8

Large-Scale Ocean Circulation

Summary: The concepts of geostrophy and Ekman layers are merged to study the large-scale barotropic circulation in the midlatitude oceans. The results lead to the Sverdrup balance for the large-scale circulation and the Stommel theory for the Gulf Stream.

8-1 SOME REMARKS ON THE OCEAN AND ATMOSPHERE

Ocean motions span a great variety of scales in both time and space. At one extreme, we find microturbulence, not unlike that in hydraulics, and at the other, the large-scale circulation, which spans ocean basins and which evolves over climatic time scale. The latter extreme is our present objective.

There are numerous reasons why the ocean water masses are set in motion: the gravitational pull exerted by the moon and sun, differences in atmospheric pressure at sea level, wind stress over the sea surface, and convection resulting from atmospheric cooling and evaporation. The moon and sun generate only the tides, and the atmospheric-pressure differences play an insignificant role. On the other hand, convection generates currents at high latitudes or other particular areas and is responsible for a very slow movement in the abyss. This leaves the stress exerted by the winds along the sea surface as the main driving force of basinwide circulations.

Ocean waters respond to the wind stress because of their low resistance to shear (low viscosity, even after viscosity magnification by turbulence) and because of the relative consistency with which winds blow over the ocean. A good example are the trade winds in the tropics; they are so steady that, shortly after Christopher Columbus and until the advent of steam, ships chartered their courses according to those winds; hence their name. Further away from the tropics are winds blowing in the opposite direction. Whereas the trades blow from the east and slightly toward the equator (they are also called northeasterlies and southeasterlies, depending on the hemisphere), the midlatitude winds blow from west to east and are aptly called westerlies (Figure 8-1).

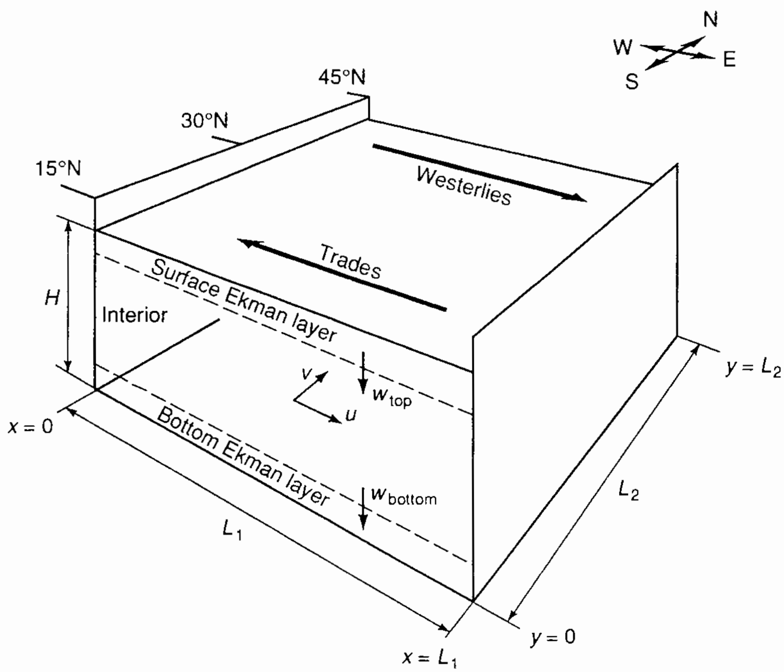

Figure 8-1 A simple model of midlatitude ocean circulation.

Generally much more variable than the trades, the westerlies have nonetheless a substantial average component, and the combination of the two wind systems drives significant circulations in all midlatitude basins: North and South Atlantic, North and South Pacific, and Indian Oceans.

8-2 A SIMPLE MODEL OF MIDLATITUDE CIRCULATION

Because our objective is not to use a mathematical model to reproduce all aspects of the large-scale circulation but only to elucidate the fundamental processes and their relations, we will once again seek the advantages of assumptions. We idealize the oceanic basin to a rectangular box aligned with the east–west and north–south axes, with a flat bottom, and filled with homogeneous water. We take the driving winds to be purely zonal and assume that the circulation is steady and governed by the beta-plane dynamics (Section 6-4). A geometrical sketch and some of the notation are displayed in Figure 8-1.

