are the same, but will be harder to implement if the FGM has a finer vertical mesh.

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EDDY DRIVEN EXCHANGE AT OCEAN FRONTS

Michael J. Follows and John C. Marshall

ABSTRACT

We explore the possibility that finite amplitude instability of ocean fronts provide a mechanism for exchange of mass and properties between the mixed layer and thermocline. Drawing parallels with studies of tropopause folding in the atmosphere, we estimate an upper limit on the exchange rate at the front due to ageostrophic axial circulations, and relate it to oceanographic observations. We deduce that eddy-driven exchange across frontal regions may be vigorous locally. Its time-mean effects, when viewed on the large scale, are potentially important, resulting in subduction (or entrainment) rates which are comparable to those associated with the steady circulation.

1. INTRODUCTION

The exchange of fluid across the mixed layer base is important to the chemical and dynamical balances of the ocean. Subduction of mixed layer water into the thermocline decorrelates the thermodynamic and chemical contact with the atmosphere and sets the properties of the ventilated thermocline (see, for example, the diagnostic study of Marshall et al., 1993). Entrainment of thermocline water into the mixed layer...
may bring supplies of nutrients into the photosynthetically active region. The possibility that significant exchanges could also be associated with radius of deformation scale, time dependent processes at ocean fronts has been noted (Woods et al., 1977) because there are vertical circulations in such regions many times stronger than Ekman pumping rates. Pollard and Regier (1992) analyse dynamical observations from the FASINEX program, and argue that at unstable fronts, strong ageostrophic, axial circulations dominate the movement of pockets of mixed layer water into the thermocline and vice versa. They suggest that unstable fronts may be important in controlling the rate of exchange of mass across the mixed layer base. Tracer studies, such as those discussed by Pollard (1986), Washburn et al. (1993) and Radoke et al., (1991), indicate that mixed layer water can be entrained into the thermocline in the vicinity of an unstable jet, although conclusive observational evidence of an irreversible exchange due to frontal processes is not yet available.

Here we estimate, by assuming that the magnitude of the ageostrophic axial circulation associated with frontogenesis is the controlling factor, an upper limit on the exchange of mass achieved by a population of fronts embedded in an ocean gyre. We show that this upper limit is significant when compared, for instance, with the subduction associated with the large-scale, climatological flow (Marshall et al., 1993). Our approach is guided by the belief that there are strong dynamical parallels between ocean front circulations and the tropopause folding processes important in stratosphere-troposphere exchange (e.g., Danilovs, 1968). A parameterisation of the exchange rate due to folding of the tropopause in the atmosphere has been developed by Follows and Austin (1992). Here we apply these ideas to the ocean.

2. CIRCULATION AT AN OCEAN FRONT

In Figure 11 we schematise the structure of a generic front meandering from west to east across the ocean, isopycnals slope sharply up towards the surface outcropping into a vertically mixed layer characterised by very low values of potential vorticity. For simplicity we suppose that this ‘mixed layer’ (low PV water) has no large scale variations in depth prior to deformation by eddy activity. However, as the jet running along the front accelerates and decelerates, then, by frontogenesis theory, an axial ageostrophic circulation is driven represented symbolically by the loops in Figure 11. Under the influence of the meandering along-stream jet (in geostrophic balance) and the axial (ageostrophic) circulation, the base of the low PV layer undulates forming ‘pockets’ of weakly stratified fluid which protrude down, along the sloping isopycnals, into the stratified fluid below and vice versa. If these pockets can be eroded by small scale processes or detached during the wave breaking process, then a net exchange of fluid can occur across the base of the mixed layer.

Consider, referring to Figure 12, a ‘jet streak’ (a zonally aligned sector of the jet, of which there will be many, whose current is anomalously large). At the entry region to the jet streak there will be convergence which spawns an ageostrophic circulation across the front leading to a concentration of cross frontal temperature gradient, frontogenesis and fast narrow currents. Let the along front scale of the jet streak be \( l \) and \( u' \) the zonal speed in the streak. At the axis of the jet the zonal momentum equation is

\[
\frac{Du}{Dt} = f u_g
\]

(1)

where subscript \( ag \) denotes the ageostrophic component. Thus in the jet entrance region, where \( Du/Dt > 0 \), a northwards ageostrophic circulation is induced at the surface. The timescale for the passage of a particle of fluid a distance \( l \) along the jet is \( \Delta t \sim l/u' \), implying that the magnitude of this ageostrophic flow is, from (1):

\[
u_g \sim \frac{u'^2}{f l}
\]

(2)

Continuity of mass suggests that

\[
w \sim \frac{u_g d}{l}
\]

(3)

where \( d \) is the depth of influence of the circulation, assumed to be of the order of a few hundred meters, somewhere in the stratified fluid beneath.

