2.2 DESTRUCTION OF POTENTIAL VORTICITY BY WINDS

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1 INTRODUCTION

In the upper-ocean mixed layer, the potential vorticity (PV) of surface waters is modified by frictional and diabatic processes until the fluid is subducted into the nearly inviscid, adiabatic interior. A particular type of water mass with PV strongly modified by upper-ocean processes that affects gyre-scale flows is mode water (Hanawa and Talley, 2000). Mode water is characterized by low PV and is associated with ocean fronts. Mode water formation is commonly attributed to diabatic processes driven by wintertime buoyancy loss from the ocean surface. In this talk it will be shown that frictional forces associated with winds can be as effective, if not more, as buoyancy loss at destroying PV at ocean fronts, and therefore may play an important role in the formation of mode water.

2 POTENTIAL VORTICITY DYNAMICS

Under what atmospheric forcing conditions is the PV of the upper ocean reduced? To address this question and to understand the dynamics of the PV, it is instructive to derive a flux form for the PV equation similar to that described in Marshall and Nurser (1992). Changes in the PV

$$q = f\boldsymbol{\omega}_a \cdot \nabla b, \tag{1}$$

where $\boldsymbol{\omega}_a = f\hat{k} + \nabla \times \mathbf{u}$ is the absolute vorticity (f is the Coriolis parameter, $\hat{k}$ is the vertical unit vector, and $\mathbf{u}$ is the velocity of the fluid) and $b = -g\rho/\rho_o$ is the buoyancy (g is the gravitational acceleration, ρ is the density, and ρ_o is a reference density), result from convergences/divergences of the PV flux, i.e.

$$\frac{\partial q}{\partial t} = -\nabla \cdot \mathbf{J}, \tag{2}$$

where the PV flux

$$\mathbf{J} = q\mathbf{u} + f\nabla b \times \mathbf{F} - f\mathscr{D}\boldsymbol{\omega}_a \tag{3}$$

has an advective constituent $q\mathbf{u}$ and nonadvective constituents that arise from diabatic processes $\mathscr{D} = \partial b/\partial t + \mathbf{u}\cdot\nabla \mathbf{b}$ and from frictional or nonconservative body forces $\mathbf{F}$.

To determine what conditions are favorable for PV destruction at a front, consider integrating (2) over a control volume that encircles the front. Application of Gauss' theorem to the integral reveals that reduction of the volume-averaged PV occurs when the net flux of PV through the surface of the control volume is outward. Consider the control volume $\mathscr{V}$ shown in Fig. 1, with an upper surface that coincides with the air-sea interface, side surfaces that are the isopycnals bounding the frontal zone, and a bottom surface that crosses isopycnals at a depth where the flow, frictional forces, diabatic processes and hence PV fluxes are weak. For such a volume, only the PV flux out of the sea-surface contributes to the change of volume-averaged PV since no PV is fluxed through the isopycnal surfaces of the control volume, in accordance with the "impermeability theorem" of Haynes and McIntyre (1987). Therefore, the PV in the control volume will be reduced if the PV flux at the sea-surface is upward.

At the sea-surface, in the limit of a rigid lid, the vertical velocity is zero or otherwise weak so that the vertical component of the PV flux is dominated by its nonadvective constituents. Under what atmospheric forcing condition is the nonadvective PV flux at the surface upward? For an inertially stable flow,

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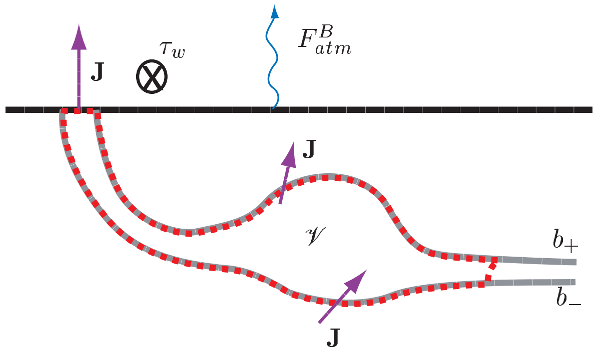

Figure 1: A control volume $\mathscr{V}$ (dashed lines) with side surfaces coincident with isopycnals (grey contours) with buoyancy b_+ and b_- that bound the frontal zone. The "down-front" wind-stress τ_w and flux of buoyancy from the ocean to the atmosphere F^B_{atm} will result in an upward PV flux $\mathbf{J}$ at the sea-surface.