The lateral dimensions of the model basin are large ($L_1 \sim L_2 \sim$ several thousands of kilometers), whereas the velocities are expected to be modest ($u \sim v \sim$ few centimeters per second). Consequently, the Rossby number is extremely small, and it is not necessary to retain the nonlinear advective terms in the momentum equations.

Note that we have not required that the fluid be inviscid. Indeed, the wind stress can be communicated to the waters only if these waters can absorb a shear stress. But, because viscosity cannot be an overriding process in most of the circulation, we infer that viscous effects are relegated to thin boundary layers, namely, a surface Ekman layer where the fluid is subject to the wind stress and a bottom Ekman layer where the fluid velocity is brought to zero.

The theory of Chapter 5 demonstrated that the combination of viscosity and rotation yields a flow transverse to the direction of the applied stress and that this transverse flow is likely to be two-dimensionally convergent or divergent. The result is a nonzero vertical velocity through the flow outside of the boundary layer.

Therefore, it becomes crucial to distinguish three layers in our model. The topmost layer is a thin surface Ekman layer, at the top of which the vertical velocity is nil (flat surface), and at the bottom of which it is prescribed by the wind-stress curl, according to (5-20):

$$w_{\text{top}} = \frac{1}{\rho_0 f_0}\left(\frac{\partial \tau^y}{\partial x} - \frac{\partial \tau^x}{\partial y}\right), \tag{8-1}$$

where τ^x and τ^y are, respectively, the zonal and meridional wind-stress components. The lowest layer is another thin Ekman layer along the bottom; at its foot, the vertical velocity is zero (flat bottom), and, at its top, it is given by (5-16):

$$w_{\text{bottom}} = \frac{d}{2}\left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right), \tag{8-2}$$

where d is the Ekman depth defined in (5-8) and u and v are the zonal and meridional velocity components in the flow above this bottom Ekman layer.

The intermediate layer, away from the top and bottom Ekman layers, comprises most of the ocean. It is sometimes called the *geostrophic interior*. In the absence of advection and viscous effects, its governing equations are

$$-(f_0 + \beta_0 y)\ v = -\frac{1}{\rho_0}\frac{\partial p}{\partial x}, \qquad (8\text{-}3)$$

$$+(f_0 + \beta_0 y)\ u = -\frac{1}{\rho_0}\frac{\partial p}{\partial y}, \qquad (8\text{-}4)$$

$$0 = -\frac{1}{\rho_0}\frac{\partial p}{\partial z}, \qquad (8\text{-}5)$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0. \qquad (8\text{-}6)$$

To validate the beta-plane approximation ($f = f_0 + \beta_0 y$ instead of $f = 2\Omega \sin \varphi$ and the need to retain the spherical geometry), we require that $\beta = \beta_0 L_2 / f_0$ be small. This implies that, in first approximation, equations (8-3) and (8-4) yield geostrophy on the f-plane, and, as noted in Section 4-1, this leads to the mutual cancellation of $\partial u/\partial x + \partial v/\partial y$ in the continuity equation. Thus, the existence of a vertical velocity in the ocean interior is intimately related to the beta effect. Obviously, this interior vertical velocity will be required to match, at both top and bottom, those vertical velocities emanating from the Ekman layers.

To recapitulate, we have the following scenario: Winds blow over the ocean and set up an Ekman layer just below the sea surface; because the wind stress has a nonzero curl, cross-wind transports in this upper layer of the ocean are convergent or divergent, and a vertical velocity is generated that is communicated to the waters below; in those waters a vertical velocity can be accommodated only by the convergence or divergence of a geostrophic flow under the influence of the beta effect; thus, a horizontal geostrophic flow is required in the interior in reaction to the vertical velocity. Finally, this geostrophic flow reaches the bottom where friction sets up a bottom Ekman layer, which generates an additional vertical velocity, with which the interior flow must reckon.

