THERMAL ACTIVITY AND RELATED PHENOMENA IN ICELAND

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ABSTRACT

The thermal activity in Iceland is grouped into two groups, the low-temperature and the high-temperature activity, in accordance with the temperature at the base of the circulation systems. The low-temperature activity includes those thermal areas where this temperature is below 150 °C. The high-temperature activity includes areas with a higher temperature. The concentration of free CO_2 in the spring gases and the concentration of dissolved SiO_2 in the thermal water furnish information about the base temperature. The isotope ratios D/H and O^{18}/O^{16} are indicative of recharge areas and the general pattern of flow. Temperature conditions in near surface layers are studied by the electric resistivity methods. The total heat transported by the low-temperature activity is estimated at 0.2 to 0.3×10^9 cal/sec and by the high-temperature activity at 0.3 to 1.5×10^9 cal/sec. Temperature conditions in 3 non-thermal wells in Iceland are studied and corrected for various effects, mainly the heavy erosion during the Pleistocene. The outward conduction of heat in Iceland appears to be of the order of 3 to 5×10^{-6} cal/sec cm^2 which is 2.5 to 4 times the normal. The abnormal conduction flow appears to be the main source of energy for the low-temperature activity whereas large volcanic intrusives of recent origin appear to supply the high temperature activity. The abnormal conduction flow may be caused partially by large magmatic intrusives in the upper 10 to 20 km under Iceland.

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I. INTRODUCTION

Natural heat has in recent years become a relatively important source of energy in Iceland, Italy and New Zealand. The development in Iceland which was initiated 30 years ago has centered around the utilization of natural hot water and steam for the purpose of domestic and industrial heating. The generation of electricity has been the topic of main interest in Italy and New Zealand.

Around 1/4 of the population of Iceland now lives in houses heated by natural heat, and there are good prospects that this ratio can be increased to 1/2 or even more within the next 10 to 20 years. In Italy, natural steam already generates $2x10^9$ kwh of electricity per year.

The efficient development of this natural resource is possible only on the basis of an understanding of the main geological and physical characteristics of the thermal areas. In all three countries a certain effort has been devoted to the exploration of these areas. Although the basic practical problems are similar, the main lines of research have differed somewhat mainly because of different geological, geographical and economical conditions. The greater variety of thermal phenomena in Iceland has required a somewhat more general approach.

Important studies of thermal activity have been made in the U.S. by Allen and Day (1), and White and collaborators (2)(3). Furthermore, in Mexico by L.de Anda (4), in

Italy by Mazzoni and collaborators (5), and in New Zealand by Grange and collaborators (6).

The work in Iceland has been carried out with essentially practical aims, that is, the aim of deriving general methods of estimating the total capacity, durability and the most economical pattern of exploitation of the individual thermal areas of the country. This work consists of theoretical studies and field exploration of the energy balance and structural control of the individual areas.

The purpose of the present paper is to present some of the more important general aspects of the results obtained so far in Iceland by the writer and under his supervision.

As in the case of most other exploration work, the studies in Iceland consist of an integrating inference from geological and physical data from different sources. Therefore, the first part of the present paper is devoted to those observable characteristics of the thermal areas that are helpful in the deriving of an integrated three-dimensional picture. This includes also an estimate of the total heat flow dissipated by the thermal activity in the country.

The second section of the paper is devoted to a study of the energy balance of the thermal activity and the subsurface temperature conditions in Iceland. An estimate of the total heat flow transported to the surface by conduction is given, and the genetical relation of this factor to the thermal activity is discussed.

The third and last section is devoted to a discussion of some of the problems of the general terrestrial temperature field in Iceland.

In order to facilitate the following discussion, a few remarks will be made about the geology of Iceland, and a few useful definitions will be listed.

Geology of Iceland

There is a considerable literature on the geology of Iceland, but the more recent works of Kjartansson (7), Einarsson (8), and Barth (9), give an adequate picture of the main conditions. Only few remarks will be made here.

Iceland (103,000 km^2), is considered to be a part of the great Thulean flood basalt area, parts of which are found in Eastern Greenland, Northern Ireland, Western Scotland, on the islands in the North Atlantic, and in the adjacent arctic region. With the exception of more recent parts in Iceland and in the surrounding region, most of the Thulean plateau is believed to be of an early Tertiary age.

Following a relatively long period of quiescence after the formation of the main plateau, the volcanism in Iceland resumed its activity probably at the end of the Tertiary. Conspicuous formations have been built up by the late Tertiary and Quaternary volcanism. The post-Glacial activity has been confined to an area which covers only about 1/4 to 1/3 of the total area of the country.

The uppermost section of the basalt plateau of Iceland is composed of lava having a somewhat lighter color than the deeper sections. This section which usually has a thickness of 300 to 400 meters, is usually defined as the grey section. Its top is an erosional peneplane.

The grey section is underlain by a sedimentary section of a variable thickness up to 200 meters. According to Peturs (10), these sediments contain glacial deposits and this author and some later geologists have therefore maintained that the grey section was of a Pleistocene age.

This result has been questioned by Einarsson (11) and others partially on the basis of the extensive denudation and dissection of the plateau. Further recent geological evidence, mainly the finding of thin layers of lignite in the grey section (personal communication by Kjartansson), appears to favor a Tertiary, possibly late-Tertiary age of the grey section. The glacial deposits found by Peturs could possibly be the result of a pre-Pleistocene glaciation.

On the other hand, recent paleomagnetic studies by Hospers (12) and by Einarsson and Sigurgeirsson (personal communication by Einarsson), have revealed a reverse magnetization of the entire grey section. The indications are that the section belongs to the last period of reversed magnetization. In view of the most recent data on paleomagnetism, this could indicate an age of the grey section of one-half to one million years and date the formation back to the Pliocene-Pleistocene boundary.

There are no reliable data on the total thickness of the Icelandic plateau. Earlier estimates of the eastern section amount to 3 km, but more recent data favor higher figures for this part at least. The conditions in the western and southwestern sections are less clear and all that can be inferred from geological observations and shallow drilling is a minimum thickness of 2 km. Geologists generally favor higher figures but without clear evidence.

The plateau of Iceland is heavily dissected and cut by a great number of fractures and faults. The volcanism and the thermal activity are obviously controlled by some of the major faults.

There are definite indications of a pre-Tertiary volcanism in the North Atlantic region. Richey (13) reports that Scotland has been subjected to recurrent volcanism since the beginning of the Paleozoic and Mesozoic volcanism has been found on Spitzbergen.

Definition of the Hydrothermal System

A hydrological system which includes the subsurface percolation and heating of meteoric water will, in the following, be defined as a hydrothermal system.

A hydrothermal system consists of (1) zones of inflow and downward percolation, defined as the recharge area, (2) zone of subsurface percolation and heating, defined as the heating zone and (3) zones of ascent and outflow, defined as thermal areas. The term thermal area will be applied to each individual zone of upflow and outflow. Accordingly, a hydrothermal system may generate many thermal areas.

Systems that are generated by dikes and faults will in the following be defined as linear systems.

Two temperature data are of main importance, that is (1) the maximum temperature of the water percolating in a system and (2) the temperature of the water as it resumes the upward movement into a thermal area. The latter figure will be defined as the base temperature of the thermal area. This figure may be equal to the maximum temperature or otherwise lower. Other important figures are the total flow of water in a hydrothermal system and the total flow that enters a thermal area. The latter figure will be defined as the total flow of the thermal area.

Extending a classification given by White and Brannock (3), thermal systems will be classified as follows in accordance with the nature of the heat source.

Thermal systems will be defined as (I) normal if the heating zone is located within subsurface regions of a normal outflow of terrestrial heat and hence normal temperature, and (II) abnormal if the heating zone is located in regions of abnormally large outflow of heat and abnormally high temperatures.

The second group is divided into three subgroups, (II.1) the non-volcanic abnormal systems where the abnormal conditions are not the result of volcanic activity, (II.2) intermediately volcanic if the abnormal heat flow and temperatures are mainly the result of nearby intrusions of magma and (II.3) direct volcanic if the heat of the circulating water is mainly derived by an intermixture with ascending magmatic water.

On the other hand, the term magmatic water or gas will be used only for matter that is derived directly from molten igneous rock. Water or gas of magmatic origin but confined in solid intrusives and their surroundings, will be defined as igneous water or gas.

Permeability conditions in the basalt plateau of Iceland

The basalt plateau is composed of a great number of lava flows of a thickness ranging from 10 to 50 meters or more. The bulk of each flow, mainly in the lower sections, appears relatively impermeable. A certain small horizontal permeability is provided by the thin layers of scoria at the top and bottom of each lava flow.

The vertical permeability of the unfractured parts of the formation appears negligible. The main vertical passages are furnished by the faults and the dikes which are quite numerous in many parts of the country. These structures which appear to cut through the whole formation, are the main structures controlling the circulation of water.

II. THERMAL ACTIVITY IN ICELAND

Thorkelsson (14)(15)(16)(17)(18), and Barth (9), have given fairly detailed descriptions of the local morphological and chemical characteristics of individual thermal areas in Iceland. Einarsson (11) on the other hand, confines himself to the local activity in northern Iceland, but presents a more comprehensive treatment of the geological and physical problems involved in this part of the country.

The following discussion will center around the more recent results obtained by the writer. The field data applied include those obtained by the writer or under his supervision, as well as a few of the earlier field data.

Earlier authors have classified the thermal springs as alkaline or acid according to the pH data of the water issued. This has led to a more general classification of the entire thermal activity in Iceland into the so-called alkaline and acid activity. This classification,which is based on a local chemical property of the water, appears somewhat confusing and will not be applied in the following. There are, on the other hand, strong geological and physical reasons for a two-group classification.

It appears more suitable to classify the individual thermal areas on the basis of one of the most important quantities involved, that is, the base temperature. There are, of course, some objections to the use of a quantity that is not directly observable, but these difficulties do not appear too serious.

In a great number of cases, there is adequate information for the classification purposes, and the difficulty may furthermore be avoided by the application of a suitable geochemical thermometer.

We are, at this juncture, not so much interested in the temperature itself as in its physical and chemical effects. As will follow from the discussion below, there are definite indications that the base temperature is one of the main factors controlling the leaching of free CO_2 and SiO_2 from the country rock.

The procedure that will be followed in the present paper consists of the application of the concentration of CO_2 in the spring gases and the concentration of SiO_2 in the thermal water as thermometers in the cases where there is no definite information on the temperature itself.

As will be shown below, the rate of leaching of CO_2 is generally negligible at base temperatures below 100°C. The rate becomes noticeable as the temperature exceeds 100°C and increases rapidly with increasing temperature.

The amount of SiO_2 found at base temperatures below 100°C is generally below 120 p.p.m., but at 200°C it has increased to more than 400 p.p.m.

In accordance with these and other observations, the classification will be carried out as follows. Areas where the base temperature is directly observed to be below 150°C will be classified as low-temperature areas. Areas with higher base temperatures are classified as high-temperature areas.

In the cases where the base temperature is not directly observable from well-data or other means, the amount of CO_2 in the spring gases will be used as a thermometer. The upper limit for the low-temperature areas will be put at 5% by volume of CO_2. Where there are no data at hand on the composition of the gas, the amount of SiO_2 may be used as a tentative thermometer, and the limit for low-temperature areas put at 200 p.p.m.

The field data obtained so far appear to indicate that there are only few thermal areas that have base temperatures in the range 150°C to 200°C. This gives a rather clear distinction between the two groups.

The classification is uncertain in some relatively few border-line cases which have to await the collection of further field and laboratory data. On the other hand, it appears consistent with the general geological and geographical distribution of the thermal areas which is a matter of main importance.

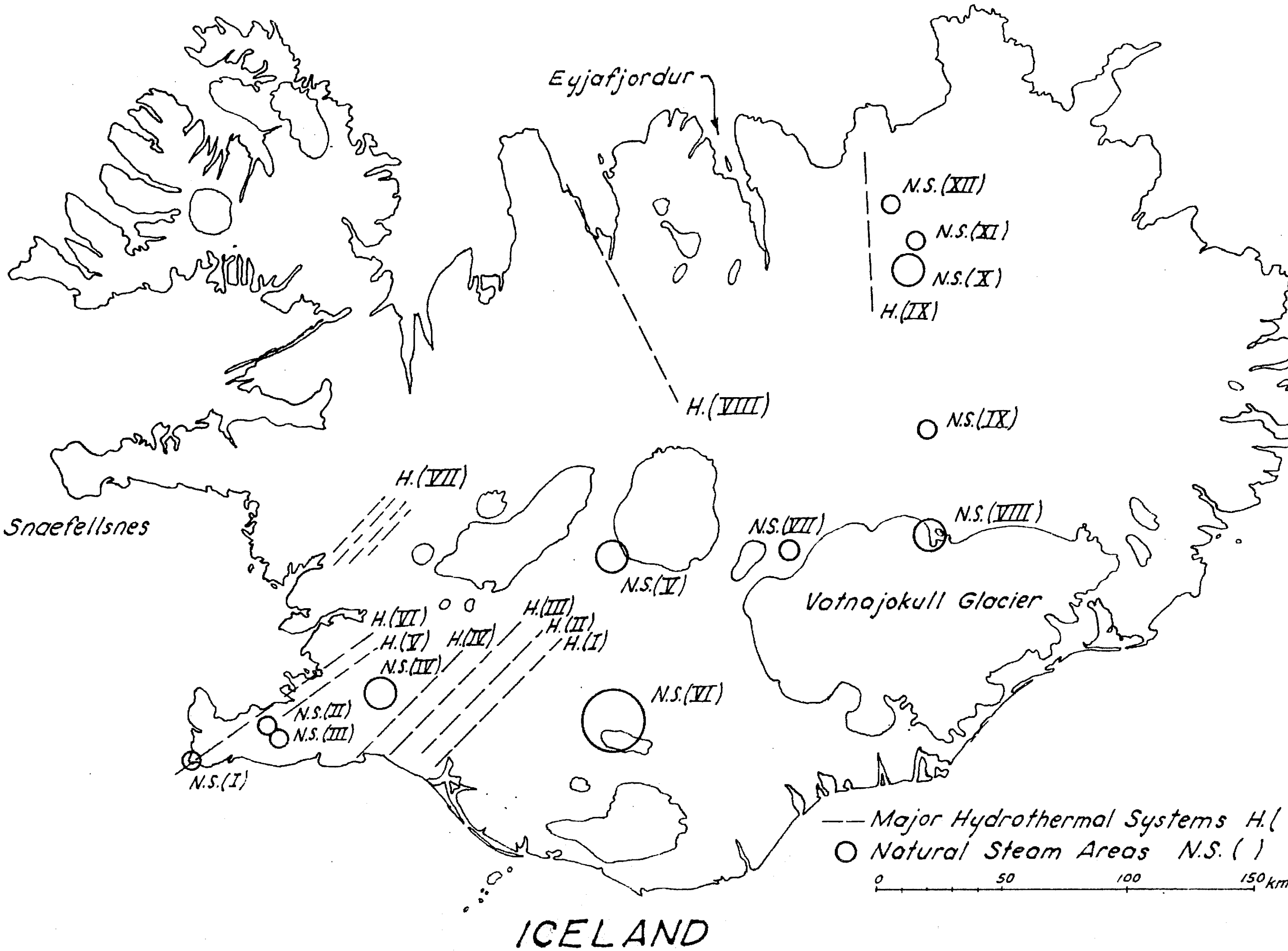

Eyjafjordur
N.S.(XII)
N.S.(XI)
N.S.(X)
H.(IX)
H.(VIII)
N.S.(IX)
H.(VII)
Snaefellsnes
N.S.(VII)
N.S.(VIII)
N.S.(V)
Votnajokull Glacier
H.(VI)
H.(V)
H.(IV)
H.(III)
H.(II)
H.(I)
N.S.(IV)
N.S.(VI)
N.S.(II)
N.S.(III)
N.S.(I)
Major Hydrothermal Systems H.( )
Natural Steam Areas N.S.( )
0
50
100
150 km.
ICELAND

Fig. 1

(II.1) The low-temperature areas

Geography and geology

There are in Iceland around 250 thermal areas that appear to belong to the low-temperature group. Nearly all of them are located in the lowlands and valleys of the southern, western and northern regions of the country. Very few are located in the central regions and only three minor areas are located in marginal areas of the eastern region. The elevation of the areas is generally below 70 meters.

A great part of the major low-temperature areas are apparently connected to 8 or 9 major hydrothermal systems as shown in Fig. 1. These systems will be defined as H(I) to H(IX). Six of these are located in the southern regions, one in the western and one or two in the northern regions of the country. There is probably one additional major system on the northwestern peninsula, but the conditions in this part of the country have not been studied in detail.

H(I) to H(VI) in southern Iceland are characterized by a clear alinement of the individual thermal areas ranging over distances up to 50 km. The systems are parallel to the general tectonic lines in this part of the country.

H(I) and H(II) have apparently no direct connection to post-Glacial or late Quaternary volcanic centra although these lie within a distance of less than 30 km measured perpendicular to the strike.

H(III) and H(IV), on the other hand, run directly through centra of post-Glacial volcanism. Furthermore, the systems H(V) and H(VI) if extended to the southwest, also run through such centra.

The extent and the clear connection to the general tectonic and volcanic lines are strong indications that the hydrothermal systems are controlled by major fault systems of repeated movement and activity. A slight scatter in the alinement of the individual thermal areas indicates that the systems are not controlled by a single major fault, but by a series of close parallel faults.

Except for the thermal activity, there are no major features that distinguish the hydrothermal faults from other major fault systems of the area. A general gravity survey carried out by Einarsson (19, 53) failed to reveal major gravity anomalies associated with these faults. Magnetic surveys by the writer have also failed to reveal general anomalies. These results do not exclude minor anomalies.

H(VII) deviates from H(I) to H(VI) as it appears to be controlled by a swarm of dikes (Einarsson (11). This system generates two of the most voluminous hot springs in Iceland. It has no direct connection to post-Glacial or late Pleistocene volcanism but a small cone of recent activity is located within a distance of 15 km.

H(VIII) in northern Iceland, which extends for about 90 km, is of a similar type as H(I) and H(II). There appears to be no connection to post-Glacial or late Pleistocene volcanism. H(IX) is of a more doubtful nature and appears to be a borderline case.

In addition to the major systems, there are minor hydrothermal systems mostly controlled by individual dikes. This is the common situation in Eyjafjordur in the northern regions.

Surface temperature and flow

According to the measurements made on behalf of the National Research Council of Iceland (20) in 1943 and 1944, the total integrated flow of all low-temperature areas is around 1.5×10^6 cm^3/sec. The total heat transported by the water, measured above 0°C, is around 1.12×10^8 cal/sec giving a mean temperature of 75°C.

The natural flow has been augmented by some 6×10^5 cm^3/sec from drilled wells having a mean temperature of 87°C. This water is generally produced from the depth of 100 to 500 meters.

The total flow, maximum surface temperature and the flow of the largest individual springs of the systems H(I) to H(IX), are listed in the following Table I.

The flow of some springs is not constant and the figures in the table are to be regarded as round figures.

TABLE I

Surface Flow of the Major Low-Temperature Hydrothermal Systems

System	Total Natural Flow	Maximum Surface Temperature	Largest Spring
H(I)	52x10^3 cm^3/sec	100°C	20x10^3 cm^3/sec
H(II)	140	100	65
H(III)	60	100	40
H(IV)	70	100	40
H(V)	120	83	
H(VI)	12	87	10
H(VII)	400	100	250
H(VIII)	70	89	9
H(IX)	145	100	27
Sum	1,069x10^3 cm^3/sec		

There is a remarkable correlation between the surface flow and the temperature of the springs. This holds for the low-temperature activity as a whole, but is probably most pronounced in the case of the many small thermal areas in the northern regions. This is shown in Fig. 2.

The temperature interval from 20°C to 100°C is divided into 8 intervals of 10°C each. The individual thermal areas in the northern regions are then grouped into the intervals according to the maximum surface temperature in each area. The mean value of the total surface flow of the thermal areas in each interval is computed and shown in Fig. 2 as a function of the temperature.

This correlation between the mean flow and the surface temperature of the low-temperature areas suggests that the two phenomena are of importance. Firstly, the thermal water ascending to some of the areas is cooled appreciably by the conduction of heat from the channels of flow. Secondly, some kind of convection is involved at least in the case of the major areas.

Well-Temperatures and Artesian Pressure

Some 300 wells have been drilled in about 40 thermal areas in Iceland. Although only few wells exceed the depth of 500 meters (max. depth 760 m), a considerable amount of temperature data and related information has been collected.

The vertical temperature distribution in the major thermal areas is generally characterized by a boundary layer at the surface with a steep temperature gradient and overlying a section of approximately constant temperature. The boundary layer is due to the effect of the surface on the temperature field and may be a few tens of meters thick. The temperature in the lower constant section is approximately equal to the temperature of the ascending water.

The minor thermal areas are generally much more localized and contain practically no section of constant temperature.

The measurement of the shut-in pressure of flowing hot-water wells has generally revealed an artesian pressure ranging from a few meters up to as much as 50 meters. This pressure appears to be related to the base temperature and increase with the temperature. Cold artesian water has never been observed in wells in Iceland.

The artesian pressure is a phenomenon of great practical importance as the recovering of unconfined hot water by pumping would present difficulties. All interconnections with adjacent channels containing cool or cold ground water would have to be sealed but this may be quite difficult and in some cases a prohibitive task.

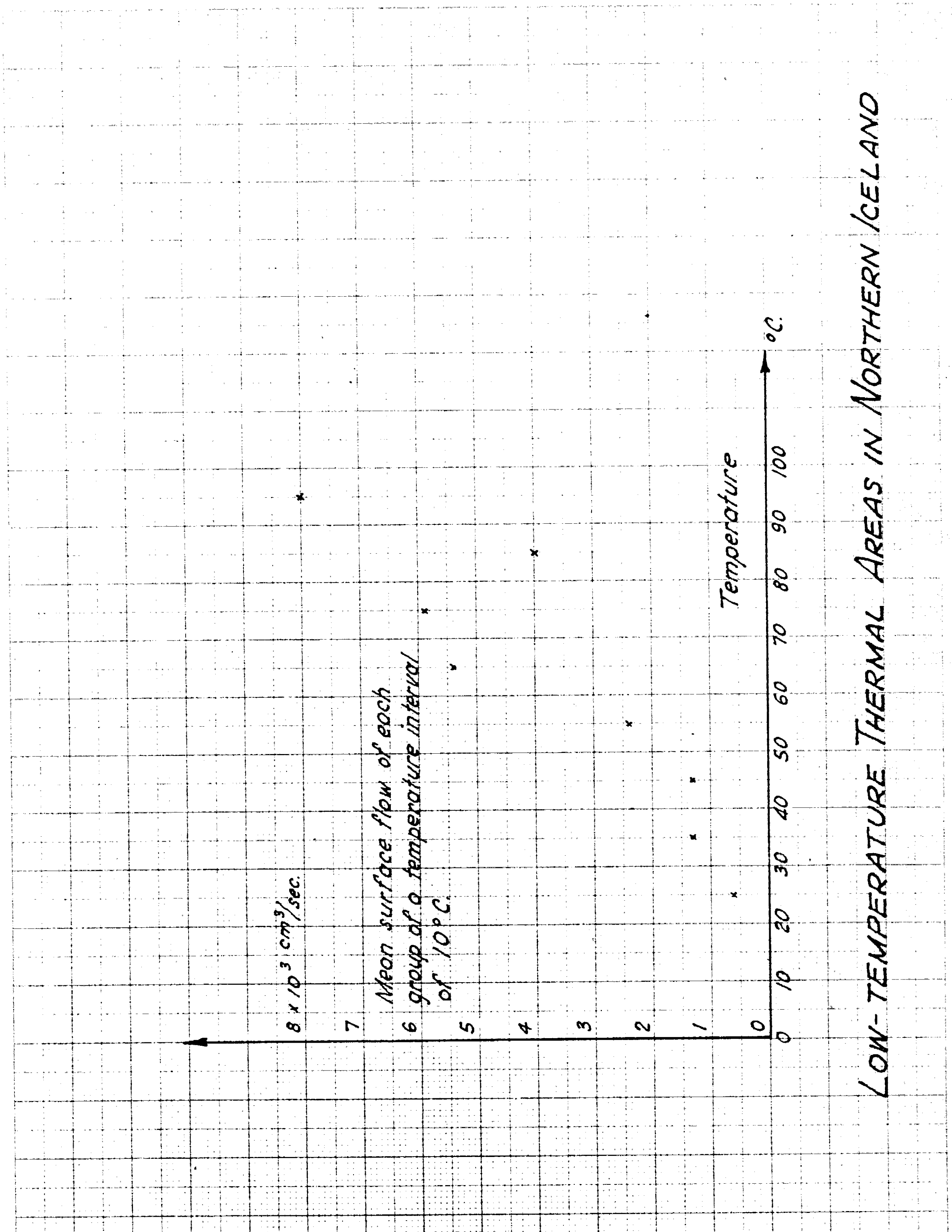
An important observation is that the artesian pressure is generally confined to the basalt plateau and is apparently substantially higher in the early Tertiary parts. The thermal water generally loses the pressure upon entering surface layers of sedimentary or late Quaternary volcanic origin.

The presence of the artesian pressure again suggests that some kind of convection may be involved in addition to a hydrostatic head. Furthermore, that the channels in the basalt plateau through which the water ascends have relatively impermeable walls. The pressure and this condition tends to counteract intermixture with local surface water.

The impression is that the ascending thermal water tends to seal off the channels by depositing silica, zeolites and calcite in the fractures in the walls.

Cooling of the Ascending Water

The water ascending through the basalt plateau and the surface layers suffers cooling in three ways, that is,



LOW-TEMPERATURE THERMAL AREAS IN NORTHERN ICELAND

Fig. 2

by (1) conduction of heat from the channels, (2) intermixture with cold ground water, and (3) self-evaporation in the case where the base temperature is above 100°C.

