Heat Flow through the Deep Sea Floor*

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1. Introduction

In all modern theories of the development of the earth's surface features, the source of the distortion of the crust is, in some sense, thermal. On one view the energy is derived from the contraction of a cooling earth, on another it comes from very sluggish convection currents driven by the force of gravity acting on the density differences caused by a non-uniform distribution of temperature. This concept of the earth as a heat engine has led to greatly increased interest in observations of temperature and heat flow near the surface of the earth.

Even on land such measurements are not easy; the process of drilling disturbs the temperature of a bore hole, and many months may elapse before the hole settles to the temperature of the undisturbed rocks. To get an estimate of the heat flow the mean conductivity of the rocks penetrated by a bore hole or shaft must be estimated. The selection of representative samples is often difficult, either because a complete core is not available or because it is impossible to estimate accurately the proportions of the different types of rock. Even if the bore hole can be left to

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reach equilibrium and a satisfactory estimate of conductivity can be made, some heat may be carried off by the circulation of surface or underground water in cracks or porous beds; the product of the temperature gradient and the thermal conductivity will then not give the total amount of heat coming from the interior of the earth.

Owing to these and other difficulties there are only a few dozen reliable measurements of heat flow on the continents. The values range between 0.5 and 3 μcal/cm^2 sec (1 μcal = 10^{-6} cal) most of them being between 0.8 and 1.4. The mean is about 1.2 μcal/cm^2 sec. The regional means show surprisingly little variation, and the measurements are not sufficiently numerous or well distributed to tell if there is any correlation with the major structures of the continents.

The continental heat flow is easily accounted for by the radioactivity of the continental rocks. In fact, the whole observed flow could be produced by the uranium, thorium, and potassium contained in about 25 km of granite. Since a large part of the material above the Mohorovičić discontinuity at a depth of about 35 km must be granite or similar rocks, and since some heat must come from below the discontinuity, the difficulty is not so much to find a source for the observed heat as to explain why the flow is not greater than it is. The radioactive material is therefore usually supposed to be strongly concentrated in the crust, the material below the Mohorovičić discontinuity being much less radioactive than most surface rocks. This conclusion appears reasonable from the point of view of petrology and geochemistry.

The petrological study of oceanic rocks and seismological work at sea have shown that the crust beneath the ocean basins is strikingly different from that underlying the continents. Granitic rocks are completely, or almost completely, absent and a discontinuity, believed to correspond to the Mohorovičić discontinuity of the continents, occurs at a depth of about 10 km below sea level. The typical oceanic rocks are basalts whose content of radioactive materials is only about 30% of that of the continental rocks.

In view of the striking differences in petrology and structure between the oceans and the continents it might be expected that the heat flowing from the oceans would be only a fraction of that from beneath the continents. In fact this is not so; the main purpose of this review is to describe how this has been demonstrated and to discuss the results. Since the oceans cover 71% of the earth's surface, a reliable estimate of the amount of heat flowing through their floors is of great importance in discussions of the earth's thermal history. The oceanic measurements are not, however, merely an addition to the continental ones; in view of the fundamental differences between the oceans and the continents the heat flow at sea poses a problem separate from that of the continental heat flow. At sea the main problem is to find a source for the heat and a way of getting it to the surface without invoking temperatures above the melting point at a depth of a few hundred kilometers. Possible explanations are discussed in Section 5, none is free from difficulties.

2. Methods of Measurement

The first attempt to measure heat flow at sea was made by Pettersson [1] on the *Albatross* expedition, his measurements were of a preliminary nature and no values of heat flow have been published. The methods by which all the data discussed in this paper have been obtained were developed by the authors and their colleagues at the Scripps Institution of Oceanography in 1949 and 1950. They have been used in the Pacific by Revelle and Maxwell [2, 3] and in the Atlantic by Bullard [4].

As on land, the temperature gradient and the conductivity have been measured separately, the former by a probe stuck into the sediments of the ocean floor and the latter by laboratory measurements on samples collected with a coring tube. It has been suggested that a more direct determination could be made by placing a slab of material on the sea floor, allowing it to come to thermal equilibrium and measuring the temperature difference across it. With a slab of material a foot thick, of ordinary specific heat and density, and of conductivity 3×10^{-3} cal/cm °C sec, the temperature would take about four days to reach 90% of the equilibrium value and the temperature difference across it would then be about a hundredth of a degree. The method is therefore almost impracticable and certainly much more difficult than that using a probe, which gives a temperature difference of a few tenths of a degree and reaches 90% of its equilibrium temperature change in about 30 min.

2.1. *Temperature Gradient*

The probes used are 10 to 15 ft in length and 1 to 1.6 in. in diameter. The temperature difference between points near the top and bottom of the probe is measured by a thermistor bridge and amplifier or by thermojunctions connected to a galvanometer. The recording equipment is carried in a pressure-tight container attached to the top of the probe, as in Fig. 1. With the thermistor bridge a pen recorder is used, while the galvanometer records photographically on film. A more detailed account of the arrangement has been given by Maxwell *et al.* [5] and by Bullard [4].

The probe is lowered on a steel dredging cable (at least ¼ inch in diameter) attached to an eye bolt at the upper end of the case of the recorder. It is stopped about 100 m above the bottom and left there for about a quarter of an hour to allow it to come to temperature equilibrium.

Thc wineh is then run out at a speed of two or three meters a second until the probe has penetrated into the bottom and some slack wire has been run out. The probe is then left in the bottom for 30 or 40 min, pulled out and raised to the surface. As the penetration of the bottom takes place out of sight at a depth of two to three thousand fathoms it is difficult to be sure exactly what happens. Wind tunnel tests suggest that if the probe is run in fast it is desirable to provide stabilizing fins at the top of the recorder case; experiments in which a recording inclinometer was fitted showed that with the fins the probe penetrates within a few degrees of the vertical.

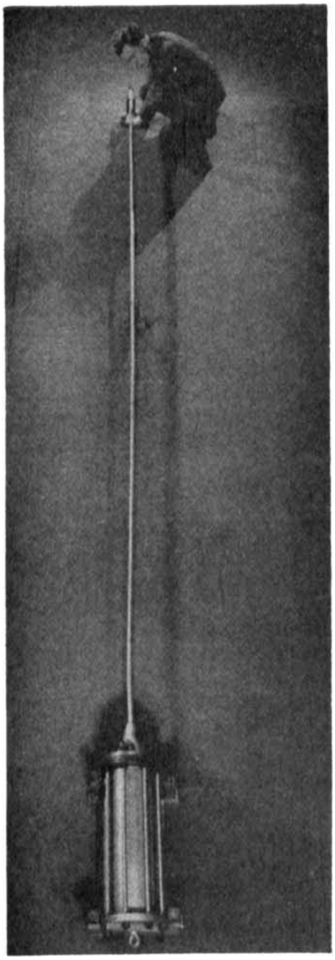

FIG. 1. Probe for measuring the temperature gradient in the ocean floor; the stabilizing fins are not shown.