where $f(f + \hat{k} \cdot \nabla \times \mathbf{u}) > 0$, diabatic processes will result in an upward PV flux

$$J^{\mathscr{D}}_z = -f(f + \hat{k} \cdot \nabla \times \mathbf{u})\mathscr{D} \qquad (4)$$

when $\mathscr{D} < 0$. Diabatic processes will reduce the buoyancy, i.e. $\mathscr{D} < 0$, when there is a net loss of buoyancy from the ocean to the atmosphere. This is not surprising, since buoyancy loss to the atmosphere triggers convective mixing which reduces the stratification. The vertical component of the nonadvective PV flux associated with frictional forces

$$J^F_z = f \nabla_h b \times \mathbf{F} \qquad (5)$$

is nonzero only if there is a horizontal buoyancy gradient $\nabla_h b$. Associating the horizontal buoyancy gradient with a vertically sheared geostrophic flow $\mathbf{u}_g$ via the thermal wind relation $\nabla_h b = f \partial \mathbf{u}_g / \partial z \times \hat{k}$ reveals that the nonadvective PV flux associated with frictional forces

$$J^F_z = f^2 \frac{\partial \mathbf{u}_g}{\partial z} \cdot \mathbf{F} \qquad (6)$$

is upward when the frictional force is in the direction of the geostrophic shear. For wind-forced flows, the frictional force is dominantly in the direction of the wind-stress. Currents at upper-ocean fronts are usually surface intensified so that the surface current of the front is oriented with the geostrophic shear. Therefore, when the wind blows in the direction of the frontal jet, i.e is in a "down-front" orientation, the conditions are favorable for PV destruction by frictional forces (i.e. $J^F_z > 0$).

What are the relative contributions to the PV reduction at a front by diabatic and frictional effects? To address this issue, consider the following scaling argument. Under conditions of atmospheric buoyancy loss, gravitational instability will result in a convective buoyancy flux that scales with the buoyancy loss to the atmosphere F^B_{atm} and that decays nearly linearly with depth through the mixed layer of thickness H. Therefore an appropriate scale for diabatic processes induced by buoyancy loss to the atmosphere is $\mathscr{D} \sim F^B_{atm}/H$, so that the scaling for (4) is

$$J^{\mathscr{D}}_z \sim f^2 \frac{F^B_{atm}}{H}. \qquad (7)$$

A wind-stress of strength τ_o will exert a frictional force of scale $\mathrm{F} \sim \tau_o/\rho_o\delta_e$ over the Ekman layer of thickness δ_e. If the magnitude of the horizontal buoyancy gradient at the front is $S^2 = |\nabla_h b|$, then the appropriate scaling for (5) is

$$J^F_z \sim f S^2 \frac{\tau_o}{\rho_o \delta_e}. \qquad (8)$$

The relative contributions of friction and diabatic effects to PV reduction can be assessed by taking the ratio of (8) and (7)

$$\frac{J^F_z}{J^{\mathscr{D}}_z} \sim \left(\frac{H}{\delta_e}\right)\left(\frac{F^B_{wind}}{F^B_{atm}}\right), \qquad (9)$$

where $F^B_{wind} = S^2\tau_o/\rho_o f$ is a wind-driven buoyancy flux (WDBF) representing the flux of buoyancy exiting the fluid beneath the Ekman layer.

On account of the strong lateral density gradients, at wind-forced fronts, the WDBF can be comparable or larger than the atmospheric buoyancy loss, making the ratio (9) greater than or equal to one. This even holds at subpolar fronts in the northern hemisphere forced by wintertime cold-air outbreaks that extract large amounts of heat from the ocean yet also force the fronts strongly with "down-front" wind-stress. For example, using hydrographic measurements combined with shipboard meteorological data taken at the subpolar front of the Japan/East Sea during a cold-air outbreak Thomas and Lee (2005) estimated the WDBF to be an order of magnitude larger than the atmospheric buoyancy flux even with the heat loss of ~ 500 W m^{-2} experienced at the front. The influence of wind-driven PV reduction at fronts may be even more pronounced in the Southern Ocean where fronts are forced by strong “down-front” winds and relatively weak air-sea fluxes of heat and freshwater (Rintoul and England, 2002). In summary, these simple scaling arguments suggest that for fronts forced by “down-front” winds, PV reduction by wind-stress can be at least as important as that due to heat loss and therefore should be accounted for in the formation of mode waters.