The ageostrophic circulation induces displacement of the base of the low PV layer (‘mixed layer’ depth) in the vicinity of the front, and horizontal advection of it, in the sense indicated in Figure 11. The magnitude of the adnuations induced by the axial circulation can be estimated thus:

\[
\Delta y \sim v \delta t \sim \frac{u'}{f}
\]

(4)

where \( \Delta y \) is the eddy induced displacement in mixed layer depth, \( \Delta y \) the eddy induced horizontal displacement and \( \delta t \sim l/u' \).

2.1 NUMERICAL ESTIMATES

Suppose that a typical flow cross-section is as shown in Figure 11. The magnitude of the key parameters may be obtained from in situ observations, such as those obtained in FASINEX, for a mid-ocean front, assuming:

\[
\begin{align*}
u' & \sim 0.3 m s^{-1} \\
f & \sim 10^{-4} s^{-1} \\
l & \sim 50 km \\
d & \sim 250 m
\end{align*}
\]

we find \( u_g \sim 0.02 m s^{-1} \) and \( w \sim 10^{-4} m s^{-1} \) (10 m day\(^{-1}\)), which compare very well with the observations of Pollard and Regier (1992). We estimate the
vertical and horizontal displacements of the boundary between low and high PV water masses, due to the axial ageostrophic flow, to be: \( \Delta h \sim 20 \text{ m}, \Delta y \sim 3000 \text{ m} \). These pocket dimensions are consistent with the “elongated tubes” observed in FASINEX (Pollard and Regier, 1992).

3. FLUID EXCHANGE ACROSS THE FRONT

The above description of distortions of the mixed-layer base is consistent with observations of ocean fronts but does not imply any irreversible exchange of water across it. There are a number of possible processes which act to control that fraction of each pocket, or fraction of the number of pockets, which becomes irreversibly entrained into the region of opposite PV anomaly. Pollard and Regier (1992) discuss some of these processes. They are poorly understood, but critical in determining the subduction/entrainment rates due to frontal scale processes. We employ a ‘dilution factor’, \( 0 \leq D \leq 1 \), to represent the efficiency of such processes to subduct/entrain the pockets of water.

There are three potentially significant processes, which we describe in, what we consider to be, order of importance:

(i) The separation of the pockets from their parent water mass during the breaking of the baroclinic wave. This may provide exchange in either or both directions. The permanent displacement of the parcels provides a consistent reduction of the larger scale available potential energy. Better understanding of the details of wave breaking are required to understand the influence of this process. Nakamura and Plumb (1993) find, using contour dynamical techniques in single layer and 1 1/2 layer systems, very complex wave breaking behaviour. Finite amplitude perturbations are found to grow and break in a manner which depends on the asymmetries in the cross-stream shear. The waves have a propensity to break on the side of the jet axis that possesses the greatest lateral shear. In Figure 13, we illustrate schematically a wave breaking event and the subsequent mass of isolated low PV water, subducted into the thermocline.

(ii) Convective readjustment of the mixed layer base may be enhanced in the upwelling branch of the vertical circulation, where the intervening, weakly stratified mixed layer is of reduced depth. This may facilitate the entrainment of water from the underlying thermocline into the mixed layer, implying an asymmetry in the action of the “dilution process” and therefore \( D \).

(iii) Small scale turbulent diffusion processes, enhanced by the strong frontal gradients (i.e. double diffusion) may cause some fraction of the intrusive pockets to be irreversibly mixed. For a mixing coefficient appropriate to such processes, \( K_{dd} \sim 10^{-5} \text{ m}^2 \text{ s}^{-1} \) (Schmitt, 1981), we can estimate the timescale for absorption of a pocket to be

\[
\Delta t \sim \frac{(\Delta z)^2}{K_{dd}}
\]

If \( \Delta z \sim \Delta h_{mb} \), then this timescale is of the order of a year. The time it takes for a particle to flow through the pocket is \( \delta t \sim l/\nu' \), only a few days. This implies a dilution factor (appropriate to this process) \( D \sim \delta t/\Delta t \sim 10^{-2} \). Thus, dilution of the pockets by small scale double diffusive turbulence is unlikely to be significant in comparison to processes (i) and (ii).

4. ESTIMATES OF THE INTEGRAL EFFECT OF FRONTAL EXCHANGE

Subduction rates induced locally by frontal processes can be very large. But we are interested in assessing the importance of frontal processes in controlling exchange rates on the large scale. In this section we attempt to estimate an upper limit on the eddy-induced subduction rates. Given the large uncertainties in the dilution factor, \( D \), we produce estimates for the limit \( D \sim 1 \).