Cooling by conduction is suggested by the mean-flow-temperature relation as shown in Fig. 2. The temperature increases with the mean flow and the relation appears roughly linear below 60°C.

A very rough quantitative picture (Bodvarsson (21)) of the situation can be obtained in the following way. We will assume a thermal area with the base temperature T_b where the water ascends through some natural channels in the basalt plateau. The normal temperature at the depth x is assumed to be gx where g is the constant gradient. The temperature of the ascending water is T.

We will now make the simplest assumption that the heat lost per unit time and unit depth by the ascending water can be written $k(T - gx)$, where k is assumed to be a constant. This is a simplification for the purpose of estimation, as k will at natural conditions depend on the depth, size and form of the channels and, furthermore, on the amount of leakage from the channels.

If steady state conditions are assumed and c_p is the specific heat of the water and q the flow per unit time, the following differential equation is obtained,

$$c_p q \, dT/dx = k(T - gx) \qquad (1)$$

which by the condition $T = T_b = gH$ at the base depth at $x = H$ has the solution,

$$T = gx + \left(1 - e^{-k(H-x)/c_p q}\right) \frac{g c_p q}{k} \tag{2}$$

The temperature at the surface is found to be

$$T_s = T_b \left(1 - e^{-kH/c_p q}\right) \frac{c_p q}{kH} \tag{3}$$

This result can be simplified in the case where the flow is small and we obtain for small q

$$T_s = gc_p q/k. \tag{4}$$

This implies that the surface temperature of small springs should be independent of the base temperature and be a linear function of the flow. This appears to be in accordance with the results in Fig. 2.

From Fig. 2 and equation (4) we are able to estimate the value of k in nature. By putting $g = 10^{-3}$ $^{o}C/cm$, * we find k = 0.034 cal/cm ^{o}C sec which should represent a rough average for the small springs.

The figure k can also be estimated theoretically on the basis of the hypothesis that most of the cooling is due to the conduction of heat from the channels. Our problem is whether this estimate can be reconciled with the observed figure.

The figure k is easily found in the case of a horizontal pipe of infinite length at the depth h and of the diameter d where d << h. The method of images gives (22)

$$k_o = 2\pi K / \log\left(\frac{4h}{d}\right) \tag{5}$$

* see page 103

where K is the constant thermal conductivity of the earth.

As k_o depends on the logarithm of the large number $4h/d$, it is relatively insensitive to changes in both h and d. Furthermore, as d is much smaller than h, the disturbance of the temperature field due to the pipe is only local. The above expression (5) is therefore also applicable as a rough approximation in the case of the vertical pipe if the depth h is chosen suitably. By using for basalt K = 0.0045 cal/cm °C sec and assuming for d a figure of the same order as the largest horizontal dimension of the channels, that is $d = 2.5 \times 10^3$ cm and furthermore, the effective mean depth $h = 2.5 \times 10^4$ cm, we find approximately $k_o = 0.008$ cal/cm °C sec. The figure k is thus roughly 4 times k_o.

There is no doubt that other factors than the conduction of heat may influence the figure k. The following factors tend to increase k above the estimated k_o. Firstly, there are no doubt areas of a low base temperature but a relatively large flow. Secondly, some of the springs may be cooled by the intermixture with surface water just below the spring vent. Furthermore, some areas may be connected to more than one channel and the channels may deviate considerably from the vertical. Finally, a transient effect could be of importance in areas that have undergone recent changes, for example, an increase in flow.

On the other hand, loss of water in near surface layers tends to decrease k. This phenomenon is probably mainly confined to areas of high artesian pressure, that is, areas of higher temperature and is therefore not to be expected to be a matter of main importance in the present case.

A separation of the various factors and a quantitative correction of the observed k is not possible. But it is a reasonable suggestion that the discrepancy between the observed figure and the semi-quantitative estimate can be understood on the basis of the various other factors that tend to lead to a higher apparent rate of cooling of the ascending water. It will therefore be expected that the above result lends support to the hypothesis that cooling by conduction is of importance in the case of some of the minor thermal areas.

The result also shows that comparatively large flows, that is, flows more than 25×10^3 cm^3/sec should be unaffected by conduction of heat from the channels. From this we may also infer that high temperature in an area of relatively small flow indicates that thermal water is lost in near surface layers. An important point is that cooling by conduction does not alter appreciably the chemical or isotopic composition of the thermal water.

On the other hand, cooling of thermal water by the intermixture with local ground water is mainly encountered in surface layers of sedimentary origin. This process alters the chemical and isotopic composition of the thermal water.

The third way of cooling, that is, cooling by the release of pressure is encountered in thermal areas where the base temperature exceeds 100°C. Self-evaporation of the ascending water is initiated at the depth where the temperature equals the boiling point. From this depth on up to the surface, the temperature of the water will be equal to the boiling temperature provided that cooling by other factors is not encountered.

Chemical Characteristics. Thermal Water in Normal Low-Temperature Areas

Data on the composition of the thermal waters of Iceland have been collected by the writer and his collaborators. Some 400 samples of water have been analyzed for their main components.

The following analytical data (1) to (3) in Table II have been selected as typical for the normal low-temperature areas. Sample (4) represents a typical composition of the surface water in the plateau basalt areas.

Sample (1) is taken from the largest spring on H(VII). Samples (2) and (3) are from small dike-controlled thermal areas in Eyjafjorour in Northern Iceland. Sample (4) was collected from a small surface stream in the same region.

The topic of main interest is the dependence of the composition on the temperature and geological factors.

The total mineral content of the thermal water is relatively low and exceeds 500 p.p.m. in rare cases only. Na is the dominant cation and K is generally very low or absent. The cations Ca and Mg are low and relatively lower than in the surface water.

TABLE II

Composition of Thermal and Surface Water in Iceland

(Analyst: Hermannsson)

	(1)	(2)	(3)	(4)
Temperature of the spring, ^{o}C	100	75	30	5
Flow of the spring, cm^3/sec	250×10^3	1.5×10^3	0.5×10^3	--
Specific resistance, Ohm m (at $25^{o}C$)	26	37	66	69
pH	9.0	9.3	9.3	7.5
Total solids, p.p.m.	361	275	140	105
Na	79	55	30	14
K	4	0	0	0
Ca	5	6	3	14
Mg	1	2	1	5
Fe	0.25	1.2	0.2	0.2
Cl	48	13	8	15
SO_4	60	49	12	10
F	1.3	0.7	0.3	0.2
SiO_2	128	110	45	27
Primary alkalinity, millival/l	0.80	0.80	0.80	0
Secondary alkalinity	0.70	0.75	0.70	1.15

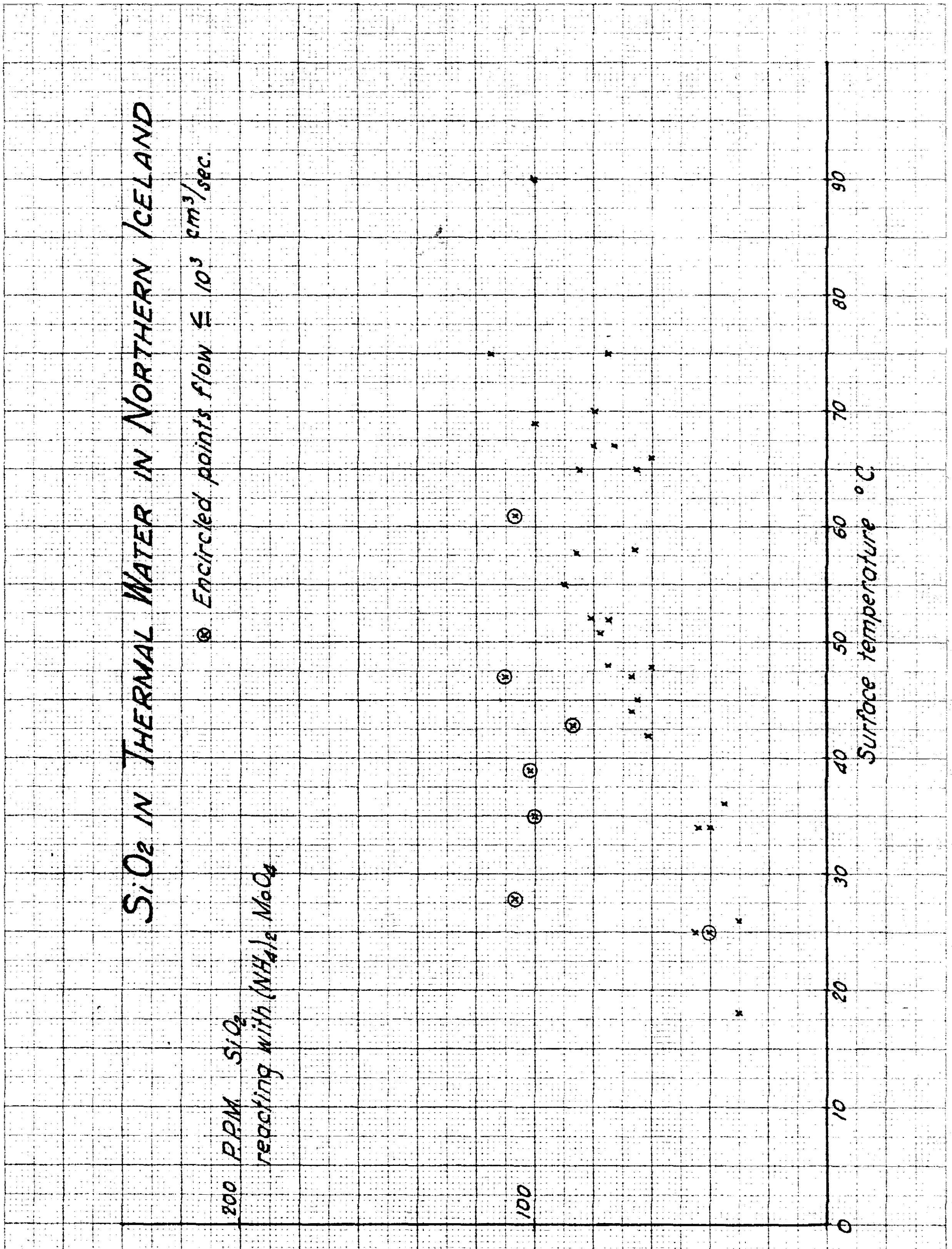

SiO2 IN THERMAL WATER IN NORTHERN ICELAND
⊗ Encircled points flow ≤ 10^3 cm^3/sec.
PPM SiO2
reacting with (NH4)2 MoO4
200
100
0
10
20
30
40
50
60
70
80
90
Surface temperature °C

Fig. 3

The anions are of a greater interest. Table II indicates quite clearly that the amount of SiO_2 depends on the surface temperature. This is the general case for all thermal areas.

This relation is illustrated further in Fig. (3) which gives the correlation of SiO_2 to the surface temperature in the case of the low-temperature areas in northern Iceland. The northern areas were selected because of their geological uniformity. In other areas, water of a surface temperature of 90°C to 95°C contains 100 to 130 p.p.m. of SiO_2.

The solubility of silica in water has been studied by Kennedy (23), Krauskopf (24) and White, Brannock and Murata (25). The experimental data of Krauskopf show that the solubility of amorphous silica in distilled water amounts to approximately 70 p.p.m. at 0°C and 350 p.p.m. at 90°C. Equilibrium is reached within a month or two. The crystalline forms of silica have lower solubilities.

The data in Fig. 3 suggest that the silica in the thermal waters of the low-temperature areas has been leached from the country rock, and that a quasi-equilibrium has been reached in a great number of cases. The ultimate solubility of silica from the basaltic minerals appears to be 25% to 35% of that amorphous silica.

The possibilities of a quasi-equilibrium are underlined by the fact that the Tritium analysis of a few samples of thermal water from low-temperature areas in Iceland have indicated a high subsurface "age".* The Tritium studies were initiated by Dr. D. E. White and the Institute for Nuclear Studies at the University of Chicago.

A few areas have abnormally high silica, that is, the encircled points in Fig. 3. These areas are all characterized by a relatively small flow of 10^3 cm^3/sec or less, which suggests that the thermal water has been cooled considerably by the conduction of heat from the channels of ascent. The base temperature of these areas should be considerably higher than the surface temperature which should explain the abnormally high silica.

The main result is that the amount of silica in the thermal water appears to depend critically on the base temperature of the areas and that a quasi-equilibrium is possible at least in a number of cases. This fact suggests the use of the amount of silica in the thermal water as a geochemical thermometer in order to estimate the base temperature.

The data in Fig. 4 show a definite correlation between the surface temperature and the amount of fluorine in the thermal water. These results suggest a leaching of the F anion from the country rock and the use of the F data as a geochemical thermometer. The amount of F in basalts has been found to be 200 to 500 p.p.m. (Correns,26) but the process of dissolving this component has not been studied in detail.

*see page 40

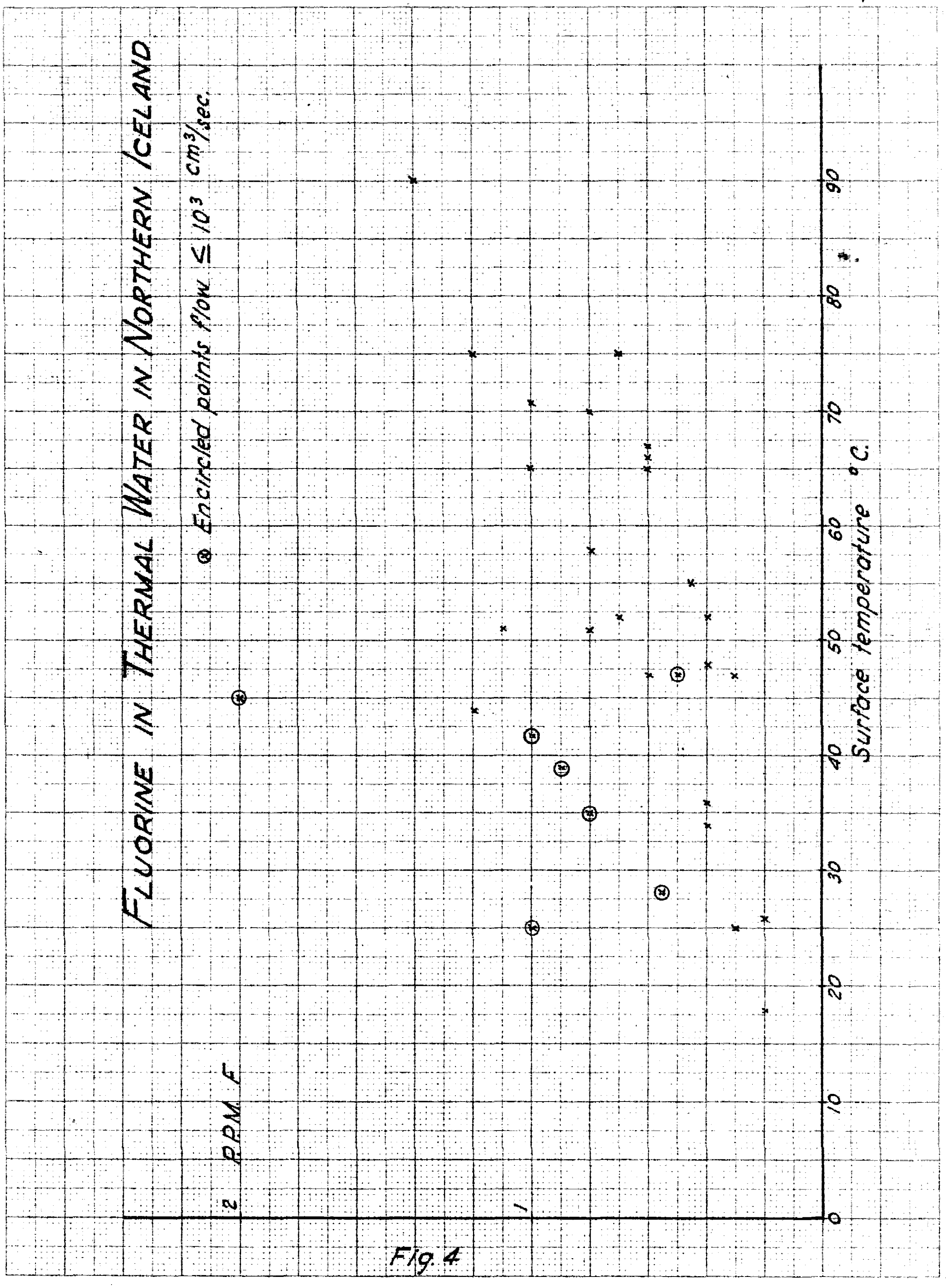

FLUORINE IN THERMAL WATER IN NORTHERN ICELAND
⊛ Encircled points flow ≤ 10³ cm³/sec.
PPM F
2
1
0
10
20
30
40
50
60
70
80
90
Surface temperature °C.

Fig. 4

There are, furthermore, some indications of a correlation of the amount of the anions Cl and SO_4 to the surface temperature and the base temperature. The correlation is, however, less clear than in the case of the silica.

According to Faber (Rankama and Sahama (27)), basalts from the Faroe Islands contain about 0.2% by weight of intercrystalline water that contains about 10% by weight of NaCl and Na_2SO_4. The leaching of these components is not excluded but the process has not been studied in detail.

Gases in normal low-temperature areas

There are traces of gases issued by most thermal areas in Iceland. The composition of these gases has been studied in detail by Thorkelsson (14 to 18). His results as to the hot water springs are given in the following Table III.

TABLE III

Composition of Gases from Hot-Water Springs in Iceland

(Analyst Thorkelsson)

Surface temperature	N_2 vol %	CO_2 vol %	CH_4 vol %	H_2S vol %
below 90°C	≥ 99.8	≤ 0.2	0	0
90°C to mild boiling	95.0 to 99.7	0.3 to 4.6	≤ 0.5	≤ 0.5
violent boiling	≤ 30	≥ 70	~ 0.5	~ 2

Thorkelsson has not measured the relative abundance of the gases in the thermal water. This has been carried out by Einarsson (11) in the case of a number of low-temperature springs in the northern regions. He found figures of 0.o1 to 0.o2 cm^3gas/cm^3 water.

The present writer suggests that the CO_2 in the thermal water is the result of the leaching of igneous or bound CO_2 from the oountry rock and that this process is strongly temperature dependent.

Shepherd (28) has found that igneous rocks in general contain free gases which are expelled when samples of the rock are heated in vacuo. Hawaiian basalts were found to expel 2 to 5 cm^3 of gases/gm of rock. About 80% by volume of these gases was found to be H_2O, 5% to 15% CO_2, and the rest being composed of CO, H_2, N_2, O_2, A, S_2, Cl_2 and F_2.

The writer has, in collaboration with Mr. Baldur Lindal, chemical engineer of the State Electricity Authority in Reykjavik, studied the content of free CO_2 in Icelandic basalts. A series of experiments was performed where about 1000 gm of crushed basalt of late Pleistocene age intermixed with 1,750 gm of water was heated in an autoclave. Each filling of the autoclave was heated at a constant temperature for 15 minutes.

At the end of the heating period, a sample of steam was taken from the apparatus and the amount of free CO_2 measured by volumetric analysis. The analytical work was carried out by Mr. Lindal.

These experiments have as yet not been completed, but the following results in Table IV can be listed.

TABLE IV

Free CO_2 Expelled by Heated Basalt

Temperature during the heating period of 15 min.	$CO_2 cm^3$/gm of rock
50^oC	0
100	0.053
150	0.093
200	0.155
250	0.273

The last figure in Table IV is of the same order as the data given by Shepherd.

The data given were obtained on fairly recent basalts and have, therefore, only a semi-quantitative significance. But it is demonstrated quite clearly that appreciable amounts of free CO_2 can be present in basalts and this may very well account for the free CO_2 issued by the low-temperature activity. In fact, the results in Table III and Table IV are parallel.

The gas is probably adsorbed on crystal surfaces or occluded in intercrystalline spaces. The expulsion is probably to a certain degree the result of a boiling of the water with which the gas is intermixed.

The main result is that free CO_2 cannot be expected in springs having a base temperature below 100^oC. Substantial amounts can be expected only when the boiling point of water is reached down to a certain depth. For example, at a base temperature of 150°C, the boiling point of water is reached only down to a depth of 40 meters. At a base temperature of 250°C, the depth is 400 meters.

The amount of CO_2 issued by the springs may thus, in normal cases, be applied as a chemical thermometer for the base temperature. This has led to the present classification of the thermal activity where a base temperature of 150°C has been selected as an upper limit to the low-temperature activity.

Chemical Characteristics. Abnormal Areas.

There are three types of deviations from the above conditions.

Firstly, a number of hot-water springs are located near to the shore and are obviously contaminated by sea water. The contamination, may in cases, amount to 500 to 1,000 p.p.m. of Cl. The pattern of contamination is quite different in the different regions of the country.

The water issued by H(I) and H(II) is contaminated by 100 to 200 p.p.m. of Cl as far inland as 30 to 40 km. H(VIII), on the other hand, shows practically no contamination.

There are, furthermore, hot-water springs on islands in the bay Breidifjordur in western Iceland as far as 30 km from the shore. Although the water issued is contaminated by 200 to 800 p.p.m. of Cl, there is evidence that this is a surface phenomenon as the springs are submerged in high tides. In any case, it is apparent that the thermal water has percolated through channels in the bottom of the bay over a distance of more than 30 km without an appreciable contamination.

This difference between the northern and the southern regions is probably related to the fact that the basalt formations at sea level in the northern regions are of a much earlier origin than those at the same level in the southern regions. The difference in age results in a different state of cementation of fractures and fissures. The older formations are less permeable.

Secondly, a much less understandable abnormality is observed at the inland ends of H(I) to H(IV) and H(VIII). The innermost springs on each system issue water that contains 50 to 100 p.p.m. of Cl and SO_4 in excess the normal.

Finally, there is a number of relatively small cold springs on the peninsula Snæfellsnes in western Iceland, that issue as much as 2 cm^3 of free CO_2/cm^3 of water. A number of these springs are located around the volcano Snaefellsnesjokull which has erupted silicic-andesitic material in the post-Glacial period.

A phenomenon probably of a similar nature was observed during the recent eruption of the volcano Hekla in 1947-1948. The ground water immediately south of the volcano became contaminated by CO_2 for a period of a few years.

This suggests that these small cold springs are possibly outlets of magmatic CO_2.

Isotope Ratios

Dr. Harmon Craig (29), Scripps Institution of Oceanography, and Dr. Don E. White, U. S. Geological Survey have initiated a study of the isotopic composition of natural thermal water. Samples from one high-temperature thermal area in Iceland, that is NS(IV) in Table VIII were collected at the request of Dr. Craig.

Additional samples were collected by the present writer and were analyzed by Dr. Craig. The results on these samples are given in Table V where also one of the results (No. 12) from NS(IV) is included.

Furthermore, the Tritium content of a few samples of surface and thermal water from Iceland has been studied by Dr. F. Begemann, University of Chicago. These samples were collected by the writer.

The location of the samples (1) to (9) in Table V which were collected from H(V) and H(VI) in the Reykjavik area, is shown in Fig. 5. The Table V furthermore includes a sample (No. 11) of thermal water from a hot spring in the Eyjafjordur area in northern Iceland. The location of sample (12) is also shown in Fig. 5.

Table V gives the deviations of the D/H and O^{18}/O^{16} ratios from a standard which gives for ocean water a D/H deviation of 4.5% and a O^{18}/O^{16} deviation of 0 o/oo. The T/H is given in the usual unit of 10^{-18}.

The table furthermore includes the SiO_2 and F content of the thermal water. These components were determined in Reykjavik.

The D/H and O^{18}/O^{16} data are plotted in the usual way in Fig. 6.

Samples (3) to (7), (9), (11) and (12) are taken from drilled wells. The flow data given are round figures.

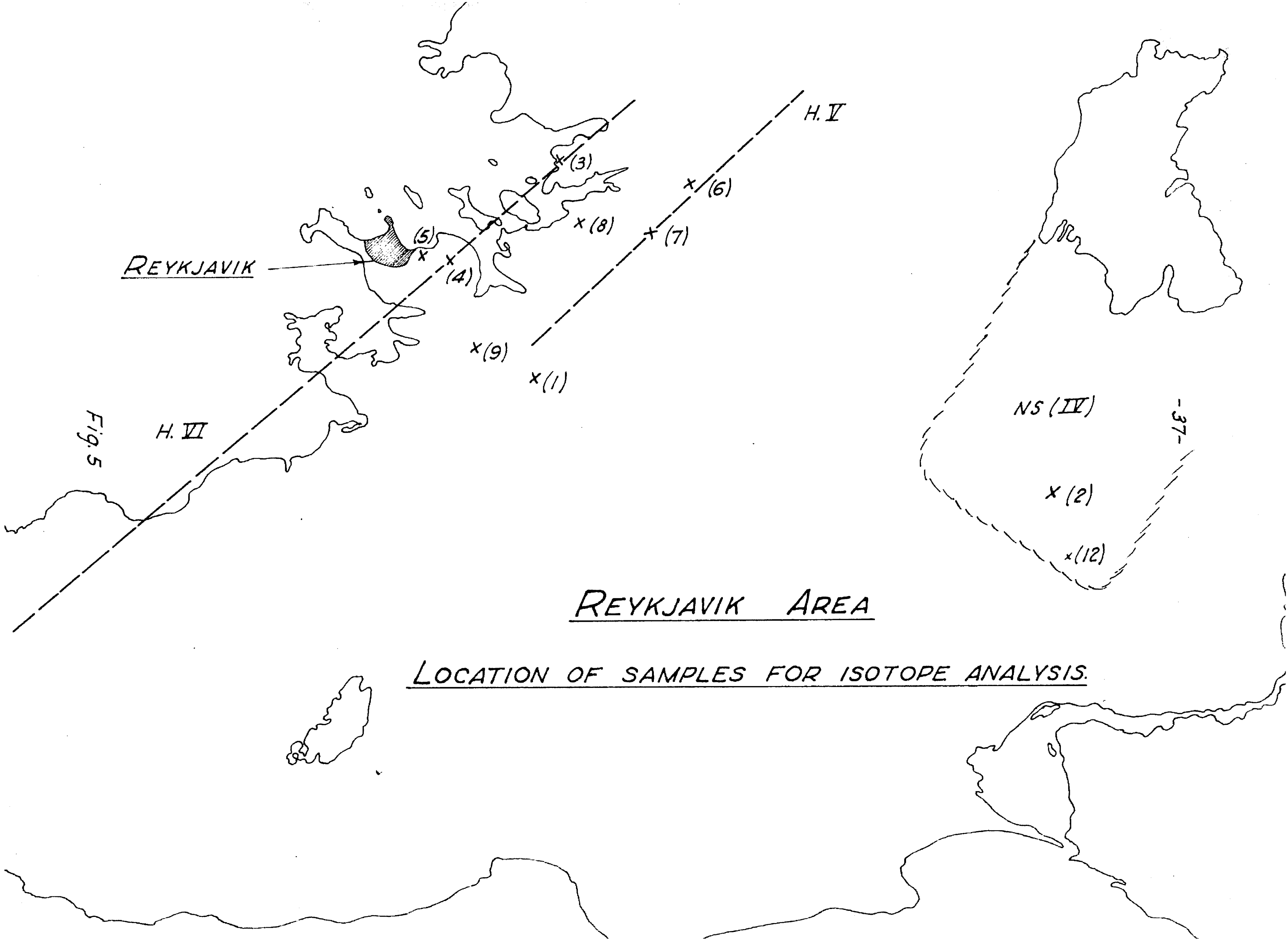
Two important general facts are suggested by the isotope data. Firstly, as pointed out by Craig (personal communication), the data indicate that the thermal water studied is of purely meteoric origin and contains apparently negligible or no magmatic component. This is in accordance with the results of Craig, Boato and White (29) in the case of similar thermal areas in other regions of the world. Furthermore, it is in accordance with the results of Einarsson (11) and the writer (Bodvarsson (30)) founded on different reasoning. The somewhat abnormal behavior of sample (12) is believed to result from an exchange of oxygen isotope with the rock.

