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WESTERN INTENSIFICATION OF BUOYANCY-DRIVEN OCEAN CURRENTS

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A laboratory study is made of a laminar convective circulation in a rotating cylindrical basin of constant depth. The buoyancy forcing is achieved by cooling a portion of the basin's sidewall and thereby causing a deep sinking in a narrow belt adjacent to the cooled sidewall. It is found that the deep sinking produces in the upper layer an anti-cyclonic basin-wide circulation with an intense concentrated current on the upstream side of the sinking region: This implies that in the norther (southern) hemisphere, a northward (southward) western boundary current can be produced by deep sinking in the northern (southern) part of a basin. So long as the fluid surface is free, the surface western boundary current is much stronger and wider then the underlying cold counter current from the sinking region. This convective process on an f-plane may be relevant to the formation of intense boundary currents in the real oceans.

1. Introduction

Strong ocean currents such as the Gulf Stream, the Brazil Current and the Kuroshio tend to occur along the western borders of the oceans. This notable feature of the gross surface-layer circulation of the oceans is known as the western intensification and has attracted much attention in the literature. A generally accepted explanation for the phenomenon is due to Stommel[1)], who showed, using a simple theoretical model, that the wind-driven component of the ocean circulation is westward intensified by the latitudinal variation of the Coriolis parameter (the so-called β-effect). As suspected by Sarkisyan[2)], however, there might be other mechanisms that are responsible for the formation of intensified currents at the west coasts of the oceans; at least, there is no

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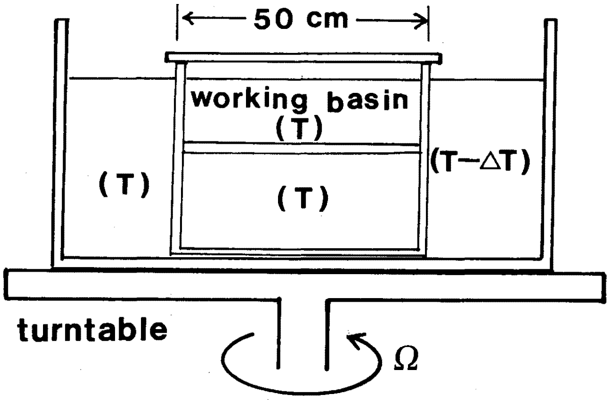
evidence that the β-effect is the only major factor causing the western intensification. More basically, the very notion that the major surface currents in the oceans are predominantly wind-driven is still open to question. The ocean surface is subject to not only wind stress but also differential cooling. Then, how does it come out that the buoyancy-driven component is not important in the surface layer? There seems to be no solid basis for that. On the contrary, recent laboratory studies by Takematsu & Kita[3] and Takematsu[4] have shown that differential cooling and resultant deep sinking give rise to a surface-layer boundary current which is much stronger and wider than underlying (cold) counter current. An additional important finding therein[3] is that such a buoyancy-driven surface current is not necessarily westward intensified by the β-effect unlike wind-driven current. This implies that the westward intensification of the buoyancy-driven component of the ocean currents cannot be explained in terms of the β-effect. The present work is a sequel to those previous laboratory experiments on the buoyancy-driven circulation. The main objective here is to demonstrate that a surface western boundary current can be produced even on an f-plane (i. e., without the β-effect) by a localized sinking of cold water.

The laboratory model consists of a cylindrical basin of constant depth on a rotating table. The buoyancy forcing is achieved by cooling only a portion (1/4) of the basin's vertical sidewall (in the previous experiment[3], the whole sidewall was cooled or heated uniformly). This is to cause a localized sinking of cold (dense) fluid similar to that in real oceans. The object of major interest here is the 'compensatory' upper-layer circulation due to the localized sinking, which seems to have eluded earlier related studies (e. g., Stommel et al.[5], Warren[6], Griffiths & Hopfinger[7] and Condie[8]). In this respect, it should be noted that surface circulations driven by deep sinking are not necessarily the same as those driven by a source or a distribution of sources; both sources and sinks would be needed to simulate the dual effect of a deep sinking.

