

Do Near Boundary Processes Control the Ocean?

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Abstract. Three ways in which near boundary processes influence the ocean are reviewed. In the first topic, solutions for the flow and stratification in a situation where mixing dies away slowly away from a sloping boundary can show a sensitivity to the eddy Prandtl number, emphasizing the importance of studying the mixing of momentum as well as buoyancy. In the second topic, the importance of near boundary mixing in the deep Canada Basin is suggested by the presence in the basin interior of a thermohaline staircase that appears not to be able to account for the vertical heat flux and is also disrupted near the sloping sides by what appears to be a region of strong mixing. The third part of the paper is a summary of some simple recent advances in hydraulic theory that may be relevant to the dynamics and influence of oceanic overflows.

Introduction

At first sight the answer to the question of the title is a simple yes. Apart from tidal forces, the ocean interior is only influenced by external forces that communicate the influence of the atmosphere or the solid earth via boundary regions that are usually rather thin compared with the ocean depth. One could therefore add the word “How” at the beginning of the title and a question mark at the end, and survey the many ways in which this control is exerted. To do so with any degree of thoroughness really requires a monograph rather than a short article, so my intent is instead to keep my general remarks rather brief and then to describe a few particular problems that either demonstrate the importance of boundary regions or raise issues that seem to warrant further attention.

Generalities

We are accustomed to handling boundary layers and the ocean interior separately in ocean models, though the choice of where to put the interface may vary. In some situations, for example, it may just be convenient to think of a bottom drag law that is quadratic in the interior speed, recognizing that there is an unresolved log layer that actually connects the interior to the solid boundary. Alternatively, it may be an Ekman layer that is parameterized. In either of these cases, or with other separations, we would hope to have boundary conditions for the ocean interior that specify things like normal velocity, stress, and either the properties of any water that may emerge from the boundary layer or other flux conditions.

In many situations there is a feedback between the boundary layer and the ocean interior. One well-known example occurs in a simple spindown situation for flows above a flat bottom: Ekman convergence or divergence changes the vorticity of interior flows and reduces the currents there, thus reducing the Ekman flux. There may be more subtle feedback situations that have not yet been investigated adequately. For example, near-inertial and other internal waves generated as a consequence of wind action on the surface mixed layer (e.g. D’Asaro 1995) may lead to increased heaving of the base of the mixed layer and increased mixing beneath its base. These in turn may have a back effect on the thickness of the surface mixed layer (Zahariev 1998) and influence the internal wave generation.

This raises the important topic of “near boundary” processes. While it may be convenient if one can just separate out a well-mixed region as a boundary layer, there are many examples of enhanced mixing extending beyond this into stratified regions of the ocean interior, for both the surface of the ocean (Alford and Pinkel 2000) and near the sea floor if it is rough (e.g. Polzin et al. 1997). It may be simplest to regard these near boundary regions outside mixed boundary layers as belonging to the ocean interior, but we need to discuss the nature of the response there as a guide to future observational and theoretical investigations.

An example of this will be discussed in the next section, drawing on earlier theory (Garrett 1991) to suggest that we should be concerned with momentum transfer as well as enhanced scalar mixing near a sloping boundary. Following that, I will discuss recent results (Timmermans 2003) from the Canada Basin in the

Arctic Ocean which suggest that the vertical transfer of geothermal heat flux in a rather quiescent basin is being accomplished near the sloping sides of the basin rather than the central region.

Overflows of water from one basin to another are a near boundary phenomenon that is attracting increasing attention. These flows often have significant speeds, leading to enhanced mixing and also raising the possibility that they are hydraulically controlled. A discussion of recent results on hydraulic control in the presence of shear and internal friction will be presented in the last main section of this short paper.

Momentum Mixing Matters

Mixing near the sloping sides of ocean basins and lakes has been the subject of much recent observational and theoretical investigation. Observationally, there is considerable evidence for stronger mixing near boundaries than well away from them (e.g. Ledwell et al. 2000) with this enhanced mixing occurring not only in a well mixed bottom boundary layer, but also in a stratified region outside it. This, in fact, draws attention to a key dynamical issue: mixing will tend to tilt isopycnals so that they are normal to the sloping boundary (giving zero normal flux), whereupon restratifying buoyancy forces will arise and tend to flatten the isopycnals again.