Secondly, the data suggest that the main recharge area of H(V) and H(VI) does not lie in the region east and southeast of the Reykjavik area. This, again, indicates

TABLE V

Isotope Ratios in Natural Water from Iceland
Isotope Analysis: Dr. Harmon Craig and Dr. F. Begemann

Location	Temp.	Flow x 10^{-3}	δD	δO^{18}	F	SiO_2	$T/H \times 10^{18}$
	°C	cm^3/sec	%	°/oo	p.p.m.	p.p.m.	
Reykjavik area							
Surface water							
(1) River Ellidaar			-1.4	- 7.9	0	10	
(2) River Varma			-1.7	- 8.0			
Thermal water							
(3) Alfsnes	30	1	-2.8	- 9.9	1.1	87	
(4) Laugarnes	87	15	-2.7	- 9.8	1.o	125	1.1 ± 0.1
(5) Raudara	91	3	-2.6	- 9.5	1.1	130	
(6) Nordur-Reykir	87	180	-2.7	- 9.6	1.1	91	
(7) Sudur-Reykir	87	160	-2.4	- 9.1			
(8) Blika	22	0.2	-2.5	- 9.3	0.2	22	
(9) Breidholt	30	1	-2.1	- 9.o	0.5	63	
Eyjafjordur area (Kristnes)							
(10) Surface water			-4.8	-11.2			18.4 ± 1
(11) Kristnes	75	1.5	-7.o	-13.7	0.7	110	6.1 ± 0.4
Hveragerdi area							
(12) Well	100		-2.4	- 6.9	1.1	340	3.5 ± 0.3



REYKJAVIK AREA

LOCATION OF SAMPLES FOR ISOTOPE ANALYSIS.

Fig. 5

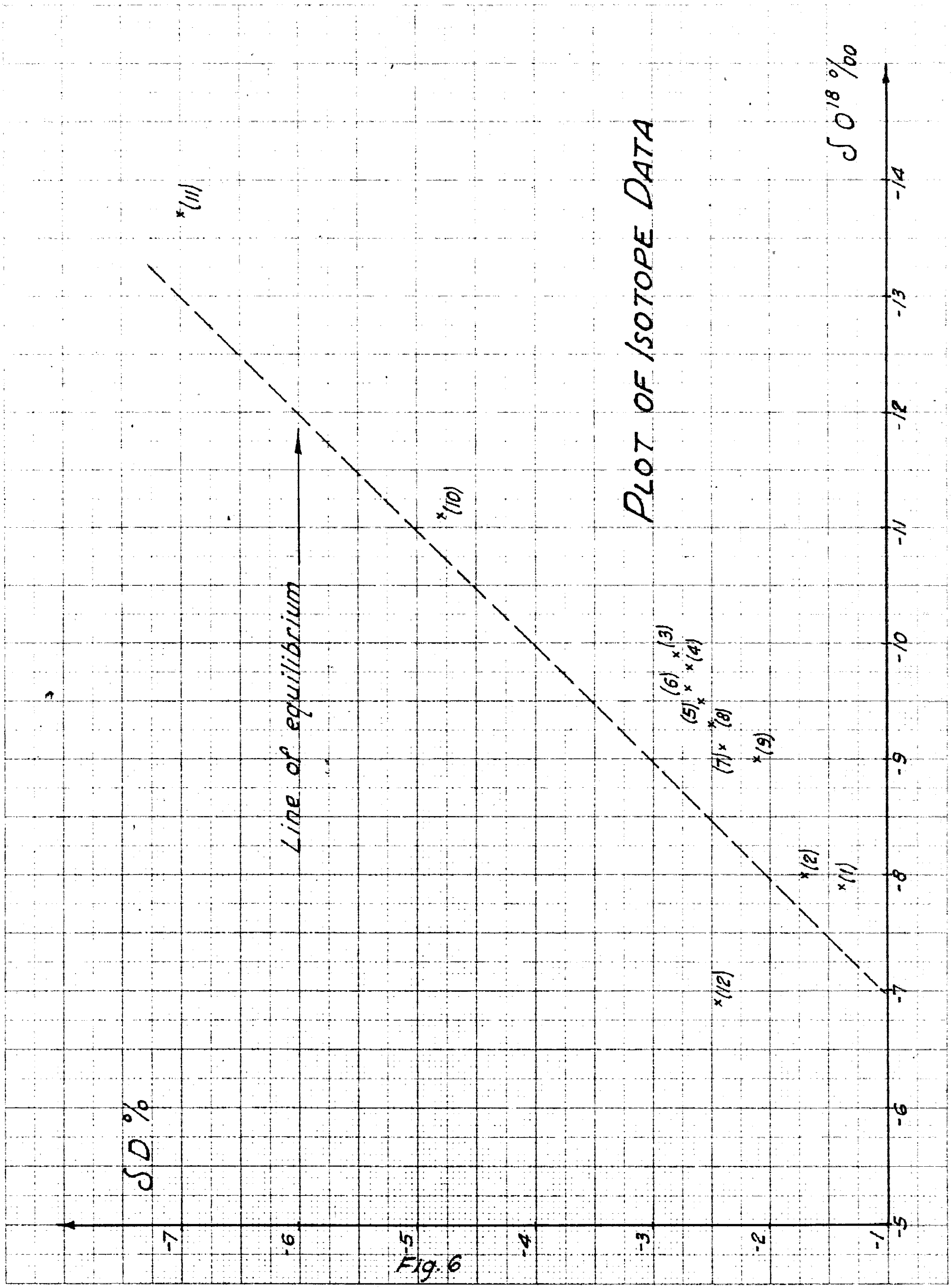

PLOT OF ISOTOPE DATA
Line of equilibrium
δD %
δO18 ‰
-1
-2
-3
-4
-5
-6
-7
-5
-6
-7
-8
-9
-10
-11
-12
-13
-14
(1)
(2)
(3)
(4)
(5)
(6)
(7)
(8)
(9)
(10)
(11)
(12)

Fig. 6

that the main flow is along the linear hydrothermal systems and that only relatively minor quantities of water enter the systems from the direction transverse to the strike. The small difference between the samples (3) and (5) indicates that a small amount of local water may have entered H(VI) between the locations.

This is in accordance with the results obtained by the drilling operations on H(V). Successful drilling at location (6) decreased the flow from wells at location (7).

An important result of local nature is obtained, that is, an indication whether the small spring at location (8) belongs to H(V) or H(VI). The isotopic and chemical data suggest that this spring is the outlet of an individual minor shallow system with a very low base or maximum temperature which draws on the same recharge area as H(V) and H(VI).

On the other hand, sample (3) appears to be water from H(VI) cooled by the conduction of heat from the channels of flow.

The results on the Tritium content are no doubt influenced by the Hydrogen-bomb test in 1954. This is indicated by the high T/H in sample (10). The pre-bomb level was probably around 5×10^{-18} but this is, of course, an uncertain figure. Based on this, there is some suspicion that sample (11) is slightly contaminated by local surface water which is possible as the sample was collected from a spring.

On the other hand, the T/H ratio of sample (4) is quite low. This sample is collected from a system of wells which produce hot water from the depth of several hundred meters. Based on a pre-bomb T/H of 4.4×10^{-18} and a half life of T of 12.5 years, the subsurface "age" of this sample is found to be 25 years which is remarkable.

Electric resistivity in low-temperature areas

The conductance of electricity in shallow igneous and sedimentary rock results in most cases from moisture in intercrystalline spaces and pores. At ordinary temperatures and pressure, dry silicates are almost complete insulators.

These conditions have been studied by the writer on basaltic drill cores that have been dried in a drying chamber.

These conditions imply that the conductivity of ordinary rock will have the same characteristics as the conductivity of natural water, that is, be approximately a linear function of the amount and concentration of the intercrystalline water and be approximately proportional to $\exp(-a/T)$, a being a constant and T the absolute temperature.

The writer has collected a number of data on the conductivity of surface and thermal water in Iceland. The data show that the SiO_2 in the water is not in ionic solution and is therefore an inactive component. Furthermore, to a usable approximation, the conductivity of normal thermal water in the temperature range from 0°C to 100°C can be expressed by the following equation (6) which is based on Walden's rule (see Moore (75), p. 434)

$$C = 4.6 \times 10^{-2} m \cdot \exp(-1,680/T) \tag{6}$$

where c is the conductivity expressed in $(\text{Ohmmeters})^{-1}$ and m is the amount of solids, less the SiO_2, expressed in p.p.m.

The writer has furthermore collected a great number of field data on the conductivity of rocks in Iceland, both at normal temperature conditions and in thermal areas. The measurements were made with a Gish-Rooney equipment and with the application of the Wenner 4-electrode configuration. (See Dobrin, (31)). The results as to normal areas and depths down to about 200 meters are as follows in Table VI.

TABLE VI

Earth conductivity at normal conditions in Iceland

Normal ground water at $5^{\circ}C$	4 - $7 \times 10^{-3} (\text{Ohm m})^{-1}$
Tertiary basalts	4 - 6
Quaternary basalts	1.5 - 2.5
Quaternary basaltic tuffs	1.5 - 2.5
Late Quaternary grey basalts	0.4 - 1.0
Post-Glacial lava below ground water	0.2 - 0.5
Post-Glacial lava above ground water	0.05- 0.2

An interesting feature of these data is the obvious dependence of the conductivity on the age of the igneous rock. This is probably the result of a progressing metamorphism.

In thermal areas, the temperature has a decisive influence on the conductivity, both through the exponential factor in equation (6) and also through the greater degree of metamorphism.

Due to the increased ion mobility, a rise in the temperature from 10°C to 100°C implies an increase of the conductivity by approximately a factor of 4. This, combined with thermal metamorphism, may bring about an overall increase of the conductivity of the rock by a factor of 10 or 20. Higher factors are encountered at higher temperatures.

This pronounced increase of the earth conductivity is a characteristic of all thermal areas and is a factor of great practical importance. It opens the exploration of the subsurface temperature by the means of the various electric methods of exploration.

The electric resistivity method with the conventional Wenner or Lee electrode configurations are best adopted for the study of temperature anomalies of shallow nature, that is, anomalies in the uppermost 200 to 300 meters.

The writer has carried out a number of resistivity surveys of this nature. A typical example of a useful result is revealed in Fig. 7. This figure gives the result of a resistivity survey in a thermal area located at an elevation of 4 meters at the lower end of H(VIII) in northern Iceland.

The basaltic basement in this area is covered by some 40 meters of sand and clay. This cover is permeable enough that hot water ascending from fissures in the basement is dispersed and mixed with ground water without leaving significant activity at the surface.

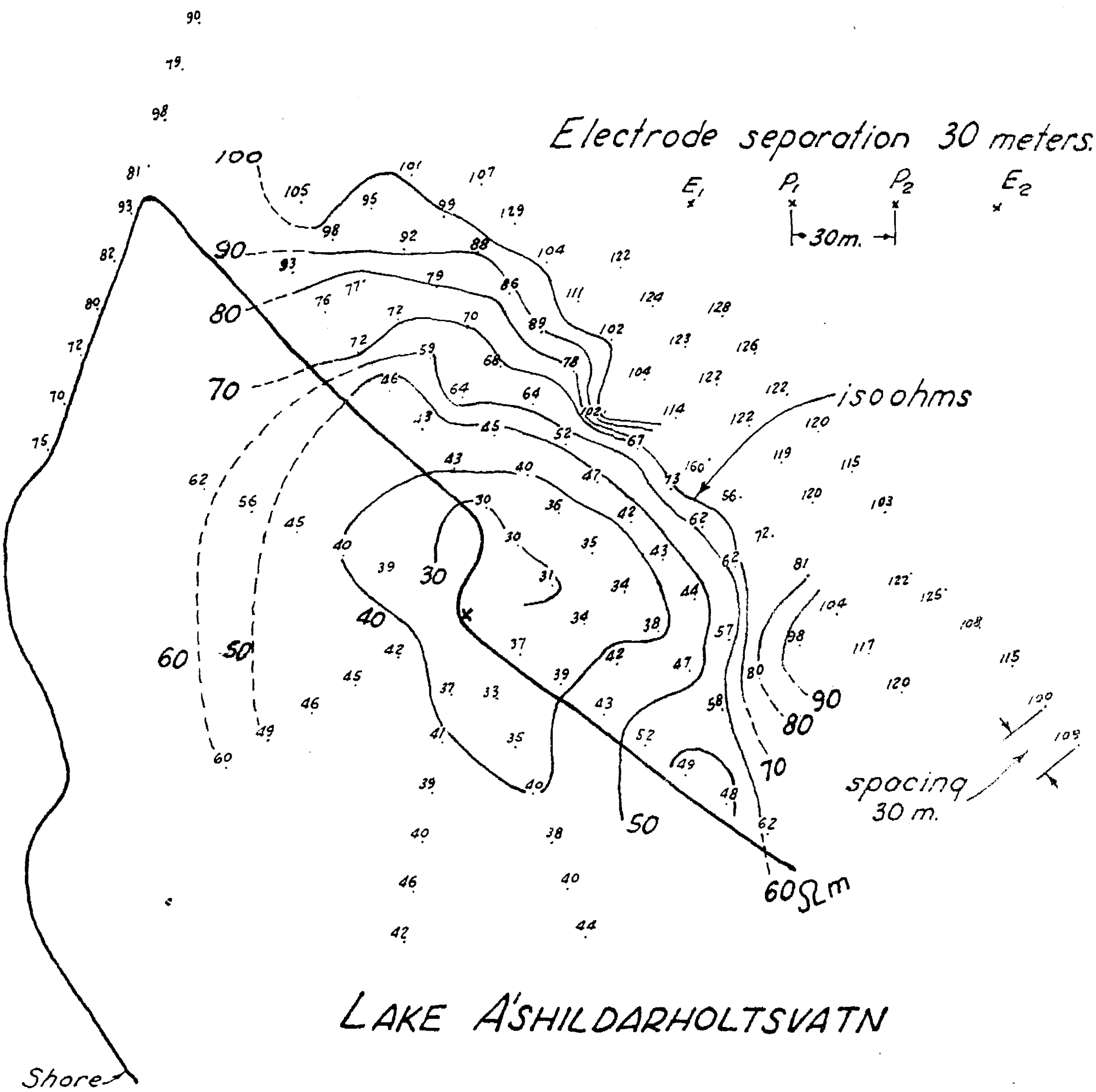
The resistivity survey revealed a clear picture of the situation. Upon detection of the approximate thickness of the sediments, a number of resistivity traverses were made with an electrode separation of 30 meters giving the result shown in Fig. 7.

The iso-ohms in the figure reveal the location of the main fractures in the basement through which the thermal water ascends. This information led to the drilling of a few successful wells, which produce about $20\times10^3 cm^3/sec$ of water at 72^oC from the depth of 130 to 150 meters. The thermal water is now utilized for domestic heating of a nearby community of 1,000 inhabitants.

The conventional resistivity method is less adapted for the study of anomalies below the depth of a few hundred meters mainly because of the greatly increased field work. An induction method of the magneto-telluric type appears to be more practical for this purpose, although the resolving power may be quite low.

The total heat flow transported by the low-temperature areas

The heat flow transported by the integrated natural surface flow of all low-temperature areas has been found to be $1.1\times10^8 cal/sec$. This figure does not include the heat flow that is dissipated by the conduction of heat and by



× Location of successful wells.

Earth resistivity in the Saudarkrokur area in Northern Iceland.

Scale ~ 1:4000

Fig. 7

subsurface drainage. The latter two components have to be estimated before it is possible to estimate the total heat flow transported.

The estimating of the heat flow that is transported in individual areas by conduction and subsurface drainage is possible only on the basis of elaborate data on the subsurface temperature and hydrological conditions. Unfortunately, extensive data of this kind are at hand in a few areas only.

The general indications obtained so far are that the anomalous temperature field in the low-temperature areas is localized around the main channels of upflow. The overall lateral extent of the temperature anomalies in the upper few hundred meters appears to amount to a few square km only even in the largest thermal areas. This is probably the result of the low lateral permeability of the Tertiary basalts and the tendency of the thermal water to deposit hydrothermal minerals in fissures and cracks.

Consequently, the subsurface losses do not appear to be excessive.

More quantitative data are at hand from the thermal areas on H(V) at Reykir (elev. 40 m) (No. (6) and (7) in Table V) in the vicinity of the city of Reykjavik. These areas are separated by approximately 3 km. Drilling was performed in the southern area in 1932 to 1947 and in the northern area in 1947 to 1956. An electric resistivity survey was performed in the northern area in 1947 prior to the drilling.

The natural surface flow in northern and southern areas was 30×10^3 and $100 \times 10^3 \, cm^3/sec$ resp. of water at $80^\circ C$. Combined the areas transport around 10% of the total heat flow transported by the surface flow of the low-temperature areas. They are therefore among the largest thermal areas.

Drilling in the southern area rapidly produced more than $200 \times 10^3 \, cm^3/sec$ of water at $87^\circ C$. In 1947 a total of 60 wells had been drilled in this part, the depth ranging from 130 to 750 meters. The total integrated flow of the wells was $280 \times 10^3 \, cm^3/sec$ but there were obvious indications that the southern part was producing at its ultimate capacity. All surface thermal activity was extinct and no substantial increase of the flow had been obtained for a number of years.

The drilling in the northern area, initiated in 1947, did not result in any great increase of the integrated flow. In 1955, after the completion of around 30 wells, the integrated flow of both areas was around $340 \times 10^3 \, cm^3/sec$. More than 1/2 of the flow is issued by the wells in the northern area, that is, a substantial part of the flow of the southern area had been diverted to the wells in the northern area.

The ultimate capacity of both areas so far as free flow had obviously been recovered. The heat flow recovered is around three times the heat dissipation by the natural springs prior to the drilling.

The obvious interaction between both areas implies that the artesian pressure in the upper part of H(V) at this location had been reduced substantially. As the wells produce at a negligible well heat pressure, it is to be expected that the total free flow of the wells represents approximately the total flow of the combined areas at the present conditions, that is, at the reduced artesian pressure.

This suggests that the total heat flow to the combined areas prior to the drilling was not more than approximately three times the heat flow transported by the surface flow. In fact, the reducing of the overall artesian pressure can have increased the total flow and the above factors may therefore be overestimated.

We may tentatively base our estimate on this result and multiply the total heat flow transported by the surface flow of all low-temperature areas by a factor of two or three in order to get the total heat transported by the group. This leads to a figure of 2 to 3×10^8 cal/sec for the entire group of low-temperature areas.

Summary of the various methods of exploration

Table VII is presented in order to give a brief summary of the various methods of exploration that have been applied by the writer.

TABLE VII

Exploration methods in the low-temperature areas

Exploration method	See page	Structural control	Base temperature and three-dim. temp. field	History of the thermal water: Cooling by conduction	Subsurface losses	Subsurface mixture
Local and general geology		Exploration of local and general structures				
Magnetic methods		Magnetic anomalies of dikes				
Electric resistivity methods	40	Location of major zones of upflow	Data on the subsurface temp. field			
Surface temperature and flow of natural springs	17			Small flow indicative of cooling	High temp. and small flow indicates losses	

TABLE VII (Con't.)

Exploration methods in the low-temperature areas

Exploration method	See page	Structural control	Base temperature and three-dim. temp. field	History of the thermal water: Cooling by conduction	Subsurface losses	Subsurface mixture
Well temperatures			Supplies data on the subsurface temp. field and on the cooling of the ascending water			
SiO_2 (and F) in the thermal water	23		Concentration of SiO_2 (and possibly F) indicative of the base temperature			
Isotope composition of the thermal water	34	Location of recharge area				Supplies data on intermixture with local water
Gases in thermal water	29		CO_2 in the gas indicative of base temperature			

(II.2) The high-temperature areas

The high-temperature activity is concentrated within 12 large thermal areas which are generally termed as the natural steam areas, or abbreviated NS-areas. The locations are shown on Fig. 1. Without exception, all are located in the close vicinity of volcanic centra of post-Glacial or recent activity. NS(I) is probably associated with H(V).

A few relatively smaller thermal areas appear to belong to the high-temperature group, mainly the Great Geyser area, which is the northernmost thermal area on H(III) and possibly also the Reykjahverfi thermal area which is the northernmost area on H(IX).

Some important data on the 12 NS-areas are given in Table VIII to which the following remarks will be made.

The total area of each thermal area represents the area within which there is visible thermal activity.

The analytical data on the composition of the natural steam were collected by the writer and Baldur Lindal, chemist of the State Electricity Authority in Reykjavik. The laboratory analysis of the gas samples were carried out by Mr. Lindal.

Samples were collected from a number of springs in each area and the data given in Table VIII represent results that are typical for each area. Some variations are encountered within the individual areas (see Appendix.)

The data show that the natural steam is characterized by a relatively high content of CO_2, H_2S and H_2 which are possibly of igneous or even partially of magmatic origin.

The figures on the magnitude, that is, the total heat flow dissipated by the individual areas, are estimates given in order to furnish a rough but useable measure of the total heat flow of the areas. The following magnitude scale is applied:

Magnitude	Range of heat dissipated
I	5 - 25×10^6 cal/sec
II	25 - 125
III	125 - 750

These data have been obtained in the following way. Four of the 12 NS-areas, that is (I), (III) and mainly (IV) and (X), which are most accessible, have been studied in greater detail by the means of surface and well data. The specific data obtained in these areas were subsequently applied to the other areas.

Heat is dissipated in each area in four ways, that is, by (1) the steam escaping from the steam holes, (2) the hot ground, (3) an abnormal gradient in a large area around the main outlets and (4) by subsurface draining of thermal water.

TABLE VIII

Natural steam areas in Iceland

Name		Elevation meters	Total area km^2	Magnitude	Analysis of steam: Gases cm^3/gr	CO_2 vol %	H_2S %	H_2 %	CH_4 %	N_2 %
(I)	Reykjanes	15	1	I	3.2	95.o	1.8	1.3	0	1.9
(II)	Trolladyngja	120	0.5	I						
(III)	Krysuvik	150	3	I	5.5	85.o	7.o	6.6	0.2	1.2
(IV)	Hengill	30 to 600	50	II	1.3	83.7	5.3	2.1	0.2	8.7
(V)	Kerlingarfjoll	900	5	II	3.3	69.1	7.4	20.9	0.3	2.3
(VI)	Torfajokull	900	100	III	3.3	88.5	3.5	5.1	0.1	2.8
(VII)	Vonarskard	1.000		I						
(VIII)	Kverkfjoll	1.500	10	II						
(IX)	Askja	1.050	25	I						
(X)	Namafjall	330	2.5	II	13.3	47.7	19.8	30.2	0.4	1.9
(XI)	Krafla	450	0.5	I	6.9	54.3	17.5	25.4	0.3	2.5
(XII)	Theistareykir	330	2.5	I	4.5	68.9	13.9	15.1	0	2.1

TABLE VIII (Con't.)

Geological remarks

(I)	Adjacent to a crater of late post-Glacial basaltic activity. Late post-Glacial opening of controlling fissure. Infiltration of sea water.
(II)	Adjacent to fissures of late post-Glacial basaltic activity.
(III)	Adjacent to fissures of late post-Glacial basaltic activity. One explosion crater within the thermal area.
(IV)	Adjacent to fissures of basaltic activity approx. 1,000 years ago. Only NS-area issuing large quantities of water. Base temp. approx. 230°C.
(V)	Located within large rhyolitic extrusions of late Quaternary or early post-Glacial age.
(VI)	Largest NS-area. Rhyolitic lava flows and explosion craters of post-Glacial age within the area.
(VII)	
(VIII)	Center of historic activity.
(IX)	Large explosion of silicic tefra in 1873
(X)	Adjacent to a fissure of basaltic activity in 1729.
(XI)	Same as (X). Adjacent to a large explosion crater formed in 1729.
(XII)	Adjacent to centra of late post-Glacial basaltic activity.

