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WIND-DRIVEN CURRENTS IN A LARGE LAKE OR SEA

G. E. Birchfield

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WIND-DRIVEN CURRENTS IN A LARGE LAKE OR SEA

by

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Applied Mathematics Division

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WIND-DRIVEN CURRENTS IN A LARGE LAKE OR SEA

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ABSTRACT

A linear model of the Ekman dynamics for a large shallow homogeneous lake or sea is constructed to examine the mass transport balance, in particular the vertical mass flux. The role of coastal upwelling is examined for steady uniform and nonuniform wind-stress forcing.

INTRODUCTION

The purpose of this report is to construct a greatly simplified model of the wind-driven currents in a large lake or sea in order to examine how the Ekman dynamics bring about horizontal and vertical mass balance. The model incorporates the earth's rotation, and the Rossby number is assumed small enough to neglect the convective terms in the equations of motion. Horizontal and vertical turbulent mixing is incorporated in the simplest way. Because the horizontal scale of wind systems is large relative to the lake size, the model wind stress is taken as either uniform or with uniform curl, and steady in time. The important feature of small mean-depth-to-width ratio of a large lake is imposed on the model. The geometry of the model has, however, been simplified to a rectangular basin of uniform depth.

Questions to be considered concern: (1) how horizontal mass transport is partitioned between the Ekman layer transports and the geostrophic transport; (2) how the surface drift transport normal to a coast is returned to the interior of the basin; (3) how mass flux balance is achieved for a divergent surface Ekman layer; and (4) whether coastal jets appear with a uniform depth and constant Coriolis parameter. These questions and the method used to pursue them are similar to those of Pedlosky* (henceforth referred to as P), who considered an ocean basin. The principal difference here is that the variation of the Coriolis parameter with latitude may be neglected. This significantly alters the dynamics and resulting circulation.

The model used here obviously lacks many important features of a real lake or sea. Because of the simplification of the turbulent effects, the

*J. Pedlosky, An Overlooked Aspect of the Wind-Driven Oceanic Circulation, J. Fluid Mech. **32**, 809-821 (1968).

details of the coastal circulations must be regarded with caution. Sufficient elements of the geophysical situation are incorporated, however, to provide some insight into the necessary viscous constraints on the mass transport away from the coastal regions.

THE DYNAMICAL EQUATIONS

The model lake or sea is a rectangular basin of width L , length ML , and uniform depth D ; the lake is homogeneous with density ρ . It will be assumed that the horizontal dimensions of the lake are large enough for the earth's rotation to be important, but small enough so that the variation of the vertical component of the angular velocity may be neglected. The explicit condition for the latter assumption is discussed below. The equations of motion for steady state are

$$\left. \begin{aligned} uu_x + vu_y + wu_z - fv &= -\frac{p_x}{\rho} + A_H(u_{xx} + u_{yy}) + A_V u_{zz}, \\ uv_x + vv_y + wv_z + fu &= -\frac{p_y}{\rho} + A_H(v_{xx} + v_{yy}) + A_V v_{zz}, \\ uw_x + vw_y + ww_z &= -\frac{p_z}{\rho} + A_H(w_{xx} + w_{yy}) + A_V w_{zz}, \\ u_x + v_y + w_z &= 0, \end{aligned} \right\} \quad (1)$$

and

where

$$f = 2\Omega_0 \left(1 + \frac{\beta_0}{2\Omega_0} y \right)$$

is the Coriolis parameter, Ω_0 and β_0 are constants, and the y -axis is directed north; z increases upward from the basin bottom; and u , v , and w are the velocity components. Eddy viscosity coefficients for the horizontal, A_H , and vertical, A_V , are assumed constant. Currents are generated by a surface wind stress

$$\vec{\tau} = \tau(x, y) \hat{j}, \quad (2)$$

where $\hat{j}$ is a unit vector along the y -axis.

