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5. *Studies of the Thermal State of the Earth. The 13th Paper: Terrestrial Heat Flow in Japan.*

(A Summary of the Terrestrial Heat Flow Measurement in Japan up to December, 1962).

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Contents

Abstract

Since 1957, terrestrial heat flow has been determined at 39 places on the Japanese Islands by the Earthquake Research Institute, University of Tokyo. Besides these, 12 sea measurements have been made in the sea around Japan as a joint program of the University of Tokyo and the Japan Meteorological Agency.

Of the total of 39 land stations, 26 are metal mines, 4 coal mines, 2 oil fields, 3 natural gas fields, 1 railway tunnel and 3 sites of other types. In each of these stations heat flux was estimated from the geothermal gradient combined with the thermal conductivity of rocks in which the thermal gradient was measured.

An outline of the method used to measure thermal gradients and thermal conductivity is described. Temperatures were measured in deep or in short bore holes for geothermal gradient assessment. The maximum depth of temperature measurement exceeded several hundred meters at most of the land stations. In the course of the temperature measurement, two types of underground isothermal surfaces were discerned and their origin investigated. Values of the thermal gradient on land varied from 0.5° to 5.0°*C*/100 *m*; except those areas which are considered to be geothermally anomalous. Thermal conductivity was determined by the divided-bar method. The apparatus was calibrated with standard substances of known conductivity. In the course of this study, measurements were made on over 240 sets of rock samples covering 36 rock types. In this paper, the results are presented in a diagram according to their rock types. These values range from 2×10^{-3} to 15×10^{-3} *cal/cm sec °C*.

Errors accompanying the calculation of heat flow values were estimated from the probable variation in the thermal gradient and the degree of ambiguity in assessing the *in situ* thermal conductivity of the strata as based on the thermal conductivities of individual samples. In most cases, the error amounts to a few tenths of 1 $\mu cal/cm^2 sec$ or 20 to 30 per centum of the observed heat flow value.

From the 39 land and 12 sea measurements, general tendency of terrestrial heat flow distribution in and around Japan can be visualized. Characteristic features of the distribution are as follows: i) A region of high heat flow exists on the inner side of the Japanese arc, where the heat flow value exceeds 2 $\mu cal/cm^2 sec$. ii) A region of low heat flow (less than 1 $\mu cal/cm^2 sec$) exists on the Pacific Ocean side of north-eastern Honshu. iii) There may be a high heat flow region in the Pacific far off the coast of north eastern Honshu. iv) In regions other than the above, almost all the heat flow values on land are between 1 and 2 $\mu cal/cm^2 sec$.

Possible distribution of heat source and temperature in the crust, which are required to bring about the heat flow distribution as observed, was examined. The steady state solution of heat conduction suggests that excess heat sources are expected beneath the high heat flow regions and the temperature at the bottom of the crust should be about 800°*C*, while it may be as low as 200°*C* in the low heat flow region. The reliability of these figures depends on the validity of the assumptions adopted for the rate of heat production, thermal conductivity and crustal thickness. The results are possibly related to other geophysical or geological phenomena.

I. Introduction

A. Scope of the present paper

Terrestrial heat flow has been determined in many places over both continental and oceanic areas. Since the earlier works in England [Benfield (1939), Anderson (1939)], heat flow measurement on land has steadily advanced. Birch's review in 1954 inferred that the average heat flow value on land is 1.2 $\mu cal/cm^2 sec$ according to the reliable measurements obtained so far. Measurements in the sea were initiated by Bullard in 1952 [Bullard (1952)]. Since his pioneering work, heat flow measurement in the seas has been promoted intensively by several institutions in USA, Great Britain etc.. Herzen's work (1959) in the south-eastern Pacific revealed a correlation between heat flow and the major topographic features of the ocean bottom. Much attention has been paid to this result and, since then, many measurements have been added to check the suggested hypothesis; including the possibility of thermal convection currents in the mantle.

To date, 90 land and 600 sea measurements of terrestrial heat flow have been accumulated. The average terrestrial heat flow in continental and oceanic areas is now known to be 1.65 $\mu cal/cm^2 sec$ and 1.48 $\mu cal/cm^2 sec$ respectively [Lee (1963)]. On the basis of this accumulated heat flow data, it becomes possible to discuss global distribution of terrestrial heat flow in relation to other geophysical features such as large changes in topography or gravity [Lee and MacDonald (1963)]. Further developments in heat flow research will bring about many interesting and geophysically important results.

In Japan terrestrial heat flow measurements have been made by the Earthquake Research Institute of the University of Tokyo since 1957. The first objective was to determine, by some suitable method, the order of magnitude of terrestrial heat flow in Japan [Uyeda, Yukutake and Tanaoka (1958)]. Secondly, we wished to then determine the general area distribution of heat flow in and around Japan whose unusual anomalies were anticipated from earlier measurements [Horai (1959), Uyeda and Horai (1960)]. Up to December 1962, a total of 51 measurements had been completed of which 39 are land measurements and 12 sea measurements. This data enables us to discuss the general tendency of heat flow distribution and possible geophysical interpretations.

In the present paper, a summary of the results of these measurements (IV) follows a description of the methods of measurement (II, III). Then, some discussion follows concerning the possible geophysical and geological significance of the heat flow pattern obtained by our survey (V). Detailed results of the measurements at each of the individual heat flow stations have been presented previously [Uyeda and Horai (1963 a, b, c), Horai (1963 a, b, c)].

*B. Brief account of the geology of Japan**

Geotectonically the Japanese Islands are divided into several provinces as shown in Fig. 1. A brief summary of the geological history of these provinces is as follows.

The area occupied by the Japanese Islands at present is considered to have been a marginal part of the Asiatic continent during pre-Silurian time. The Honshu marginal geosyncline which developed in this region during the subsequent period until late Paleozoic time, then experienced a period of extremely intense regional orogeny which formed the framework of the Japanese Islands. Dissected metamorphic zones now exposed in south-western Japan are the products of this orogeny. Region I (Hida and Sangun metamorphic zones) which lies on the north side of south-western Japan is older than regions II and III (Ryoke and Sambagawa metamorphic zones) which lie to the south. Region IV (Kitakami and Abukuma land masses) is considered to be an extension or a marginal phase of the Ryoke zone.

In central Hokkaido, the Hidaka orogeny (Region V) started later

* The author owes the original idea of this section to Dr. A. Sugimura, Geological Institute, Faculty of Science, University of Tokyo. Further consultations were made with the following papers: H. Kuno, Outline of the geology of Japan, *unpublished manuscript* (1955); M. Minato, K. Yagi and M. Funahashi, Geotectonic synthesis of the Green Tuff regions in Japan, *Bull. Earthq. Res. Inst.*, **34**, 237-264 (1956); A. Miyashiro, Evolution of metamorphic belts, *J. Petrology*, **2**, 277-311 (1961); A. Sugimura, T. Matsuda, K. Chinzei and K. Nakamura, Quantitative distribution of late Cenozoic volcanic materials in Japan, *Bull. Volc.*, *XXVI*, 125-140 (1963).

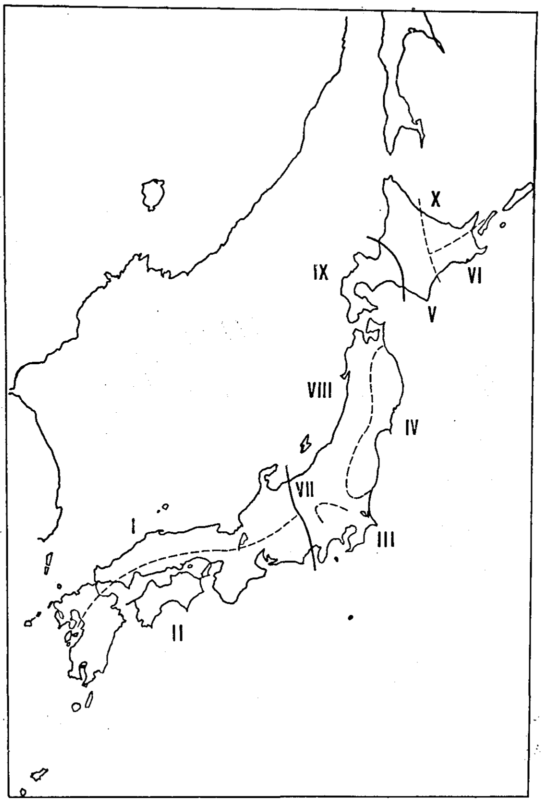

Fig. 1. Tectonic provinces of Japan.

than the Honshu orogeny and completed almost all of its formation as a mountain range by early Neogene time. In this region, Hidaka and Kamuikotan metamorphic zones are now exposed. Region VI (Southeastern part of Hokkaido) is the marginal phase of region V.

In regions VII (*Fossa Magna*) and VIII (Uetsu district), a younger intense disturbance (Mizuho orogeny) commenced from Neogene time. The violent volcanism which was associated with this orogeny is generally called "green tuff" volcanism. Region IX (Southwestern Hokkaido) is an extension of this orogenic belt in Hokkaido. Region X (North-eastern Hokkaido) was also characterized by the Neogene orogenic activity similar to that described in regions VII, VIII and IX.

II. Description of the measurement

To measure terrestrial heat flow coming up through the crust to the surface of the earth, it is necessary to know two separate quantities; the geothermal gradient and the thermal conductivities of rocks in the strata where the geothermal gradient is to be measured.

In this and subsequent sections, a brief survey of methods will be given.

A. Measurement of geothermal gradient

Subsurface temperatures had been measured in 8 localities in Japan including bore-holes, oil wells, a railway tunnel and a mine to obtain thermal gradient data in our earlier heat flow studies prior to June 1960. The methods and the results of these 8 measurements were described in the previous papers [Uyeda, Yukutake and Tanaoka (1958), Horai (1959), Uyeda and Horai (1960)].

Since July 1960, heat flow has been measured at 31 localities in Japan utilizing bore-holes and mines. Results of these measurements at individual localities have been given in previous papers [Uyeda and Horai (1963, a, b), Horai (1963, a, b)].

i) Temperature measurements in bore holes.

Two sets of thermistor thermometers with a depth capacity $500\,m$ and $1200\,m$ were used to measure the temperature distribution in bore holes. Cables used in the study were two-ply armored wires, each having $6.6\,gr/m$ of linear density and $2\,mm$ in diameter. A thermistor sensor contained in a thin metal tube was fixed at the end of the cable. The sensors were calibrated with a standard mercury thermometer. The probe was lowered by hand-drive winch down the hole and temperatures were logged, with an accuracy of $0.1°C$, at 10, 25 or $50\,m$ intervals. Usually, 1 minute was sufficient for the thermistor sensors to attain temperature equilibrium with the ambient when they were in water, but 5 minutes or more was often necessary when measurements were made in the air column of a hole.

