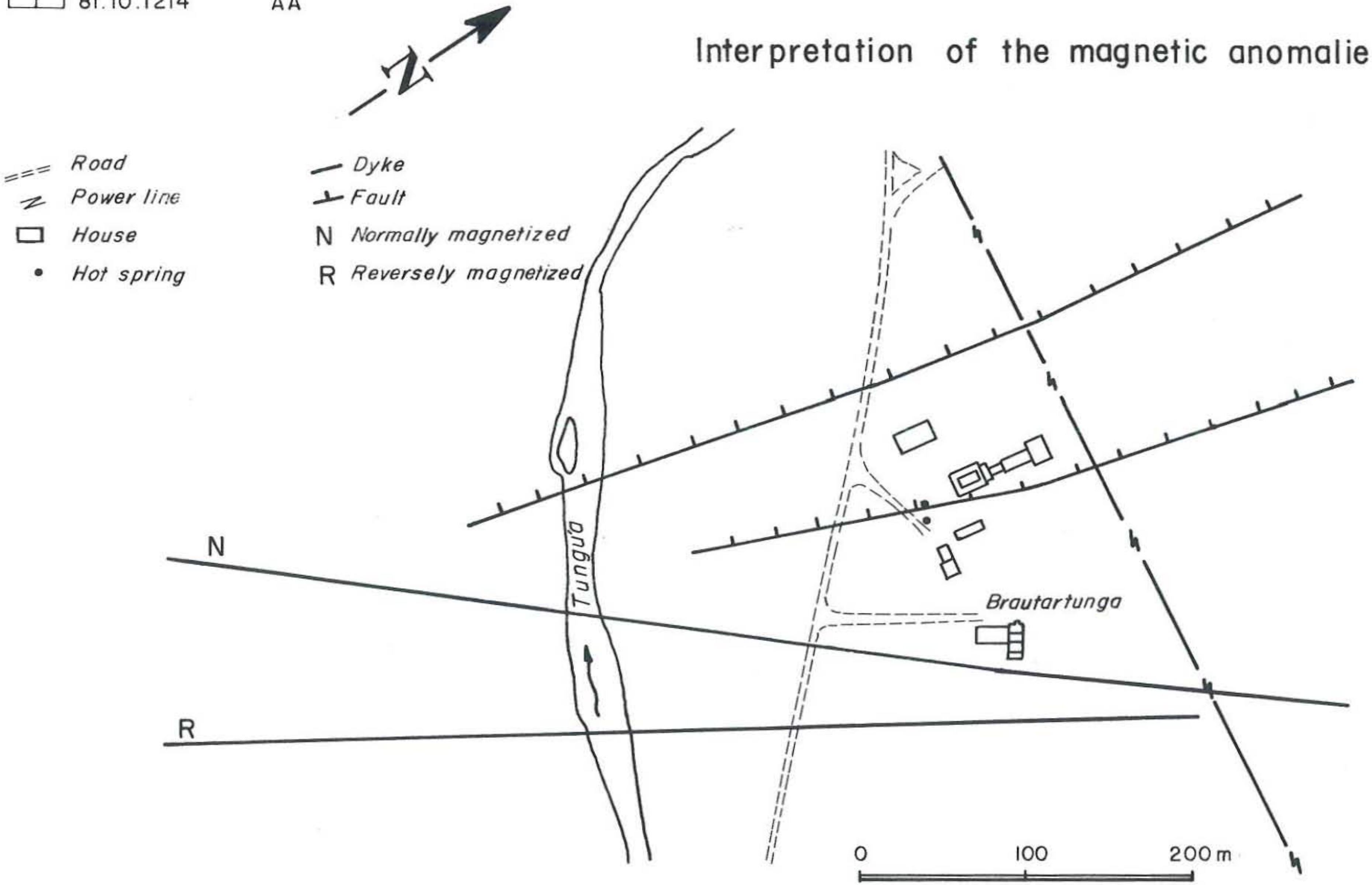
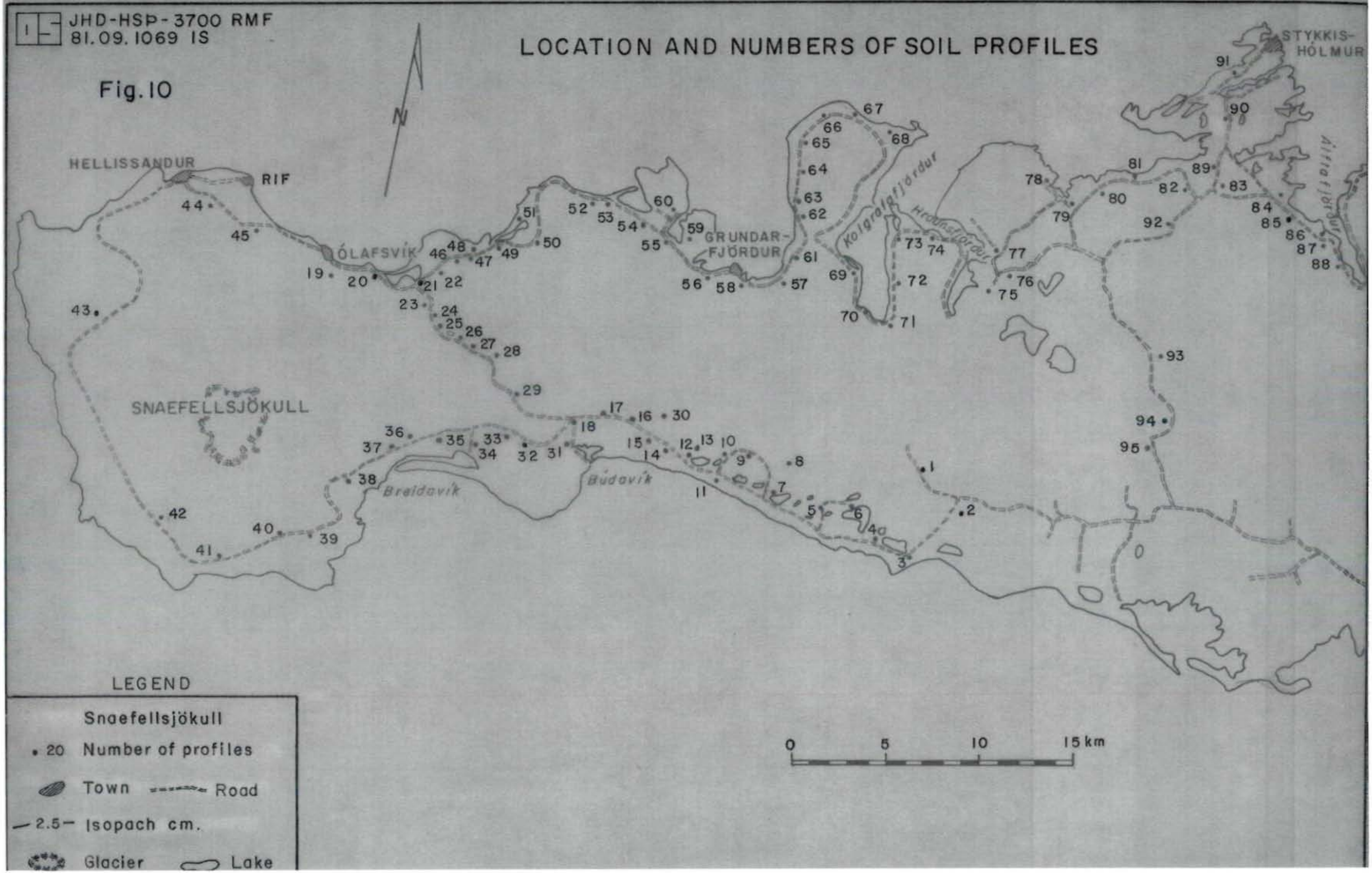