If the mixing is represented by eddy viscosity ν and eddy diffusivity κ , then for constant values of these coefficients there is a rather well mixed boundary layer on a slope but this is thinner than the Ekman layer on a flat bottom by an amount that depends on the slope and the eddy Prandtl number (Thorpe 1987). A more interesting situation arises if the mixing coefficients decrease away from the boundary over a larger distance than the boundary layer thickness based on the mixing values right at the boundary. In this case, there is a tendency for the well mixed region to merge into an outer region in which the stratification is finite but reduced from the value much farther away in the ocean interior (Garrett 1991). The isopycnals are tilted in this outer region (Figure 1), with a slope given by

$$\tan \alpha = \frac{1 + SPr}{\kappa_\infty/\kappa + SPr} \tan \theta \quad (1)$$

where α is the angle between the isopycnal and the sloping bottom which is at an angle θ to the horizontal, $S = (N^2/f^2) \sin^2 \theta$ is the slope Burger number, with N the buoyancy frequency well away from the slope and f the Coriolis frequency, and $Pr = \nu/\kappa$ is the eddy Prandtl number, assumed constant even though ν and κ decrease away from the boundary with κ_∞ the eventual eddy diffusivity.

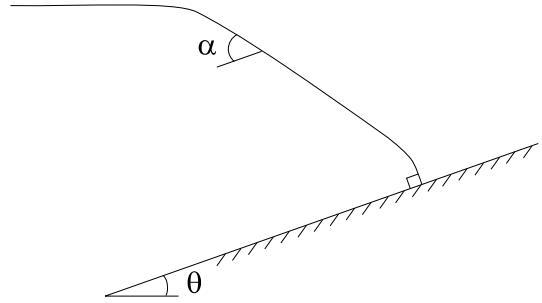


Figure 1. Schematic of an isopycnal near a plane sloping boundary, for a situation in which mixing rates fall off slowly away from the boundary.

The dynamical balances in this region are interesting. There is an alongslope shear flow in thermal wind balance with the lateral buoyancy gradient associated with the tilted isopycnals. The divergence of the viscous force acting on this alongslope flow is then balanced by the Coriolis force on a weak downslope flow which acts to restratify the region and balance the influence of mixing.

As the far ocean interior is approached, with $\kappa \rightarrow \kappa_\infty$, $\alpha \rightarrow \theta$ so that the isopycnals become flat, but before that, with $\kappa \gg \kappa_\infty$, they can have a significant slope away from the horizontal if SPr is not large. In fact, if SPr is small, as for a small slope, the isopycnals tend to be nearly normal to the boundary over this whole extended region.

The interesting aspect of this solution is its dependence on the Prandtl number Pr and hence on the eddy viscosity ν . Of course, the concept of an “eddy viscosity”, with its connotation of a strictly local dependence of the eddy momentum flux on the mean shear, may not be an appropriate representation of the physical processes, but operationally one can just define the eddy viscosity as the momentum flux divided by the mean shear. It is clear that, in a situation with mixing falling off slowly away from a plane boundary, we need to know the eddy momentum flux as well as the scalar mixing rate.

An objection to this scenario is, of course, that in many of the situations where mixing is enhanced near a sloping boundary, the topography is not at all well represented by a plane slope, but is instead furrowed by deep canyons (St. Laurent et al. 2001). In this situation, isopycnals can be tilted across the canyon, setting up a pressure gradient along the slope. This pressure gradient, rather than the divergence of the viscous force on the alongslope flow, then balances the Coriolis force on the upslope flow. The solution for this situation is as for a flow with $f = 0$, or $S \rightarrow \infty$ in (1); outside the boundary layer immediately adjacent to the bottom,

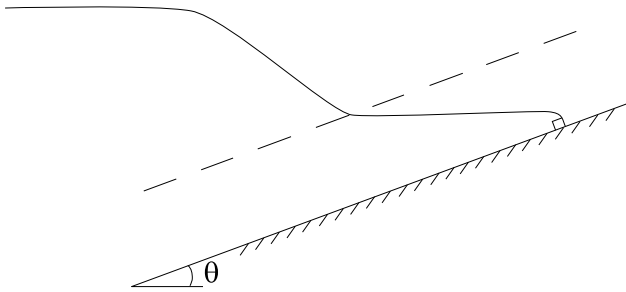


Figure 2. Schematic of the possible shape of an isopycnal near a sloping boundary with canyons and ridges, for a situation in which mixing rates fall off slowly away from the boundary.

complete restratification has occurred.