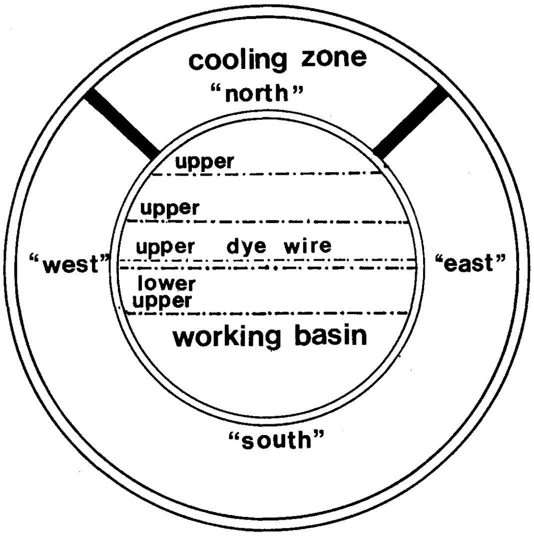
On an f-plane, unlike on a β-plane, there is no definite rule for introducing cardinal points. One natural and meaningful way of doing so may be to regard the cooled portion of the basin as "north" ("south") for northern (southern) hemisphere basic rotation. Other directions are then automatically determined. In what follows, we will follow this rule and directions thus introduced will be distinguished by attaching quotation marks " "

2. Experiment

The experimental set-up on a turntable is sketched in Fig. 1. A plexiglass cylindrical container (50 cm in diameter) was placed centrally in a larger cylindrical tank (80 cm in diameter) filled with tap water. The working fluid with a free surface occupied the upper portion of the inner cylinder partitioned by a thermally insulating false bottom. A portion (1/4) of the annulus region



(a)



(b)

Fig. 1 Sketch of the experimental set-up on a turntable. a) Side view b) Plan view

between the two cylindrical containers was partitioned by two insulated vertical plates so that the water emperature in this portion could be varied independently of the remainder. Initially all the fluid temperatures in the apparatus were set at a fixed value, say T°C, and the top of the inner basin was covered with a lucite plate to prevent the air from exerting a stress on the free surface of the working fluid. The whole apparatus was then set rotating in the counter-clokwise direction (northern hemisphere rotation) about its vertical axis of symmetry.

After the working fluid was fully spun up, the water temperature in the 1/4 portion of the outside annulus was lowered, by ΔT say, by adding chilled water to the portion and was kept at the lowered value (T-ΔT) throughout the duration (about 3 hours) of each experimental run. This allowed us to impose a well-controlled local cooling on the vertical sidewall of the working basin. Experiments were conducted only when the room temperature was slightly higher (by less than 0.5°C) than the initial temperature T°C of the working fluid; otherwise, the top cover of the working basin would be dimmed with vapor. Therefore the working fluid was subjected to heating from above. Repeated checks by releasing dye lines showed that the uniform heating from above alone did not induce any appreciable motion in the working basin.

For the sake of flow visualization, a diluted solution of Thymol blue (a pH indicator) was used as the working fluid and several electrodes of 0.005 cm platinum wire were stretched horizontally across the basin as in Fig. 1 b); the upper dye wires were 1.5 cm below the frec surface and the lower one was 1.5 cm above the bottom. Dye patterns swept off each electrode by the flow were photographed by a camera mounted on the turntable. The temperature of the working fluid was measured by a thermistor at the foot of the cooled ("northern") sidewall and the "western" sidewall.

The controllable external parameters are the basic rotation rate Ω, the hight of the working fluid H and the applied temperature difference ΔT. Values of these parameters were

$$\Omega = 0.1 \sim 0.5 \text{ rad/s}, \ H = 8 \text{ cm},$$
$$\Delta T = 0.5 \sim 1.5^\circ\text{C}.$$

Here it should be noted that the effective temperature difference imposed on the working fluid was much smaller than ΔT due to highly insulating nature of the sidewall material (1 cm-thick plexiglass). Temperature measurements showed that the largest temperature difference in the working fluid was about one fourth of ΔT. Then we will use $\Delta T' = \Delta T/4$, rather than ΔT itself, as a measure of the thermal forcing. In terms of these external parameters, a propagation speed of the density difference, the Rossby radius of deformation and the buoyancy frequency are given respectively as

$$c = \sqrt{g \cdot \alpha \cdot \Delta T' \cdot H}, \quad \lambda = c/2\Omega$$

and

$$N = \sqrt{g \cdot \alpha \cdot \Delta T' / H},$$

where g is the gravitational acceleration and α the thermal expansion coefficient (which depends on T). Values of these quantities were typically

$$(for\ \Delta T = 1.3°C,\ T = 15°C,\ \Omega = 0.3\,\mathrm{rad/s})$$
$$c = 0.6\ \mathrm{cm/s},\ \lambda = 1\ \mathrm{cm},\ N=0.08S^{-1}$$

