Geothermal measurements in five small lakes of northwest Ontario

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The heat flow through the floors of five small lakes of known thermal history on the Canadian Shield was measured with a modified Bullard probe. A small correction for seasonal bottom water temperature variations was applied to temperature gradient measurements, and the heat flows are corrected for glaciation, lateral temperature gradients, sedimentation rates, and lateral thermal conductivity changes. Four lakes have an average heat flow of 49 ± 4 mW/m² (1.2 ± 0.1 μcal/cm² s). A high heat flow in the fifth lake is thought due to unusual refraction effects. The heat generation – heat flux combination yields a point that falls near accepted lines for the Canadian Shield.

Le flux thermique à travers le fond de cinq petits lacs du Bouclier canadien dont l'histoire thermique est connue a été mesuré à l'aide d'une sonde Bullard modifiée. Une petite correction pour les variations saisonnières des températures de l'eau de fond a été appliquée aux mesures de gradient de température, et les flux thermiques ont été corrigés pour la glaciation, les gradients de température latéraux, les taux de sédimentation et les changements latéraux dans les valeurs de conductivité thermique. Quatre lacs ont un flux thermique moyen de 49 ± 4 mW/m² (1.2 ± 0.1 μcal/cm² s). Un flux thermique élevé dans le cinquième lac semble causé par des effets de réfraction inhabituels. La combinaison production de chaleur-flux thermique donne un point qui se situe près des courbes établies pour le Bouclier canadien.

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Introduction

Successful heat flow measurements have been made in many of the world's large, deep lakes using the standard oceanic probe technique. However, attempts in several small meromictic lakes in the east and northeast U.S.A., which also have relatively constant bottom water temperatures, have proved less satisfactory (Reitzel 1966; Diment 1967; Johnson and Likens 1967; Likens and Johnson 1969). Very large corrections are required for the temperature differences at their margins (the 'warm rim effect'), and such lakes have been considered unreliable for regional heat flow studies (Roy *et al.* 1972).

The Canadian Shield abounds in small- and moderate-sized lakes (< 5 km across, < 50 m deep). A mean annual temperature close to the bottom water temperature, particularly in northwest Ontario, causes the warm rim effect to be much less pronounced. The purpose of this study was to see whether some of these lakes could be used to obtain reliable heat flow values.

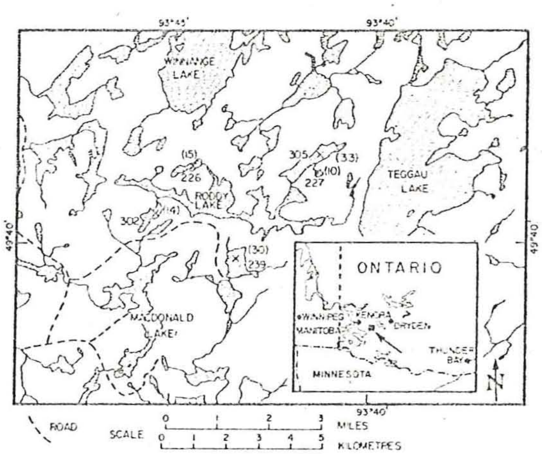


FIG. 1. Location of the lakes used in this study; black square on inset map identifies map-area. Figures are identification numbers of lakes assigned by Fisheries Research Board; numbers in parentheses are depths, in metres; crosses indicate measurement sites.

Five lakes in proximity to each other were selected in the Experimental Lakes Area of northwest Ontario (Fig. 1). Maximum depths ranged from 10 to 32 m, and temperature records kept by the Freshwater Institute (Fisheries Research Board of Canada) for the four years previous to these measurements showed that the annual variation of their bottom water temperature was less than 2 °C (Fig. 2). Measurements were made at two times of the year to verify that the correction to sediment temperatures for these annual variations gave consistent heat flows.