Bullard (1947) has presented a calculation regarding the effect of drilling on the temperature distribution along a bore-hole. He concluded that a fairly long period must elapse before a bore-hole will regain an undisturbed temperature gradient. For example, on suitable assumptions regarding the size of bore and the thermal constants of surrounding rocks, a year or so is necessary for a bore-hole drilled in one month to restore its original temperature (at the top) within 1 percent of errors.

But, recently, J. C. Jaeger (1961) indicated that, when a method such as diamond drilling was used, a day of quiescence after drilling is sufficient to obtain an approximate geothermal gradient within an error of 5 percent.

The discrepancy between the views of Bullard and Jaeger lies, as Jaeger himself stated, in the effect of the drilling system. Bullard based his calculations on up and down circulation of water at high velocities along the whole length of a hole as in rotary drilling; while Jaeger observed temperature variations during drilling which was accompanied by a small amount of circulating water.

During our heat flow study in Japan, there were cases in which Bullard's assumptions and results were confirmed [Uyeda and Horai (1960)], and an effort has been made to select bore-holes as old as possible to measure reliable temperature distributions. Actually, temperature measurements were made in several holes while drilling was underway and there were some instances in which temperature logs could be better explained by Jaeger's model. If the disturbance Bullard discussed had really occurred, the discrepancy between disturbed (i. e. observed) and undisturbed temperatures must be largest near the mouth of the hole. On the other hand, mean surface temperatures extrapolated from undisturbed subsurface temperatures in an equilibrium state are said to be approximately equal to the mean surface air temperature. So, a comparison of extrapolated surface temperature with the mean air temperature should afford a means of checking whether measured temperatures are really in equilibrium.

An example of this is the temperature record logged at Ashibetsu, Hokkaido. There, temperatures were measured in the bore-hole of Ashibetsu No. 5, a day and a half after drilling. The result shows, as indicated in the previous paper [Fig. 12 in Horai (1963)], a curve describing a linear temperature-depth relationship without any appreciable kink or curvature. The average thermal gradient and the extrapolated surface temperature were $3.08°C/100\,m$ and $8.9°C$ respectively. The mean air temperature in Ashibetsu for comparison with the extrapolated surface temperature was not available. But at Akabira, about 18.5 km to the west of Ashibetsu, the extrapolated surface temperature obtained from the bore-hole and the drifts of the mines, becomes $7.8°C$. This figure is derived from temperature data representing equilibrium conditions, and it approximates the mean air temperature in the area concerned. The coincidence between the two extrapolated surface temperatures, therefore, seems to prove that the temperature gradient, in Ashibetsu No. 5, did not suffer any large disturbance from drilling.

Another example was obtained in a bore-hole at Yanahara, Okayama Prefecture. As described in the previous paper [pp. 116–118 in Uyeda and Horai (1963 b)], temperature logs were made in No. 247 bore-hole down to the depth of 950 m after one day of quiescence. The temperature-depth relationship from a depth of 250 m to 750 m (i. e. in the slate layer) gives a thermal gradient of $1.92°C/100\,m$, and the extrapolated surface temperature as $14.4°C$. This value again cannot be compared to a suitable standard. The extrapolated surface temperature obtained in No. 248 bore-hole near No. 247 becomes $12.8°C$. No. 248 was drilled on the north flank of a hill, while No. 247 was situated in the swamp-like area at the foot of the hill. So, the discrepancy between the two values seems to be rather natural. The mean annual air temperature observed at Okayama, 35 km south of Yanahara is $14.6°C$; at Tottori, 62 km north of Yanahara, $13.9°C$. This suggests that the mean annual temperature at Yanahara will be similar. If so, it may be said that the temperature distribution in bore-hole No. 247 which gave the extrapolated surface temperature $14.4°C$ had not suffered any large amount of disturbance, hence the thermal gradient derived from it should also be reliable.

In Table I, some extrapolated surface temperatures obtained from

Table I. Comparison of extrapolated surface temperature and mean annual air temperature observed at the nearest weather station.

Extrapolated surface temperature			Mean annual air temperature observed at the nearest weather station	
Tokyo Univ.	(TU)	$15.6°C$	$14.3°C$	Tokyo W. S.
Mobara	(MB)	$13.7°C$	$14.8°C$	Choshi W. S.
Haboro	(HB)	$6.7°C$	$6.9°C$	Haboro W. S.
Hidaka	(HD)	$18.2°C$	$15.3°C$	Wakayama W. S.

the undisturbed bore-holes are compared with the mean annual air temperature observed at the nearest weather station.

ii) Temperature measurement in drifts of mines.

The rate of increase of subsurface temperatures with depth can be assessed from the temperatures measured in mineworkings at increasing depths from the surface. In order to obtain undisturbed rock temperatures, a thermometer must be inserted some distance into a thin hole drilled into the wall of the drift.

A. D. Misener first pointed out, working from temperature loggings along the holes, that if the measurements were taken 70 feet ($\simeq 21\,m$) inside the wall no effects due to ventilation in the drifts were apparent [Misener (1949), Misener, Thompson and Uffen (1951)].

The same experiment was conducted by the author in the Hitachi mines [Horai (1959)] and it has been repeated in other places several times since. It seems that in most cases more than $10\,m$ of insertion into a narrow (usually $3\,cm$ or more in diameter) hole, is sufficient to avoid artificial thermal effects.

iii) Two types of isothermal surfaces.

Temperature measurements in the drifts of the mines revealed two types of isotherm patterns.

In some mines, temperature variations were more coherent in a diagram when the altitude rather than the depth of the temperature stations was selected as the ordinate. This indicates that the subsurface isothermal surfaces may be considered to lie as horizontal planes regardless of the surface topography. We shall tentatively call this type of isothermal pattern "L type".

In other mines, temperatures could be better arranged adopting the depth as the ordinate. In this case, the isothermal surfaces appeared to be parallel to the surface topography (D type).

The difference of isotherm patten is probably related to the erosional history of the topography. Consider the progress of erosion as illustrated in Fig. 2. At the first stage, isotherms are considered to be running horizontally under a flat plain which had not yet suffered the action of erosion. At the second stage, the action of erosion has started and reached a maximum. The topography has changed its morphology rapidly and the isotherms were unable to modify their shapes to the new equilibrium positions (L type). In the last stage, erosion has subsided and the surface topography remains nearly unchanged. Then, the isotherms would gradually regain their equilibrium positions and ultimately attain them (D type).

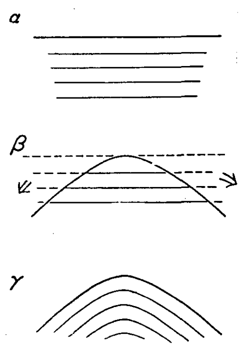

Fig. 2. Variation of isotherm type.

In Japan, most topographical features are said to have been formed during the latest glacial ages in Pleistocene time. Therefore, the difference in isotherm patterns is not attributed to the difference in the time which had actually elapsed since the action of erosion started, but to the difference in the time necessary for the isothermal surfaces to recover their equilibrium positions. The time necessary for recovery will depend on the magnitude of the topographical features and the thermal properties of the ground. Rough evaluations of the time necessary for recovery in Osarizawa (L type) and Ikuno (D type) seem to support the above hypothesis. However, an exhaustive study will be necessary to confirm this theory.

iv) Disturbances of the thermal gradient.

Disturbance of subsurface temperature distribution may be induced from various causes. One cause is the circulation of underground water.

Several examples will be given. In Tanna and Oshima [Uyeda and Horai (1963 a)], temperatures were measured in the vertical bore-holes and no thermal gradient was observed because temperature remained constant everywhere in the water columns of the bore-holes. The gradient observed in these places was interpreted to be caused by the water which was moving through the permeable medium, for example, coarse materials such as gravel in the Tanna Basin, or volcanic ash or porous lava flows as in Oshima.

In the Seikoshi mine, temperatures were measured in the drifts extended to various depths, and an increase of temperature with depth was observed as in many other mines except the deepest one. In the deepest drifts, however, the temperatures measured in several places on the same level were not exceptionally low, completely contradicting the tendency obtained from the measurements in the upper drifts as shown in Fig. 3. One explanation is water flowage in fissures or faults induced by drainage in the mines or the natural flow of underground water. In general, water flow in fissures or fault zones results in a complicated irregular temperature field such as observed in Kushikino [Horai (1963 b)] rather than temperature distribution with a null gradient as observed in Oshima and Tanna [Uyeda and Horai (1963 a)].

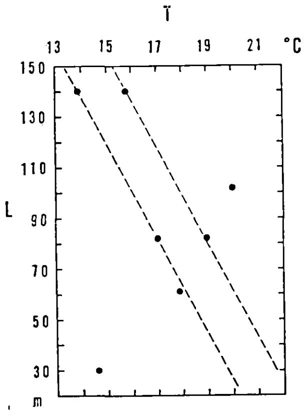

Fig. 3. Temperature-depth relation at Seikoshi, Shizuoka Prefecture.

It is known that changes in the surface environment affect subsurface temperature distributions. The effect of factors, such as climatic change and upheaval or subsidence of the ground in geologically recent years on the subsurface temperature distribution has been investigated in detail by some workers. It seems advisable to check the possible ammount of disturbance in order to see whether substantial correction would be necessary for the observed gradient.

Birch (1948) calculated the effect of climatic changes based on an investigation of climatic fluctuations during the Pleistocene. His conclusion was that "the climatic correction to the geothermal gradient may never exceed $3°C/km$, with a still smaller maximum correction more probable".

As for the effect of subsidence (sedimentation) and upheaval (denudation), Benfield (1949) carried out a model calculation and presented the result in tables. In order to apply his result, certain quantities such as the rate of subsidence or upheaval, its duration and the change of the surface temperature, are necessary. For example, a $3\times10^{-2}\,cm/yr$ rate of subsidence, lasting 10^7 *yrs*, would induce a $3.7°C/km$ variation in the $20°C/km$ of thermal gradient, if the sedimentation took place at the same rate as subsidence with constant temperatures. This would hold in such a region as the geosynclinal basins which lasted from Miocene to Pliocene on north-western Honshu [Matsuda (1963), private communication], But if upheaval of the same magnitude followed a period of subsidence as was the case in the abovementioned area, a compensating effect with regard to the geothermal gradient will occur; and it can be expected that these variations in the thermal gradient may tend to cancel one another.