In a flat calm, with no wind or current, the operation of the probe is simple, but even a slight wind or current causes the ship to drift and to pull the probe out of or through the bottom. To allow for this drift, it is necessary to continue to pay out the dredging cable while the probe is in the bottom. By skillful maneuvering the wire running down from the ship can be kept within a degree or two of the vertical, but this merely ensures that the ship is stationary relative to the top few hundred fathoms of water. Observations of a moored buoy have shown that it may be drifting at half a knot relative to the bottom. In practice, satisfactory results can usually be obtained with a wind up to 15 or 20 knots.

In penetrating the sediment the probe is heated by friction and by the work done in the plastic distortion of the material penetrated. If it were not for this heating, the probe would initially be colder than the sediment. It usually happens, however, that the heating is sufficient to raise the probe temperature above that of the sediment (Fig. 2). This rise in temperature is rapidly communicated to the thermistors or thermojunctions, and the recorded temperature rises for about 3 min. The rise is followed by a fall as the surplus heat spreads out in the sediment. It is impracticable to wait until a close approach to equilibrium has been attained, and a small correction is needed for the residual effect of the initial heating remaining at the end of the record.

The cooling of the probe and the estimation of the correction were studied in some detail by Ferris and Bullard [6]. They showed that, to a close approximation, the heat diffuses radially, and that longitudinal conduction in both the probe and the sediment may be neglected. It is also permissible to neglect radial temperature gradients in the probe after the first few minutes from the instant of penetration. With these assumptions

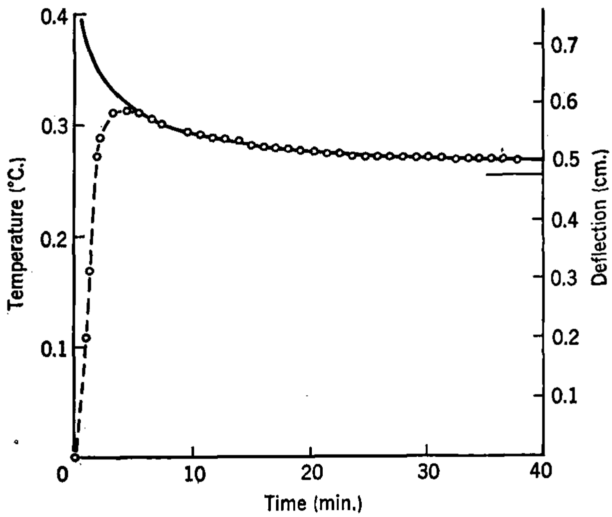

FIG. 2. Variation of the temperature difference between the top and bottom of the probe as a function of time. The full-line curve represents the theoretical expression (1).

the course of the cooling depends on $\kappa t/a^2$ and on a constant e, where κ is the thermometric conductivity of the sediment, t is the time since penetration, a is the radius of the probe, and $1/e$ is the ratio of the water equivalent of the probe (i.e., the number of calories required to raise its temperature by 1°C) to that of twice the volume of sediment. The temperature, T, at time t may be written

$$T = T_1 + T_0F\,(e, \kappa t/a^2) \qquad (1)$$

where T_1 is the final temperature when the initial disturbance, T_0, has subsided. The function F starts at unity and, as t increases, decreases to zero. F has been expressed as an integral involving Bessel functions and has been tabulated [4, 6]. When the disturbance has sunk to a small fraction of its initial value, equation (1) becomes approximately

$$T = T_1 + T_0 m/4\pi kt \tag{2}$$

where m is the water equivalent of the probe per unit length and k is the thermal conductivity of the sediment. In some experiments in the Atlantic with a probe 2.7 cm in diameter, m was 4.5 cal/cm °C, and k was 2.5×10^{-3} cal/cm °C sec. This gives

$$T = T_1 + 150T_0/t \text{ (sec)} \tag{3}$$

The disturbance will therefore be reduced to 10% of its initial value in 25 min, but would take 4 hr to sink to 1%. The observed cooling follows

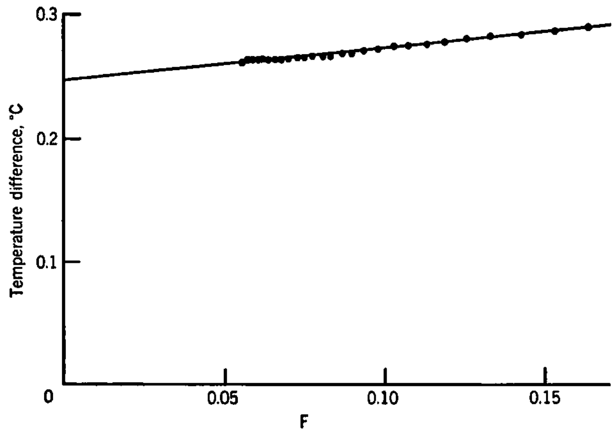

FIG. 3. Correction for lack of equilibrium. The temperature difference between the top and bottom of the probe is plotted against the theoretical function, F, of (1). The intercept of the best straight line through the points gives the temperature difference that would be attained after an infinitely long time.

that calculated from equation (1) very closely as is shown in Fig. 3 where T is plotted against F for the observations shown in Fig. 2. In such a plot the intercept on the T axis is the equilibrium temperature, T_1, freed from the effects of the initial disturbance T_0.

The temperature disturbance, T_0, is proportional to Q/m where Q is the heat generated per unit length of the probe during penetration. From equation (2), the time for the disturbance to subside to a specified level is thus proportional to Q/k. As k is a property of the sediment the only controllable factor affecting the time is Q. If the heating is largely by friction on the sides of the probe Q will be approximately proportional to the radius of the probe, a. Since m is proportional to a^2, the temperature disturbance will decrease as a increases, but the time needed to settle will increase about in proportion to a. If the heating is by plastic distortion of the clay near the point of the probe the temperature disturbance will be independent of radius while the heat generated and the time to settle will be proportional to the square of the radius. Since the principal experimental difficulty is to keep the probe stationary in the ocean bottom for a sufficient time, it is important to reduce the time necessary for an observation, and therefore to use a probe of as small a diameter as possible. It is doubtful if the diameter of the main part of the probe could be reduced below one inch without risk of buckling on entry, but some experiments have been made with a probe carrying an extension 0.25 in. in diameter at its lower end. The temperature near the surface of the sediment was measured by a similar short probe running parallel to the main one near its upper end. These showed that the initial temperature disturbance was greater for this probe than for a 1-in. probe, but that the time to settle was less. Unfortunately, the extension to the probe was broken every time it was used and the arrangement is therefore impracticable for routine use. It seems that the heat generated in penetration is more nearly proportional to a than to a^2 and that a large part of it must therefore be due to friction rather than to plastic distortion of the sediment.

Under favorable circumstances the errors in the temperature gradient caused by uncertainty in the measurement of the temperature and in extrapolation to equilibrium conditions probably do not exceed three per cent. It sometimes happens that the probe does not penetrate fully into the sediment; if this is not detected a large error will be made. Fortunately, if the probe does penetrate fully, mud is left on the bottom of the recording case, while mud is not present with partial penetration; incomplete penetration is also usually indicated by a sharp bend in the probe near the level of the surface of the sediment. It is convenient to provide an extra measuring point a few feet below the top of the probe so that a reliable gradient is obtained even if the probe does not penetrate fully; the extra measuring point also provides a check on the linearity of the temperature gradient.

