8. HEAT-FLOW MEASUREMENTS DSDP LEG 37[1]

R.D. Hyndman, Victoria Geophysical Observatory, Earth Physics Branch,
Department of Energy, Mines, and Resources, Victoria, B.C., Canada
R.P. Von Herzen, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts
A.J. Erickson, Geology Department, University of Georgia, Athens, Georgia,
and
J. Jolivet, Institute de Physique du Globe, Université de Paris VI, Paris, France

ABSTRACT

Heat-flow measurements have been made in five deep crustal holes drilled into the Mid-Atlantic Ridge by *Glomar Challenger* that provide new information on the hydrothermal circulation that probably controls the crustal temperature distribution near ocean ridge spreading centers. Nearly constant very low heat flow of 0.6 $\pm 0.1\ \mu cal\ cm^{-2}\ sec^{-1}$ ($25 \pm 4\ mW\ m^{-2}$) was found for three holes and three ocean probe measurements across one 3-km-wide sediment pond on 3.5 m.y. old sea floor and to a depth of 400 meters into the underlying basaltic basement. The heat flow predicted for theoretical models of plate accretion is much higher, about 6.4 (269). Heat must be carried to the surface by extensive hydrothermal circulation that may extend to the lower part of the crust. Temperature measurements in the basement part of one borehole into 13 m.y. old sea floor indicate that seawater was flowing rapidly down the hole into a permeable horizon near the bottom. The heat flow of 0.5 (21) measured in the overlying sediments suggests that before the hole was drilled, water may have been flowing horizontally in this horizon at about 10°C.

INTRODUCTION

There is a serious discrepancy between the geothermal heat flux measured and that predicted by theoretical models for spreading ocean ridges. Over a thousand heat-flow measurements have been made on ocean ridges using the standard ocean probe technique. They show a very large scatter with some high values, but the smoothed means are little above the averages for older ocean basins (e.g., Langseth and Von Herzen, 1971). In contrast, all theoretical conductive cooling models for the formation of the oceanic lithosphere by upwelling under ocean ridges predict very high heat flow near the ridge crests, decreasing smoothly with increasing age of the sea floor. This pattern is obvious in the simple cooling plate models (Langseth et al., 1966; McKenzie, 1967; Sleep, 1969; Sclater and Francheteau, 1970; Parker and Oldenburg, 1973; Davis and Lister, 1974). The pattern is similar, although the ridge crest values predicted are somewhat lower, if the detailed geometry of the ridge crest volcanism is considered (e.g., Palmason, 1973) and if the effects of the melt crystallization and phase changes are included (e.g., Bottinga, 1974). There is strong support for the general form of the theoretical models from the elevations of ridges and from their gravity field, which are readily explained by thermal expansion. Thus, the major source of the discrepancy must be sought in the heat-flow data.

The normal ocean probe measurements involve penetration of only 2 to 5 meters into the sea floor sediments, and these may be subject to the disturbing effects of short-term variations in bottom water temperatures, small-scale bottom topography, and local sea floor processes. However, there are now over 20 well-determined heat-flow measurements in the deep holes drilled several hundred meters into the sea floor sediments by *Glomar Challenger* (Erickson et al., 1975) which agree well with nearby ocean probe measurements, although the deep holes have less scatter of values.

Confirmation of the validity of ocean probe values of heat flow also comes from a comparison with the heat flow found in the preliminary Mohole site (Von Herzen and Maxwell, 1964) and with the heat flow found in a deep hole drilled into the island of Bermuda (Hyndman et al., 1974b). Thus the source of the disturbance to the thermal field must be deeper than several hundred meters.

It is now clear that the low and variable heat-flow values from ridge crests arise because of hydrothermal circulation in the oceanic crust. Heat is lost in the region of the ridge crests by water convection in the fractured and porous crustal rocks rather than by conduction and thus is not measured by the normal sedi-

[1] Contribution of the Earth Physics Branch No. 596.

ment heat probe (Palmason, 1967; Langseth and Von Herzen, 1971; Talwani et al., 1971; Lister, 1972; Hyndman and Rankin, 1972; Anderson, 1972; Williams et al., 1974; Sclater et al., 1974). A recent compilation of all heat-flow values taken where there is a uniform blanket of sediment to seal off the water circulation, so that all heat must be conducted to the surface, although circulation may continue in the crustal rocks (J.G. Sclater, personal communication), shows good agreement with theoretical models. The hydrothermal circulation is of vital importance not only to thermal models, but also to our understanding of the genesis of the oceanic crust and of metalliferous sediments. Hydrothermal circulation is needed to explain the observed alteration in crustal rocks (e.g., Aumento et al., 1971; Muehlenbachs and Clayton, 1971; Hekinian, 1971; Hart, 1973; Spooner and Fyfe, 1973) and is undoubtedly the source of the metal-bearing submarine hydrothermal solutions required to produce many of the metallic sediments observed (e.g., Corliss, 1971; Cronan et al., 1971; Dymond et al., 1973; Anderson and Halunen, 1974).

Leg 37 provided an opportunity to obtain additional heat-flow measurements in deep sediment holes near the crest of a spreading ridge and for the first time to make measurements in deep crustal holes that might outline the nature and extent of the hydrothermal circulation. Heat-flow values were obtained in the sedimentary sections of five holes and in the deeper basaltic basement in two holes, with crustal ages from 3.5 to 13 m.y. One sediment pond was studied in detail with three holes and three standard ocean probe measurements.

In the following discussion the conductivity and heat flux values are given first in units of mcal cm^{-1} sec^{-1} $°C^{-1}$ and μcal cm^{-2} sec^{-1}, respectively, and secondly, in brackets, in units of Wm^{-1} K^{-1} and mW m^{-2}. Gradients are given in units of °C km^{-1} or equivalently of $mK^{1}m^{-1}$.

HEAT-FLOW MEASUREMENTS IN BASEMENT HOLES

Near spreading ridge crests, temperature and thermal conductivity measurements in deep crustal holes should permit a check on the hydrothermal circulation hypothesis. The best method for confirming the hypothesis is to measure temperature in the rock where circulation may be occurring since zones of significant circulation penetrated should be defined by the temperature-depth profile. A uniform gradient result also may be used for comparison with models of general deep circulation and conduction.

Heat-flow measurements near spreading ridges vary by more than two orders of magnitude, so that even quite low accuracy results can give very useful information. A gradient with an error of 20% is a useful result. Higher accuracy is needed for useful interpretation away from ridges, for example, to define variations with sea floor age and depth. The temperature measurement technique used in sediment holes cannot be used in crystalline basement so that temperatures can be measured only in the hole itself where there is a large disturbance from the drilling fluid. Thus measurements must be made some time after drilling circulation has stopped. A large part of the disturbance must have decayed for an accurate extrapolation to equilibrium. A gradient accuracy of ±20% may be obtained by extrapolation using a single measurement if about 50% of the disturbance has decayed (see below). With three or more temperature measurement repetitions to define the decay of the disturbance, a gradient accuracy of ±20% may be obtained if about 30% of the disturbance has decayed. If less than 30% of the disturbance had decayed even repeated measurements will not permit a gradient of useful accuracy to be obtained, because of the limit of accuracy of temperature measurement and because of the necessary approximations in the models used for extrapolation to equilibrium.

TEMPERATURE MEASUREMENT TECHNIQUES

There are three possible approaches to obtaining estimates of equilibrium temperatures in deep crustal holes.

1) The bottom-hole temperature may be recorded at several depths during the drilling. The duration of the disturbance at the bottom is short so that the return to equilibrium is rapid. However, at least several hours of recording after the drilling circulation has stopped are required for sufficient decay of the disturbance. The recording can be done with the instrument used for sediment temperatures (Erickson et al., 1975). Temperatures must be recorded as close as possible to the bottom where the disturbance time is shortest. Ideally, the temperature sensor would rest on the bottom and be connected to the instrument in the core barrel by thin wires so that it is not moved by the vertical motion of the ship and drill pipe. Reasonable results might be obtained with the sensor inside the bit, and the bit resting on the bottom, but vertical ship's motion then may pump water through the drill bit and there may be a problem of the bit becoming stuck in "sticky" holes if it rests on the bottom without circulation or rotation. Near hole bottom temperature measurements were attempted at Sites 334 and 335.

2) A completed hole, or a hole temporarily suspended for bit replacement, may be logged over its whole depth. At least 1 day will be required after termination of drilling for sufficient approach to equilibrium temperatures and then only temperatures from near the bottom of the hole will be useful. Repeated logging will help define the decay curve. A bit change and re-entry into the hole requires 1 to 2 days, so immediately after a re-entry before circulation is commenced is the obvious time for temperature measurement. The present downhole temperature recorders are quite suitable for logging temperature in the drill pipe but not in the open hole. Fortunately, the thermal mass of the pipe is relatively small compared to that of the fluid in the hole so the pipe will equilibrate sufficiently with the fluid and the surrounding rock in about 1 hr. Slightly longer times may be required for the drill collar assembly, which makes up the bottom 80 meters or so of the drill string. Thus, the drill pipe may simply be lowered to the bottom and a temperature log made inside it. Lowering

the temperature recorder in the pipe attached to a core barrel disturbs the temperatures in the pipe slightly so it is better to hold the instrument for about 2 min at depth intervals, say 20 meters, rather than to move it continuously. Measuring during instrument lowering also will give less disturbance than during raising the instrument. Temperatures were measured after a bit change and re-entry in Hole 332B.

3) A string of temperature sensors may be placed in the hole through the re-entry cone after completion, leaving a recoverable recorder or an acoustic telemeter to transmit the data to the surface. This method would give the best results, but is the most difficult and has a high chance of complete failure. It has not yet been attempted.

Decay of Drilling Disturbance in Glomar Challenger Boreholes

As described above, it usually will not be possible to wait for temperatures to reach equilibrium in *Glomar Challenger* deep crustal holes. Thus, the measurements must be extrapolated to equilibrium using the geometry of the hole, the thermal properties of the rock, and drilling fluid and the disturbance history. The extrapolation is of critical importance to the accuracy of the heat-flow result so it is considered in some detail below.

