Temperature Microstructure in Powell Lake

T. R. Osborn

Institute of Oceanography, University of British Columbia, Vancouver, B. C., Canada

(Manuscript received 4 October 1972, in revised form 23 January 1973)

ABSTRACT

Measurements of the vertical component of the temperature gradient in Powell Lake show phenomena similar to those observed in the ocean. The gradient is an irregular function of depth, with temperature inversions indicating static instability of the water column on the centimeter scale.

The lower portion of the lake contains old sea water. Doubly diffusive layers were found only near the very bottom of the lake and another form of thermohaline circulation may exist for 60 m above the layers.

1. Introduction

Powell Lake was a fjord; with the melting of the last ice age and rebound of the land, its outermost sill rose above sea level. The lake consists of a series of deep flat-bottomed basins separated by sills of various depth. Mathews (1962) studied the bathymetry of Powell Lake. Fig. 1 shows the outlines of the South Basin, the hatched areas representing the essentially flat basin floor. Williams *et al.* (1961) showed that the bottom of the lake contained old sea water with a low oxygen and high sulphide concentration. Fig. 2 shows their measured profiles of temperature, salinity and dissolved oxygen (U. B. C., 1962). Geological considerations indicate the sea water is around 11,000 years old (Mathews, personal communication).

The temperature microstructure in the lake is of interest for two reasons. The first is that measurements by Gregg and Cox (1972) have shown the ocean to be unstably stratified on the centimeter scale, while Simpson and Woods' (1970) study of Loch Ness revealed no temperature inversions and thus no density inversions present in that lake. It is of interest to see if microstructure in lakes is fundamentally different from that of the ocean. The second reason for study at Powell Lake is to look for microstructure in the salt water that is specifically related to the doubly diffusive convection described by Turner (1965).

2. Instrumentation

The instrument consists of a plastic shell 1 m in length containing two glass spheres 25 cm in diameter. The lower glass sphere is an instrument housing containing the electronics and battery supply. There are two thermistors: one mounted directly below the instrument body and the second $\frac{1}{2}$ m outboard but level with the first thermistor. The three wings on the instrument slow its descent and cause it to rotate at a frequency near $\frac{1}{3}$ Hz. Since the instrument rotates, the outboard thermistor cuts a helical path $\frac{1}{2}$ m in radius about the vertical path of the central thermistor.

The data are treated in a manner similar to that described by Osborn and Cox (1972). The data for calculating the temperature gradient are produced by passing the temperature signal through a high-pass filter with a 3-dB point at $\frac{1}{3}$ Hz and then amplifying 1600 times. Instead of recording internally, the signals are telemetered to the surface using a Sippican expendable wire length. Five signals are telemetered from the instrument to the surface: the pressure as measured by a vibrotron pressure gage, the temperatures from each thermistor, and two signals which correspond to the high-frequency part of the temperature signal for each thermistor. A simple convolution converts the high-frequency part of the temperature signals to the temperature gradient as a function of time as observed by each thermistor.

Depth and fallspeed are calculated from the vibrotron pressure gage. The error in depth is ±5% due largely to the temperature dependence of the vibrotron. Since absolute value of pressure is much more sensitive to temperature than the relative value, the velocity is known to ±2%. The fallspeed is about 25 cm sec^{-1} and is calculated from the data for each drop. The instrument descends until a Richardson-type pressure release drops the weight. A salt block is used as a back-up release and drops the weight after 30 min of immersion in water if the pressure release has failed.

3. Observations

Five profiles were taken in Powell Lake during the period 30 September to 2 October, 1971. Drop no. 2 on the morning of 1 October and drop no. 4 on the morning of 2 October have been analyzed in most