Let us now proceed with the mathematical developments. To bring forth the relation between the vertical velocity and the beta effect in the interior, let us eliminate the f-plane geostrophic dynamics by combining the momentum equations (8-3) and (8-4) to cancel the pressure derivatives. The result is

$$(f_0 + \beta_0 y)\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) + \beta_0 v = 0.$$

Since the beta correction to the Coriolis parameter is required to be small ($\beta_0 y \ll f_0$), by virtue of the continuity equation, (8-6), it follows that

$$\beta_0 v = f_0 \frac{\partial w}{\partial z}. \qquad (8\text{-}7)$$

This is an extremely important result, which deserves a comment. As anticipated, if there is no beta effect ($\beta_0 = 0$), there is no vertical velocity gradient, and the vertical velocity cannot vary gradually to meet the required values (8-1) and (8-2) at both top and bottom. Equation (8-7) can also be viewed as a result of conservation of volume, of circulation, and, hence, of potential vorticity (Figure 8-2). If $\partial w/\partial z$ is positive, a water parcel is stretched vertically, which implies that it must shrink laterally to conserve its volume; but, as it shrinks laterally, it must increase its vorticity to conserve circulation. At large scales, the only significant contribution to vorticity is the ambient

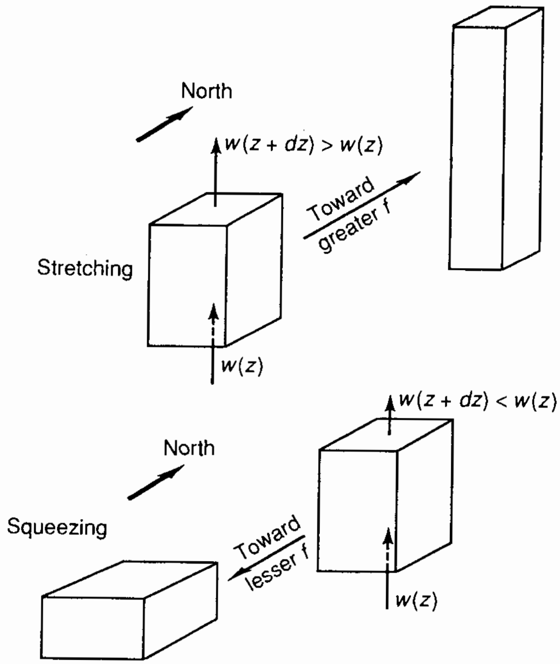

Figure 8-2 Meridional migration of fluid parcels induced by vertical stretching or squeezing in the large-scale oceanic circulation.

vorticity $f = f_0 + \beta_0 y$, and the parcel has no choice but to migrate meridionally in search of a better f; the term $\beta_0 v = df/dt$ represents this rate of variation. Note that nothing has been said regarding the zonal velocity, u, which need not vanish.

Vertical derivatives of (8-3) and (8-4) with the use of (8-5) indicate that the horizontal velocities in the interior do not change with depth. According to (8-7), this implies that $\partial w/\partial z$ is also constant with depth. The top and bottom values of the vertical velocity are given by the Ekman layer dynamics, (8-1) and (8-2), and the vertical gradient of vertical velocity is thus known:

$$\frac{\partial w}{\partial z} = \frac{w_{\text{top}} - w_{\text{bottom}}}{H}$$

$$= \frac{1}{\rho_0 f_0 H}\left(\frac{\partial \tau^y}{\partial x} - \frac{\partial \tau^x}{\partial y}\right) - \frac{d}{2H}\left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right), \tag{8-8}$$

where H is the thickness of the geostrophic interior or, with a slight approximation, the ocean depth.