Description of the Problem

The process of drilling causes disturbance of temperature because of the heat exchange between the drilling water and the rock surrounding the hole. When drilling is terminated, if there is no water flow in the hole, the temperature in the hole gradually approaches the in situ, equilibrium temperature of the rock. The rate of return to equilibrium depends on the duration and intensity of the disturbance, on the diameter of the hole, and on the thermal properties of the rock and of the material in the hole. Because boreholes on land frequently cannot be kept open sufficiently long for equilibrium temperatures to be reached, this problem has been studied extensively (Bullard, 1947; Lachenbruch and Brewer, 1959; Cooper and Jones, 1959; Cheremenski, 1960; Jaeger, 1961; Oxburgh et al., 1972).

Temperature measurements in the drill pipe show that during drilling by *Glomar Challenger*, the hole is maintained at a temperature very close to that of the deep ocean. There is sufficient heat exchange through the drill pipe in the seawater column to bring the drilling water to the external deep-sea temperature, and flow rates are sufficiently high that this temperature is not significantly changed in the borehole; Jaeger (1961) shows this theoretically. The heat generated at the bit may be important in diamond drilling of small holes, but it can be neglected in *Glomar Challenger* holes because of the high drilling water flow rates.

When the drilling fluid circulation ceases, the temperature disturbance gradually dissipates by radial conduction, the fluid in the hole warming toward the original rock temperature. Two simple idealizations are possible for the subsequent behavior of the fluid in the borehole: (1) There is sufficient small-scale convection or turbulence for the water column to have uniform temperature across its cross-section. The water can be considered to be a perfect conductor with known heat capacity. The slow settling of cuttings in the hole may facilitate such local mixing, as will high lateral temperature gradients; (2) The fluid in the hole is completely stable with no mixing or convection. In this case, the fluid in the hole can be considered as a solid of known thermal properties that may be different from those of the rock. This approximation is appropriate if there is any viscous drilling mud or a slurry of settled fine drill cuttings, particularly from the sedimentary part of the hole. The difference between the two approximations is negligible for "long" disturbance times of more than several hours, for which most of the heat of disturbance is contained in the rock, but is important for "short" disturbance times for which most of the disturbance heat is contained in the fluid in the hole. The limit of the latter case is the instantaneous emplacement of the fluid of a different temperature to that of the rock, the fluid then being allowed to warm undisturbed.

All of the temperature measurements in the *Glomar Challenger* holes were sufficiently far from the bottom of the hole for end effects to be neglected. Also, the drilling rates generally are much faster than the rate of heat conduction in the rock so cylindrical symmetry can be assumed. Some error is involved in assuming a constant hole diameter equal to that of the drill bit because the holes probably frequently are enlarged in zones of fractured basalt.

Mathematical Solutions

A) The simplest solution of the drilling disturbance problem represents the drilling operation by a line source of heat in an infinite medium (Bullard, 1947; Lachenbruch and Brewer, 1959; Cheremenski, 1960). This solution neglects the effect of the presence of the borehole filled with fluid of different thermal properties to that of the rock, during the decay of the disturbance. The magnitude of the line source of heat is not known, but is taken to be constant over the disturbance period with intensity equal to that required to give the observed, or estimated temperatures in the hole at the finish of drilling. As discussed above, in *Glomar Challenger* holes, constant internal hole temperature during the drilling is a better approximation than a constant heat supply. However, for long times after the disturbance period, the form of the decay of the disturbance is very similar for both approximations, constant hole temperature, or constant disturbance heat. For this model the temperature disturbance decays approximately as (e.g., Jaeger, 1965):

$$\frac{T}{T_o} = \frac{\ln\,(1 + \mathrm{t_o}/t)}{\ln\,(4k\mathrm{t_o}/a^2 - 0.577)}$$

where:

T_o is the temperature on the inner surface of the hole at the end of the time of disturbance,

t_o is the duration of the disturbance,

t is the time since the disturbance ceased,
k is the diffusivity,
a is the radius of the hole.

The thermal properties of the fluid in the hole and the surrounding rock are assumed to be the same and it is assumed that temperatures are measured at the surface of the hole, i.e., radius a. No mixing of the fluid in the hole is implied (see comments above).

B) The minimum disturbance decay time for the drill hole temperatures is given by the instantaneous emplacement of the cold, deep-sea temperature, drilling fluid into the hole with no circulation. Bullard (1954) discussed the theory involved, in connection with the dissipation of the frictional heating of ocean bottom sediment gradient probes, and Cooper and Jones (1959) use the theory for the extrapolation of the temperatures measured near the bottom of boreholes where the disturbance times were very short. The same theory is appropriate for the extrapolation of transient temperatures measured with the *Glomar Challenger* sediment temperature probes (e.g., Erickson et al., 1975). The solution assumes that the fluid in the borehole is a perfect conductor, i.e., well-mixed laterally (see comments above). The temperature of the fluid after a time t is given by (Carslaw and Jaeger, 1959):

$$\frac{T}{T_o} = \frac{4}{\pi^2} \int_0^\infty e^{-ktu^2/a^2} \frac{du}{u\Delta u}$$

where:

T_o is the temperature of the fluid at emplacement

$$\propto = \frac{2\,(\rho C)_{\text{rock}}}{(\rho C)_{\text{fluid}}} \quad \text{i.e. twice the ratio of rock to fluid heat capacity}$$

$$\Delta u = \left[uJ_o(u) - \propto J_1(u)\right]^2 + \left[uY_o(u) - \propto Y_1(u)\right]^2$$

and J_o, J_1, Y_o, Y_1 are Bessel functions.

This approach permits the incorporation of different heat capacities for the fluid and the surrounding rock, but is valid only for short disturbance times, times for which at the termination of the disturbance, the disturbance heat in the rock is much less than in the fluid in the hole. In the *Glomar Challenger* holes, this theory is a good approximation for disturbance times up to about 30 min. Von Herzen et al. (1971) discuss this model for *Glomar Challenger* holes drilled rapidly into soft sediments. Carslaw and Jaeger (1959) provide plots giving the decay of the disturbance.

C) The most complete theory for the decay of drilling disturbance has been given by Jaeger (1956, 1961). It is valid both for short and for long drilling disturbance times. Its only serious failure is that it assumes the thermal properties of the fluid to be the same as those of the rock during the decay of the disturbance, however, the theory probably could be extended to include different thermal properties. Thus, the theory may be a poorer approximation than the instantaneous emplacement model above for very short disturbance times. The theory assumes that the hole is held at constant temperature T_o for a time t_o, the drilling disturbance time, then allowed to warm undisturbed. The fluid in the hole is assumed to be stable, i.e., has finite thermal conductivity (see comments above). The temperature at a time t after a disturbance of duration t_o at the center of the hole $r=0$, is (Jaeger, 1956):

$$\frac{T}{T_o} = \left[1 - e^{-1/(4n\tau_o)}\right] + \frac{1}{2n\tau_o} \int_1^\infty \left(e^{-R^2/4n\tau_o}\right) f(R)\, dR$$

where

$$f(R) = 1 - \frac{2}{\pi} \int_0^\infty e^{-\tau_o u^2} \frac{C_o(u, Ru)\, du}{u\left[J_o^2(u) + Y_o^2(u)\right]}$$

and $C_o(u, Ru) = J_o(u)\, Y_o(Ru) - Y_o(u)\, J_o(Ru)$ are Bessel functions

$$\tau = \frac{kt}{a^2} \quad R = \frac{r}{a} \quad n = \frac{t}{t_o}$$

Physically, this solution represents the maintenance of the cylindrical surface of the hole at constant temperature for a time t_o, producing a radial temperature distribution $f(R)$, then with the hole filled with material all initially at temperature t_o, allowing the disturbance to decay. Jaeger (1961) has provided plots of T/T_o as a function of n, the number of disturbance times since the end of the disturbance, for various values of $\tau_o = kt_o/a^2$. The first term in the expression for T corresponds to the effect of the heat stored in the fluid in the hole. It can be neglected for long times since the disturbance ceased, e.g., $n\tau_o > 25$. The second term corresponds to the effect of the heat stored in the rock. It can be neglected for very short times, e.g., $n\tau_o < 0.025$.

D) In addition to these analytical solutions the problem of the decay of the drilling temperature disturbance can be solved readily numerically. However, unless the nature of the thermal disturbance or the physical geometry is very complicated and can be specified very precisely, it is doubtful if the decay of the disturbance could be computed more exactly numerically than by Jaeger's (1956) theory (C above).

In the analysis of Leg 37 temperature measurements only Jaeger's theory, C above, has been used.

LEG 37 TEMPERATURE GRADIENT AND CONDUCTIVITY MEASUREMENTS IN BASEMENT

Temperature Measurements in Basement

Temperatures were measured in the basement parts of three holes, 332B, 334, and 335. In Hole 332B, temperatures were measured immediately after a bit change and upon re-entry when the hole was at 551 meters depth. The last drilling circulation was 67 hr before the latter measurement. The regular sediment temperature instrumentation was used to log temperatures in the drill pipe. Temperature first was recorded for 15 min with the sensor extending through the bit and the bit 10 meters above the hole bottom. Then recordings were made inside the drill pipe for 2 min each, at 20-meter intervals from the bottom (Figure 1).

