NOTES AND CORRESPONDENCE

Ventilation of Eastern Subtropical Gyres

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ABSTRACT

A simple mechanism for the generation of flow in the eastern region of the oceans is presented. It is shown that even a small amount of upwelling-favorable wind stress parallel to the boundary makes the subducted geostrophic contours veer into the eastern boundary. This sweep toward the wall ventilates the top few hundred meters of the eastern subtropical gyre. The latitudinal extent of the eastern ventilation is calculated. Along a meridional boundary the subsurface flow is eastward and appears to provide the mass source for the poleward undercurrents observed at the eastern boundary of the ocean.

In this short communication, we examine the conditions for the existence of a "shadow zone" in the eastern limb of a subtropical gyre. The shadow zone is a central ingredient of both the Rhines and Young (1982) and the Luyten et al. (1983; hereafter LPS) circulation theories. In both of these models it is a motionless, unventilated region whose geostrophic contours connect to the eastern boundary.

In LPS, the trajectory of a particle is followed southward from its latitude of subduction. LPS argue that there is a zone defined by "blocked" geostrophic contours within which the fluid is stagnant. The blocked contours emanate from the eastern boundary and are directed westward and southward. The argument works as long as the geostrophic contours do indeed veer westward as they subduct. In the following, we show that even an infinitesimal amount of equatorward wind stress causes the geostrophic contours subducted near the eastern boundary to veer *eastward,* that is, into the wall. The return flow associated with the southeastward sweep of the subducted trajectories can support eastern boundary currents with mass transport.

Consider the dynamics of a single moving layer of constant density ρ_2 and thickness h_2 overlying a motionless layer of density ρ_3. North of the subduction latitude, $y = y_0$, the flow is forced by a prescribed wind stress and is governed by

$$fu_2 = \tau^y/h_2 - g''\partial_y h_2,$$
$$-fv_2 = \tau^x/h_2 - g''\partial_x h_2,$$
$$0 = \partial_x(h_2u_2) + \partial_y(h_2v_2), \qquad (1.1)$$

where $g'' \equiv g(\rho_3 - \rho_2)/\rho_2$. Notice that we have included the Ekman flow in the velocity field for this layer. Some authors prefer to keep the Ekman flow separate from the upper-layer flow, but we treat them together. The formulation (1.1) assumes that inertia can be neglected so that the vorticity equation is

$$\frac{\beta}{2f}g''\partial_x h_2^2 = -f\partial_y\left(\frac{\tau^x}{f}\right) + \partial_x\tau^y. \qquad (1.2)$$

The requirement of no-normal flow through the eastern wall determines the depth of the layer at that location up to a constant. Specifically, evaluating (1.1a) at $x = x_e$, one has

$$g''\partial_y h_2^2(x_e, y) = 2\tau^y(x_e, y). \qquad (1.3)$$

The balance (1.3) states that the pressure gradient at the coast is proportional to the wind stress. The neglect of the meridional wind stress in LPS implies that the depth of the interface is constant along the eastern boundary. Here, equatorward wind stress piles up the water toward the equator, and the depth of the (only) moving layer increases in that direction. Tide gauge measurements along the North Pacific coast confirm that the latitudinal gradient of sea level elevation is correlated with the meridional wind stress, in qualitative agreement with (1.3) (Enfield and Allen 1980).

Finally, with the additional (but not essential) simplification that τ^x is independent of x and τ^y is independent of y, we obtain

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$$g''\frac{h_2^2}{2} = (x_e - x)\frac{f^2}{\beta}\partial_y\left(\frac{\tau^x}{f}\right) + y\tau^y(x) - y_0\tau^y(x_e) + g''\frac{H_0^2}{2}. \quad (1.4)$$

The arbitrary constant value of h_2 in the "corner" $(x, y) = (x_e, y_0)$ is indicated by H_0. The expression (1.4) reduces to that in LPS [their Eq. (2.9)] when $\tau^y = 0$. Notice that the zonal flow, u_2, obtained by substituting (1.4) into (1.1a), is actually independent of τ^y and is therefore identical to that in LPS. The meridional flow, given in (1.2), is proportional to the curl of the wind stress and is dependent on $\partial_x\tau^y$.

At $y = y_0$, the water of density ρ_2 is subducted under a layer of density ρ_1. Thus, the dynamics of this layer south of y_0 are described by

$$fu_2 = -g''\partial_y h,$$
$$-fv_2 = -g''\partial_x h,$$
$$0 = \partial_x(h_2u_2) + \partial_y(h_2v_2), \quad (1.5)$$

where $h \equiv h_1 + h_2$ is used to indicate the depth of the subducted layer. Because the lower layer is not forced, it conserves potential vorticity and Bernoulli function, and the relation between these conserved quantities is determined at subduction:

$$\frac{f}{h_2} = \frac{f_0}{h}, \quad (1.6)$$

where $f_0 \equiv \beta y_0$ is the rotation rate at the subduction latitude. Of course (1.6) applies only to those potential vorticity contours that intersect the outcrop latitude, y_0. To calculate the path of the geostrophic contours, the upper-layer flow must be determined. The upper layer is directly forced by the wind stress, so for $y < y_0$,