When this result is used in (8-7) and the horizontal velocity components are approximated by their f-plane geostrophic values $[u = -(1/\rho_0 f_0)\ \partial p/\partial y,\ v = +(1/\rho_0 f_0)\ \partial p/\partial x]$, we obtain a single equation for the pressure:

$$\frac{f_0 d}{2\beta_0 H}\nabla^2 p + \frac{\partial p}{\partial x} = \frac{f_0}{\beta_0 H}\left(\frac{\partial \tau^y}{\partial x} - \frac{\partial \tau^x}{\partial y}\right). \tag{8-9}$$

This equation governs the flow pattern in the ocean interior, that is, away from the surface and bottom Ekman layers. Notice how subtle the dynamic balance is and thus how indirectly the atmospheric winds drive the oceanic circulation.

8-3 SVERDRUP TRANSPORT

The ratio of the first two terms of the preceding equation is on the order of $f_0 d/\beta_0 H L_1$. For a turbulence-enhanced viscosity of $\nu = 10^{-2}\ \mathrm{m^2/s}$ and a mean Coriolis parameter $f_0 = 8 \times 10^{-5}\ \mathrm{s^{-1}}$, the Ekman depth $d = (2\nu/f_0)^{1/2}$ is about 15 m. With $\beta_0 = 2 \times 10^{-11}\ \mathrm{m^{-1}\cdot s^{-1}}$, $H = 3000$ m, and $L_1 > 1000$ km, the preceding ratio is at most 0.02 and can thus be considered small. Physically, this implies that the Ekman pumping induced by the surface winds is much more vigorous than that caused by bottom friction. Mathematically, it implies that, in a first approximation, the first term of (8-9) can be neglected. This leaves

$$\frac{\partial p}{\partial x} = \frac{f_0}{\beta_0 H}\left(\frac{\partial \tau^y}{\partial x} - \frac{\partial \tau^x}{\partial y}\right). \tag{8-10}$$

Rewritten in terms of the meridional velocity, this equation takes the form

$$v = \frac{1}{\rho_0 \beta_0 H}\left(\frac{\partial \tau^y}{\partial x} - \frac{\partial \tau^x}{\partial y}\right), \tag{8-11}$$

which states that the meridional velocity is given at all locations as a function of the local wind-stress curl. This remarkable result, first obtained in 1947 by H. U. Sverdrup, is called the *Sverdrup transport*.

In the Northern Hemisphere midlatitude oceans (Figure 8-1), the main wind pattern, consisting of the trades to the south and the westerlies to the north, provides $\tau^y \approx 0$ and $\partial \tau^x/\partial y > 0$. Consequently, the Sverdrup transport is directed southward. This southward flow exists in the band between the maximum trade winds and the maximum westerlies (i.e., the latitudes where $\partial \tau^x/\partial y$ vanishes). The latitudes of these two maxima can be taken as the meridional limits of our model ($y = 0$ and $y = L_2$).

In terms of the pressure field, the solution to equation (8-10) is

$$p = P_1(y) - \frac{f_0}{\beta_0 H}\frac{d\tau^x}{dy}\,x, \tag{8-12}$$

where the winds are assumed to be zonally uniform for the sake of an easy integration. The "constant" of integration, $P_1(y)$, remains to be determined by imposing boundary conditions in the x-direction. In terms of the velocity components, the solution is

$$u = \frac{-1}{\rho_0 f_0} \frac{\partial p}{\partial y} = \frac{-1}{\rho_0 f_0} \frac{dP_1}{dy} + \frac{1}{\rho_0 \beta_0 H} \frac{d^2 \tau^x}{dy^2} x \tag{8-13}$$

$$v = \frac{1}{\rho_0 f_0} \frac{\partial p}{\partial x} = \frac{-1}{\rho_0 \beta_0 H} \frac{d\tau^x}{dy}, \tag{8-14}$$

as the first f-plane approximation. Ideally, we would like the zonal velocity to satisfy a no-flow condition on the meridional boundaries of the ocean domain (i.e., $u = 0$ at both $x = 0$ and $x = L_1$), but this is impossible because there is only one function $P_1(y)$ to be determined. Physically, the problem arises because the meridional velocity is directed southward everywhere except at the two limits $y = 0$ and L_2, where it vanishes; no northward return flow is allowed, and, consequently, the domain must remain open in the zonal direction on at least one side so that a volume flux can be provided.