Fig.10

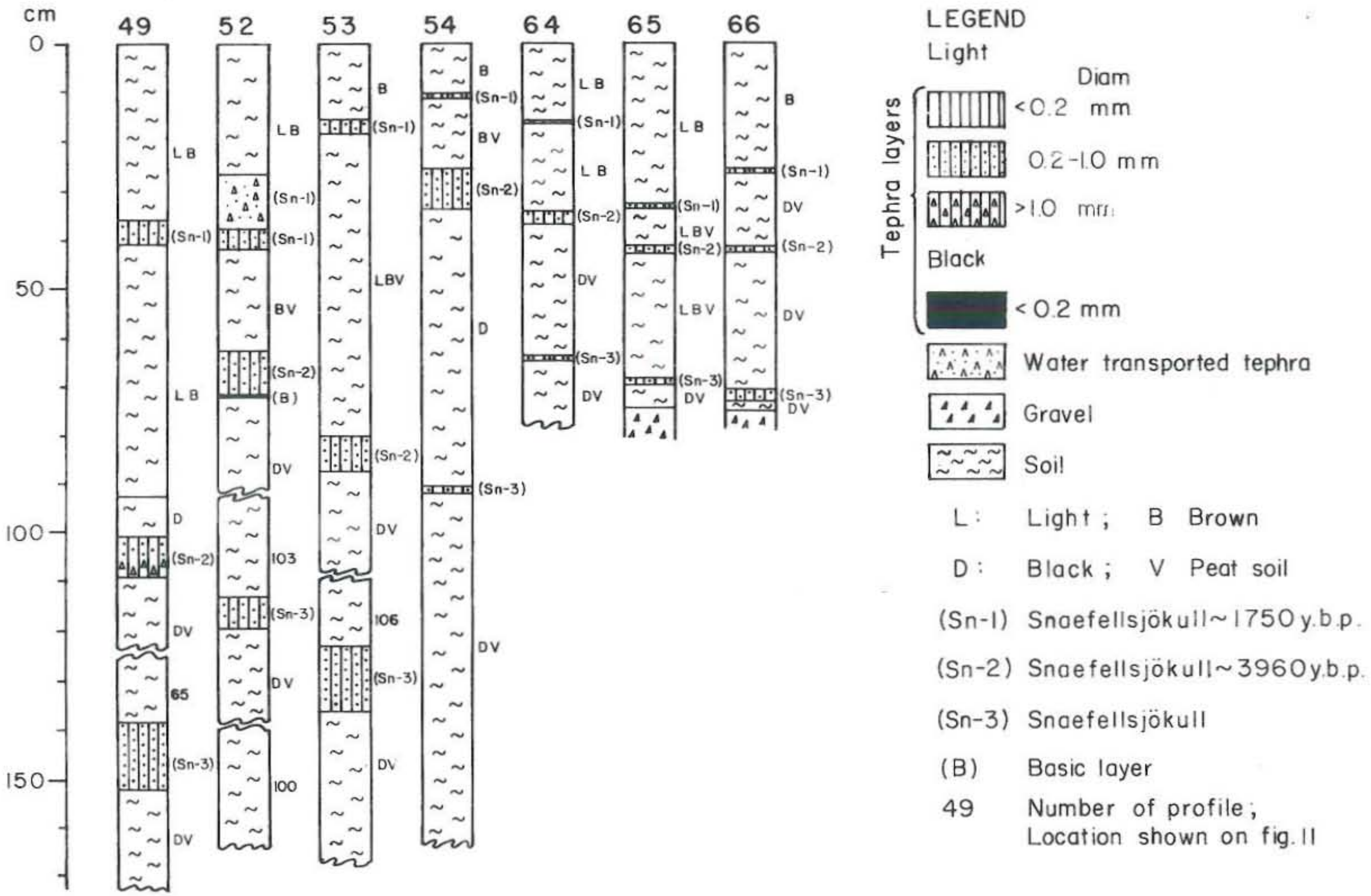
LOCATION AND NUMBERS OF SOIL PROFILES



JHD-HSÞ 3700 RFM
81.09 1087 ÁA

Fig. II

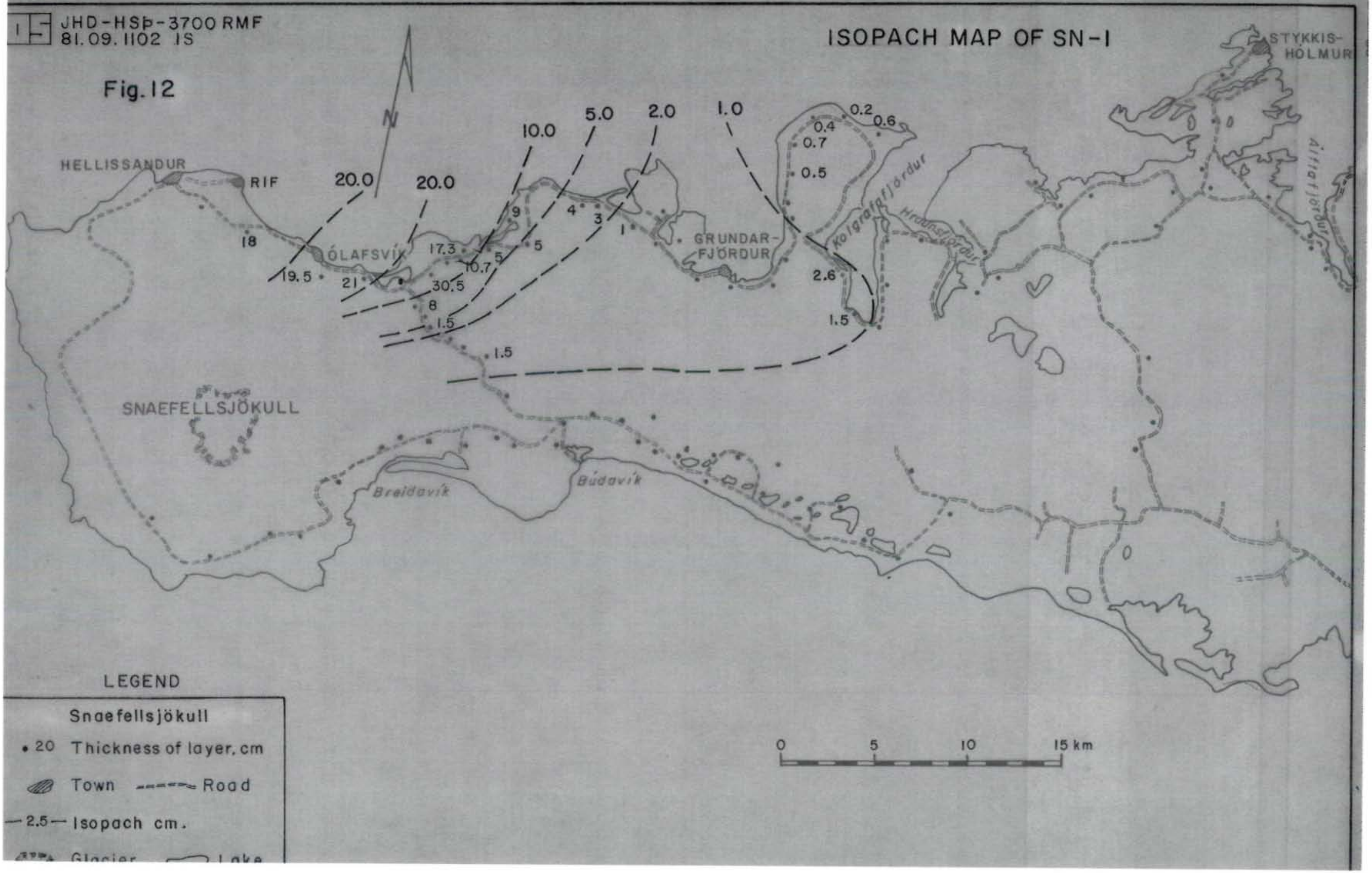
Soil profiles in Snaefellsnes peninsula



JHD-HSÞ-3700 RMF
81.09.1102 IS

Fig. 12

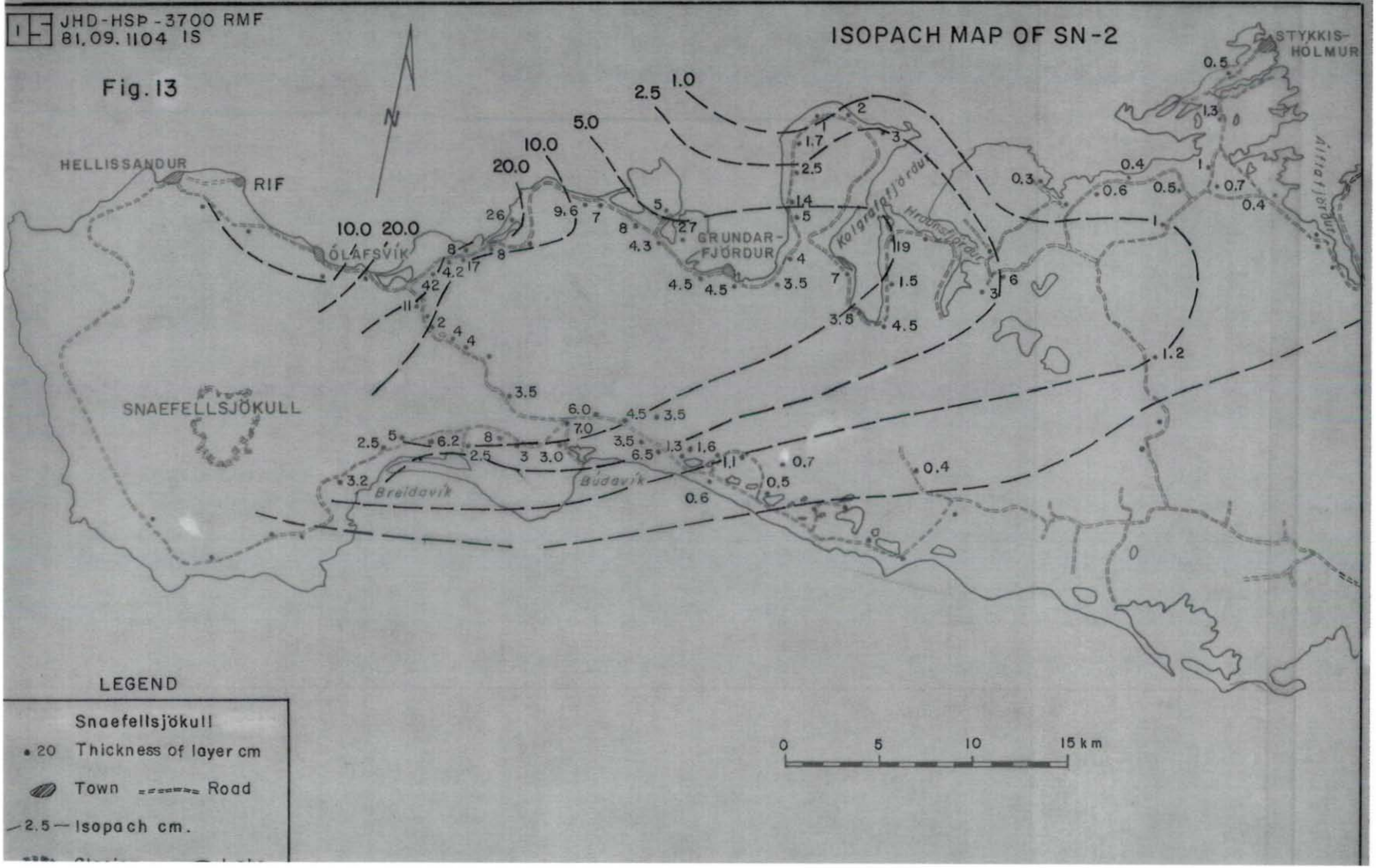
ISOPACH MAP OF SN-1



JHD-HSÞ-3700 RMF
81.09.1104 IS

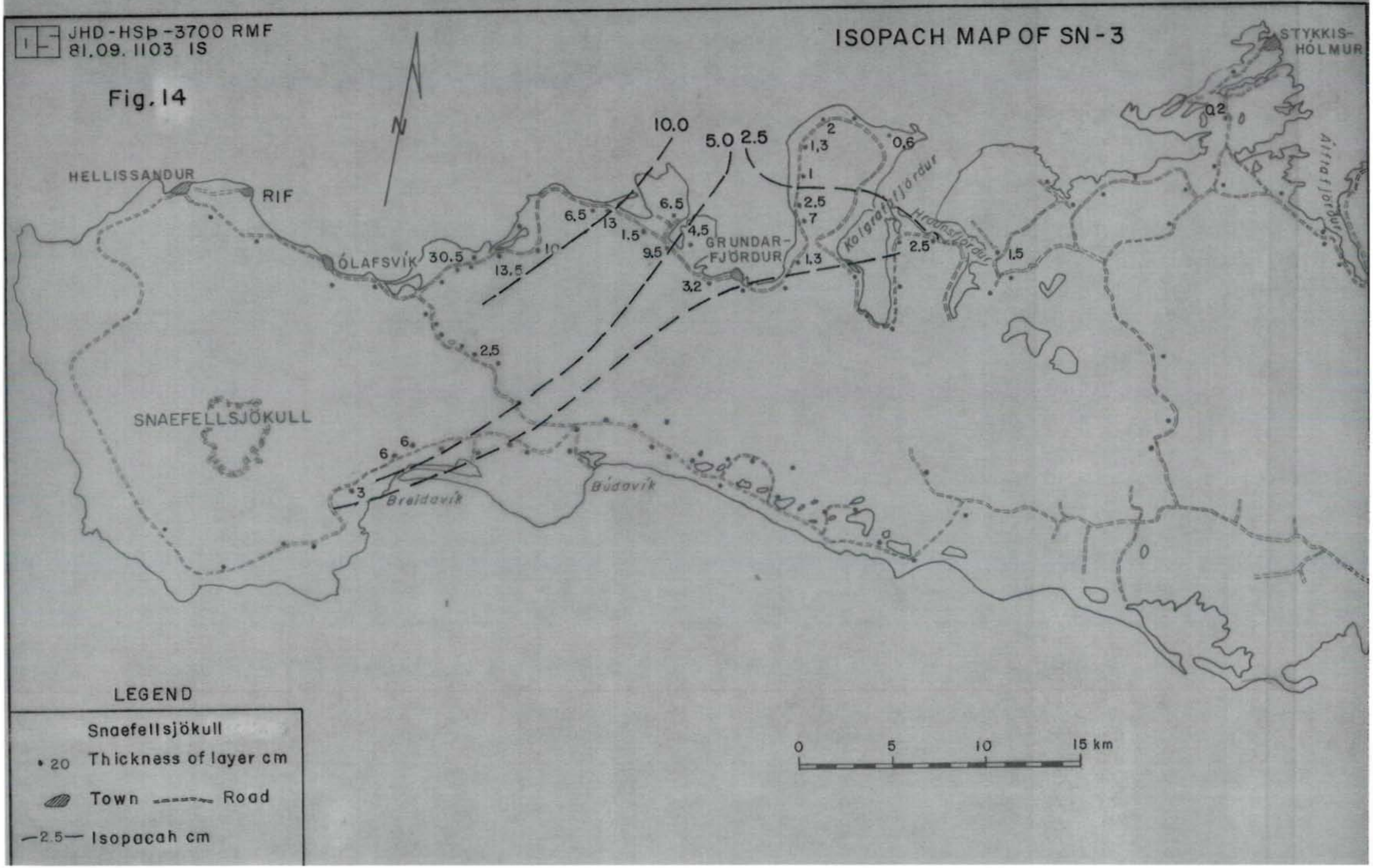
Fig. 13