In the present study, corrections for the effects of changes in topography were not attempted. This work must be attempted in the future.

v) *Summary of the result.*

In Fig. 4 and Fig. 5, 39 temperature-depth (or elevation) relationships are presented. The results of measurements in the bore-holes or in the mines which show the D-type isothermal patterns are shown in Fig. 4, while those in the mines which show the L-type isothermal patterns are shown in Fig. 5. The general tendency of the subsurface temperature distribution and the average geothermal gradient in each of these localities can be estimated by the inscribed line on the diagrams. Individual values of the thermal gradient are listed in Table IV (page 116).

A histogram of the 39 thermal gradient values, excluding 2 which

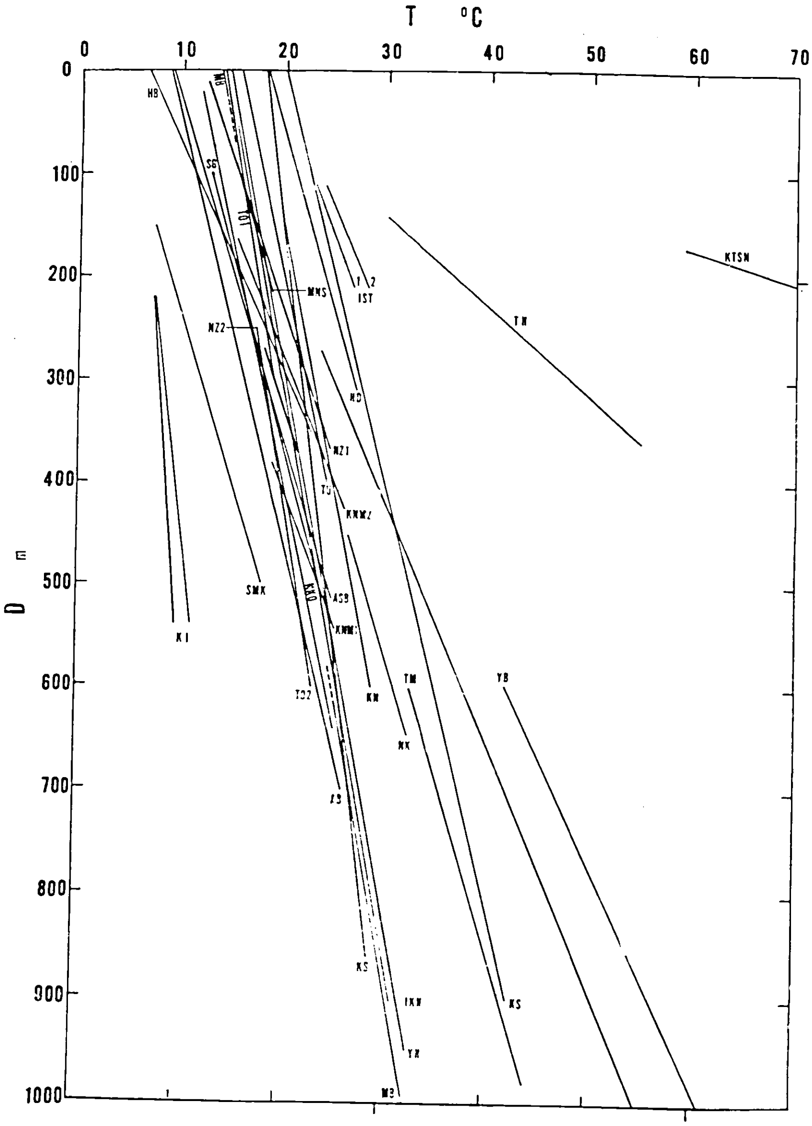

Fig. 4. Temperature-depth relations observed in Japan. Letters attached to each line are abbreviations of heat flow stations. (See Table IV)

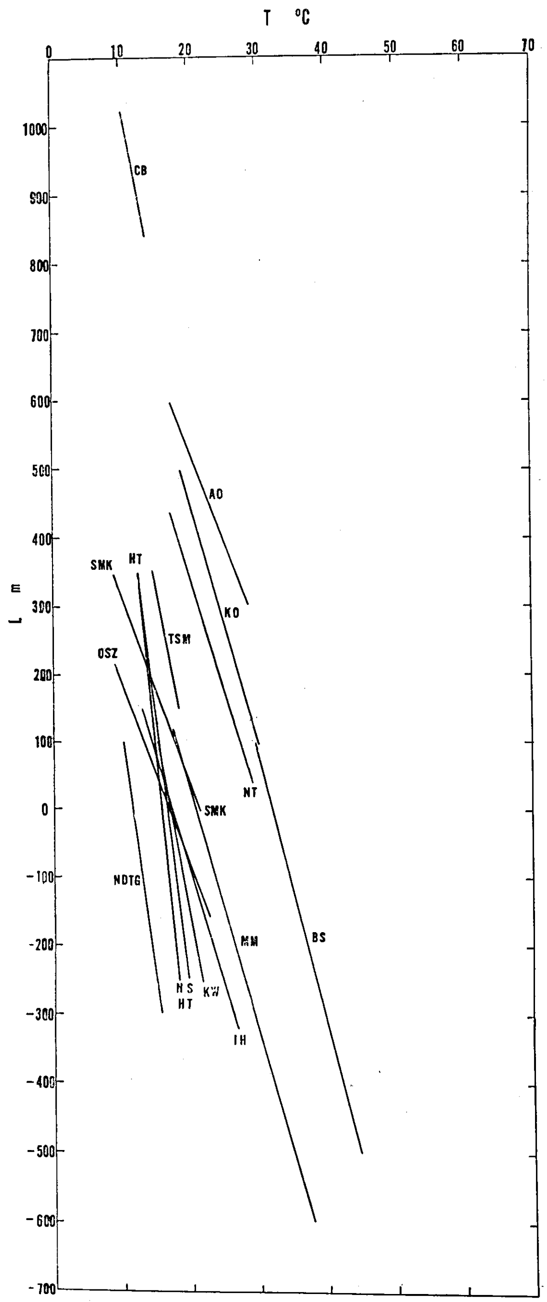

Fig. 5. Temperature-altitude relations observed in Japan. Letters attached to each line are abbreviations of heat flow stations. (See Table IV).

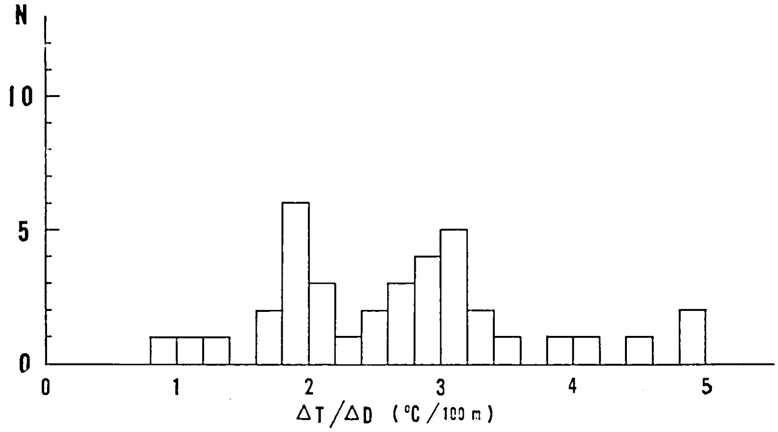

Fig. 6. Histogram of thermal gradient.

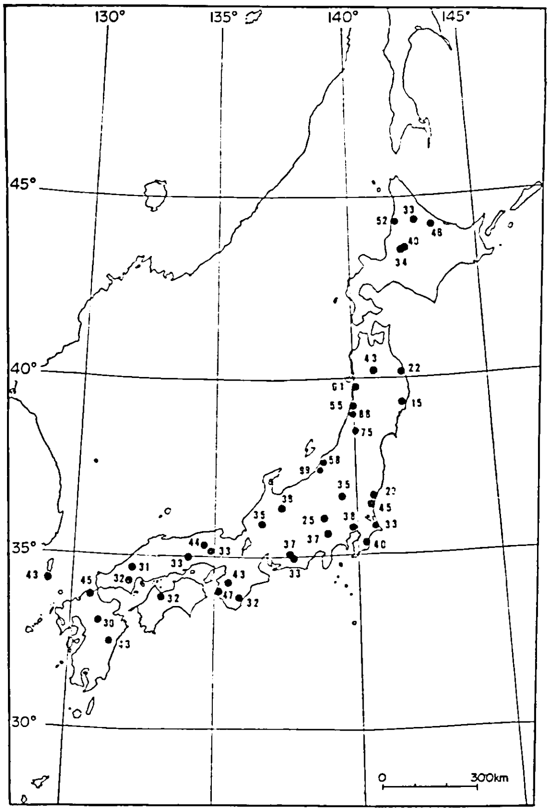

Fig. 7. Distribution of temperature at 1 *km* depth (°*C*).

are considered to belong to geothermal areas, is shown in Fig. 6. As seen in the diagram, the values of the thermal gradient in geothermally normal areas in Japan range from $0.5°C/100\,m$ to $5°C/100\,m$. They show no systematic areal distribution as seen in the heat flow values.

Extrapolated temperatures at a depth of $1000\,m$ from the surface can be calculated from the recorded thermal gradients and temperature distributions measured in most cases down to depths of several hundred meters. They are shown in Fig. 7 for reference. At most of the sites, these temperatures range between $30°C$ and $40°C$. Exceptions include the Niigata oil field, where the temperature exceeds $70°C$ [Uyeda, Yukutake and Tanaoka (1958)] and the area along the east coast of Tohoku district where the temperatures are as low as $10°C$.

B. Measurement of thermal conductivity

i) Rock samples.

Rock samples were collected from the localities where the thermal gradient was measured. Major rock types prevailing in the areas concerned were selected by consulting the geological map or well and bore-hole logs.

Usually core samples were used. When they were not available, samples were cut from the walls of the mineworkings or from surface outcrops.

Rock samples were shaped into disks with a diamond drill and saw and their faces were ground parallel with carborundom on an iron plate for the thermal conductivity measurement.

Assessment of composite thermal conductivity, which the underground layers display as a heterogeneous mass, from the thermal conductivity of individual rock samples is a fairly difficult problem. For simple structure such as that composed of horizontally layered beds, composite thermal conductivity can be calculated, for example, as a serial combination of the individual thermal conductivities. But no better way was found than applying the average thermal conductivity of all the samples to the composite one when the underground structures were more complicated due to intrusion, folding, fault, etc..

ii) Method of Measurement.

A divided-bar apparatus similar to Benfield's (1939) was constructed to measure thermal conductivities of rock samples. Later, another apparatus similar to A. Beck's style (1957) was used to reduce the time necessary for the measurement. The principle of measurement used by these instruments has been discussed in detail by Benfield (1939), Beck

(1957), A. E. Beck and J. M. Beck (1958), and is well known.