3 NUMERICAL EXPERIMENTS

In the previous section, it was shown that forcing of fronts by “down-front” winds leads to a reduction of the volume-averaged PV. However, the utility of volume averages is limited because they do not shed light on the detailed structure of the PV and how it is destroyed. In order to investigate in detail how frictional forces by “down-front” winds destroy PV at a front, two-dimensional, high-resolution, numerical experiments of a wind-forced frontal zone were performed using the Regional Oceanic Modeling System (ROMS).

For the numerical experiments, the idealized frontal zone configuration of Thomas and Lee (2005) was used. The frontal zone consists of a laterally-homogeneous, baroclinic, zonal flow $U = S^2/fz$ in a thermal wind balance with a buoyancy field B characterized by uniform lateral gradient $\partial B/\partial y = -S^2$ and a vertically varying stratification that is constant in a layer from the surface to a depth H and that increases linearly beneath this depth. The increase in stratification beneath $z = -H$ is meant to be a simple representation of a pycnocline, a region of high PV that can be eroded away by PV-reducing wind-forcing. The frontal zone is forced by a spatially uniform “down-front” wind-stress of strength τ_o turned on impulsively at $t = 0$. The domain width and depth of the experiments was $L = 12000$ m and $D = 250$ m, respectively. So as to adequately resolve frontal features and the Ekman layer, horizontal and vertical grid spacings of $\Delta y = 120$ m and $\Delta z = 5$ m were used.

A solution representative of the numerical experiments is shown in Fig 2. As illustrated in the

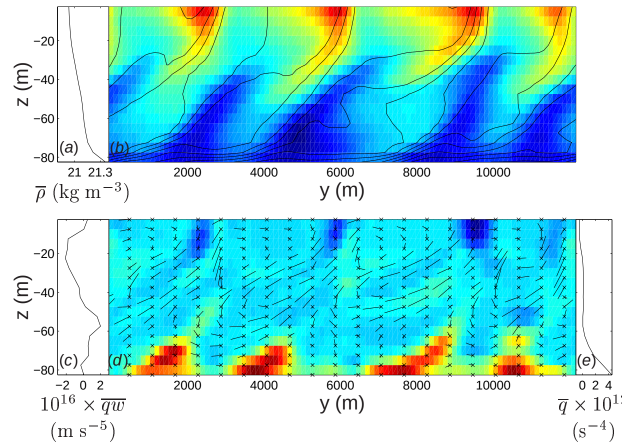

Figure 2: Solutions from a numerical experiment at $t = 8.6$ inertial periods. (*a*) The meridionally averaged density (black); (*b*) zonal velocity (shades) and density (contours) fields; (*c*) the meridionally averaged advective PV flux; (*d*) PV (shades) and secondary circulation (vectors); and (*e*) the meridionally averaged PV. In both (*b*) and (*d*), warm (cool) shades indicate higher (lower) values.

meridional section of the density and zonal velocity (Fig. 2*b*), forcing of the frontal-zone by “down-front” winds leads to the formation of multiple fronts with sharp, frontal jets. The location of the fronts is marked by a frontal interface of enhanced lateral and vertical buoyancy gradients. The fronts form within several inertial periods as a result of frontogenetic secondary circulations characterized by upwelling along the frontal interface and downwelling of the Ekman flow (i.e. southward flow in the upper 20 m in between the fronts) down the dense side of the front, Fig. 2*d*.