Dimensionless variables are introduced with primes:

$$\left. \begin{aligned}
 (u, v) &= U(u', v'), \quad w = \frac{D}{L} U w', \\
 (x, y) &= L(x', y'), \quad z = Dz', \\
 p &= \rho U L \Omega_0 p', \quad \tau = \tau_0 \tau', \quad \Omega' = 1 + \beta' y', \quad f' \equiv 2\Omega',
 \end{aligned} \right\} \quad (3)$$

and

$$U \equiv \sqrt{1/\Omega_0 A_v \tau_0 / \rho}, \quad \beta' \equiv \beta_0 L / 2\Omega_0,$$

where τ_0 is a characteristic wind stress. The dimensionless equations become (dropping primes)

$$\left. \begin{aligned}
 \epsilon \vec{v} \cdot \nabla u - fv &= -p_x + E_H \nabla_2^2 u + E_v u_{zz}, \\
 \epsilon \vec{v} \cdot \nabla v + fu &= -p_y + E_H \nabla_2^2 v + E_v v_{zz}, \\
 \delta^2 \epsilon \vec{v} \cdot \nabla w &= -p_z + E_H \delta^2 \nabla_2^2 w + E_v \delta^2 w_{zz},
 \end{aligned} \right\} \quad (4)$$

and

$$u_x + v_y + w_z = 0,$$

where $\nabla_2^2 = \partial^2/\partial x^2 + \partial^2/\partial y^2$ and $\vec{v} = u\hat{i} + v\hat{j} + w\hat{k}$. The boundary conditions are

$$\text{at } x = 0, 1; y = 0, M; z = 0: u = v = w = 0; \quad (5a)$$

$$\text{at } z = 1: u_z = 0; v_z = E_v^{-1/2} \tau(x, y); w = 0. \quad (5b)$$

Five nondimensional numbers appear:

$$\epsilon = U/\Omega_0 L, \text{ the Rossby number,}$$

$$E_H = A_H/\Omega_0 L^2, \text{ the horizontal Ekman number,}$$

$$E_v = A_v/\Omega_0 D^2, \text{ the vertical Ekman number,}$$

$$\delta = D/L, \text{ the aspect ratio,}$$

and

$$\beta = \beta_0 L / 2\Omega_0, \text{ nondimensional } df/dy.$$

Only small amplitude motions will be considered; that is, terms multiplied by ϵ will be neglected. As done in P, E_v and E_H will be

replaced by E in view of the lack of reliable information as to their magnitudes. The qualitative character of the flow should not be altered. It is further assumed that L is sufficiently small that

$$\beta \ll E^{1/2} \ll 1$$

may be assumed; that is, to a first approximation, f may be replaced by 2, except possibly where differentiated.

THE SURFACE EKMAN LAYER

Away from the side walls of the basin, we assume the flow variables to be represented by the sum of three parts; letting $\vec{v} = u\hat{i} + v\hat{j}$, we obtain

$$\vec{v} = \vec{v}_1(x, y, z) + \vec{v}_2(x, y, \zeta) + \vec{v}_3(x, y, \zeta'),$$

$$w = w_1(x, y, z) + w_2(x, y, \zeta) + w_3(x, y, \zeta'),$$

and

$$p = p_1(x, y, z) + p_2(x, y, \zeta) + p_3(x, y, \zeta').$$

These represent the interior or inviscid part (subscript 1), the surface Ekman corrections (subscript 2), and the bottom Ekman corrections (subscript 3). The stretched variable in the surface layer is $\zeta \equiv E^{-1/2}(z - 1)$ and at the bottom $\zeta' = E^{-1/2}z$. Each component solution represents an asymptotic series in E , in particular:

$$\left. \begin{aligned} \vec{v}_1(x, y, z) &= \vec{v}_{11} + E^{1/4}\vec{v}_{12} + E^{1/2}\vec{v}_{13} + \dots; & w_1(x, y, z) &= E^{1/2}w_{13} + \dots; \\ p_1(x, y, z) &= p_{11} + E^{1/4}p_{12} + E^{1/2}p_{13} + \dots; \\ \vec{v}_2(x, y, \zeta) &= \vec{v}_{21} + O(E); & w_2(x, y, \zeta) &= E^{1/2}w_{23} + \dots; & p_2 &= O(E); \\ \vec{v}_3(x, y, \zeta') &= \vec{v}_{31} + E^{1/4}\vec{v}_{32} + E^{1/2}\vec{v}_{33} + \dots; & w_3 &= E^{1/2}w_{33} + \dots; \\ p_3 &= O(E). \end{aligned} \right\} \quad (6)$$