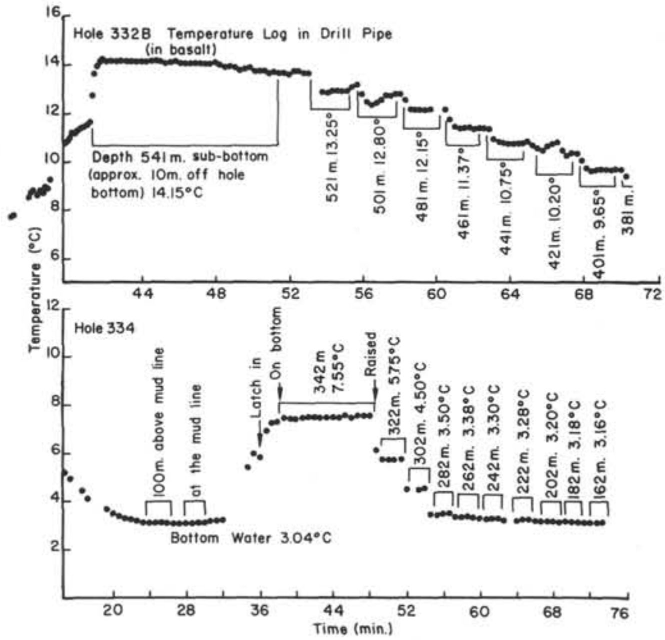

Figure 1. *Records of temperature versus time for basement parts of holes. The instrument was kept stationary for 2 min at each depth. The interpreted temperatures also are given.*

The temperatures in the drill pipe form a nearly linear gradient, but one that is considerably greater than the in situ gradient. The drilling disturbance time and thus temperature disturbance increase roughly linearly up the hole. Thus, these temperatures may be extrapolated to the bottom of the hole, giving an approximate temperature of 14.6°C. The theory of Jaeger (1956, 1961) (see above section for discussion) has been used to extrapolate the measured temperatures to in situ equilibrium values. The disturbances remaining for each measurement were taken from the plot of Jaeger (1961), which gives limited but sufficient accuracy. The parameters used in the computation and the results are summarized in Tables 1 and 2. The temperature during the disturbance period was taken to be that of the deep ocean water. The duration of the disturbance and the time since the disturbance were taken from the drilling log. The duration of the disturbance was taken to equal the time interval between when the hole reached the measurement depth and the termination of drilling, minus the times during the interval, when there was no circulation e.g., the period when core was being brought to the surface. Again, this is a rough approximation sufficient for the data, that could be improved. The results are not greatly different if the non-circulation times are not subtracted. The diffusivity of the rock was taken to be 6.0×10^{-7} m^2 sec^{-1}. The theory requires that the diffusity of the material that fills the hole be the same as surrounding rock. Fortunately, the combination of low diffusivity drilling water and cuttings and high diffusivity drill pipe probably gives an effective diffusivity not too different from the value taken for the rock. The thermal effect of the hole also becomes negligible several hours after the disturbance. The diameter of the hole was taken to be 40 cm. The drill bit has a diameter of about 25 cm but this hole probably is significantly larger, because of the fractured porous formation, particularly in rubble zones. The hole diameter is a major uncertainty in the computation, and variations in hole diameter may be the major source of the scatter in computed equilibrium temperatures. Repeated temperature measurements would permit removal of most of this uncertainty. The estimates of accuracy are subjective. The computed equilibrium temperatures extrapolate well to the estimated temperature at the sediment-basalt contact from the sediment temperature measurements. The resulting gradient in the basement and an estimated error are 18.3 ±0.7. The temperature at the bottom of the hole at 551 meters is 14.9°C, very close to the 14.6°C estimated from the uncorrected temperatures.

At Site 334, temperatures were measured when the hole depth was 348 meters during a brief stop in drilling. The first measurement was to 250 meters depth in the hole, where the core barrel containing the instrument jammed in the pipe, 1 hr after drilling circulation had stopped and the second at 20-meter intervals to 342 meters depth, 3 hr after circulation (Figure 1). The measurement on the first lowering was 100 meters above the bottom and still in sediment. It showed only a very small decay of the drilling disturbance, the temperature being about 0.3°C above the ocean bottom water temperature, or less than a 5% decay of the drilling disturbance. The second lowering reached to 5 meters from the hole bottom. The temperatures are almost identical to those of the first lowering at 250 meters depth, but the return to equilibrium increases rapidly toward the bottom of the hole. The more rapid decay of the disturbance below 250 meters comes mainly from the shorter disturbance time and partly because of the higher diffusivity of the basement rocks compared to the sediments. At 5 meters from the bottom, about 30% of the disturbance had decayed. This amount of disturbance decay, with only one measurement, permits only a very rough estimation of equilibrium temperatures, and that only for near the bottom of the hole. The extrapolation was carried out as for Hole 332B except that a hole diameter of 35 rather than 40 cm was used since the rocks are less fractured. The resulting temperatures agree well with the gradient defined in the sediment, but the accuracy is not sufficient to estimate a gradient in the basement rocks alone.

The results from this hole show that useful hole bottom temperatures can be obtained after only 2 or 3 hr from last circulation if temperatures are recorded continuously or at 2 or 3 times during that period, and if

TABLE 1
Histories of Borehold Drilling Disturbance for Holes 332B and Site 334 Prior to Basement Temperature Measurements

Depth (m)	Time Bit Reached Depth	Time of End of Circulation	Measurement Time	Disturbance Interval (h)	Non Circ. Time (h)	Disturbance Duration (t_o) (h)	Time Since Disturbance (h)	$\frac{\kappa t_o}{a^2}$	n	Fraction of Disturbance Remaining
Hole 332B										
401	16:00 June 26	16:00 June 29	14:10 July 2	72	41	31	70	1.67	2.26	0.29
421	22:00 June 26		14:10	66	40	26	70	1.40	2.69	0.27
441	10:00 June 28		14:10	30.0	6	24	70	1.30	2.92	0.25
461	17:30 June 28		14:05	22.6	5	17.5	70	0.95	4.00	0.24
481	01:00 June 29		14:05	15.0	4	11.0	70	0.59	6.36	0.18
501	05:30 June 29		14:00	10.5	3	7.5	70	0.41	9.33	0.14
521	10:00 June 29		14:00	6.0	2	4.0	70	0.22	17.5	0.10
541	15:00 June 29		13:55	1.0	0	1.0	70	0.054	70.0	0.06
Site 334										
302	08:15 July 16	18:00 July 16	21:35 July 16	9.7	2.5	7.2	3.6	0.51	0.50	0.87
332	11:45 July 16		21:35	6.2	1.5	4.7	3.6	0.35	0.77	0.78
342	17:00 July 16		21:30	1.0	0	1.0	3.6	0.071	3.50	0.64

TABLE 2
Correction to Basement Borehole Temperature for Drilling Disturbance, Hole 332B and Site 334

	Hole 332B								Site 334		
Depth in hole (m)	401	421	441	461	481	501	521	541	302	332	342
Duration of disturbance (h)	33	26	24	17.5	11.0	7.5	4.0	1.0	7.0	5.0	1.0
Time since end of disturbance (h)	70	70	67	67	67	67	67	67	3.0	3.0	3.0
Disturbance decay (%)	71	73	76	76	82	86	90	94	12.5	21.5	36
Disturbance temperature	4.45	4.45	4.45	4.45	4.45	4.45	4.45	4.45	3.0	3.0	3.0
Measured temperature (°C)	9.65	10.20	10.75	11.37	12.15	12.80	13.25	14.15	4.5	5.75	7.55
Computed equilibrium temperature (°C)	11.8	13.2	12.9	13.6	13.8	14.2	14.2	14.8	15.0	15.8	15.6
Estimated accuracy	0.5	0.5	0.4	0.4	0.3	0.3	0.2	0.2	4.0	3.0	2.0

Note: Diffusivity (k), 0.006 cm^2 sec^{-1}; hole diameter ($2a$); 332B = 40 cm; 334 = 35 cm.

the sensor is held as close as possible to the bottom. Ideally, the hole should be cleared thoroughly when drilling is stopped, then temperatures measured with the drill bit very near or on the bottom and the sensor extending through or just inside the bit.

At Site 335, three successive temperature logs at 20 meters depth intervals were measured in an attempt to define the decay of the drilling disturbance after short times. They were approximately 1, 2, and 3 hr after drilling circulation was terminated. In each case the complete hole, even in the sediments, was at virtually constant temperature, slightly above, but within 0.5°C of the temperature of the deep ocean water (Figure 1). In fact there was a small but significant decrease in temperature with time, the successive measurements each more closely approaching the bottom ocean water temperature. The only way that this surprising result can be explained is by ocean water flowing down the hole and escaping into a permeable horizon at the bottom. This explanation is supported by the drilling record, which indicates a very soft easily drilled zone near the bottom of the hole. At the very bottom of the hole, drilling was terminated in a very hard formation from which no core was recovered. It is difficult to guess what would drive such a downward flow, even if the water is able to escape at the bottom of the hole. The flow rates must be quite high, probably at least 1 m/min downhole, to maintain constant temperature. One interesting possibility is that a pressure deficiency is produced by a thermally driven hydrothermal upwelling in the surrounding topographic highs (e.g., Lister, 1972). The recharge then would be in the valleys wherever the water flow is not blocked by low permeability sediments.

Thermal Conductivity of Basement Samples

The thermal conductivity of 21 basement samples, 19 basalts, 1 gabbro, and 1 serpentinite, were measured using a divided bar apparatus using solid brass bars with constant temperature ends (e.g., Beck, 1957). See Jolivet (1966), for a detailed description of the instrument used. Calibration was with quartz and fused silica using the values of Ratcliffe (1959). The mean sample temperature was 21°C, slightly higher than the in situ temperatures.

Three discs 1, 3, and 6 mm thick were measured from each sample to compute the contact resistance and to provide a better average conductivity. A total of 63 discs thus was measured. The faces of the discs were silver-plated and polished and a pressure of 75 kg cm^{-2} applied to minimize the contact resistance between the brass bars and the samples. A correction was applied for a difference between the diameter of the divided bars and the discs (Jaeger and Beck, 1955). The samples all were water saturated before measurement. Corrections to in situ temperature and pressure should be negligible. The accuracy of the measurements is better than ±5%. The values are given in Table 3.

