Convection in Geophysical Flows

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Abstract

Cooling at the ocean surface produces small scale convective turbulence and mixing of near-surface waters. If the cooling or buoyancy flux is sustained, the small-scale turbulence will erode the underlying density-stratified fluid. The combination of a non-uniform distribution of the surface buoyancy flux and this convectively driven deepening generates lateral density gradients between the convective "chimney" and the ambient fluid and, in turn, a lateral mean flow at the periphery of the convecting region. If the scales are large enough, this mean flow will be affected by rotation, taking the form of a swirling rim current which can become unstable to large-scale baroclinic instabilities that eventually break down the convective "chimney" and mix heat laterally. Recent studies have provided insight into these flows and have considered cases ranging from open ocean convection, where the effects of the earth's rotation is of primary importance, through to exchange flows between semi-enclosed marginal seas and the coastal ocean where topographic control is of primary importance. These studies will be reviewed with emphasis on describing how the buoyancy forcing, the ambient stratification, the rotation rate, the depth and the lateral length scales determine both the timescales of evolution of the flow to steady state as well as the final steady state water mass property distribution and circulation.

Introduction

The ocean has a stable vertical density gradient which inhibits the vertical exchange of fluid as well as chemical and biological tracers between the surface layers and the deep regions of the ocean abyss. In certain localised regions of the ocean, however, episodes of intense heat loss to the atmosphere in combination with relatively weak background density stratification, results in intense turbulent convection and the mixing of surface waters to great depth, often to depths of 2000 m or more. The resulting deep convection is not exclusive to high latitudes and is observed to occur both in the open ocean and in coastal regions where in the latter case topography can play an important role in the convective transport between marginal seas and the open ocean.

Understanding the dynamics of these convective flows is of central importance in quantifying ocean circulation, the budgets of tracers such as oxygen and carbon dioxide and in global climate prediction, but the dynamics of these convective flows has been poorly understood. In part this lack of understanding results from the complex nature of the convective flows which depend upon the intensity, duration and spatial extent of the surface heat or buoyancy loss, the effects of ambient density stratification, the effects of the earth's rotation, and the effects of bottom topography in restricting or controlling the flows. While some field observations of these flows have been reported (see the review by Marshall and Schott [1] for example), they are notoriously difficult to obtain due to the often hostile field environments and the remote nature of many of the known sites of convective activity (Schott et al. [2]). Recent laboratory have, however, been able to able to provide insight into these flows and in this paper we examine this work and some recent numerical studies and suggest a simple framework useful for describing these convectively-driven geophysical flows.

Convective Motions and Scales

In the ocean convection is driven by a combination of heat loss and salinity input, the latter associated with either the brine rejection process associated with the formation of sea ice or evaporation in shallow coastal systems (e.g. Burling et al. [3]). Heat fluxes can be as high as 1000 Wm^{-2} and the overall loss out of the free surface can be described by a buoyancy flux B which is observed to have values often in excess of $10^{-7}\,m^2s^{-3}$. In response to this surface forcing, convective cells form at the surface and begin to penetrate into the interior and a fundamental question is the influence of the rotation of the earth on the convection.

The effects of the earth's rotation are quantified by the Coriolis parameter $f = 2\Omega \sin\varphi$, where φ is latitude and $\Omega = 7.3x10^{-5}\,s^{-1}$ is the rotation rate of the earth. Rotation *controls* the turbulence if the depth is greater than a critical depth scale

$$z_C \geq \alpha\left(B/f^3\right)^{1/2} \tag{1}$$

where the value of the constant α has to be measured. For a fluid of depth H, the ratio of these two *vertical* lengthscales can defined as a Rossby number

$$Ro^* = \frac{z_c}{H} = \frac{\left(B/f^3\right)^{1/2}}{H} \tag{2}$$

From the measurements of Coates and Ivey [4], only if $Ro^* < 0.1$ does rotation *control* the small scales of convective turbulence. For example, using $B = 10^{-7}\,m^2s^{-3}$, $f = 10^{-4}\,s^{-1}$, and $H = 2000$ m, yields $Ro^* = 0.16$. Thus even at such great depth, the conclusion is that while rotation influences the flow, the small scale convective turbulence is not controlled by rotation and is best described by well known non-rotating scaling laws (e.g. Adrian et al [5]).