The upper Ekman correction solutions are of the standard form and are simply stated:

$$\left. \begin{aligned} u_{21}(x, y, \zeta) &= \frac{1}{2}\tau(x, y)e^{\zeta}(\cos \zeta - \sin \zeta); \\ v_{21}(x, y, \zeta) &= \frac{1}{2}\tau(x, y)e^{\zeta}(\cos \zeta + \sin \zeta); \\ w_{23}(x, y, 0) &= -\frac{1}{2}\hat{k} \cdot \text{curl } \vec{\tau} = -\frac{1}{2}\frac{\partial \tau}{\partial x}. \end{aligned} \right\} \quad (7)$$

THE INTERIOR AND BOTTOM FLOW

If we assume $\beta = bE^n$, where b and n are constants greater than or equal to zero, the vorticity equation for the uniform depth model is found to be, from Eq. 4,

$$-2(1 + bE^n y) \frac{\partial w}{\partial z} + 2bE^n v = E \nabla^2 \omega; \quad \omega = v_x - u_y. \quad (8)$$

To accept the vertical mass flux from the surface layer, $E^{1/2} w_{23}(x, y, 0)$, an interior vertical velocity $O(E^{1/2})$ is required. For a large lake or sea, n is greater than $1/2$, and hence, for the interior flow, $\partial w / \partial z$ is zero. A horizontally nondivergent bottom Ekman layer with $O(E^{1/2})$ vertical velocities is required. This then requires $O(1)$ interior geostrophic velocities.

For the assumed solution (Eq. 6), it is found that at least up to $O(E^{1/2})$ the interior variables satisfy equations of the form

$$\left. \begin{aligned} -2v_{11} &= -p_{11x}, \\ +2u_{11} &= -p_{11y}, \\ 0 &= -p_{11z}, \end{aligned} \right\} \quad (9a)$$

and

and

$$u_{11x} + v_{11y} = 0; \quad w_{13z} = 0. \quad (9b)$$

The interior flow is hydrostatic, geostrophic, and nondivergent. The vertical velocity is determined from matching with the surface layer:

$$w_{13}(x, y) = -w_{23}(x, y, 0) = \frac{1}{2} \hat{k} \cdot \text{curl } \vec{\tau}.$$

The bottom boundary layer is a conventional Ekman layer. From Eq. 6, it is found that

$$\left. \begin{aligned} u_{31}(x, y, \zeta') &= -(u_{11} \cos \zeta' + v_{11} \sin \zeta') e^{-\zeta'}, \\ v_{31}(x, y, \zeta') &= -(v_{11} \cos \zeta' - u_{11} \sin \zeta') e^{-\zeta'}, \\ w_{33}(x, y, 0) &= -\frac{1}{2} \hat{k} \cdot \text{curl } \vec{v}_{11}. \end{aligned} \right\} \quad (10)$$

and

If the interior velocity is represented by a stream function,

$$u_{11} = -\frac{\partial \psi_{11}}{\partial y}, \quad v_{11} = \frac{\partial \psi_{11}}{\partial x},$$

then using the bottom boundary condition to match the interior vertical velocity, we have

$$\nabla^2 \psi_{11} = \hat{k} \cdot \text{curl } \vec{\tau} = \frac{\partial \tau}{\partial x}. \quad (11)$$

This is a statement that the generation of vorticity by the wind stress is balanced by the dissipation at the bottom. Since $\psi_{11} = 0$ on the coast, to bring the normal velocity to zero, the interior flow is determined to this order.

THE COASTAL BOUNDARY LAYERS

The interior horizontal mass transport is $(U_1, V_1) = (u_{11}, v_{11}) + O(E^{1/4})$. To the lowest order, the transport normal to the lateral boundaries vanishes, as in the usual small β transport theory. To discuss the mechanics of the $O(E^{1/2})$ Ekman transport, the surface drift transport and the vertical Ekman suction transport, however, we must extend the solution up to and including $O(E^{1/2})$. This requires analysis of coastal viscous layers, which for this model means the wall boundary layers.

