Double-Diffusive Convection and its Implications for the Temperature and Salinity Structure of the Ocean and Lake Vanda

HERBERT E. HUPPERT AND J. STEWART TURNER

Dept. of Applied Mathematics and Theoretical Physics, Silver Street, Cambridge, England

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ABSTRACT

Recently, Fedorov has attempted to interpret some temperature and salinity profiles in the northeast Atlantic, as reported by Zenk, in terms of the results of laboratory experiments in double-diffusive convection. We argue here that the laboratory experiments do not adequately cover all the oceanographic phenomena contained in these records and that Fedorov's interpretation is inappropriate.

Before laboratory results can be used at all in interpreting oceanographic data, however, confirmation that the results of small-scale experiments can be transferred quantitatively to large-scale motions is needed. Such a confirmation is attempted here using the observed profiles in Lake Vanda. A steady-state model of Lake Vanda is considered which determines that heat flux through each double-diffusive interface which is necessary to maintain the lake in a steady state. Using an empirical formula connecting the heat flux and the temperature jump across an interface, we calculate a temperature profile for that part of the lake which is in a state of double-diffusive convection. The agreement between the temperature profile so calculated and the one observed is a test of the derived steady-state heat flux formula. The agreement is surprisingly good and we conclude that, under the assumptions incorporated in the model, this particular experimental result can be used in describing large-scale motions.

1. Introduction

In some recently obtained oceanographic measurements there has been evidence of extensive layers of uniform temperature and salinity separated by comparatively thin horizontal interfaces, across which there were very large temperature and salinity gradients. Areas where such layering has been observed include the Arctic Ocean between 200 and 400 m depth, as probed from the drifting Ice Island T-3 (Neal *et al.*, 1969); the Mediterranean outflow into the northeast Atlantic (Tait and Howe, 1968, 1971; Howe and Tait, 1970); the bottom depths of the Red Sea (Degens and Ross, 1969); the main thermocline in the central Atlantic near Barbados (Amos[1]); and the upper three-quarters of Lake Vanda in Victoria Land, Antarctica (Hoare, 1966, 1968; Shirtcliffe and Calhaem, 1968). Readers are referred to the individual articles for specific details, but the general characteristic of all these observations was the appearance of a series of layers and interfaces with simultaneous jumps in temperature and salinity across the interfaces.

Concurrent with these observations, laboratory experiments of double-diffusive convection have displayed qualitative features very similar to those recorded in the ocean (see, e.g., Turner and Stommel, 1964; Turner, 1965, 1967; Stern and Turner, 1969).

In a recent paper, Fedorov (1970) attempted to use the results of these experiments to explain three particular oceanographic observations. Fedorov's work represents the first major attempt to employ the experiments quantitatively in analyzing oceanographic situations. Such an exercise can only be successful, however, if executed with an eye to the similarities and differences in the two situations and an appreciation of the effects, if any, of the differences. In our opinion, Fedorov did not adequately consider these differences, and his contribution needs reappraisal.

Further, before the experimental results can be used as standard tools in the interpretation of oceanographic data, an intermediate step is required: verification that the results of these small-scale laboratory experiments can be applied, quantitatively, on a larger scale. Though we question whether there is at present sufficient knowledge of any situation in the ocean for which this is possible, such a verification can be attempted in the simpler case of Lake Vanda, a permanently ice-covered antarctic lake. A model of Lake Vanda is derived herein which suggests that the results of the particular laboratory experiment involved can be transferred quantitatively to larger scale phenomena.

A careful description of the experiments invoked by Fedorov and used in our model of Lake Vanda is presented in the next section, followed by a discussion of Fedorov's investigation. Lake Vanda is then discussed in greater detail. The temperature, density and conductivity measurements which have been undertaken

[1] Private communication.

are described, and a calculation is then made which incorporates the quantitative interpretation of the laboratory data.

2. The relevant experiments

Double-diffusive convection is a generic term identifying the type of convection in fluids in which there are two components of different molecular diffusivities which make opposing contributions to the vertical density gradient. In the ocean the components are heat and salt, and they contribute to the density in an opposing sense if the temperature and salinity either both increase with depth or both decrease with depth.