Related controllable non-dimensional quantities are the Ekman number (E), the Burger number (S) and the Rayleigh number (Ra), which are respectively defined as

$$E = \nu/2\Omega H^2$$
$$S = (N/2\Omega)^2 \cdot (H/L)^2 = (\lambda/L)^2$$

and

$$Ra = g\alpha\Delta T' H^3/\nu\kappa,$$

where L is the diameter of the working basin, ν the kinematic viscosity and κ the thermal diffusivity. Most experiments were conducted under the following conditions:

$$E = 10^{-4} \sim 10^{-3},\ S=10^{-4} \sim 10^{-3}$$
$$Ra = 10^6 \sim 10^7.$$

The convective motion due to the differential cooling is inherently time-dependent, and so it is necessary to refer to relevant time scales. The shortest time scale is $\tau_1 = L/c$, which was of the order of 1 min in the present experiment. The phenomenon under consideration is a typical 'filling-box' process (Turner[9]). The most important time sale for general 'filling-box' problems is the time required to pump all the working fluid through vertical boundary layers along cooled (or heated) sidewalls (see, e. g., Worster & Leitch[10]) and Condie[8]). This 'stratification' time scale is given for the present configulation as

$$\tau_2 = LH/\kappa Ra^{1/4},$$

and takes values of order 1 hour. The longest time scale is the thermal diffusion time

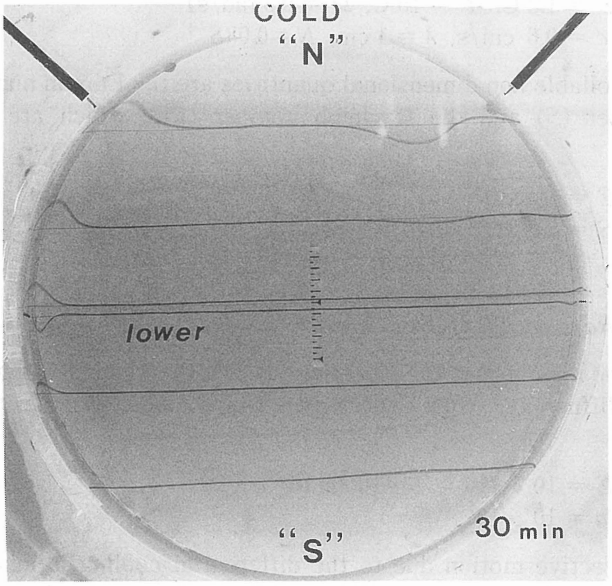
$$\tau_3 = H^2/\kappa,$$

which is about 12 hours in the present case. The duration of each experiment was about 3 hours ($\sim 2\tau_2 < \tau_3$).

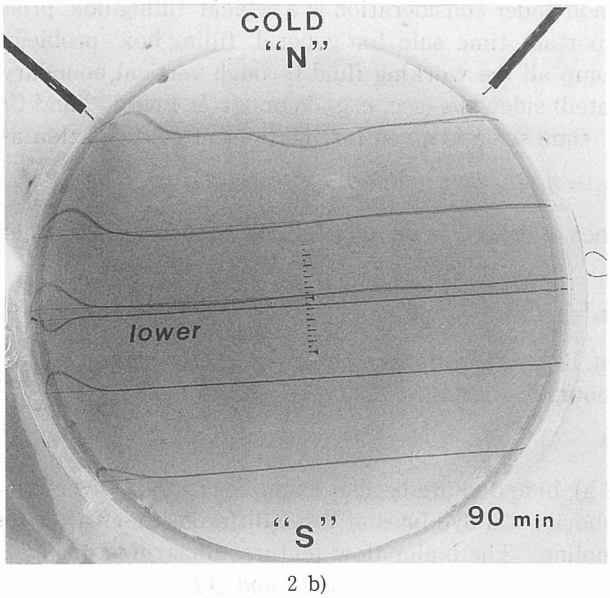
3. Results

Figures 2 a), b) and c) are typical examples of the convective circulation as visualized by horizontal dye lines at three different times after the onset of the differential cooling. The major flow features illustrated in Fig. 2 were similar in the other experiments for different Ω and ΔT.