If the mixing extends farther above the bottom than the top of the ridges, it is possible that the solution for a plane boundary becomes relevant again. Maybe the isopycnals are now as depicted in Figure 2, although this is not seen in the data presented by St. Laurent et al. (2001) for the Brazil Basin. Of course, in that situation and most others, the flow is influenced by more than just mixing, but the simple dynamics reviewed here does point to the need for more attention to the diapycnal transfer of momentum as well as buoyancy.

Enhanced Boundary Mixing in the Canada Basin

As mentioned earlier, there are many situations where the diapycnal mixing for a basin appears to be dominated by regions near the sloping sides. The topic is reviewed by Wüest and Lorke (2003) who also present evidence for the same effect in lakes. In this section I shall just report briefly on a recent suggestion that this also happens in a deep Arctic basin (Figure 3). The main points presented by Timmermans et al. (2003) are

- The bottom water in the Canada Basin appears from ^{14}C dating to be about 500 years old (Schlosser et al. 1997), and in that time has responded to geothermal heating by developing a bottom boundary layer that is many hundreds of metres thick and warmer than the water above it.
- Surveys over more than a decade show that this bottom layer is no longer warming, so that the geothermal heat must be escaping.
- In the central basin, the thick bottom layer is surmounted by a staircase of isothermal layers separated by thin interfaces (Figure 4), and formulae based on laboratory studies predict vertical heat fluxes that seem to be consistent with the expected geothermal flux of

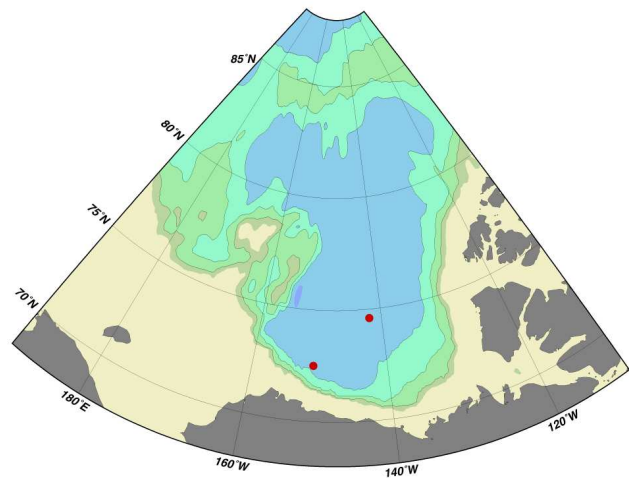


Figure 3. The Canada Basin of the Arctic Ocean. The isobaths are 1,000, 2,000, and 3,000 m and the red dots show the locations of stations for which deep temperature profiles are shown in Figure 4.

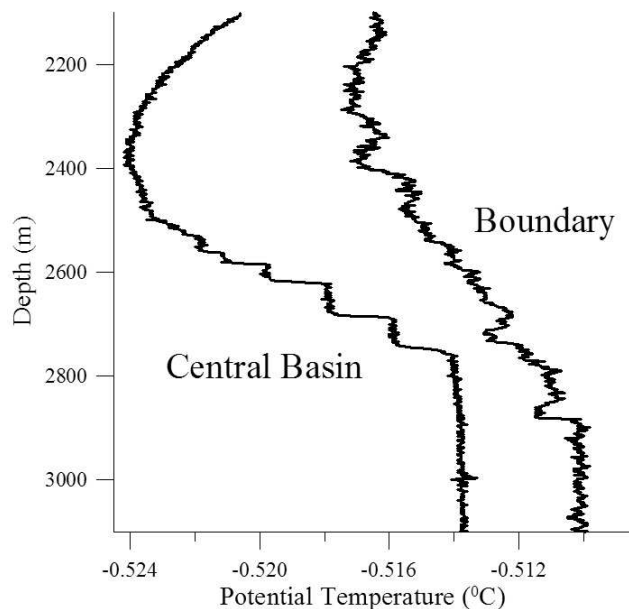


Figure 4. The deep potential temperature profiles for the stations shown in Figure 3. See Timmermans et al. (2003) for details.

about 50 mW m^{-2} .