A disk-shaped rock specimen is inserted between two vertically held brass bars, and heat is transmitted by conduction from the upper to the lower bar through the specimen. In our apparatus a constant temperature, say $40°C$, is maintained at the upper end of the top bar by an electric heater with a variation of less than $0.1°C$, and the base of the lower bar is held at a constant temperature (say $10°C$) by a water bath with no appreciable temporal fluctuation.

A set of specimens, each with a diameter of 1 inch (the same as that of the brass bars) of increasing thicknesses (usually $1\,mm$, $4\,mm$, $7\,mm$ and $10\,mm$) are used to determine the thermal conductivity of each rock.

In the thermally steady state, the thermal gradient in the brass bars and the temperature difference between the faces of the specimen are measured. Then thermal conductivity is derived from the equation:

$$T/P_B = M + \frac{K_B}{K_R} \cdot \delta\,, \tag{1}$$

where

P_B: thermal gradient in the brass bars,
T : temperature difference between the faces of the rock specimen,
K_B: thermal conductivity of brass,
K_R: thermal conductivity of rock specimen,
δ : thickness of rock specimen,
M : thermal resistance at the contact interfaces between the bars and the sample.

(The temperature difference occurring at the interfaces between the specimen and the bars (described as $M \cdot P_B$) is included in T).

By plotting the measured quantity T/P_B against δ, one can get K_B/K_R as a slope, whence K_R can be found if K_B is known. More than 200 sets of specimens were measured in this manner during the course of the present study.

Specimens are smeared with vaseline on their contact faces, and pressure of $30\,kg/cm^2$ is applied during measurement on the upper bar to ensure stable thermal contact across the specimen-brass bar interfaces.

Even under these conditions, thermal resistance M shows a variance as illustrated in Fig. 8, averaging about $15\,mm$ with a standard deviation of $8\,mm$. In this $M = 15\,mm$, about $9\,mm$ is the sum of the distance between the thermojunctions hidden in the bars and the end-faces of the bars. So about $6\,mm$ or so of the resistance really corresponds to the interfacial temperature drops induced in the vaseline film. The temperature drop amounts to $0.4°C$ when P_B is $2°C/30\,mm$.

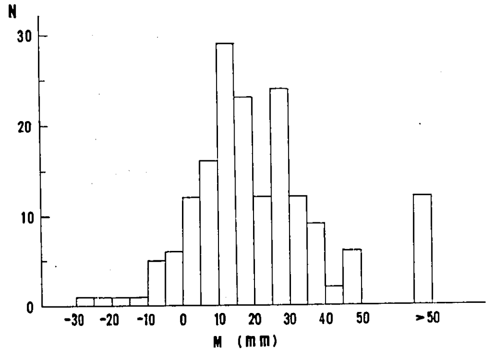

Fig. 8. Histogram of M.

The origin of the scattering of the resistance was discussed by A. E. Beck and J. M. Beck (1958). They applied the average thermal resistance to estimate the amount of correction associated with the determination of the thermal conductivity when only a single specimen was used. In our study, there were several instances, from more than 200 series, when we had to apply this method. This was the case for the friable rocks such as sandstone, shale and schist.

iii) *Calibration of the apparatus.*

To determine K_B, specimens of standard snbstances, i. e., fused silica, crystalline quartz, dense flint glass and borosilicate crown glass with known thermal conductivities, were used. On all of these substances, A. E. Ratcliff of the National Physical Laboratory of England had determined the thermal conductivities as listed in Table II. Fused silica and crystalline quartz samples were purchased from the Thermal Syndicate Co., England, with the assurance that they were the same

Table II. Thermal conductivity of standard substances for k_B calibration.

Standard substance	Density (*gm/cc*)	Thermal conductivity (k in *cal/cm sec °C*, t in *°C*)	Reference
Fused Silica	2.18	$10k = 31600 + 46t - 0.16t^2$ $(-150 < t < 50)$	Ratcliff (1958)
Crystalline Quartz	2.64	$1/k = 60.7 + 0.242t$ $(0 < t < 100)$	Ratcliff (1958)
Dense Flint Glass	3.54	$k = 0.00182$ at $t = 20$ $k = 0.00185$ at $t = 30$	Ratcliff (1959, private communication)
Borosilicate Crown Glass	2.45	$k = 0.00255$ at $t = 20$ $k = 0.00259$ at $t = 30$	Ratcliff (1959, private communication)

materials as Ratcliff's. Other specimens, borosilicate crown glass and dense flint glass, were donated by Dr. Ratcliff for our geothermal studies in Japan.

Procedures for thermal conductivity measurement were repeated at various temperatures on each of these standard samples, and the empirical formula for K_B as a function of temperature was determined as follows:

$$10^3 K_B = 146.4 + 2.24\,t - 0.0166\,t^2, \qquad (2)$$
$$K_B \text{ in } cal/cm\ sec\ ^\circ C,$$
$$t \text{ in } ^\circ C.$$

Every value of K_B determined from any of these standard substances falls within the 5 per cent range of the value calculated by this formula. So we are justified in concluding that the accuracy of the divided bar method used is about the same.

iv) Some remarks on the measurement and the result.

Here, some attention should be paid to the difference of the thermal conductivity of rocks in situ and that measured in the laboratory.

We will first consider the effect of temperature. As has already been discussed in the previous chapter the in situ rock temperature lies, in most of the locations, on the 30°C or 40°C level and seldom exceeds 50°C in geothermally normal areas. The temperature of the rock specimens was maintained at about 25 to 35°C during ordinary measurements.

According to the work of Birch and Clark (1940), ultrabasic rocks such as dunite and hypersthenite are most temperature sensitive in their thermal conductivity, and the rate of decrease in thermal conductivity with the temperature amounts to 10 % per 50°C in the range between 0°C and 100°C. According to this rate, 10°C of temperature variation will cause 2 % of the variation in thermal conductivity.

For many other rocks common in the earth's crust, the variation of thermal conductivity with temperature is not so remarkable as ultrabasic rocks, so in view of the fact that the accuracy of the measurement itself is 5 %, any corrections regarding the temperature would be meaningless.

As for the effect of pressure, Bridgeman (1949) stated in his book on high pressure physics that "in general the change of thermal conductivity with pressure does not appear to be so large" which statement was based on his measurement of thermal conductivity of several rocks and minerals such as basalt, limestone, talc, pipestone etc.. His results show that for the harder rocks such as basalt and limestone, the change of thermal conductivity per 1000 kg/cm^2 is 0.5 % or less, while for the softer substances such as talc and pipestone, it is greater as much as 1.57 and 3.0 %.

If we may assume that the pressure dependence of thermal conductivity of our rocks does not differ much from those obtained by Bridgeman, it can be safely concluded that the effect of the pressure is negligible, as the pressure difference we are concerned with (30 kg/cm^2 in the laboratory and less than 300 kg/cm^2 in the natural state) is far less than that of Bridgman's experiment.

As it is known that the water content of porous sedimentary rocks has a considerable effect on their thermal conductivity, care must be taken to maintain the state of the specimen as close as possible to the natural state during the measurement. To restore the specimen to its natural state an appropriate amount of water must be added when specimens of partially dried porous rocks are to be measured.

Since Bullard and Niblett's work (1951), the effect of the water content of porous sedimentary rocks on their thermal conductivity has been noticed and A. E. Beck and J. M. Beck (1958) discussed the problem in detail. We also investigated the relationship between the thermal conductivity of sedimentary rock and the water content [Horai and Uyeda, (1960)]. The point of the argument was that porous sedimentary rock has its maximum thermal conductivity value when it is saturated with water, and this maximum would not differ much from the in situ value. As for low porosity sedimentary rocks, the water saturated state was assumed by soaking the specimen in water, while the soft sedimentary rocks were treated by the method described in the previous paper [Horai and Uyeda, (1960)].

v) Summary of the result.

During the course of our heat flow study, thermal conductivity was measured on 242 rock samples including several different rock types. Individual thermal conductivity values were tabulated together with the description of each heat flow station [Uyeda and Horai (1963 a, b) and Horai (1963 a, b, c)]. In Fig. 9, the resultant measurements are presented in a diagram in accordance to rock type.

No systematic investigations on the relationship between thermal

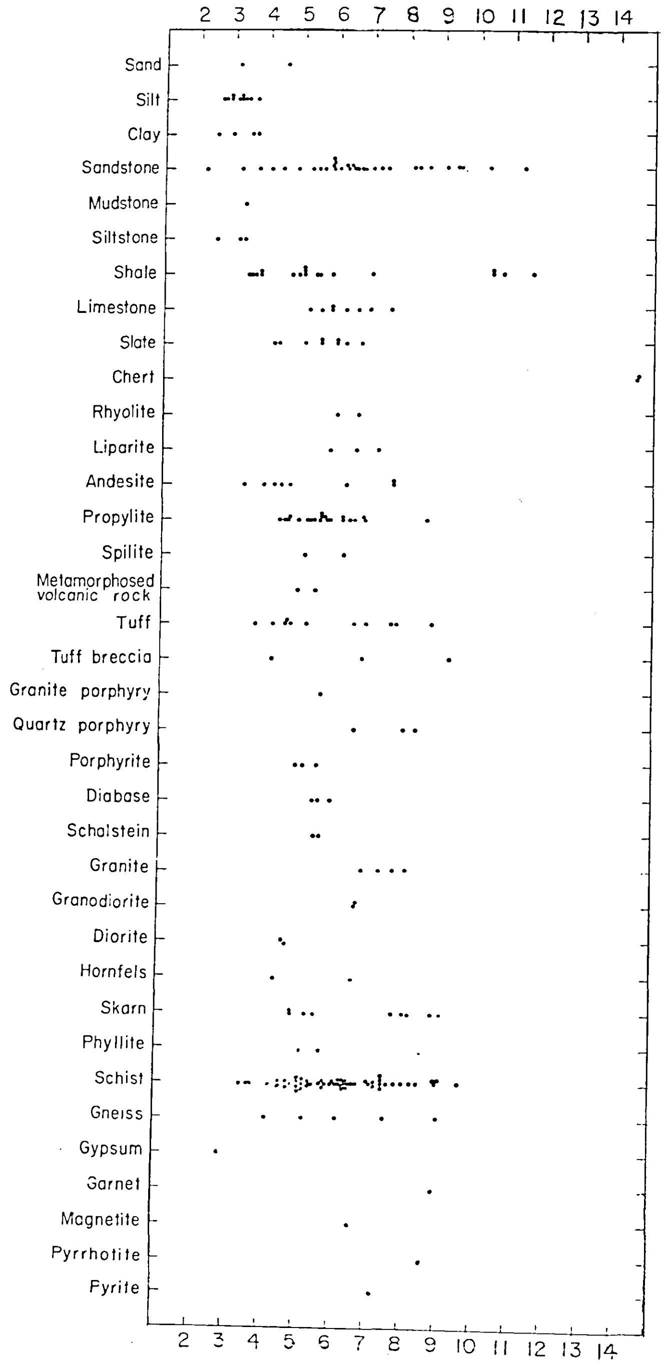

Fig. 9. Thermal conductivity of various rocks. Unit: 10^{-3} cal/cm sec °C.

conductivity and the rock type, based on the mineralogy or fabric of the rocks, were attempted. Only a few remarks from the practical point of view will be mentioned.