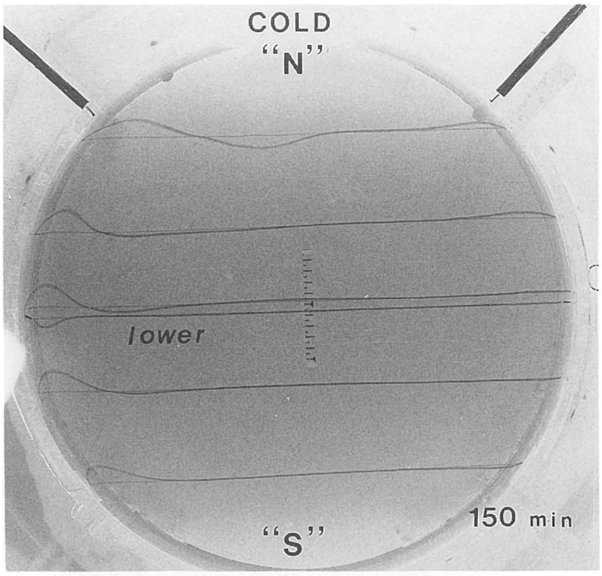
As soon as the sidewall cooling is initiated, working fluid adjacent to the cooled ("northern") portion of the basin's sidewall begins to descend towards the bottom in a thin vertical boundary layer. The cold (dense) bottom fluid thus



2 a)



2 b)

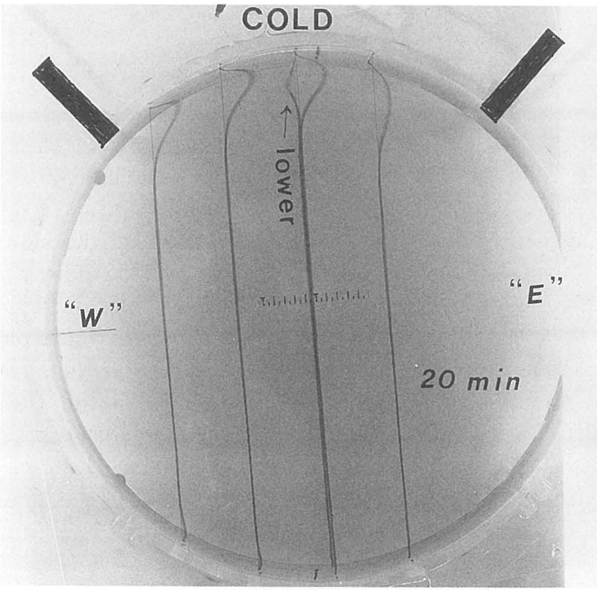


2 c)

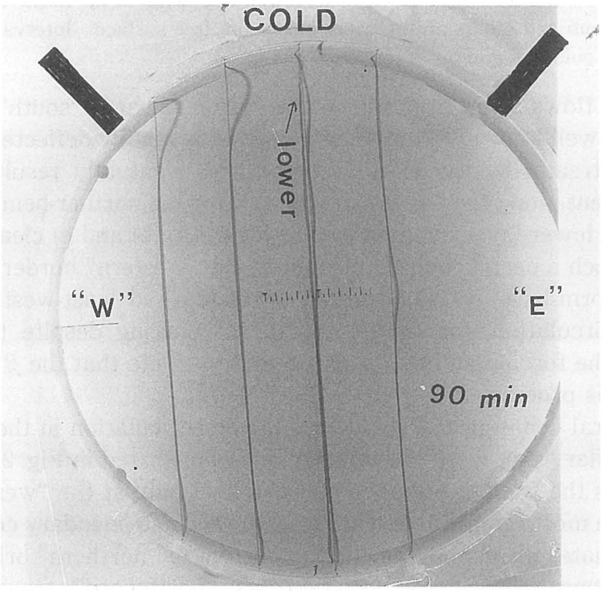
Fig. 2 Horizontal circulation patters at a) 30 min b) 90 min and c) 150 min $\Omega=0.209$rad/s, $a\Delta T=1.3\times10^{-4}$. The lower dye wire is at 1.5cm above the bottom and others are at 1.5cm below the free surface. Interval between dye pulses is 1 min.

produced then flows away from the source region towards "south" in a density current. As is well known[7),8)], the density current is readily deflected to the right (facing downstream) by the Coriolis force and eventually results in a deep boundary current along the "western" sidewall for the norther-hemisphere basic rotation. The lower-layer dye patterns in Figs. 2 a), b) and c) clearly show the existence of such a deep boundary current at the "western" border of the basin. Thus, the uniform basic rotation itself causes a marked "east-west" asymmetry in the deep circulation due to the "northern" sinking despite the complete symmetry of the forcing and the basin geometry. Note that the β-effect is not essential in this process.