Mathematically, the problem finds its root in equation (8-10), which is of first order in the x-direction and thus admits only one boundary condition. With the bottom friction retained, the full equation (8-9) is of second order and should permit specification of two boundary conditions. From this, we conclude that, somewhere, friction must be important. Because the coefficient of the highest x-derivative is small, the term can be significant only if that derivative is large, and we thus expect that friction will be important in a relatively narrow boundary layer. Whether this boundary layer must lie on the eastern or western side of the domain is the object of the next section.

8-4 WESTWARD INTENSIFICATION

This section title gives us the answer to where the boundary layer must lie, but let us verify it. Assume first that the boundary layer lies along the eastern wall ($x = L_1$). As shown in Figure 8-3a, the northward flow required to return the Sverdrup transport must be accompanied by a positive $\partial v/\partial x$. This derivative is large because the boundary layer is narrow and also because the velocity must be large to accommodate the entire Sverdrup transport. By comparison, $\partial u/\partial y$ is very small. Thus, according to (8-2), the vertical velocity produced by the bottom Ekman layer is upward. This creates a problem: The winds are such that they induce an influx of fluid from the surface Ekman layer into the interior, and if the bottom Ekman layer does likewise, where is the fluid ultimately going? So, we reject this possibility.

This implies that the boundary layer must lie along the western boundary (Figure 8-3b). Indeed, the term $\partial v/\partial x$ is then large and negative, the vertical velocity of the bottom Ekman layer is directed downward, and the influx from the top can be accommodated at the bottom.

To obtain the solution in the boundary layer, we note that the x-derivatives are much larger there than they are in the rest of the domain, and equation (8-9) now takes the approximate form

$$\frac{f_0 d}{2\beta_0 H} \frac{\partial^2 p}{\partial x^2} + \frac{\partial p}{\partial x} = 0, \tag{8-15}$$

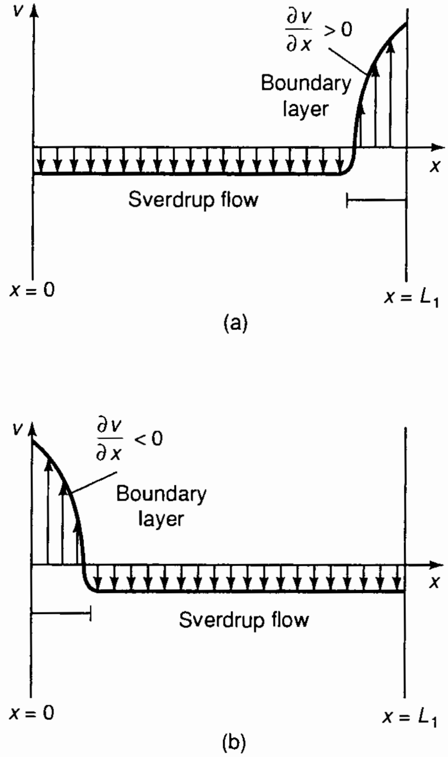

Figure 8-3 Possible configurations for a northward boundary current to return the southward Sverdrup flow that exists across most of the ocean basin: (a) boundary current on the eastern side, (b) boundary current on the western side. The former is to be rejected on dynamic grounds, leaving the latter as the correct configuration.

of which the solution is

$$p = P_2(y) + P_3(y)\ \exp\left(-\frac{2\beta_0 H}{f_0 d}\ x\right). \tag{8-16}$$