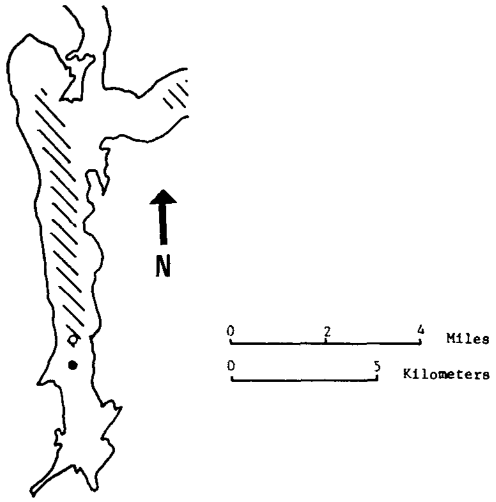

FIG. 1. Outline of the South Basin of Powell Lake, 100 km northwest of Vancouver, Canada. The hatched area indicates flat-bottomed basin floors. The location of drops no. 2 and no. 4 is shown by the open circle and the location of drop no. 5 by the solid circle.

detail. Drop 2 extends from the surface to 200 m depth while drop 4 extends from the surface to the bottom. The position of drops 2, 4 and 5 are shown in Fig. 1.

Fig. 3 shows the temperature gradient vs depth from drop 2 for the 20 m starting at a depth of 50 m. The curves represent a convolution of the data. The left-hand trace is the gradient as seen by the nose probe which travels straight down. The right-hand trace is from the outboard probe which descends along a helical path one revolution per 74 cm. The line of zero gradient was determined from the regions of low gradient between 65 and 72 m depth and by examining the original data record. If the lake were perfectly horizontally layered, the two traces would be identical. There are negative temperature gradients at depths of 50, 56.5, 62, 64, and 65.5 m. In fresh water above 4C these negative gradients mean a temperature increase with depth which would be statically unstable in a fresh water lake. These inversions appear in both traces showing their horizontal scale to be greater than 50 cm.

The region of intense thermal activity between 56 and 58 m represents a scene of recent stirring activity, perhaps a broken internal wave similar to those pictured by Woods and Fosberry (1967) in the Mediterranean. This 2 m thick region does not show the lateral similarity evident in the rest of the record. Examinations of

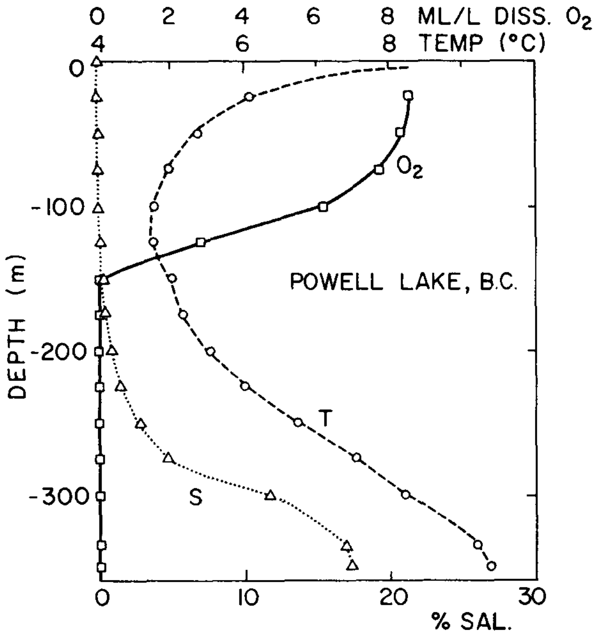

FIG. 2. Profiles of temperature, salinity and dissolved oxygen in Powell Lake as measured by Williams *et al.* (1961). Data from 1961 University of British Columbia Data Report.