If the probe penetrates at an angle θ with the vertical the quantity measured will be the temperature difference multiplied by $\cos\theta$. The error reaches 2% for an inclination of 12°. Experiments with a probe including a recording inclinometer have shown that there is no difficulty in inserting the probe within a few degrees of the vertical.

Comparison of results from probes of various diameters, of lengths from 10 to 15 ft, and with the measuring points at various positions give strong reason to believe that the measurements do give the true temperature gradient in the undisturbed sediment immediately beneath the ocean floor with an uncertainty of a few per cent.

2.2. Thermal Conductivity

Samples of the sediment were obtained by standard coring techniques [7, 8] at each station where the heat flow was measured. These core samples were normally from 6 to 30 ft in length and had diameters between 1.5 and 2.5 in. As soon as the core was retrieved it was either hermetically sealed for later determination or measured before any appreciable drying took place. Measurements of the moisture content of samples immediately after they were taken compared with measurements made on the same samples approximately two years later indicate no appreciable loss of moisture by evaporation during storage, although there was a tendency for the water to migrate under the influence of gravity. This migration may be minimized by storing the cores horizontally and by occasionally rotating them.

The properties of the sediment usually measured were the thermal conductivity, moisture content, and wet density, the latter two quantities being necessary for the determination of the thermometric conductivity, which is required for the correction of the measured thermal gradient for lack of complete equilibrium.

Thermal conductivities have been measured by both steady-state and transient methods. In the former [9], heat is produced at a known rate by a heater enclosed in a copper disk. The heat flows from the faces of the disk through two disk-shaped specimens to two water-cooled disks. The heat leaving the heater and that received by the water can both be measured which enables a good estimate to be made of the heat lost or gained from the sides of the disks. The effect of imperfect thermal contact between the specimens and the disks can be investigated by using disks of varying thickness. Loss of water and distortion of the specimens can be avoided by the use of a thin ring of Lucite around the periphery of the specimen. The method is straightforward and should be free from serious systematic error, it is however tedious, as it is necessary to wait for the apparatus to attain thermal equilibrium.

In order to make measurements more quickly a transient method was developed. In this a thin cylinder, in which heat is generated at a constant rate, is stuck into the specimen and the temperature of the cylinder is measured. After a time, t, large compared to b^2/κ, where b is the radius of the cylinder, the cylinder can be regarded as a line source. The temperature, T, is then given by

$$T = \frac{q}{4\pi} \frac{\ln t}{k} \tag{4}$$

where k is the thermal conductivity of the specimen and q is the rate of generation of heat per unit length of the cylinder.

The method is an adaptation of that discussed by van der Held and van Drunen [10] and by Mason and Kurtz [11]. By greatly reducing the dimensions of Mason and Kurtz's instrument it has been possible to remove the undesirable corrections for the finite diameter of the probe and for end losses, and to adapt it for use in core samples. The heating coil and temperature sensitive element are encased in a 20-gage stainless steel hypodermic needle with a diameter of 0.034 in. and a length of 2½ in. A ten-minute record gives sufficient information to determine the slope of the temperature time curve. Due to the very small mass of the needle and the excellent thermal contact between the moist sediment and the needle, the record is linear on semi-log paper for times from a few seconds to over ten minutes. Measurements of the thermal conductivity of naphthalene molded to the needle gave average results which agree within 0.2% with the value listed in the International Critical Tables [12], and a maximum deviation of 1.5%. The method therefore has been adopted for the rapid measurement of the sediment conductivity aboard ship immediately after the core sample is retrieved. It would be possible to adapt the method to the determination of conductivities *in situ* in the ocean bottom. The transient method does not give the conductivity along the axis of the natural heat flow; however, measurements have shown that the sediments have no anisotropy in conductivity.

The conductivities determined in the laboratory must be corrected so that they refer to the temperature and pressure occurring when the material is on the ocean floor. The temperature coefficient obtained by Butler [9] appears to be nearly independent of both the absolute conductivity and the moisture content of the sample. The correction to the conductivity measured at room temperature for the ambient temperature of the ocean bottom is a reduction of between 5 and 10%. Since the effect of pressure on the conductivity of minerals is very small its effect on the conductivity of the sediments is not likely to exceed that on the conductivity of water. For pure water at 30°C, the conductivity is increased by less than 3% by the pressures encountered at the ocean floor [12]. No corrections for pressure have been applied to the observed values.

Measurements of the moisture content of the sediment samples indicate that the thermal conductivity depends primarily on the moisture present, and only to a small extent on the mineral composition of the solid particles. It is possible to visualize the variation of conductivity with water content by a very simple model. A deep sea sediment consists of mineral particles separated by water, the particles having a conductivity several times greater than that of the water. The conductivity of the aggregate will depend on the proportion of water by volume, on the conductivity of the two components, and on the shape and size distribution of the particles. An arrangement of plates parallel to the direction of heat flow gives the maximum possible conductivity, while an arrangement of plates perpendicular to the heat flow gives the minimum possible conductivity. Such orientated arrangements of plates are highly anisotropic. A sediment is better represented by a random arrangement of plates or by spheres. The exact result naturally depends on the details of the distribution and size of the particles, but any reasonable assumptions enable the observed variation to be reproduced. In such a comparison the ratios

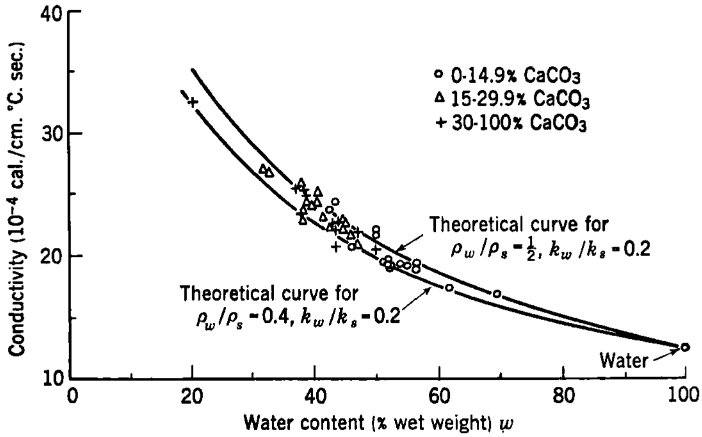

FIG. 4. Variation of conductivity with water content.

of the conductivity and density of the particles to that of water can, within limits, be treated as adjustable constants. A comparison of some observed conductivities with calculations for an assembly of widely spaced spheres is given in Fig. 4. The theoretical curves have been calculated from an expression due to Maxwell [13]:

$$k = k_w \frac{3 - 2v(1 - k_w/k_s)}{3k_w/k_s + v(1 - k_w/k_s)} \tag{5}$$

where k_s and k_w are the conductivities of the solid particles and of water, k is the conductivity of the sediment, and v is the proportion of water by volume; v is given in terms of the densities of particles and water, ρ_s and ρ_w by

$$v = w/[w + (1 - w)\rho_w/\rho_s] \tag{6}$$

where w is the proportion of water by weight in the sediment. More elaborate solutions for close packed particles of various shapes have been derived by Rayleigh [14] and by de Vries [15], they all give curves closely similar to those shown in Fig. 4. This figure also shows that sediments of varying carbonate content all lie near the same curve.