The negative sign in the exponential indicates eastward decay (or westward growth) and confirms that the boundary layer must lie on the western side of the domain. It now remains to match the preceding boundary solution to that in the interior. Taking the limit of (8-16) toward large x (i.e., leaving the boundary layer) and the limit of (8-12) toward small x (i.e., approaching the western boundary) yields $P_1(y) = P_2(y)$. Combining both solutions, we then obtain the solution that covers the entire zonal extent of the domain:

$$p = P_1(y) - \frac{f_0}{\beta_0 H}\ \frac{d\tau^x}{dy}\ x + P_3(y)\ \exp\left(-\frac{2\beta_0 H}{f_0 d}\ x\right). \tag{8-17}$$

In the western boundary layer ($x \lesssim f_0 d/\beta_0 H$, i.e., much less than L_1), the middle term is negligible and the solution reduces to (8-16), whereas in the rest of the domain ($x \sim L_1$, i.e., much greater than $f_0 d/\beta_0 H$), the last term is negligible and the solution reduces to (8-12).

With two arbitrary functions, $P_1(y)$ and $P_3(y)$, we are now in a position to impose the no-flow conditions on both eastern and western walls ($u = 0$ and thus $\partial p/\partial y = 0$ at $x = 0$ and L_1). On the eastern side ($x = L_1$), solution (8-12) is used to find

$$P_1(y) = \frac{f_0 L_1}{\beta_0 H} \frac{d\tau^x}{dy} + P_{10},$$

whereas on the western side ($x = 0$), solution (8-16) is used to find

$$P_3(y) = -P_1(y) + P_{30}.$$

The constant P_{10} does not affect the flow field in any way and can be set to zero; the constant P_{30} adds an arbitrary meridional flow inside the boundary layer, which is in no way related to the flow in the rest of the domain. It, too, can be dropped. The final solution is

$$p = \frac{f_0 L_1}{\beta_0 H} \frac{d\tau^x}{dy} \left[1 - \frac{x}{L_1} - \exp\left(-\frac{2\beta_0 H}{f_0 d} x\right)\right], \tag{8-18}$$

$$u = \frac{-L_1}{\rho_0 \beta_0 H} \frac{d^2\tau^x}{dy^2} \left[1 - \frac{x}{L_1} - \exp\left(-\frac{2\beta_0 H}{f_0 d} x\right)\right], \tag{8-19}$$

$$v = \frac{1}{\rho_0 \beta_0 H} \frac{d\tau^x}{dy} \left[-1 + \frac{2\beta_0 H L_1}{f_0 d} \exp\left(-\frac{2\beta_0 H}{f_0 d} x\right)\right]. \tag{8-20}$$

From the graphical representation of this solution (Figure 8-4), it is evident that the circulation takes the form of an asymmetric gyre, with a slow southward flow (the Sverdrup transport) occupying most of the domain and a swift boundary-layer current

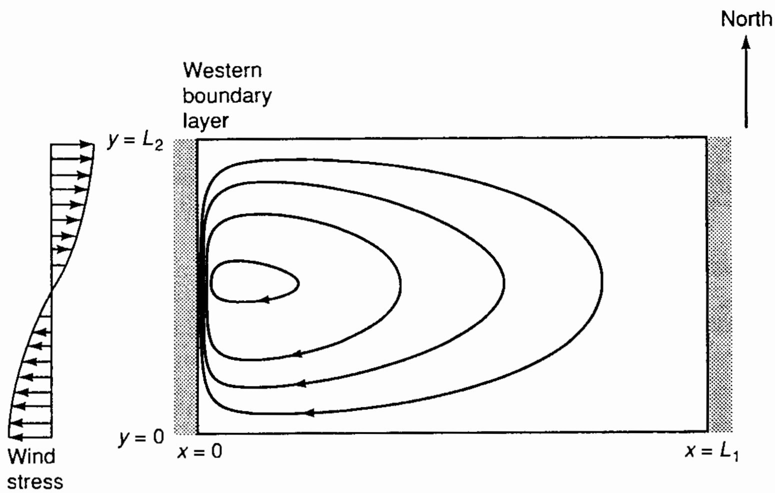

Figure 8-4 Wind-driven barotropic circulation at midlatitudes according to the Stommel model.

on the western side that returns the water masses northward. The latter current is to be identified with the Gulf Stream of the North Atlantic Ocean and the respective western-boundary currents in the other ocean basins (Kuroshio in the North Pacific, for example). The circulation closes with zonal currents, which flow eastward to the north and westward to the south.