In response to surface forcing, there is then a well defined sequence of events: penetration stage, unstable stage, and long term stage [6]. While rotation is unimportant in the dynamics of the small scale turbulent convection, it can become important later. Specifically, the convection from the free surface tends to generate a well mixed water mass or "chimney" separated from the surrounding ambient water by a vertical density front. The horizontal density contrast will drive a mean flow and rotation can now becomes important via its influence on this mean flow. Jacobs and Ivey [7] argued that a second Rossby Number Ro_L, based on the ratio of two *horizontal* scales, is now needed to characterise the flow where Ro_L is defined as

$$Ro_L = \frac{R_D}{L} = \frac{(g'H)^{1/2}}{fL} \tag{3}$$

In this definition the buoyancy forcing B is applied on a characteristic horizontal scale L, R_D is the Rossby deformation radius and $g' = (g\Delta\rho/\rho)$ is the horizontal buoyancy contrast between the fluid within the convective region and the exterior ambient fluid. For large Ro_L rotation is likely to affect the large scales of the motion. We can therefore define two possibilities based on the value of Ro_L.

Case 1: ($Ro_L = \left(R_D/L\right) < 1$).

In this case, corresponding to open ocean and shelf convection, rotation does not influence the small scale turbulence but does influence the mean flow, ultimately leading to baroclinic instabilities and horizontal transport of heat by turbulent large scale eddy motion at long time, as depicted in the schematic in Figure 1.

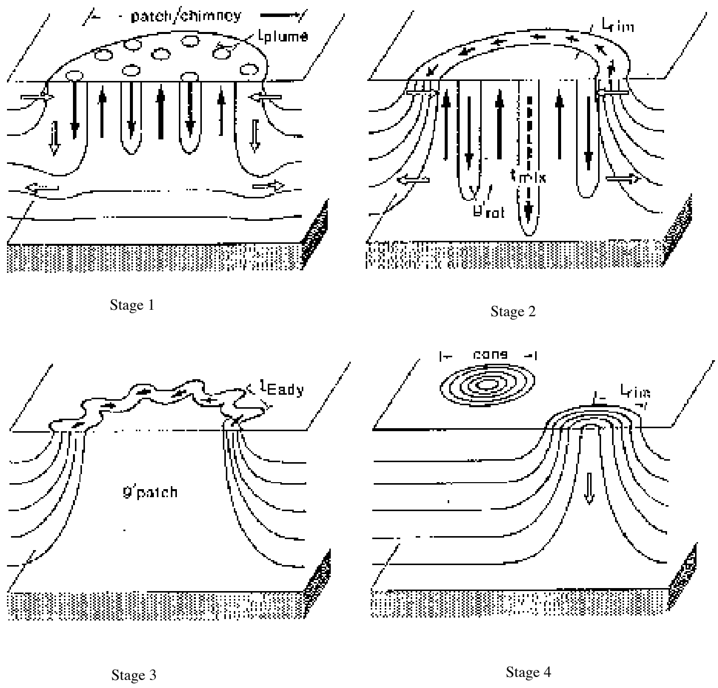

Figure 1: Schematic of stages of flow development after initiation of cooling at the surface in a density stratified fluid. In Stage 1, convective cells penetrate down from the surface into the underlying stratified fluid. Stage 2 corresponds to times longer than the rotation period $T = (2\pi/f)$ where a mean flow develops around the rim of the convective region. In stage 3 this rim current becomes unstable and large scale eddies develop which can mix heat laterally. In Stage 4 there is a final break up after the surface cooling event stops.