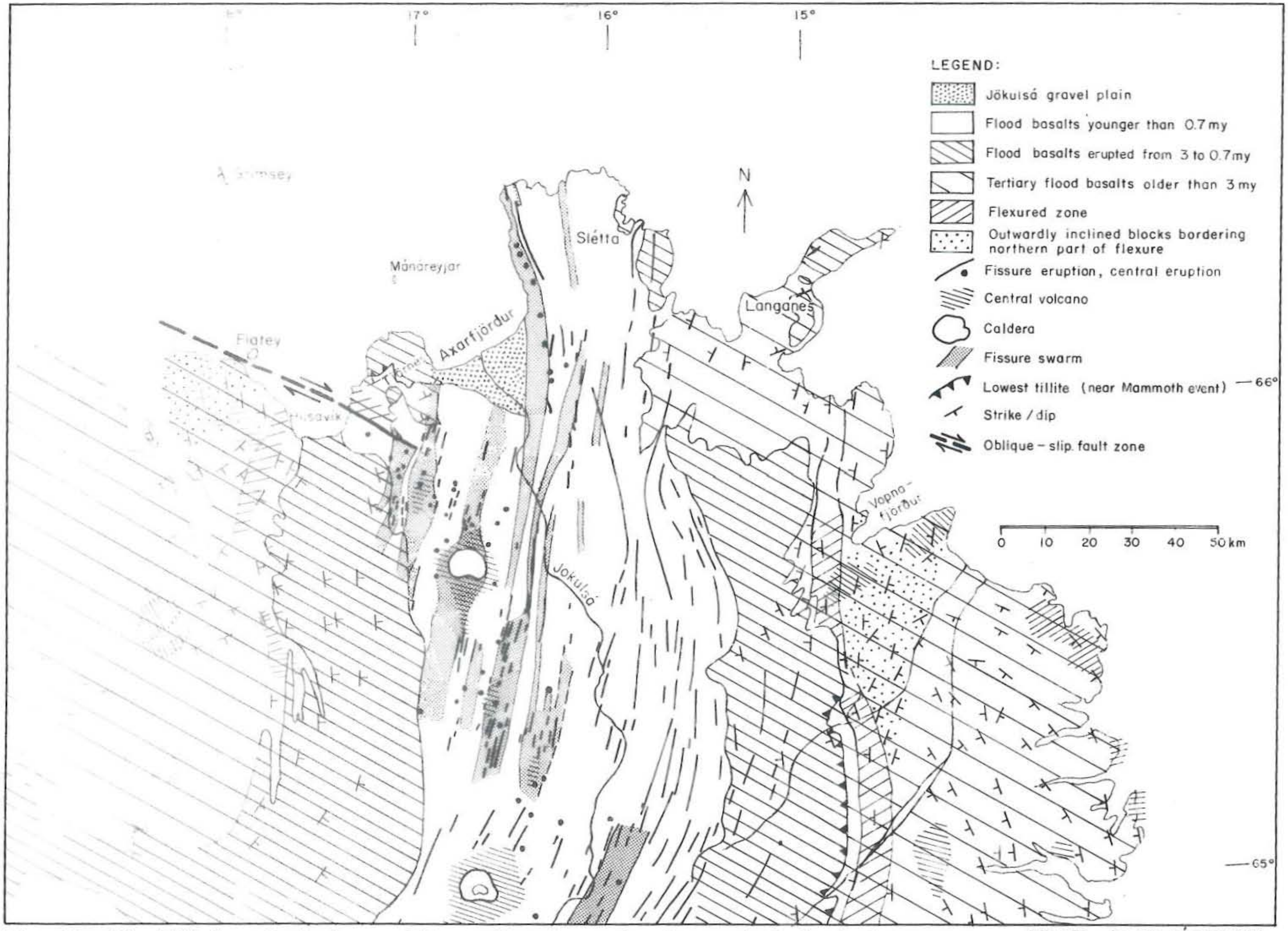
ISOPACH MAP OF SN-2



ISOPACH MAP OF SN-3

Fig. 14





9. 3. 1972 Tnr. 306 J-Ým. Fnr. 10451

Fig. 15 Simplified geological map of the northeastern volcanic zone in Iceland. Two volcanoes; Krafla and Askja have developed calderas (Saemundsson, 1974).

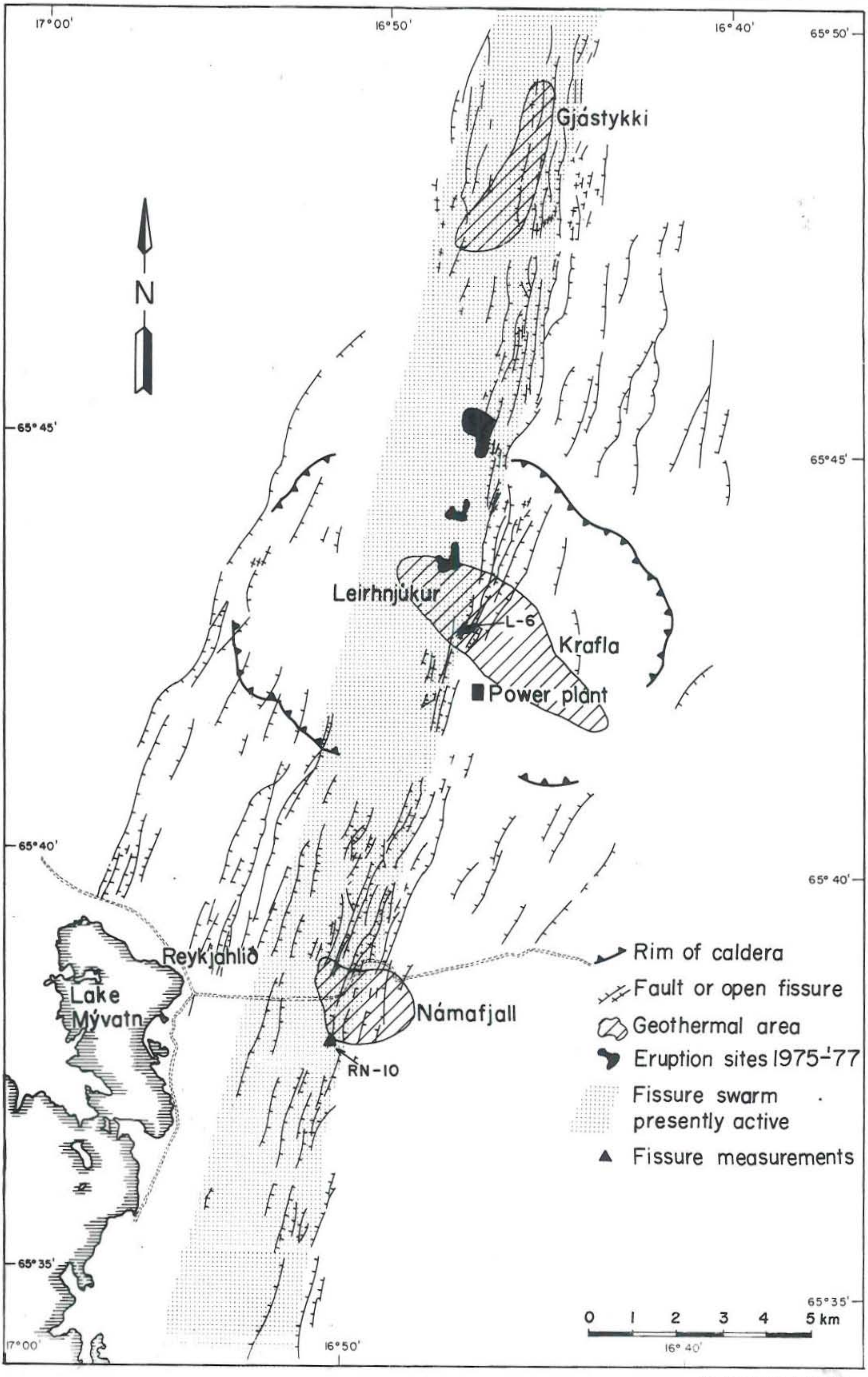


Fig. 16 Tectonic map of Krafla area (Bjornsson et al 1979)