$$fu_1 = \tau^y/h_1 - \partial_y(g''h + g'h_1),$$
$$-fv_1 = \tau^x/h_1 - \partial_x(g''h + g'h_1),$$
$$0 = \partial_x(h_1u_1) + \partial_y(h_1v_1), \quad (1.7)$$

where $g' \equiv g(\rho_2 - \rho_1)/\rho_2$. Substituting (1.6) into (1.7) and eliminating u_1 and v_1 gives an equation in the single variable h

$$[g'' + g'(1 - f/f_0)^2]h\partial_x h = \frac{f}{\beta}\partial_x\tau^y - \frac{f^2}{\beta}\partial_y\left(\frac{\tau^x}{f}\right). \quad (1.8)$$

In the subtropical gyre both terms on the right-hand side are negative: the meridional wind stress τ^y is negative and its magnitude increases eastward. Thus, the depth-integrated meridional transport is negative. The requirement of no depth-integrated flow through the eastern wall yields

$$\partial_y[g'' + g'(1 - f/f_0)^2]h^2(x_e, y) = 2\tau^y(x_e) \quad (1.9)$$

so that integration of (1.8) and (1.9) finally gives

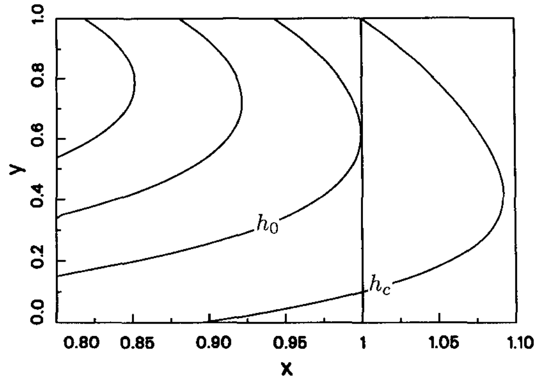

FIG. 1. Lines of constant h south of the subduction line. The y axis is in units of y_0 and the x axis in units of $g'H_0^2\,\tau^{-x}$. There is a critical contour entering the eastern boundary that separates the ventilated region from the shadow zone. North at the subduction latitude of the critical contour $h = h_c$, the subducted water is in motion and flows into the boundary, south of h_c is the lower layer shadow zone.

$$[g'' + g'(1 - f/f_0)^2]\frac{h^2}{2} = (x_e - x)\frac{f^2}{\beta}\partial_y\left(\frac{\tau^x}{f}\right) + y\tau^y(x) - y_0\tau^y(x_e) + g''\frac{H_0^2}{2}. \quad (1.10)$$

It can be checked that h, given in (1.10), is continuous at $y = y_0$ with h_2, given in (1.4). It also reduces to the expression obtained by LPS [their Eq. (2.20)] when $\tau^y = 0$.

With a meridional wind stress, it is not possible to satisfy the no-flow condition in each layer separately. Indeed, the zonal flow at the corner $(x, y) = (x_e, y_0)$ is given by

$$f_0u_1h_1 = \tau^y(x_e),$$
$$f_0u_2h_2 = -\tau^y(x_e). \quad (1.11)$$

The reason for this failure can be understood by examining the lower-layer geostrophic contours, which are parallel to lines of constant h. The geometry of the geostrophic contours near the eastern boundary can be obtained by a Taylor expansion of the meridional wind stress in the vicinity of that longitude: $\tau^y(x) \approx \tau^y(x_e) + (x - x_e)\partial_y\tau^y$. Equation (1.10) then gives

$$(x - x_e)\left[y\partial_x\tau^y - \frac{f^2}{\beta}\partial_y\left(\frac{\tau^x}{f}\right)\right] = g'\left(1 - \frac{f}{f_0}\right)^2\frac{h^2}{2} + \frac{g''}{2}(h^2 - H_0^2) + (y_0 - y)\tau^y(x_e). \quad (1.12)$$

In the vicinity of the outcrop the first term on the right-hand side is negligible, and if τ^y is negative, lines of constant h sweep southeastward, as illustrated in Fig. 1. Specifically, there is a critical contour $h = h_c$ that enters the eastern boundary at the outcrop and then reemerges from it farther to the south. In the region to the north of the geostrophic contour $h = h_c$ the subducted flow is into the wall and mass balance cannot be satisfied. Notice that the critical contour separating the motionless region from the ventilated one is not the contour $h = h_0$ tangent to the eastern boundary. The expression for the latitude y_c at which the geostrophic contour $h = h_c$ emerges from the eastern boundary is given by

$$y_c \approx y_0\left(1 + \frac{2\tau^y y_0}{g' H_0^2}\right). \quad (1.13)$$

In LPS, y_c coincides with the subduction latitude, y_0, and the shadow zone extends from y_0 to the equator. With equatorward wind stress the shadow zone must begin south of y_c, where geostrophic contours emerge from the eastern wall. The Hellerman and Rosenstein (1983) maps of monthly wind stress indicate that the annual average of τ^y is equatorward near the eastern boundaries of the subtropical oceans (except in the northern Indian Ocean). The magnitude varies considerably from ocean to ocean, but it is no less than about 0.5×10^{-4} m^2 s^{-2}. If $H_0 = 200$ m, $y_0 = 3000$ km, and $g' = 10^{-2}$ m s^{-2}, then y_c is a considerable fraction of y_0: there is no shadow zone in the northern portion of the eastern subtropical gyre, and the subducted Sverdrup flow cannot satisfy the mass balance at the solid boundary.