The initial heat budget in the convective chimney is a balance between surface heat loss and heat content of the convecting region, i.e.

$$BL^2 t = g' L^2 H \quad \text{or} \quad g' = \frac{Bt}{H} \tag{4}$$

At steady state, the chimney has become baroclinically unstable and the surface heat loss is balanced by a lateral heat flux by large scale eddies, i.e.

$$BL^2 = \overline{vg}(LH) = Rv'g'LH \tag{5}$$

where R is a turbulent correlation coefficient. Laboratory experiments confirm this sequence of events and have shown that at steady state

$$g' = (0.8-0.9)(Bf)^{1/2}(L/H)\,,\; v' = (1.0-1.2)(B/f)^{1/2}\,,\; \tau_S = (1-1.5)(f/B)^{1/2}L \tag{6}$$

where v' is the horizontal eddy velocity scale, and τ_S the time to steady state (Jacobs and Ivey, [7, 8]). Note that while the Coriolis parameter f does not appear in the heat balance in (5), it does influence the final water mass properties in (6).

Initial attempts at numerical simulation of these flows used hydrostatic models (e.g. Jones and Marshall [9]), but Qui and Street [10] report results from a non-hydrostatic, large eddy simulation using a dynamic sub-grid scale closure scheme in a three dimensional domain with up to 6 million grid points. Figure 2 shows a comparison between a typical experimental run from Jacobs and Ivey [8] and the corresponding numerical simulations from Qui and Street [10], clearly demonstrating the initial linear variation of buoyancy contrast with time predicted by equation (4). For longer time, the buoyancy contrasts asymptotically approach the steady state in (6). This work confirms a number of aspects of the flow, in particular the dependence of the density contrast in (5) on the aspect ratio. More fundamentally, the agreements suggest that such numerical model could be used to investigate effects which are difficult to study in the laboratory, such as time dependent boundary conditions and surface wind stress (e.g. Rosman et al, [11]) and spatial variability in the surface heat loss, factors which are known to be important in the field.

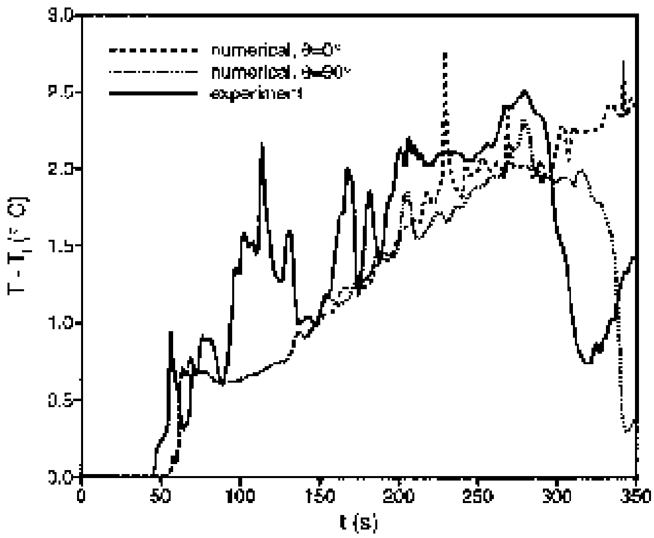

Figure 2. Comparison of laboratory and numerical model results for temperature and hence buoyancy contrast following the initiation of cooling (from [10]).
Parameters are $B = 1.57\mathrm{x}10^{-6}\mathrm{m}^2\mathrm{s}^{-3}$, $\Omega = 0.2\mathrm{rs}^{-1}$.

Case 2: $\mathrm{Ro_L} = (\mathrm{R_D}/\mathrm{L}) > 1$

In this case, corresponding to convectively driven exchange between semi-enclosed marginal seas and the ocean (e.g. Maxworthy [12]) as depicted in Figure 3, rotation is not of primary importance but topographic control is important and horizontal transport of heat occurs by mean flows with strength Q.