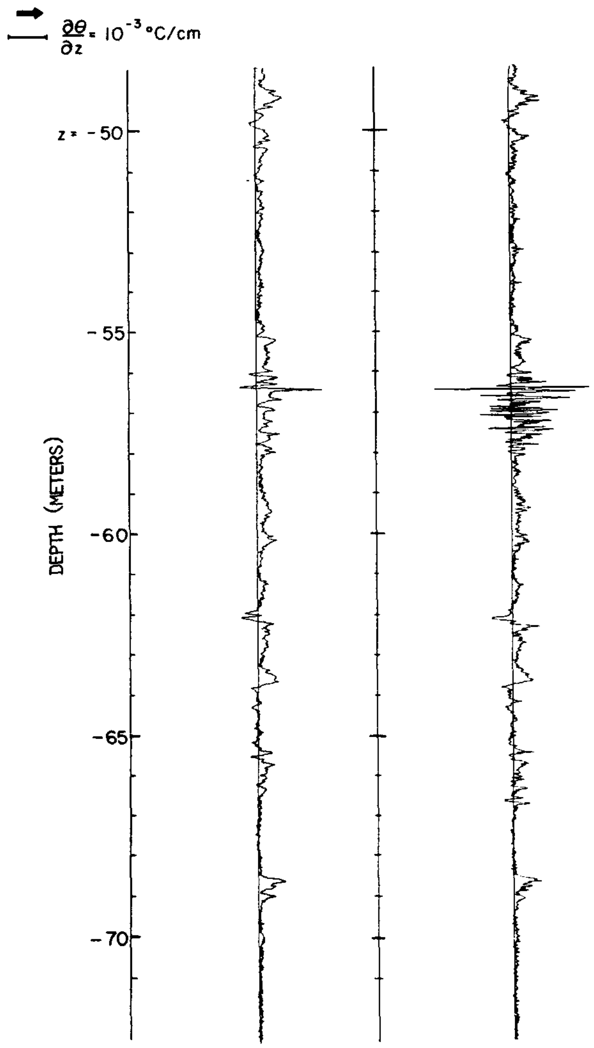

FIG. 3. Temperature gradient between 50 and 70 m in Powell Lake. The scale applies only to the left trace which is from the nose probe. The right trace is from the outboard thermistor which cuts a helical path.

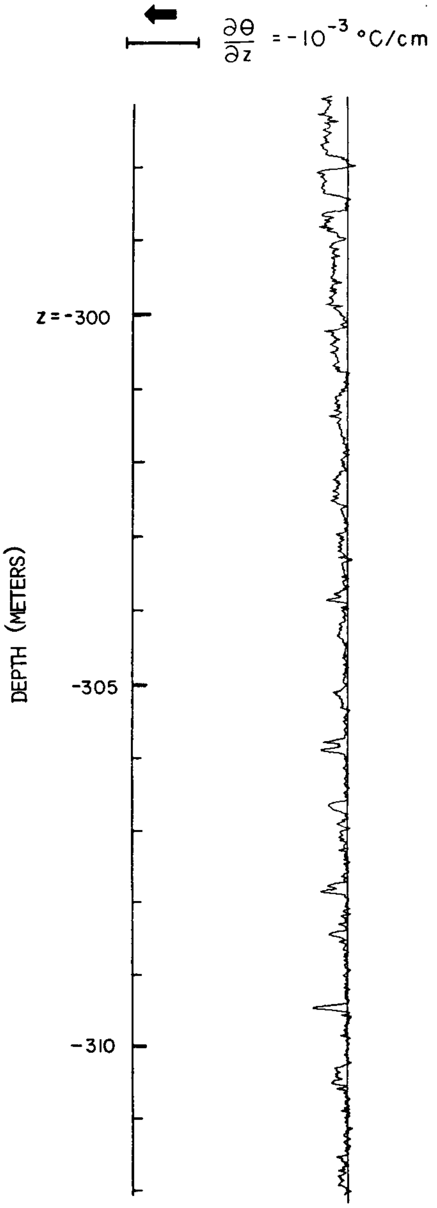

FIG. 4. Vertical component of the temperature gradient near the bottom of Powell Lake.

the other four records taken on this trip show three similar occurrences of intense microstructure below the thermocline. One of these was $\sim$3 m thick and the other two were $\sim$ 1 m.

The rest of the record shows that the irregular positive temperature gradient continued from 70 m down to approximately 110 m where there is a 10 m thick isothermal region ($\partial\theta/\partial z<10^{-4}$C cm^{-1}) below which the gradient becomes irregular and negative with depth. However, the gradient is not uniformly of one sign in either region. We have already observed inversions in the upper fresh water portion of the lake. The upper 100 m of the salt water contains positive gradients although the mean gradient is negative.

In these data the transition from positive to negative mean temperature gradient appears around 110 m. Williams *et al.* found a higher temperature at 125 m than at 150 m, but only by 0.02C. Their salinity values show greater salinity at 125 m than at 100 m, indicating there is salt water above 125 m. It is not possible to decide how much, if any, the salt water interface has mixed upward in the last 10 years.