The flow of all steam wells and a few natural springs in NS (IV) and (X) was measured by Pitot tubes or condensation. The flow of the steam holes in the individual areas was subsequently estimated on the basis of a visual comparison with these known sources. The method is as a matter of course quite inaccurate but gives nevertheless the order of magnitude.

The hot ground which is characterized by an elevated temperature of the surface, is formed in the spots where the upflow of steam per unit area and time is so small that the steam condenses within a short distance from the surface.

The temperature of the surface of the hot ground usually varies from 25°C to 75°C. It is furthermore characterized by small deposits of sulphur 10 to 50 cm below the surface, and a color different from that of the normal ground. It is, therefore, easily distinguished and its area easily measured. These measurements are preferably carried out when there is a light snow-fall as the hot ground melts a thin snow cover.

The measurement of the temperature down to a depth of 25 cm within the hot ground has revealed a temperature gradient in the surface which averages to roughly 5 to 10 °C/cm and which corresponds to a heat flow of roughly 0.o2 to 0.o4 ca./cm^{2}sec. The product of the heat flow and the area gives the total heat dissipated by the hot ground.

The temperature gradient at the surface outside the main outlets varies considerably in the individual areas. Figures of $5x10^{-3}$ to $250x10^{-3}$ $^{o}C/cm$ are found. The latter figure is found in the close vicinity of the hot ground and the main outlets. For the entire area of the individual NS-areas mean figures of $10x10^{-3}$ to $50x10^{-3}$ $^{o}C/cm$ are found. These figures combined with the heat conductivity of the ground and the total area of activity give the total heat flow that is dissipated by conduction to the surface.

The fourth component, the subsurface drainage of thermal water, is again the least accessible factor. However, in the case of NS(X), most of the drainage appears to go into an adjacent lake. Studies of the temperature of the lake at the intake from the area have shown that only a minor part of the heat goes this way.

Furthermore, the drainage area of NS (III) has been investigated by electric resistivity measurements with the result that only minor temperature anomalies were detected indicating that the amount of thermal water drained is only small.

The ratio total steam flow/total flow of a NS-area should be a function of the base temperature. For example, for T_b = 300^{o}C, this ratio should be at most 0.4. In high-temperature areas of relatively low base temperature, the total steam flow should therefore carry only a minor fraction of the heat. This is the case of NS (IV).

The surface outlets of the NS-areas are generally spread over relatively large areas. This is probably due to the tendency of the steam to dissolve and soften the country rock and hence open new outlets.

The fact that two of the greatest NS-areas, that is, NS (V) and NS (VI), furthermore also NS (IX) and possibly NS (X) are associated with silicic intrusives of recent or late Quaternary origin, is remarkable. Silicic volcanics are, indeed, quite rare in Iceland, so this appears to be more than a mere coincidence.

It is in this respect also remarkable that no thermal activity is closely associated with the many large post-Glacial and recent basaltic eruptions. For example, the center which erupted the largest known lava flow, the Thjorsarhraum in southern Iceland which measures about 22 km^3 is not associated with any high-temperature activity. Neither is there activity associated with the Laki fissure which erupted 12 km^3 of basaltic lava in 1783.

The computed total heat flow of all high-temperature areas in Iceland is 0.3 to 1.5×10^9 cal/sec. It is of interest to note that Grange (6) reports a figure of 0.5×10^9 cal/sec for the total activity on the northern islands of New Zealand.

III. THE SOURCE OF ENERGY OF THE THERMAL ACTIVITY

A few of the physical and chemical characteristics of the thermal areas in Iceland were discussed in some detail in the preceding section. The results obtained so far allowed us to carry out an approximate estimate of the total heat flow dissipated by the two groups of thermal areas. Although the discussion has been based on rather definite models, there has been no reference to the possible sources of heat that constitute the ultimate causes of the thermal phenomena.

The question of the heat supply of thermal springs has been discussed by a number of authors. A general discussion with particular reference to the Steamboat Springs, Nevada, is given by White and Brannock (3). Furthermore, the conditions in Northern Iceland have been discussed on a general basis by Einarsson (11).

The general conclusion reached by these and other authors is that the heat supply of thermal springs is either derived from the normal outward flow of terrestrial heat, or from volcanic sources. Exothermal chemical processes, very localized high concentrations of radioactive materials and heat released by mechanical processes,appear inadequate sources. The definitions given on page 5 and 6 were actually based on these general results.

In order to discriminate between a non-volcanic or a volcanic supply of heat, White and Brannock (3) use the evidence furnished by the chemical composition of the thermal water and the gases issued, and furthermore, the temperature gradients in and around the thermal areas. In a later paper, White (32) emphasizes, however, that there is no reliable chemical method of identifying volcanic water or gas.

Einarsson (11) stresses the evidence furnished by the overall heat balance of the individual thermal areas and comes to the conclusion that the low-temperature activity in Northern Iceland is essentially normal, that is, of non-volcanic origin (see p.6). Barth (9), on the other hand, concludes that all thermal activity in Iceland is directly volcanic, that is, the heat is supplied by magmatic water expelled from cooling intrusives.

In the present section, the problem of the heat supply of the thermal activity in Iceland will be discussed with reference to the more recent results obtained by the writer.

Data of general importance for the treatment of this problem are the subsurface temperature-depth relation and the outward flow of terrestrial heat in Iceland. The writer has made an attempt to infer these data from the observations of temperatures in a few shallow wells drilled in areas that appear to be largely unaffected by local thermal anomalies. This will be discussed in the following two sections.

(III.1) Observed well-temperatures in non-thermal areas.

Almost all wells in Iceland that exceed the depth of 50 meters have been drilled in thermal areas for the purpose of the production of hot water or natural steam. There are, in fact, only three apparently non-thermal locations where wells deeper than 90 meters have been drilled and that appear to give some information on the temperature-depth relation and the outward flow of terrestrial heat as unaffected by significant local thermal anomalies or volcanism.

These locations, listed as (I) to (III), are shown in Fig. 8. Only one well has been drilled in (I) and (II) but there are two wells in location (III). The temperature data measured in these wells are given in Fig. 9 to 11 where the computed gradients are also given.

There is furthermore a borderline case, listed as location (IV) in Fig. 8, where one well has been drilled giving the data shown in Fig. 10. This well appears only slightly affected by local upflow of low-temperature water and is therefore worth mentioning.

None of these wells was drilled under the supervision of the writer.

The temperature data measured by the maximum thermometers in locations (I) and (II) should be accurate within $\pm$ 0.1°C. Three thermometers were used at each point. The data obtained by the thermistor equipment in location (III) should be accurate within $\pm$ 0.05 °C. The measurements were all performed a considerable time after the completion of the wells. The exact time in the case of location (I) is unknown to the writer, but more than a year elapsed in the cases of (II) and (III).

On the other hand, the measurements at location (IV) were made by the means of maximum thermometers placed at the well-bottom during periods of shut-down of the drilling equipment. This introduces an uncertainty because of the influence of the drilling fluid which cools the well and leads to a transient temperature effect. In general, however, the measurements were made at times when the drilling fluid had not been circulating for more than 50 hours. Experience shows that the errors in this case will generally not exceed 2°C if one or two meters had been drilled in the hours before the shut-down.

A further disturbance in the case of location (IV) is introduced by the upflow of 300 cm^3/sec of water at 24°C from the depth of 150 meters. This water may be flowing from the same system that is associated with the little spring at location (VIII) in Fig. (5). This local disturbance appears to raise the temperature level in the well by about 6°C and cause further uncertainty as to the reliability of the temperature data.

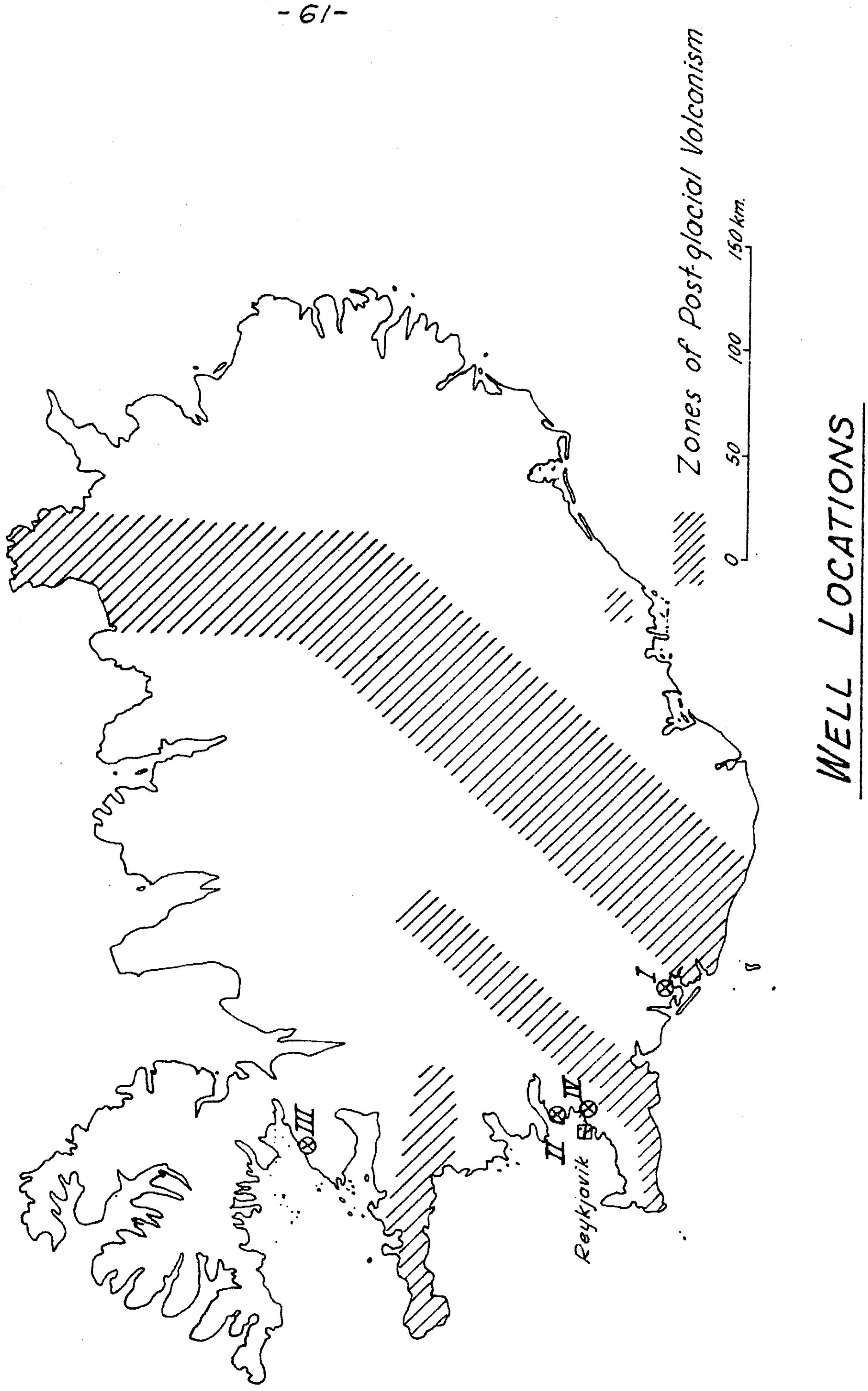
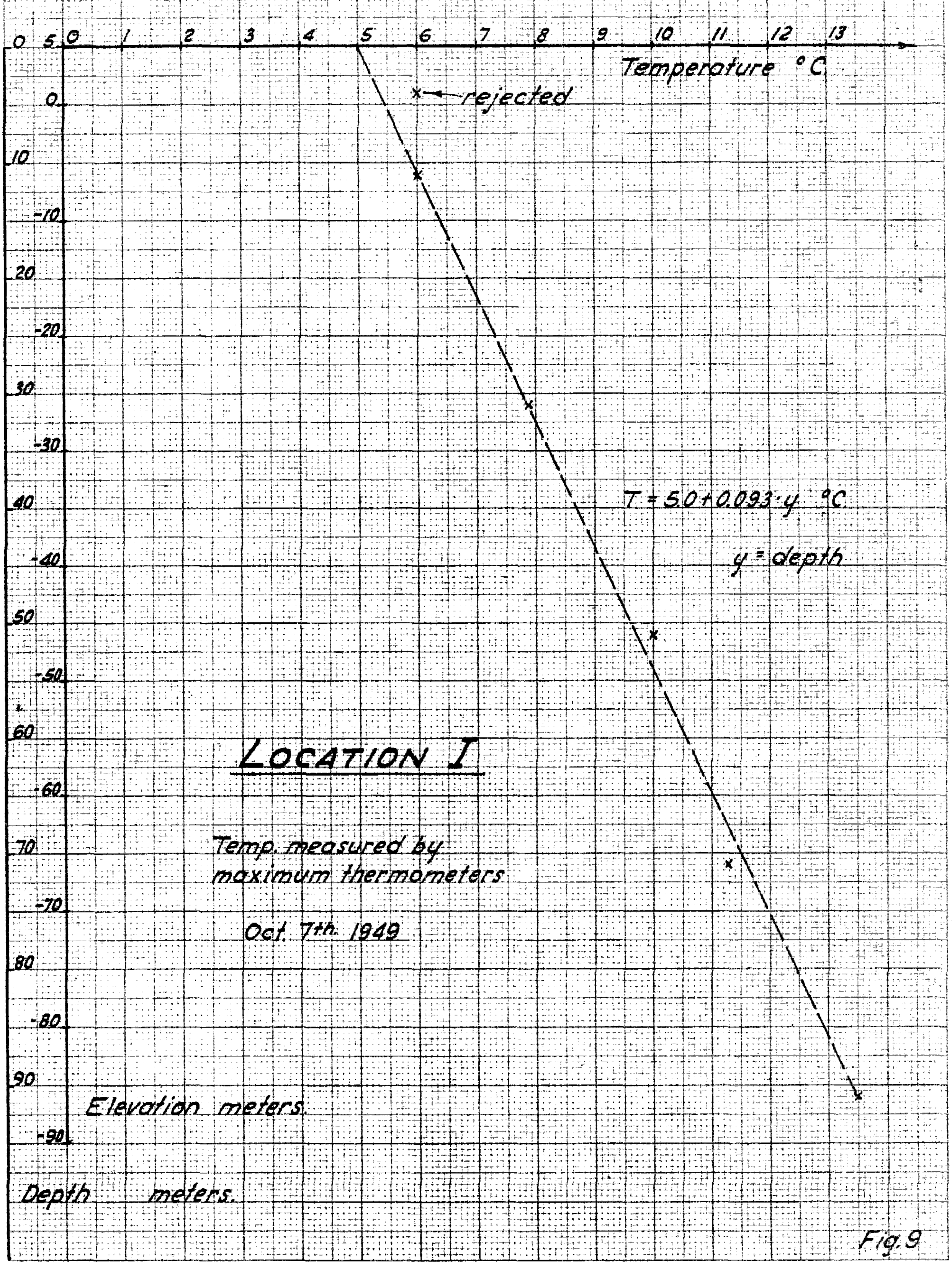

WELL LOCATIONS
Zones of Post-glacial Volcanism
0
50
100
150 km.
Reykjavik
I
II
III
IV

Fig. 8



0
5
0
1
2
3
4
5
6
7
8
9
10
11
12
13
Temperature °C
rejected
0
10
-10
20
-20
30
-30
40
-40
50
-50
60
-60
70
-70
80
-80
90
-90
T = 5.0 + 0.093 · y °C
y = depth
LOCATION I
Temp. measured by maximum thermometers
Oct. 7th. 1949
Elevation meters.
Depth meters.

Fig. 9

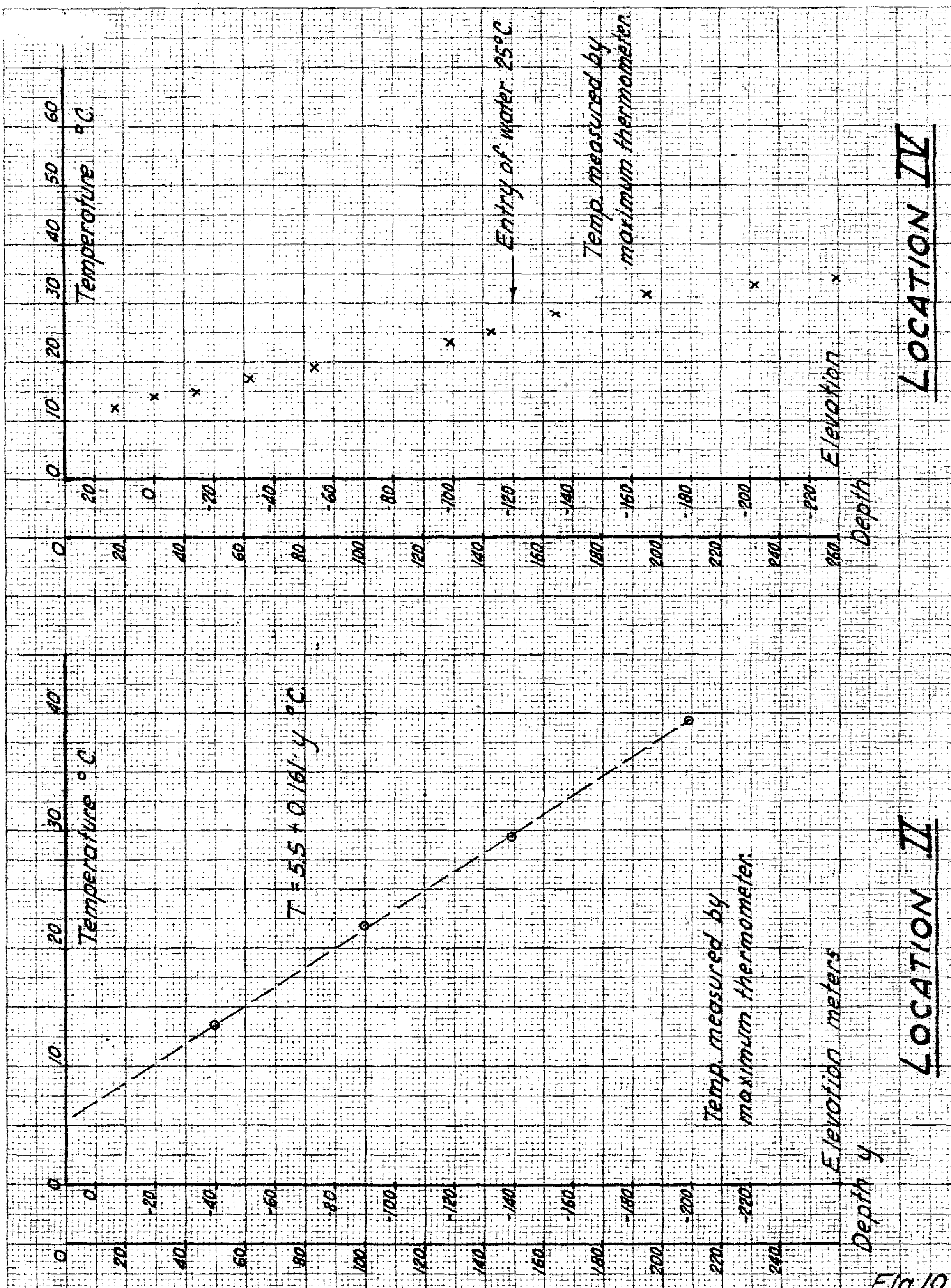

LOCATION IV
Temperature °C.
Entry of water 25°C.
Temp. measured by maximum thermometer.
Elevation
Depth
LOCATION II
Temperature °C.
T = 5.5 + 0.161 · y °C.
Temp. measured by maximum thermometer.
Elevation meters
Depth y

Fig. 10

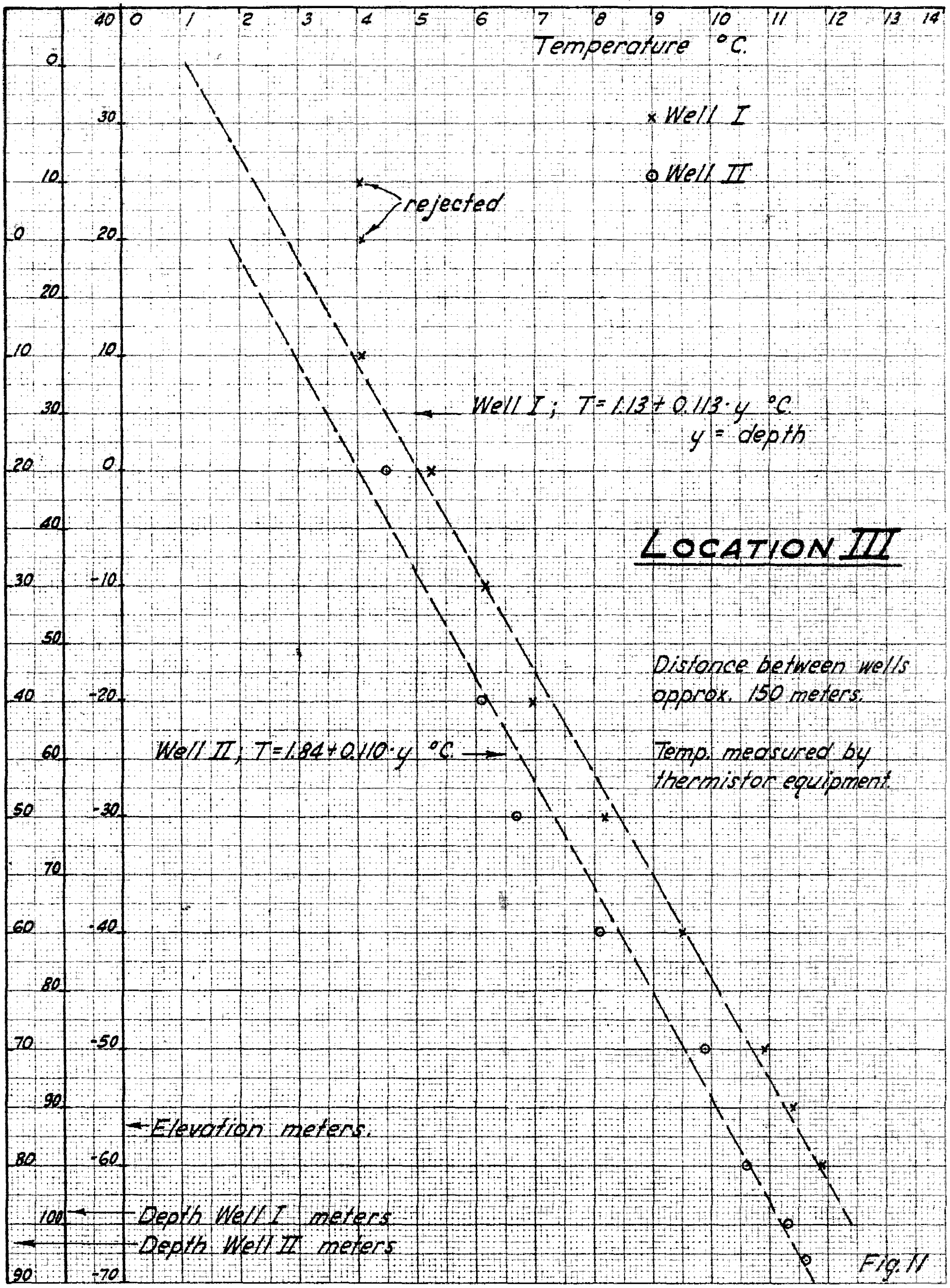

Temperature °C
× Well I
⊙ Well II
rejected
Well I; T = 1.13 + 0.113·y °C
y = depth
LOCATION III
Distance between wells approx. 150 meters.
Temp. measured by thermistor equipment.
Well II; T = 1.84 + 0.110·y °C
Elevation meters.
Depth Well I meters
Depth Well II meters

Fig. 11

(III.2) Interpretation of the well-temperature data

At this juncture, we are mainly interested in approximate data on the terrestrial temperature field above the depth of 5 kilometers and the regional specific outflow of terrestrial heat in Iceland. The present problem consists of inferring these data from the above well-temperatures.

The shallow temperature field is affected by various factors of more or less local nature. In the extrapolation of the above temperature data, a number of corrections will therefore have to be made in order to eliminate the local disturbances. Some of these factors are listed in the following Table (IX) and the reference is given in the cases where special investigations have been made elsewhere.

The factors that appear to the writer to be of importance in the above case of the locations (I) to (IV) are selected in Table (IX).

The Icelandic plateau basalts are considered quite uniform as to the thermal conductivity and item (1) refers only to materials different from the normal plateau basalts.

As shown in Fig. 9 to 11, all well sites are at a low elevation. Locations (I), (II) and (IV) are on a sufficiently large area of flat land that corrections for topography can be neglected.

On the other hand, location (III) is on a slope at the mouth of a small valley of glacial origin and a

TABLE IX

Factors affecting the shallow temperature field

	Factors	Reference	Corrections to be made in locations: (I)	(II)	(III)	(IV)
(1)	Thermal constants of local material	Birch and Clark (33) Birch (34) (35)	x			
(2)	Topography	Birch (34) Jeffreys (36)			x	
(3)	Sedimentation	Jeffreys (37)	x			
(4)	Erosion	Benfield (38)(39)	x	x	x	x
(5)	Annual temperature wave	Ingersoll e.a. (40)	x	x	x	x
(6)	Climatic changes	Birch (41)	x	x	x	
(7)	Movement of fluids and gases	Bodvarsson (21)				x
	Precipitation				?	
	Convection in porous media	Lapwood (42)				
(8)	Intrusion of magma	Bodvarsson (43)	x	x	x	x
(9)	Dissipation of mechanical energy	Bullard (44)				
(10)	Tectonic movements Chemical processes					

correction for topography appears necessary in this case.

Sedimentation is encountered only in location (I) where approximately 50 meters of post-Glacial fine-grained sand has accumulated.

Heavy glacial erosion is a general feature of the geomorphology of Iceland. Corrections for this factor are, therefore, necessary in all four locations.