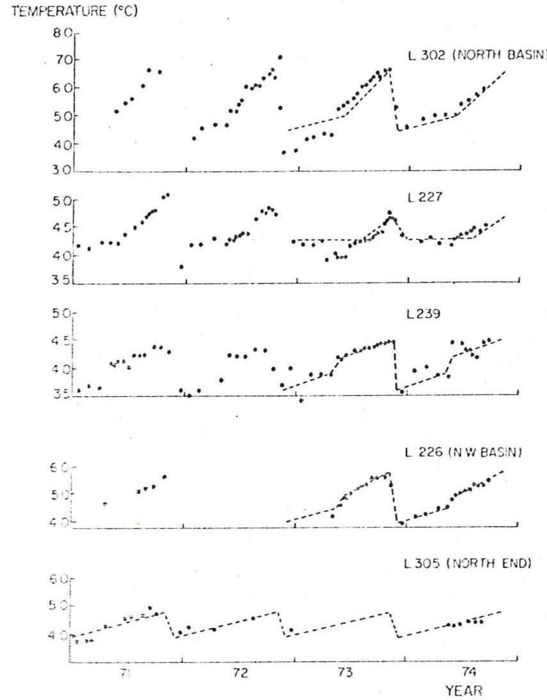


FIG. 2. Bottom water temperatures in the five lakes for four years. Broken line shows the thermal model assumed for each lake.

Probe

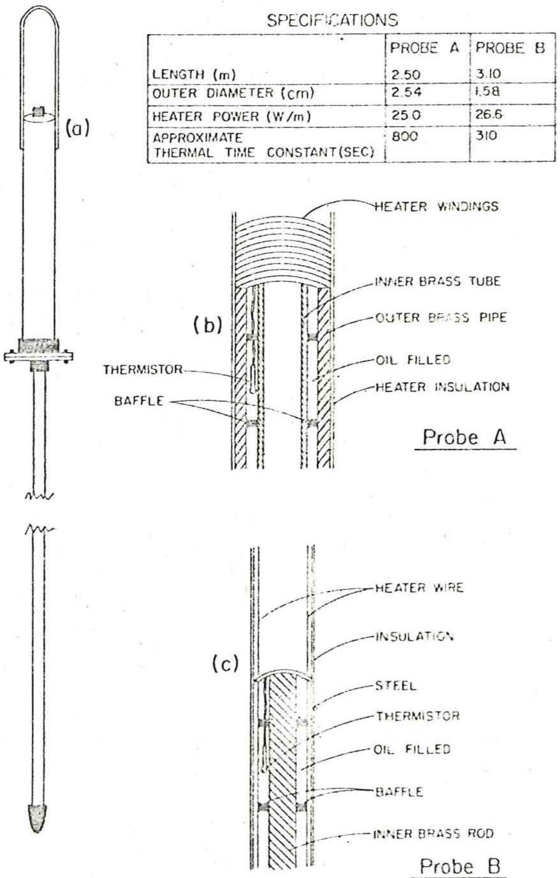
Two modified Bullard probes, designated probe A and probe B, were built for the work (Fig. 3). Temperatures were measured by six (probe A) or nine internally mounted thermistors. An external heater wire down the full length of the probes was designed to provide multiple *in situ* thermal conductivity determinations.

The thermistors were commercially available 4 kΩ Fenwall assemblies. These were monitored consecutively at the lake surface by a simple D.C. Wheatstone bridge capable of measuring 10 kΩ to $\pm \frac{1}{2}$Ω. Allowing for self-heating effects and calibration procedures, the relative accuracy was better than 0.005 °C.

The power source for the heater was a 12 V lead–acid battery connected to a 125 watt inverter with an output of 120 V.A.C. The power output from this system remained stable ($\pm$ 1%) for the requisite heating time of 30 min.

Measurements

All measurements were made in the deepest part of each lake basin. The probe was raised



	PROBE A	PROBE B
LENGTH (m)	2.50	3.10
OUTER DIAMETER (cm)	2.54	1.58
HEATER POWER (W/m)	25.0	26.6
APPROXIMATE THERMAL TIME CONSTANT (SEC)	800	310

FIG. 3. The two probes used in this study.

and lowered from a 4 m boat fitted with a portable derrick and manual winch. Sediment clinging to the top of the probe indicated complete penetration of the apparatus on all drops.