The mean thermal conductivity of 19 basalt samples, comprising 57 discs, from the holes, mainly 332B, is 3.97 ±0.04 (1.66 ±0.02) with no obvious systematic variation with depth (see Hole 332B), or between holes. This compares with the mean of 3.70 (1.55) for six basalt samples from a DSDP Leg 26 hole into 100-m.y. old sea floor in the eastern Indian Ocean (Hyndman et al., 1974a), a mean of 3.9 (1.63) for 16 basalts from very shallow drill holes, and a mean of 4.2 (1.76) for 11 dredged basalts all from the Mid-Atlantic Ridge near 45°N (Hyndman and Jessop, 1971). The Leg 26 rocks were subjected to surface weathering and the conductivity increases with depth. An average value of 4.0 (1.67) is thus reasonable for young sea floor layer 2 basalts. It should be pointed out, however, that only solid basalts are recovered by DSDP drilling. The remainder of the upper layer 2 section frequently has very extensive voids, rubble, and some sediment. The amount of solid basalt, such as that measured, is estimated to average 50% in the upper 500 meters of Hole 332B (see Hyndman, this volume). The remainder will have a conductivity of about 3.0 (1.3), e.g., a mix of water, 1.5 (0.63); consolidated carbonate sediment, 3.0 (1.3); and basalt, 4.0 (1.7). Thus a reasonable estimate for the mean conductivity and error of the upper 500 meters of Hole 332B is 3.5 ±0.3 (1.5 ±0.1) approaching 4.0 (1.7) at greater depths. In regions of the sea floor where layer 2a is not present, e.g., Site 335, the mean

TABLE 3
Thermal Conductivity of Basement Rocks

Sample (Interval in cm)	Depth Into Basement (m)	Rock Type	Thermal Conductivity mcal $cm^{-1}sec^{-1}{}^{\circ}C^{-1}$ ($Wm^{-1}K^{-1}$)
Hole 332A			
21-1, 131-134	143.8	Basalt	4.04 (1.69)
22-1, 37-40	152.4	Basalt	3.92 (1.64)
29-1, 74-77	219.2	Basalt	4.04 (1.69)
33-2, 92-95	258.9	Basalt	4.11 (1.72)
37-1, 116-119	295.7	Basalt	3.96 (1.66)
Mean			4.01 ±0.03 (1.68 ±0.01)
Hole 332B			
2-2, 86-89	11.9	Basalt	4.04 (1.69)
2-6, 129-132	18.3	Basalt	3.99 (1.67)
3-4, 10-13	62.6	Basalt	4.42 (1.85)
9-2, 104-106	192.6	Basalt	3.96 (1.66)
14-2, 81-83	249.3	Basalt	3.70 (1.55)
25-4, 47-49	356.5	Basalt	3.99 (1.67)
26-2, 7-10	362.6	Basalt	4.06 (1.70)
28-2, 23-26	381.7	Basalt	3.92 (1.64)
33-1, 62-65	426.9	Basalt	4.08 (1.71)
36-5, 40-42	462.4	Basalt	4.01 (1.68)
43-2, 94-96	525.0	Basalt	3.58 (1.50)
Mean			3.98 ±0.06 (1.67 ±0.03)
Site 334			
21-1, 39-41	50.9	Gabbro	4.71 (2.39)
26-2, 19-22	99.7	Serpentinite	7.61 (3.19)
Site 335			
5-2, 36-38	1.0	Basalt	4.06 (1.70)
9-2, 74-76	33.0	Basalt	3.65 (1.53)
10-5, 101-103	48.5	Basalt	3.94 (1.65)
Mean			3.88 ±0.12 (1.63 ±0.05)
Mean all 19 basalts			3.97 ±0.04 (1.66 ±0.02)

conductivity should be close to 4.0 (1.7), throughout basaltic layer 2. One gabbro from Site 334 has a conductivity of 5.7 (2.4). This is somewhat higher than some previously measured gabbros, e.g., 4.60 (1.9) (Birch and Clarke, 1940). One serpentinized peridotite has a conductivity of 7.6 (3.2). The conductivity of the latter is higher than most serpentinites, e.g., two dredged samples from 45°N have a mean 4.3 (1.8) (Hyndman and Jessop, 1971), reflecting the low degree of serpentinization of the Site 334 sample.

MEASUREMENTS IN SEDIMENTS

Successful heat-flow measurements have been made in over 20 deep sediment holes on previous Deep Sea Drilling Project legs, and the technique has been described in detail (Erickson, 1974; Erickson et al., 1975). Temperature-depth profiles are obtained by forcing a sensor that is attached to and extends below the drill bit into the undisturbed sediments below the bottom of the hole, at intervals during drilling. The temperature of the probe thermistor is recorded in a self-contained instrument in the core barrel. Forcing the probe into the sediment produces frictional heating which gradually decays (e.g., Figures 2-5). The temperatures can be extrapolated to equilibrium using the theory outlined by Bullard (1954) or Jaeger (1956). If the probe is undisturbed for at least several minutes, the "long time" asymptote may be used. The disturbance then decays with 1/time (Figure 6). The temperature resolution and accuracy of relative temperatures are about 0.02°C, and the absolute temperatures are accurate to 0.1°C. Temperatures can be obtained only at intermediate depths in the holes. At shallow depths the sediments frequently are too soft to support the drill bit assembly so the sensor cannot be kept stationary in the sediments. At great depths the sediments are too consolidated to permit penetration of the sensor probe. A maximum of four temperature points have been obtained in any one previous DSDP hole.

On Leg 37, all measurements in the sediments were made with the temperature sensor locked into the core barrel. Temperatures were obtained in all five of the holes drilled. The temperature-time records are given in Figures 2-5, the extrapolation to equilibrium in Figure 6 and the temperature-depth profiles in Figures 7-10 and Table 4.

The thermal conductivity of the sediments penetrated by the holes is measured in the laboratory onboard ship using the transient needle probe technique (Von Herzen and Maxwell, 1959) on the cores recovered. The cores are allowed to equilibrate at laboratory temperature for at least 4 hr before measurement. Any residual temperature drift is assumed to be linear over the measurement period of 5 min and is removed by least-square analysis. The results are corrected to in situ temperature and pressure using the factors given by Ratcliffe (1960) (see also Erickson, 1974). The accuracy of measurement is about 5%. The measured thermal conductivities generally increase systematically with depth in the holes because of compaction. Carbonate

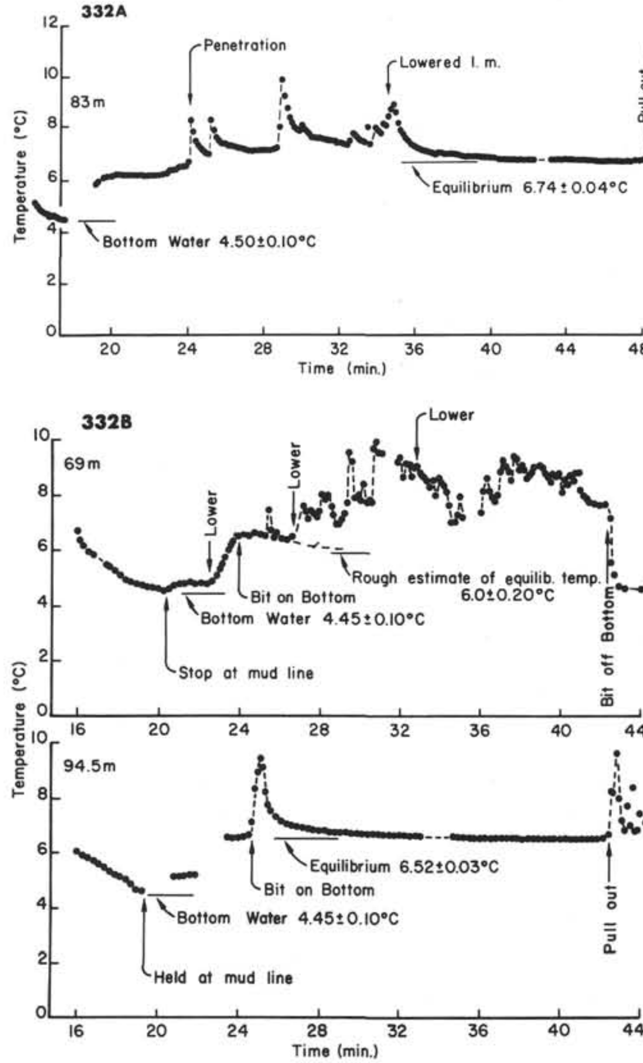

Figure 2. *Records of sediment temperature versus time, Site 332.*

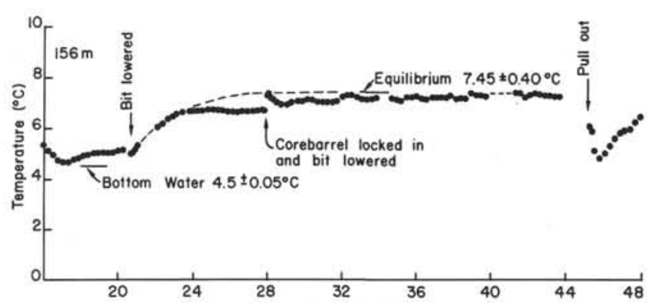

Figure 3. *Records of sediment temperature versus time, Site 333.*

sediments have significantly higher conductivity than terrigenous sediments, with average conductivity 2.65 +0.002 per meter of hole depth (1.11 +0.0008) (e.g., Hyndman et al., 1974a). The values measured on Leg 37 are in general agreement with this relation. The measured thermal conductivities are given in Figure 11 and in Table 5.