The data from the salt water portion of the lake have been examined for gradients indicative of doubly diffusive convection. Only drops 4 and 5 which reached the bottom in the flat southern basin show the layering. Fig. 4 shows the profile of temperature gradient vs depth for drop 4 from the central thermistor. The outboard thermistor produced a very similar profile indicating lateral homogeneity. On this drop and drop 5 the instrument actually reached the lake bottom. Drop 5 was about $\frac{1}{4}$ mi away from drop 4 and in slightly

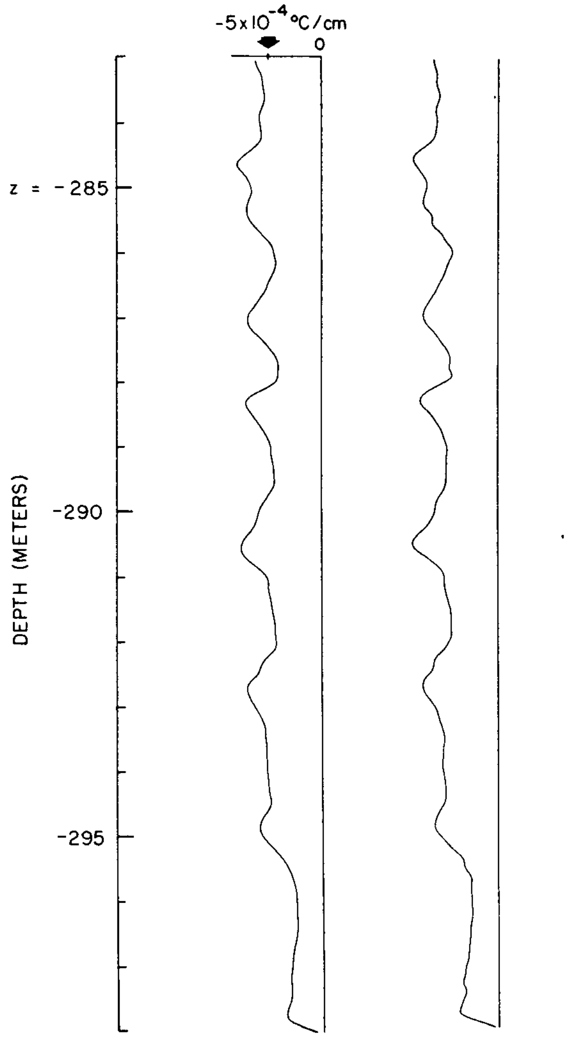

FIG. 5. Temperature gradient profiles between 283 and 298 m in Powell Lake. The left-hand trace is from the centrally mounted thermistor, the right-hand trace from the outboard thermistor.

shallower water, and shows the characteristic doubly diffusive layers in the region between 3–5 m above the bottom. The records for drop 4 end above the bottom of the lake because the tape ran out on the recorder; thus, we cannot set the true height above the bottom by using the fallspeed and time to the bottom. Depth in this region is estimated from Mathews (1962) to be 320±10 m. Between 304 and 311 m Fig. 4 shows a series of high-gradient regions separated by fairly quiet regions. The signal-to-noise ratio of the convoluted data is quite low due to the low level of the signal.

Fig. 5 shows the data for the 15 m region above that shown in Fig. 4. The trace on the left is the vertical gradient as measured by the central thermistor. The trace on the right is the signal from the outbound thermistor. The two traces are plotted with the same gain. The similarity of the signals shows the gradient to be almost vertical. The signal undulates between -2.5×10^{-4} and -8×10^{-4}C cm^{-1}, with high gradient peaks about $\frac{1}{2}$ m in thickness and $2\frac{1}{2}$ m apart. The undulating region extends upward for 60 m and the gradient then becomes irregular.