The water content and conductivity of a core are not uniform throughout its length, it is therefore necessary to adopt some average value. The most convenient method of allowing for small variations is to determine the water content of the whole core, or of a longitudinal section cut from it, and to use the conductivity corresponding to this; if a core is seen to contain two types of sediment these should be considered separately and the resistivities of the two sections added. Figure 5 shows the variation of conductivity with depth for two cores from the Pacific, one of green mud and the other of red clay.

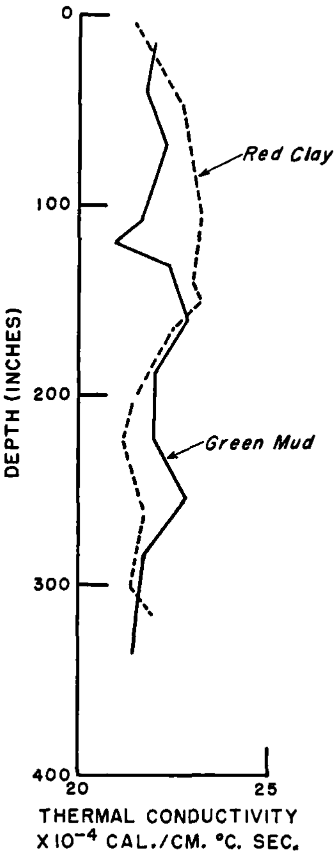

FIG. 5. Variation of conductivity with depth in two cores from the Pacific.

A rough value of the thermometric conductivity, κ, is needed for the extrapolation of the observed temperature gradients by the use of equation (1). The thermometric conductivity is $k/\rho c$, where c is the specific heat of the sediment; k and ρ are known and c may be obtained with sufficient accuracy by calculation from the specific heats of the components of the sediment. The specific heats of dry calcium carbonate, quartz, and red clay are 0.186, 0.170, and 0.181 cal/gm °C, respectively. No appreciable error will be incurred if the specific heat at zero water content is taken as 0.182, and the specific heat of the sediment calculated from this and the specific heat of water.

The variation of conductivity, density, specific heat, and thermometric conductivity with water content are given in Table I. These figures can only be taken as a general indication of the values usually found and will not apply accurately to all types of sediment. The densities in Table I were obtained by assuming that the solid particles had two and a half times the density of water, the conductivities are from Fig. 4.

TABLE I

100 w % wet wt	$10^4 k$ cal/cm°C sec	ρ gm/cm^3	c cal/gm°C	$10^4 \kappa$ cm^2/sec
20	33	1.92	0.35	49
30	28	1.72	0.43	38
40	24	1.56	0.51	30
50	20	1.43	0.59	24
60	18	1.32	0.67	20
70	16	1.22	0.75	18

3. Local Disturbances

The oceanic measurements of heat flow will be disturbed by topographic irregularities and should be corrected in the same way as are land observations [16]. This correction is usually small, if very rough topography is avoided. For example, a circular depression of radius d and depth h, having the form of a paraboloid, will give a heat flow at the center $(1 + 2h/d)$ times the undisturbed value. If the average slope, h/d, is 5% the error will be 10%. Calculations of the corrections to measurements of heat flow on land suggest that this is a somewhat extreme example [16, 17]. The topography at sea will rarely be known well enough to make a correction. Ripples on the bottom could produce an effect of a few per cent, but it is believed that large ripples are not common and the effect is not systematic [3].

If an irregular ocean floor composed of rocks of good or moderate thermal conductivity is buried by sediments of poor conductivity, the heat will flow out preferentially through the areas where the sediment is thinnest; that is, through the elevated areas of the original rock surface. The only case that has been worked out in detail is that of a series of parallel simple harmonic ridges of height l measured from crest to trough and of wavelength λ. The heat flow through the crests then exceeds that through the troughs by an amount δH given by

$$\frac{\delta H}{H} = \frac{2\pi l(k_r - k)}{\lambda k_r} \tag{6}$$

where H is the mean heat flow, k_r is the conductivity of the rock, and k that of the sediment. Since k_r may be over twice k the heat flow may vary by a large fraction of itself between crests and troughs; for example, ridges 10 km apart and 1 km high would give a 30% variation of heat flow if k_r/k is two. It is probable that examples such as this, chosen for geometrical simplicity, tend to give larger corrections than do natural topographic forms, but the example taken does show that the disturbance of heat flow by irregularities in sediment thickness is likely to be important and should be studied further. Somewhat extreme examples of similar variations on land have been investigated with an electrolytic tank and found to reach 30% [17, 18].

Convective motion of water through the sediment, or the churning activities of boring organisms could produce a transport of heat. If V is the vertical component of velocity in the sediments, it can be shown [3] that to a good approximation, when R is the radius of the convecting columns,

$$\frac{H_v - H_0}{H_0} = \frac{V^2}{\kappa^2} \pi R^2 \tag{7}$$

Where H_v is the actual heat flow and H_0 is the heat flow computed from the measured temperature gradient and thermal conductivity of the sediments on the assumption that $V = 0$. To account for as much as 10% of the heat flow, when $\pi R^2 = 1$ cm^2 the velocity would have to be at least 50 cm/day. This appears to be an impossibly high value.

Representative measurements of the outflow of heat would not be obtained if the measurements were made in the proximity of recent volcanic activity. The area of thermal disturbance would be related to the lateral and vertical dimensions of the activity and would also be a function of time. On the average, under the ocean, volcanism probably accounts for far less than 10% of the heat flow.

Other possible disturbances to the normal heat flow are rapid sedimentation or erosion. If the sediment is being rapidly deposited, some of the heat is utilized in raising its temperature, hence reducing the heat flow. Conversely, the heat flow might be increased by rapid erosion. The average rate of deposition of deep sea sediments is believed to be less than a centimeter per thousand years, therefore exercising a negligible influence on the heat flow [3].

Additional environmental factors affecting the heat flow could be suggested, and like the preceding factors probably could not be adequately detected. An evaluation of some of the environmental errors could be obtained by making several measurements in a restricted area and comparing the results. A first attempt in this connection has been made. The temperature gradient was measured at two points less than 8 mi apart and the thermal conductivity measured on three cores taken in an area, about ten mi in diameter, which included the above gradient measurements. The thermal gradients were 0.0483 and 0.0492 °C/m, a difference of less than two per cent. The average of the conductivities was 2.30×10^{-3} cal/cm °C sec with a maximum deviation of 1.3%. The computed heat flows are 1.11 and 1.13 μcal/cm^2 sec. The close agreement is reassuring.

4. Summary of Measured Values of Heat Flow

4.1. Pacific Ocean

Twenty-five measurements of the heat flow have been made in the Pacific Ocean on four separate expeditions of the Scripps Institution of Oceanography. Six measurements made on the joint U.S. Navy Electronics Laboratory—Scripps Institution Mid-Pacific Expedition and nine obtained on the Scripps Capricorn Expedition have been reported previously [2, 3, 19]. The remaining values were determined on two subsequent expeditions in the vicinity of the Acapulco Trench and in the East Pacific area. Table II lists all values obtained in the Pacific Ocean to date. The large deviation of many of the values from the mean makes it difficult to

Table II. Results Obtained in the Pacific Ocean.