As a natural consequence, the compensatory circulation in the upper layer exhibits a similar "east-west" asymmetry as is illustrated in Fig. 2 : The major current twords the sinking region is concentrated only at the "west" border of the basin. The most remarkable feature of the surface boundary current is that it far predominates the underlying counter current of "northern" origin. It is not a mere compensating current. Indeed, its "northward" transport (which increases downstream) is larger than the total rate of sinking estimated from the formula (see, e. g., Condie[8)]) $\pi/4\cdot\kappa LRa^{1/4}$. Another feature worth noting is that the surface "western" boundary current naturally separates from the sidewall as



3a



3b

Fig. 3 Circulation patterns as visualized by "north-south" trending dye lines at a) 20 min and b) 90 min. $\Omega=0.140$rad/s,$\alpha\Delta T=1.1\times10^{-4}$. Dye interval, 1min.

the Gulf Stream and the Kuroshio do. This feature can be seen more clearly in Fig. 3, where circulation patterns as observed by "north-south" trending dye lines are shown. Fig. 3 as well as Fig. 2 show that the separation point gradually moves towards "west" as time goes on.

The "western" boundary current has a net (vertically integrated) horizontal transport so long as the upper surface of the working fluid is free. This was found to be generally true of the slow interior flow ; the interior flow is too slow to be visible in either Fig. 2 or Fig. 3 due to shortness (1 min) of the dye interval (at least 5 minutes were needed to adequately visualize the slow motion). In passing it should be noted that when the upper surface of the working fluid was in contact with a rigid lid, the buoyancy-driven circulation was almost anti-symmetric with respect to the mid-depth and had zero net horizontal transport.

Dye profiles released from the two (upper and lower) electrodes along the "east-west" diameter were graphicall integrated to obtain the "northward" volume transports per unit depth of the "western" boundary current, V_u and V_l, at the two specific depths. A mean vertical shear of the boundary current between the two depths may be defined as

$$S = (V_u - V_l)/h \cdot b,$$

where h is the spacing between the upper and lower electrodes and b is a breadth of the "western" boundary current. If the boundary current satisfies the thermal wind relation, the mean vertical shear can be expressed in the form

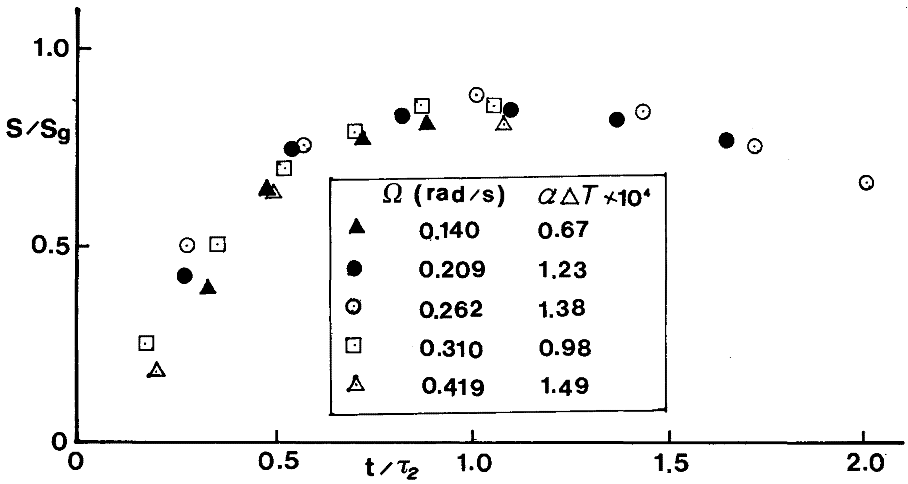


Fig. 4 Normalized mean vertical shear of the "western" boundary currents plotted against the non-dimensional time t/τ_2. S_g is a geostrophic mean vertical shear.