The collection of the southbound Sverdrup transport into a westward current on the southern side and the resulting increase of the latter downstream has been termed *westward intensification* by H. M. Stommel, who proposed the preceding model as the first correct theory for the existence of the Gulf Stream (Stommel, 1948).

8-5 DISCUSSION

With the solution now at hand, let us recapitulate the results. As we are becoming aware, the mechanisms of ocean circulation are intricate and certainly less direct than a simple torque exerted by a surface stress on a viscous fluid. The chief reasons are that viscosity is weak, and the Coriolis effect (including its variation with latitude) is strong.

The scenario is as follows. The large-scale atmospheric winds, comprising essentially the trades and westerlies, generate a stress along the ocean surface. Because seawater is only slightly viscous and rotation is strong, the direct effect of the stress is limited to a thin (15 m or so) layer of the ocean. The earth's rotation also generates a component of the upper-layer flow transverse to the winds, which converges, resulting in a downward flow into the ocean interior below. Although relatively weak (10^{-6} m/s), this vertical flow squeezes water parcels vertically. In reaction, fluid parcels flatten and widen, and to conserve their circulation, their ambient vorticity must decrease. They thus migrate southward. As they progress southward, these waters run into a region of slower southward flow and hence must turn laterally. Bottom friction is the factor determining the direction of this zonal flow. Waters veer westward and gather into a zonal flow that intensifies downstream. Upon arriving at the western boundary, this flow turns into a swift northward flow, so swift that bottom friction has a controlling influence. This friction generates a bottom Ekman layer, in which the transverse flow is divergent, resulting into a downward pumping of waters. This downward pumping stretches water parcels vertically, thus enabling them to regain their original thicknesses and eventually to complete the circuit.

Obviously, all of our assumptions have eliminated a considerable number of additional processes that can all affect the ocean circulation in one way or another. Inertia (represented by the nonlinear advection terms) is important in the western boundary layer where the flow is swift and narrow (Rossby number becoming of order 1). The consequence is a detachment of this intense current from the coast and its penetration into the ocean interior, where it starts to meander rather freely. Barotropic instabilities (see Chapter 7) are likely. Stratification is another aspect that requires ample consideration. Briefly, the effect of stratification is to decouple the flow in the vertical and thus to make it respond less to bottom friction. On the other hand, a reserve of potential energy due to the presence of stratification causes baroclinic instabilities (Chapter 16).

Barotropic and baroclinic instabilities generate eddies, and these in turn create a net horizontal mixing of momentum. In other words, lateral friction may overcome vertical friction. Finally, because swift poleward currents, such as the western-boundary current (Gulf Stream) modeled here, bring warm water masses to higher latitudes, an air–sea heat flux is created, resulting in the cooling of the ocean and a distortion of the circulation pattern. The interested reader will find additional information on ocean-circulation dynamics in the review article by Veronis (1981), the book by Abarbanel and Young (1987), and the article by Cushman-Roisin (1987).

PROBLEMS

8-1. From the Stommel theory of the western boundary current in the large-scale oceanic circulation, derive analytical expressions for the scale of the width of the boundary layer and the scale of the current speed within it. Derive these scales using the following quantities: Coriolis-parameter mean value and gradient f_0 and β_0, mean ocean depth H, zonal width of basin L_1, zonal wind-stress amplitude τ_0, meridional distance between two latitudes where the wind-stress curl vanishes L_2, reference water density ρ_0, and Ekman-layer depth d.

8-2. Using the numerical values listed at the beginning of Section 8-3 together with $\tau^x = 0.15$ N/m^2 and $L_2 = 3300$ km, estimate the width and speed of the western boundary current. How do the values compare with those of the Gulf Stream? [For ample information on the Gulf Stream, consult the book by Stommel (1965).]