Lat.	Long.	Depth in m	Heat flow μcal cm^{-2} sec^{-1}
20 48 N	159 42 W	4500	1.16
18 18 N	173 23 W	3900	0.72
19 28 N	174 35 W	4900	1.29
16 45 N	176 24 W	5040	1.19
19 02 N	177 19 W	4750	1.09
32 35 N	122 30 W	4000	1.27
0 40 N	169 17 E	4310	1.88
9 04 S	174 51 E	5000	1.35
18 59 S	177 36 E	2700	1.51
21 56 S	178 33 E	3900	2.58
17 28 S	158 40 W	4880	1.58
12 48 S	143 33 W	4300	0.36
14 45 S	112 11 W	3020	5.25
5 52 N	123 55 W	4100	1.65
14 59 N	124 12 W	4350	2.43
13 08 N	91 57 W	6170	0.47
11 55 N	91 37 W	3600	0.76
9 49 N	93 02 W	3730	0.25
12 14 N	98 44 W	3500	0.69
10 52 N	105 04 W	3300	>3.57
10 54 N	104 25 W	2950	2.73
12 12 N	111 04 W	3600	0.93
20 44 N	115 42 W	3910	1.19
25 01 N	123 04 W	4300	1.11
24 54 N	123 05 W	4200	1.13

assign significance to an average value. For example, the average of all 25 measurements* is 1.53 μcal/cm^2 sec with a standard deviation of 1.07 or 70 per cent. If the highest value is excluded the average becomes 1.37 μcal/cm^2 sec with a standard deviation of 0.78, which is 57 per cent. Sixteen of the 25 values lie within 50 per cent of 1.2 μcal/cm^2 sec, the average continental value. Of the remaining nine, 6 fall above, 3 below this range. The average of the 16 is 1.16 μcal/cm^2 sec having a standard deviation of 0.28 or 24 per cent.

Figure 6 shows the geographic distribution of the Pacific measurements. Although the dearth of adequate measurements prevents the correlation of the heat flow data with topography or the supposed boundary of the central Pacific basin, the andesite line, two important features are noted. The first is that the highest value yet obtained in the Pacific was measured on the presumably continental side of the andesite line, on the very broad topographical feature known as the Albatross Plateau. This plateau extends over large areas of the eastern Pacific Ocean both north and south of the equator. Subsequent measurements of the heat flow on the Albatross Plateau some 1500 mi to the north also yielded values considerably above the average, suggesting the entire topographic feature to have an anomalous heat flow. Secondly, a marked departure from the average heat flow has been found in the region of the Acapulco Trench. Here the values are considerably less than the average.

If the Trench and the surrounding area were covered with an equal thickness of sediment the topographic correction would give a higher heat flow in the Trench than elsewhere. But in the region of measurement the Trench probably contains a smaller thickness of sediment than the sea floor to the westward and the difference in thermal conductivity between the sediment and the underlying rocks would increase the heat flow in a manner similar to the increase for troughs parallel to ridges, discussed above. There remains a possible reduction in heat flow within the Trench because of rapid sedimentation. However, two measurements to the seaward of the Trench, though above the value for the Trench itself, are still well below the average for the Pacific Ocean. (A third very low value measured near the Trench may have resulted from either incomplete penetration of the probe or penetration at a large angle from the vertical.) Seismic refraction measurements, by R. W. Raitt, George Shor, and R. L. Fisher, of the thickness of sediments beneath the sea floor westward of the Trench strongly indicate that the rate of sedimentation is about the same here as elsewhere in the deep Pacific. Although more measurements are needed, we may tentatively conclude that the low heat

*For purposes of computation the value at 10°52′ N. Lat. and 105°04′ W. Long. is taken as 3.57 μcal/cm^2 sec.

flow in the neighborhood of the Acapulco Trench reflects a real difference in processes occurring far beneath the sea floor. The measured heat flows increase as one proceeds seaward until a maximum is found on the northern extension of the Albatross Plateau. Westward of the Plateau the value again returns to the average.

4.2. Atlantic Ocean and Mediterranean

Seven measurements have been made in the Atlantic Ocean and one in the Mediterranean. These are shown in Fig. 7 and Table III. The Mediterranean result was obtained at a point about 70 mi east of Gibraltar and less than that distance from the coasts of Spain and Africa. No seismic measurements have been made in this area, but it seems likely

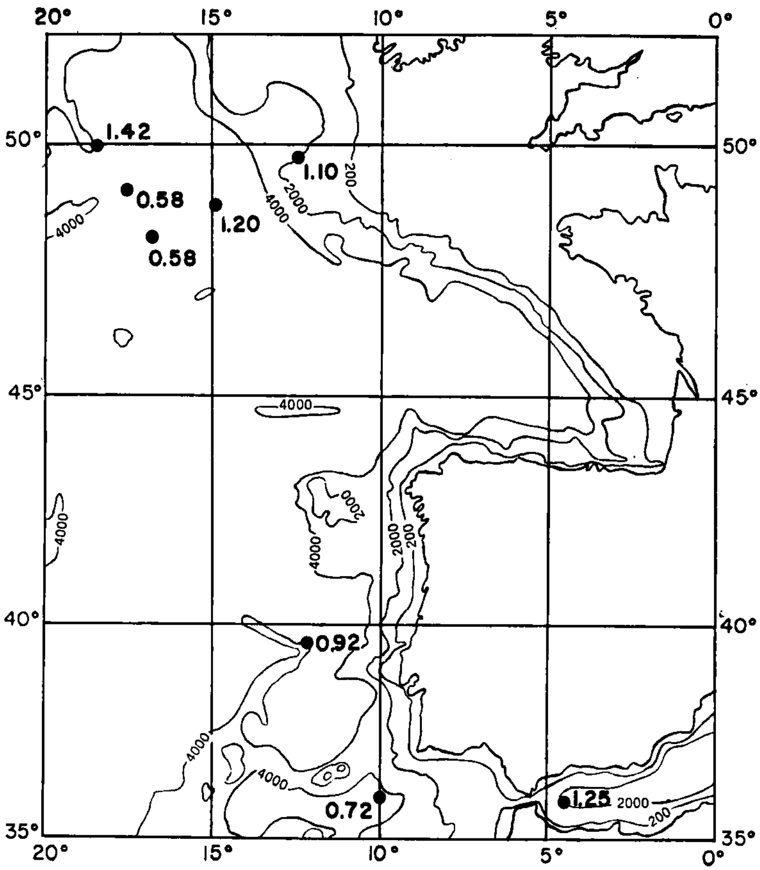

FIG. 7. Measurements of heat flow in the Atlantic Ocean and Mediterranean, heat flow in $\mu cal/cm^2$ sec, depths in meters.

that the structure of the crust is characteristic of the continents rather than the oceans. The agreement of the measured heat flow with the continental average is gratifying.

The average of the 7 Atlantic stations is 0.93 μcal/cm^2 sec, which is substantially less than the Pacific average. The average is considerably affected by two low results; these, like all the 5 northern stations were taken with a probe not provided with fins. Further measurements should

TABLE III. Results Obtained in the Atlantic Ocean and in the Mediterranean.