Using the scale estimates of velocities and displacements from Section 2, we can estimate the flux of water into the ‘pockets’ of the distorted PV interface. This flux is assumed to be the product of the horizontal speed of flow into the pocket, and the vertical cross sectional area of the mouth of the pocket, following the method of Follows and Austin (1992). Multiplied by the dilution factor, \( D \), this flux determines the exchange rate (subduction/entrainment) of water, \( E \), for one such eddy. Thus:

\[
E \sim D \Delta h \Delta \nu \sim D \frac{\nu^3 d}{f^3 l}
\]

Using parameter scales as in Section 3, appropriate for an ocean frontal region, we estimate a local exchange rate of fluid, over a single eddy, of \( D \times 14000 \text{ m}^3 \text{ s}^{-1} \). Clearly the estimate is very sensitive to the magnitude of \( D \).

It is possible to reinterpret the exchange rate, \( E \), in terms of a horizontal eddy transfer coefficient, \( K \), since we can consider the exchange process as the horizontal displacement of eddy induced anomalies in the depth of the mixed layer. Reinterpreting (6) in this way we can write

\[
E \sim D \Delta h \Delta \nu \sim l \alpha K
\]
where \( \alpha = d/l \sim \Delta h/\Delta y \) is the aspect ratio of the transfer process; the slope of the isentropes near the front, and the eddy transfer coefficient is

\[
K \sim D u_{ag} \Delta y \sim D \frac{u^3}{f^2 l^2} .
\]

\( K \) is estimated to be of order 50 m² s⁻¹ for the typical parameters used in section 3, and assuming the limit \( D \sim 1 \). This coefficient is an order smaller than that usually ascribed to stirring on isopycnal surfaces by geostrophic eddies in the ocean. The coefficient determined here characterises the exchange of fluid across a PV discontinuity in the ocean, and so may be expected to be somewhat smaller.

What does this exchange rate imply for the large scale subduction and entrainment rates? We can extrapolate the estimate to consider an area of ocean, \( A = L^2 \), containing a front (or fronts) of total along front dimension \( C \) (see Figure 14). Let us consider the potential subduction rate from such a system. We will consider some limiting cases to appraise the likely significance of frontal subduction in comparison to other (larger scale) processes. We will assume the maximum possible number of eddies along the front, \( n = C/l \). The area average subduction rate, \( S \), is the product of the exchange rate for one eddy, \( E \), with the number of eddies on the front, \( n \), normalised by the total area of ocean containing the front(s), \( A \):

\[
S \sim \frac{E n}{A} \sim \frac{D n u^3 d}{A f^3 P} .
\]

The first limit considered is that where \( C \sim l \), and \( L \sim l \), giving the local subduction rate for a single eddy. Substituting suitable parameter values (appropriate to the FASINEX front) from above, we find a local subduction rate at the front of order 150 m year⁻¹. We should again note that this assumes the maximising limit \( D \sim 1 \).

Another interesting limit case is the minimum along front dimension, \( C \), in some ocean region of area \( L^2 \), required to support an area average subduction rate comparable to the large scale estimates of Marshall et al (1993). The minimum \( C \) is found if we assume maximum along-front eddy density, \( n = C/l \), and maximum eddy exchange efficiency, \( D \sim 1 \). Rearranging (9) we find

\[
C \sim \frac{S l L^2}{E} .
\]

Substituting appropriate values for the parameters, assuming an area of side \( L = 1000 \) km, and a subduction rate, \( S = 50 \) m year⁻¹, leads to a minimum necessary along front dimension of the order of 5000 km. Thus, in order for such frontal processes as described to make a significant contribution to the subduction rate, we require \( C/L \sim 5 \). In other words, there must be at least 5000 km of active front in an ocean area of 1000 km side. This (minimum) ratio is not totally implausible when considered in the light of the FASINEX data (Halliwell et al., 1991). Unstable ocean fronts may indeed play a significant role in the exchange of mass and properties across the base of the mixed layer.

Such eddy processes may also transfer thermocline water into the mixed layer, depending on the nature of the wave breaking processes which probably control \( D \).

5. SUMMARY AND DISCUSSION

The exchange of fluid across the base of the mixed layer is important in setting the exchange of trace constituents, for example dissolved forms of carbon and nutrients, as well as physical quantities such as \( T/S \), and potential vorticity. Pollard and Regier (1992) show that fronts are places where strong axial circulations may drive fluid to and from the mixed layer. While there is clear observational evidence for the strong vertical circulations, there is as yet little firm evidence for irreversible exchange of fluid across the mixed layer base at ocean fronts.

Here we have considered the mechanisms by which exchange may occur at a front, and estimated associated upper bounds on the local and large scale influence on the exchange rate. We deduce that an active front of many thousands of kilometres is required to drive a large scale subduction rate comparable to that of the mean, gyre-scale circulation. Rates local to the front, however, can greatly exceed these average values.