4. Discussion

a. Fresh water time scales

Osborn and Cox (1972) discuss two time scales which can be applied to microstructure data. The first is related to the period of internal waves, the Brunt-Väisälä frequency N (Eckart, 1960):

$$N^2 = -\frac{g}{\rho}\frac{d\rho}{dz} - \frac{g^2}{c^2},$$

where ρ is the density, g the acceleration of gravity, and c the speed of sound. Since the density of fresh water is dependent only on temperature and pressure, we can write

$$N^2 = -\frac{g}{\rho}\frac{\partial\rho}{\partial\theta}\left[\frac{d\theta}{dz} - \left(\frac{d\theta}{dz}\right)_A\right].$$

The subscript A indicates the adiabatic temperature gradient, which in this case is negligible. Thus,

$$N^2 = -\frac{g}{\rho}\frac{\partial\rho}{\partial\theta}\frac{\partial\theta}{\partial z} = g\alpha\frac{d\theta}{dz},$$

where α is the coefficient of thermal expansion.

Since fresh water is close to its temperature of maximum density below 6C, α is thus very dependent upon temperature. Although the instrument measures and transmits the temperature, the calibrations before the trip are not sufficient to allow accurate calculations of the temperature profile. The loss of the instrument during the last drop makes further calibrations impossible. However, since examination of the temperature records and the limited calibrations gives good agreement with the profile shown in Fig. 1, α will be determined using those temperatures.

TABLE 1. The Väisälä period as a function of the temperature gradient.

$\partial\theta/\partial z$ (10^{-3}C cm^{-1})	Period (min)
0.5	33
1	23
1.5	16
2	12

The value of α is 2×10^{-5} (°C)$^{-1}$(Cox, 1965). Table 1 shows the Väisälä period as a function of the temperature gradient. The large positive spike in the vertical gradient at −56.6 m has a Väisälä time scale of ~14 min, whereas the small inversion preceding it has a time scale of ~37 min.

The other time scale is from molecular diffusion. The gradient associated with a step change in temperature decays in a Gaussian manner, i.e.,

$$\frac{\partial\theta}{\partial z} = \frac{\Delta\theta}{4\pi\kappa t}\exp\left[-\frac{(z-z_0)^2}{4\kappa t}\right].$$

The width of a region of high gradient at the point where the gradient is 63% of the maximum value gives an estimate of the time since the gradient was infinite:

$z_{1/e}-z_0$	1 mm	1 cm	10 cm	1 m
Time	1.8 sec	3 min	5 hr	21 days

The large gradient at −56.5 m has an interface half-thickness around 5 cm, giving a diffusive time scale of 18 min, about 25% longer than the time scale estimated from the Väisälä frequency.

b. Salt water time scales

It is possible to use the diffusive scaling to look at the large-scale temperature and salinity profiles. The temperature interface between salt and fresh water has a $(z-z_0)\approx100$ m, giving a time scale of 500 years. The temperature profile is probably maintained by heat flow through the bottom of the lake.

The salt change is concentrated into a thinner region, but since the molecular conductivity of salt is lower we obtain an estimated age from the salt gradient profile of ~4000 years. Pickard (personal communication) suggests that the salt profile is more likely due to turbulent diffusion eroding the top of the salt water layer and forcing the region of high gradient deeper and deeper. This idea also explains the small overall salt content of the lake for if the lake were a two-layer system of pure fresh water over mixed salt water, the lower salt layer would be less than 80 m thick and the fresh water 250 m thick.

TABLE 2. Value of T, S, β/α, $\Delta S/\Delta T$ and $\beta\Delta S/(\alpha\Delta T)$ for the region from 125 m to the bottom. Temperature and salinity data are from the Institute of Oceanography, University of British Columbia Data Report for 1961. The values for β/α are approximate and calculated from a table of specific gravity anomaly in Cox (1965).