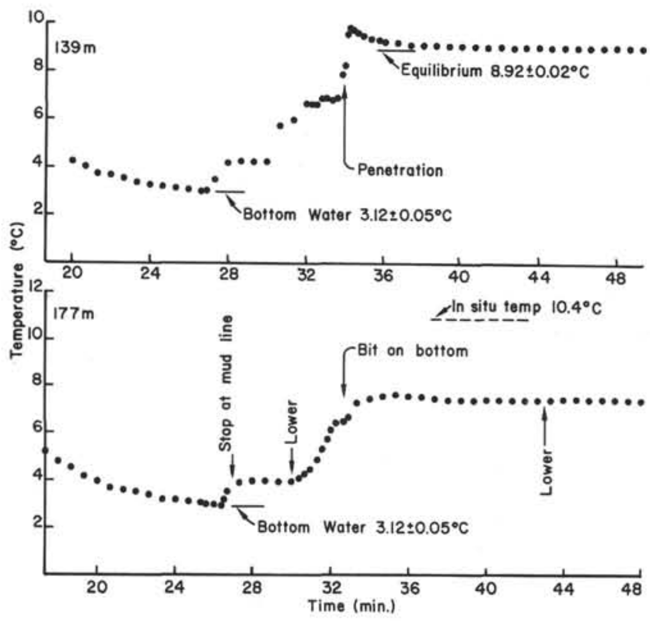

Figure 4. *Records of sediment temperature versus time, Site 334.*

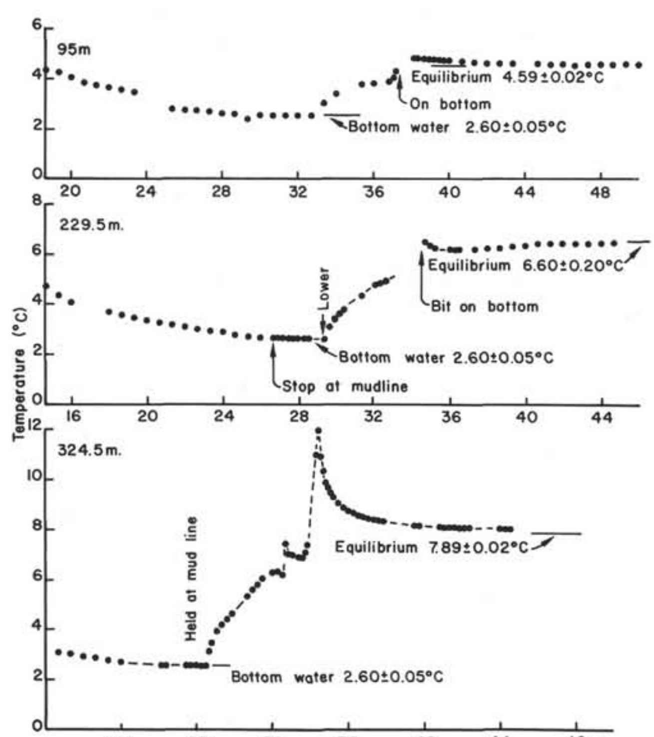

Figure 5. *Records of sediment temperature versus time, Site 335.*

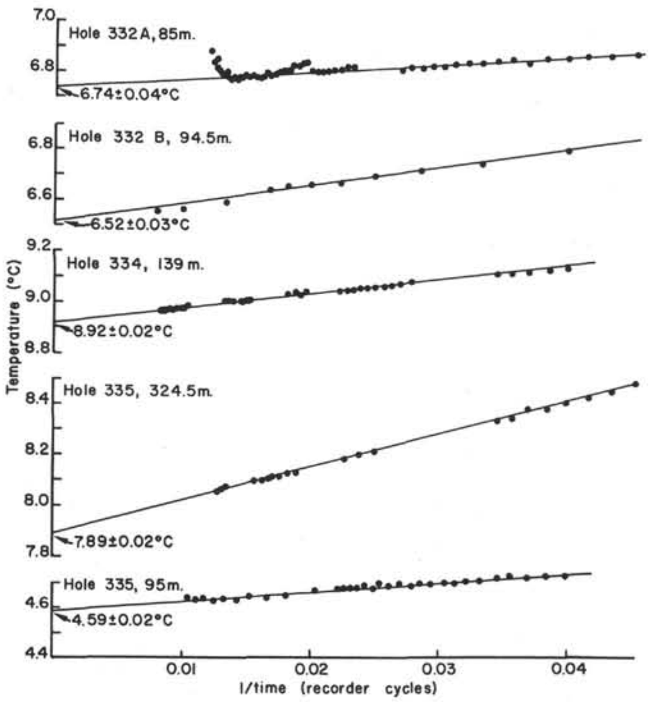

Figure 6. *Extrapolation to equilibrium of recorded sediment temperatures.*

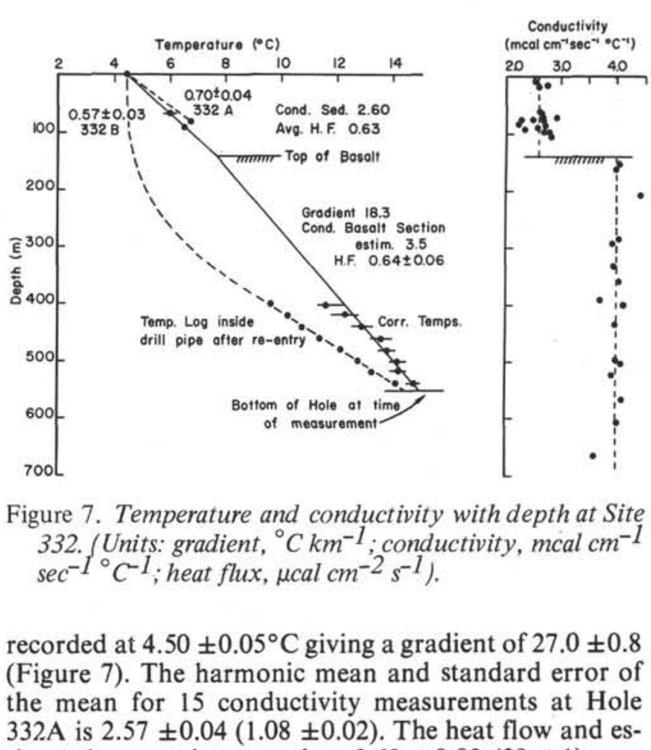

Figure 7. *Temperature and conductivity with depth at Site 332. (Units: gradient, °C km^{-1}; conductivity, mcal cm^{-1} sec^{-1} $°C^{-1}$; heat flux, µcal cm^{-2} s^{-1}).*

LEG 37 HEAT-FLOW RESULTS

Hole 332A

One good sediment temperature of 6.74 ±0.04°C was measured at 83 meters depth in Hole 332A (Figure 2). The ocean bottom water temperature also was well recorded at 4.50 ±0.05°C giving a gradient of 27.0 ±0.8 (Figure 7). The harmonic mean and standard error of the mean for 15 conductivity measurements at Hole 332A is 2.57 ±0.04 (1.08 ±0.02). The heat flow and estimated uncertainty are then 0.69 ±0.03 (29 ±1).

Hole 332B

Two sediment temperatures were measured in Hole 332B (Figure 2) one at 69 meters giving a rough estimate of 6.0 ±0.2°C and the second at 94.5 meters giving a well-determined value of 6.52 ±0.03. The bottom water temperature is 4.50 ±0.05°C. The three values

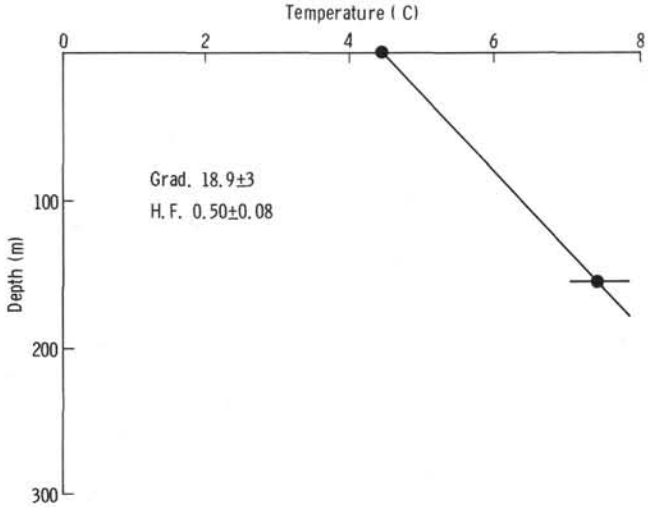

Figure 8. *Temperature with depth at Site 333.*

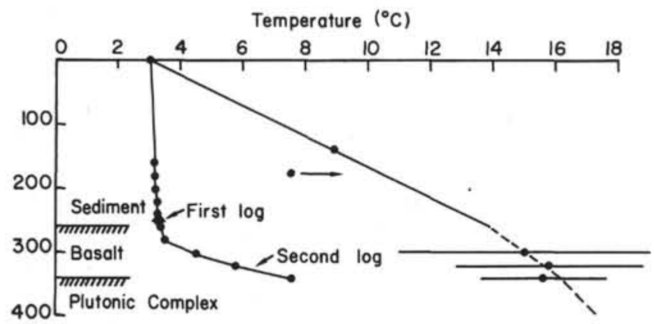

Figure 9. *Temperature with depth at Site 334.*

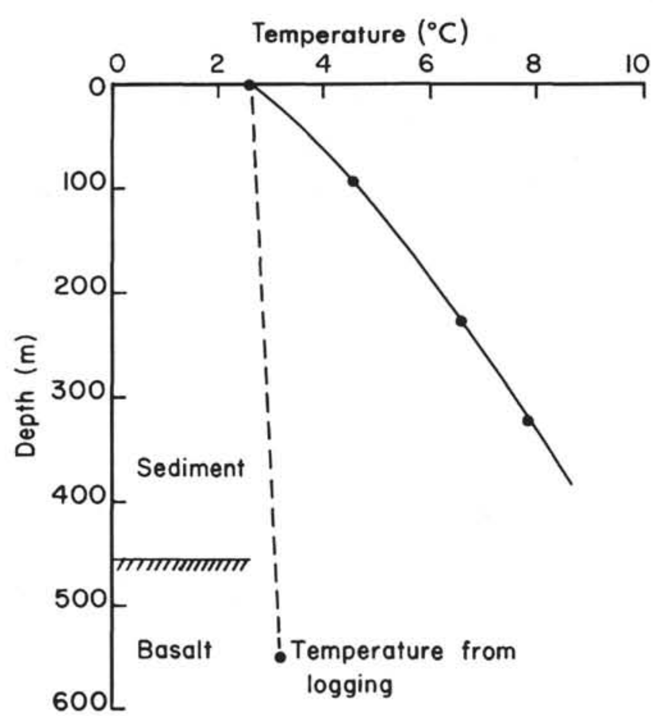

Figure 10. *Temperature with depth at Site 335.*

form a very linear gradient of 21.5 ±0.3 (Figure 7). The thermal conductivity estimate of 2.57 ±0.04 (1.08 ±0.02) gives a heat flux of 0.55 ±0.02 (23 ±1).

Reliable temperatures were obtained in the basement part of the hole between 401 and 541 meters depth (see section above) (Figure 1). This depth range is too small for an accurate gradient estimate but least-squares value of 20.6 ±1.8 agrees well with the more precise value of 18.3 ±0.7 obtained also using the extrapolated temperature at the sediment-basalt interface of 7.56°C at 142 meters from the sediment measurements (Figure 7). The previously estimated basement (see above) conductivity of 3.5 ±0.3 (1.5 ±0.1) gives a heat flux and error estimate for the basalt part of the hole of 0.64 ±0.06 (27 ±3). This value is within the error limits of less than ±10% of the heat fluxes for the sedimentary part of this hole and of the adjacent Hole 332A. It is actually equal to the mean of the heat-flows in sediments for the two holes. Thus, there is a uniform heat flux within ±10% to a depth of 541 meters, 400 meters into basement, at this site.

Hole 333A

One temperature measurement was made in the sediment at 156 meters depth in Hole 333A (Figure 3). The record is very disturbed probably because there was some difficulty in getting the core barrel and instrument locked into the bit. We have taken the maximum temperature reached several times on the later part of the record of 7.45 ±0.40°C as the best estimate at 156 meters and 4.50 ±0.05°C as the bottom water temperature giving a gradient of 18.9 ±3.0 (Figure 8). The mean conductivity is 2.90 ±0.04 (1.21 ±0.01) for nine measurements. The heat flux is then 0.55 ±0.09 (23 ±4).