8-3. Given that the North Pacific Ocean is about twice as wide as the North Atlantic Ocean and that both basins are subjected to equally strong winds, compare their boundary-layer widths and boundary-current speeds.

8-4. What do you think is the fate of the waters absorbed by the bottom Ekman layer under the western boundary current?

8-5. Show that, as in the Northern Hemisphere, the boundary currents in ocean basins of the Southern Hemisphere are along the western boundaries. Which way are they directed?

8-6. Imagine that a single ocean were covering the entire globe, as the atmosphere does. With no western wall to support a boundary current returning the equatorward Sverdrup flow, what would be the circulation pattern? Relate your results to the existence of the Antarctic Circumpolar Current. [For a succinct description of this major current, sea Section 7.2 of the book by Pickard and Emery (1990) or some other oceanography textbook.]

SUGGESTED LABORATORY DEMONSTRATION

Equipment

Rotating circular tank with slanted bottom, transparent floating disk with overhanging wings, neutrally buoyant particles.

Experiment

Fit a circular tank with a slanted false bottom of uniform slope, fill it to near capacity with water and a collection of neutrally buoyant particles, and cap it with a transparent floating disk, which should fit loosely within the tank. On the upper side of that disk, attach a series of protrusions (such as cardboard wings). These protrusions should preferably overhang in order not to obstruct visualization of the flow from above and should be designed to maximize air drag. Bring the tank so fitted to rotation.

The air drag on the protrusions forces the floating disk to rotate slightly slower than the tank, thus exerting a retarding (anticyclonic) surface drag on the water. This surface drag simulates the wind stress over the ocean. The sloping bottom models the beta effect (deep is toward the equator, shallow is toward the pole).

Observe the broad and slow Sverdrup flow across isobaths in most of the tank (is it directed toward the deeper side?) and the narrow and fast boundary current (is it on the "western" side?). (For a detailed account of the motions in the sliced cylinder, see Pedlosky and Greenspan, 1967.)

Harald Ulrik Sverdrup

1888 – 1957

Born in Norway, Harald Ulrik Sverdrup first studied meteorology at the University of Bergen as an assistant to Vilhelm Bjerknes and, upon completion of his doctorate, took charge of the scientific work of Roald Amundsen's North Polar Expedition with the *Maud* (1917–1925). During those years, his interest shifted to oceanography and he began to propose improvements for oceanic instrumentation. In 1936 Sverdrup accepted, at first for a limited period, the directorship of the Scripps Institution of Oceanography at the University of California (USA). He stayed for twelve years. It was there that he developed the theory of the general equatorward flow in the midlatitude oceans that now bears his name. In his later years, Sverdrup returned to Norway and became increasingly involved in the activities of international scientific bodies. Throughout his life, he strived to accomplish every task he undertook. A unit for volumetric ocean transport bears his name: 1 sverdrup = 1 Sv = 10^6 m^3/s. (*Photo from the archives of the Scripps Institution of Oceanography.*)

Henry Melson Stommel

1920 – 1992

At an early age, Henry Melson Stommel considered a career in astronomy but turned to oceanography as a way to make a peaceful living during World War II. Having been denied admission to graduate school at the Scripps Institution of Oceanography by H. U. Sverdrup, then its director, Stommel never obtained a doctorate. This did not deter him; having soon realized that, in those years, oceanography was largely a descriptive science almost devoid of physical principles, he set out to develop dynamic hypotheses and to test them against observations. To him, we owe the first correct theory of the Gulf Stream (1948), theories of the abyssal circulation (early 1960s), and a great number of significant contributions on virtually all aspects of oceanography. Unassuming and avoiding the limelight, Stommel relied on a keen physical insight and plain common sense to develop simple models that clarify the roles and implications of physical processes. He was generally wary of numerical models. Particularly inspiring to young scientists, Stommel continuously radiated enthusiasm for his chosen field, which, as he was the first to acknowledge, is still in its infancy. (*Photo by George Knapp.*)