As evident in Fig. 2*d*, the secondary circulation is most intense above the strongly-stratified pycnocline, yet exists in a region where the stratification of the meridionally averaged density $\bar{\rho} = -\rho_o/g\bar{b}$ (where $\overline{(\ )} = (1/L)\int_0^L \ dy$) is non-zero (Fig. 2*a*). The region that the overturning motions occupy is therefore not characterized by a homogeneous density layer as with typical mixed layers, but is characterized by a “zero-PV layer” with a nearly zero meridionally-averaged PV (Fig. 2*e*). Although it may seem somewhat paradoxical that the fluid has zero PV yet non-zero stratification, it must be realized that for a strongly baroclinic current such as

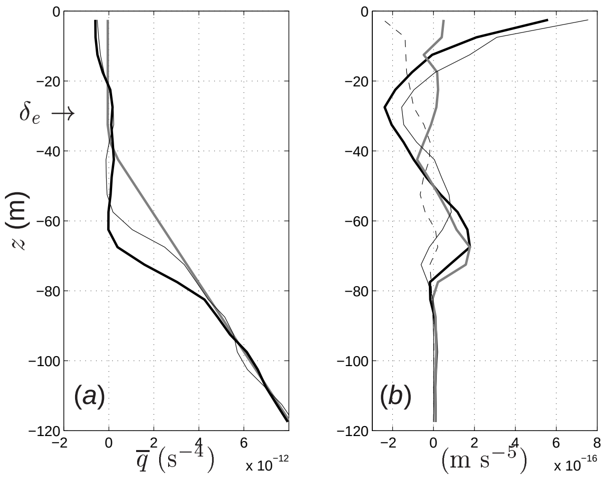

Figure 3: (*a*) Meridionally averaged PV at $t = 0$ (gray), 5 (thin black), and 10 (thick black) inertial periods. (*b*) The meridionally averaged vertical component of the PV flux (thick black) and its constituents: the advective PV flux (gray), the frictional nonadvective flux (thin black), and the nonadvective flux associated with diabatic processes (dashed). All PV fluxes have also been averaged in time from $t = 8 - 10$ inertial periods.

that associated with a front, the horizontal vorticity (i.e. vertical shear) and horizontal buoyancy gradients significantly contribute to the PV (1), so that the combined conditions of $q = 0$ and $\partial b/\partial z > 0$ can be met if $f\boldsymbol{\omega}_a \cdot \nabla_h b < 0$, a condition that is always satisfied for a geostrophically-balanced current.

The "zero-PV layer" deepens with time, as high PV is eroded from the pycnocline (Fig. 3*a*). This erosion of PV indicates a net reduction of the total PV in the domain, a result that was to be expected since frictional forces by "down-front" winds drive a net PV flux out of the surface of the ocean and hence reduce the volume averaged PV. PV destruction occurs at the base of the "zero-PV layer" and not within the Ekman layer where frictional forces are expected to be largest, therefore, it is not obvious that friction is directly responsible for the fluxing of PV out of the pycnocline and the deepening of the layer. To determine exactly what mechanism is responsible for the PV flux out of the pycnocline, the meridionally averaged, vertical component of the advective and nonadvective constituents of the PV flux were calculated (Fig. 3*b*). The figure illustrates that the total PV flux is upward at the base of the "zero-PV layer" and decays rapidly with depth, yielding a divergent flux that therefore decreases the PV. At this depth, the PV flux is dominated by the advective PV flux, indicating that it is the secondary circulation and not friction, nor diabatic processes that is responsible for the deepening of the "zero-PV layer". The way in which the secondary circulation reduces the PV of the pycnocline is illustrated in Fig. 2*d*. Near the base of the "zero-PV layer", upward motions advect high PV from the pycnocline while downward motions draw low PV waters from the surface into the pycnocline. This leads to a positive correlation between the PV and the vertical velocity, yielding an upward advective PV flux $\overline{qw} > 0$ (Fig. 2*c*) that extracts PV from the pycnocline and deepens the "zero-PV layer." Although the frictional nonadvective PV flux plays only a minor role at the base of the "zero-PV layer", in the Ekman layer its value is large and positive, resulting in PV flux divergence that maintains the low PV values at the surface. This frictionally induced source of low PV is necessary to sustain the deepening of the "zero-PV layer", since the secondary circulation acts only to exchange PV vertically and therefore cannot decrease the volume-averaged PV of the frontal zone. For the experiments presented here, diabatic processes do not contribute greatly to the total PV flux, a consequence of the lack of forcing by atmospheric buoyancy fluxes.