1) The number of rock types listed in Fig. 9 amounts to 36. Of these, sedimentary and metamorphic rocks are dominant and only a few volcanic rocks are included. This fact reflects in part the characteristic geology of Japan, and partly the location of thermal investigation sites.

2) Rocks such as shale, sandstone or schist, in which the classification is in part influenced by the mode of genesis (and not by mineralogical or chemical composition) show a large range of conductivity value, including most of the values that ordinary rocks provide.

This fact indicates that one cannot foresee conductivity values, when given only the names of such rocks.

3) In rocks, such as granite, granodiorite, or limestone, the definitions impose special mineralogical, structural or chemical conditions and the values of conductivity are, as a matter of course, much more restricted. In this case one can correlate to some extent the rock type with representative thermal conductivity values.

vi) Thermal conductivity of ocean sediments.

To obtain the values of heat flow from the ocean floor, as well as on the land, it is necessary to measure thermal conductivities of the ocean sediment, instead of hard rocks. Sediment samples were collected with the aid of a piston or gravity corer from the ocean bottom where thermal gradient measurement were made.

As a change of water content in the sediment strongly affects the thermal conductivity, sediments recovered on board ship were sealed in plastic barrels and stored for later laboratory measurement.

In our earlier conductivity measurements of deep sea sediments, the divided-bar method was used. Sediment samples were placed in a plastic ring and inserted between the bars instead of the rock samples. As the thermal contact between the bars and the specimen is perfect owing to the moisture of the sediment specimen, the process of eliminating the interfacial temperature drop was unnecessary.

Since the summer of 1962, the needle-probe method has been used to measure the thermal conductivities of the sediments. Our apparatus was constructed by Dr. S. Uyeda during his stay in the Scripps Institution of Oceanography. The apparatus is quite similar to those developed by R. P. von Herzen and A. E. Maxwell (1959).

Table III. Comparison of thermal conductivity of ocean sediment measured by divided-bar method and needle-probe method. (Yasui, Hôrai, Uyeda and Akamatsu (1962)).

Site and date of collection	Distance from the top of the core (*cm*)	Water content (%)	Thermal conductivity (10^{-3} *cal/cm sec °C*)	Method
39°30′N, 143°28′E 22 August 1962	80— 85	46.6	2.16	D. B.
	85— 95		2.20	N. P.
	100—105	47.3	2.15	D. B.
	105—115		2.18	N. P.
	120—125	49.7	2.06	D. B.
	125—135		2.09	N. P.
	140—145	45.7	2.22	D. B.
39°22′N, 150°03′E 16 August 1962	10— 15	66.2	1.87	D. B.
	15— 20		1.80	N. P.
	25— 30	66.3	1.92	D. B.
	35— 40		1.83	N. P.
	40— 45	63.7	2.07	D. B.
	55— 60	63.6	2.07	D. B.
	60— 70		1.85	N. P.

No direct calibration of the needle-probe method was made. Instead, a series of thermal conductivity measurements were made on the same sediment samples, collected in August 22, 1962 at 39°30′N, 143°28′E and in August 16, 1962 at 39°22′N, 150°05′E by both the needle-probe method and the divided-bar method and the results compared as shown in Table III. They are in fair agreement proving the consistency between the two methods.

C. Probable errors accompanying the heat flow determination

As the terrestrial heat flow is obtained from two separate quantities, i. e., thermal gradient and thermal conductivity, the probable error accompanying the determination of heat flow arises from the ambiguities of these two quantities. As stated in the previous report [Uyeda and Horai (1963 a)], they are assessed respectively as the standard deviation in the least squares method applied to the temperature-depth relation; and as half the range of the thermal conductivies of rocks which compose the succession.

The results are contained in the columns of Table IV. As seen from the table, errors are, in most cases, of the order of a few tenths of 1 $\mu cal/cm^2$ *sec*, amounting to 20 to 30 % of the observed heat flow value.

III. Terrestrial heat flow measurement in the sea

The instrumentation and methodology of terrestrial heat flow measurement in the seas around Japan were described by Uyeda, Tomoda, Horai, Kanamori and Futi (1961) and some of the results of actual measurements were reported [Uyeda, Horai, Yasui and Akamatsu (1962), Yasui, Horai, Uyeda and Akamatsu (1963)]. Here we omit detailed descriptions on this subject, and only refer to the results of these measurement in the following sections.

Thermal conductivities of the sediment samples collected from the ocean bottom for the heat flow study in the seas were measured mainly in our laboratory. This was described in a previous section.

IV. Terrestrial heat flow in Japan

In Tables IV and V, all the heat flow values ever measured in and around Japan are listed with appropriate references. The number of heat flow values totals 51 including 39 measurements on land and 12 measurements in the seas.

A histogram of the heat flow values listed in Tables IV and V is shown in Fig. 10. Heat flow values at Shirane and Toyoha which are considered to belong to geothermal areas are excluded. In the figure

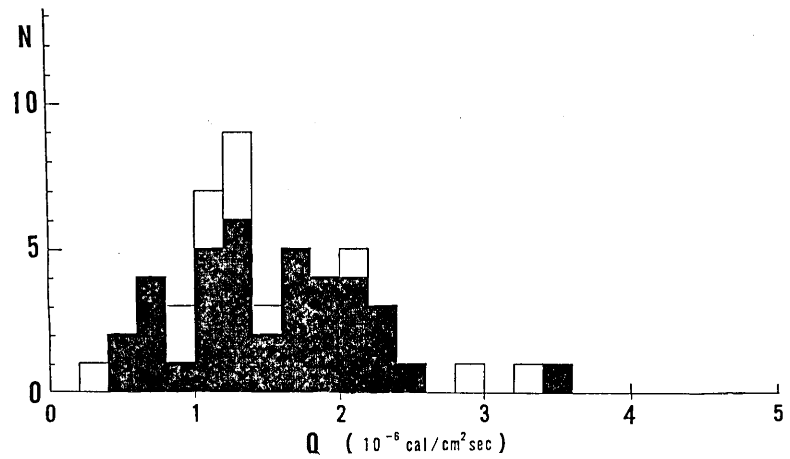

Fig. 10. Histogram of heat flow values in Japan.

Table IV. Heat flow

Station	Abbreviated name of station	Latitude (N)	Longitude (E)	Site*	Maximum depth, m	Temperature logging †
Haboro	HB	44°21′	141°52′	CM	350	1
Shimokawa	SMK	44 14	142 41	MM	533	2
Konomai	KNM	44 08	143 21	MM	524	1,2
Akabira	AB	43 32	142 02	CM	700	2
Ashibetsu	ASB	43 33	142 12	CM	500	1
Tayoha	TH	42 54	141 05	MM	400	2
Yabase	YB	39 44	140 06	OF	1700	1
Innai	IN	39 16	139 58	OF	1050	1
Osarizawa	OSZ	40 11	140 45	MM	300	1,2
Noda-Tamagawa	NDTG	40 04	141 50	MM	400	2
Kamaishi	KI	39 16	141 42	MM	529	2
Hitachi	HT	36 38	140 38	MM	550	2
Katsuta	KT	36 24	140 30	GF	900	1
Kashima	KS	35 57	140 41	GF	900	1
Mobara	MB	35 24	140 20	GF	1900	1
Tokyo	TU	35 42	139 46	OS	385	1
Ashio	AO	36 39	139 27	MM	796	2
Kusatsu-Shirane	KTSN	36 37	138 34	GA	250	1
Chichibu	CB	36 01	138 48	MM	401	1
Sasago	SG	35 37	138 48	RT	479	2
Kamioka	KO	36 21	137 19	MM	720	1,2
Nakatatsu	NT	35 52	136 35	MM	642	1,2
Kune	KN	35 05	137 50	MM	560	2
Nako	NKO	35 03	137 52	MM	640	2
Minenosawa	MNS	35 00	137 51	MM	190	1
Ikuno	IKN	35 10	134 50	MM	880	2
Nakaze	NZ	35 21	134 57	MM	365	1,2
Yanahara	YH	34 57	134 04	MM	940	1
Isotake	IST	35 11	132 26	MM	170	1
Tsumo	TSM	34 34	132 00	MM	309	1
Kawayama	KY	34 15	132 59	MM	517	1
Naka	NK	34 15	135 25	MM	640	1
Hidaka	HD	33 57	135 05	OS	310	1
Kiwa	KW	33 50	135 53	MM	413	1
Besshi	BS	34 01	133 09	MM	1600	1
Izuhara	IH	34 13	129 14	MM	480	1
Takamatsu	TM	33 52	130 43	CM	1000	2
Taio	TO	33 07	130 52	MM	560	2
Makimine	MM	32 38	131 27	MM	845	2

* MM: Metal mine, CM: Coal mine, OF: Oil field, GF: Gas field, RT: Railway

† 1: in borehole, 2: in drifts of mines.

data on land.