Site 334

Two sediment temperatures were attempted at Site 334 (Figure 4), one at 139 meters giving a reliable value of 8.92 ±0.02°C and a second at 177 meters that apparently did not penetrate undisturbed sediment. The good value with a bottom-water temperature estimate from two lowerings of the instrument in the drill pipe of 3.12 ±0.05°C gives a gradient 41.7 ±0.4 (Figure 9). The thermal conductivity mean for 28 measurements is 2.60 ±0.06 (1.08 ±0.03). The heat flux is then 1.08 ±0.04 (45 ±2).

The temperatures measured in the basement part of the hole (Figure 1) are discussed above. Very approximate equilibrium temperatures, to ±2°C, were obtained for the bottom part of the hole. They are in good agreement with the gradient defined in the sediment, but are of too low accuracy and are over too small a depth range to define an independent gradient in the basement part of the hole. The mean conductivity for the upper 60 meters basalt portion of basement is estimated to be about 4.0 (1.7) and for the underlying plutonic complex about 6.0 (2.5). Thus, the illustrated gradients of 29 and 19 in Figure 9.

TABLE 4
Summary of Leg 37 Heat-Flow Data

Site	Location N	Location W	Crustal Age (m.y.)	Water Depth (m)	Subbottom Depth (m)	Temperature (°C)	Gradient (°C km^{-1})	Conductivity mcal cm^{-1} sec^{-1} °C^{-1} (W m^{-1} K^{-1})	Heat Flux μcal cm^{-2} sec^{-1} (mW m^{-2})
332A	36°52.7′	33°38.5′	3.5	1818	0 83	4.50 ±0.05 6.74 ±0.04	27.0 ±0.8	2.57 ± 0.04 (1.08 ± 0.02)	0.69 ± 0.03 (29 ± 1)
332B	36°52.8′	33°38.6′	3.5	1806	0 69 94.5	4.50 ±0.05 6.00 ±0.20 6.52 ±0.03	21.5 ±0.3	2.57 ± 0.04 (1.08 ± 0.02)	0.55 ± 0.02 (23 ± 1)
					401-541	(Basement)	18.3 ±0.7	3.5 ± 0.3 (1.5 ± 0.1)	0.64 ± 0.06 (27 ± 3)
333	36°50.5′	33°40.1′	3.5	1666	0 156	4.50 ±0.05 7.45 ±0.40	18.9 ±3.0	2.90 ± 0.04 (1.21 ± 0.01)	0.55 ± 0.09 (23 ± 4)
334	37°02.1′	34°24.9′	8.9	2632	0 139 177 302-342	3.12 ±0.05 8.92 ±0.02 >7.5 (Basement)	41.7 ±0.4 ~25 ~25	2.60 ± 0.06 (1.08 ± 0.04) ~5.0 (2.1)	1.08 ± 0.04 (45 ± 2) ~1 (50)
335	37°17.7′	35°11.9′	~13	3198	0 95 229.5 324	2.60 ±0.05 4.59 ±0.02 6.60 ±0.20 7.89 ±0.02	16.1 ±0.01	3.03 ± 0.08 (1.26 ± 0.03)	0.49 ± 0.02 (21 ± 1)

Site 335

Three sediment temperature measurements were attempted at Site 335. (Figure 5). The results are: 4.59 ±0.02 at 95 meters; 6.60 ±0.20 at 229.5 meters; 7.89 ±0.02 at 324 meters depth. The record at 229.5 meters has a warming toward equilibrium so the temperature is uncertain. The two other values are well determined (Figure 10). The bottom water temperature was well recorded in the drill pipe at 2.60 ±0.05. The temperatures define a gradient that increases smoothly with depth. The least-squares gradient is 16.1 ±0.1. Taking the mean conductivity of 3.03 ±0.08 (1.26 ±0.03) for eight measurements gives a heat flow of 0.49 ±0.02 (21 ±1). Constant heat flux with depth implies that conductivity increases at a rate of +0.003 (0.0013) per meter depth, which is not apparent in the measured values. The rate also is slightly more rapid than the 0.002 (0.008) suggested by Hyndman et al. (1974).

As described above, temperatures were measured in the basement part of the hole three successive times after the completion of drilling (Figure 1). All recorded nearly constant temperature in the hole which is interpreted as resulting from seawater flowing down the hole into a permeable horizon at the bottom. Equilibrium temperatures in the rock cannot be predicted.

The heat flow results from Leg 37 are shown with nearby shallow oceanographic probe values in Figure 12.

DISCUSSION

The most important geothermal result of DSDP Leg 37 is the constancy and low value of heat flow 0.6 ±0.1 (25 ±4) for three holes across one sediment pond on 3.5-m.y. old sea floor and to a depth of 400 meters into the basaltic basement beneath the pond. The results for Hole 332B represent the first heat-flow measurement into normal oceanic basement, layer 2. The data for the pond are summarized in Figure 13. Also shown are three measurements of heat flux in the pond using a standard 2.5-meter long oceanographic probe (Lewis, 1975; Table 6), which gave almost the same heat flow as the drill holes. Thus, our data show that the heat flow measured at the surface represents not only that deep in the sediment, but also that at least 400 meters into the underlying basement. This pond is only one example, but it is very near a spreading ridge crest where there are very large variations in the values measured by the standard oceanographic probe (e.g., Williams, 1974) and where also all measured values are much less than those predicted by theoretical models.

To show that the low flow into the sediment pond cannot arise through thermal refraction by low conductivity sediments or by the effect of topography, the thermal field of the region has been computed using a two-dimensional finite difference solution for steady-state heat conduction (Figure 13). The deviation from an assumed regional flux of 0.7 (29) is shown for the surface, e.g., oceanographic probe, and at 500 meters depth, e.g., deep borehole. The variation in flux is small, less than 0.1 (4) except at the surface near the steep west side of the pond. The simple approximation of a hemi-ellipsoidal body of sediment (Von Herzen and Uyeda, 1963) indicates that the flux in the pond may be less than the regional value, but the reduction is not more than 10%. The numerical solution shows that the effect of the topography is of similar magnitude and of opposite sign.

The heat flow of 0.6 (25) is very low compared to the world average of 1.5 (63) and especially compared to the values of about 6.4 (269) predicted by theoretical models for 3.5-m.y. old ocean floor. However, as discussed in the introduction, such low values are very common near ridge crests and are ascribed to heat transport by hydrothermal circulation. Important questions now are: over what time period or distance

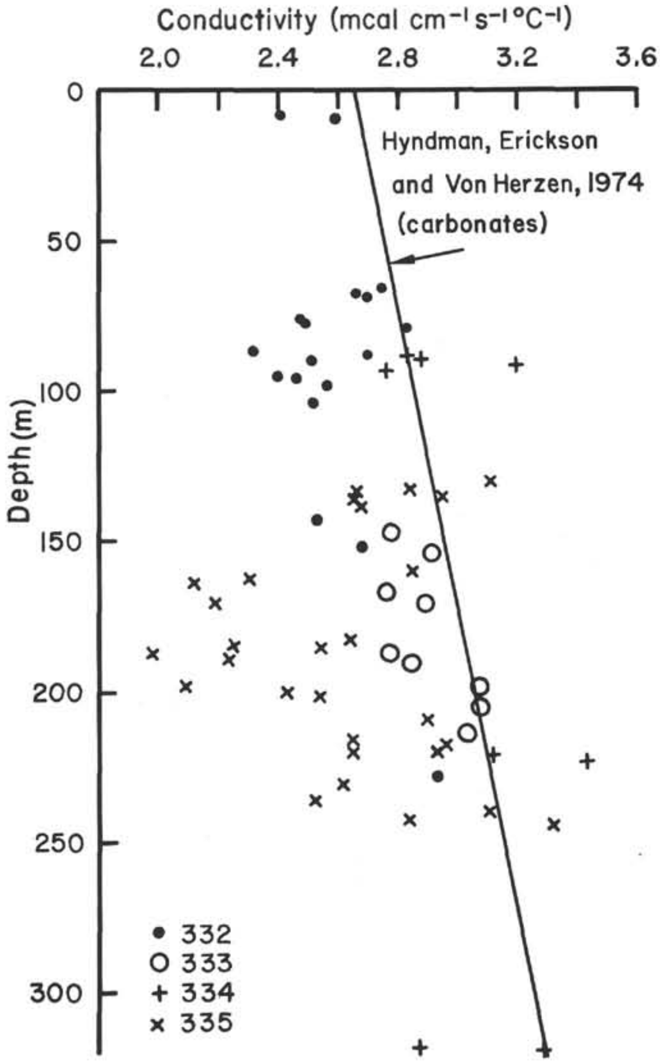

Figure 11. *Thermal conductivity of Leg 37 sediments.*

from the axis of spreading does the circulation persist, and to what depth does it extend. Lister (1971, 1972), Davis and Lister (1974) and Williams et al. (1974) suggested that circulation and cooling may extend to the base of the crust although it seemed reasonable to expect most of the circulation to be concentrated in the fractured and porous upper part of layer 2. Water circulation probably is excluded from the mantle by the expansion associated with serpentinization that will fill and close any cracks and fissures. However, Hart (1973) suggests that the low temperature alteration products in the crust will have high permeability and will facilitate further circulation.

Having at the same time such a low and such a uniform heat flow is a puzzle. The low values require circulation to have continued at least to within 0.5 m.y. of the present since the thermal time constant of the crust is only about 0.5 m.y. But the nearly constant heat flux from six measurements over a horizontal distance of 3 km and to a depth of 540 meters requires that, either the circulation stopped a long time ago so that small-scale temperature variations have been smoothed out or that the circulation has a very deep and regular character with large horizontal scale. We suggest that the circulation is concentrated in the lower part of the crust, perhaps in layer 3, but at least in the lower part of layer 2. This possibility, contrary to first expectation, can be explained readily. The upper part of layer 2 primarily consists of many thin successive flows, each one cooled before the subsequent one arrives. Fractures and cooling cracks in a flow tend to be sealed off by the following flows. In contrast, the deeper crust is produced in a region that is relatively hot and that cools slowly. Seawater penetrating to this hot region through fractures some time after its formation will produce rapid cooling and cracking from thermal contraction. The process of a cracking front in such a hot region penetrated by seawater has been discussed in detail by Lister (1974). The schematic diagram, using the observed bathymetry, of Figure 14 illustrates the proposed type of water flow in this region. The flow directions undoubtedly are highly variable but there may be a systematic pattern of sources to the west and discharge to the east toward the axis of accretion where temperatures and elevations are higher. It is important to note that for the observed heat flow the crust must be maintained by water flow at very low temperatures, e.g., 15°C at 1 km and as low as 75°C at 5 km depth.