The surface drift transport from Eq. 7 is

$$\left. \begin{aligned} U_2(x, y) &= E^{1/2} \int_{-\infty}^0 u_{21} d\zeta = \frac{1}{2} E^{1/2} \tau(x, y) + \dots; \\ V_2(x, y) &= E^{1/2} \int_{-\infty}^0 v_{21} d\zeta = 0 + \dots \end{aligned} \right\} \quad (12)$$

The mass flux associated with viscous flow at the bottom, from Eq. 10, is:

$$\left. \begin{aligned} U_3(x, y) &= -\frac{1}{2} E^{1/2} (u_{11} + v_{11}) + \dots; \\ V_3(x, y) &= \frac{1}{2} E^{1/2} (v_{11} - u_{11}) + \dots \end{aligned} \right\} \quad (13)$$

The total horizontal mass transport is then

$$U = \sum_{i=1}^3 U_i, \quad V = \sum_{i=1}^3 V_i.$$

Since $\beta \ll 1$, there will be no $E^{1/3}$ boundary layer on the western coast. In fact, with the absence of β , the structure will be essentially the same on all four coasts. There exists an outer layer of thickness $E^{1/4}L$, which is hydrostatic and geostrophic. It brings the interior longshore velocity to zero. It accepts the bottom Ekman flux $O(E^{1/2})$ and transports it up the wall. A thinner viscous region of thickness $E^{1/2}L$, identical to that occurring in P, is needed to connect the interior flow with the surface Ekman layer.

As in P, a much thinner layer of thickness δL is required to bring the vertical velocity to zero at the wall; this layer plays no role in the mass balance to lowest order.

The only difference between the viscous layers occurring at $\bar{x} = 0, 1$ and those at $y = 0, M$ will arise from there being no normal surface-drift transport at the latter two coasts, since the surface stress is directed in the y direction. It is therefore sufficient to discuss in detail the viscous layers occurring only at $x = 0$. By a simple transformation, solutions at the remaining coasts may be obtained.

In the vicinity of $x = 0$, let $x = E^{1/4}\eta$; the dependent variables are

$$\left. \begin{aligned} u &= u_1 + E^{1/4}\hat{u}(\eta, y) + \dots, \\ v &= v_1 + \hat{v}(\eta, y) + \dots, \\ w &= w_1 + E^{1/4}\hat{w}(\eta, y, z) + \dots, \\ p &= p_1 + E^{1/4}\hat{p}(\eta, y) + \dots \end{aligned} \right\} \quad (14)$$

and

The correction variables denoted by a $\hat{}$ satisfy the equations

$$\left. \begin{aligned} 0 &= -\hat{p}_\eta + 2\hat{v}, \\ 0 &= -\hat{p}_y - 2\hat{u} + E^{1/4}\hat{v}_{\eta\eta}, \\ 0 &= -\hat{p}_z, \\ 0 &= \hat{u}_\eta + \hat{v}_y + E^{1/4}\hat{w}_z. \end{aligned} \right\} \quad (15)$$

and

The correction flow has a horizontal divergence $O(E^{1/4})$. The second momentum equation may be written

$$\hat{u} \equiv \hat{u}_a + E^{1/4}\hat{u}_b \equiv -\frac{1}{2}\hat{p}_y + \frac{1}{2}E^{1/4}\hat{v}_{\eta\eta}. \quad (16)$$

That is, the normal correction velocity in the $E^{1/4}$ region has two parts: a geostrophic part $O(E^{1/4})$ and a divergent part $O(E^{1/2})$. Only the latter plays a role in the mass balance.

The vorticity equation is

$$0 = -2 \frac{\partial \hat{w}}{\partial z} - \hat{v} \eta \eta \eta.$$

Since $\hat{v}$ is independent of z , $\hat{w}$ is a linear function of z . For the bottom boundary condition to be satisfied, an Ekman layer is required in the independent variables $x = E^{1/4} \eta$, $z = E^{1/2} \zeta$; again this is a standard divergent Ekman layer near $z = 0$. If $\hat{w}(\eta, y, z) = (1 - z) \hat{w}(\eta, y, 0)$, the Ekman suction at $z = 0$ requires

$$\hat{w}(\eta, y, z) = \frac{1}{2} \hat{v} \eta (1 - z).$$

The vorticity equation reduces to

$$\hat{v} \eta \eta \eta - \hat{v} \eta = 0;$$

at $x = 0$ we must have $\hat{v}_{11} + \hat{v} = 0$. Hence,

$$\hat{v}(\eta, y) = -v_{11}(0, y) e^{-\eta}, \quad (17)$$

$$\hat{w}(\eta, y, z) = \frac{1}{2} (1 - z) v_{11}(0, y) e^{-\eta}, \quad (18)$$

and

$$\hat{u}(\eta, y) = \frac{-\partial v_{11}}{\partial y} (0, y) e^{-\eta} - \frac{1}{2} E^{1/4} v_{11}(0, y) e^{-\eta}. \quad (19)$$