• However, the interfaces are too thick to transport the heat by molecular conduction, and apparent over-turns in the temperature profile appear to be the consequence of noise rather than being a signature of turbulent transfer. It thus seems that the heat is not escaping through the staircase.

- The heat may be escaping vertically near the

boundaries of the basin (Figure 4), where the staircase vanishes and the temperature profile seems to offer evidence of overturns of sufficient scale to indicate turbulent transport of the required magnitude.

- The energy required for this near boundary turbulent mixing is not large, given the weak stratification, and seems not to be inconsistent with expectations of small energy input into the internal wave field by wind and tides (Halle and Pinkel 2003).

It is perhaps worth mentioning that the conclusions on the lack of mixing in the interfaces and the vigorous mixing near the boundaries came from Thorpe scale analysis. This was hampered by the fact that in this deep basin the salinity gradient is too weak to allow reliable estimates of the density profile on vertical scales of just metres, so that temperature profiles alone were used in the hope that intrusions are not a major complication. With this assumption, distinguishing between real overturns and spurious ones arising from temperature sensor noise was aided by use of the run length criterion proposed by Galbraith and Kelley (1996). The point to be made here is that, in regions of the ocean where mixing matters, the estimation of Thorpe scales from routine CTD data may be adequate and a simple alternative to the sophistications of microstructure measurement.

Hydraulic Control of Overflows

While near boundary regions in the ocean seem to be of general significance, straits and channels may be of particular importance. In some situations, such as the exchange through the Strait of Gibraltar (e.g. Bryden and Kinder 1991), there are two active layers, whereas in others, such as the Denmark Strait overflow (e.g. Käse and Oschlies 2000), one can think in terms of just one active layer. In all these situations, concepts of hydraulic control can be useful. The simplest situation is for the flow of a single homogeneous layer in a channel of rectangular cross-section and constant width, but with varying bottom height $H(x)$, where x is the downstream coordinate (Figure 5).

The momentum and continuity equations are

$$u \frac{du}{dx} + g \frac{d}{dx}(h + H) = 0, \quad hu = Q \quad (2)$$

where u is the current speed, h is the thickness of the layer, and Q is the volume flux per unit width. The equations integrate to

$$\frac{1}{2}Q^2/h^2 + gh + g(H - a) = 0 \quad (3)$$

where a is the height of an upstream reservoir. If the level of a downstream basin is lowered, the upstream

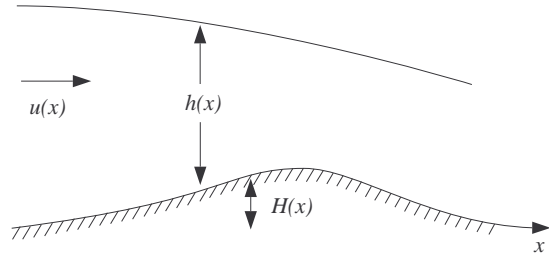


Figure 5. Depth independent flow of a homogeneous fluid over a bottom of varying height.

reservoir height a follows until it is no longer possible to have a solution at the highest point of the channel where $H = H_{\max}$. Then

$$a = H_{\max} + (3/2)(Q^2/g)^{1/3} \quad (4)$$

and the flow is said to be “controlled” in the sense that, for a specified flow rate, further changes in the downstream basin height cannot affect the height of the reservoir. (This is consistent with the idea that disturbances cannot propagate upstream as the Froude number $F = u/(gh)^{1/2}$ is equal to 1 at the control section at the crest of the ridge.) Alternatively, Q can be predicted from (4) and a knowledge of $a - H_{\max}$. Either interpretation of this standard hydraulic result can be of considerable value in oceanographic situations. To be sure, extensions are required to allow for reduced gravity in overflows, varying channel width, different channel shapes and rotation. Recent theoretical studies have also addressed questions about the role of bottom friction, mixing and vertical shear of the velocity profile. One motivating example from the deep ocean is shown in Figure 6, which hints at the possibility of hydraulic control, though the nature of this and the location of the control section are not clear because of the intense mixing (again apparent from a Thorpe scale analysis) as well as the stratification and irregular topography.