Lister (1972), Williams et al. (1974), and Williams (1974) showed that high heat flows on ridges generally occur on topographic highs and frequently but not always near steep scarps. However, for the Galapagos Ridge the horizontal wavelength between heat-flow highs or lows is about 6 km and does not seem to be controlled by the topography which has shorter characteristic wavelengths. The regular heat-flow pattern found by Williams et al. (1974) on the Galapagos spreading center suggests a penetration of circulation to 3 to 4 km depth if the permeability is uniform with only the detailed location of the upwelling and downwelling being controlled by topography and perhaps fractures.

Bodvarsson and Lowell (1972) however suggest, based on studies in Iceland, that the permeability of ridge crest crustal rocks is highly variable. The horizontal permeability in the upper crust is furnished mainly by tubular and irregular openings at the contacts of the lava flows, while the vertical permeability is provided primarily by open spaces between and within dikes, and by deep vertical fractures. Extensive fractures must occur in the crust from thermal contraction as the lithosphere cools moving away from the spreading center. They show that low heat flows such as we observed can be produced by a horizontal crack or layer only a few millimeters thick connected to the surface by vertical fractures.

The heat flow of 1.08 (45) at Site 334 is higher than the values near the ridge crest but still below the theoretical value of about 4.0 (167) for 9-m.y. old ocean floor. Thus, hydrothermal circulation must have been important in this region although it may not be still active at present.

The heat flow of 0.49 (21) measured at Site 335 also is much below the theoretical value for 11 to 13 m.y. old ocean floor. Hydrothermal circulation must be still active in the region of this hole. The flow of seawater down the hole into a horizon 100 meters into the base-

TABLE 5
Thermal Conductivity of Leg 37 Sediments

Sample (Interval in cm)	Subbottom Depth (m)	Thermal Conductivity (mcal $cm^{-1}sec^{-1}°C^{-1}$) Measured	Corrected
Hole 332A			
1-1, 115	8.1	2.54	2.41
1-2, 88	9.4	2.74	2.59
2-2, 85	66.4	2.88	2.74
2-3, 85	67.8	2.79	2.66
2-4, 84	69.3	2.84	2.53
3-2, 75	75.7	2.59	2.47
3-3, 75	77.2	2.62	2.49
3-4, 75	78.7	2.97	2.83
4-3, 75	86.7	2.43	2.32
4-4, 75	88.2	2.83	2.70
4-5, 75	89.7	2.63	2.51
5-2, 75	94.7	2.52	2.40
5-3, 75	96.2	2.58	2.46
5-4, 75	97.7	2.68	2.56
6-2, 75	104.2	2.64	2.52
Hole 332B			
1-1, 75	142.7	2.65	2.53
2-1, 75	152.2	2.80	2.68
4-1, 75	228.2	3.05	2.93
Harmonic mean = 2.57 ±0.04			
Site 333			
2-2, 75	147.7	2.91	2.78
2-6, 75	153.7	3.05	2.91
3-2, 75	166.7	2.89	2.76
3-5, 75	171.2	3.02	2.89
4-3, 75	187.2	2.89	2.77
4-5, 75	190.2	2.98	2.85
5-4, 75	198.2	3.21	3.07
6-2, 95	204.9	3.21	3.08
7-2, 75	214.2	3.16	3.03
Harmonic mean = 2.90 ±0.04			
Site 334			
2-1, 75	130.2	3.23	3.11
2-2, 75	131.7	2.95	2.84
2-3, 75	133.2	2.76	2.66
2-4, 75	134.7	3.06	2.95
2-5, 75	136.2	2.75	2.65
2-6, 75	137.7	2.78	2.68
5-2, 75	160.2	2.95	2.85
5-3, 75	161.7	2.39	2.31
5-4, 75	163.2	2.19	2.12
6-2, 75	169.7	2.27	2.19
7-4, 75	182.2	2.73	2.64
7-5, 75	183.7	2.32	2.25
7-6, 75	185.2	2.63	2.55
8- , 75	187.2	2.04	1.98
8-2, 75	188.7	2.30	2.23
9-2, 75	198.2	2.16	2.09
9-3, 75	199.7	2.50	2.43
9-4, 75	201.2	2.62	2.54
10-3, 75	209.2	2.99	2.90
11-1, 75	215.7	2.73	2.65
11-2, 75	217.2	3.05	2.96
11-3, 75	218.7	3.01	2.93
11-4, 75	220.2	2.73	2.65
12-4, 75	229.7	2.69	2.62
13-2, 75	236.2	2.59	2.52
13-4, 75	239.2	3.19	3.11
13-6, 75	242.2	2.91	2.84
14-1, 100	244.5	3.40	3.32
Harmonic mean = 2.60 ±0.06			

TABLE 5 – *Continued*

Sample (Interval in cm)	Subbottom Depth (m)	Thermal Conductivity Measured	($mcal\ cm^{-1}sec^{-1}°C^{-1}$) Corrected
Site 335			
1-1, 100	88.0	2.97	2.83
1-2, 75	89.2	3.02	2.88
1-3, 75	90.7	3.29	3.19
1-4, 75	92.9	2.89	2.76
3-1, 30	220.3	3.57	3.43
3-1, 120	221.2	3.25	3.12
4-3, 30	318.7	2.99	2.87
4-3, 120	319.2	3.41	3.29
Harmonic mean = 3.03 ±0.08			

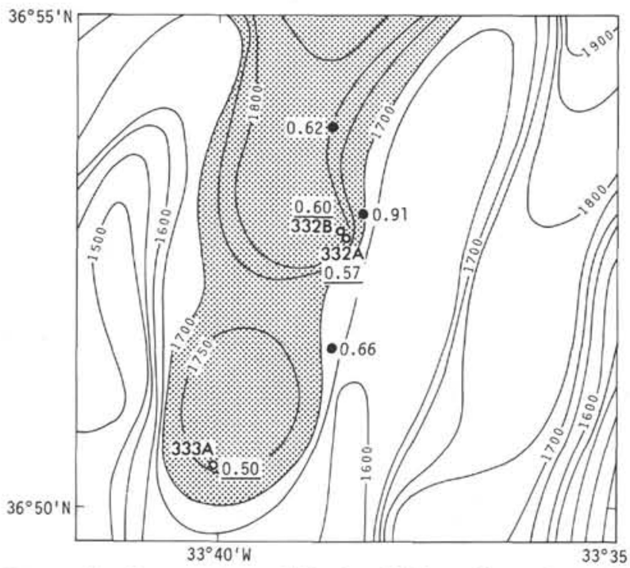

Figure 12. *Comparison of the Leg 37 heat flow values with nearby shallow ocean probe data from Lewis (1975). The hatched area is approximate extent of the sediment pond.*

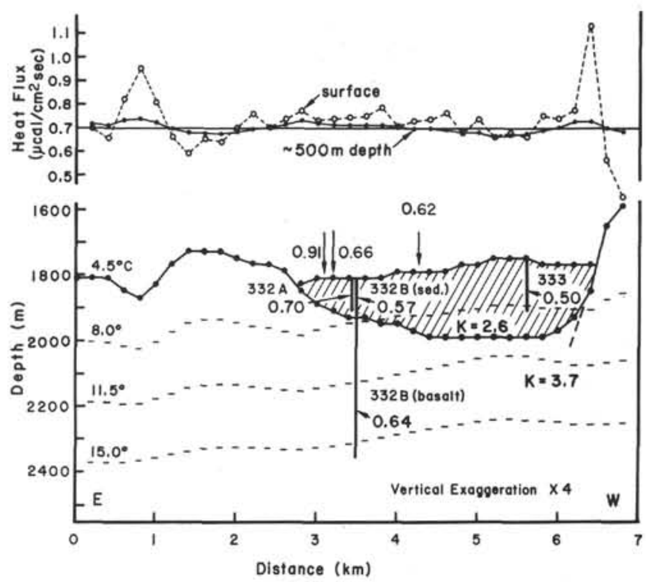

Figure 13. *Cross-section of sediment pond penetrated by Holes 332A, 332B, and 333 on 3.5-m.y. old sea floor. The heat flows measured in the holes along with three shallow probe values from Lewis (1975) are shown. The computed thermal refraction effect of the sediment pond and topography on heat flow is shown in the upper graph and the effect on deep temperature by the isotherms on the cross-section.*

ment that was deduced from the temperature measurements supports such a circulation. The observed surface heat flow indicates that there may be water flowing in this horizon at about 10°C.

ACKNOWLEDGMENTS

We wish to express our appreciation to Mr. A. Porter, Heat Flow and Electronics Technician onboard D.V. *Glomar Challenger* for his very able efforts in obtaining downhole temperatures. This work was done while one of us (R.D. Hyndman) was at Department of Oceanography, Dalhousie University, Halifax, N.S., Canada.

REFERENCES

Anderson, R.N., 1972. Petrologic significance of low heat flow on the flanks of slow-spreading mid-ocean ridges: Geol. Soc. Am. Bull., v. 83, p. 2947-2956.

Anderson, R.N. and Halunen, J., Jr., 1974. Implications of heat flow for metallogenesis in the Bauer Deep: Nature, v. 251, p. 473-475.

Aumento, F., Loncarevic, B.D., and Ross, D.I., 1971. Hudson geotraverse: Geology of the Mid-Atlantic Ridge at

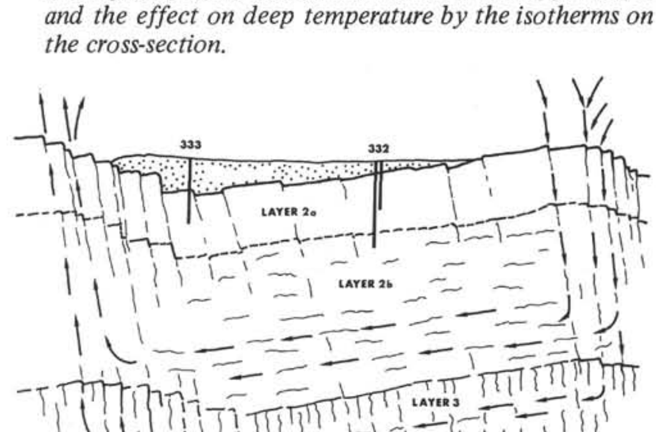

Figure 14. *Schematic diagram of possible hydrothermal circulation in the region of Sites 332 and 333.*

45°N: Phil. Trans. Roy. Soc. London, Sci. A. v. 268, p. 623-650.