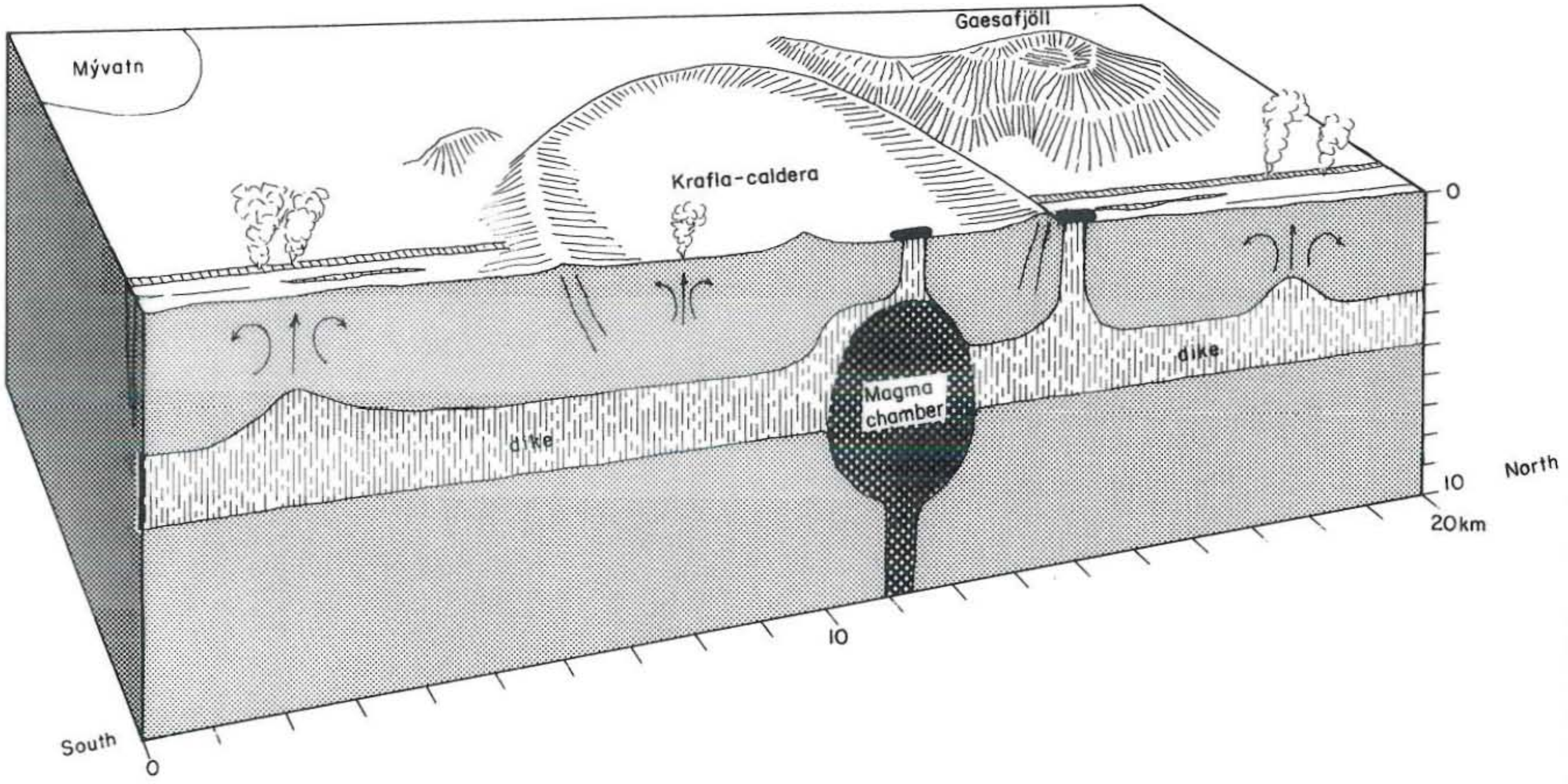


Figure 17. A block diagram showing schematically the magma body below the Krafla caldera and the dyke formed in the present tectonic episode. From Björnsson et al.(1979).

ORKUSTOFNUN

Simplified geological model of the
magma flow in the Krafla-caldera

T-538
Krafla
F16729

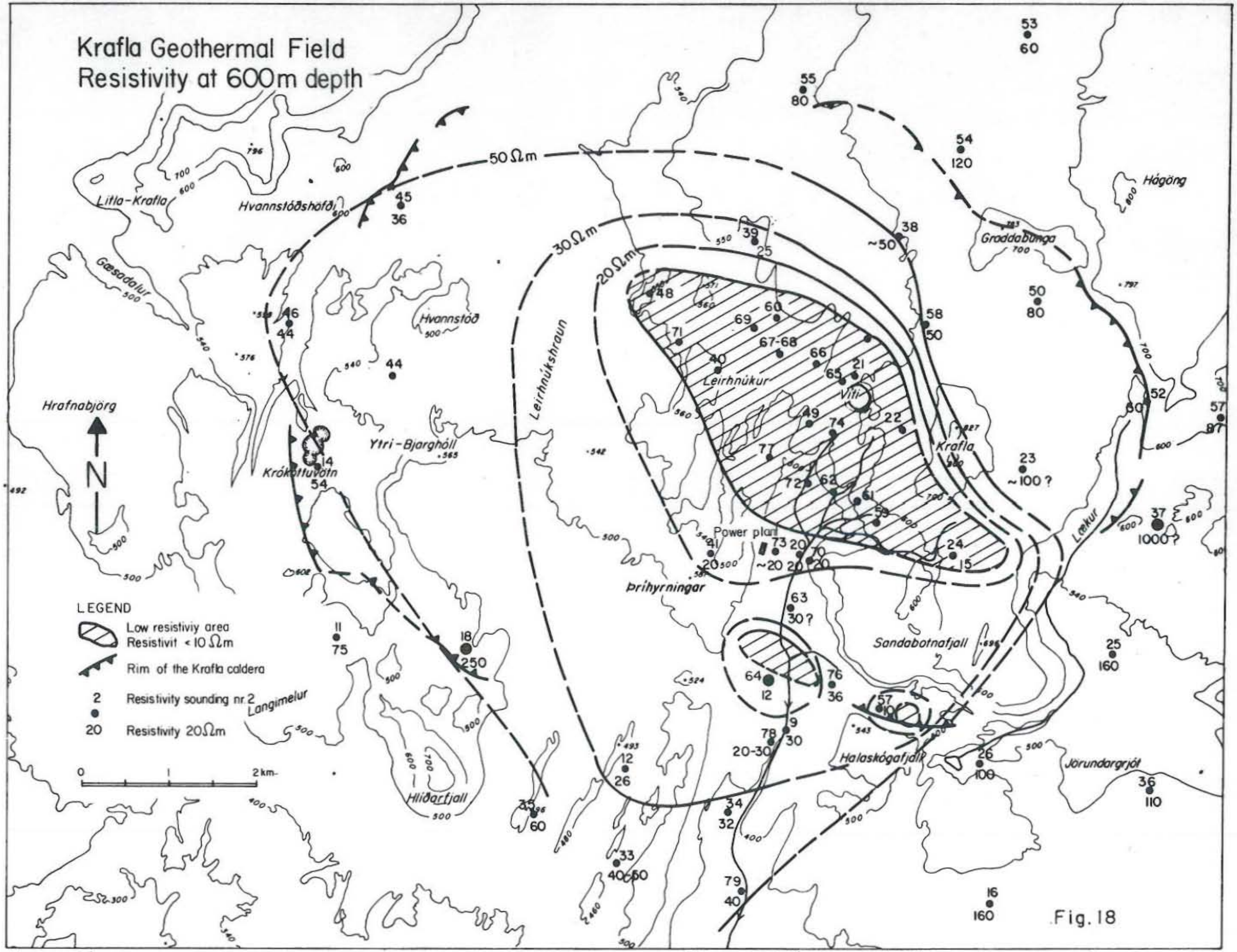


Fig. 18

'78.11.22 RK/AÁ

(From Karlsdóttir et.al., 1978)

Krafla viðnám F 17865

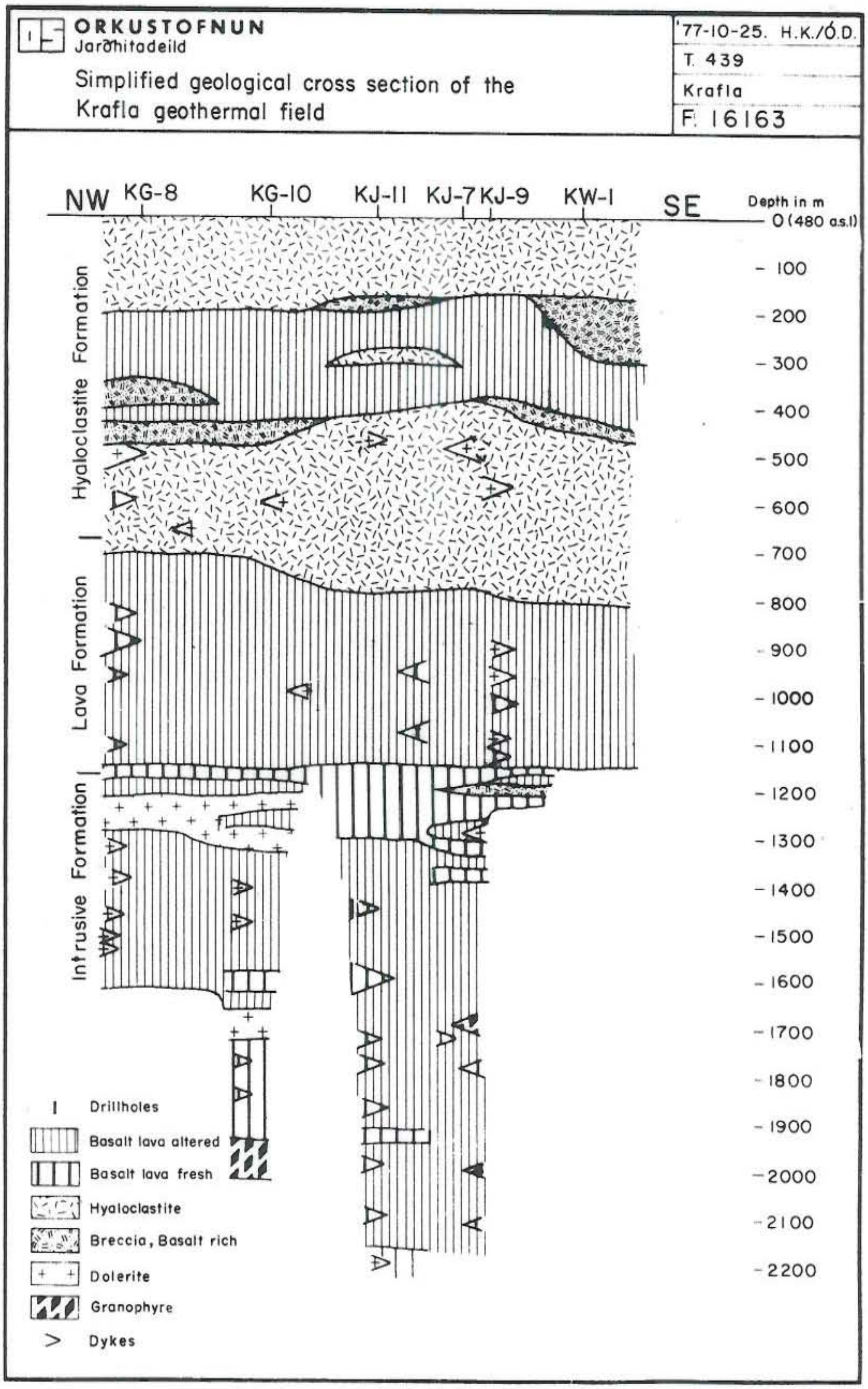


Fig. 19 Simplified geological cross section of the Krafla geothermal field (Kristmannsdóttir, 1978)