A first step in including friction was reported by Pratt (1986). He added a quadratic bottom drag term $-C_d u^2/h$ to the right hand side of the momentum equation in (2) and showed that control still occurs where the Froude number equals 1, but that the location of the control section is shifted downstream from the crest of the ridge to a location where $dH/dx = -C_d$. This could be a significant shift if C_d is several times 10^{-3} . One way to derive this is to write the momentum equation as

$$\frac{d}{dx} \left[\frac{1}{2} \frac{Q^2}{h^2} + gh + g(H - a) \right] = 0, \quad (5)$$

where $g da/dx = -C_d u^2/h$, as if the head a varies downstream. Then the minimum of the function $\frac{1}{2}Q^2/h^2 + gh$, giving the location of the control section where also

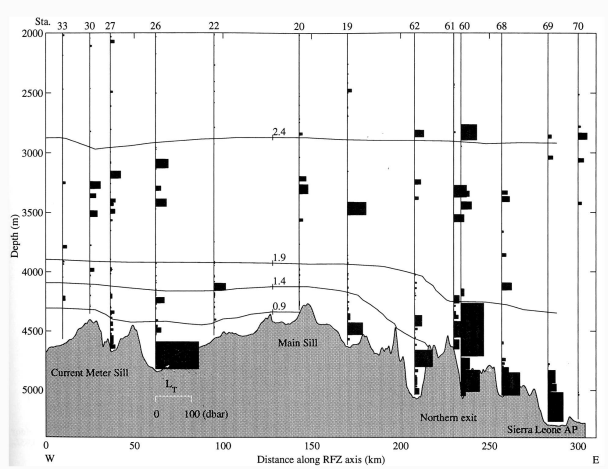


Figure 6. Potential temperature contours in the Romanche Fracture Zone, with the black bars showing the Thorpe scale as a measure of mixing. From Ferron et al. (1998).

$F = 1$, occurs where $d(H - a)/dx = 0$, or $dH/dx = -C_d$ using $F = 1$.

Even for a homogeneous fluid this ignores the likely presence of shear in the velocity profile, a factor recently investigated by Garrett and Gerdes (2003). They converted the trivial equation $h = \int_H^{H+h} dz$ to

$$\int_0^Q \frac{d\psi}{u} = h \quad (6)$$

where the streamfunction ψ describes the flow via $u = d\psi/dz$, $w = -d\psi/dx$. For an inviscid flow, the momentum equation can be written

$$\frac{1}{2}u^2 + gh + g[H - a(\psi)] = 0, \quad (7)$$

where the head, or Bernoulli constant, a is now a function of the streamfunction rather than constant as it is for an unsheared flow. The value of u from this may now be substituted into the continuity equation (6) to give the single equation

$$J(h; H) = \int_0^Q \frac{d\psi}{(2g)^{1/2}[a(\psi) - (H + h)]^{1/2}} - h = 0 \quad (8)$$

for the layer thickness h as a function of the prescribed heads $a(\psi)$ and the varying bottom height $H(x)$. As for the unsheared flow (which can be included in this formulation by taking $a = \text{constant}$), $J(h; H)$ has a minimum at some intermediate value of h so that (8) does not have a solution if H exceeds some critical value. A nondimensionalised example is shown in Figure 7 for a linear shear; we see that for $b = a_{\min} - H(x) < 0.993$ there are no solutions, so that the critical value of H is

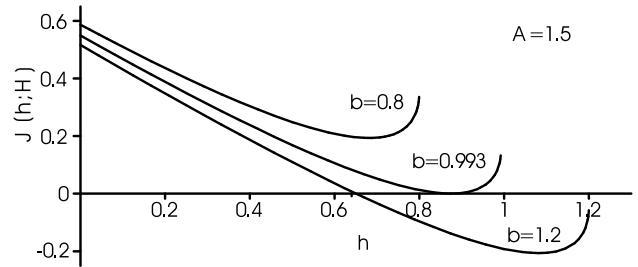


Figure 7. The function $J(h; H)$ from (8), for the case of a linear shear with $a(\psi) = a_{\min} + A\psi$, where $A = 1.5$, for various values of $b = a_{\min} - H$. The streamfunction ψ has been non-dimensionalized with Q , and a, H, h with $(Q^2/g)^{1/3}$. From Garrett and Gerdes (2003).

just $a_{\min} + 0.993$. For H less than this, there are likely to be two solutions for h , though in the example here we see that for small H one solution can disappear, as for the curve for $b = 1.2$. This corresponds to reversal of flow near the bottom, violating the assumption of unidirectional flow.