The thermistors were monitored for about an hour after the probe entered the sediments, by which time the probe was sufficiently close to thermal equilibrium to permit the necessary extrapolation to be made. The heater was then turned on for the conductivity determination, and the thermistors were monitored for another 30 min.

The theory for the thermal response of a heat flow probe in sediments was established for Bullard (1954) and extended by Jaeger (1956). Extrapolation of the cooling curves of the thermistors is necessary to obtain the equilibrium sediment temperatures. This was accomplished with the cylindrical plot method of Lister (1970), in which the time axis is scaled in such a way that

TABLE 1. Heat flow measurements and corrections

Lake	226	227	239	302	305
Temperature gradient* ($\times 10^{-4}$ K/m)					
Measured between:					
28 May and 3 June 1974	5.5	9.3	5.6; 6.0	4.2	1.3
12–16 September 1974	5.7	11.2	5.1	4.0	1.1
MEAN	5.6	10.3	5.6	4.1	1.2
Thermal conductivity (W/mK)	0.71	0.71	0.71	0.71	0.71
Uncorrected heat flow (mW/m^2)	40	73	40	29	85
Corrections (mW/m^2)					
Glaciation	8	8	8	8	8
Lateral temperature gradients	−19	−37	−8	−10	−12
Sedimentation	4	2	2	1	5
Lateral conductivity gradients	16	8	12	12	24
Corrected heat flow (mW/m^2)	49	54	54	40	110

*After correction for annual bottom-water temperature variation.

the theoretical cooling curve plots as a straight line. It was found that temperature readings for each thermistor, taken between five minutes and one hour after penetration, did define straight lines, and that the temperature intercept at infinite time could be determined within 0.005 °C.

All sediment temperature profiles showed some curvature with depth. This was completely removed when the effect of the annual variation of bottom water temperatures was calculated. In making this correction, the exact position of the thermistors relative to the sediment surface was obtained by shifting the measured profiles vertically, relative to the computed curves (*e.g.* Fig. 4) until the best fit was obtained. Temperature records for the lakes indicated that the annual variation in water temperature could be divided into four parts: the autumn circulation period, winter heating, spring circulation, and summer heating. The effect of spring circulation was usually very small, because atmospheric temperatures had usually risen to above 4 °C by the time of ice break-up. A characteristic 'saw-tooth' temperature variation curve for the bottom water temperature therefore resulted (Fig. 2). The correction to the temperature gradient in the sediments was obtained by computing the Fourier components of the adopted saw-tooth curve for each lake, and calculating, by the well known expression, the temperature–depth variation resulting from each harmonic. A typical set of correction curves is shown in Fig. 4. As Table 1 indicates, temperature gradients measured in the spring, and in late summer, after correction, were in very satisfactory agreement.

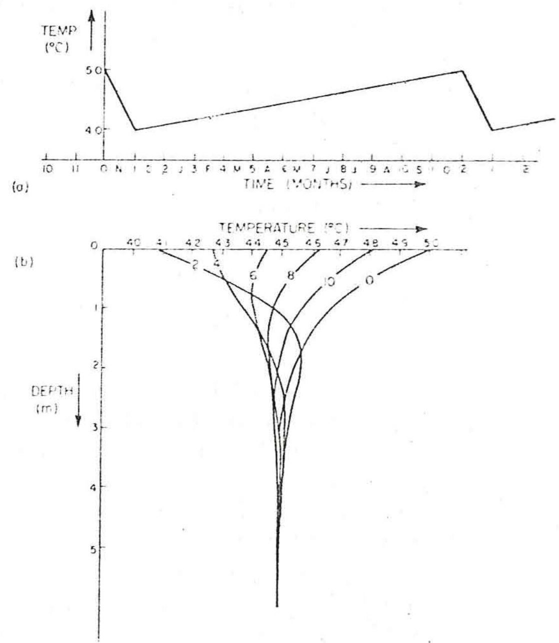


FIG. 4. For the annual temperature variation at the sediment-water interface shown in (a), the temperature–depth curves in the sediment at different times of the year are as shown in (b). (Sediment thermal diffusivity is taken to be 1.5×10^{-7} m^2/s.)