OS JHD-HSÞ 3400 HJ.RFM
81.10.1204 AÁ

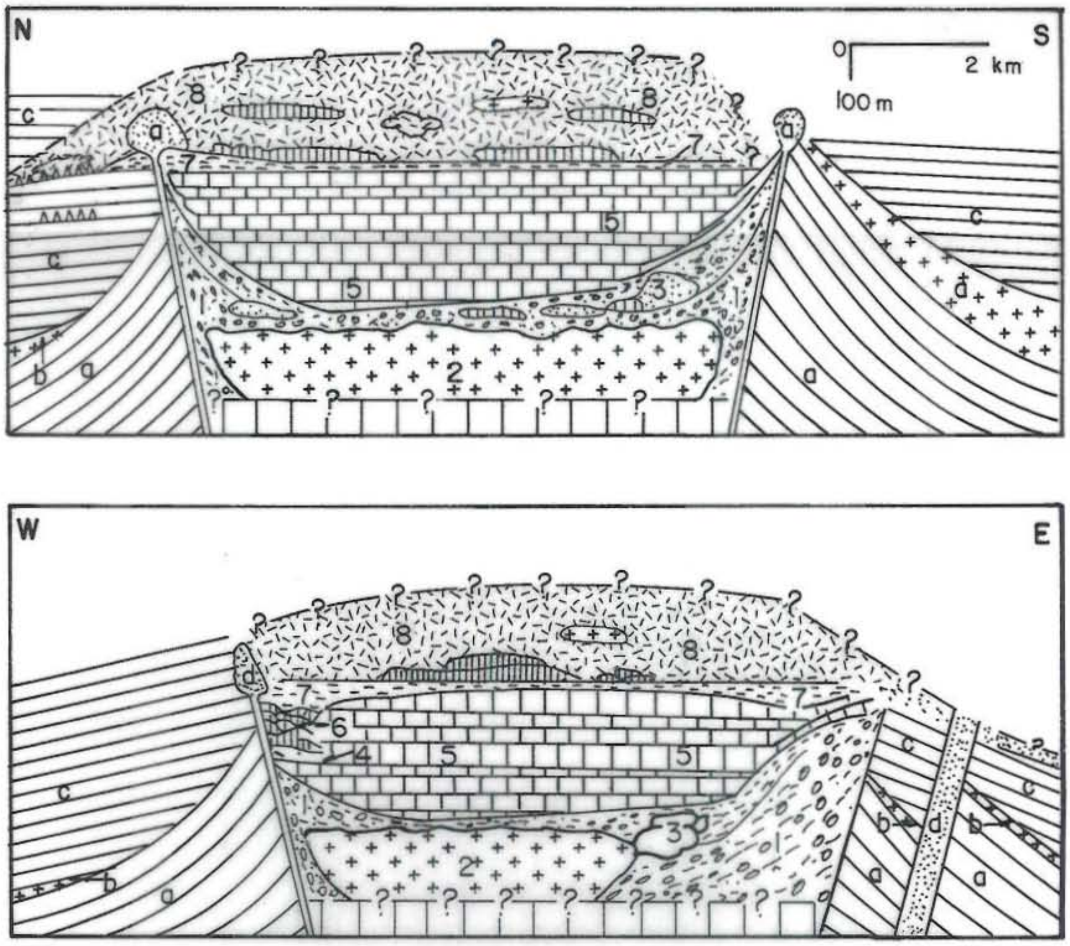


Fig. 20 Two schematic sections across the caldera. The units of the caldera filling are numbered as in the text (1 to 8); the reykjadalur thick-layered series (a), Main Phase of differentiated extrusives (b), Reykjadalur thin-layered series (c), and intermediate and acid cone sheets and plugs (d)

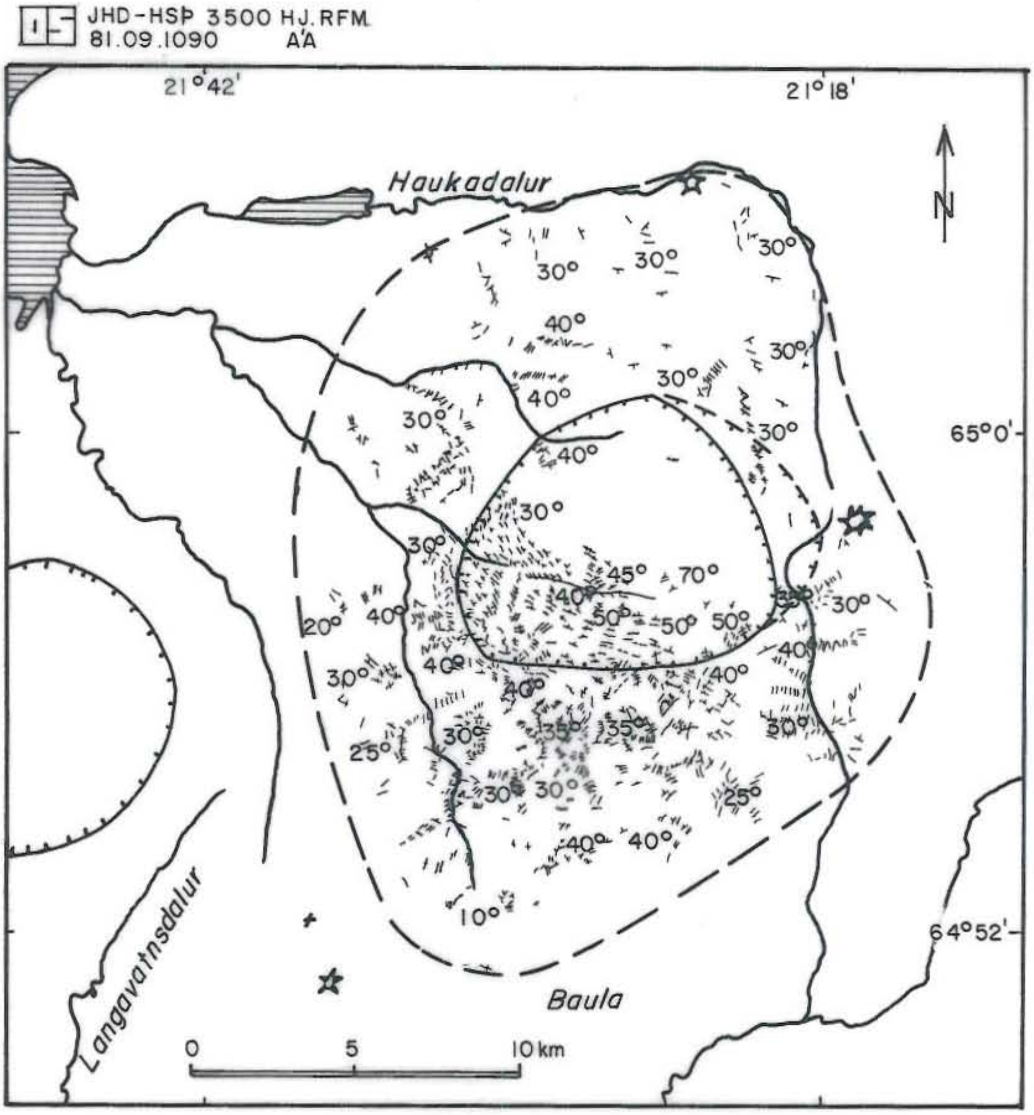


Fig. 21 The average strike and dip of cone sheets. The Reykjadalur and Laugardalur calderas are indicated and the broken line marks the outer limits of cone sheet occurrence: star -like symbols are basic plugs. (Jóhannesson, 1975)

JHD-HSÞ 3500 HF. RFM.
81.09.1089 A'A

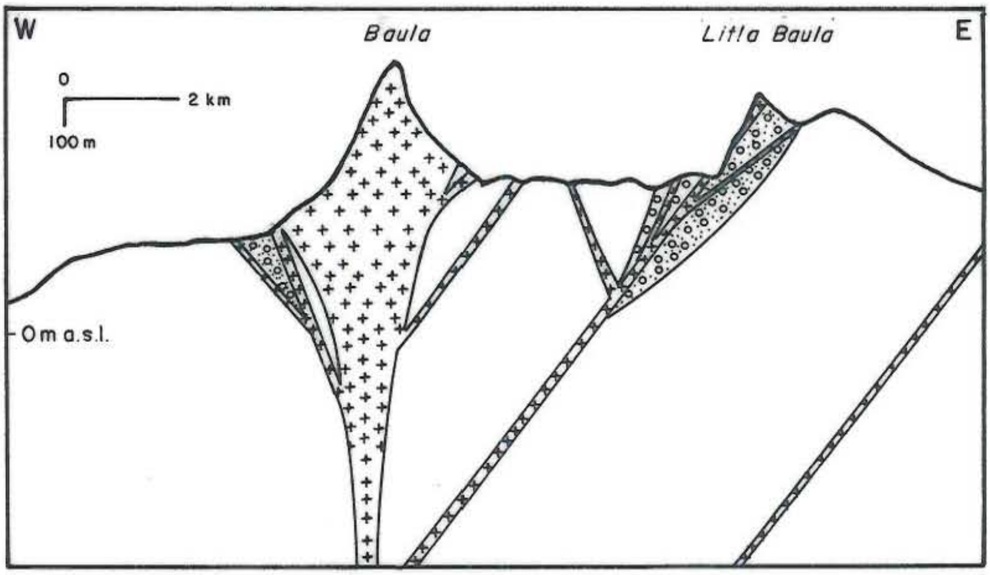


Fig. 22 A schematic section across the Baula complex

(+ + +) Rhyolitic intrusion and dykes

(o o) Rhyolitic vent breccias

(Johannesson, 1975)