The annual temperature wave as propagated by the conduction of heat penetrates only the upper 10 meters of the ground. Heavy precipitation can, in cases, have a slight influence on this process. In spite of the very shallow penetration, the annual wave is of importance in the uppermost part of the wells.

The correction for the climatic change at the end of the glaciation is necessary, although of little importance in the present case.

The influence of ground water on the general subsurface temperature field in Iceland is a major problem. That conspicuous amounts of heat are transported by moving water is shown by the widespread thermal activity. However, the thermal water is transported mainly along and up through major fractures as shown by the distribution of the thermal activity.

In general, the vertical permeability of the bulk of the plateau basalts appears negligible, especially that of the earlier parts. This situation is clearly implied by the great number of small lakes located at the edge of higher lying plateaus or in bowls at the tops of single mountains.

It will therefore be assumed that the influence of the ground water on the temperature field is generally confined to the vicinity of the larger water carrying fractures. Corrections are therefore not made in the cases of locations (I) to (III).

There is a possibility that small amounts of rain water may enter at least one of the wells in location (III).

A great number of dikes, sills and other forms of intrusions are encountered in the icelandic plateau basalts. Some of the more recent intrusive bodies could have an influence on the general terrestrial temperature field.

There appear no reasons for expecting that tectonic movements or heat produced by mechanical processes have any influence on the very shallow temperature field. Corrections for these factors therefore appear unnecessary.

The theory of the shallow temperature field

The next step consists of an attempt of a quantitative evaluation of the influence of the various above factors on the shallow temperature field. A general theory will have to take into account the transport of heat by conduction and percolating ground water, transient effects at the surface due to erosion, sedimentation and temperature variations, and finally the heat input by intruded magma and radioactive processes.

A general theory taking into account space-time inhomogenities of these processes would be extremely complex and practically unmanageable. The following treatment will therefore be confined to a much simplified model which, however, appears adequate for the present purpose.

A one-dimensional model will be assumed consisting of a semi-infinite permeable solid with a horizontal boundary. The temperature T' is assumed to be a function of the depth x and the time t only, and the thermal conductivity K, the specific heat c_p and the density ρ are assumed to be constant. The surface boundary is at $x = 0$ at the time $t = 0$.

Water is injected into the solid either at the surface resulting in a downward flow of q units mass per unit time and unit area, or at a great depth x_o, resulting in an upward flow where q has to be counted negative. The temperature of the water injected at the surface is assumed the same as that of the surface. The flow q is assumed constant in space and time.

Furthermore, the flow q is assumed relatively small and the distributions of the channels so dense that a complete temperature equilibrium is reached between the water and the rock, that is, both assume the same temperature. The specific heat of the water s_p is assumed to be constant. The vertical mass transport of heat is therefore $s_p qT'$ units heat per unit area and unit time.

Erosion of the surface or sedimentation on the surface are assumed to proceed at a constant rate v. Due to the erosion, the surface boundary moves into the solid at the velocity v. In the case of sedimentation, the figure v is negative.

Finally, it is assumed that magma is being intruded into the solid at various times and depths. The size of each intrusive is assumed relatively small and their distribution relatively dense so that it is in the first approximation possible to work with a mean rate of intrusion measured in M' mass units of magma per unit time and unit volume country rock. This figure is assumed constant in time and a function of the depth only. The figure M' is also assumed so small that the volume of the country rock is not greatly affected.

Let L be the latent heat of melting of the magma, T_m its temperature at the time of intrusion and $\bar{c}_{pm}$ its mean specific heat in the interval T' to T_m where T' is the temperature of the country rock at the location of the intrusion. The heat input per unit volume country rock and unit time will be

$$M'(L + \bar{c}_{pm}(T_m - T')). \tag{7}$$

Adding to (7) the heat produced by radioactive processes measured in R' units heat per unit volume and unit time the total heat input becomes

$$H = M'(L + \bar{c}_{pm}(T_m - T')) + R'. \tag{8}$$

The figure R' is assumed constant in time.

Upon these assumptions, the heat transport equation can be written

$$-\frac{\partial}{\partial x}\left(-K\frac{\partial T'}{\partial x} + s_p q T'\right) + H = c_p\rho\frac{\partial T'}{\partial t} \qquad (9)$$

or rewritten

$$KT'_{xx} - s_p q T'_x + H = c_p \rho T'_t \qquad (10)$$

where T'_x and T'_t represent partial derivatives with respect to x and t respectively.

This equation has to be solved with the boundary and initial conditions

$$\begin{aligned} x' - vt &= 0 \qquad & T' &= h(t) \\ t &= 0 \qquad & T' &= f(x) \end{aligned} \qquad (11)$$

where x' is the position of the top surface.

The first step toward the solution of (10) consists of the transformation

$$y = x - vt' \qquad t = t' \qquad (12)$$

where t' has been substituted for t in equation (10). This transforms the derivatives as follows:

$$T'_x = T_y \qquad T'_{t'} = -vT_y + T_t \qquad (13)$$

$$T'_{xx} = T_{yy}$$

Equation (10) is transformed into

$$aT_{yy} + wT_y - bT + A = T_t \tag{14}$$

where the following abbreviations have been made

$$a = K/c_p\rho \qquad w = v - s_\rho q/c_\rho\rho$$
$$b = M\bar{c}_{pm}/c_p\rho \qquad A = (ML + M\bar{c}_{pm}T_m + R)/c_p\rho \tag{15}$$

and M and R are the transforms of M' and R' respectively.

The boundary and initial conditions (11) are in the moving system (y,t)

$$y = 0 \qquad T = h(t)$$
$$t = 0 \qquad T = f(y) \tag{16}$$

The derivation of equation (14) is based on the assumption that the figures a and w were constant in space and time and the figures M and R constant in time. This is still rather general for the present purpose.

In the following, we will mainly be concerned with the computation of short time temperature transients in near surface layers, that is, in the uppermost few km. The heat released in this limited region by the radioactive processes is generally very small compared to the total outward flow of heat. Furthermore, the influence of magmatic intrusives in this region is in many cases small or negligible. It is, therefore, in many cases possible to simplify equation (14) further, either by disregarding the latter two terms in the l.h.s. or carry out the computation on the basis of a semi-infinite solid with constant b and A. A general solution of (14) with the conditions (16) can be given for the latter case.

The solution for this case is obtained by the means of the transformation (see Frank-Mises (45) page 605, Carslaw (46) page 111)

$$T = A/b + u(y,t)\exp(-wy/2a - w^2t/4a - bt) \tag{17}$$

which transforms (14) into the simple heat conduction equation

$$au_{yy} = u_t. \tag{18}$$

The boundary and initial conditions are transformed into

$$\begin{aligned} y &= 0 \qquad u = (h(t) - A/b)\exp(w^2t/4a + bt) \\ t &= 0 \qquad u = (f(y) - A/b)\exp(wy/2a). \end{aligned} \tag{19}$$

The solution of (18) at the conditions (19) is given by Carslaw (46) page 46).

$$\begin{aligned} u &= \int_0^\infty \left(f(z) - \frac{A}{b}\right) e^{\frac{wz}{2a}} G_1(t,y;z)\,dz \\ &+ y\int_0^t \left(h(z) - \frac{A}{b}\right) e^{\left(\frac{w^2z}{4a} + bz\right)} G_2(t,y;z)\,dz \end{aligned} \tag{20}$$

where $G_1(t,y;z)$ and $G_2(t,y;z)$ are defined

$$G_1(t,y;z) = \frac{1}{2\sqrt{\pi a t}}\left(e^{-\frac{(z-y)^2}{4at}} - e^{-\frac{(z+y)^2}{4at}}\right) \tag{21}$$

$$G_2(t,y;z) = \frac{1}{2\sqrt{\pi a}\,(t-z)^{3/2}}\, e^{-y^2/4a(t-z)}$$

The term A/b is to be dropped from equations (17) and (19) in the case where A = b = 0.

<u>Special case: Short period of erosion with the initial condition f(y) = gy where g is constant</u>. The temperature field in near surface layers at the end of a relatively short period of erosion can be computed on the basis of the assumption of b = A = 0. The case of a constant initial temperature gradient is of main interest, that is, f(y) = gy, where g is the constant gradient.

Assuming for further simplification that the surface is held at zero temperature and that no upflow or downflow of water occurs, that is h(t) = 0 and q = 0 or w = v, we find by an evaluation of the first integral in equation (20) and a transformation back to T by equation (17)

$$T = \frac{g}{2}(y+vt)\left(1+\operatorname{erf}\left(\frac{y+vt}{2\sqrt{at}}\right)\right) + \frac{g}{2}(y-vt)\left(1-\operatorname{erf}\left(\frac{y-vt}{2\sqrt{at}}\right)\right)e^{-vy/a} \tag{22}$$

The derivative of (22) gives the temperature gradient at the surface

$$(T_y)_{y=0} = g\left((1+v^2t/2a)\left(1+\operatorname{erf}\left(\frac{v}{2}\sqrt{\frac{t}{a}}\right)\right) + v\sqrt{\frac{t}{\pi a}}\,e^{-v^2t/4a}\right) \tag{23}$$

The temperature gradient at the surface depends therefore on $R = r/2\sqrt{at}$, where r is the portion eroded in the time t, that is, r = vt.

<u>Approximation methods in the case of two or more short periods of erosion</u>. The problem of estimating the effect of more than one period of erosion will be encountered in the following. In principle, this problem is solved by

by the application of the general solution (20) to each period. The temperature distribution found at the end of a period serves as the initial distribution for the subsequent period. The parameter v is zero for periods of no erosion.

The computations involved in this procedure become quite lengthy when several periods with different external parameters are encountered.

In the present case, we are only interested in the erosion during the Pleistocene glacial stages. As will be mentioned below the indications are that each of the 4 major glacial stages were relatively short compared to the total length of the Pleistocene. Each stage may have covered less than 10% of the total length of the Pleistocene.

The dating of the various periods of the Pleistocene is still very inaccurate and this will introduce a considerable uncertainty. Estimates of the glacial erosion are furthermore quite inaccurate. Accurate theoretical evaluations of the expressions involved are therefore not warranted. The following approximations may consequently be introduced.

Let Δt be the length of a period of erosion and t the time that has passed since the middle of the period. In the case where $t >> \Delta t$ the effects of the erosion at the time t may be approximated on the basis that the total erosion occurred instantaneously at the middle of the period.

Let the initial conditions before the erosion be $T = gy$, where g is a constant gradient, and furthermore $q = b = A = h(t) = 0$. If r is the total thickness of the layer eroded, the approximation consists of solving the equation

$$aT_{yy} = T_t \tag{24}$$

with the initial and boundary conditions

$$\begin{array}{ll} t = 0 & T = g(y + r) \\ y = 0 & T = 0 \end{array} \tag{25}$$

The solution of this simple problem is (Carslaw (46) page 41).

$$T = g(y + r \operatorname{erf} \frac{y}{2\sqrt{at}}) \tag{26}$$

and the temperature gradient is

$$T_y = g(1 + \frac{r}{\sqrt{\pi at}} \exp(-y^2/4at)). \tag{27}$$

The surface gradient may be written

$$(T_y)_{y=0} = g(1 + C_e(t)) \quad , \quad C_e(t) = r/\sqrt{\pi at} \tag{28}$$

where $C_e(t)$ is the correction factor at the time t. This holds for $t >> r/v$ only.

Let a new short period of the length Δt and a total erosion r start at the time $t = t_1$ after the middle of the first period. If $t_1 >> \Delta t$ the initial temperature distribution for the second period is given approximately by equation (26). The temperature distribution at the end of the second period can therefore be found approximately by inserting the expression (26) into the general solution (20).

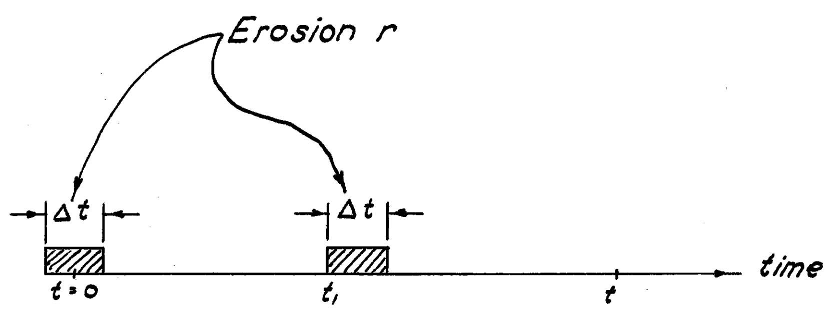


Fig. 12

The computation of the integral (20) can often be avoided by the fact that it is easy to furnish adequately accurate estimates of the upper and the lower limits to the cumulative effects of the erosional periods.

The upper limit is easily derived from the fact that the gradient T_y in equation (27) has its maximum at the surface. The effect of the first period of erosion is therefore overestimated if the expression (28)

$$T = gy(1 + C_{e,1}(t_1)) \tag{29}$$

is applied as the initial temperature distribution to the second period of erosion. The subscript e,1 applies to the first period.

As the initial temperature gradient is constant, the effect of the second period of erosion can also be expressed by a correction factor $C_{e,2}(t - t_1)$ where t is a time after the beginning of the second period (see Fig.12). If $t - t_1 < \Delta t$ the correct expression for this factor is given in the r.h.s. of equation (23). On the other hand, if

$$t - t_1 >> \Delta t$$

the factor can be expressed in good approximation according to equation (28)

$$C_{e,2}(t - t_1) = r/\sqrt{\pi a(t - t_1)} \tag{30}$$

The surface gradient at the time t, where t is a time at or after the end of the second period, is therefore subject to the inequality

$$(T_y)_{y=0} < g(1 + C_{e,1}(t_1))(1 + C_{e,2}(t - t_1)) \tag{31}$$

The lower limit to the cumulative effect is obtained in the following way. According to the general solution (20), the solution at the end of the second period of erosion , that is, for $t = t_1 + \Delta t$, may be written

$$T = \int_0^\infty \left(gZ + F(Z,t_1)\right) e^{\frac{wZ}{2a}} G_1(\Delta t, y; Z)\, dZ \tag{32}$$

where $F(z,t_1)$ represents the effect of the first period of erosion at the beginning of the second period.

The integral (32) consists of two parts. The first part is due to the first term in the parenthesis and leads to a solution of the form (22) and its contribution to the surface gradient can therefore be written

$$g(1 + C_{e,2}(\Delta t)). \tag{33}$$

In computing the second part due to the second term in the parenthesis, it is to be observed that the factor exp(wz/2a) is always greater than or equal to unity and this part of the integral is therefore underestimated if this factor is replaced by the factor one.

The function F(y,t) is a solution of the heat conduction equation (24) and hence

$$F(y,t_1+\Delta t) = \int_0^\infty F(z,t_1)\, G_1(\Delta t, y; z)\, dz \tag{34}$$

The surface gradient due to this part is therefore

$$F_y(0,(t_1 + \Delta t)) = gC_{e,1}(t_1 + \Delta t) \tag{35}$$

The surface gradient at the end of the second period is therefore underestimated if the following expression is applied

$$g(1 + C_{e,1}(t_1 + \Delta t) + C_{e,2}(\Delta t)) \tag{36}$$

Accordingly, by the proper choice of the second correction factor the following inequality holds for any time t after the second period

$$(T_y)_{y=0} > g(1 + C_{e,1}(t) + C_{e,2}(t - t_1)) \qquad (37)$$

which furnishes the lower limit to the surface gradient.

The expressions (31) and (37) represent the results of the present estimations. Both are easily extended to the case of more than two periods of erosion. The correction factors are, in most practical cases, small compared to unity and both expressions therefore give similar values.

Approximations in the case of a short period of no erosion following a period of erosion. In the case where a period of erosion is followed by a much shorter period of no erosion, the changes of the surface gradient in the second period will be relatively small and can be regarded as a small correction to the surface gradient at the end of the first period. A useable approximation to this correction can be derived in the following way.

It will be assumed again that $q = b = A = h(t) = 0$ and that the initial temperature distribution at the beginning of the erosional period is $T = gy$ where g is a constant gradient. The temperature distribution at the end of this period is therefore given by equation (22). In order to obtain the temperature distribution at the end of the second period, this expression has to be used as the initial temperature distribution for the second period.

An approximation in the present case is found by assuming that the total erosion during the first period occurred instantaneously at the time t_o before the end of this period. The time t_o can be found by the condition that the approximation should give the exact surface gradient at the end of the period.

Expression (23) can in most practical cases be written

$$(T_y)_{y=0} = g(1 + 1.1r/\sqrt{at_o}) \tag{38}$$

where t is the time since the beginning of the erosion. On the other hand, the surface gradient at the time t in the case of an instantaneous erosion of the same amount at the time t - t_o is given by expression (27)

$$(T_y)_{y=0} = g(1 + r/\sqrt{\pi a t_o}) \tag{39}$$

By equating (38) and (39) we find

$$t_o = t/1.21\pi = 0.26t \tag{40}$$

The surface gradient at the end of the second period of the length Δt where $\Delta t << t$ can therefore be approximated by

$$(T_y)_{y=0} = g\left(1 + \frac{r}{\sqrt{\pi a(0.26t + \Delta t)}}\right) \tag{41}$$

The surface gradient at the end of the second period is therefore obtained by the multiplication of the surface gradient at the end of the first period by the factor

$$\sqrt{\frac{0.26t}{0.26t+\Delta t}} \qquad (42)$$

<u>Uniform intrusion of magma at constant rate into the upper parts of the crust.</u> The next case of particular interest consists of that where magma is being uniformly intruded at a constant rate into the upper parts of the crust, that is, into the upper 10 to 15 km. This case will be treated in the following with the further simplification that no erosion, sedimentation or flow of ground water occurs.

Due to the linear character of the heat conduction equation, the temperature field in this case can be written T + u where T is the unperturbed terrestrial temperature field and u is the perturbation due to the magmatic processes. This may be inserted in equation (14) and we get for u

$$au_{yy} - bu - bT + A' = u_t \qquad (43)$$

where

$$A' = (ML + M\bar{c}_{pm}T_m)/c_p\rho$$

and

$$b = M\bar{c}_{pm}/c_p\rho$$

The rate of intrusion M will, as before, be assumed constant in space and time and relatively small. Equation (43) can then be simplified further in the present case of relatively shallow intrusions.

The heat content of magma, reckoned from $0^{\circ}C$ will, in general, be 400 to 500 cal/gm (Verhoogen (47), Birch (48)). On the other hand, the temperature of the country rock at the depth of 15 km will, in most cases, not exceed $500^{\circ}C$. The average heat content of the country rock above the depth of 15 km is therefore not in excess of 60 cal/gm. This implies that the term A' will be substantially greater than the term b(T + u). The latter can therefore be disregarded in the first approximation mainly if the heat content of the magma is replaced by the average sensible heat content, that is, the heat content in excess of that of the country rock.

Equation (43) can therefore, in the present case, be replaced by

$$au_{yy} + B = u_t \tag{44}$$

where B represents the term A' as corrected for the average sensible heat content of the country rock. The term B is

$$B = ME_o/c_p\rho \tag{45}$$

where E_o is the average sensible heat content per unit mass of the magma. It will now be assumed that B is constant down to a certain depth d and zero below this depth. The present problem can then be treated as the case of equation (44) for a semi-infinite solid with the boundary and initial conditions

$$y = 0 \qquad u = 0$$
$$t = 0 \qquad u = 0 \tag{46}$$

The solution of this simple problem is given by Carslaw (46) page 61)

$$0 < y < d$$

$$u = \frac{aME_o t}{K}\left(1 - 4i^2\text{erfc}\,\frac{y}{2\sqrt{at}} + 2i^2\text{erfc}\,\frac{d+y}{2\sqrt{at}} - 2i^2\text{erfc}\,\frac{d-y}{2\sqrt{at}}\right) \tag{47}$$

$$d < y$$

$$u = \frac{aME_o t}{K}\left(2i^2\text{erfc}\,\frac{y-d}{2\sqrt{at}} + 2i^2\text{erfc}\,\frac{y+d}{2\sqrt{at}} - 4i^2\text{erfc}\,\frac{y}{2\sqrt{at}}\right)$$

where $i^2\,\text{erfc}(x)$ represent the second integral of the erfc(x) and is

$$i^2\text{erfc}\,x = \tfrac{1}{4}\left((1 + 2x^2)\,\text{erfc}\,x - \frac{2}{\sqrt{\pi}}\,x\,e^{-x^2}\right).$$

The temperature gradient at the surface is

$$(u_y)_{y=0} = \frac{2ME_o\sqrt{at}}{K}\left(\frac{1}{\sqrt{\pi}} - i\text{erfc}\,\frac{d}{2\sqrt{at}}\right) \tag{48}$$

where

$$i\text{erfc}\,x = \frac{1}{\sqrt{\pi}}\,e^{-x^2} - x\,\text{erfc}\,x$$

The after-effects of a period of magmatic activity.

The approximate computation of the thermal effects of a period of magmatic activity at a time after its end which is long compared to the duration of the period is of certain interest. This case may be treated as an instantaneous input of heat at zero time.

It will be assumed that the intrusion occurs between the depths y_1 and y_2 and that it is uniform in space as in the above case. If a total of M_o mass units of magma with the sensible heat content E_o are intruded per volume country rock at time zero, the instantaneous temperature increase between the depths y_1 and y_2 will approximately be $M_oE_o/c_p\rho$. It is furthermore assumed that this figure is approximately constant.

Following the same lines as in the preceding paragraph, the computation of the temperature field due to the magmatic activity consists of that of solving equation (44), for the semi-infinite solid in the case of B = 0 and the boundary and initial conditions

$$y = 0 \qquad u = 0$$

$$t = 0 \qquad u = \begin{cases} M_oE_o/c_p\rho \text{ for } y_1 < y < y_2 \\ 0 \text{ elsewhere} \end{cases} \tag{49}$$

The solution of this simple problem is given by Carslaw ((46) page 43)

$$u = \frac{M_o E_o}{2 c_p \rho}\left(\operatorname{erf}\frac{y_2 - y}{2\sqrt{at}} - \operatorname{erf}\frac{y_2 + y}{2\sqrt{at}} - \operatorname{erf}\frac{y_1 - y}{2\sqrt{at}} + \operatorname{erf}\frac{y_1 + y}{2\sqrt{at}}\right) \tag{50}$$

the surface gradient is

$$(u_y)_{y=0} = \frac{M_0 E_0}{c_p \rho \sqrt{\pi a t}} \left(\exp(-y_1^2/4at) - \exp(-y_2^2/4at) \right) \qquad (51)$$

The terrain corrections

The temperature field in the near surface layers is affected by the forms of the earth's surface. The temperature data obtained in shallow wells have, therefore, to be corrected for nearby departures of the surface from the horizontal. These corrections will be made along the lines given by Jeffreys (36).

The surface forms are generally quite irregular and can not, in most cases, be described by simple mathematical expressions. In order to arrive at useable results without excessive numerical computations, the surface forms will have to be approximated by simple geometrical forms.

The case that will be treated here is the case where the horizontal dimensions of the surface irregularities are much larger than the vertical dimensions. The effects of these irregularities on the temperature field are, in general, small with the exception of the effects in the close neighborhood of discontinuities of the surface slope. These exceptions are, however, unimportant.

Only the two-dimensional steady state case will be treated in the following, that is, the case of long mountain ridges of a steady form.

Suppose a semi-infinite solid with surface irregularities as described above. Let the x-axis be horizontal and perpendicular to the strike of the surface forms. Furthermore, the y-axis be vertical and the plane $y = 0$ at sea level. Let $h(x)$ be the height of the surface forms which is a function of x only.

The temperature field will now be expressed by $T + u$ where T is the field in the case $h(x) = 0$ everywhere, and u is the perturbation due to the surface irregularities.

Let g be the temperature gradient at $y = 0$, K be the thermal conductivity and $Q = gK$ be the outflow or heat per unit area and unit time in the case when $h(x) = 0$ everywhere. Then in the first approximation, the surface irregularities lead to a temperature of $gh(x)$ in the plane $y = 0$.

The first approximation of the perturbation temperature u is therefore obtained by solving the steady state equation

$$u_{xx} + u_{yy} = 0 \tag{52}$$

in the case of a semi-infinite solid with horizontal surface and the boundary conditions $u = gh(x)$ for $y = 0$.

The solution of equation (52) is given by the Poisson integral (see e.g. Courant-Hilbert, (49) page 245)

$$u = \frac{y}{\pi}\int_{-\infty}^{\infty} \frac{gh(z)dz}{(x-z)^2 + y^2} \tag{53}$$

The solution (53) will now be evaluated in a special case which consists of a single surface ridge of the form shown in Fig. (13).

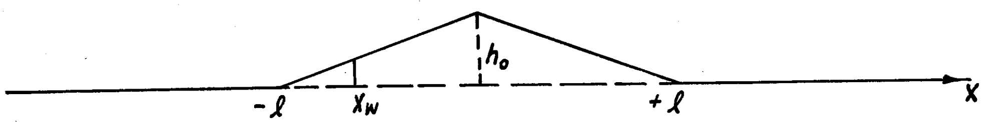


Fig. 13

In this case, the function h(x) is

$$h(x) = \begin{cases} 0 & \text{for } |x| > \ell \\ h_o(\ell + x)/\ell & -\ell < x < 0 \\ h_o(\ell - x)/\ell & \ell > x > 0 \end{cases}$$

Assuming that g is constant, the integral (53) is found to be

$$u = \frac{gh_o}{\pi\ell}\left((\ell+x)\left(\tan^{-1}\frac{x+\ell}{y} - \tan^{-1}\frac{x}{y}\right) + (\ell-x)\left(\tan^{-1}\frac{x}{y} - \tan^{-1}\frac{x-\ell}{y}\right)\right.$$
$$\left. + \frac{y}{2}\log\frac{(x^2+y^2)^2}{((x-\ell)^2+y^2)((x+\ell)^2+y^2)}\right) \tag{54}$$

The gradient is

$$u_y = \frac{gh_o}{\pi\ell}\left[(\ell+x)\left(\frac{x}{x^2+y^2} - \frac{x+\ell}{(x+\ell)^2+y^2}\right)\right.$$
$$+ (\ell-x)\left(\frac{x-\ell}{(x-\ell)^2+y^2} - \frac{x}{x^2+y^2}\right)$$
$$+ y\left(\frac{2}{x^2+y^2} - \frac{1}{(x-\ell)^2+y^2} - \frac{1}{(x+\ell)^2+y^2}\right)$$
$$\left. + \frac{1}{2}\log\frac{(x^2+y^2)^2}{((x-\ell)^2+y^2)((x+\ell)^2+y^2)}\right] \tag{55}$$

Putting y = 0 we find the surface gradient at the location x

$$(u_y)_{y=0} = \frac{g h_0}{\pi \ell} \log \frac{x^2}{|x^2 - \ell^2|} \qquad (56)$$

Expression (56) is not applicable in the close neighborhood of the logarithmic singularities at x = - , 0 and . This difficulty can be circumvented by giving y a small positive value at these points.