A simple calculation illustrates that whenever the component of the wind stress parallel to the coast is upwelling favorable, then the mass balance cannot be satisfied at every depth within the context of Sverdrup dynamics. For instance, now consider the wind stress to be perfectly zonal, that is, $\tau^y = 0$, but tilt the coast at an angle θ with respect to north (positive anticlockwise). The eastern boundary is then

$$x = x_e - y \tan\theta. \quad (1.14)$$

If we require that there is no depth-integrated flow across the boundary, the depth of the moving water south of the subduction latitude is given by

$$\left[g'' + g'\left(1 - \frac{f}{f_0}\right)^2\right]\frac{h^2}{2} = (x_e - x - y\tan\theta)\frac{f^2}{\beta}\partial_y\left(\frac{\tau^x}{f}\right) - \tan\theta\int_{y_0}^{y}\tau^x(\eta)d\eta + g''\frac{H_0^2}{2}. \quad (1.15)$$

The lower-layer flow normal to the eastern boundary is

$$u_{2n} \equiv u_2\cos\theta + v_2\sin\theta. \quad (1.16)$$

Near the subduction latitude and at the eastern boundary, the subducted transport normal to the solid wall is approximately given by

$$f_0 h_2 u_{2n} \approx \sin\theta(\tau^x(y_0) - y_0^2\partial_y(\tau^x/y)|_{y_0}). \quad (1.17)$$

The second term on the right-hand side is the lower-layer contribution to the meridional Sverdrup transport. Unless the Ekman pumping vanishes at the "eastern" boundary, the condition of no-normal flow cannot be enforced in each separate layer. On the other hand, unless the boundary is exactly zonal ($\theta = \pi/2$), the vertically integrated transport vanishes at the "eastern" wall. The first term on the right-hand side is the transport due to the alongshore component of the wind stress; it is into the wall when τ^x and the coast tilt, θ, have coincident signs; that is, when the wind stress is upwelling favorable.

The failure of Sverdrup dynamics to accomodate the mass balance is a familiar result in eastern boundary dynamics with upwelling-favorable wind stress (e.g., Salmon 1987). This is only problematic for the baroclinic component of the flow: when a single layer is active, the no-flow condition at the wall can be satisfied [see (1.3) and the surrounding discussion].

In traditional models of eastern boundary flows, subsurface motion is set up by cross-isopycnal diffusion of density (McCreary 1981; Salmon 1987): in a layer model this is equivalent to mass exchange between layers. In the model presented here we exclude diapycnal exchange, and the lower layer is set in motion by ventilation along those geostrophic contours that trace back to the subduction latitude.

The calculations presented here show that whenever the wind stress has an upwelling-favorable component parallel to the eastern boundary the propagation of information is toward the boundary in the ventilated region and mass balance cannot be satisfied within the context of Sverdrup dynamics. The subsurface water flowing into the boundary is returned in a non-Sverdrupian boundary current, which is identified with the poleward eastern undercurrent observed in most oceans.

In summary, the concept of "shadow zone" or "blocked" geostrophic contours must be revised in the vicinity of the outcrop whenever the wind stress has a component parallel to the boundary that is upwelling favorable. Because the geostrophic contours veer into the solid boundary, this region is ventilated and not blocked. This result has important consequences for the ventilated thermocline theory because it removes the constraint that all outcropping layers must have zero thickness on the eastern boundary.

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REFERENCES

Enfield, D. B., and J. S. Allen, 1980: On the structure and dynamics of monthly mean sea level anomalies along the Pacific coast of North and South America. *J. Phys. Oceanogr.*, **10**, 557–578.

Hellerman, S., and M. Rosenstein, 1983: Normal monthly wind stress over the world ocean with error estimates. *J. Phys. Oceanogr.*, **13**, 1093–1104.

Luyten, J. R., J. Pedlosky, and H. Stommel, 1983: The ventilated thermocline. *J. Phys. Oceanogr.*, **13**, 292–309.

McCreary, J. P., 1981: A linear stratified ocean model of the coastal undercurrent. *Phil. Trans. Roy. Soc.*, London, **A302**, 385–413.

Rhines, P. B., and W. R. Young, 1982: A theory of wind-driven ocean circulation. I. Mid-ocean gyres. *J. Mar. Res.*, **40**, (Suppl.) 559–596.

Salmon, R., 1987: A simplified ocean circulation theory. *J. Mar. Res.*, **44**, 695–711.