Depth (m)	T (°C)	S (‰)	β/α [(°C)(‰)$^{-1}$]	$\Delta S/\Delta T$ [‰(°C)$^{-1}$]	$\beta\Delta S/(\alpha\Delta T)$
100	4.76	0.08			
125	4.73	0.18	32	3.3	106
150	5.01	0.34	32	0.6	18.3
175	5.14	0.47	32	1	32
200	5.52	0.79	32	0.8	26.9
225	6.00	1.42	16	1.3	21
250	6.73	2.69	16	1.7	27.8
275	7.51	4.63	16	2.5	39.8
300	8.19	11.55	8	50.2	81.4
334	9.19	16.92	8	5.4	43
353	9.33	17.07	8	1.1	8.6

c. Doubly diffusive phenomena

Why is the doubly diffusive process active only at the bottom of the lake? Turner (1965) shows the important parameters in this phenomenon are the contributions to the density difference across the interface by the temperature and salinity differences. He uses the temperature difference across the interface multiplied by α (the coefficient of thermal expansion) and the salinity difference multiplied by β (the coefficient of saline contraction) as non-dimensional parameters. Turner finds that for ratios of $\beta\Delta S/\alpha\Delta T>2$ there is a constant ratio of 0.15 between the density flux due to salt and the density flux due to heat, i.e.,

$$\frac{\beta F_S}{\alpha F_H}=0.15 \quad \text{for} \quad \frac{\beta\Delta S}{\alpha\Delta T}>2,$$

where F_S is the flux of salt and F_H is the flux of heat divided by the specific heat at constant pressure in order to make the final quantity non-dimensional.

It is reasonable to expect that as the stability of the interface increases (increasingly large values of $\beta\Delta S/\alpha\Delta T$) the fluxes will approach molecular conduction.

Neshyba *et al.* (1971) point out that there is a maximum value for $\beta\Delta S/(\alpha\Delta T)$ if the ratio $\beta F_S/(\alpha F_H)$ is to be 0.15 and the transports are to be by molecular conduction; thus,

$$\frac{\beta F_S}{\alpha F_H}=\frac{\beta K_S\Delta S}{\alpha K_T\Delta T}=0.15$$

implies

$$\frac{\beta\Delta S}{\alpha\Delta T}=\frac{0.15}{\kappa_S/\kappa_T}\approx 15 \quad \text{since} \quad \frac{\kappa_S}{\kappa_T}\approx 0.01.$$

Therefore, when $\beta\Delta S/(\alpha\Delta T)>15$ there must be an increase in the diffusion of salt relative to temperature due to the molecular effects.

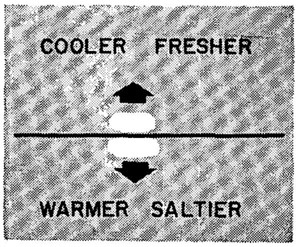

FIG. 6. An interface where relatively cool fresh water overlies warm salty water.

While my data do not permit calculation of $\beta\Delta S/(\alpha\Delta T)$, those of Williams *et al.* can be used. ΔS and ΔT will be evaluated as the difference in salinity between the bottles and we see if this approach is valid. Table 2 shows the values of β/α, $\Delta S/\Delta T$ and $\beta\Delta S/(\alpha\Delta T)$ for the region from 125 m to the bottom, the values of S and T being taken from the U. B. C. Data Report for 1961. The values of β and α are estimated from Cox (1965). Only in the bottom section is the mean density ratio under the "critical" value of 15 and it is there that we might expect to find an extensive region of doubly diffusive convection.

A physical explanation is helpful. Fig. 6 shows an interface with a parcel of cooler fresher water in contact with a parcel of warmer saltier water. Salt and heat are both transferred upward. Since heat is transferred 100 times faster than salt, the lower parcel loses heat and salt but mostly heat (becoming heavier and sinking), while the upper parcel gains more heat than salt (becoming lighter and rising). The coefficients of thermal expansion and saline contraction are used to bring the comparisons of salt and heat flows onto the same dimensional basis. However, in Powell Lake the water is close to its temperature of maximum density. Even though heat is transferred upward much faster than salt, it does not affect the density as much as the small amount of salt that is also transferred upward. Table 2 brings the point out quite clearly that it is the ratio of β/α that is so very large, thereby dominating the ratio of $\beta\Delta S/(\alpha\Delta T)$.