JHD- HSÞ 3400 HJ. RFM.
81.10.1203 A'A

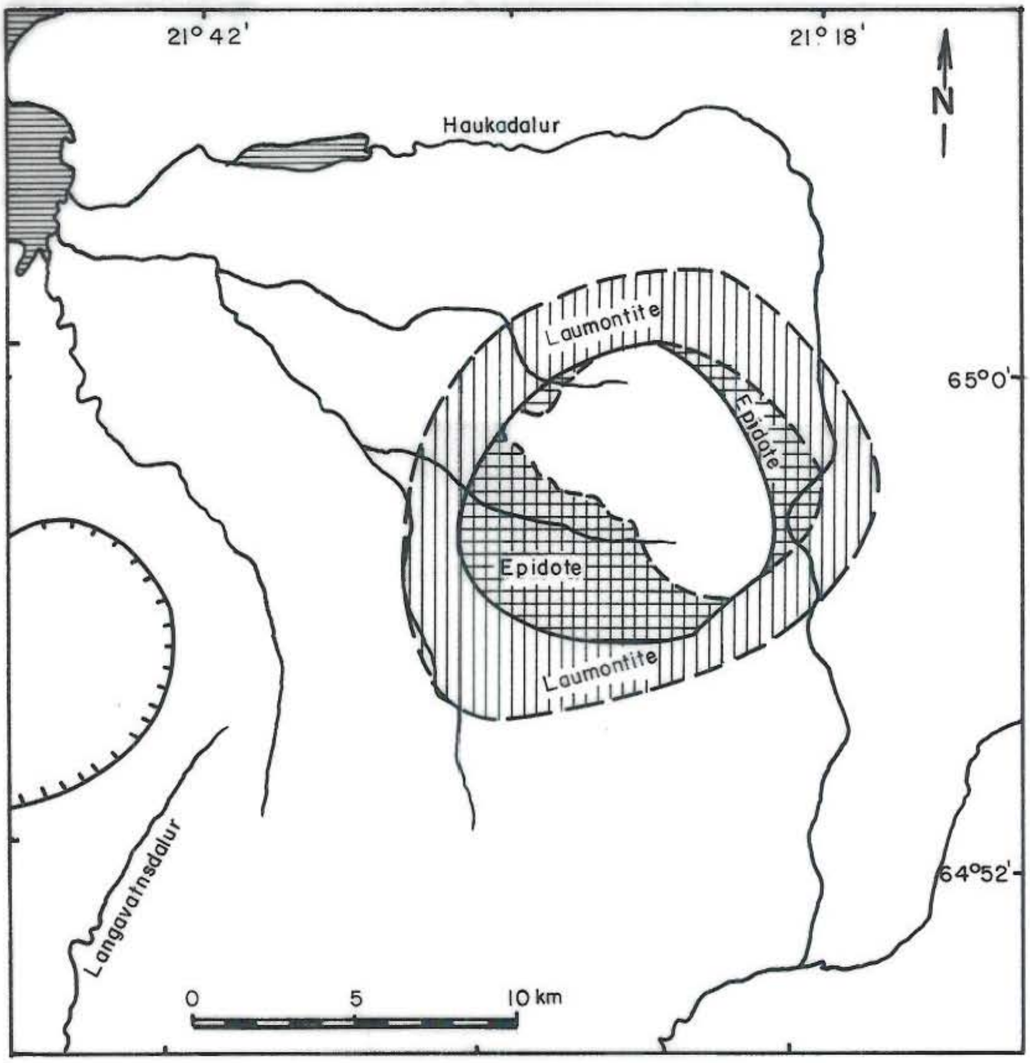


Fig. 23 a Distribution of laumontite and epidote in the Reykjadalur central volcano

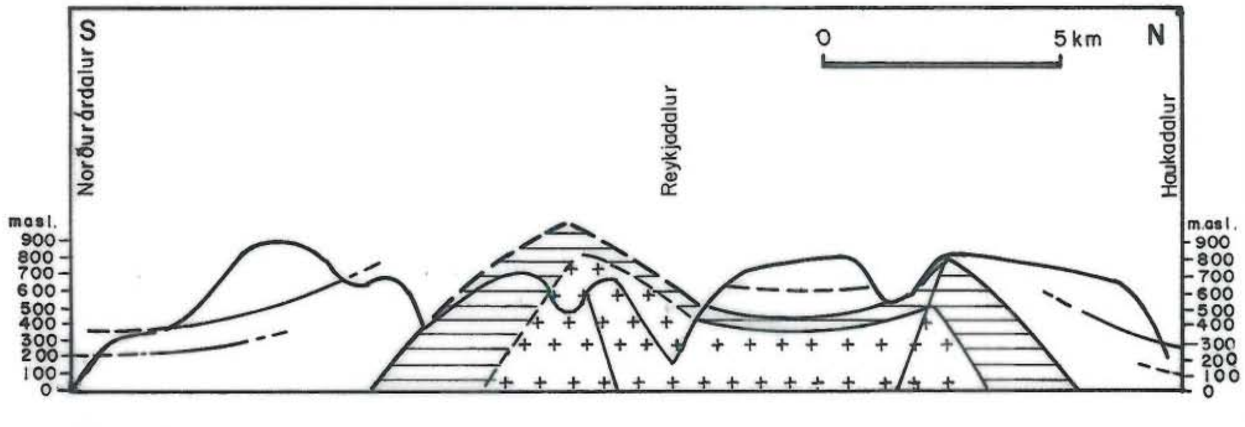


Fig. 23 b A section through the Reykjadalur central volcano, western Iceland, showing the metamorphic zones around it. Large dots, chabazite-thomsonite zone; medium dots, analcine zone; small dots, mesolite-scolecite zone; horizontal lines, laumontite zone; crosses, epidote zone. From Jóhannesson, 1975.

Fig. 24

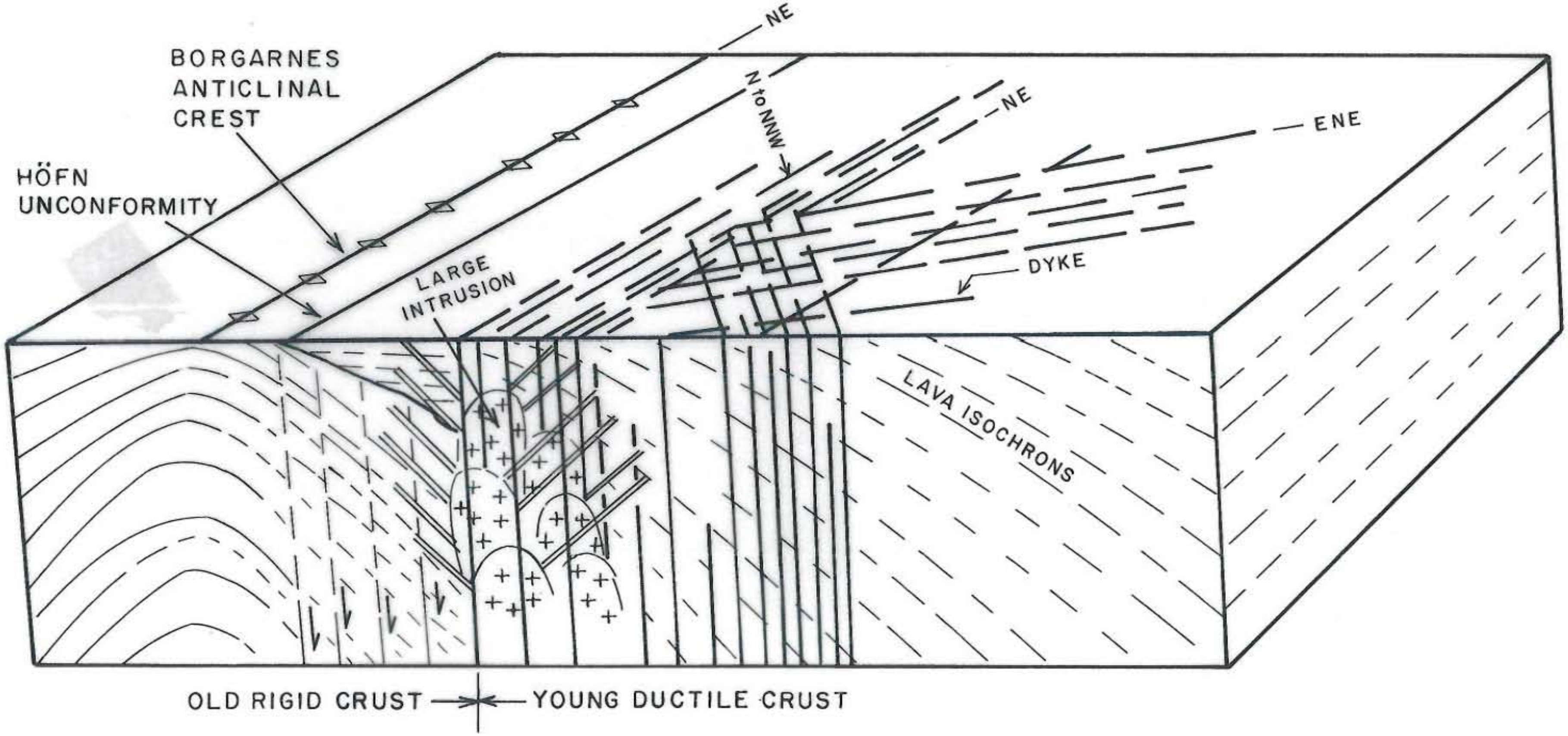
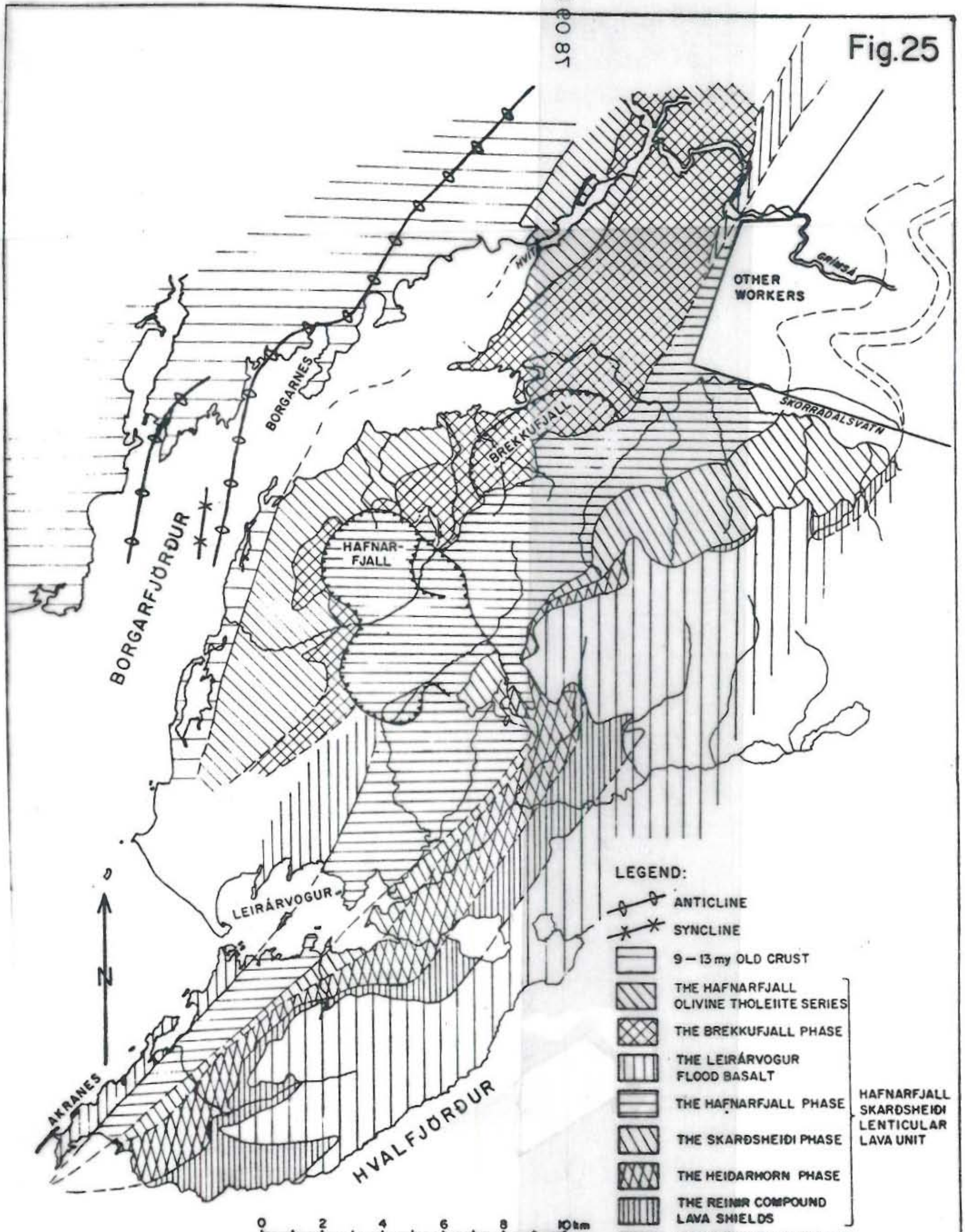