Isothermal pattern	Geothermal gradient, °C/100 *m*	Thermal conductivity, 10^{-3} *cal/cm sec °C*	Terrestrial heat flow 10^{-6} *cal/cm^2 sec*	Reference
	4.54±0.05	4.12 (0.61)	1.87 (0.30)	
L	3.04	5.63	1.71	Hôrai (1963c)
	{3.24±0.06			
	2.84±0.12}			
	3.96	6.41 (1.90)	2.54	
	{4.10±0.06		{2.63 (0.83)	
	3.81±0.09}		2.44 (0.80)}	
	2.49±0.04	4.31 (0.72)	1.07 (0.20)	
	3.08±0.03	4.38 (0.72)	1.35 (0.24)	
	11.3	~5	>5	
	4.80	4.19	2.01	
	4.80	3.11	1.49	Hôrai (1963a)
L	3.34±0.09	6.70 (2.91)	2.24	
L	1.38±0.11	8.28 (4.81)	1.14 (1.05)	
D	0.91	5.66 (2.22)	0.52 (0.81)	
	{0.66±0.05		{0.37 (0.19)	
	1.16±0.07}		0.66 (0.31)}	
L	0.94~1.21	6.73~6.50	0.63~0.78	
	3.01	3.02	0.91	Uyeda and Hôrai (1963a)
	2.1	3.57	0.76	
	1.85±0.02	2.94 (0.57)	0.54 (0.11)	
	2.20	3.36	0.74	
L	3.57±0.31	6.25 (0.29)	2.23 (0.31)	
	24.2	4.48	10.8	
L	1.90±0.07	7.06 (1.04)	1.34 (0.25)	
D	2.71	7.61	2.06	
L	2.77±0.04	6.49 (2.37)	1.80 (0.69)	
L	2.90±0.06	6.71 (1.26)	1.95 (0.41)	
D	1.97±0.31	8.14 (1.62)	1.60 (0.62)	
D	2.17±0.12	6.65 (1.01)	1.44 (0.31)	
	2.92±0.15	6.13 (1.05)	1.79 (0.40)	
D	1.88±0.09	7.33 (0.39)	1.38 (0.14)	
D	3.39		2.21	Uyeda and Hôrai (1963b)
	{3.35±0.07	6.31 (1.66)	{2.11 (0.61)	
	3.43±0.22}	6.72 (1.94)	2.30 (0.86)}	
	2.03	5.89 (0.80)	1.20	
	{1.92±0.02		{1.13 (0.17)	
	2.14±0.07}		1.26 (0.22)}	
	4.03	8.65 (3.75)	3.49	
L	1.80±0.05	6.08 (0.76)	1.09 (0.17)	
	1.49~1.98	5.76~5.94	0.86~1.14	
	3.04±0.06	5.90 (0.44)	1.79 (0.17)	
	2.87±0.02	7.40 (0.91)	2.12 (0.28)	
L	1.85±0.11	7.06 (1.79)	1.31 (0.43)	
L	2.49±0.04	4.89 (0.96)	1.22 (0.26)	
L	2.93±0.04	7.41 (2.33)	2.17 (0.72)	
	3.06±0.14	6.27 (1.59)	1.92 (0.60)	Hôrai (1963b)
D	1.71	6.16 (1.95)	1.05	
L	2.57±0.07	6.95 (1.97)	1.79 (0.57)	

tunnel, GA: Geothermal area, OS: Other type of site.

Table V. Heat flow data at sea.

Station	Latitude (N)	Longitude (E)	Water depth (m)	Geothermal gradient ($^{\circ}C/100m$)	Thermal conductivity $\left(\frac{10^{-3} cal/cm}{sec\ ^{\circ}C}\right)$	Terrestrial heat flow $\left(\frac{10^{-6}\ cal/}{cm^2\ sec}\right)$	Reference
E 1	38°09′	142°58′	1710	1.30	2.10	0.27	Uyeda, Horai, Yasui, Akamatsu (1962)
E 2	37 59	143 58	7345	5.42	2.11	1.14	
E 6	38 12	147 55	5631	10.5	1.95	2.05	
F 20	33 39	161 39	5605	6.81	2.00*	1.36*	Yasui, Hôrai, Uyeda, Akamatsu (1963)
F 23	34 23	142 15	7490	6.30	2.21	1.39	
F 24	34 04	142 56	5110	5.98	2.07	1.24	
F 25	33 53	145 26	5770	5.49	1.81	0.99	
Akko 7	39 22	150 03	5480	17.39	1.90	3.30	
Akko 8	39 30	143 28	2800	4.68-6.24	2.16	1.01-1.35	
MYJ 1	34 32	139 46	1710	5.74	2.54	1.46	
Akko 11	29 53	137 56	3960	3.97	2.06	0.82	
Akko 12	32 35	138 06	3970	12.0	2.42	2.88	

* Heat flow value at F 20 is tentative as no sediment was obtained.

solid columns show the values measured on land and hollow columns indicate the values in the sea.

The average of 37 heat flow values on land, is 1.55 $\mu\, cal/cm^2\ sec$. As the localities of measurement are distributed rather uniformly throughout the Japanese Islands, this figure could be regarded as the average value of heat flow upward through the crust to the surface in geothermally normal areas in Japan.

Birch (1954) deduced after reviewing the earlier works of geothermal heat flow on land, that the average heat flow on the continent would be near 1.2 $\mu\, cal/cm^2\ sec$. Subsequently, some authors such as Bullard slightly increased the mean heat flow on land to 1.4 $\mu\, cal/cm^2\ sec$, taking into consideration the heat flow determinations which had been presented since Birch's review.

Bullard and Day (1961) stated in their paper that the average heat flow on the sea floors of the Pacific and the Atlantic, excluding areas near the submarine ridges which are considered to be a sort of geothermal areas, is $1.08 \pm 0.054\ \mu\, cal/cm^2\ sec$, based on their own measurements and the results obtained by Bullard, Maxwell and Revell (1956), and von Herzen (1959).

Before we started the present geothermal work in Japan we had expected to observe higher heat flow than the world's average. Our reasoning was based on the high seismicity and volcanicity of Japan, as a part of the encircling Circum Pacific Orogenetic belt. But it has been shown that average heat flow values in Japan do not differ much from the standard values on the continents or the ocean floors.

Nevertheless, heat flow distribution in Japan has unique characteristics as will be described below. The interpretation of these characteristic features may reveal some facts regarding the thermal state of the crust in an orogenetic region.

Fig. 11 shows the geographic distribution of the measurement sites and the appropriate heat flow values. Several important features are

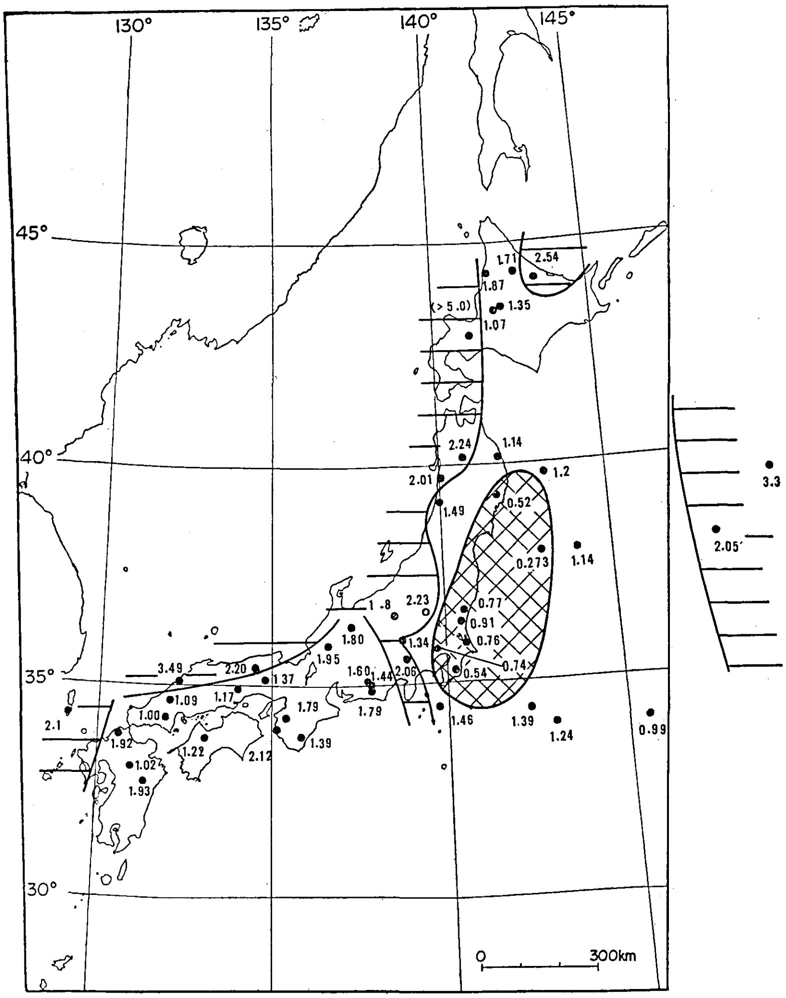

Fig. 11. Heat flow distribution in Japan. Unit: 10^{-6} cal/cm^2 sec.

noted in the figure.

i) There exists a region of high heat flow ($>2\ \mu\ cal/cm^2\ sec$) on the inner side of the Japan and Kuril arcs and a banch from it seems to extend along the Fossa Magna towards the Izu-Mariana arc.

This zone, shown in Fig. 11 by bold hatching, coincided with the region where volcanic activity has been prominent since middle Tertiary as described in Chapter I.

ii) Another notable feature is the existence of a region of low heat flow ($<1\ \mu\ cal/cm^2\ sec$) localized in the outer zone of north-eastern Japan. This region, shown in the figure by cross hatching, covers the Kitakami and Abukuma land masses and Kanto basin in the west. Its eastern margin is limited by the trough of the Japan trench, one of the most remarkable physiographic features in Japan.

iii) In regions other than the two mentioned above, heat flow values are on the 1 $\mu\ cal/cm^2\ sec$ level. This area includes the central part of Hokkaido and a large part of south-western Japan including central Honshu west of the Fossa Magna.

iv) Another high heat flow region seems to exist in the Pacific about 500 *km* off the east coast of north-eastern Japan.

The significance of this distribution will be discussed in the following chapter.

V. Discussion

As described in the previous chapter, the distribution of heat flow in and around Japan has distinct characteristics, *i. e.*:

i) high heat flow regions on the inner side of the Honshu and Kuril arcs, and the Fossa Magna.

ii) a low heat flow region on the outer zone of northeastern Japan.

iii) an intermediate heat flow region in the rest of the Japan Islands.

iv) a high heat flow region in the Pacific about 500 *km* off the coast of northeastern Japan.

The regional distribution of heat sources and temperature in the crust, as suggested by these data will now be discussed.

A. *Distribution of heat sources*

Terrestrial heat flow of 2 $\mu\ cal/cm^2\ sec$ is not a large value when compared to values obtained in regions other than Japan.

S. Clark and Niblett (1956) and S. Clark (1961), have recorded heat flow values ranging from 1.8 to 2.2 $\mu cal/cm^2\ sec$ in the mountaneous regions of the Austrian and Swiss Alps. Birch (1950) has also observed heat flow values of 1.7 $\mu cal/cm^2\ sec$ in the Rocky Mountain Range of North America. But in these cases, as described in the original papers, deep mountain roots composed of crustal materials are sufficient to supply the heat sources producing heat flows ranging 1.8 to 2.2 $\mu cal/cm^2\ sec$. So, in order to interpret the observed heat flow, one must take into account the distribution of heat sources and the crustal thicknesses.