An example of the former case occurs when a layer of uniform temperature and salinity underlies a layer of lower (uniform) temperature and salinity with the total density of the upper layer less than that of the lower. For a sufficiently large temperature difference, turbulent convection occurs in both layers, maintaining the uniformity while both heat and salt are transferred from the lower to the upper layer through the thin interface separating the two layers. A quantitative experimental investigation of this situation was undertaken by Turner (1965) and his results were shown by Huppert (1971) to be consistent with the formulae

$$H_I=0.32\kappa(\alpha g/\kappa\nu)^{\frac{1}{3}}(\Delta T)^{\frac{4}{3}}(\beta\Delta S/\alpha\Delta T)^{-2}, \tag{1}$$

$$F_S=\begin{cases}\alpha\beta^{-1}H_I(1.85-0.85\beta\Delta S/\alpha\Delta T), & 1<\beta\Delta S/\alpha\Delta T\leqslant 2 \quad \text{(2a)}\\ 0.15\alpha\beta^{-1}H_I, & \beta\Delta S/\alpha\Delta T\geqslant 2 \quad \text{(2b)}\end{cases}$$

where H_I is the heat flux across the interface, κ the thermal diffusivity, α the coefficient of thermal expansion, g gravity, ν the kinematic viscosity, ΔT the (positive) temperature difference across the interface, β the proportional density change produced by a unit salinity change, ΔS the (positive) salinity difference across the interface, and F_S the accompanying salt flux.

In another experiment Turner (1968) supplied a constant heat flux uniformly to the bottom of a tank containing water in which the salinity initially increased linearly with depth. After a short time a convecting, well-mixed layer appeared at the bottom, and the thickness of this layer gradually increased until a second convecting layer appeared above it. While the thickness of the first layer now remained constant, the thickness of the second layer increased until a third layer appeared above it; and so on. Due to the difficulty of completely insulating the side walls of the tank and the size limitation of the equipment, only three or four layers were generally created by this mechanism. The increase in thickness of the last-formed layer was due to turbulent entrainment by this convecting layer of the (less dense) fluid above. At the same time, heat was conducted ahead of the advancing front resulting in a temperature profile which eventually caused the fluid ahead to become unstable and mix to form another layer. The interface so formed then transferred heat and salt in the same manner as in the experiment with cold, fresh water over hot, salty water described above. For an initially linear salinity gradient, Turner showed that the thickness of the first layer before it stopped increasing could be expressed as

$$h=H_*^{\frac{1}{2}}S_*^{-\frac{1}{2}}t^{\frac{1}{2}}, \tag{3}$$

and that the final thickness (attained when the second layer formed) was

$$h_1=\left(\frac{\nu\mathrm{R}}{64\kappa^2}\right)^{\frac{1}{4}}H_*^{\frac{3}{4}}S_*^{-1}, \tag{4}$$

where

$$H_*=\alpha gH_B/(\rho c),\quad S_*=-\tfrac{1}{2}g\beta(dS/dz), \tag{5a,b}$$

R, a Rayleigh number, was experimentally evaluated as 2.4×10^4, t represents time, H_B the heat flux supplied at the bottom of the tank, ρ the mean density, c the specific heat, and dS/dz the (negative) salinity gradient.

This experiment describes just one of the many ways a series of layers and interfaces can be created from initially smooth distributions. Many others are described by Turner (1973), to which the reader is also referred for a more comprehensive and detailed discussion of double-diffusive convection.

3. Fedorov's analysis of *Meteor* station 58

One section of Fedorov's paper was aimed at explaining the temperature and salinity profiles reported by Zenk (1970) from *Meteor* station 58 and reproduced in Fig. 1. In particular, Fedorov sought to use Turner's

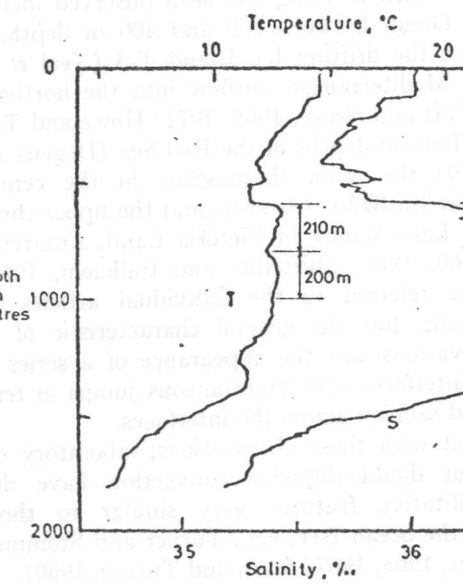

Fig. 1. The temperature and salinity profiles recorded at *Meteor* station 56, as reported by Zenk (1970).