Fig. 24 A hypothetical boundary between the two crustal sections west of Hafnarfjall-Skardsheidi central volcano, as indicated by unconformity, basalt flexuring, larger intrusives, sheet swarm and disappearance of dyke trends. (Franzson, 1978).



(Franzson, 1978)

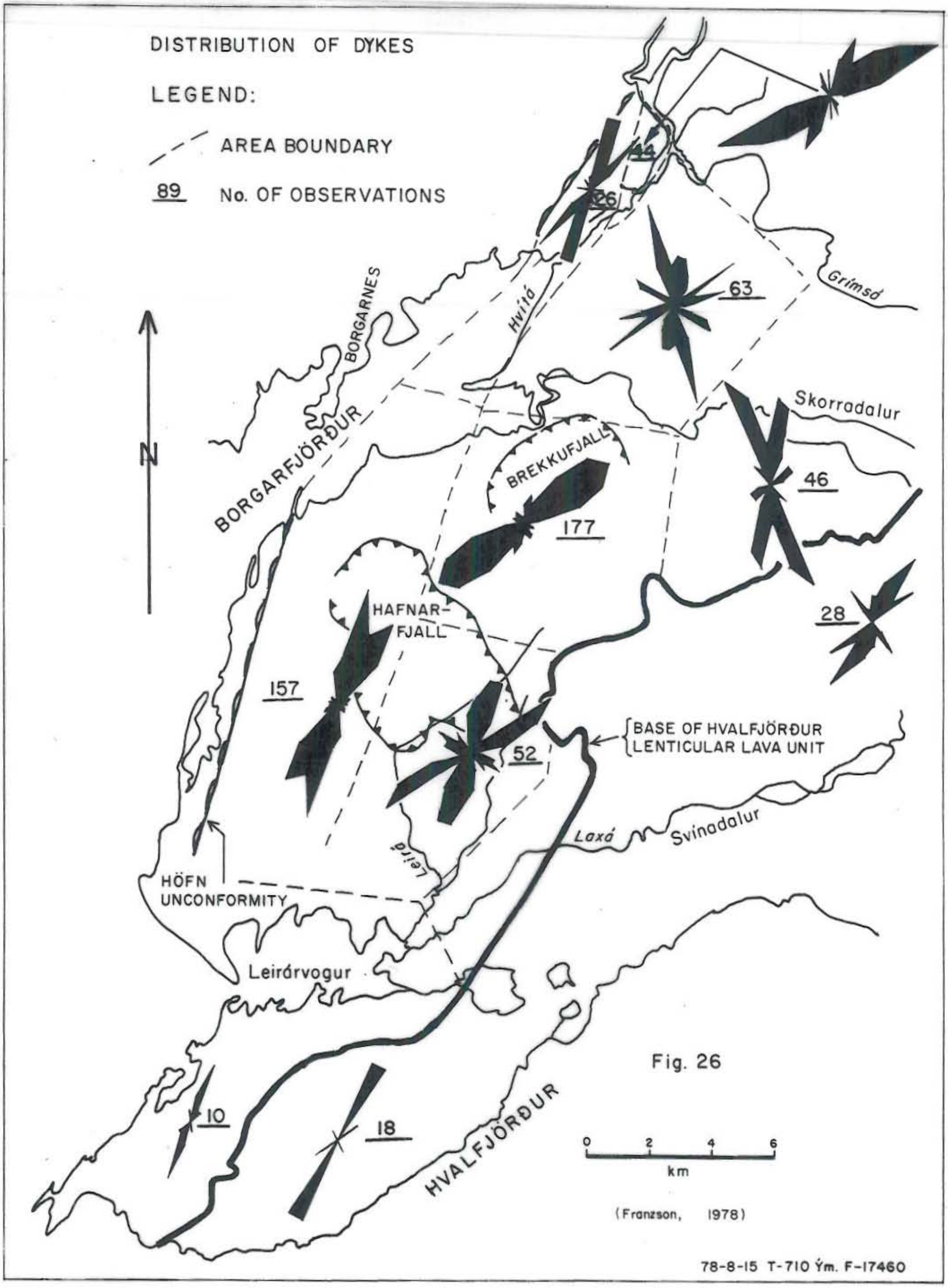
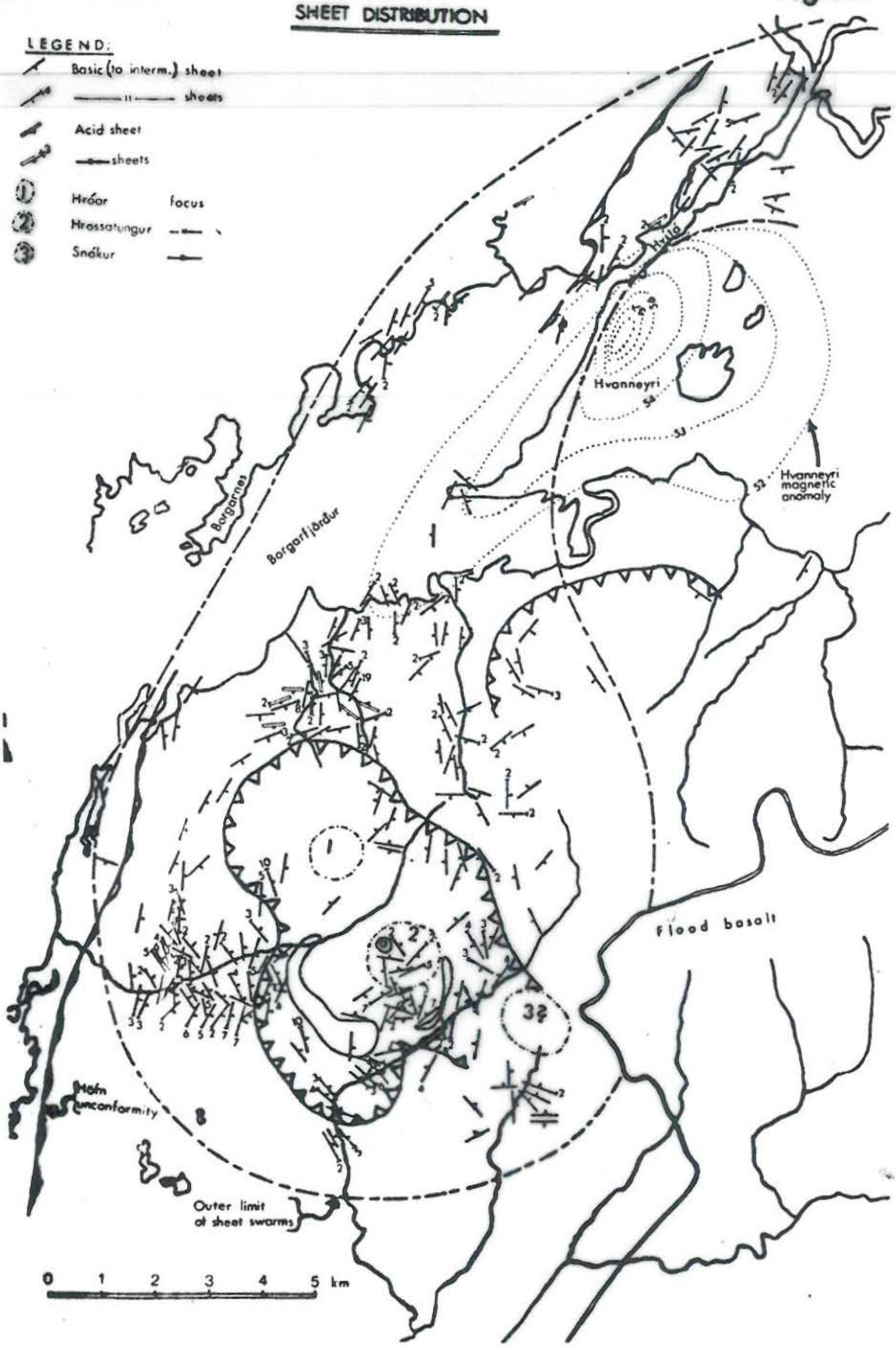


Fig. 26

(Franzson, 1978)

Fig. 27



(Franzson, 1978)