Equation (56) gives the first approximation to the terrain correction of the surface temperature gradient.

The thermal constants of the Icelandic basalts

Data on the thermal constants of the icelandic basalts are necessary in order to evaluate the various equations given above. Unfortunately, no drill cores from the above wells have been available to the writer and a direct study of the conductivity has therefore not been possible.

This difficulty appears, however, not too serious. Basaltic effusives are rather uniform in character, and a great scatter of the conductivity values within this class of rock is not to be expected. For the present purpose, it should therefore be possible to rely on data obtained on basaltic effusives elsewhere.

The Handbook of Physical Constants (50) lists the thermal conductivity of basalts 4.3 to $5.3\text{x}10^{-3}$ cal/sec ^{o}C cm. The recent studies of Mossop and Gaffner (51) on dolerites in the Orange Free State, Union of South Africa, give values of 4.1 to $5.0\text{x}10^{-3}$ and Bullard and Niblet (52) give the value of $4.5\text{x}10^{-3}$ for dolerite in Kelham Hills, Nottinghampshire, England.

In view of these data, the thermal conductivity of the icelandic plateau basalts will be estimated at $4.5\text{x}10^{-3}$ cal/sec ^{o}C cm. The specific heat can be estimated at 0.21 cal/^{o}C gm (see Mossop and Gaffner (52)) and the average density at 2.7 gm/cm^{3} (see Einarsson (53)). This gives a thermal diffusivity of $8\text{x}10^{-3}$cm^{2}/sec, or in a more convenient unit 25 km^{2}/10^{6} years.

Corrections of the well-temperature data

In order to arrive at the necessary corrections of the observed well-temperatures, the main lines of the late Tertiary and Quaternary geological history of the well locations have to be known.

As to the geological history, the point of main interest is the age of the uppermost section of the basalt plateau, that is, the age of the grey section. It has already been stated that recent geological evidence favors a Tertiary, probably late Tertiary age. This conclusion would also have to apply to the age of the peneplane at the top of the grey section. Erosion by Pleistocene glaciers could in this case only account for a part of the total erosion and dissection of the plateau.

On the other hand, however, the earlier theories of Peturs (10) and his followers, still find support by some geologists. According to this view, the grey section should be of an early Pleistocene age, and erosion by middle and late Pleistocene glaciers would therefore have to account for the whole of the impressive erosion and dissection.

An attempt to solve this controversy will not be made. The latter possibility can not be excluded and the present corrections will therefore be made along both lines separately. There is a possibility that the thermal data may lend some support to the one or the other hypothesis.

Accordingly, the theory of a Tertiary age of the grey section will be referred to as Hypothesis I, and that of an early Pleistocene age as Hypothesis II. The data for the corrections will be selected such as to represent somewhat extreme conditions in each direction.

The data selected for the corrections of the well-temperatures, are given in Table (X). The data on the Pleistocene chronology have been selected with regards to the works of Emiliani (54) and Antevs (55). These data furnish the basis for the estimates of the age of the grey section.

In Hypothesis I, the mean age of the grey section is estimated at $5x10^6$ years which places it back into the mid-Pliocene. This is several times less than the current estimates of the age of the lower sections of the basalt plateau. The difference in age is suggested by the existence of considerable sedimentary horizons underlying the grey section (Einarsson (11)).

On the other hand, the age of $5x10^6$ years is roughly 10 times the age that is arrived at on the basis of the theories of Peturs (10) who interpreted the underlying sediments as glacial deposits.

The data on the erosion are estimated with regard to the local topography and the relation of the well-locations to the peneplane at the top of the grey section.

In Hypothesis I, it is assumed that the peneplane is pre-Glacial and represents an average pre-Glacial erosion of 200 to 300 meters. The glacial valleys and fiords would then be the work of the whole Pleistocene period. It is estimated that each stage accounts for roughly 1/4 of the total dissection of the valleys and fiords.

In Hypothesis II, it is assumed that the peneplane and the valleys as well, as the fiords, are the result of the latter half of the Pleistocene. The third and the fourth glacial stages are each accounted for 1/2 of the total peneplanation and dissection.

As to Item (14), it is assumed that the building up of the grey section was accompanied by the formation of

TABLE X

Estimated data for the corrections of the well-temperatures

Item		
(1)	End of the glaciation of the lowlands of Iceland	8,000 y.b.p.
(2)	Total length of fourth glacial stage	50,000 y.
(3)	Total length of the last interglacial stage	100,000 y.
(4)	Total length of third glacial stage	50,000 y.
(5)	Total length of first and second glacial stages	100,000 y.
(6)	Total length of the Pleistocene	600,000 y.
(7)	Climatic changes	Temperature rise by 5°C about 8,000 y.b.p.
(8)	Sedimentation and conductivity of the sediments Location I	50 meters in 8,000 y. conduct. lower than for basalts
(9)	Annual temperature wave	Rejection of the data in the upper 10 to 20 meters of the wells
(10)	Terrain correction Location III	Situated on a slope 1:10

TABLE X (Con't)

Estimated data for the corrections of the well-temperatures

Item	Hypothesis I	Hypothesis II
(11) Estimated mean age of the grey section	5×10^6 y.	500,000 y.
(12) History of the glacial erosion	Pre-Glacial erosion of the plateau. Glacial valleys formed in the entire Pleistocene.	Total erosion of the plateau occurred during the third and the fourth glacial stage.
(13) Estimated erosion during each of the glacial stages		
Location I	75 meters	250 meters
Location II	225 meters	600 meters
Location III	150 meters	400 meters
(14) Intrusion of magma in the upper 5 km		Early Pleistocene Intrusives not in excess of 5% of the volume of the country rock above the depth of 5 km.

a number of small intrusives as dikes, etc. In order to indicate a possible order of magnitude of this term, it will in the first instance be assumed that the intrusives amount to an average of 5% by volume of the country rock above the depth of 5 km. This figure is, as a matter of course, very uncertain.

On the other hand, intrusives of this kind formed in the upper 5 km several millions of years ago would be without any influence on the present temperature field.

The data obtained in well-location (IV) will be regarded as uncertain and will not be included in the following discussion.

The correction for sedimentation in Location (I) is necessary, as the well penetrated 50 meters of dense sand. It is furthermore to be taken into account that the sand will have a slightly lower thermal conductivity than the basalts.

Well-location II is located on a slope averaging 1:10 and its distance from the lower corner is about 1/8 of the total length of the slope.

Upon arriving at these data, the correcting of the well-temperatures is straight forward. All steps can be expressed in one equation.

It will now be assumed that the unperturbed outflow of terrestrial heat in Iceland has remained approximately constant for a relatively long period, that is, for more than one million years. The term, perturbation, refers to the surface effects and the magmatic activity. Disregarding the relatively slow pre-Glacial erosion, it is thus being assumed that the regional heat flow at the beginning of the Pleistocene was equal to the unperturbed flow.

Furthermore, it is assumed that the local temperature gradient is approximately constant down to a depth of more than 200 meters. This is actually possible for all processes that involve no changes of the parameters for a period of 5,000 years or more.

In the present case, the only exception is found in the sedimentation process in Location (I). As will be seen below, the correction for this process is quite small and this exception is unimportant.

Accordingly, it is assumed that it will be possible to operate solely with the surface heat flow, that is, with the surface gradient $g_o = (T_y)_{y=o}$ because the thermal conductivity of the basalts is assumed to be constant.

The temperature gradient observed in the wells will be defined as g_o and the unperturbed regional gradient as g_{or}.

The evaluation of equation (51) and the latter term of equation (20), shows that in the present cases the corrections for the magmatic activity and the climatic changes are small. The same applies to the terrain correction in location (III) expressed in equation (56), and the sedimentation in location (I) expressed in equation (38). The figures will be computed below.

Disregarding second order corrections, we may thus take advantage of the linearity of the heat conduction process and write in the case of no erosion

$$g_o = g_{or} + g_{om} + g_{otr} - g_{os} - g_{ocl} \tag{57}$$

or

$$g_o = g_{or}(1 + C_{tr} - C_s) + g_{om} - g_{ocl} \, . \tag{58}$$

The figure g_{om} is the additional gradient due to the magmatic activity and g_{otr} the additional surface gradient due to the terrain anomaly in location III. Furthermore, g_{os} and g_{ocl} are the gradients that have to be subtracted because of the sedimentation in location I and the climatic change in the post-Glacial respectively. The gradients g_{otr} and g_{os} are proportional to g_{or} and can therefore be expressed in terms of the correction factors C_{tr} and C_s as shown in equation (58).

The corrections due to the glacial erosion are relatively greater and can therefore, at the first instance, not be treated in the same simple manner. Although the pre-Glacial temperature gradient g_{or} can be assumed to be constant down to a great depth each period of erosion will introduce second order terms and the gradient cannot be assumed to be completely constant at the beginning of the second to fourth stages. Equation (22) and (23) will therefore not apply as they are based on a constant temperature gradient as initial condition.

On the other hand, expressions for the case of a variable initial gradient are a great deal more complicated than equations (22) and (23). This computational difficulty can, however, be avoided by the method given on pages 74 to 82, where the upper and the lower limits to the erosional effect are estimated. These estimates are accurate enough for the present purpose. The method will be carried out as follows.

In view of the great uncertainty in the dating of the first and the second glacial stages, these will be added to one single stage, the middle of which will be dated back to 450,000 y.b.p.

In correcting for the erosion during the combined first and second stages, and during the third stage, it will be assumed that the erosion occurred instantaneously at the middle of the stages, that is, the correction factor given in equation (28) will be applied. On the other hand, the correction factor for the fourth stage will have to be computed by equation (23).

The various instances of time during the Pleistocene will be denoted as shown in Fig. 14.

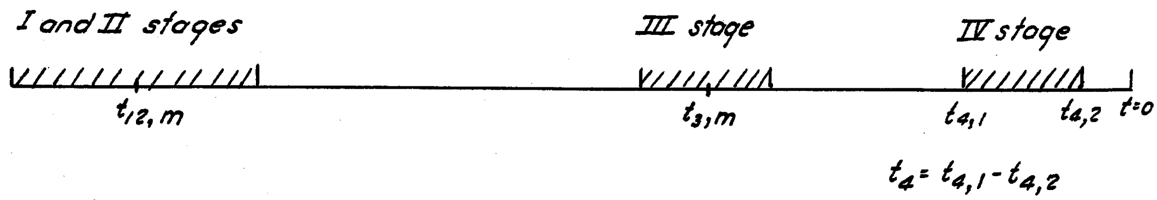


Fig. 14

By disregarding other factors than erosion, the upper limit to the surface gradient at the present is given by the expression $g_{or}P$ where the factor P is found by the inequality (31)

$$P = \left(1 + C_e^{1}(t_{12,m} - t_{3,m})\right)\left(1 + C_e^{1}(t_{3,m} - t_{4,1})\right)\left(1 + C_e^{2}(t_4)\sqrt{\frac{0.26t_4}{0.26t_4 + t_{4,2}}}\right) \quad (59)$$

The corresponding lower limit is expressed by $g_{or}Q$ where the factor Q is found on the basis of the inequality (37)

$$Q = \left(1 + C_e^{1}(t_{12,m}) + C_e^{1}(t_{3,m}) + C_e^{2}(t_4)\sqrt{\frac{0.26t_4}{0.26t_4 + t_{4,2}}}\right) \quad (60)$$

The beginning of the third glacial stage has, in these expressions, been replaced by the time at the middle of the stage, $t_{3,m}$, which will not introduce any further error. The correction factors are according to equations (28) and (23)

$$c_e^{1}(t) = \frac{r}{\sqrt{\pi \alpha t}}$$

$$c_e^{2}(t) = 2R^2 + (1 + 2R^2)\,\mathrm{erf}\,R + \frac{2R}{\sqrt{\pi}}\,e^{-R^2}$$

where $R = r/2\sqrt{at}$ and r is the total erosion during each stage which is assumed to be the same for all stages.

In accordance with Table X, the following time data have been adopted

$$t_{12,m} = 450{,}000 \text{ y.} \qquad t_{3,m} = 190{,}000 \text{ y.}$$

$$t_{4,1} = 58{,}000 \text{ y.} \qquad t_{4,2} = 8{,}000 \text{ y.}$$

A correction for erosion of the surface gradient due to the magmatic activity, g_{om}, will be regarded as a second order term and will be omitted.

The observed gradient in Location (I) corrected for the lower thermal conductivity of the sand will be denoted by g_o'. The average conductivity of the rock penetrated by this well is assumed to be $4\text{x}10^{-3}$ cal/sec °C cm.

The figure g_{om} is in Hypothesis II computed on the basis of an instantaneous input of heat at the time 500,000 y.b.p.

In the case of the sedimentation in Location (I), the approximation (38) can be applied

$$C_s = \frac{1.1 r_s}{\sqrt{a t_{4,2}}} \tag{61}$$

where r_s is the total sedimentation during the post-Glacial. As mentioned above, the thermal diffusivity, a, will have to be corrected in this case for the lower conductivity of the basaltic sand. An average value of $a = 20 \text{ km}^2/10^6$ years will be adopted for equation (61).

Furthermore, according to equation (56)

$$C_{tr} = \frac{h_o}{\pi \ell} \quad \log \frac{x_w^2}{|x_w^2 - \ell^2|} \tag{62}$$

where h_o/ℓ is the slope and $\ell - x_W$ the distance of the well-location from the lower end of the slope as shown in Fig. 13.

Using the same notation as in equation (51), we find

$$g_{om} = \frac{M_o E_o}{c_p \rho \sqrt{\pi a t_m}} \left(1 - \exp(-y_2^2/4at_m)\right) \tag{63}$$

where t_m is the time since the heat input by the magmatic activity and y_2 will be put equal to 5 km.

Finally, the evaluation of the latter term in equation (20) with w = 0 gives for $q_{oc\ell}$

$$u_{cl} = T_{cl} \operatorname{erfc} \frac{y}{2\sqrt{at}}$$

and hence

$$g_{ocl} = T_{cl}/\sqrt{\pi a t_{4,2}} \tag{64}$$

where T_{cl} is the sudden rise of the mean air temperature at the beginning of the post-Glacial. It is to be observed that in the present notation, g_{ocl} is counted positive. A figure of $T_{cl} = 5^oC$ is adopted.

The computation of all corrections are given in Table(XI). The table is based on a value of the thermal diffusivity of the basalts a = 8×10^{-3} cm^2/sec = 25 km^2/10^6y and a thermal conductivity K = 4.5×10^{-3} cal/cm ^{o}C sec. except for Location (I) where corrections for a lower conductivity and diffusivity at the surface are made.

The final computations in Table XI are based on the mean between the upper and the lower estimates of the unperturbed gradient. The difference between both estimates is small and unimportant.

The results of Table (XI) show that the erosion is the main factor affecting the temperature data, and that there is a rather clear correlation between the observed gradients and the amount of erosion as estimated from the topographic map. Other factors are of less importance.

The two hypotheses applied lead to results that differ considerably and it is therefore reasonable to enquire whether the temperature data furnish any clues to the problem of the age of the grey section. This will be discussed below.

The main results are that the unperturbed gradient in the basalts appears to be between 70 and 110 ^{o}C/km and the unperturbed outward flow of heat appears to be between 3×10^{-6} and 5×10^{-6} cal/cm^{2}sec. The term, perturbation, refers to the effects of surface processes and of magmatic activity in the upper 5 km.

Another main result is that the temperature at the depth of 3 km appears to be roughly between 200^{o}C and

TABLE XI

Correction of the well temperatures

		Hypothesis I Location			Hypothesis II Location		
		I	II	III	I	II	III
(1)	g_o °C/km	93	160	110	93	160	110
(2)	g_o'	83	160	110	83	160	110
(3)	g_{ocl}	6	6	6	6	6	6
(4)	g_{om}	0	0	0	7	7	7
(5)	(2) + (3) - (4)	89	166	116	82	159	109
(6)	c_s	0.14	0	0	0.14	0	0
(7)	c_{tr}	0	0	0.04	0	0	0.04
(8)	(5)/(1 - (6) + (7))	103	166	112	95	159	105
(9)	$c_e^1(t_{12,m} - t_{3,m})$	0.02	0.05	0.03	0.06	0.13	0.09
(10)	$c_e^1(t_{3,m} - t_{4,1})$	0.02	0.07	0.05	0.08	0.19	0.13
(11)	$c_e^2\ (t_4)\sqrt{\frac{0.26t_4}{0.26t_4 + t_{4,2}}}$	0.05	0.18	0.12	0.20	0.61	0.36
(12)	P	1.09	1.33	1.21	1.37	2.17	1.68
(13)	$c_e^1(t_{12,m})$	0.01	0.04	0.03	0.04	0.10	0.07
(14)	$c_e^1(t_{3,m})$	0.02	0.06	0.04	0.06	0.15	0.10
(15)	Q	1.08	1.28	1.19	1.30	1.86	1.53
(16)	(P + Q)/2	1.09	1.31	1.20	1.34	2.02	1.61
(17)	(8)/(16)	95	127	93	71	79	65
(18)	Mean (I),(II),(III)		105			72	
(19)	Heat flow cal/cm^{2}sec		4.7×10^{-6}			$3,2\times10^{-6}$	
(20)	Temperature at y = 3 km		315°C			216°C	

300°C. This would probably correspond to the temperature in the lowest part of the basalt plateau.

The above figures are all considerably in excess of what is considered normal. The outward flow of heat appears not less than 3 to 5 times the value in normal stable areas.

(III.3) The source of energy of the thermal activity

Having collected the above information on the temperature-depth relation in the upper few kilometers, we are now in the position of discussing the possible sources of energy of the thermal activity.

(III.3a) The low-temperature activity

The highly abnormal flow of terrestrial heat in Iceland as indicated by the above data, appears to preclude any possibility that the thermal systems of the country are of the normal type as defined on page 6.

Furthermore, the extensive volcanism in Iceland, both during the Tertiary and the Quaternary, suggests that the low-temperature activity is either of the intermediate volcanic type, or of the direct volcanic type.

This conclusion is, however, not as straight forward as indicated. It is quite possible that the low-temperature activity is mainly the result of the abnormal outflow of heat from deep sources, and that volcanic intrusives are of less importance. For example, there is a possibility that there has been no active volcanism in the northern parts of the country for several millions of years but there is, nevertheless, abundant low-temperature activity in this part of the country.

The data obtained so far do not furnish decisive clues to this problem. The question whether the abnormal heat flow is the direct result of the volcanism, or whether both are the result of a deep-seated anomaly, will be discussed later.

The present problem is, therefore, mainly to decide whether the low-temperature activity could possibly be of the direct volcanic type, that is, whether the heat is mainly supplied by upflowing magmatic water.

It has already been stated that neither the chemical composition of the dissolved solids or gases nor the isotopic ratios in the thermal water of the low-temperature activity suggest intermixture with magmatic water. There are no components that can not be accounted for by normal leaching from the basaltic country rock. Furthermore, the isotopic ratios are in the same range as those of normal surface water in Iceland.

The only suspected exceptions are the few small and cool carbonated springs on the Snæfellsnes peninsula as indicated on page 34. The isotope ratios of these springs have not been studied.

The next question is whether it is possible to maintain the energy supply of the low-temperature activity without an intermixture with upflowing volcanic water.

The total heat transported by the low-temperature group has been estimated at 2 to 3×10^8 cal/sec. On the other hand, the outward flow of terrestrial heat has, according to Table (XI) been estimated at 3 to 5×10^{-6} $cal/cm^2 sec$ which gives a total of 1.5 to 2.5×10^9 cal/sec for the entire western half of the country. As nearly all low-temperature areas are located in this part of the country, the total heat flow transported by this group appears to be roughly 10% of the total available outflow of heat, if steady state conditions are assumed.

This ratio appears quite high mainly because of the fact that a greater part of the heat flow is transported by a few large low-temperature areas that are located in the south-western parts of the country. This indicates some difficulties in maintaining the heat supply on the basis of steady state conditions.

The question therefore arises whether the low-temperature activity could possibly be a more or less transient process drawing upon the heat content of the rock of the heating zones of the hydrothermal systems rather than a steady state process maintained by the steady outward flow of terrestrial heat.

The effects of the deglaciation

There is no doubt that considerable hydrological and geological changes resulted from the last deglaciation of the country.

The ground water conditions during a period of total glaciation can be expected to be different from the present conditions. The total circulation of ground water was probably increased at the deglaciation. The pattern of flow may also have been changed considerably by the glacial erosion.

Finally, the upcoming of Iceland after deglaciation may have resulted in the opening of fractures and the formation of new faults and fractures. The upcoming amounted to 50 to 80 meters in coastal areas but probably to some 200 meters in the center of the country. It may be remarked that the first part of the post-Glacial appears to have been a period of frequent large landslides (personal communication by Thorarinsson).

Although these arguments are somewhat speculative, there are, nevertheless, reasons for believing that the deglaciation may have resulted in an increased circulation of ground water, and hence an increased low-temperature activity. The effect of the erosion on the subsurface temperatures has already been discussed. We are, therefore, arriving at the conclusion that the Pleistocene glaciation may have been one of the promoters of the widespread thermal activity in Iceland.

Transient heating of the water

The question whether there is a transient component in the heat transport of the low-temperature activity may now be discussed on the basis of a simple model.

It has already been mentioned that there are good reasons for assuming that the low-temperature activity is mainly maintained by a deep flow of ground water along the systems of fissures that control the individual hydrothermal systems. The depth of the fissures is not known, but they probably reach at least the bottom of the basalt plateau, that is, a depth of 2 to 4 kilometers. The slight scatter of the alinement of the surface springs indicates that each hydrothermal system is controlled by a system of parallel fissures.

The four hydrothermal systems H(I) to H(IV) shown in Fig. (1) appear to be controlled by a rather dense system of fissures. The distance from H(I) to H(IV) is only about 20 km. These four systems issue in surface springs an integrated flow of $320 \times 10^3 cm^3/sec$ and transport at least $3 \times 10^7 cal/sec$ of heat. Due to subsurface losses, the actual flow may be greater, and we will tentatively assume that all four systems carry a flow of $600 \times 10^3 cm^3/sec$ of water at $100^\circ C$.

The flow of water along a number of deep densely distributed vertical fissures may, from the thermal point of view, in a rough approximation, be regarded as equivalent to the flow along an extensive horizontal fissure at a constant depth as shown in Fig. (15)

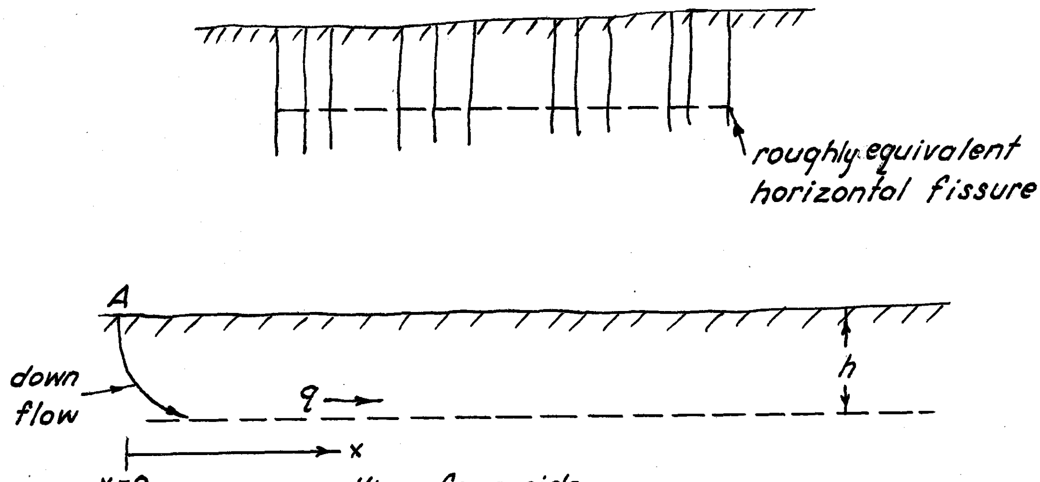


Fig. 15

It will now be assumed that the water enters the horizontal fissure at the location A and that q mass units of water flow per unit time and unit breadth of the fissure. This flow shall be equal to the integrated flow through the vertical fissures. The depth of the horizontal fissure is assumed to be h . The flow q is assumed constant and relatively large so that the temperature gradient along the fissure is much less than the vertical temperature gradient in the area.

At steady state conditions, the following approximate equation applies

$$s_p q \frac{dT}{dx} = Q - KT/h \qquad (65)$$

where s_p is the specific heat of the water, T its temperature at the point x, Q the steady outward flow of terrestrial heat per unit area and unit time and K the thermal conductivity of the overlying rock.

If T_o is the temperature at the depth h, the above equation has the solution

$$T = T_o(1 - \exp(-Kx/s_p qh)) \tag{66}$$

where it has been assumed that the water enters the fissure at zero temperature.