While we have indicated an upper bound to the exchange rate produced at fronts, firmer estimates will only be possible with significant improvements in our knowledge of the process of Rossby wave breaking at fronts and the frequency of occurrence of eddies and dynamically active fronts in the ocean.

While the eddy driven contribution to exchange may turn out to be considerably less than that achieved by the mean flow, its intermittent nature and localisation at fronts, which may be quasi-stationary features, will have important implications for tracer transports, particularly locally. These effects may manifest themselves, for example in a concentrated transient injection of nutrients into the mixed layer all year round, as opposed to the annual injection in the early spring implied in the mean flow calculations (Marshall et al., 1993). In the modelling of ocean chemistry and biology these issues may be very important.

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AN ENERGY-BALANCE BOUNDARY CONDITION FOR OCEAN CLIMATE MODELS

Stefan Rahmstorf

Traditionally, the surface heat flux in ocean models is calculated using a Newtonian restoring law, \( Q = k(T - T_0) \) (Haney 1971). This implies that the ocean is in contact with a fixed atmospheric state which does not respond in any way to changes in the ocean. In particular, it implies that the atmospheric temperature field is constant in time. For this reason, the Haney boundary condition is not suitable for model experiments involving large scale changes in the ocean circulation. El Niño modellers have been aware of this problem for some time (Schopf 1983), and it equally applies to studies of the variability of the thermohaline circulation.

In reality, the atmospheric state is not independent of the ocean circulation. Oceanic heat transport is a major factor influencing the surface air temperature. For example, the heat transport of the conveyor belt circulation is generally thought to warm the northern North Atlantic by about 4–5°C (in contrast to comparable latitudes of the North Pacific), causing the relatively mild climate of Western Europe. A collapse of this circulation would therefore lead to a 4–5°C drop in surface temperature in the northern North Atlantic region, and this is indeed what coupled GCM experiments predict (Manabe and Stouffer 1988). This temperature drop has important implications for deep water formation and the stability of the deep circulation, but it cannot be modelled in uncoupled ocean models with Haney restoring.

In order to create a thermal boundary condition suitable for climate change experiments with ocean models, we have replaced the assumption of fixed atmospheric temperature, using an atmospheric energy balance model instead. This energy balance model allows the atmospheric temperature to respond in a realistic way to changes in the oceanic heat transport. It can be condensed into the following form:

\[
Q = \gamma(T - T_0) - \mu \mathbf{\nabla}^2 (T - T_0). \]  

(1)

\( Q \) is the heat flux at the ocean surface, \( \gamma \) a radiative relaxation constant, \( \mu \) a constant related to atmospheric heat diffusion, \( T_0 \) the ocean surface temperature and \( T \) is a restoring temperature. For details
FIGURE 11a
(Follows and Marshall)
Schematic of a section of ocean containing a front.

The “unperturbed front”, generated by the large scale wind stress pattern for instance. Thin black lines indicate isopycnals, large solid arrows indicate the sense of the along front geostrophic flow. The thick black line indicates the unperturbed base of the “mixed layer” (low potential vorticity (PV), weakly stratified water near the surface). Beneath this boundary lies the strongly stratified, high PV thermocline.

FIGURE 11b
(Follows and Marshall)
Schematic of a section of ocean containing a front.

The unstable front, which undulates with perturbation length scale, $l$, of order of the deformation radius. Confluent and diffuent regions of the jet, associated with this perturbation length scale, induce ageostrophic circulations in the $y$, $z$ plane, indicated by dashed lines with arrows. These circulations have rapid vertical velocities associated with them (tens of metres per day), inducing perturbations, of magnitude $\Delta h$, $\Delta y$ in the PV interface.
FIGURE 12
(Follows and Marshall)

The displacement of surface isopycnals in a region of confluence/diffuence, of length scale $l$, along a front with straight mid-jet isopycnal. Solid arrows indicate the nature of the along front (geostrophic) flow, while the dashed arrows indicate the nature of the surface horizontal ageostrophic flow, induced by the confluence.
FIGURE 13a
(Follows and Marshall)

Schematic representation of the isopycnal subduction of a pocket of mixed layer (low PV, low stratification) water into the thermocline, following the separation of a cold eddy ring from the jet. This is a result of the breaking of a finite amplitude baroclinic wave. Isopycnal flow associated with the breaking wave drags the pocket of water (formed by the acceleration of the jet streak) into the thermocline.

FIGURE 13b
(Follows and Marshall)

Vertical cross-section following the wave breaking event. A slug of low PV water has been introduced into the thermocline, and separated from the water mass of its origin. Over long timescales the water may become assimilated into its new environment through smaller scale processes.
FIGURE 14
(Follows and Marshall)
Meandering fronts, total along front length $C$, in an ocean area, $L^2$. The length scale of the undulations, $l$, is assumed to be of order of the radius of deformation.