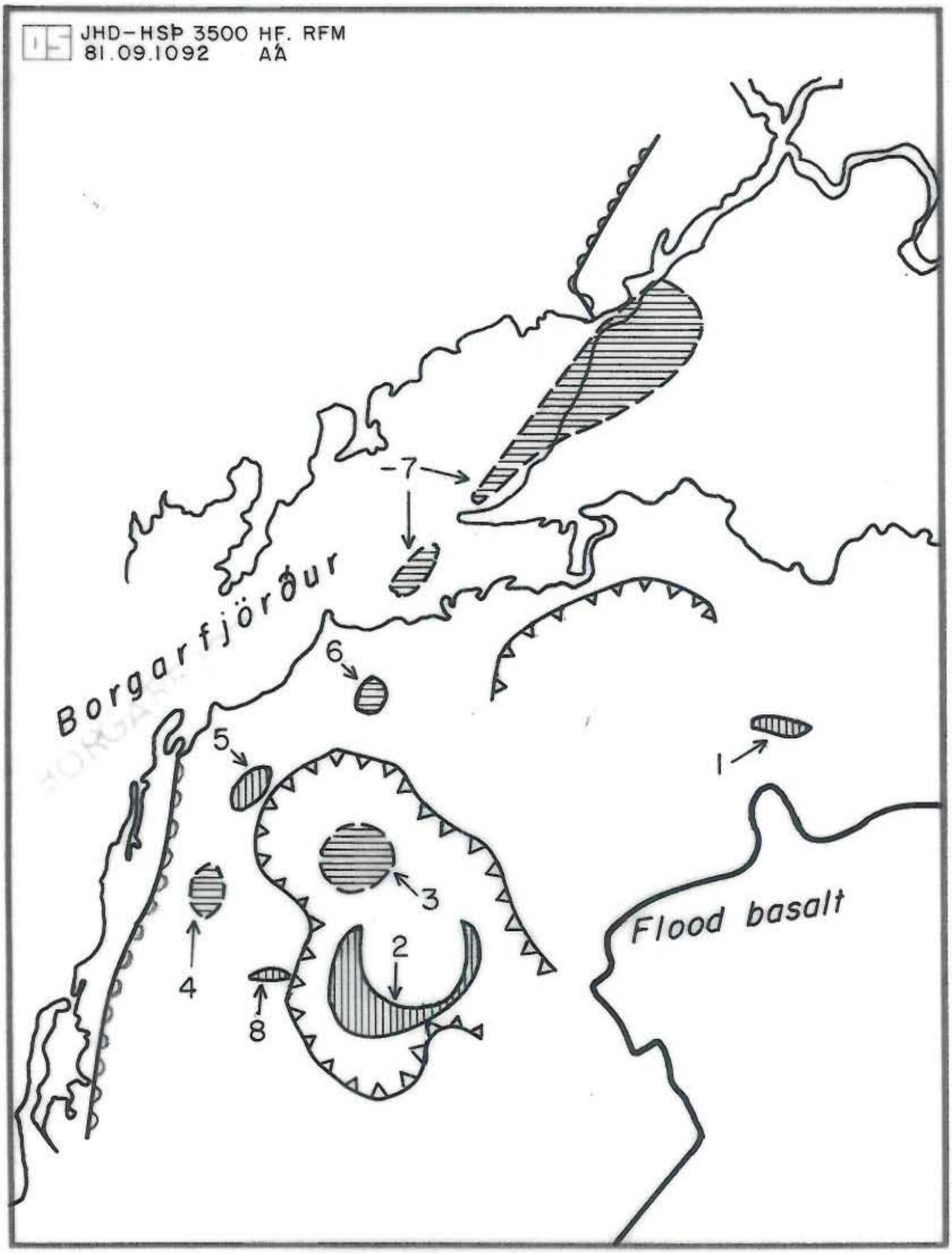


Fig. 28 Approximate location of major intrusions. Vertical lines = exposed. Horizontal lines = inferred intrusion.

1. Skessuhorn gabbro
2. Hrossatungur gabbro
3. Hroar gabbro
4. Inferred gabbro, west Hafnarfjall
5. Flydrur granophyre
6. Tungukollur granophyre
7. Hvanneyri intrusion
8. Hafnardalur gabbro

(Franzson, 1978)

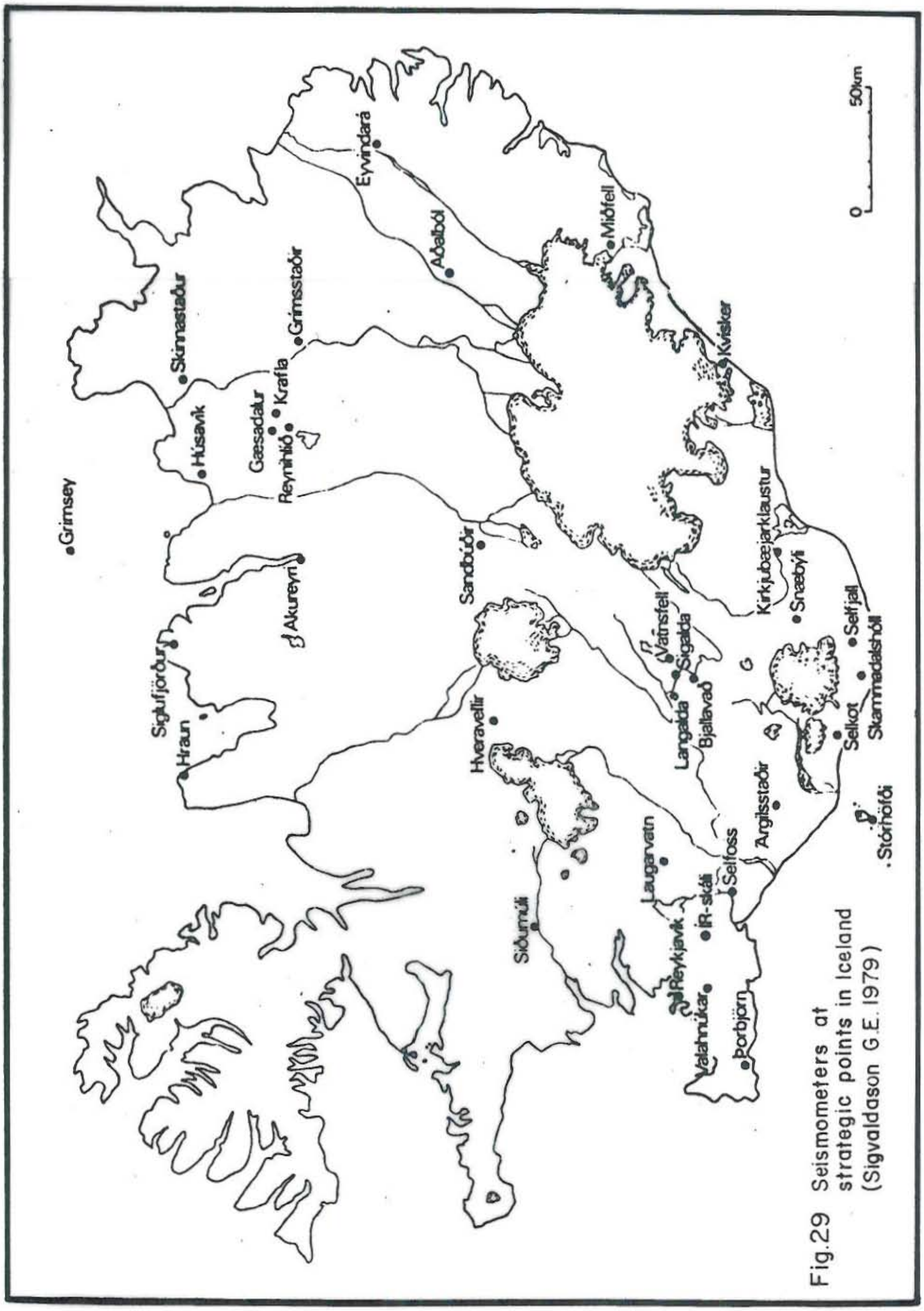


Fig.29 Seismometers at strategic points in Iceland (Sigvaldason G.E. 1979)

Fig.29 Seismometers at strategic points in Iceland (Sigvaldason G.E. 1979)

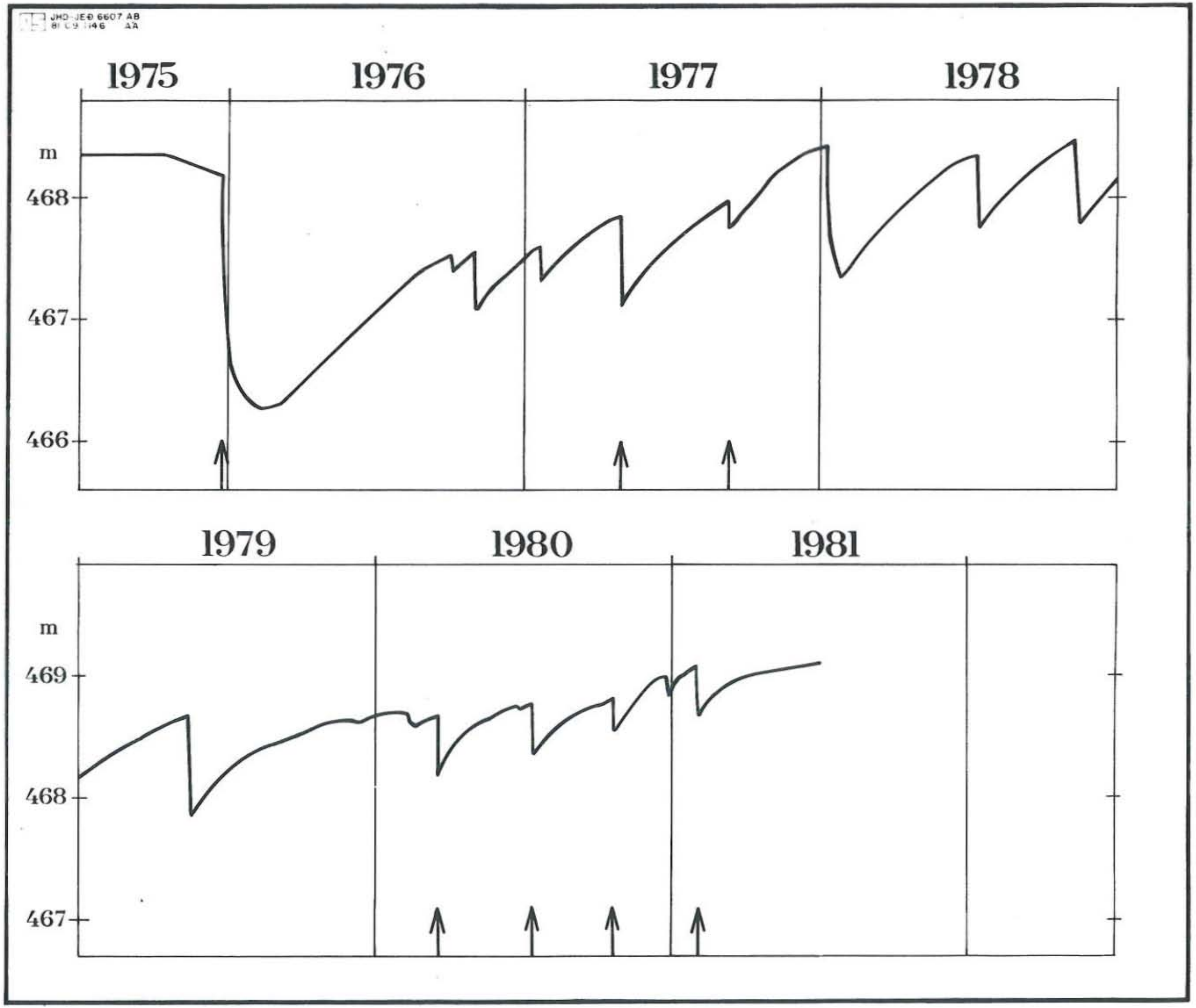


Fig. 30

Changes in land elevation at the centre of the Krafla caldera during the current rifting episode that has been in progress since 1975. Arrows indicate volcanic eruptions.