experimental investigation of heating a salinity gradient from below to calculate the thickness of the layer which, according to his interpretation of the record, existed between 800 and 1000 m. He envisaged the outflow taking 4×10^6 sec ($\sim1\frac{1}{2}$ months) to travel the 400 km from Gilbraltar to station 58 while continuously losing heat. Fedorov calculated the quantity of heat so lost, which he regarded as equivalent to the input in Turner's experiment, from

$$H=\epsilon\rho c\Delta hu(\Delta T/L), \tag{6}$$

where ϵ, an efficiency factor, allowed for the fact that not all the heat was "lost in an upwards direction"[2] and was arbitrarily set equal to $\frac{2}{3}$; Δh "the thickness of the main body of Mediterranean water" was taken as 300 m; u was the (assumed constant) velocity of the water mass set equal to 10 cm sec^{-1} ($L=400$ km in 4×10^6 sec); and ΔT "the drop in temperature within the main body of water at distance L" was taken as 1C. Eq. (6) was regarded as the finite-difference formulation of

$$H=\rho c\Delta hu\frac{\partial T}{\partial x}, \tag{7}$$

where $\partial T/\partial x$ is the *horizontal* gradient of temperature within the water mass. Clearly there need be no relationship whatever between this horizontal gradient and the value of the *vertical* temperature difference divided by an (arbitrary) length scale. Nevertheless, from H, calculated[3] as 0.5×10^{-2} cal cm^{-2} sec^{-1}, Fedorov calculated H_* to be 10^{-3} cm^{-1} sec^{-3}, and taking $dS/dz=10‰$ cm^{-1} (see Fig. 1) calculated S_* to be 0.4×10^{-5} cm^{-3} sec^{-2}, which, with $t=4\times10^6$ sec in (3), yielded $h=320$ m.

Leaving aside the quantitative inaccuracies (better estimates could be obtained if necessary), was the situation Fedorov envisaged related to the laboratory experiment and associated theory? Clearly not. In the laboratory experiment heat was supplied at a constant rate *from below*, and during the time for which (3) was valid, the one and only layer was advancing into a constant, stable salinity gradient. Fedorov's outflow consisted of two layers [for which (3) should no longer be valid], heat was being lost from within the layers, and there was a far from constant, stable salinity gradient directly above. Furthermore, no account was taken of the vertical shear experienced in the 400 km distance. Turner's experiment and results are not here directly relevant.

What is the correct explanation of the record? With our present knowledge and the sparse data available this must remain uncertain, but it seems likely that the superposition of different branches of the current flowing along the bottom from the Straits of Gibraltar play a significant role. That is, parcels of water which have followed different paths, with independent mixing histories, could produce a wide variety of layered structures when they rejoin off the coast. A careful re-inspection of the temperature profile makes one wonder if both salinity and temperature layers are present. The temperature hardly changes across the possible interfaces and is not uniform within the layers. Not all temperature and salinity profiles are caused by double-diffusive convection; possibly not this one either.

Fedorov's paper also contains an interpretation of the observed temperature structure in Lake Vanda. We consider Fedorov's investigation in the Appendix, after describing our own model for Lake Vanda (which is quite different from his) in the next section.

4. Lake Vanda

Lake Vanda (77°34′S, 161°36′E), a permanently ice covered antarctic lake, is an ideal place to test hypotheses on double-diffusive convection. The lake is approximately 5 km long and $1\frac{1}{2}$ km wide, with a 3–4 m thick ice cover. The temperature structure has been measured at least five times since December 1961 (Shirtcliffe and Calhaem, 1968) and showed definite layering. A series of measurements taken from different positions on the lake surface by Wilson and Wellman (1962; herein referred to as WW) indicated that the structure is virtually independent of horizontal position and can be considered a function of depth alone. The detailed measurements of Hoare (1966) taken in January 1964 lead to the temperature structure presented in Fig. 2. Two distinct regions are present: the upper region, from 4 to 46.9 m, which consists of thirteen uniform layers; and the lower region, from 46.9 to 64.8 m (the bottom), throughout which the temperature gradually increases to a maximum of 24.8C. The measured density increased very slightly with depth in the upper region, from 1.004 to 1.007 gm cm^{-3}. The increase was considerably greater in the lower region wherein the density increased smoothly up to 1.10 gm cm^{-3} at the bottom. This increase with depth could only be due to an increasing presence of salt. The conductivity measurements showed abrupt increases, at those depths at which the temperature also increased abruptly, from 0.5×10^{-3} to about 6×10^{-3} Ω^{-1} cm^{-1} at the bottom of the upper region, the limit of the measurements. Measurements taken at other times showed only slight differences; the overall description has not changed since (at least) 1961. It appears, then, that the lake is in an almost steady-state situation.