Comparison of the interface in the doubly diffusive region in Powell Lake with data I took in the arctic from T-3 in November 1970, shows the interfaces to be much thicker in Powell Lake. Thickness in the arctic was 1.5 to 10 cm, with many around 5 cm in thickness but in Powell Lake 10 cm is the minimum thickness observed. Also the distance between interfaces in the arctic was up to several meters while in Powell Lake the distance was generally less than 1 m. The thicker interfaces and smaller separations both imply a lower level of convection in Powell Lake than the arctic. The theoretical attempts at explanation of doubly diffusive convection by Turner (1965) and Huppert (1971) start with the relationship $\mathrm{Nu}=c\,\mathrm{Ra}^{1/3}$ which automatically implies the heat flow is independent of the layer thickness. Perhaps there is a relationship between the heat flow and layer thickness.

It is possible to calculate the Rayleigh number for the interface observed in Powell Lake and the result is $\mathrm{Ra} \approx 2 \times 10^6$. Malkus (1964) suggests a critical value around 10^6. The bottom water of Powell Lake is close to this transition point. Using the relationship

$$\mathrm{Nu} = c\,\mathrm{Ra}^{\frac{1}{3}}$$

and the value of $c = 0.085$ quoted by Huppert (1971), the heat flow through these layers is estimated at 1.2×10^{-6} cal cm^{-2} sec^{-1}, only three times larger than the heat flow calculated from the data assuming molecular transport in the region from 200 to 300 m depth. If the transfer of heat through the double diffusivity interfaces is molecular, the heat flow would be 0.4–0.7 μcal cm^{-2} sec^{-1}. Measurements by R. D. Hyndman (personal communication, 1972) show the heat flow through the bottom of the south basin to be 0.65 ± 0.05 μcal cm^{-2} sec^{-1}.

d. Undulating temperature gradients

In the region between 240 and 300 m, the temperature gradient is a very smooth function of depth. Its value does not vary by more than a factor of 3 from the local average over, say, 10 m. Nowhere in this does the gradient change sign as it does in the region above. In this portion of the record, regions with above average values of the temperature gradient are on the order of half a meter thick, considerably thicker than in the region above or below. The cause of this smooth gradient is unknown; it should be noted, however, that this is the region where there is the largest mean salinity gradient. Perhaps what we see here is some form of thermohaline convection.

Acknowledgments. This work was supported by the Marine Sciences Branch, Department of the Environment, and the National Research Council of Canada.

REFERENCES

Cox, R. A., 1965: The physical properties of sea water. *Chemical Oceanography*, J. P. Riley and G. Skirrow, Eds., New York, Academic Press, 73–120.

Eckart, Carl, 1960: *Hydrodynamics of Oceans and Atmospheres.* New York, Pergamon Press.

Gregg, M. C., and C. S. Cox, 1972: The vertical microsctructure of temperature and salinity. *Deep-Sea Res.*, **19**, 355–376.

Huppert, Herbert E., 1971: On the stability of a series of double-diffusive layers. *Deep-Sea Res.*, **18**, 1005–1021.

Malkus, Willem V. R. (Ed.), 1964: Notes on the 1964 Summer Study Program in Geophysical Fluid Dynamics at The Woods Hole Oceanographic Institution. *Geophys. Fluid Dyn.*, **1**, 241 pp.

Mathews, W. H., 1962: Bathymetry of Powell Lake, British Columbia. Ms. Rept. No. 13, Institute of Oceanography, University of British Columbia.

Neshyba, Steve, V. T. Neal and W. W. Denner, 1971: Temperature and conductivity measurements under Ice Island T-3. *J. Geophys. Res.*, **76**, 8107–8120.

Osborn, Thomas R., and Charles S. Cox, 1972: Oceanic fine structure. *Geophys. Fluid Dyn.*, **3** (in press).

Simpson, J. H., and J. D. Woods, 1970: Temperature microstructure in a fresh water thermocline. *Nature*, **226**, 832–835.

Turner, J. S., 1965: The coupled turbulent transports of salt and heat across a sharp density interface. *J. Heat Mass Transfer*, **8**, 759–767.

U. B. C., 1962: Inlet cruises, 1961. Data Rept. 19, Institute of Oceanography, University of British Columbia.

Williams, P. M., W. H. Mathews and G. L. Pickard, 1961: A lake in British Columbia containing old sea-water. *Nature*, **9**, 830–832.

Woods, J. D., and G. G. Fosberry, 1967: The structure of the thermocline. Underwater Association Report 1966–67.