The vertical mass flux is

$$\hat{W} = E^{1/2} \int_0^\infty \hat{w} d\eta = \frac{1}{2} E^{1/2} (1 - z) v_{11}(0, y). \quad (20)$$

This layer, having nonvanishing normal velocities at the coast, in turn forces higher-order interior velocities. To fully discuss the mass balance, we must consider these terms. The $O(E^{1/4})$ interior flow satisfies equations of the form of Eq. 9. Since there is no additional interior vertical velocity required, the matching bottom Ekman layer is nondivergent, and hence the second-order interior flow satisfies

$$\nabla_2^2 \psi_{12} = 0, \quad (21)$$

where ψ_{12} is the stream function for the $O(E^{1/4})$ flow. Boundary conditions force the flow, e.g.,

$$x = 0: u_{12}(0, y) = \frac{-\partial \psi_{12}}{\partial y} = -\tilde{u}(0, y) = \frac{\partial v_{11}}{\partial y}(0, y). \quad (22)$$

The third-order interior flow $O(E^{1/2})$, is forced not only by the $E^{1/4}$ layer, but the thinner $E^{1/2}$ layer discussed next.

Since the $E^{1/2}$ layer is so similar to that in P (see P, Sec. 5 and appendix), only an abbreviated derivation will be given. If we let $x = E^{1/2}\xi$, the expansion for the flow variables (Eq. 14) will have the following added terms:

$$\left. \begin{aligned} u &= \dots + E^{1/2}\tilde{u}(\xi, y) + \dots, \\ v &= \dots + E^{1/2}\tilde{v}(\xi, y) + \dots, \\ w &= \dots + \tilde{w}(\xi, y, z) + \dots, \\ \text{and} \\ p &= \dots + O(E^{5/4}) + \dots \end{aligned} \right\} \quad (23)$$

The tilde variables satisfy

$$\left. \begin{aligned} 0 &= 2\tilde{v} + \tilde{u}\xi\xi, \\ 0 &= -2\tilde{u} + \tilde{v}\xi\xi, \\ \text{and} \\ 0 &= \tilde{u}\xi + \tilde{w}_z. \end{aligned} \right\} \quad (24)$$

The flow is hydrostatic, but not geostrophic. If $\tilde{u} = \partial\tilde{\psi}/\partial z$, $\tilde{w} = -\partial\tilde{\psi}/\partial\xi$, these reduce to

$$\frac{\partial^2}{\partial z^2} (\tilde{\psi}\xi\xi\xi\xi + 4\tilde{\psi}) = 0.$$

The solution is found to be

$$E^{1/2}\tilde{\psi}(\xi, y, z) = ze^{-\xi}[U_2(0, y) \cos \xi + V_2(0, y) \sin \xi] = zU_2(0, y)e^{-\xi} \cos \xi,$$

$$E^{1/2}\tilde{u}(\xi, y) = U_2(0, y)e^{-\xi} \cos \xi,$$

$$E^{1/2}\tilde{v}(\xi, y) = U_2(0, y)e^{-\xi} \sin \xi,$$

and

$$E^{1/2}\tilde{w}(\xi, y, z) = zU_2(0, y)e^{-\xi}(\cos \xi + \sin \xi),$$

where use has been made of the fact that $V_2 \equiv 0$. The vertical mass flux is

$$\tilde{W}(y, z) = E^{1/2} \int_0^\infty \tilde{w}(\xi, y, z) d\xi = E^{1/2}\tilde{\Psi}(0, y, z) = zU_2(0, y). \quad (25)$$

The forcing for the third-order interior flow $O(E^{1/2})$ now becomes apparent. Since the $O(E^{1/2})$ interior flow is also geostrophic and nondivergent, we have again

$$\nabla_2^2 \psi_{13} = 0, \quad (26)$$

where ψ_{13} is the stream function for (u_{13}, v_{13}) . The boundary forcing comes from imposing that the total normal $O(E^{1/2})$ velocity vanish at the coast; that is,

$$\left. \begin{aligned} x = 0: \quad \hat{u}_b(0, y) + \tilde{u}(0, y) + u_{13}(0, y) &= 0, \\ \text{or} \quad -\frac{1}{2}v_{11}(0, y) + E^{-1/2}U_2(0, y) - \frac{\partial \psi_{13}}{\partial y} &= 0. \end{aligned} \right\} \quad (27)$$