Assuming that the breadth of the fissure is 20 km and that the total flow is $600 \times 10^3 \, cm^3/sec$, we find $q = 0.3 \, cm^3/cm \, sec$. Assuming furthermore that h = 3 km and $T_o = 250°C$, we find that for x = 50 km which is a reasonable length of the system that by $K = 0.0045 \, cal/cm°C \, sec$

$$Kx/s_p qh = 0.25 \qquad \text{and} \qquad T = 55°C$$

which is only about 1/2 the temperature of the springs of the systems which we are dealing with.

The next step consists of assuming that the main flow of these sytems was initiated at the deglaciation of the country and of calculating the transient component of the heat supply.

This computation can be facilitated very much by using a solution of a similar problem given by Carslaw ((46), page 329).

Carslaw solves the following problem. Given a semi-infinite solid in the space $x > 0, z > 0, -\infty < y < \infty$, with the constant thermal conductivity K and constant thermal diffusivity a. The surface x = 0 is insulated from the surroundings. A fluid which is insulated from the outer surroundings moves along the surface z = 0 in the direction x. The flow starts at x = 0 and is constant in time and space. The amount of fluid in contact with the unit area of the plane z = 0, is M, and the velocity of the fluid in the direction x is U, that is, the flow is q = UM mass units per unit time and unit length perpendicular to the x axis.

The temperature of the fluid at x = 0 is constant and equal to unity. At the time t = 0, the fluid elsewhere and the solid have the temperature zero.

It is assumed that the flow q is relatively large and that the temperature gradient with respect to x is much less than the gradient with respect to z. In this case, the approximate solution of this problem is

$$u = \operatorname{erfc}\left[\frac{Kx}{2s_p UM\sqrt{a(t-x/U)}}\right] \tag{67}$$

where u is the temperature of the fluid and s_p its specific heat.

The form of the solution (67) is not directly applicable to the present problem. The conditions will now be changed so that the temperature of the fluid at x = 0 is zero, but the initial temperature of the solid and of the fluid elsewhere is T_o'. It will furthermore be assumed that M is relatively very small but U relatively large. The approximate solution for this case is easily derived from (67) and is

$$u = T_o' \operatorname{erf}\left[\frac{Kx}{s\rho qh} \frac{h}{2\sqrt{at}}\right], \tag{68}$$

where the constant h has been introduced for practical purposes.

The steady state solution (66) for the model treated above showed that at equilibrium conditions, the water was heated from zero to 55°C only over a distance of 50 km. The depth of flow was assumed at h = 3 km and the unperturbed temperature at this depth was assumed T_o = 250°C.

In seeking the solution of the transient problem for the system shown in Fig. (15), it is in the first rough approximation possible to disregard the heat capacity of the rock overlying the fissure, and also three-dimensional effects at the ends of the fissure. These assumptions will only minimize the transient effect.

Using these simplifications, and putting T'= 250 - 55 = 195°C, we find that the sum of the expressions (66) and (68) represents an approximate solution to the transient problem where it is assumed that the flow started at the time t = 0.

Inserting h = 3 km and assuming that the flow was initiated at the beginning of the post-Glacial, that is,

$t = t_{4,2} = 8{,}000$ years, we find the following present temperatures of the water in the horizontal fissure

x	25	50 km
T	30	55 °C
u	86	148
T + u	116	203

As a number of the data applied are quite arbitrary, the above results can, as a matter of course, only give at most a semi-quantitative picture of the actual conditions. However, if the flow in the low-temperature systems was initiated not much earlier than at the time of the last deglaciation, the importance of the transient component appears evident.

Field evidence of transient conditions

It is at this juncture appropriate to inquire whether there are other indications of a transient nature of the low-temperature activity. In fact, there is such evidence.

The writer has observed that there is a slight temperature decrease in the lower parts of a few wells that have been drilled through horizontal strata carrying thermal water. In general, the drilling is discontinued when the wells have been drilled into such horizons and the data available therefore very scarce.

An example is shown in Fig. (16) which represents the temperature measurements in a well in Location (9) in Figure (5). This well was drilled by the Municipality of Reykjavik around 1940. The temperature data are taken from the drillers notes and were obtained by the means of an enclosed maximum thermometer. The measurements were made at the well-bottom at intervals during the drilling of the well and are therefore not very accurate.

A thin water-bearing horizon was encountered at the depth of 130 meters. A flow of 650 cm^3/sec at a surface temperature of 43°C was obtained from this horizon. The drilling was continued down to a depth of 190 meters where it was discontinued because of the decrease of the bottom-temperatures. The negative gradient between 130 and 190 meters amounts to roughly 130 °C/km.

The writer suggests that the temperature decrease be interpreted as a transient effect resulting from the fact that the flow of thermal water through the horizon at the depth of 130 meters was initiated relatively recently.

The "age" of the flow may be estimated on the basis of this hypothesis. It will be assumed that the temperature of the horizon was suddenly raised from 15°C to 45°C at the time $t = 0$. This case may tentatively be treated as the case of a semi-infinite solid with an initially constant positive temperature gradient of 70 °C/km and a sudden rise of the surface temperature from zero to 30°C. Equation (64) can be applied and gives in the present case

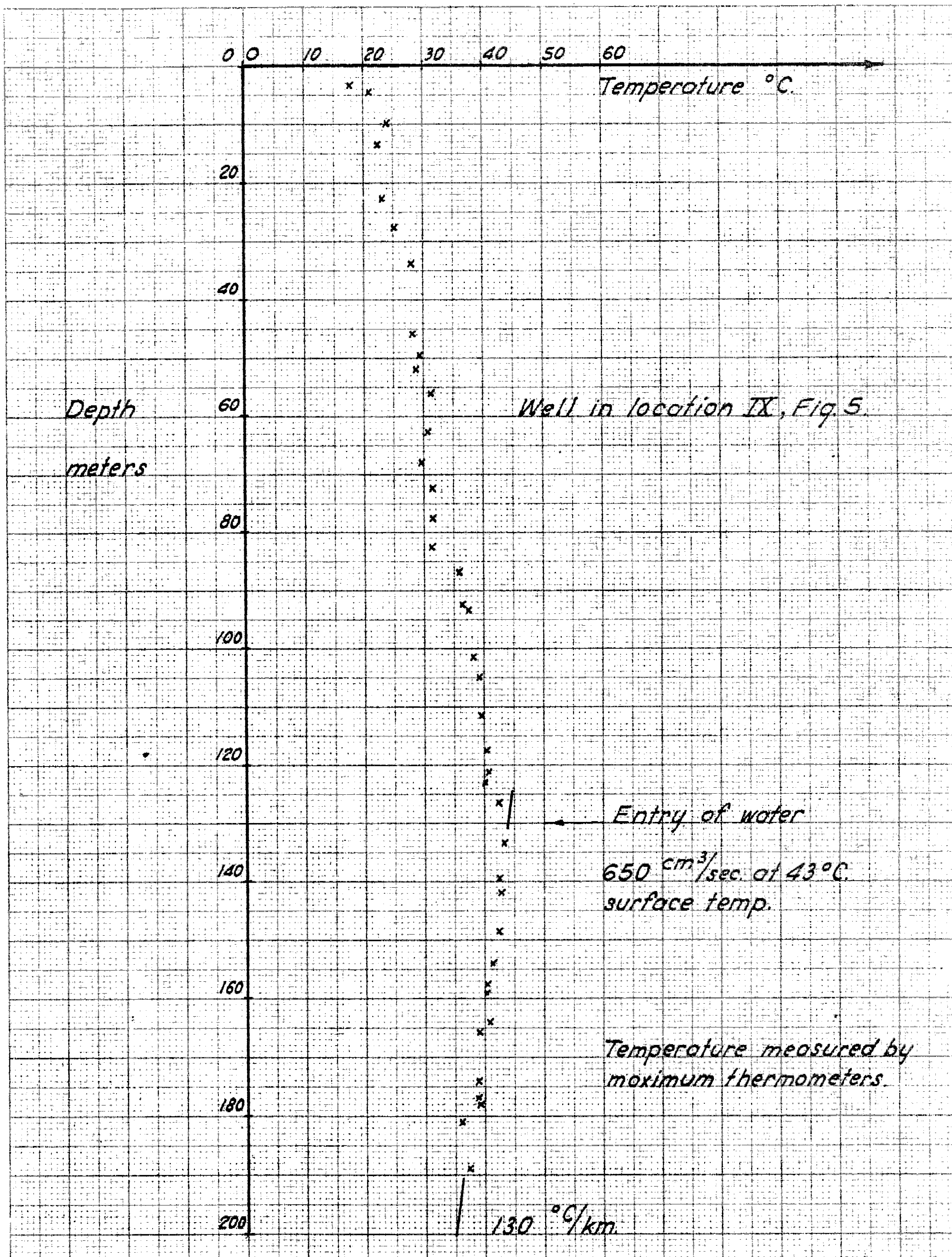

0
0
10
20
30
40
50
60
Temperature °C
20
40
Depth
60
meters
Well in location IX, Fig. 5.
80
100
120
Entry of water
650 cm³/sec at 43°C
surface temp.
140
160
Temperature measured by
maximum thermometers.
180
200
130 °C/km

Fig. 16

$$70 + 130 = 30/\sqrt{\pi a t} \qquad (69)$$

where t is the time since the temperature rise. Putting $a = 25\ km^2/10^6 y$ we find t = 300 years.

This surprisingly low "age" can probably be attributed to the fact that the case was treated on the basis of a one-dimensional model. Taking into account three-dimensional effects, that is, the effect of a limited extent of the water-bearing horizon would increase the estimated age substantially. There are, however, no data at hand on the extent of the horizon. This is, therefore, only an example of a qualitative nature showing the possible existence of a phenomena that can be interpreted as a transient effect.

The above interpretation is, as a matter of course, not unique. By a relatively very small extent of the water bearing horizon, the above temperature conditions are possible even at steady state conditions. Further evidence must come through the collection of additional data.

Transport of heat by magmatic water

The chemical and isotopic data have furnished negative evidence as to the presence of any magmatic components in the water issued by the low-temperature springs. The data on the heat balance and the temperature conditions in these areas lend further support to this result.

One of the hydrothermal systems in northern Iceland, for example H(VIII), may be considered. This system issues a total of $70x10^3$ cm^3/sec of water at a maximum surface temperature of 89^oC. Because of possible subsurface losses the maximum flow of the system may be greater. A figure of $140x10^3$ cm^3/sec will therefore be assumed which gives a total heat transport of $1.3x10^7$ cal/sec.

It is reasonable to assume that magmatic water that possibly enters this sytem is near the critical point and has, therefore, an enthalphy of some 500 cal/gm, that is, some 400 cal/gm are available for heating of the circulating meteoric water. The above heat transport therefore corresponds to a flow of magmatic water of some $3x10^4$ gm/sec.

Very little is known about the amount of water that can be contained in magma. Ingersoll (56) reviews this question and expects an average of some 3% by weight. Einarsson (57) has studied the water content of the basaltic-andesitic magma that was erupted by the volcano Hekla in Iceland in 1947-48. He reports a water content of less than one % by weight. Fries (58) finds a figure of 1.1% for the lava of the Parichutin in Mexico.

It may therefore be expected that 5% by weight is a relatively high estimate. This corresponds to some 0.15 gm water/cm^3 magma and H(VIII) would thus have to consume the water from $2x10^5$ cm^3 magma/second. In the post-Glacial period, this adds up to some 50 cubic kilometers of magma.

There are no signs of any recent magmatic activity in the vicinity of H(VIII) and in particular, no signs of any recent intrusion of this size. The last period of volcanism in this area dates back to the middle Pleistocene, at least.

Another point of interest is that intrusions of the more common forms, dikes and sills for example, solidify relatively very rapidly. The temperature at the center plane of a dike intruded at the temperature T_o into an infinite solid of zero temperature is found to be (Carslaw (46) page 43) very roughly

$$T = T_o \, \mathrm{erf} \, \frac{h}{2\sqrt{at}} \tag{70}$$

where 2h is the thickness of the dike and t the time after its intrusion. This equation is based on heat conduction alone, but the transport of heat by other processes, ground water convection for example, will increase the rate of cooling.

Putting a = 10 $km^2/10^6$years, where the latent heat of melting has been taken into account, and T_o = 1,100 ^oC gives the following time for the cooling of the center plane down to 600^{o}C in an environment at 300^{o}C

2h	t
10 meters	5 years (approx.)
100	500
1,000	50,000

Any intrusive active in the expulsion of supercritical water must therefore be of a very recent age. As already underlined, there is no sign of any such magmatic activity in the vicinity of H(VIII).

Conclusion

The low-temperature activity appears to be associated with the abnormal conduction heat flow in Iceland. It appears to be the result of a circulation of meteoric water in the deep fissures and fractures of the basalt plateau, the base depth being probably several kilometers. A considerable part of the heat may be supplied by the contact of the water with the hot rock at this depth. Transient phenomena initiated by the geological changes at the end of the last glaciation of the country appear to be of considerable importance.

(III.3 b) The high temperature activity

The integrated heat transport of the high-temperature activity has been estimated at 0.3×10^9 to 1.5×10^9 cal/sec which is considerably in excess of that of the low-temperature activity. In fact, the figure may be of the same order as the integrated conduction heat flow in the western half of the country.

These conditions imply that the high-temperature activity cannot be associated with the abnormal conduction flow and the resulting abnormal subsurface temperature alone.

It appears to draw upon additional heat sources. The fact that all high-temperature areas are in the immediate neighborhood of volcanic centra of recent activity suggests that the bulk of the heat is supplied by large recent intrusives, that is, intrusives of late Pleistocene or post-Glacial age.

It has already been estimated that subsurface magma could possible supply 0.15 gm water/cm^3 and thus supply an amount of heat of some 60 to 120 cal/cm^3 of magma. On the other hand, the total heat of melting of magma is probably not less than 1,200 cal/cm^3, that is, about 10 to 20 times the possible heat content of the supercritical water that may be expelled by the magma.

These conditions suggest that the bulk of the heat transport of the high-temperature activity is acquired by a contact of the circulating meteoric water with recent intrusives. This result is further supported by the fact that neither the chemical nor the isotopic composition of the water from one of the major high-temperature areas, NS (IV), indicates the presence of any magmatic component. (See Table V, sample (12)).

The present result is therefore that the high-temperature activity appears to be a transient phenomenon drawing upon the heat content of large intrusives of recent age. The underlying mechanism of heat supply, therefore, appears quite similar to that of the low-temperature activity, that is, both groups acquire the heat by a contact of meteoric water with rock of high temperature. In the case of the low-temperature activity, the abnormal temperatures are associated with the abnormal conduction flow of heat, whereas the upflow of magma into near surface layers appears to be the main cause in the case of the high temperature activity.

The heat transport of the high-temperature activity may possibly have been steady throughout the post-Glacial period. The transport of 10^9 cal/sec corresponds to a total of 2.5×10^{20} cal during this period. It may tentatively be assumed that some 500 cal/cm^3 are available from the intrusive rock for the heating of the circulating water. The total heat transport of the high-temperature activity, if assumed steady during the post-Glacial, would thus correspond to the available heat from 500 km^3 of intrusives. This is, in fact, equal to the total post-Glacial production of magma by volcanos in Iceland. (See below).

The discussion above has mainly been concerned with the heat supply of the thermal activity in Iceland. The role of the structural control has also to be underlined. It appears to be the unique combination of an adequate heat supply and multiple systems of deep fractures, dikes and other structures that gives rise to the very widespread thermal activity in Iceland.

(IV) THE TERRESTRIAL HEAT BALANCE IN ICELAND AND RELATED PROBLEMS

There are three processes that contribute to the outward transport of terrestrial heat in Iceland, that is, outward conduction, thermal activity and volcanism. It has been concluded that the thermal activity is not an independent process. A part of it is probably associated with the highly abnormal outward conduction, and another part appears to be associated with the volcanism.

The heat flow that is transported by conduction and thermal activity has already been estimated. The heat that has been transported to the surface of Iceland by magma in post-Glacial time can be estimated on the basis of Einarsson's (53) estimate of the total amount of lava erupted during this period. His conclusion is that the post-Glacial lava if averaged over the total area of the country, amounts to approximately 5 meters. Estimating the post-Glacial period at 8,000 years, the total heat content of the lava at 450 cal/gr and its density at 2.7, we find an average in space and time of 2.5×10^{-6} cal/cm^2 sec.

If the results of Table (XI) are assumed representative for the whole country, the heat balance expressed in round figures becomes

Outward conduction	3 to 5×10^{-6}	cal/cm^2 sec
Thermal activity	0.5	2
Volcanism	2.5	2.5
Sum	6 to 9.5×10^{-6}	cal/cm^2 sec

These are estimated averages in space and time for the post-Glacial period. The figures are exceptionally high. The average outward conduction of terrestrial heat in normal areas is 1.2×10^{-6} (Birch (48)).

In fact, according to the recent review of Birch (48) and Bullard, Revelle and Maxwell (59), there is only one location where a higher outward conduction of heat has been reported. A figure of 5.25×10^{-6} cal/cm^2 sec was found by Revelle and Maxwell in one location of the Albatross plateau west of South America.

It is of interest to compare the conduction component of the icelandic figures to that of other volcanic areas. Data from only two other locations are at hand.

Anderson (60) reports three figures for the Tertiary basalt areas of Scotland. All are based on temperature measurements in wells located in the close vicinity of the Mull dike-swarm. Anderson reports figures of 2.1×10^{-6} and 1.8×10^{-6} from the neighborhood of Glasgow and 1.8×10^{-6} at Durham.

The same author (Anderson (60)) reports figures of 1.2×10^{-6} and 1.3×10^{-6} from a non-volcanic area some 80 kilometers east of the Mull swarm.

For various parts of England, Benfield (61), Bullard and Niblett (52), and Chadwick (62), report figures ranging from $1.1x10^{-6}$ to $1.7x10^{-6}$. Bullard found figures up to $2.9x10^{-6}$ in one location in Nottinghampshire, but this anomaly is believed to be caused by the upflow of ground-water.

Anderson's figures for the volcanic areas of Scotland appear 0.3 to $0.9x10^{-6}$ cal/cm^2 sec higher than the average for the areas of Great Britain that have not been subjected to Tertiary volcanism. The figures are especially interesting for the fact that the western parts of Scotland have been subjected to recurrent periods of volcanism since the beginning of the Paleozoic. The last period of volcanism was probably in the middle or early Tertiary, that is, 30 to 50 million years ago. It can be shown on the basis of equation (51) that middle, or early Tertiary intrusives above the depth of 25 km, can only play an insignificant part in this anomaly.

Revelle and Maxwell (59) report a figure of $1.17x10^{-6}$ for a location west of the Hawaiian Islands.

The comparison of Anderson's figures for the conduction heat flow in the volcanic districts of Scotland to the Icelandic figures, suggests that the excessive conduction of terrestrial heat in Iceland is closely associated with the present period of volcanism in the country. The term present applies to the volcanic period that started by the formation of the grey section. This relation will be discussed below.

(IV.1) <u>The cause of the abnormal outward conduction of heat in Iceland</u>.

The conduction heat flow in Iceland was estimated at $3x10^{-6}$ to $5x10^{-6}$, that is, an average $4x10^{-6}$ cal/cm^{2}sec. A normal component of $1x10^{-6}$ may be deducted from the average figure giving a residue of $3x10^{-6}$. The normal component represents approximately the average normal heat generation by radioactive processes in the upper few tens of kilometers.

It may now be assumed that the concentration of radioactive elements causing the normal component is evenly distributed down to the depth h. The temperature at this depth due to this component alone is therefore $T_{nh} = Q_n h/2K$ where Q_n is the normal outflow of heat and K the conductivity of the rock which is assumed constant (see Carslaw (46), page 60).

Assuming h = 25 km, K = 0.006 cal/cm°C sec and putting $Q_h = 1x10^{-6}$ cal/cm^2 sec, we find a temperature T_{nh} = 210° C or approximately 200° C.

Furthermore, the abnormal component $3x10^{-6}$ cal/cm^2 sec gives by K = 0.006 a temperature gradient of 50 °C/km. This implies that a temperature of 1,500 °C would already be reached at the mere depth of 26 km if the total abnormal component originates below this depth. As the melting point of the material at this depth cannot be much in excess of 1,500 °C, this would by steady state conditions imply a state of general fusion below the depth of 25 to 30 km.

The state of general fusion at this level is highly improbable and can be ruled out on various grounds. The observation by Einarsson (53) that the post-Glacial upcoming of Iceland lagged a few thousand years behind the deglaciation, is probably the most direct evidence against a large fused layer under the country.

It is improbable that the heat conductivity of the material has been underestimated. At this relatively small depth, the pressure is not far from the conditions that can be simulated in laboratories. The experiments of Bridgeman (63) have shown that the lattice conductivity of basalt increases only about 0.2 to 0.5% for a pressure change of 1,000 bars, that is, for a change in depth of 3 km.

The immediate conclusion is therefore that a great part of the abnormal conduction flow must either originate above the depth of 25 to 30 km, or it cannot be transported through this level by conduction alone.

Five factors that may contribute to the solution of this problem may be listed

(1) Abnormally high concentration of radioactive elements in the upper few tens of kilometers.

(2) Heat release in the upper few tens of kilometers by the dissipation of mechanical energy.

(3) Radiative transfer of heat.

(4) Convection currents under the region around Iceland.

(5) Heat transport by upflowing magma.

Abnormal Radioactivity

Jeffreys (64) has studied the concentration of radioactive elements in the Icelandic effusives. He reports no abnormal values for the icelandic samples, and reports them in the same groups as the samples from other parts of the North Atlantic area, that is, Scotland, Ireland and Greenland.

Mechanical energy

According to Gutenberg and Richter (65), the total energy released annually by earthquakes is in the average around 10^{25} erg, that is, around 7×10^{9} cal/sec. This is of the same order as the total outward transport of terrestrial heat in Iceland. The possibilities that the dissipation of mechanical energy is of any importance are therefore rather small.

Radiative transfer of heat

Clark (66) has recently pointed out on the basis of experiments of McQuarrie (67), that the radiative transfer of heat through rocks may be of significance at temperatures above 1,300 °C to 1,500 °C. He expresses the effective conductivity (see van der Held (68))

$$K_e = K + bT^3 \tag{71}$$

where K is the lattice conductivity and b a constant which is estimated at 0.5×10^{-12} to 1.1×10^{-12}. The temperature is measured in absolute degrees.

The radiative contribution to the conductivity is negligible below 1,000 °C. The increase of the mean conductivity in the interval 0 to 1,500 °C due to the radiative term is probably only 0.001 to 0.002 cal/cm °C sec. The radiative transfer of heat can therefore not be of primary importance in the present case. It may increase the computed level of fusion from 26 km to 35 km.

Convection

Various authors (Verhoogen (69), Griggs (70)) have stressed the possibility of large scale convection currents in the mantle. Bullard, Revelle and Maxwell (59) apply this hypothesis in order to explain some of the anomalous data on the heat flow in the Pacific. Birch (71) has, on the other hand, pointed out some serious arguments against the existence of convection currents in the earth's present state of development.

Additional difficulties are encountered in applying the convection hypothesis in order to explain the anomalous conduction heat flow in Iceland. In order to prevent fusion in the present case, the currents have to transport an appreciable amount of heat up through the 30 to 40 km depth level. This is improbable in view of the large scale of movement necessarily involved.

Heat transport by ascending magma

There appear no serious difficulties in assuming that relatively large amounts of magma have been intruded into the upper 10 to 20 km under Iceland during the present period of volcanism.

For example, Richey (13) reports that the Tertiary dike-swarms of Scotland may have originated partially from plutons at a depth of a few miles, and partially from a regional magma layer at a greater depth.

The basalt plateau of Iceland has been invaded by a conspicuous number of dikes, sills and other intrusions which may in a similar way originate in large plutonic masses at a relatively shallow depth. The solidifying and cooling of these masses could account for the abnormal conduction flow of heat.

A semi-quantitative test of this hypothesis is possible on the basis of equation (48). At the first instance, it may be assumed that Iceland has been subjected to a period of volcanism of a length of 2 million years which could represent an average between Hypothesis I and II. This period would include the present period of volcanism in the central and southern parts and possibly, but not necessarily, also the formation of the grey section in the western, northern and eastern parts.

It will be assumed that large amounts of magma have, during this period, been intruded into the upper 15 km of the crust under Iceland. An estimate will be made of the amount necessary in order to lead to the present abnormal conduction heat flow.

The integrated effect of this period of multiple intrusion may be approximated very roughly by assuming that the intrusives have been formed at a constant rate, and that they are evenly distributed in space. Equation (48) would thus apply as a first approximation.

Equation (48) may in this case be written

$$g_{om} = \frac{2M_o E_o}{c_p \rho \sqrt{at}} \left(\frac{1}{\sqrt{\pi}} - \operatorname{ierfc} \frac{d}{2\sqrt{at}} \right), \qquad (72)$$

where g_{om} is the surface temperature gradient at the time t since the beginning of the period of intrusions. M_o is is the total mass of magma intruded into the unit volume country rock, and E_o the sensible heat content of the magma. Finally, d is the depth of the layer of intrusions, that is, 15 km in the present case.

The abnormal conduction flow of 3.0×10^{-6} cal/cm^2 sec gives on the basis of an average conductivity in the upper 15 km of 0.006 cal/cm ^{o}C sec an average gradient of 50 ^{o}C/km. Assuming furthermore that a = 30 km^2/10^{6}y, we find that the latter term in the parenthesis amounts to about 0.05 and the parenthesis to 0.5. This gives

$$50 = M_o E_o / c_p \rho \sqrt{2 \cdot 30} \qquad (73)$$

or $M_o E_o$ = 240°C if c_p = 0.21 cal/°C gm and ρ = 2.9 gm/cm^3.

If the average temperature of the region above the depth of 15 km during the last 2 million years is assumed at 500°C, the sensible heat content of the magma may be estimated at E_o = 300 cal/gm and hence M_o = 0.8 gm/cm^3, that is, about 1/4 of the rock above the 15 km level must consist of intrusions formed in the last 2 million years. This is based on conditions that average between Hypothesis I and II.