The theory for the heating of a cylindrical probe, required for the thermal conductivity measurements, has been developed in great detail by Blackwell (1953, 1954, 1956) and Jaeger (1956, 1958, 1959). The long-term solution is the basis of the familiar "needle probe method" (Von Herzen and Maxwell 1959) used in determining the thermal conductivity of sediment cores. *In situ* conductivity measurements have been made by Sclater *et al.* (1969), Christoffel and Calhaem (1969), and Lister (1970).

Strength requirements dictated relatively large diameters for the long probes used in this work, and therefore produced large thermal time constants (see probe specifications, Fig. 3). Even after 30 min of heating, the temperature change of probe A was 30%, and of probe B 12%, less than the change predicted by the long-term logarithmic asymptote. Consistent thermal conductivity values were obtained both with depth in any lake, and between the lakes. The 36 values from probe A lay within 0.75 ± 0.08 W/mK and the 63 values from probe B were within 0.71 ± 0.04 W/mK.

The mean water content of the surface sediments in these lakes is 94.3 ± 2.7% of wet weight (Brunskill *et al.* 1971), and Stockner (1971) found that the water content decreased from 90 to 80% over a 1.5 m core retrieved from one of the lakes. The empirical curves of Ratcliffe (1960) and Steinhart *et al.* (1969) indicate that sediment with a water content of 80% should have a thermal conductivity of 0.63 ± 0.02 W/mK. This is significantly less than the values measured by the probes.

A consequence of having a relatively short heating time of thermal conductivity probes is that the response of the thermistors is affected by thermal contact resistances between the sensors and the sediments. Initially it was thought that contact resistance between the heater and the sensors was causing the discrepancy in the conductivities. To avoid this problem, a heat flow probe of new design was built after the field work described in this paper had been completed. This apparatus has 3 outrigged thermal conductivity miniature probes that have very short time constants. Preliminary tests made in February 1976 in lakes of northwest Ontario similar to those of this study have yielded conductivities close to 0.74 W/mK, with the conductivity increasing by 7% of its surface value through the upper 3 m of sediments. This tends to confirm the conductivities obtained from probes A and B. A thermal conductivity of 0.71 ± 0.04 W/mK has therefore been adopted for the heat flow calculation.

Corrections to the Heat Flow

Several corrections must be applied to the local, uncorrected heat flows shown in Table 1 to obtain values representative of the regional flux. These are for the effects of glaciation, lateral temperature gradients, sedimentation, and lateral thermal conductivity gradients (Allis 1975).

The effect of glaciation was calculated using the method of Birch (1948) and Jessop (1971). Prest (1969) indicated that the final retreat of the ice sheet over the Experimental Lakes area occurred 12.5 thousand years ago. For earlier climatic variations the chronology of Jessop (1971) was used, with an ice base temperature of −1 °C assumed.

Both topography around a lake and the presence of the lake itself cause lateral temperature gradients in the sediments and country rock. The correction was calculated from Lachenbruch (1957), using data on the annual water temperature pattern in each lake, a mean annual soil temperature at lake level of 6.5 °C (based on 7-year records from Winnipeg and Atikokan), and the assumption that the lateral temperature gradients existed since the end of the last glaciation.

Cores from adjacent lakes indicate a two-stage sedimentation history (J. McAndrews, personal communication 1975). The upper 5–8 m of sediment is gyttja (organic ooze) that has accumulated since the last glaciation at 0.05 cm/y. Below is at least 3 m of clay-fill. Topography and bathymetry profiles suggest 5–15 m of these glacial sediments in the lake basins. The thermal effect of this sedimentation pattern was calculated by the method of Von Herzen and Uyeda (1963).

The relatively small correction for sedimentation (Table 1) shows that sedimentation rates are sufficiently low that the sediment body is very close to thermal equilibrium. Steady state conditions can therefore be assumed for the correction for lateral thermal conductivity gradients. The sediments were assumed to lie in oblate, hemispheroidal depressions, with a conductivity contrast of 3–1 at the bedrock contact. The appropriate expression of Von Herzen and Uyeda (1963) was used for the calculation.