DISCUSSION

The case of vanishing curl $\vec{\tau}$ is simplest to discuss. If $\tau = \tau(y)$, $\text{curl } \vec{\tau} \equiv 0$. The Ekman suction vanishes from the surface layer; with no vertical flux of mass in the interior, no $O(1)$ bottom Ekman layer is required, hence no interior velocity $O(1)$. Further, with no $O(1)$ interior velocities, no coastal layers of thickness $O(E^{1/4})$ are required. There remain: the uniform (in x) horizontal Ekman transport,

$$U_2(y) = \frac{1}{2}E^{1/2}\tau(y); \quad (28)$$

the $O(E^{1/2})$ geostrophic interior flow,

$$E^{1/2}u_{13}(y) = -U_2(y)$$

and

$$E^{1/2}v_{13} = 0,$$

and the viscous $E^{1/2}$ upwelling coastal layer at $x = 0, 1$. The surface current is essentially that of the Ekman layer, that is, 45° to the right of the wind stress. Lines of constant free surface height are perpendicular to the wind stress and with maximum setup downwind. The surface drift transport is returned via the interior at each value of y , with downwelling at $x = 1$, upwelling at $x = 0$ ($\tau > 0$).

In discussing the case of nonvanishing $\hat{k} \cdot \text{curl } \vec{\tau}$, let

$$\tau(x) = \bar{\tau} + \Delta(x - \frac{1}{2}), \quad (29)$$

where $\bar{\tau} > 0$, $\Delta < 0$ are constants. Then the interior $O(1)$ flow is found from Eq. 11:

$$\nabla^2 \psi_{11} = \Delta, \quad (30)$$

where

$$\psi_{11} = 0 \text{ on all coasts;}$$

the interior flow is a single symmetric clockwise gyre. There is a uniform downward exit of mass from the surface Ekman layer to the bottom Ekman layer. The total flux is

$$W_I = E^{1/2} \int_S w_{11} dS = \frac{1}{2} E^{1/2} \int_S \hat{k} \cdot \text{curl } \vec{\tau} dS = \frac{1}{2} E^{1/2} M \Delta < 0, \quad (31)$$

where S is the area of the basin. This mass flux begins its return from the bottom to the surface Ekman layer at the coasts in the $E^{1/4}$ viscous layer. From Eq. 20, the total vertical flux in the $E^{1/4}$ layer is found by integrating around the four sides of the basin. It is found that

$$\begin{aligned} \hat{W}_T &= -\frac{1}{2} E^{1/2} (1-z) \oint_C \vec{v}_{11} \cdot d\vec{s} = -\frac{1}{2} E^{1/2} (1-z) \Gamma \\ \text{and} \quad \Gamma &\equiv \oint_C \vec{v}_{11} \cdot d\vec{s} \end{aligned} \quad (32)$$

where C is a horizontal contour encircling the basin, and Γ is the geostrophic circulation $O(1)$ at the coast. By application of Stokes' theorem to Eq. 32 and use of Eq. 11, Eq. 29 yields

$$\begin{aligned}
 \Gamma &= \int_S \hat{k} \cdot \text{curl } \vec{\tau} \, dS = M\Delta \\
 \text{and} \\
 \hat{W}_T &= -\frac{1}{2}E^{1/2}(1-z) \int_S \hat{k} \cdot \text{curl } \vec{\tau} \, dS = -\frac{1}{2}E^{1/2}M\Delta(1-z)
 \end{aligned}
 \quad \left. \vphantom{\begin{aligned} \Gamma &= \int_S \hat{k} \cdot \text{curl } \vec{\tau} \, dS = M\Delta \\ \hat{W}_T &= -\frac{1}{2}E^{1/2}(1-z) \int_S \hat{k} \cdot \text{curl } \vec{\tau} \, dS = -\frac{1}{2}E^{1/2}M\Delta(1-z) \end{aligned}} \right\} \quad (33)$$

Note that this upwelling occurring on all coasts depends only on $\text{curl } \vec{\tau}$ and not directly on the stress itself. The vertical flux decreases from a maximum at $z = 0$, to zero at $z = 1$.