The solution shown in Fig.2 demonstrates that the extraction of PV from the pycnocline by the advective PV flux is responsible for the deepening of the "zero-PV layer." To determine how this result depends on the relevant parameters of the flow, namely, the strength of the wind-stress τ_o, lateral buoyancy gradient of the frontal zone S^2, and the Coriolis parameter f, ten experiments with different values of τ_o, S^2, and f were performed. The key parameter for PV destruction by winds is the scaling of the frictional nonadvective flux (8). This is illustrated in Fig. 4, that shows that for all the experiments, the advective PV flux at the base of the "zero-PV layer" increases linearly with $fS^2\tau_o/(\rho_o\delta_e)$, being smaller than this scaling by about 80%.

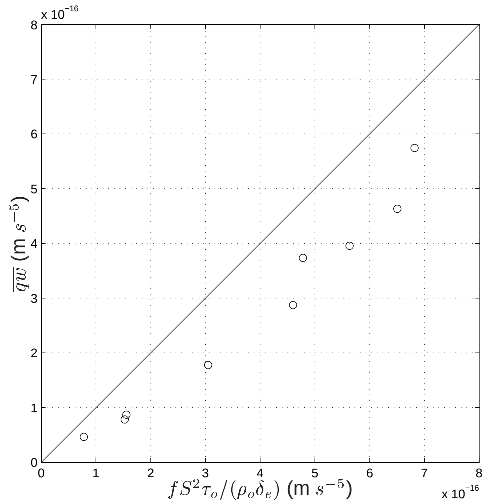

Figure 4: Comparison of the advective PV flux $\overline{qw}$ (circles) with the scaling parameter for the frictional nonadvective PV flux (8). The advective PV flux is evaluated at its maximum near the base of the zero PV layer and has been averaged in time from $t = 8 - 10$ inertial periods. A line with slope one is also plotted for reference.

4 CONCLUSIONS

Many ocean fronts such as those associated with the Gulf Stream and Kuroshio currents, coastal upwelling systems, and the Antarctic Circumpolar Current are forced by winds that blow in the direction of the frontal jet. Forcing a front by "down-front" winds drives a nonadvective frictional PV flux that is upward at the sea surface. This PV flux extracts PV from the fluid and leads to the formation of an oceanic boundary layer with nearly zero PV and non-zero stratification. Diabatic processes associated with buoyancy loss from the ocean to the atmosphere also extract PV from the ocean, but on account of the surface frictional PV flux's dependence on the lateral density gradient, at wind-forced fronts the PV flux driven by friction can be comparable or larger than that due to buoyancy loss.

The PV reducing effect of the frictional PV flux directly affects the Ekman layer but is transmitted through the "zero-PV layer" via secondary circulations that exchange low PV from the Ekman layer with high PV from the pycnocline. Averaged over the frontal-zone, this PV exchange yields an upward advective PV flux that peaks at the base of the "zero-PV layer" and extracts PV out of the pycnocline, deepening the boundary layer. The strength of the advective PV flux and the PV deficit in the pycnocline scales with $fS^2\tau_o/(\rho_o\delta_e)$, a quantity that depends on parameters that do not vary over lengthscales smaller than the frontal zone. Therefore, it should be possible to parameterize PV destruction by winds in coarse grid numerical models that cannot resolve the order kilometer-width, frontogenetic secondary circulations but can resolve the frontal zone. A possible parameterization might involve an algorithm that: one, sets the PV in the boundary layer to zero; two, modifies the PV in the pycnocline by applying a divergent PV flux that scales with $fS^2\tau_o/(\rho_o\delta_e)$ and has a prescribed vertical shape function; and three, adjusts the depth of the "zero-PV layer" depending on how much PV has been eroded from the pycnocline.

References

Hanawa, K. and Talley, L. D. (2000). Mode waters. In Siedler, G. and Church, J., editors, *Ocean Circulation and Climate*. Academic Press.

Haynes, P. and McIntyre, M. (1987). On the evolution of vorticity and potential vorticity in the presence of diabatic heating and frictional or other forces. *J. Atmos. Sci.*, 44.

Marshall, J. C. and Nurser, A. J. G. (1992). Fluid dynamics of oceanic thermocline ventilation. *J. Phys. Oceanogr.*, 22:583–595.

Rintoul, S. R. and England, M. H. (2002). Ekman transport dominates local air-sea fluxes in driving variability of Subantarctic Mode Water. *J. Phys. Oceanogr.*, 32:1308–1321.

Thomas, L. N. and Lee, C. M. (2005). Intensification of ocean fronts by down-front winds. *J. Phys. Oceanogr.* in press.