WW found that the only significant heat source was solar radiation. Records taken at Scott Base, 130 km due east of Lake Vanda, indicate a solar energy flux of 91,000 cal cm^{-2} year^{-1} incident on top of the ice sheet. The measurements by WW indicated that approximately

[2] All quotes in this section are from Fedorov, p. 706.

[3] Fedorov's stated 0.25×10^{-2} cal cm^{-2} sec^{-1} is one of the arithmetical slips occuring in his analysis; the arithmetically correct results are quoted herein without further comment.

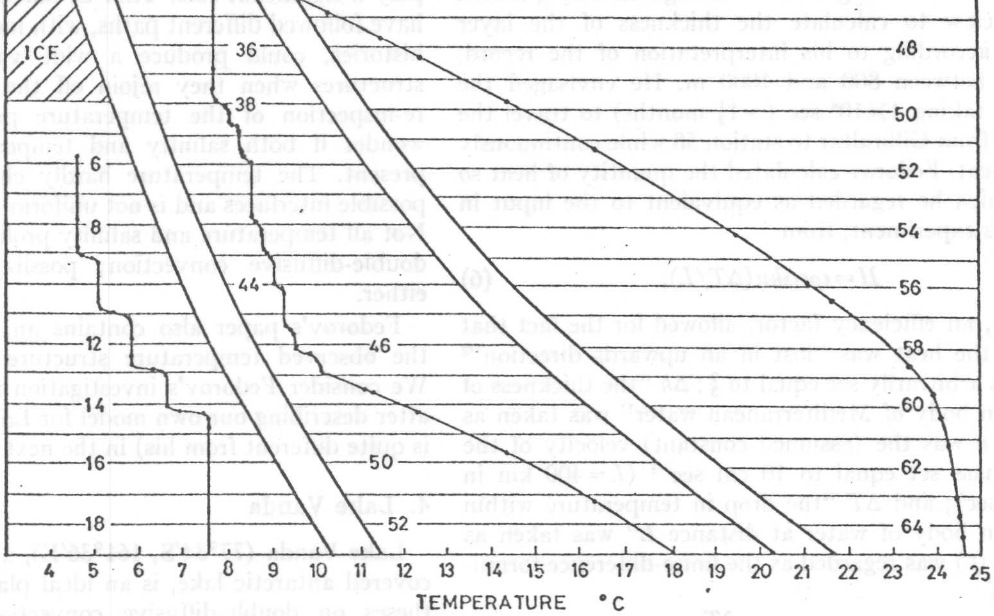

FIG. 2. The temperature profile in Lake Vanda with the depth marked in meters. The measurements shown were made in January 1964 and reported by Hoare (1968). Note the layer of uniform temperature (7.6C) between 14.2 and 46.9 m which has been partially omitted from this figure.

6% of this penetrated the ice sheet, and that 13.7 m of water reduced the flux to a half. Hoare's measurements (1966, 1968) indicated 12.3 m and 12.9 m for this value. Two measurements of the thermal gradients in the sediments at the bottom of the lake have been made. In December 1961 WW used a geothermal gradient probe to obtain a value of -0.13C m^{-1}, indicating a heat flow into the sediments. In December 1962 Ragotzkie and Likens (1964) used a thermistor probe to obtain a value of $+18\text{C m}^{-1}$, indicating a heat flow from the sediments.

Our principal aim is now to construct a model of Lake Vanda within a steady-state framework, with a temperature structure as depicted in Fig. 2. Incorporated in the model is formula (1), and the degree of self-consistency of the model is used as a test for the validity of (1) in large-scale situations. No attempt is made at explaining the *formation* of the present structure; we do not here seek the initial structure of the lake from which the present structure evolved, nor do we determine why the interfaces are at the depths observed. These are problems beyond the scope of this paper.

A major assumption of the model is the use of mean values for the solar radiation. Measurements carried out during the southern summer indicate that the temperature structure of the lake changes very little. The solar radiation, however, does vary significantly during the year, being virtually zero for the months from April to August and increasing and then decreasing nearly linearly during the other months with a peak in December (New Zealand Meteorological Service). This seems to indicate a rather long time constant for the lake, thus partly justifying a model with a mean solar input. A more rigorous justification would require a fully time-dependent model, a refinement which goes beyond the purpose of this present work.