Discussion

The accuracy of the temperature gradients is estimated to be 10%, the uncertainty being due to doubt over the exact penetration of the probe, and the relative accuracy of the temperature measurements. This probably accounts for the small differences in gradient measured in each lake at the two times of the year.

With the exception of Lake 305, the spread in the uncorrected heat flows is considerably reduced when the corrections are applied. The largest corrections, for lateral temperature and conductivity gradients, are opposite in sign. Uncertainty in both the shape of the bedrock–sediment interface and the conductivity contrast at the interface makes the latter correction the most poorly determined. This could explain the high heat flow in Lake 305. If the bedrock is not the ideal basin shape assumed, but is ridge-like, or if the sediments contain large erratics, then a significant reduction in the measured heat flow would be required. Thus, the high heat flow in this lake is not regarded as representative of the regional heat flow. The mean heat flow and probable error for the other four lakes is 49 ± 4 mW/m^2 (1.2 ± 0.1 μcal/cm^2 s).[1]

Radioelement concentrations in 44 outcrops around the lakes were determined with a portable gamma ray scintillation counter (McPhar TV5). Outcrops were all typical shield granite-gneiss. The concentrations varied considerably with outcrops, and sometimes varied by more than 100% within an outcrop. The deduced mean heat generation for the area is 1.3 μW/m^3 with a standard error of 0.1 μW/m^3 (3.2 ± 0.2 × 10^{-13} cal/cm^3 s).

In Fig. 5, the heat flow–heat generation measurements are compared with available geothermal data from the Canadian Shield (from Rao and Jessop 1975). As none of the points on the figure has a glacial correction, the heat flow of 41 ± 4 mW/m^2 (0.98 ± 0.10 μcal/cm^2 s) is plotted for the lakes. This point falls close to both lines I and II of Rao and Jessop (1975).

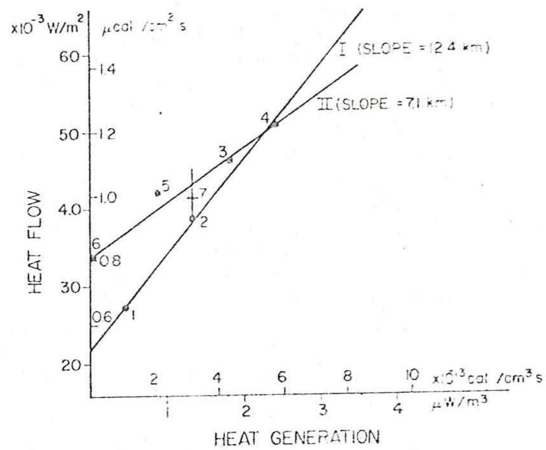


FIG. 5. Heat flow *vs.* heat generation for the Canadian Shield (after Rao and Jessop 1975). Points 1–3 are from Cermak and Jessop 1971; 4 and 6, from Roy *et al.* 1968; 5, from Sass *et al.* 1971; 7, this paper.

Conclusions

Small lakes on the Canadian Shield can be used to determine the regional terrestrial heat flow. An initial judicious choice of lakes is necessary so that the perturbation of the steady-state temperature gradient in the sediments by annual temperature variations of the bottom water is minimized. To obtain the regional flux, significant corrections are required for lateral temperature gradients that tend to increase heat flow through the deep central regions of a lake floor, and lateral thermal conductivity gradients that decrease the heat flow in the sediments.

Reliable *in situ* determinations of the sediment thermal conductivity are possible using a Bullard probe with a heat source over its full length. However, the necessarily large diameter of the probe requires both a long heating time in the sediment and a relatively large correction to obtain the long-term asymptote. For more accurate *in situ* conductivity measurements (error < 5%), small-diameter outrigger probes with short time constants are recommended.

Acknowledgments

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[1]The working units used throughout were those the authors consider most appropriate for geophysics, the cgs system. Values given in the SI system were derived from the cgs quantities.

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