Lat. N.	Long. W.	Depth m	Heat flow μcal/cm^2 sec
49° 58′	18° 33′	4017	1.42
49° 49′	12° 30′	2032	1.10
49° 09′	17° 38′	4532	0.58
48° 52′	15° 00′	4710	1.20
48° 14′	16° 58′	4670	0.58
39° 39′	12° 08′	3020	0.92
35° 59′	9° 59′	4534	0.72
35° 58′	4° 34′	1251	1.25

be made to determine whether these low results are genuine or are due to the probe penetrating obliquely; without them the average is 1.07 μcal/cm^2 sec.

The measurements show no obvious correlation with topography, but are hardly numerous enough to detect one even if it existed.

5. INTERPRETATIONS

5.1. *Possible Sources of Heat*

Unlike the surface of the continental areas the ocean floor is believed to experience only minor temperature fluctuations. For example, the diurnal and seasonal variations of temperature in continental regions require that measurements of the heat flow be made at depths of several hundred feet to avoid the surface temperature disturbance. The water immediately above the ocean floor, on the contrary, appears to have a constant temperature over very large areas and has probably remained in this state for many thousands of years. An illustration of this constancy can be seen from an examination of water temperatures below 3000 m throughout the Pacific Ocean given by Sverdrup *et al.* [20, p. 746]. A temperature difference of less than 0.3°C exists between the bottom waters off the west coast of North America and those off Japan. Further, these temperatures have been observed to be constant over the past three decades. The fact that observed changes in the world wide sea level over the past hundred years can be accounted for almost exactly by melting of ice caps and glaciers indicates that the deep water temperatures have not changed appreciably for a century [21, 22]. This large thermal inertia of the ocean water makes the ocean floor an ideal laboratory in which observations of the heat flow through the earth's surface may be obtained.

An estimate of the disturbance in the heat flow caused by past fluctuations in the temperature of the water can be obtained by considering a change ΔT to have occurred in the temperature at some time in the past. If a measurement is made at a depth x after a time t has elapsed since the change in water temperature, the disturbance, $T(x)$ (that is the difference between the temperature at time t and that after an indefinitely great time) will be [23]

$$T(x) = \Delta T \operatorname{erf}\left[\frac{x}{(4\kappa t)^{1/2}}\right]. \tag{8}$$

The gradient of the disturbance will therefore be

$$\frac{dT}{dx} = \frac{\Delta T}{(\pi\kappa t)^{1/2}} \exp\left(\frac{-x^2}{4\kappa t}\right) \tag{9}$$

which gives $\Delta T/(\pi\kappa t)^{1/2}$ near the surface. Table IV gives the values of ΔT necessary to account for the whole of the observed gradient (taken as 6×10^{-4} °C/cm) in a 10-ft probe assuming κ to be 20×10^{-4} cm^2 sec^{-1}.

TABLE IV

Time (years)	$\Delta T'$ °C
1	0.30
10	0.89
100	2.7
1000	8.4
10,000	26.7

The existing data on deep water temperatures, although scarce, suggest that the temperature changes in the past have never been sufficiently large to account for any significant proportion of the observed gradient.

Assuming that the temperature of the bottom water has had a negligible effect on the observed temperature gradient, the measured outflow of heat must have its origin within the solid part of the earth. Many exothermic processes occur within the sediments; it is desirable to evaluate these in order that the oceanic heat flow may be compared with the continental heat flow. The two processes producing the most heat in the sediments are biological activity and radioactivity. Upper limits to the amount of heat produced by each of these are readily computed.

If it is assumed that all of the organic carbon in the deep sea sediments is completely oxidized at a rate equal to its deposition an upper value will be obtained for the biological activity. The fact that some organic material is still present in deep cores justifies this assumption. With a rate of deposition of 0.5×10^{-3} cm yr^{-1}, of which not more than 1% is organic carbon [20, pp. 1009 and 1036], we find that 0.0029 μcal/cm^2 sec is the maximum heat produced by biological processes [3].

The heat from the radioactivity of the sediments may be estimated from measurements of the uranium content in deep sea cores by Petterson [24]. Petterson finds uranium present in the sediments in quantities of about 1 gm/ton. Using a thorium/uranium ratio equal to that in continental rocks, a potassium content of 3.5×10^{-2} gm/gm with the rate of heat generation given by Birch [25] and assuming a sediment thickness of 400 meters, somewhat above the average found by Raitt [26], the amount of heat generated in the sediments by radioactive decay is found to be of the order of 0.004 μcal/cm^2 sec [3].

Other processes occurring within the sediments which contribute to the heat flow are chemical weathering, compaction, compression, and possibly, the isostatic adjustment imposed by the sedimentary load and the resulting loss of potential energy. However, it can easily be shown that the energy involved in these processes can never be sufficiently large to account for any appreciable fraction of the observed heat flow [3]. Thus, the total heat liberated within the oceanic sediments will not exceed one per cent of the observed values and for the purposes of the remaining discussion may be neglected.

The oceanic as well as the continental heat flow appears to have its origin below the sediments. In addition, the average heat flow in different regions is roughly the same, and it remains to be shown why this similarity exists. As mentioned previously, the whole observed continental flow could be produced by the uranium, thorium, and potassium in 25 km of granite; consequently, the heat flow there is predominantly a product of the continental crust. On the other hand, the oceanic crust is relatively thin, e.g., 5 to 10 km [26], and is largely composed of basaltic materials which have a much lower concentration of radioactive substances than has granite. The radiogenic heat from basaltic materials may be taken as 5.3×10^{-14} cal/gm sec [27] from which it is calculated that the oceanic crust can produce, at most, less than 10% of the observed value. The remaining heat must emanate from the mantle beneath the oceans. Such a large amount of heat arising from the mantle requires a striking difference between the mantle below the oceans and continents.

Heat sources within the crust and mantle that may appreciably affect the observed heat flow are radioactivity, the energy released by earthquakes, and the loss of potential energy that would accompany the growth of the core, continents, and oceans over geologic time. Gutenberg and Richter [28] have estimated the amount of energy radiated as seismic waves by shallow, intermediate, and deep earthquakes from 1904 to the present time. They report an average energy release per sec per square centimeter of the earth's surface to be 0.2 μcal/cm^2 sec. However, they have since [29] revised their estimates of the energy and suggest that the 1954 values are high by a factor of about 100. With this recent estimate, the earthquakes would produce only a few tenths of one per cent of the observed heat flow. In an earthquake zone the energy turned into heat will much exceed the world average of energy absorbed from seismic waves, it is possible that in these regions earthquakes may make an important contribution to the observed heat flow. It is more difficult to assess the energy involved in the growth of the core, continents, and oceans over geologic time, although Birch [25] has estimated the energy released if the ocean were produced as steam from the earth's interior over 3×10^9 years to be 0.002 μcal/cm^2 sec, for the water and 0.04 μcal/cm^2 sec for the lava. It appears that radioactivity is the primary source of the heat that is observed to escape at the earth's surface at the present time.

Of course, a fraction of the observed heat flow may be attributed to the original heat of the earth, this part will depend on the initial conditions and temperatures throughout the earth, but the low conductivity and large size of the earth preclude the possibility that this can be greater than 25%. The problem then is to distribute the radioactivity under the oceans in some manner such that the observed heat flow will be possible, the radioactivity will not be improbably high and the resulting temperatures will not be so great as to disagree with gravimetric or seismic evidence. Many attempts have been made to determine such an ideal distribution of the radioactive materials, but all are confronted with the common difficulty that with the available information the problem has no unique solution.