Consider first the lower region of Lake Vanda, from 46.9 m to bottom. The continuous decrease of the temperature gradient with depth indicates that in this region the transfer of heat is by conduction, the heat being supplied in part by solar radiation and in part by the flux (positive or negative) from the bottom sediments. An attempt was made to fit the temperature measured by Hoare, at thirteen points between 46.9 and 59.6 m,[4] to an expression of the form

$$T(z)=Az+Be^{-Cz}+D, \tag{8}$$

where z is depth (m), A the thermal gradient accounting for the heat flux from the bottom sediments, Be^{-Cz} the effect of the heat imparted to the water by solar radiation, and D a normalizing constant. However, the least-squares problem of fitting the data is ill-conditioned. Measured values of C^{-1}, 19.8 m by WW, 22.0 m by Hoare in January 1964, and 21.0 m by Hoare in November 1964, indicate that the comparatively short length of record available is not sufficient to permit accurate evaluation[5] of A, B, C and D. We

[4] The sharp decrease in the temperature gradient below 59.6 m is probably due to muddy water near the bottom, with a significantly larger thermal conductivity. Since no proper estimate can be given for this effect, we exclude this region from further consideration.

[5] Explicitly, large variations in A, B, C and D lead to very small variations in the minimizing function.

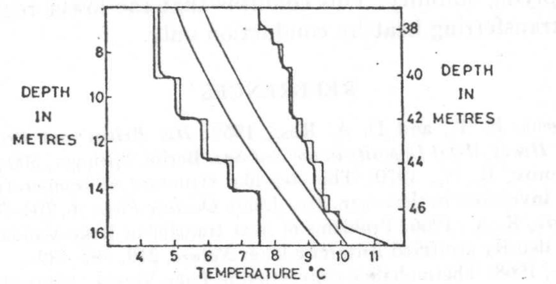

FIG. 3. Comparison between the measured and calculated temperature profiles. The measured profile (thick curve) is taken from Fig. 2, and the calculated profile (thin curve of straight segments) is obtained by the procedure described in Section 4.

hence set $C=0.048$ m^{-1} (the mean of the above) and determined A, B and D for each record.

Although the variations in the values of A, B and D were not small, the value of dT/dz evaluated at 46.9 m was fairly uniform: 1.40C m^{-1} (WW record #1); 1.45C m^{-1} (WW record #2); 1.43C m^{-1} (WW record #2N); 1.37C m^{-1} (WW record #4); 1.56C m^{-1} (Hoare, January 1964); and 1.53C m^{-1} (Hoare, November 1964). To these values, A contributed never more than 10%, though for some records A was positive (indicating heat flow from the bottom sediments) and for some negative (heat flow into the bottom sediments). Accordingly, one cannot distinguish between these two possibilities by this method.

The upper region (0 to 46.7 m) with thirteen layers transfers heat and salt by double-diffusive convection. The heat input to these regions is from two sources: solar radiation and conduction from the top of the lower region. From the measurements of WW and taking $C=0.048$ m^{-1}, the heat imparted to the water between the depths d_1 and d_2 below the ice ($d_1<d_2$) by solar radiation is 1.73×10^{-4} $(e^{-0.048d_1}-e^{-0.048d_2})$ [cal cm^{-2} sec^{-1}]. A thermal conductivity of 1.4×10^{-3} cal cm^{-1} sec^{-1} (°C)$^{-1}$ together with a thermal gradient of 1.46C m^{-1} (the average of the six values calculated above) implies that the heat transfer from the lower region is 2.2×10^{-5} cal cm^{-2} sec^{-1}.

Knowing the profiles of temperature and salinity in the upper region, we can build a model involving a consistency calculation which tests the steady-state heat flux relationship (1) as follows: If the lake is in a steady state, for each layer the heat flux through the lower interface plus the heat imparted to the layer by solar radiation equals the heat flux through the upper interface. Successive application of this principle to each layer starting at the lowest, for which the heat input at the bottom is by conduction, will lead to a value for the heat flux through each interface. The observed temperature and salinity profiles are now used only to evaluate the stability ratio $r\equiv\beta\Delta S/\alpha\Delta T$ for each interface, which, on substitution with the calculated heat flux into (1), leads to the necessary temperature difference across each interface. Then, commencing with the temperature of the top of the lower region, a temperature profile in the upper region can be constructed from these evaluated temperature differences. The comparison between this calculated profile and that observed (Fig. 2) represents a test of (1) which uses the observed profiles only in establishing the stability ratio across each interface (and the depths of the individual interfaces).