Controlled flow occurs if the the maximum bottom height equals the critical value, i.e. if the minimum of J as a function of h , where $\partial J/\partial h = 0$, occurs on the abscissa. Using (7) as well, it is easy to show (Garrett and Gerdes 2003) that this occurs when

$$\int_H^{H+h} \frac{g dz}{u^2} = 1. \quad (9)$$

This is then the condition on the flow profile for controlled flow, and, in fact, still corresponds to long waves being arrested, though now with a depth averaged flow $\int_H^{H+h} u dz$ greater than the value $(gh)^{1/2}$ for unsheared flow.

The extension to allow for friction can again proceed by allowing the Bernoulli head $a(\psi)$ to vary downstream, with $\partial a/\partial x = \partial \tau/\partial z$ where $\tau(x, z)$ represents internal friction. Garrett (2003) shows that control still occurs when (9) is satisfied, i.e. when long waves are arrested, but that the location is shifted downstream of the ridge crest to a location where

$$\frac{dH}{dx} = -C_d + \int_H^{H+h} \frac{2\epsilon}{u^3} dz; \quad (10)$$

with bottom friction represented as $C_d u^2$, as before, and ϵ the internal dissipation given by

$$\epsilon = \tau \frac{\partial u}{\partial z}. \quad (11)$$

This is just $\nu(\partial u/\partial z)^2$ if the stress is represented with an eddy viscosity ν . The opposite signs of the two terms in (10) are a puzzle at first, but occur because bottom

drag removes momentum as well as energy, whereas internal friction just redistributes momentum in the vertical; on a flat surface, bottom friction would need to be balanced by a downhill slope whereas a lessening of vertical shear by internal friction requires an uphill slope to conserve the momentum flux. While (10) reduces to Pratt's (1986) result that $dH/dx = -C_d$ in the absence of shear, the presence of shear moves the control section back towards the crest of a ridge. Garrett (2003) shows that typical parameterizations of internal friction will lead to a prediction of control where $dH/dx \simeq -0.5C_d$.

This result is hardly something that one would take too seriously in trying to say where particular reduced gravity flows in the ocean might be controlled, particularly as it applies only to the flow of a homogeneous fluid, and situations such as that shown in Figure 6 are clearly influenced by the presence of stratification and entrainment into the active layer. Once again, however, a simple, idealized, model can help to build intuition. Gerdes et al. (2002) investigated the effects of entrainment on the hydraulics of a vertically homogeneous and unsheared reduced gravity flow. In a channel of constant width, the control section is now shifted to a location where

$$\frac{dH}{dx} = -C_d - \frac{3}{2} \frac{u_e}{u} \quad (12)$$

where u_e is the entrainment speed.

Combining the various different influences on hydraulic flows is likely to require a numerical model, but analytical results for idealized cases which treat just one or two effects at a time should aid in the understanding of numerical output as well as data, and help in the prescription of parameterizations that are simple enough to use when representing overflows in large scale models.

Discussion

The three topics discussed in this brief review are rather different but do perhaps illustrate some common points. One is that more mixing seems to occur near sloping boundaries than in the ocean interior, but we need to remember that the nature of the response is likely to depend on the transfer of momentum as well as buoyancy. Another point is that the physics is further complicated by the heterogeneity of bottom topography. This needs to be taken into account in observational programs and theoretical and modeling studies, both for mixing that is caused by things like internal tides and other internal waves, and for flows driven by large-scale buoyancy differences in the ocean. In all of these topics, it seems that insights can still be obtained from the investigation of idealized models.

Quite apart from the subtleties involved in determining suitable parameterization schemes of near boundary

processes for use in large-scale models, there are still many basic questions about the physical processes associated with the influence of all scales of bottom topography.

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