It is reasonable to suppose that the radioactivity under the oceans is in some degree concentrated upwards. Owing to the shallowness of the oceanic Mohorovičić discontinuity it cannot be concentrated above the discontinuity. A distribution through the upper 200 km of the mantle would require a radioactive heat production of about 4×10^{-14} cal/cm^3 sec, which is much greater than that of ultra-basic rocks or stony meteorites. If this solution is to be accepted it would be necessary to assume that the material beneath the oceans, down to a depth of some hundreds of kilometers, has a radioactivity nearly one third that of basalt. This is not inconceivable, but would require the material to be widely different from the ultra-basic rocks occurring at the surface. The temperatures at depths of 300 km would, on this assumption, be 300 to 600°C higher than at corresponding depths under the continents which would be difficult to reconcile with the gravity anomalies measured at sea and in the continental regions.

A distribution of the radioactivity to depths greater than 200 km with the usually assumed thermal conductivity of about 5×10^{-3} cal/cm °C sec leads to temperatures above the melting point of the rocks. The rigidity of the earth and the transmission of shear waves throughout the mantle require the mantle to be solid. Hence if the radioactivity is assumed to be distributed to great depths it is necessary to provide a means of getting the heat generated by radioactive decay to the surface that requires a lesser temperature gradient than does conduction with the value given above for thermal conductivity. The difficulty is made worse by the greater rate of heat generation that must be assumed to have occurred in the past. The half-lives of U-235 and K-40 are both below 10^9 years and detailed calculation [27, 30, 31] has shown that the relatively large amounts of these nucleides that must have been present 3×10^9 years ago would have more than doubled the present rate of generation of heat.

It is possible that the thermal conductivity of the mantle has been underestimated. Increase of temperature will decrease it below the laboratory value, but increase of pressure will increase it. Uffen [32] has estimated that on balance it may be eight times greater within the earth than has usually been assumed. Revelle and Maxwell [3] have shown that if the earth were initially "cool" a factor of seven or eight would prevent melting in an earth in which the radioactivity was distributed throughout the mantle.

5.2. *Convection*

An alternative method of bringing heat to the surface without excessive temperature gradients is by convection in the material of the mantle. Very slow convective motions suitable for this purpose have been suggested by many authors as a mechanism of orogenesis [33, 34, 35].

If a fluid is stirred by convection currents that are just able to maintain themselves, the temperature gradient must be nearly that due to the adiabatic expansion and contraction of the material occurring during the motion. If a particle of matter moves from a point A to a higher point B, its temperature will be reduced by expansion, and this reduction must be such that when it gets to B it is at the same temperature as the material surrounding B. If it were cooler than the surroundings it would be denser and would not have risen as far as B; if it were hotter the particle would be accelerated upwards, the convection would not have been "only just maintained" and would become more vigorous until the temperature difference was reduced to satisfy the adiabatic condition. The adiabatic temperature gradient β, is easily shown to be given by,

$$\beta = \frac{\alpha T_m g}{c_p} \tag{10}$$

where α is the volume coefficient of expansion of the fluid, c_p its specific heat at constant pressure, and T_m the temperature. Verhoogen [36] showed that α/c_p can be estimated by combining the seismic velocities and computed densities with extrapolations of laboratory data using relations derived from the theory of the solid state; it probably varies from about 4×10^{-12} sec^2/cm^2 near the surface to 1×10^{-12} at the base of the mantle. Equation (10) can be regarded as a differential equation which can be integrated to give the temperature at any depth if it is known at a single depth [37, 38, 39]. If the adiabatic gradient existed throughout the mantle and the temperature at a depth of 100 km were 700°C, the temperature at the base of the mantle would be 1300°C and the gradient would vary from about 0.4°C/km near the surface to 0.1°C/km at the base of the mantle.

The transmission of S waves and the amplitude of the bodily tides show the mantle to be a solid and not a fluid. There is, however, evidence, from isostasy and from the agreement of the ellipticity of the sea level surface with that calculated for a fluid body, that the material of the mantle can flow under the influence of long continued stresses. Flow at moderate depths is also indicated by the structure of the mountain ranges. It would clearly be difficult to suppose the intensely compressed structures seen in folded mountains to have been formed on a rigid base. The nature of the flow in rocks at the temperatures and pressures occurring in the mantle is not clear. They might yield like a fluid of very high viscosity; if this were so, motion would start at a temperature gradient only slightly greater than that given by equation (10). If the material were plastic it would require a finite stress difference to start movement, but once started this stress would produce an indefinitely great displacement. By considering the stress and density differences associated with a heated column of rock it can be shown that the stress difference in kg/cm^2 is of the same order as the associated gravity anomaly in milligals. As the mean gravity anomaly over large areas of the oceans rarely exceeds 50 mg, the stresses available to drive convection currents must be small compared with the breaking strength of rocks, measured in the laboratory, which is of the order of 10^3 kg/cm^2 at ordinary pressures and temperatures. The breaking strength is increased by rise of pressure and decreased by rise of temperature; it is therefore uncertain how it will vary within the earth. At high temperature, however, rocks show a tendency to "creep" in a way analogous to metals. This phenomenon has only recently been observed in igneous rocks and can occur far below the melting point [40]. Its laws have not been studied in detail, but it seems probable, by analogy with metals, that large movements can occur far below the melting point under small stress differences, provided sufficient time is available. Much further study, both experimental and theoretical, is required in order to estimate what rates of movement can reasonably be expected from the available stresses.

The possible temperature differences are limited by the gravity anomalies that would be associated with them. A vertical cylindrical column of rock of height h_c, radius r, density ρ_c and coefficient of volume expansion α at a temperature T_c above its surroundings would produce a gravity anomaly of about

$$2\pi G\rho_c\alpha T_c s$$

where s is the smaller of h_c and r, and G is the constant of gravitation. Putting $\alpha = 4 \times 10^{-6}$, $\rho_c = 3$, and $s = 300$ km, this gives $0.14T_c$ mgal. Temperature differences of one- or two-hundred degrees are therefore possible without the production of excessive gravity anomalies.

The rate of vertical transport of heat by a column at excess temperature T_c moving upwards with a velocity V is $\rho_c c_v V T_c$ cal/cm^2 sec, where c_v is the specific heat at constant volume. If T_c is 100°C and $c_v\rho_c$ is 1 cal/cm^3, this gives 100 V cal/cm^2 sec. A transport of 10^{-6} cal/cm^2 sec therefore needs a velocity of 10^{-8} cm/sec or 3 mm/yr; this is of the same order as the velocities usually assumed by proponents of convection in the mantle. If the velocity gradient is spread over 100 km, it implies a rate of strain of 10^{-15} sec^{-1}; this is only about a hundred times less than that measurable in prolonged laboratory tests, so that relevant experiments are not impossible. So far as our very inadequate information goes, it seems that there is no insuperable objection to the existence of convection currents from the point of view of the likely mechanical properties of rocks.