Four regions were selected as representative provinces for an examination of the relationships between observed heat flow at the earth surface and subsurface distribution of heat sources. They are listed in Table VI, as A, B, C, D, and represent the following 4 regions respectively: A) Pacific coast of north-eastern Japan (Low heat flow region), B) Japan Sea coast of north eastern Japan (High heat flow region), C) Central part of Honshu (Region of moderate heat flow with thick crust) and D) Central part of Chugoku and Shikoku (Region of moderate heat flow with moderate crustal thickness).

Table VI. Crustal thickness and heat flow value in 4 representative regions in Japan.

Model	A	B	C	D
Site	Pacific coast of northeastern Japan	Japan Sea coast of northeastern Japan	Central part of Honshu	Central part of Chugoku and Shikoku
Crustal thickness C (km)	23	27	39	29
Terrestrial heat flow Q_0 ($\mu\ cal/cm^2\ sec$)	0.65	2.20	1.80	1.20
Temperature at 1 km of depth T_0 ($°C$)	20	40	35	30
Q_0/C ($\mu\mu\ cal/cm^3\ sec$)	0.28	0.81	0.46	0.41

In Table VI, the values of heat flow (Q_0) and the crustal thicknesses (C) in the regions A, B, C, D, are cited. Crustal thickness in Japan as proposed by Aki and Kaminuma (1963) was used in our calculations. In the 4th column of Table VI, the ratio Q_0/C is calculated. This quantity represents the rate of production of heat per unit volume of the crust when the observed heat flow is wholely attributable to the steady state of heat flow produced from sources uniformly distributed within the crust. The value of Q_0/C is 0.28 $\mu\mu cal/cm^3\ sec$ in region A, 0.81 $\mu\mu cal/$

*cm*³ *sec* in region B, 0.46 $\mu\mu cal/cm^3$ *sec* in region C, and 0.41 $\mu\mu cal/cm^3$ *sec* in region D.

Heat sources in the earth are usually attributed to the radioactive decay of elements such as U, Th, K contained in the rocks. In Table VII, Bullard's summary of the standard values for the radioactive heat

Table VII. Thermal conductivity, content of radioactive elements and rate of heat production of rock (after Bullard).

Rock type	Termal conductivity	Content of radioactive element in rocks (*gm/gm*)			Rate of heat production ($cal/gm \times 10^{-16}$)				Total rate of heat production
	(*cal/cm sec °C*)	U (10^{-5})	Th (10^{-5})	K (10^{-2})	U	Th	K	Total	($cal/cm^3\ sec \times 10^{-16}$)
Granitic rocks	~0.007	400	1300	3.5	940	820	300	2100	5500
Basaltic rocks	~0.005	60	200	1.3	140	140	110	380	1100
Ultrabasic rocks	~0.01	0.1	?	0.001	0.2	?	0.1	0.3	1.0

production of common rocks is reproduced. As seen in the table, the radiogenic heat production of granitic or basaltic rocks is far greater than that of ultrabasics, thus supporting the current concept that a large part of the heat source within the earth is concentrated in the upper layers of the earth, especially in the crust. A crust having a composite structure of granitic and basaltic rocks in any ratio must have a rate of heat production per unit volume which ranges between 0.11 to 0.55 $\mu\mu cal/cm^3$ *sec*.

For example, in the Austrian Alps, as described by S. Clark (1961), 1.85 $\mu\, cal/cm^2$ *sec* of heat flow at the surface can be explained by the underlying crust composed of a layer of granitic rocks, 35 *km* thick, with a rate of heat production of 0.5 $\mu\mu cal/cm^3$ *sec* and a layer of basaltic rocks, 10 *km* thick, with a rate of heat production of 0.1 $\mu\mu cal/cm^3$ *sec*. The average rate of heat production becomes 0.411 $\mu\mu cal/cm^3$ *sec* according to this model. This rate of heat production does not contradict the current concept of concentration of heat sources in the crust.

Our work in Japan shows that, Q_0/C in region B amounts to 0.81 $\mu\mu cal/cm^3$ *sec*, which exceeds the imposed upper limit of 0.55 $\mu\mu cal/cm^3$ *sec*. This fact indicates that under conditions of steady heat flow if the whole crust in region B has a granitic composition, the crust can produce only 64 % of heat flow observed at the surface of the earth. In consequence, at least 36 % of heat flow must be supplied from beneath the crust *i. e.* the upper mantle.

There is no reason to believe that the granitic rocks in Japan have an anomalously higher concentration of radioactive elements. For example, according to Dr. S. Okada (1961; private communication), the uranium content of the 266 granitic rock samples collected from several outcrops in Japan does not display any anomalous concentration. His result indicates a mean concentration of uranium in granitic rock of 6 ppm, with a standard deviation of about 3 ppm.

So, one must assume, the existence of some phenomena in the upper mantle under the high heat flow regions, which produce the additional heat. This phenomenon may possibly play an important role in maintaining this volcanic activity of the region. Further clarification of this process remains as a future task. But it may be possible to say that anomalous heat sources in the upper mantle are charactristic of orogenic regions.

The values for Q_0/C in regions C and D are moderate, *i. e.* 0.41 $\mu\mu cal/cm^3\, sec$ and 0.46 $\mu\mu cal/cm^3\, sec$. This suggests that we are possibly observing heat flows produced mainly in crustal materials of typical sialic composition, similar to the stable regions on the continents.

In region A, the value of Q_0/C amounts to 0.28 $\mu\mu\, cal/cm^3\, sec$, which does not violate the limitation imposed for a sialic crust. But the value is considerably less than the values in regions C and D. This fact may require an *ad hoc* explanation. Either the crustal concentration of heat sources is really anomalously small, or there is some mechanism operating underneath the crust which extracts the heat from the crust. For the latter, a downward convection current, a chemical process such as regional metamorphism etc., may be a possibility.

B. Distribution of temperature

Under the condition that the heat flow observed at the surface is flowing up through the crust by "steady state" conduction, one can calculate the temperature distributions within the crust. As we know the heat flow and the temperature near the surface of the crust, we can derive the temperature at any depth of the crust by integrating the equation of heat conduction in the steady state down to the desired depth. To achieve this, it is necessary to make assumptions regarding the distribution of beat source and thermal conductivity within the crust. We assume that the heat sources are distributed uniformly throughout the crust. Thermal conductivity was assumed to be a function of temperature as

$$K=\left(\frac{600}{300+T(^{\circ}C)}+4\right)\times 10^{-3}\, cal/cm\, sec^{\circ}C\,. \tag{3}$$

To set this expression, the experimental results of Birch and Clark (1940) on temperature dependence of thermal conductivities of rocks were consulted.

To carry out the calculations, the crust was divided into thin horizontal layers of $2\,km$ thickness, and the temperature was determined successively by the following formula:

$$T_{i+1}=T_i+\frac{Q_i}{K_i}\Delta D-\frac{H}{2K_i}(\Delta D)^2\,, \tag{4}$$

$$Q_{i+1}=Q_i-H\Delta D\,, \tag{5}$$

where

T_i : temperature at the upper face of the ith layer.
Q_i : heat flowing through the upper face of the ith layer.
K_i : thermal conductivity within the ith layer. K_i is assumed to be constant in the ith layer and is arranged to be $K(T_i)$ by the formula (3).
ΔD: thickness of the layer ($=2\,km$).
H : rate of heat production.

For the sake of comparison, temperature distribution is calculated for the regions A, B, C, D as selected in the previous section. Calculations were made from a depth of $1\,km$ taking the values of T_0 and Q_0 as listed in Table VI downward to the bottom of the crust, the depth of which is also indicated in the corresponding column in Table VI.

The calculation was made twice for each site, assuming two different values for H, *i.e.* H_1 and H_2. H_1 represented the value for the average rate of heat production of basaltic rocks, *i.e.*, 0.11 $\mu\mu\, cal/cm^3\, sec$ as listed in Table VII, assuming that this is the minimum rate of heat production of sialic crust. H_2 was equal to the rate of heat production which is obtained in regions A, C, D, if all heat sources were concentrated in the crust. In region B, H_2 was made equal to the average rate of heat production of granitic rocks, *i.e.*, 0.55 $\mu\mu\, cal/cm^3\, sec$ as listed in Table VII, assuming that this rate is the maximum heat production rate of sialic crust.

The results of the calculations are shown in Fig. 12. The curves

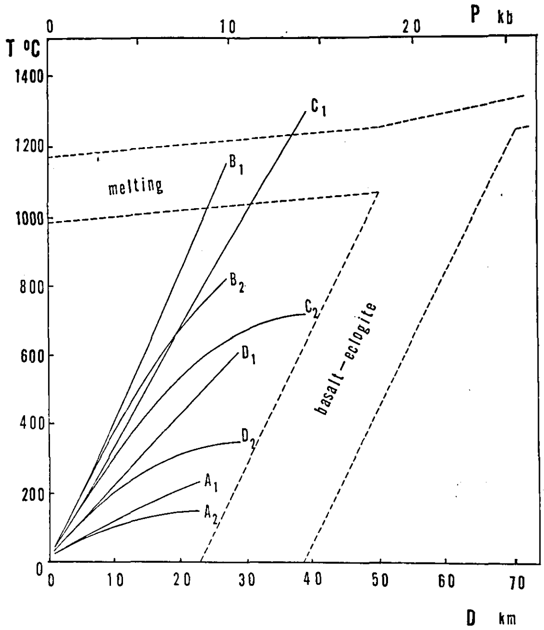

Fig. 12. Temperature-depth curves in the crust. Dotted lines are the phase diagram of basaltic rocks after Yoder and Tilley.

represent the temperature-depth (or pressure) relationships in each site with the two extreme values of H. As seen from the figure, the temperature distribution in the crust is markedly different at different sites. In Table VIII, calculated temperatures at the bottom of the crust are listed. In site A, the temperature is 150°C and 230°C according to the two values of H. In site B, the temperature exceeds 800°C at the bottom of the crust. At site D, the temperature at the bottom of the crust is moderate, *i. e.*, 350°C and 600°C as calculated for each value of H. At site C, the temperature at the bottom of the crust reaches 1300°C when $H_1=0.11$ $\mu\mu$ *cal/cm*^{3} *sec* is assumed as the rate of heat production. But this rate is probably too small for the rate of heat production of the deep mountain root of Central Honshu. When $H_2=0.55$ $\mu\mu$ *cal/cm*^{3} *sec* is used, the temperature at the bottom of the crust is near 700°C, and this may be more closely related to the real crustal tempearture under this site. If H_2 is more realistic than H_1 at sites A and D, the temperature listed in the second column of Table VIII, will be more accurate.