$$S_g = g \cdot \alpha \cdot \Delta T_b / 2\Omega \cdot b,$$

where ΔT_b is a representative temperature difference across the boundary current. Temperature measurements showed that ΔT_b along the "east-west" diameter was about one tenth of the impressed temperature difference except at the initial stage of the development. In Fig. 4, values of S/S_g are plotted against the non-dimensional time t/τ_2 with $\Delta T_b = \Delta T/10$ to illustrate evolution of the convective circulation. The mean vertical shear reaches its maximum at around τ_2, and then gradually decreases towards a final equilibrium value which may depend on the room temperature as well as the thermall boundary conditions (insulated or isothermal). Flow features of the fully developed circulation (at τ_2 and a later stage) are typically seen in Figs. 2 b) and c). Note in particular that the major anti-cyclonic surface circulation at this stage is confined only to the "western" half of the basin. A typical width of the fully developed "western" boundary current is about 5 times the Rossby radius of deformation (λ). The plott in Fig. 4 suggests that the boundary current is nearly in geostrophic balance.

4. Conclusions and discussion

We have revealed some aspects of a laminar convective motion produced by a localized sinking in a rotating basin. The cold water mass in the bottom layer supplied by the sinking makes its way looking at the lateral boundary on the right (left) -hand side for the northern (southern) hemisphere basic rotation. Then a sinking at the high latitude eventually produces a western bottom current even in the absence of the β-effect. In this respect it should be noted that a northward (southward) cold water mass in the northern (southern) hemisphere would form an eastern bottom boundary current. Indeed, such an eastern boundary current is known to exist along the Mid-Atlantic Ridge (Fuglister[11]). The most important finding may be that the localized sinking gives rise to an intense western boundary current also in the upper layer.

The upper-layer western boundary current is much wider (about 5 λ) and stronger than the underlying counter current and is nearly in geostrophic balance. It separates naturally from the lateral boundary and exhibits a downstream increase in transpot. Moreover, the anti-cyclonic gyre involving the western boundary current is confined to the western part of the basin. In these essential respects, the buoyancy driven circulation is very similar to the real ocean circulations; the similarity seems to be better than that of the existing wind-driven circulation model to the real oceans.

The experiments have been conducted only within the laminar-flow regime and for rather restricted ranges of parameters, but the major flow features are expected to be similar for wider ranges of flow conditions. In conclusion, the simple convective process on an f-plane may be another important factor causing the western intensification of the ocean currents. At least the process may be worth studying in more detail as a promising explanation for the phenomenon of western intensification. Quantitative aspects of the buoyancy-driven circulation including its temperature field are now studying. Those results will be published in a subsequent paper.

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References

1) Stommel, H. M.: *The westward intensification of wind-driven ocean currents*, Trans. Amer. Geophys. Union **29** (1948) 202-206.
2) Sarkisyan, A. S.: *Theory and compulation of ocean currents*, Israel Program for Scientific Translations Ltd., 1969.
3) Takematsu, M. and T. Kita: *Some aspects of thermally induced circulation in a rotating basin*, Dyn. Atmos. Oceans **13** (1989) 149-164.
4) Takematsu, M.: *A laboratory study of the Antarctic Circumpolar Circulation*, preprint Volume of Third. Interational Conf. on Southern Hemisphere Met. and Oceanogr., 1989 (Buenos Aires), pp. 252-253.
5) Stommel, H. M., A. B. Arons and A. J. Faller: *Some examples of stationary planetary flow patterns in bounded basins*, Tellus **10** (1958) 179-187.
6) Warren, B. A.: *Structure of deep western boundary currents*, Deep-Sea Res. **23** (1976) 129-142.
7) Griffiths, R. W. and E. J. Hopfinger: *Gravity currents moving along lateral boundary in a rotating fluid*, J. Fluid Mech. **134** (1983) 357-399.
8) Condie, S. A.: *A laboratory model of a convectively driven ocean*, Dyn. Atmos. Oceans **13** (1989) 77-93.
9) Turner, J. S.: *Buoyancy Effects in Fluids*, Cambridge Univ. Press, 1979.
10) Worster, M. G. and A. L. Leitch: *Laminar free convection in confined regions*, J. Fluid Mech. **156** (1985) 301-319.
11) Fuglister, F. C.: *Atlantic Ocean Atlas of temperature and salinity profiles and data from the internatinal geophysical year of 1957-1958*, Woods Hole Oceanographic Institution Atlas Series 1, 1960.

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