In a steady state, the amount of heat brought to the surface by a particle taking part in convection cannot greatly exceed that generated by radioactivity while it is traveling once round its path. This gives a rate of arrival of heat at the surface of about 2 h_cW per unit area, where h_c is the depth of the convection cell and W is the rate of generation of heat per unit volume. It is interesting that this is independent of the velocity; a slowly moving current brings less material to the surface than a fast one, but has longer to heat up.

The rate of heat generation within the earth is unknown. It has been suggested that the mantle may be composed of material similar to that of stony meteorites or to the ultra-basic rocks found at the surface of the earth. The amounts of uranium, thorium, and potassium in such materials are uncertain; from such data as exist the rate of heat generation has been estimated as 6.3×10^{-15} cal/gm sec for stony meteorites [41] and 1.2×10^{-15} cal/gm sec for ultra-basic rocks [27]. If all the heat generated in a mantle of stony meteorites were transported to the surface by convection, the heat flow would be 4.3 μcal/cm^2 sec; for a mantle of ultra-basic rock it would be 1.0 μcal/cm^2 sec.

If convection through the whole thickness of the mantle were occurring at a near-adiabatic temperature gradient and the temperature at the top of the convection cells were 700°C, the temperature at the base of the mantle would be only 1300°C which is below the melting point of iron at the core boundary, estimated by Simon [42] to be 2700°C. The discrepancy could be resolved in one or more of three ways: the temperature at the top of the convection cells may be higher than 700°C, as suggested by Verhoogen [43] from consideration of the probable temperatures of production of basaltic lava; the average temperature gradient required to maintain convection in materials of finite strength may be two or three times the adiabatic value [43]; or convection may be possible only in the outer parts of the mantle, where the temperatures are closer to the melting point of silicate materials than in the interior. (The melting point gradient in the mantle, though uncertain, is probably at least four or five times the adiabatic gradient.)

If convection were confined to the upper 1000 km of the mantle, a radioactive heat generation of about 3.5×10^{-15} cal/gm sec would be necessary over this range of depth. The radioactivity at greater depths would have to be small enough to avoid a near approach to melting but sufficient to maintain the temperature at the base of the mantle above the melting point of iron. It appears that the observed heat flow could be accounted for by convection not extending to the base of the mantle without requiring an impossibly high rate of heat generation and that, at the same time, temperatures of a few thousand degrees can be supposed to exist in the interior.

Bullen [44] has argued that the mantle can not be made of the same material at all depths. His computations show that an earth with a uniform material having the elastic properties indicated by the velocity of seismic waves, the correct total mass, and a reasonable surface density, would not have the observed moment of inertia. This discrepancy can only be removed by supposing the density in the upper part of the mantle to increase more rapidly with depth than would be expected from the compressibility judged from seismology. Birch [37] has arrived at a similar conclusion by comparing the observed rate of increase of the seismic velocities with depth with that expected from a theoretical equation of state. He places the inhomogeneous region at depths between 400 and 1000 km.

A variation in chemical composition with depth would be inconsistent with convection, since the convection currents would produce mixing and destroy the variation. Also, any substantial increase in the heavier materials with depth would make the system stable and prevent convection. Convection cells confined to the region shallower than 400 km might be possible, but it seems unlikely that sufficient heat is produced in this region to provide the observed flow at the surface.

If the change required by Bullen and Birch were a change in crystal structure induced by pressure, the difficulty might be avoided. It is possible for such a change to be spread over a range of depth and to provide, in effect, an increase in compressibility. The change would not, however, take place instantaneously and would not follow the pressure changes associated with the passage of a seismic wave; it would not, therefore, form part of the seismological compressibility. It is not clear if a convection current could persist through a layer in which such a transition occurred. For this to be possible, the change must occur without appreciable lag as the material rises. As the motion may take about a hundred thousand years to go a kilometer, there may well be sufficient time for this.

In brief, it seems not impossible that the objections to the existence of convection currents in the mantle can be overcome by reasonable assumptions about the properties of the material of which it is composed. That the mantle does really possess these properties is not self-evident, in fact, the main reason for supposing that it does is the desire to have convection currents to account for the oceanic heat flow.

Gold has postulated the existence of liquid channels in an otherwise solid mantle. If these exist and are large enough to transfer a useful amount of heat by convection, they may produce a detectable scattering and absorption of seismic waves, but no detailed discussion has been published.

If the heat flow at sea is largely brought to the surface by rising convection currents, there should be corresponding regions of low heat flow in places where the material is sinking. The Acapulco Trench area may be an example of such a region. In laboratory experiments with liquids, convection cells usually have hot rising cores and cool sinking peripheries; the hot regions of the upper surface are therefore completely surrounded by cool ones. The reason for this distribution is not understood, and it cannot be assumed that the mantle of the earth will necessarily behave in this way. It might be that the positions and forms of the convection cells would be determined largely by temperature disturbances due to features near the surface, such as continental edges. If the sinking currents lie beneath the continents, the low heat flow would be obscured by heat coming from the crust. The form of the currents will not be appreciably affected by the rotation of the earth.

The rough equality of the oceanic and continental heat flows can scarcely be a coincidence. The simplest explanation would be that the radioactivity originally in the upper part of the mantle has been concentrated in the crust, allowing the continental heat to escape by conduction, while beneath the oceans the same amount of radioactivity is still distributed through the mantle and the heat is brought to the surface by convection or by unexpectedly high thermal conductivity.

List of Symbols

a radius of temperature probe
b radius of cylinder used in conductivity measurements
c specific heat of sediment cal/gm °C
c_p specific heat of rock at constant pressure
c_v specific heat of rock at constant volume
d radius of circular depression in the sea floor
e $2\rho c$ times probe volume divided by water equivalent of probe
F function symbol
g acceleration of gravity
G constant of gravitation
h depth of circular depression in the sea floor
h_c height of rock column in mantle convection cell
H average heat flow
H_o heat flow assuming no vertical movement of material in sediments
H_v heat flow from both conduction and convection in sediments
δH excess of heat flow through crests over that through troughs
k thermal conductivity of sediments cal/cm °C sec
k_w thermal conductivity of water
k_s thermal conductivity of solid particles in sediment
k_r thermal conductivity of rock
l height from crest to trough
m water equivalent of temperature probe per unit length
q rate of heat generation per unit length in cylinder used for conductivity measurement
Q heat generated per unit length in temperature probe
r radius of rock column in mantle convection cell
R radius of convecting columns of sediment
s smaller of h_c and r
t time
t_0 starting time

T temperature at probe thermal junction at time, t
T_0 initial temperature disturbance at probe thermal junction
T_1 final temperature at probe thermal junction
T_c temperature difference between rock column and surroundings
T_m temperature in mantle
$T(x)$ temperature difference at depth x between temperature at time t_0 and that at an indefinitely great time
ΔT past change in temperature of ocean bottom water
v proportion of water by volume
V vertical component of velocity
w proportion of water by weight
W rate of generation of heat per unit volume
x depth below sediment surface
α volume coefficient of expansion
β adiabatic temperature gradient
θ angle with the vertical
κ thermometric conductivity $= k/\rho c$, cm^2/sec
λ distance between crests or troughs
ρ density of sediment
ρ_c density of rock
ρ_s density of particles in sediment
ρ_w density of water

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