Beck, A.E., 1957. A steady state method for the rapid measurement of the thermal conductivity of rocks: J. Sci. Instrum., v. 34, p. 186-189.

Birch, F. and Clark, H., 1940. The thermal conductivity of rocks and its dependence upon temperature and composition: Am. J. Sci., v. 238, p. 529-558, 613-635.

Bodvarsson, G. and Lowell, R.P., 1972. Ocean-floor heat flow and the circulation of interstitial waters: J. Geophys. Res., v. 77, p. 4472-4475.

Bottinga, Y., 1974. Thermal aspects of sea-floor spreading, and the nature of the suboceanic lithosphere: Tectonophysics, v. 21, p. 15-38.

Bullard, E.C., 1947. The time necessary for a borehole to attain temperature equilibrium: Mon. Not. Roy. Astr. Soc., Geophys. Suppl., v. 5, p. 127-130.

_______, 1954. The flow of heat through the floor of the Atlantic Ocean: Proc. Roy. Soc. London, Ser. A, v. 222, p. 408-429.

Carslaw, H.S. and Jaeger, J.C., 1959. Conduction of heat in solids, 2nd ed.: New York (Oxford University Press).

Cheremenski, S.A., 1960. Time of re-establishing the thermal conditions disturbed by drilling a borehole: Izv. Akad. Nauk SSSR, Ser. Geofiz., v. 1960, p. 1801-1805.

Cooper, L.R. and Jones, C., 1959. The determination of virgin strata temperatures from observations in deep survey boreholes: Geophys. J., v. 2, p. 116-131.

Corliss, J.B., 1971. The origin of metal-bearing submarine hydrothermal solution: J. Geophys. Res., v. 76, p. 8128-8138.

Cronan, D.S., van Andel, T.H., Heath, G.R., Dinkelman, M.G., Bennett, R.H., Bukry, D., Charleston, S., Kaneps, A., Rodolfo, K.S., and Yeats, R.S., 1971. Iron rich basal sediments from the eastern equatorial Pacific: Leg 16, Deep Sea Drilling Project: Science, v. 175, p. 61-63.

Davis, E.E. and Lister, C.R.B., 1974. Fundamentals of ridge crest topography: Earth Planet. Sci. Lett., v. 21, p. 405-413.

Dymond, J., Corliss, J.B., Heath, G.R., Field, C.W., Dasch, E.J., and Vech, H.H., 1973. Origin of metalliferous sediments from the Pacific Ocean: Geol. Soc. Am. Bull., v. 84, p. 3355-3372.

Erickson, A.J., 1974. Initial report on downhole temperature and shipboard thermal conductivity measurements, Leg 19. *In* Creager, J.S., Scholl, D.W., et al., Initial Reports of the Deep Sea Drilling Project, Volume 19: Washington (U.S. Government Printing Office), p. 643-656.

Erickson, A.J., Von Herzen, R.P., Sclater, J.G., Girdler, R.W., Marshall, B.V., and Hyndman, R.D., 1975. Geothermal measurements in deep-sea drill holes: J. Geophys. Res., v. 80, p. 2515-2528.

Hart, R.A., 1973. A model for chemical exchange in the basalt-seawater system of oceanic layer II: Canadian J. Earth Sci., v. 10, p. 799-816.

Hekinian, R., 1971. Chemical and mineralogical differences between abyssal hill basalts and ridge tholeiites in the eastern Pacific Ocean: Mar. Geol., v. 11, p. 77-91.

Hyndman, R.D. and Jessop, A.M., 1971. The Mid-Atlantic Ridge near 45°N. XV Thermal conductivity of dredge and drill core samples: Canadian J. Earth Sci., v. 8, p. 391-393.

Hyndman, R.D. and Rankin, D.S., 1972. The Mid-Atlantic Ridge near 45°N. XVIII. Heat flow measurements: Canadian J. Earth Sci., v. 11, p. 809-818.

Hyndman, R.D., Erickson, A.J., and Von Herzen, R.P., 1974a. Geothermal measurements on DSDP Leg 26. *In* Davies, T.A., Luyendyk, B.P., et al., Initial Reports of the Deep Sea Drilling Project, Volume 26, Washington (U.S. Government Printing Office), p. 451-563.

Hyndman, R.D., Muecke, G.K., and Aumento, F., 1974b. Deep Drill 1972. Heat flow and heat production in Bermuda: Canadian J. Earth Sci., v. 11, p. 809-818.

Jaeger, J.C., 1956. Numerical values for the temperature in radial heat flow: J. Math. Phys., v. 34, p. 316-321.

_______, 1961. The effect of the drilling fluid on temperatures measured in boreholes: J. Geophys. Res., v. 66, p. 563-569.

_______, 1965. Application of the theory of heat conduction to geothermal measurements. *In* Lee, W.H.K. (Ed.), Terrestrial heat flow: Am. Geophys. Un. Monogr. Ser. no. 8, p. 7-23.

Jaeger, J.C. and Beck, A.E., 1955. The calculation of heat flow through disks and its application to conductivity measurements: Brit. J. Appl. Phys., v. 6, p. 15-16.

Jolivet, J., 1966. Mesure de la conductibilité thermique des roches au laboratoire: Cahier IPG No. 19, Inst. Phys. Globe, Univ. Paris.

Lachenbruch, A.H. and Brewer, M.C., 1959. Dissipation of the temperature effect in drilling a well in Arctic Alaska: U.S. Geol. Surv. Bull. 1083-C, p. 73-109.

Langseth, M.G., Jr., and Von Herzen, R.P., 1971. Heat flow through the floor of the worlds oceans: *In* Maxwell, A.E. (Ed.), The sea, v. 4, pt. 1: New York (Wiley-Interscience), p. 299-352.

Langseth, M.G., Jr., Le Pichon, X., and Ewing, M., 1966. Crustal structure of the mid-ocean ridges, 5. Heat flow through the Atlantic Ocean floor and convection currents: J. Geophys. Res., v. 71, p. 5321-5355.

Lewis, J.F., 1975. Oceanic heat flow measurements over the continental margins of Eastern Canada: Unpublished M.Sc. Thesis, Dalhousie University, Halifax, Canada.

Lister, C.R.B., 1971. Heat flow and sea-floor spreading off the coast: *In* Symposium on tectonism of the Pacific Northwest. *In* Moen, D. (Ed.), EOS: Trans. Am. Geophys. Union, v. 52, p. 628-630.

_______, 1972. On the thermal balance of a mid-ocean ridge: Geophys. J., v. 26, p. 515-535.

_______, 1974. On the penetration of water into hot rock: Geophys. J., v. 39, p. 465-509.

McKenzie, D.P., 1967. Some remarks on heat flow and gravity anomalies: J. Geophys. Res., v. 72, p. 6261-6273.

Muehlenbachs, K. and Clayton, R.N., 1971. The oxygen isotope geochemistry of submarine greenstones (abstract): EOS, Trans. Am. Geophys. Union, v. 52, p. 365.

Oxburgh, E.R., Richardson, S.W., Turcotte, D.L., and Hsui, A., 1972. Equilibrium borehole temperatures from observation of thermal transients during drilling: Earth Planet. Sci. Lett., v. 14, p. 47-49.

Palmason, G., 1967. On heat flow in Iceland in relation to the mid-Atlantic ridge. *In* Bjornsson, S. (Ed.), Iceland and Mid Ocean Ridges: Reykjavik (Soc. Sic. Islandica), p. 111-127.

_______, 1973. Kinematics and heat flow in a volcanic rift zone with application to Iceland: Geophys. J., v. 33, p. 451-481.

Parker, R.L. and Oldenberg, D.W., 1973. Thermal model of ocean ridges: Nature, v. 242, p. 137-139.

Ratcliffe, E.H., 1959. Thermal conductivities of fused and crystalline quartz: Brit. J. Appl. Phys., v. 10, p. 22-25.

_______, 1960. The thermal conductivities of ocean sediments: J. Geophys. Res., v. 65, p. 1535-1541.

Sclater, J.G. and Francheteau, J., 1970. The implications of terrestrial heat flow observations on current tectonic and geochemical models of the crust and upper mantle of the earth: Geophys. J., v. 20, p. 509-542.

Sclater, J.G., Von Herzen, R.P., Williams, D.L., Anderson, R.N., and Klitgord, K.D., 1974. The Galapagos spreading center: heat flow low on the north flank: Geophys. J., v. 38, p. 609-626.

Sleep, N.H., 1969. Sensitivity of heat flow and gravity to the mechanism of sea floor spreading: J. Geophys. Res., v. 74, p. 542-549.

Spooner, E.T.C. and Fyfe, W.S., 1973. Sub-sea-floor metamorphism, heat and mass transfer: Contrib. Mineral. Petrol., v. 42, p. 287-304.

Talwani, M., Windish, C.C., and Langseth, M.G., Jr., 1971. Reykjanes ridge crest: a detailed geophysical study: J. Geophys. Res., v. 76, p. 473-517.

Von Herzen, R.P. and Maxwell, A.E., 1959. The measurement of thermal conductivity of deep-sea sediments by a needle probe method: J. Geophys. Res., v. 64, p. 1557-1563.

———, 1964. Measurements of heat flow at the preliminary mohole site off Mexico: J. Geophys. Res., v. 69, p. 741-748.

Von Herzen, R.P. and Uyeda, S., 1963. Heat flow through the eastern Pacific Ocean floor: J. Geophys. Res., v. 68, p. 4219-4250.

Von Herzen, R.P., Fiske, R.J., and Sutton, G., 1971. Geothermal measurements on Leg 8. *In* Tracey, J.I., et al., Initial Reports of the Deep Sea Drilling Project, Volume 8: Washington (U.S. Government Printing Office), p. 837-849.

Williams, D.L., 1974. Heat loss and hydrothermal circulation due to sea-floor spreading: Unpubl. Ph.D. thesis, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts.

Williams, D.L., Von Herzen, R.P., Sclater, J.G., and Anderson, R.N., 1974. The Galapagos spreading center: lithospheric cooling and hydrothermal circulation: Geophys. J., v. 38, p. 587-608.