Table VIII. Calculated temperature at the bottom of the crust.

Site	Temperature at the bottom of the crust (°C)	
	Minimum rate of heat production (H_1).	Maximum rate of heat production (H_2).
A	233	147.
B	1146	816.
C	1297	716
D	605	346

In Fig. 12, the phase diagram for basalt is reproduced after Yoder and Tilley (1961). Zones in the figure represent the transition zones from basalt to eclogite, and from basalt and eclogite to their liquid phases. It is noted in the figure that, at site B, temperatures at the bottom of the crust are near the melting zone of basalt. This implies that if the basaltic materials are in existence at the lower crust or the upper mantle, thermal conditions there are favorable for them to melt and form basaltic magma. Some cause, such as a supply of excess heat coming from the lower part of the mantle, or a release of pressure may cause the formation of magma. In this case magma is generated under the state of $T=1000°C$ and $P=10$ to $15\,kb$. The possibility of magma formation under such pressures has some bearing on the problem regarding the origin of two types of basaltic magma [Kushiro (1962)]. In the site such as A or D, temperature at the bottom of the crust is too low to form basaltic magma in the uppermost layers of the mantle. These facts seem to be in accordance with the geological characteristics of each site, *i. e.*, site B is on the geologically active region as described in the previous chapters, while sites A and D are in the inactive regions.

Differences of temperatures at the bottom of the crust in various areas in Japan may bring about other consequences. For example, the magnetic, electric, or mechanical properties of the crust and the upper mantle can be expected to vary considerably, if the temperature distribution in the crust is markedly different as shown in Fig. 12. Experimental studies of these properties on rocks, and the interpretation of the internal constitution of the crust and the mantle based on the experiment and the analysis of observations would serve to confirm or reject the temperature distributions presented here.

Whether the M discontinuity is a phase transition or a compositional boundary between different materials is an important and interesting problem. As shown in Fig. 12, temperatures at the bottom of the crust coincide, in the sites A, C and D, to a precision of 2 to 3 *km*, with the basalt-eclogite transition temperature, if H_2 is assumed for the rate of heat production. This is against the phase transition hypothesis, because the assumption adopted here regarding the heat production rate in the crust requires a substantial concentration of radioactive elements into the crust. This is hardly compatible with the phase-boundary hypothesis in which the substances above and below M discontinuity is identical in chemical composition.

VI. Conclusion

Based on heat flow measurements at 51 localities in and around Japan, a general distribution of terrestrial heat flow has been obtained. The mean terrestrial heat flow value observed for the Japanese Islands is 1.53 $\mu\, cal/cm^2\, sec$. This value does not substantially differ from the average heat flow values recorded to date for the continents (1.4 $\mu\, cal/cm^2\, sec$). Since the Japanese Islands are one of the most active arcs of the Pacific, with high seismicity and volcanicity, it was first anticipated that the general heat flow might be much greater than in other areas. But resultant value failed to support this prejudice.

The heat flow distribution presented in this paper, however, seems to show some remarkable features which may be characteristic of an orogenic zone:

i. A region of high heat flow ($>2.0\ \mu\, cal/cm^2\, sec$) along the inner zone of the Honshu Arc.

ii. This high heat flow region seems to cross the Hoshu Arc at the Fossa Magna and extends down to the Izu-Mariana Arc.

iii. It is also possible that a similar high heat flow zone exists in the inner zone of the Kurile Arc.

iv. All of these regions of high heat flow precisely coincide with those of Cenozoic active volcanism and orogeny. The so-called "Green Tuff Region" corresponds to these regions.

v. Off the coast of north-eastern Japan near 150°E, there seems to be an area of high heat flow. The extent of this area is not known yet.

vi. Low heat flow ($<1.0\ \mu cal/cm^2\, sec$) prevails in the Pacific coast side of north-eastern Japan and in the oceanic area directly east of it. The eastern margin of the low heat flow region seems to be limited by the Japan Trench. A low heat flow region does not exist in south-western Japan.

vii. The area bounded by the above-mentioned high and low heat flow regions has a mean heat flow value which is normal.

Possible distribution of heat sources and temperatures in the crust, which are required to bring about the heat flow distribution as observed on the surface, were examined. The "steady state" solution of the heat conduction equation suggests that excess heat sources can be expected beneath high heat flow region. The temperature at the bottom of the crust should be about 800°C in the high heat flow region, while it is about 200°C in the low heat flow region. These results depend on the assumptions regarding the rate of heat production, thermal conductivity and the crustal thickness adopted in calculation. It seems clear, however, that even with some extreme values for the parameters mentioned above, the difference in the heat sources and temperatures at the Moho under the high and low heat flow regions cannot be denied. The high temperature which generally exists under the region of high heat flow would favor a production of magma in the upper mantle or lower crust regions. This harmonizes well with the fact that the high heat flow region is the region of recent volcanism. At the same time, the low temperature of the Moho on the eastern side of northeastern Japan ensures that magma can hardly be generated there.

Further development in the heat flow studies in Japan will contribute to the solution of a number of important geophysical and geological problems regarding the basic nature of the crust and the upper mantle, especially of the orogenic zone. The following investigations are scheduled as future tasks for geothermal studies in Japan.

i. To examine the validity of the assumption that the conduction of heat in the crust is in a steady state.

ii. The determination of the rate of heat production of rocks forming the crust and the mantle underlying Japan.

iii. Experimental studies of rocks, regarding their thermal properties under high temperature and pressure.

iv. Continued precise measurements of heat flow in the seas surrounding the Japanese Islands as well as on the land.

Postscript

This paper was originally written by the present author as a thesis for the University of Tokyo submitted in December 1962. The original paper consists of the text and the appendixes. According to the standing rules of the University, the author presents here the text of his thesis in a slightly modified form as the 13th paper of our series of papers on the thermal studies of the earth. The appendixes were already published in this bulletin as the 8th to 12th papers of the series.

Studies of the terrestrial heat flow in Japan were commenced in 1957 by Dr. S. Uyeda and others under the general direction of Prof. T. Rikitake of the Earthquake Research Institute of the University of Tokyo. The present author joined the project in 1958 and the studies as reported in the volumes of this bulletin and in other journals have been the author's joint work with Dr. S. Uyeda and others.

The summary presented here reviews the results of our studies up to December 1962. Stress was mainly placed on the details of measurement with possible interpretations of the results. Dr. S. Uyeda and the present author are planning to publish the reviews of their studies with more extended discussions in the near future.

The author thanks the former Directors of the Earthquake Research Institute, Dr. Nobuji Nasu and Dr. Ryutaro Takahashi, for their encouragement. Thanks are also extended to Prof. Tsuneji Rikitake and Dr. Seiya Uyeda for their kind guidance and cooperation. Prof. Hitoshi Takeuchi first suggested to the author the need of researches in this field of geophysics for which suggestion the author is deeply indebted. Thanks are due also to Prof. Robert B. Forbes, University of Alaska, for reading the manuscript and valuable advices.

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5. 地球熱学 第 13 報 日本列島の地殻熱流量

(1962 年 12 月における測定結果の要約)

地震研究所 宝 来 帰 一

東京大学地震研究所によつて 1957 年以来行なわれてきた地殻熱流量の測定は 1962 年 12 月までに，日本各地の 39 地点で行なわれた．他に気象庁と協同して行なつた海洋観測によつて，海域にも 12 地点での測定結果がある．

陸上で行なった 39 の測定のうち，26 は金属鉱山，4 は炭砿，2 は油井，3 は天然ガス井，1 は鉄道隧道，他の 3 はその他の目的のために掘鑿された試錐孔を利用したものである．地殻熱流量は地下温度増加率と地層の熱伝導率の積として求めた．

温度測定は主として，地表からの垂直ボーリングおよび鉱山の坑道内外のボーリングを用いて行なつた．測定のおよぶ深さは一般に地表から数 100 m 乃至 1000 m である．鉱山における地下温度測定の結果，地下等温面が地表の起伏に応じてほぼ相似の起伏を示す場合と，地表の形状に関係なくほぼ水平になつている場合とがあることにきづき，この相違は侵食による地形の変化と山体の温度伝導率の差によつてもたらされるものではないかと推論した．日本各地の地下温度増加率は，熱的に異常な地域 (地熱地帯) 以外では 0.5°C/100 m～5.0°C/100 m の範囲にある．

岩石試料の熱伝導率は divided-bar method によつて測定した．この方法は比較法であるので，その較正を，既知の熱伝導率をもつ標準物質 (crystalline quartz その他) によつて行なつた．本研究中測定した岩石試料数は 242 組で，36 種の岩石にわたつている．本文中では測定値を岩石の種類によつて分類し，一枚の図に記入して示した．岩石の熱伝導率は，その鉱物組成や組織にしたがつて $2\sim15\times10^{-3}$ cal/cm sec°C の範囲にわたる種々の値を示している．

地殻熱流量測定値に伴なう誤差は，温度測定値から地下温度増加率を推定するとき，および岩石試料の熱伝導率から地層の熱伝導率を推定するときのおのおのに伴なつて生ずる．その大きは一般に測定値の 20 乃至 30% 程度である．

日本列島およびその近海の地殻熱流量測定によつて，日本列島の地殻熱流量分布の一般的傾向が明らかになつた．すなわち，i) 日本列島弧 (千島弧・伊豆マリアナ弧を含む) の内帯に沿つて地殻熱流量の大きい ($>2\,\mu$cal/cm² sec) 地域があること．ii) 東北日本の外帯に地殻熱流量の小さい ($<1\,\mu$cal/cm² sec) 地域があること．iii) 本州東方海域に地殻熱流量の大きい地域があるらしいこと．iv) 上に述べた地域以外の地域では，地殻熱流量は $1\sim2\,\mu$cal/cm² sec の範囲にあること．なお，熱的に異常な地域に属すると考えられるものを除いた地殻熱流量測定値の平均は，陸上においては 1.54 μcal/cm² sec，海域のものについて 1.48 μcal/cm² sec である．

上に求められた地殻熱流量が，地殻の中の定常な熱伝導によつてもたらされるものと仮定すると，地下の温度および熱源の分布に関する推定を行なうことができる．地殻の発熱量，熱伝導率を適当に仮定して計算すると，高熱流量地域の地下では余剰の熱源の存在が期待されると同時に，Moho 面附近における温度も低熱流量地域の地下におけるよりも相当高温であることが予測せられる．今後測定を精密にすると同時に，岩石の熱的物性など関連諸分野の研究をも伴わせ行なつて，これらの推論が妥当であるかどうかを検討する方向に進むことが望ましい．