The total vertical mass flux in the $E^{1/2}$ layer is found in a similar way by integrating around the basin; using Eq. 25, we have

$$\tilde{W}_T = -z \oint_C \vec{V}_2 \cdot \hat{n} \, ds = \frac{1}{2}E^{1/2}z \oint_C \hat{k} \times \vec{\tau} \cdot d\vec{s} = -\frac{1}{2}E^{1/2}zM\Delta > 0, \quad (34)$$

where $\vec{V}_2 = U_2\hat{i} + V_2\hat{j}$, and $\hat{n}$ is the outward normal to the coast. Since $\vec{V}_2 \cdot \hat{n} = 0$ on $y = 0, M$, the $E^{1/2}$ vertical flux occurs only on the coasts parallel to the wind stress: upwelling at $x = 0$, downwelling at $x = 1$; the flux depends directly on the wind stress. Since $U_2(0)$, the normal surface drift transport at $x = 0$, is larger than $U_2(1)$, $\tilde{W}_T$ is positive. For uniform stress, $\tilde{W}_T = 0$; the upwelling exactly balances the downwelling, and the return flow was via the $O(E^{1/2})$ interior flow. With nonvanishing $\text{curl } \vec{\tau}$, the $O(E^{1/2})$ geostrophic interior flow performs the same task; but an additional source of mass influx to the interior flow is needed to supply the extra mass at $x = 0$. This flux comes from the convergence of the vertical mass flux $\hat{W}_T$ in the $E^{1/4}$ layer; the horizontal outflow into the $E^{1/2}$ interior field is brought about by the divergent part of the $E^{1/4}$ layer normal velocity. Extending Eq. 12 to all sides of the basin and integrating, we find that the flux entering the interior horizontal flow from the $E^{1/4}$ layer is

$$-E^{1/2} \oint_C \vec{v}_b \cdot \hat{n} \, ds = -\frac{1}{2}E^{1/4} \oint_C \vec{v}_{11} \cdot d\vec{s} = -\frac{1}{2}E^{1/2}\Gamma, \quad (35)$$

which is exactly the flux entering the $E^{1/4}$ layer at the bottom of the coastal walls. The horizontal $O(E^{1/2})$ interior flow streamlines may be constructed schematically. Streamlines emanate from the three coasts at $x = 1$, $y = 0, M$ and converge on the coast at $x = 0$, returning horizontally the downwelling flux at $x = 1$, and the upwelling flux from the bottom Ekman layer, which in turn originates from the Ekman suction flux from the surface layer. With the $O(E^{1/2})$ horizontal flow constructed, the mass flux balance is complete.

SUMMARY

For the rectangular uniform-depth basin, with no gradient of earth's vorticity present ($\beta = 0$), and for a simple wind stress $\vec{\tau} = \tau(x)\hat{j}$, there is downwelling $\frac{1}{2}E^{1/2}\tau(1)$ on the coast to the right of the wind stress and upwelling $\frac{1}{2}E^{1/2}\tau(0)$ on the left coast via an $E^{1/2}$ wall layer. The downwelling mass is returned horizontally across the basin to the upwelling coast by an $O(E^{1/2})$ geostrophic interior flow. The vertical flux of mass out of the divergent surface Ekman layer (Ekman suction) is transported to the bottom friction layer and thence horizontally to the coasts. Although this mass begins its return to the surface by ascending the coastal walls (via the $E^{1/4}$ vertical layer), it does not continue directly to the free surface, instead being forced again out into the interior. As part of the $O(E^{1/2})$ horizontal geostrophic return flow, it moves to the upwelling coast and thence to the surface friction layer. Large $[O(1)]$ geostrophic interior flow is required to produce the necessary stress for the bottom friction layer.

The simple wind stress was chosen solely for its effectiveness in permitting visualization of the circulation; a general steady stress can be treated with no more mathematical complexity. The artificiality of the rectangular geometry can also be easily removed to an arbitrary smooth closed coastline, thereby eliminating the special corner regions. The constraint of depth uniformity cannot be removed as easily. Preliminary analysis indicates, however, that much of the circulation mechanics is unchanged for the basin containing closed depth contours; the most significant alteration, arising from the Taylor-Proudman constraint, appears to be that the horizontal $O(E^{1/2})$ geostrophic return flow is pushed out of the interior and instead occurs in viscous coastal layers of thickness $O(E^{1/4})$.

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