We may now, as an extreme alternative, assume the heat flow in accordance with Hypothesis II to be 3.6×10^{-6} cal/sec cm^2. This figure has been obtained by omitting correction (4) in Table XI and taking into account the shorter length of the volcanic period applied below. The deduction of a component of 1.8×10^{-6} in accordance with Anderson's data (60) gives an abnormal component of 1.8×10^{-6}. Furthermore, by reducing the present period of volcanism to 500,000 years, we find M_o = 0.22 gm/cm^3. This amounts to about 8% by volume of the country rock, that is, an average section of intrusives of 1.2 km in the region above the depth of 15 km.

The latter result is not unacceptable. It indicates that the present theory can be accepted mainly on the basis of Hypothesis II in combination with somewhat abnormal radioactive generation of heat. The above figure of the volume fraction of intrusives may be overestimated because of the influence of convection in larger unsolidified plutons.

Conclusion

The discussion above indicates that the abnormal conduction flow of heat in Iceland can be understood on the basis of the presence of relatively large magmatic intrusives in the upper 10 to 20 km in combination with slightly abnormal generation of heat.

The heat flow figures of Hypothesis II appear more acceptable. This indicates a considerable erosion during the Pleistocene, and favors an early Pleistocene age of the grey section. This result is in accordance with the recent paleomagnetic data (see page 4).

(IV.2) The heat balance of the Tertiary period

The total thickness of the basalt plateau of Iceland has been estimated 3,000 to 6,000 meters. An average thickness of 4,500 meters appears to be a reasonable guess. The section may contain some 10% of sediments and the total effusives should thus average to 4,000 meters.

The lowest section appears to be of Eocene age, that is, the total plateau has been built up during the entire Tertiary period which, according to recent data, lasted for approximately 60 million years.

The heat content of basaltic lava is about 1,200 cal/cm^3. The mean outward transport of heat by surface lava in Iceland has therefore during the Tertiary amounted to some 0.25×10^{-6} cal/sec cm^2 if averaged over the total area of the country.

There are good reasons for believing that the lava originates at a depth where the normal temperature is not far from the melting point of the material. The above figure is therefore an overestimate of the heat directly involved in the melting. It appears reasonable to assume that only 1/3, that is, some 0.08×10^{-6} cal/sec cm^2 were used in the process of melting.

On the other hand, it is to be realized that the data in the present paper indicate that the surface lava is only a fraction of the total magma created. To this we must add the possibility that a longer time than the Tertiary period may have been available for the creation of the magma. Taking both factors into account, the total mean output may be of the order of 0.1 to 0.2×10^{-6} cal/sec cm^2 instead of the above mentioned 0.08×10^{-6}. This is 10% to 20% of the outflow of terrestrial heat in normal areas.

The heat flow data in western Scotland found by Anderson were 1.8 to 2.1×10^{-6} cal/sec cm^2, that is, some 0.3 to 0.9×10^{-6} above the average for other parts of Scotland and England. Although a marginal area, there is a possibility that western Scotland reflects conditions in the North Atlantic flood basalt area as unaffected by volcanism during the last 20 to 40 million years.

Anderson's data are markedly above the average, but they are not exceptional. Data of the same order are reported from the Pacific by Bullard, Maxwelle and Revelle (59). Similar data are also reported by Birch (34) for the Front Range, Colorado. The conditions in the Front Range can be understood in terms of mountain roots.

A comparison of the data from western Scotland to those from the Front Range raises the question whether the extensive volcanism in Scotland, and in the whole North Atlantic area, can be understood in terms of the moderately abnormal outflow of heat observed in western Scotland.

This is possible if the bulk of the radioactive elements in the region below Scotland is distributed to a greater depth than in the non-volcanic areas.

Let the concentration of radioactive elements be constant down to a depth h and the total heat flow produced in the section down to this depth be Q_h. The steady state temperature at the depth h due to the radioactivity is then

$$T_h = Q_h h/2K \tag{74}$$

where K is the thermal conductivity which is assumed constant (Carslaw (46) page 60).

Equation (74) shows that the temperature is proportional to the depth. As an example we may mention that by $Q_h = 2.0 \times 10^{-6}$ cal/sec cm^2 and $K = 8 \times 10^{-3}$ cal/sec °C cm, a temperature of 1,600 °C is possible only if h = 130 km. This indicates that a depth of not less than of the order of 100 km is necessary in order to lead to melting on the basis of excess radioactivity.

It may be inferred on the basis of equation (47) that temperature transients at this depth are extremely slow and may require times of the order of several hundred millions of years. Volcanism of the present kind based on excess radioactivity would thus have to be a relatively very slowly progressing phenomenon. This is actually indicated by the situation in Scotland where 7 periods of volcanism have been recorded since the beginning of the Paleozoic, that is, during some 500 million years. Each period appears to have been quite brief, and the intervals of quiescence very long. For example, the Mesozoic which may have lasted some 150 million of years appears to have been a period of complete quiescence.

It may at this end be mentioned that Uffen (72) has proposed a method of estimating the melting point of the earth's material on the basis of the equation

$$T_m = Ck/\rho \qquad (75)$$

where T_m is the melting point in absolute degrees, k the bulk modulus, ρ the density and C a figure which should be approximately constant for the class of materials involved.

The theoretical foundation of Uffen's equation is very weak and there are as yet very few experimental data at hand in order to test his proposition.

However, Uffen's equation may possibly indicate that the discontinuity of the seismic gradients found by Gutenberg (73) at the depth of some 200 km, is associated with a discontinuity of the melting point gradient. The melting point gradient may increase rather abruptly at this depth. This could indicate a possible lower limit for the formation of magma.

APPENDIX

Data on the chemical composition of thermal water and natural steam

In addition to Tables II and VIII, the following Tables XII and XIII present further data on the chemical composition of thermal waters and natural steam in Iceland. The material listed is selected from the data collected by the writer and Mr. Baldur Lindal. Most of the water samples and all gas samples were analyzed in the laboratory of the State Electricity Authority in Reykjavik. Some water samples were analyzed by the University Research Institute in Reykjavik.

The present collection of data includes mainly those components that are of importance in the study and exploration of the thermal areas. The locations of the springs where the water samples were collected are given in Fig. 17 and 18. The analytical methods are listed below.

The temperature is generally measured at the vents by ordinary thermometers. Most of the flow data are given in the references (11) and (20).

The pH data were measured by a Beckman pH-meter. The conductivity was measured by the means of a Leeds & Northrup AC-bridge which is accurate to ± 1%.

The data on the F were obtained by a colorometric method given by Snell ((74) p. 583). The reagents used are Zirconium Nitrate and 1,2,5,8 Tetrahydrooxyantraquinone.

The SO_4 anion was generally measured turbidometrically by the method given by Snell ((74) p. 611). This involves precipitation in acid solution with BaCl.

The SiO_2 data were obtained colorimetrically by the standard ammonium molybdate method. (Snell (74) p. 517). Gravimetric data on the SiO_2 were obtained in a few cases. Two of these are given in Table XII, that is, H(III)(1) and H(IV)(2) where the second figure in the column for the SiO_2 represents the gravimetric data.

Hardness is given in p.p.m. CaO and was tested by the standard soap-method.

The alkalinity was measured by the standard HCl-titration method with a 1/10 N acid solution. The data are given in millivalence/liter, that is, the amount of cm^3 of 1/1 N acid required in order to neutralize a liter of the water sample.

The gas samples were analyzed by a Fisher gas-analyzer delivered by the Eimer & Amend Co., New York, N.Y.

TABLE XII

Composition of thermal water

Hydrothermal System (I)

Location	Temp. °C	Natural Flow x 10^{-3} cm^3/sec	pH	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/1
(1)	67	0.7	9.1	0.057	49	3.0	84	8.2	118	2.2
(2)	100	2.5	9.4	0.054	41	2.5	72	5.0	181	3.0
(3)	43	20	9.0	0.022	21	1.0	28	10.0	46	1.0
(4)	100	27	9.0	0.038	33	1.5	56	5.9	131	1.8
(5)	66	0.2	9.4	0.054	98	2.5	48	13.5	62	0.9
(6)	73	1.0	9.5	0.081	178	2.5	64	24.4	58	0.8
(7)	71	1.0	9.4	0.082	176	2.5	68	16.8	64	0.9

TABLE XII (Con't.)

Composition of thermal water

Hydrothermal System (II)

Location	Temp °C	Natural Flow x 10^{-3} cm^3/sec	pH	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkal-inity mval/l
(1)	100	10	9.3	0.061	98	2.0	60	8.4	150	1.3
(2)	100		8.1	0.052	55	2.5	62	10.4	130	2.2
(3)	55	2	8.6	0.042	52	3.0	48	13.8	74	1.4
(4)	100	68	9.2	0.040	51	2.5	61	9.2	113	1.4
(5)	100	4	9.2	0.042	58	2.5	62	11.8	98	1.1
(6)	90	40	8.5	0.090	204	2.0	66	24.0	83	0.7
(7)	43	1	9.3	0.050	107	1.5	30	11.3	54	1.0
(8)	44	15	9.4	0.093	200	1.2	102	21.6	48	0.9

TABLE XII (Con't.)

Composition of thermal water

Hydrothermal Systems (III) and IV)

Location	Temp °C	Natural Flow x 10^{-3} cm^3/sec	pH	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/l
(1)	100	12	9.0	0.111	142	12.0	114	6.6	225/530	4.7
(2)	91	0.5	8.9	0.048	60	3.0	54	5.2	145	1.7
(3)	100	40	8.9	0.058	54	2.5	54	8.0	165	1.6
(4)	62	0.2	8.9	0.040	53	3.0	50	10.8	126	1.3
(5)	51	1	8.5	0.057	114	0.5	42	17.5	56	1.5

Hydrothermal System (IV)

Location	Temp °C	Natural Flow x 10^{-3} cm^3/sec	pH	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/l
(1)	95	10	8.9	0.042	53	2.5	74	10.5	125	1.2
(2)	100	60	8.6	0.041	25	1.4	53		169/206	1.5

TABLE XII (Con't.)

Composition of thermal water

Hydrothermal System (V)

Location	Temp. °C	Natural Flow x 10^{-3} cm^3/sec	ph	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/l
(1)	87	30	9.7	0.022	16	1.1	21	10.o	91	1.5
(2)	80	100	9.3	0.022	16	0.8	36	8.o	79	1.4

Hydrothermal System (VI)

Location	Temp. °C	Natural Flow x 10^{-3} cm^3/sec	ph	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/l
(1)	75	1	8.1	0.041	32	2.3	53	17.o	122	
(2)	30	1	10.o	0.029	21	1.1	41	12.o	87	
(3)	87	10	9.9	0.027	27	1.o	21	8.o	125	
(4)	91		9.7	0.027	29	1.1	22	7.o	130	

TABLE XII (Con't.)

Composition of thermal water

Hydrothermal System (VIII)

Location	Temp. °C	Natural Flow x 10^{-3} cm^3/sec	pH	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/l
(1)	70		9.2	0.027	24	0.8	37	8.o	55	1.3
(2)	90	7.o	9.3	0.037	32	1.4	52	7.o	100	1.7
(3)	44		9.5	0.021	11	0.9	9	4	102	2.o
(4)	48	1.8	9.6	0.023	11	1.2	12	4	67	1.9
(5)	42	1	9.3	0.023	16	1.o	17	6	62	1.6
(6)	58	1.o	9.6	0.022	12	0.8	10	4	66	1.8
(7)	65	1.o	9.4	0.032	33	1.o	32	5	85	1.5

TABLE XII (Con't.)

Composition of thermal water

Hydrothermal System (VIII)-(Con't.)

Location	Temp. ^oC	Natural Flow x 10^{-3} cm^3/sec	pH	Conductivity at 25^oC $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/l
(8)	63	30	9.6	0.024	19	1.0	14	5	72	1.8
(9)	18	1.4	8.8	0.016	10	0.2	7	7	30	1.3
(10)	34	1.5	9.3	0.015	10	0.4	6	7	44	1.5
(11)	66	11	9.2	0.030	77	0.6	26	8	60	1.2
(12)	65	10	9.5	0.032	50	0.6	27	8	65	1.4
(13)	71	2	8.6	0.056	82	1.0	78	16	72	0.9

TABLE XII (Con't.)

Composition of thermal water

Dike-controlled spring in the Eyjafjordur area

Location	Temp. °C	Natural Flow x 10^{-3} cm^3/sec	pH	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/l
(1)	48	12	9.1	0.016	12	0.2	12	6	61	1.4
(2)	52	4	9.5	0.019	11	0.4	14	6	75	1.6
(3)	52	0.7	9.5	0.030	36	0.7	39	9	81	1.8
(4)	51	1	9.2	0.036	56	0.8	59	9	111	1.7
(5)	25	5	9.3	0.015	8	0.3	12	6	45	1.4
(6)	75	3	9.3	0.028	23	1.2	36	10	110	1.9
(7)	48	3	9.3	0.022	16	0.4	36	6	75	1.4

TABLE XII (Con't.)

Composition of thermal water

Dike-controlled spring in the Eyjafjordur area (Con't.)

Location	Temp. °C	Natural Flow x 10^{-3} cm^3/sec	pH	Conductivity at 25°C $(Ohm\ m)^{-1}$	Cl p.p.m.	F p.p.m.	SO_4 p.p.m.	Hardness p.p.m. CaO	SiO_2 p.p.m.	Alkalinity mval/l
(8)	75	1.5	9.3	0.027	13	0.7	49	11	110	1.6
(9)	58	1.0	9.3	0.024	10	0.7	41	6	86	1.5
(10)	35	0.5	9.2	0.027	18	0.8	49	7	100	1.5
(11)	55	1.5	9.3	0.024	14	0.9	34	6	90	1.5
(12)	67	2	9.3	0.023	13	0.6	36	6	80	1.4
(13)	43	0.5	9.6	0.042	46	2.8	79	12	87	

TABLE XIII

Composition of natural steam

No.	Gases cm^3/gm	CO_2 vol %	H_2S %	H_2 %	CH_4 %	Rest (N_2,A) %	Source
			Natural steam area (I)				
(1)	2.9	95.5	1.8	0.7	0	2.0	spring
			Natural steam area (III)				
(1)	5.7	89.3	5.7	3.6	0.1	1.3	well 15
(2)	7.6	83.9	9.6	5.4	0.1	1.0	well 16
(3)	5.0	81.3	12.5	4.5	0	1.7	well 14
			Natural steam area (IV)				
(1)	0.43	72.3	16.5	4.7	1.1	5.4	well
(2)	0.48	83.3	10.7	1.8	0.1	4.1	well
			Natural steam area (V)				
(1)	3.0	60.8	10.0	27.6	0.2	1.4	spring
(2)	3.7	74.0	8.0	16.4	0.1	1.5	spring
(3)	3.8	71.6	9.7	17.2	0.1	1.4	spring
			Natural steam area (VI)				
(1)	4.2	87.3	4.4	5.1	0.1	3.1	spring western part
(2)	2.6	69.4	8.3	19.5	0.4	2.4	spring eastern part
(3)	0.3	60.9	17.4	18.6	0.6	2.5	spring eastern part
(4)	1.1	63.4	11.2	21.4	0.3	3.7	spring eastern part

TABLE XIII (Con't.)

Composition of natural steam

Natural steam area (X)

No.	Gases cm^3/gm	CO_2 vol %	H_2S %	H_2 %	CH_4 %	Rest (N_2,A) %	Source
(1)	8.3	31.0	16.4	49.7	0.8	2.1	spring eastern part
(2)	7.10	35.3	18.3	44.2	0.9	1.3	spring eastern part
(3)	13.8	37.2	17.2	42.7	0.8	2.1	spring eastern part
(4)	13.4	34.2	17.2	46.2	0.6	1.8	spring eastern part
(5)	14.3	63.0	18.3	14.9	0.3	3.0	spring eastern part
(6)	19.2	52.6	11.3	26.9	0.4	8.8	spring eastern part
(7)	14.3	63.2	21.0	9.3	0.2	6.3	spring eastern part
(8)	8.5	63.9	18.5	14.8	0.4	2.5	spring eastern part
(9)	7.5	50.7	17.5	30.0	0.1	1.7	well eastern part
(10)	6.1	53.0	18.5	26.0	0.1	2.4	well eastern part

Natural steam area (XI)

No.	Gases cm^3/gm	CO_2 vol %	H_2S %	H_2 %	CH_4 %	Rest (N_2,A) %	Source
(1)	6.0	66.9	15.8	15.2		2.1	spring

Natural steam area (XIII)

No.	Gases cm^3/gm	CO_2 vol %	H_2S %	H_2 %	CH_4 %	Rest (N_2,A) %	Source
(1)	5.9	77.5	10.9	6.7	0.1	4.8	spring
(2)	4.0	60.9	14.4	20.6		4.1	spring

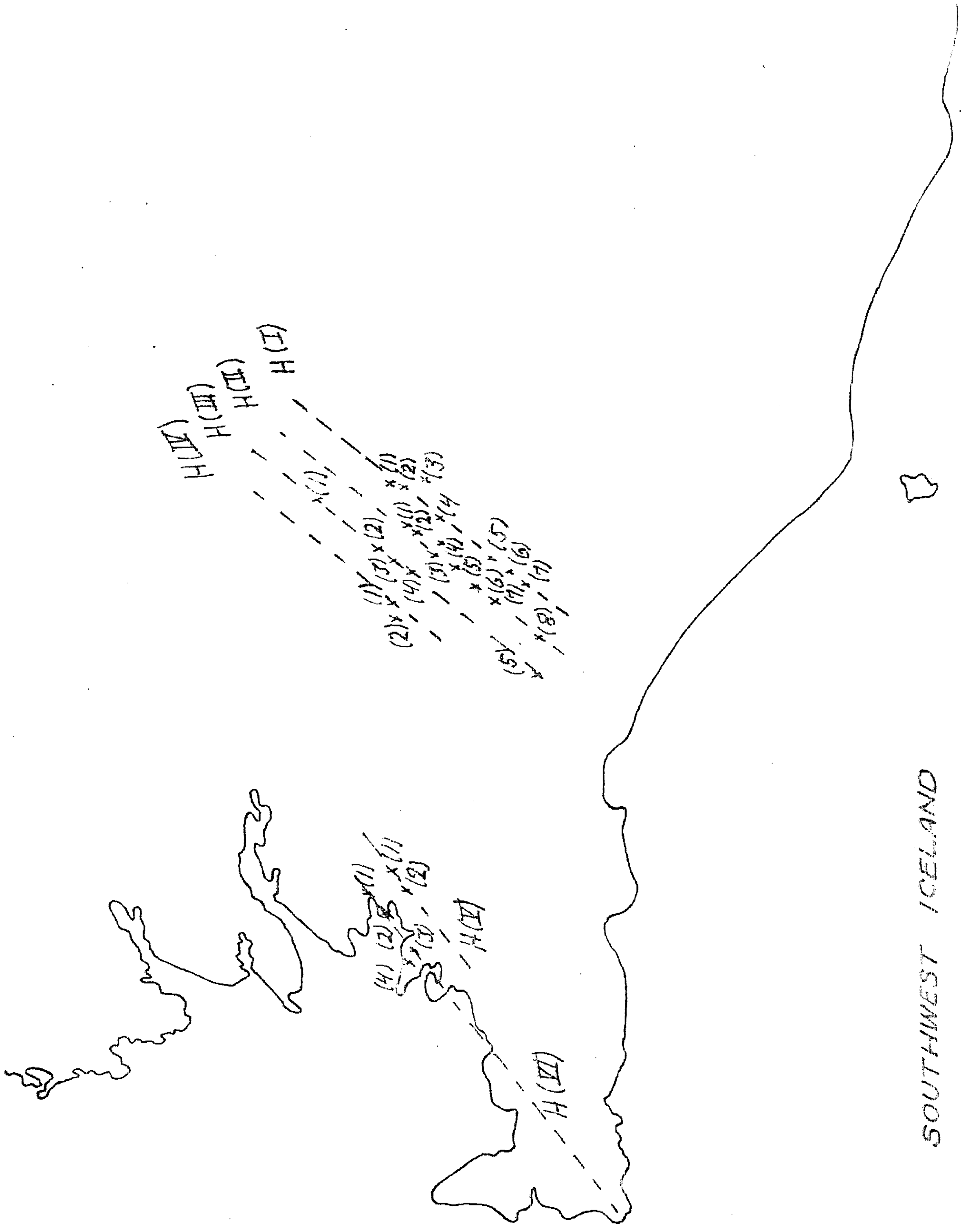


Fig. 17

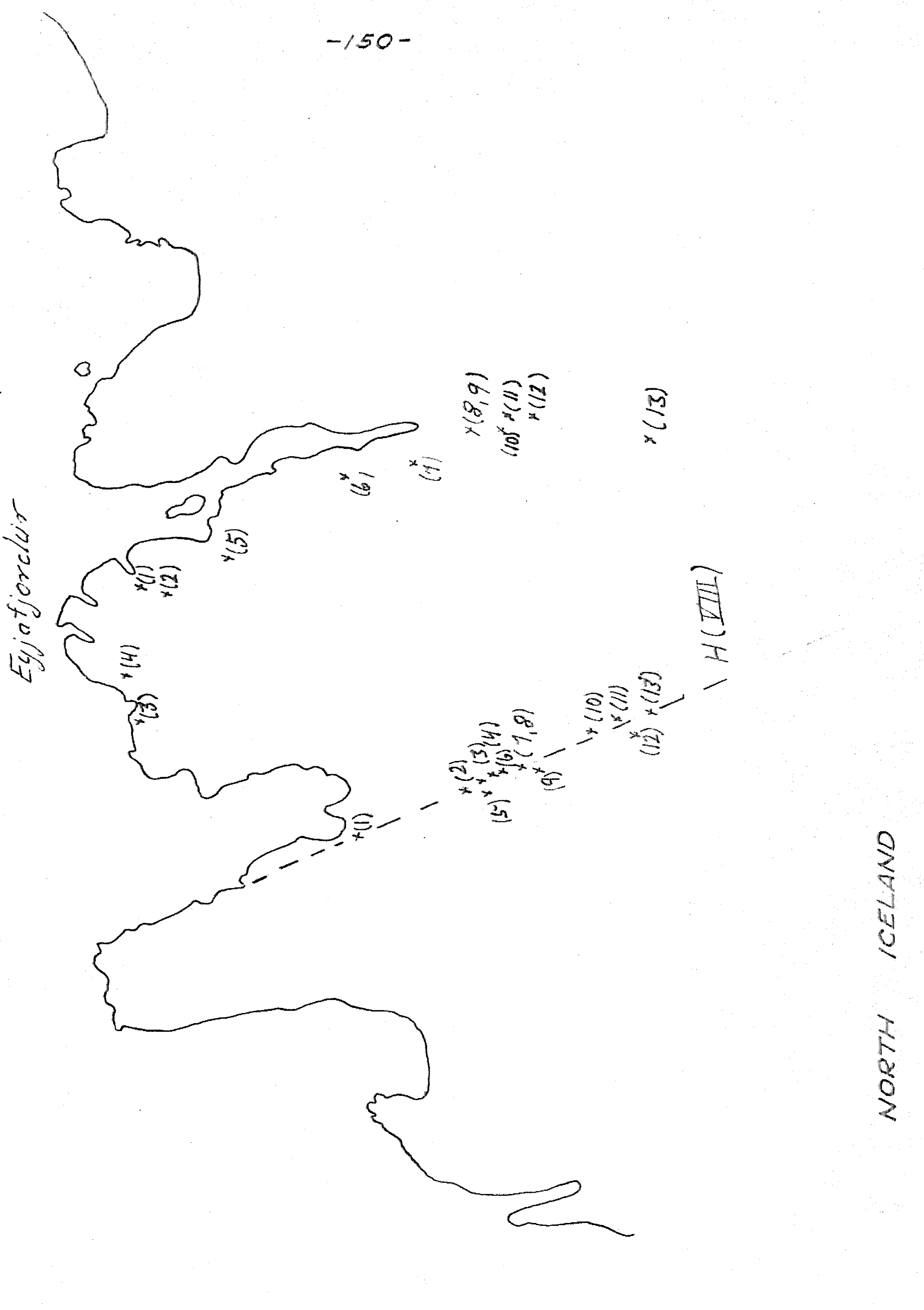


Fig. 18

Legend to Fig. 17 and 18

H(I)
- (1) Skipholt
- (2) Laugar, Rkjd.
- (3) Hrunalaug
- (4) Gröf, Grafarbakki
- (5) Reykir, Skeiðum
- (6) Húsatóptir
- (7) Hlemmiskeið

H(II)
- (1) Reykholt
- (2) Reykjavellir
- (3) Spóastaðir
- (4) Laugarás
- (5) Skálholt
- (6) Sólheimar
- (7) Ormsstaðir
- (8) Laugar, Hraunghr.

H(III)
- (1) Geysir
- (2) Efri-Reyirr
- (3) Syðri-Reykir
- (4) Böðmóðsstaðir
- (5) Laugabakkar

H(IV)
- (1) Útey
- (2) Laugarvatn

H(V)
- (1) Norður-Reykir
- (2) Suður-Reykir

H(VI)
- (1) Kollafjörður
- (2) Álfsnes
- (3) Þvottalaugar
- (4) Rauðará

H(VIII)
- (1) Sauðárkrókur
- (2) Varmahlíð
- (3) Vallalaug
- (4) Saurbær
- (5) Kríthóll
- (6) Vindheimar
- (7) Reykjafoss
- (8) Steinsstaðir
- (9) Hamarsgerði
- (10) Tunguháls
- (11) Bakkakot
- (12) Goðdalir
- (13) Hof

Eyjafjörður
- (1) Ólafsfjörður
- (2) Reykir, Ólafsf.
- (3) Lambnesreykir
- (4) Reykjahóll, Bk.
- (5) Sundl. Svarfd.
- (6) Laugarland, Hgd.
- (7) Glerárgil
- (8) Reykhús
- (9) Kristnes
- (10) Hrafnagil
- (11) Laugarland, Eyjaf.
- (12) Brúnalaug
- (13) Hólsgerði

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