Our calculation was slightly less complete than this. Salinity has not been measured in Lake Vanda; only the related quantity, conductivity. Pingree (1970) has described an empirical formalism to convert measurements of conductivity, temperature and depth in the ocean to salinity. Its explicit use at each individual interface seemed unwise here, where the density range is much larger than that on which his formalism is based, and the composition of the lake water may not be the same as the ocean. Instead, Pingree's formulae were used with the temperature and conductivity data of Shirtcliffe and Calhaem (1968) to obtain a mean value for r of 7.5 for each interface. Using this figure in the calculation outlined above and taking the values 10^{-4}C^{-1} for α, 1.4×10^{-3} cal cm^{-1} sec^{-1} (°C)$^{-1}$ for κ, and 1.4×10^{-2} cm^{2} sec^{-1} for ν, we obtained the temperature profile shown in Fig. 3.

The agreement with the observed profile is surprisingly good. Calculations including an arbitrary multiplicative factor of 2 or $\frac{1}{2}$ in front of the right-hand side of (1) yielded results bearing much less similarity to the observed profile, as is shown in Fig. 4. This suggests that the steady-state heat flux relation (1) may be profitably carried over from the laboratory to larger scale situations.

From the above model the heat and salt flux into the first layer, beneath the ice, can be calculated. This heat flux, 1.16×10^{-4} cal cm^{-2} sec^{-1}, continues into the ice, it is envisaged, and combines with external heat fluxes to control the thickness of the ice. The presence of a salt flux, calculated as 2.6×10^{-9} gm cm^{-2} sec^{-1}, using (2b), implies that the model cannot be exactly in a steady state. However, this salt flux is extremely small,

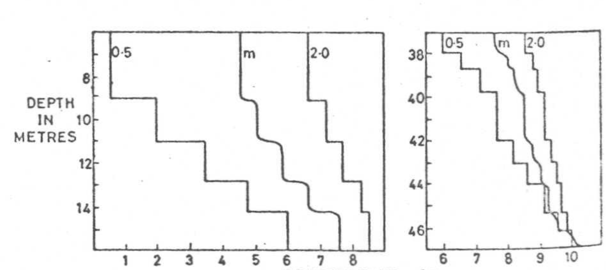

FIG. 4. Comparison between the measured temperature profile and the calculated temperature profiles obtained by including an arbitrary multiplicative factor of 2 or 0.5 in Eq. (1).

[6] The WW record #2S extended only to a depth of 56 m and #3 seems to contain an inconsistent measurement at 53 m; hence, these two records were not considered.

resulting in a change of only 0.14‰ year^{-1} in the salinity of the first layer and can safely be neglected.

5. Conclusion

A steady-state model of Lake Vanda incorporating heat transfer by conduction in the lower region and by double-diffusive convection in the upper region has been considered. Using Turner's laboratory results to quantify the double-diffusive convection leads to a temperature profile in good agreement with that observed. We hence conclude that this particular laboratory result can be applied to large scale motions.

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APPENDIX

Fedorov's Model

Fedorov's model of Lake Vanda differs very significantly from the one presented in Section 4.

Choosing representative values of h and S_* for the double-diffusive layers between 9 m and 14.3 m and also for those between 37.6 and 46.9 m, Fedorov substituted these into (4) to determine the ratio of the heat transfer at the top of the upper region to that at the bottom of the upper region. From the value so obtained, 0.99, Fedorov dismissed the influence of solar radiation. Such a calculation is, however, quite invalid since (4) is correct only for the *first* layer formed when heating a salinity gradient from below, and it cannot be extended to other layers in a steady-state situation. In fact, the first layer in Turner's experiments was very much thicker than subsequent ones.

Having dismissed solar radiation, Fedorov conjectured that the lower region was in a state of double-diffusive convection. From (4), inappropriate for this situation, he calculated that convection with a layer thickness of 2 cm was occurring. Such layering, if it existed, could not be resolved by the measuring instruments used. Now, the lower region is 18 m in depth and had an increase in density with depth due to salt of 9.6×10^{-2} gm cm^{-3} and a decrease due to temperature of 2.6×10^{-3} gm cm^{-3}. Thus, the (destabilizing) thermal Rayleigh number of 3.7×10^{14} is very much less than the (stabilizing) solutal Rayleigh number of 1.4×10^{16}—implying stability. This confirms that the lower region is transferring heat by conduction only.

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