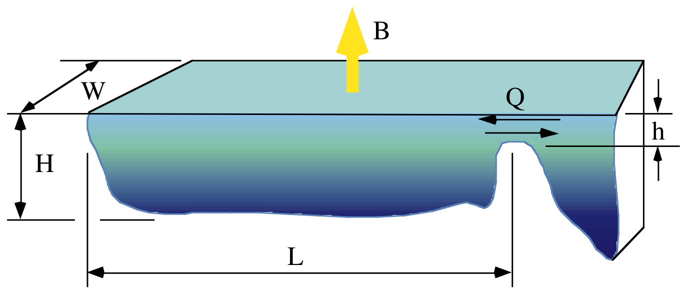

Figure 3: Schematic of topographically controlled convection in semi-enclosed sea.

At steady state the heat balance in Figure 3 is

$$\mathrm{BLW} = \mathrm{Q}g' \quad \textbf{(7)}$$

where experiments have shown (Finnigan and Ivey, [13], [14]) that

$$Q = 0.4(BL)^{1/3} hW\,,\; g' = 2.5\left(B^2/h\right)^{1/3}(L/h)^{2/3}\,, \tau_s = 0.8LhW/Q \tag{8}$$

The depth scale h at the sill is important as hydraulic controls apply since the internal Froude number us locally unity. Numerical modelling of this flow has also been conducted by Finnigan et al. [15] using a three dimensional, non-hydrostatic model, compares well with the earlier laboratory studies, and enables the effect of time varying surface buoyancy flux to be investigated - corresponding to the seasonal variation of the B expected in such water bodies as Shark Bay or the Red Sea - classic geophysical examples of this process.

Discussion

All the various flow regimes possible can be conveniently described in terms of a single regime diagram shown in Figure 4, where each regime is delineated by the value of the two Rossby numbers Ro^* and Ro_L. Jacobs and Ivey [7, 8] have argued that transitions occur when $Ro_L = 0.2$ and $Ro^* = 0.1$, and this suggests the regime diagram shown in Figure 5.

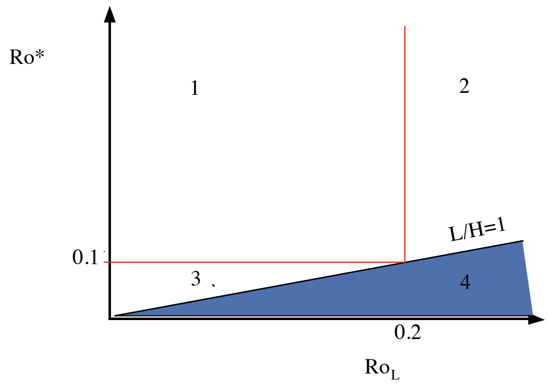

Figure 4: Regime Diagram

Regime 1 corresponds to the case of deep ocean convection or continental shelf convection where rotation is important in affecting only the large scales of motion, leading to instability of the water masses (e.g. Marshall and Scott [1]). Regime 2 is the case where rotation does not affect the convective motion at either the small turbulent scales or the large scale of the mean motion, corresponding to convective motions in small water bodies like lakes, for example (e.g. Sturman et al. [16]). Conversely, Regime 3 corresponds to the case where rotation affects the convective motion at both the small turbulent scales and the large scales; while theoretically possible, there are no known geophysical applications in this regime. Finally, regime 4 corresponds to convective situations where the convection occurs in cells where the depth scale is considerably greater than the horizontal scale; this is the situation in convection under leads in polar regions where convection occurs underneath narrow fissures in the sea ice where the underlying water is directly exposed to atmospheric cooling.

Conclusions

A combination of laboratory experiments and, more recently, numerical experiments, have been able to provide for the first time an understanding of the physics of the convectively driven motions in the oceanic environment. While there are a wide range of processes, four regimes can be identified which appear to capture all possible flows. Within in each regime, the major properties of the water masses and the times of formation can be specified, giving predictive capability for use in refining field experiments and for large scale modelling of the oceanic circulation.

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