# WIND DRIVEN CIRCULATION

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## Introduction

Winds represent a dominant source of energy for driving oceanic motions. At the ocean surface, such motions include surface gravity waves, which are familiar as the waves that break on beaches. Winds are also responsible for small-scale turbulent fluctuations just beneath the ocean surface. Turbulent motions can be created by breaking waves or by the nonlinear evolution of currents near the air–sea interface. Subsurface processes such as these can lead to easily observed windrows or scum lines on the sea surface. Winds also generate other complex and varied small-scale motions in the top few tens of meters of the ocean. However, the surface/wind-driven circulation described here refers instead to considerably larger-scale motions that compare in size to the ocean basins and extend as much as a kilometer or more below the surface.

The textbook notion of the surface/wind-driven circulation includes most of the well-known surface currents, such as the intense poleward-flowing Gulf Stream in the western North Atlantic (**Figure 1**). Analogues of the Gulf Stream can be found in each of the major ocean basins, including the Kuroshio in the North Pacific, the Brazil Current in the South Atlantic, the East Australian Current in the South Pacific, and the Aghulas in the Indian. These 'western boundary currents' are not isolated structures. Rather, they represent the poleward return flow for basin-scale motions that occupy middle latitudes in all major oceans. Each of the major ocean basins has an analogous set of large-scale current systems. The western boundary currents are quite intense, reaching velocities in excess of 1 $\mathrm{m\,s^{-1}}$, while the interior flow speeds are considerably smaller in magnitude.

The basin-scale patterns in the mid-latitude surface circulation are referred to as subtropical gyres. The gyres extend many hundreds of meters below the surface, reaching the bottom in some locations. Subtropical gyres rotate anticyclonically, that is, they rotate in a sense that is opposite to the sense of the earth's rotation (clockwise in the northern hemisphere and counterclockwise in the south). In the North Atlantic and North Pacific, subpolar gyres reside to the north of the subtropical gyres. They too include intense western boundary currents. However, the subpolar gyres rotate cyclonically, in the opposite sense of the subtropical gyres and in the same sense as the earth. Rotation of the wind-driven gyres is related to the rotation of the earth through a simple, though nonintuitive, physical mechanism. This mechanism is fundamental to understanding how the wind drives large-scale flows.

Our present understanding of the dynamics associated with the surface/wind-driven circulation developed largely during a 30-year period starting in the late 1940s. Before that time, oceanographers were aware of the gyre-scale features of the surface circulation. But it was not until the major theoretical advances in geophysical fluid dynamics beginning around 1947 that the surface circulation was conceptually linked to the winds.

## Observations

Oceanic wind systems exhibit a large-scale pattern that is common to the major ocean basins (**Figure 2**). Near the equator, trade winds blow from east to west. Near the poles, westerly winds blow from west to east. The ocean gyres have similar distributions of east–west flow. But the reasons for this are quite subtle. Moreover, there are profound differences between oceanic and atmospheric motion. North–south flows in the ocean are much more strongly pronounced than they are in the atmosphere, and winds fail to exhibit analogues of the intense poleward western boundary currents found in the ocean.

Western boundary currents such as the Gulf Stream were evident in estimates of time-averaged surface circulation obtained over a century ago with 'ship-drift' data (**Figure 3**). Ship drift is the discrepancy between a ship's position as obtained with dead reckoning and that obtained by more accurate navigation. Thus, ship drift can be attributed at least in part to ocean currents. The patterns of the surface circulation that emerge after averaging large numbers of such measurements are qualitatively correct. In general, however, accurate measurements of currents are difficult to obtain, particularly below the surface.

Temperature and salinity measurements are relatively abundant, and they provide an alternative resource for estimating large-scale currents. Temperature and salinity determine seawater density. The density distribution provides information about the pressure field, which in turn can be used to diagnose currents. For many decades, oceanographers have been routinely measuring vertical profiles of temperature and salinity. Consequently, detailed maps of the three-dimensional density structure exist for all the ocean basins.

Such maps clearly show a region of anomalously large vertical gradients in temperature, salinity, and density known as the main thermocline. The main thermocline divides two regions of relatively less stratified water near the surface and bottom. Thermocline depth varies substantially on the gyre scale, and can exceed 700 m depth in some regions. Lateral variations in the density field can be quite large as well, and these are associated with the

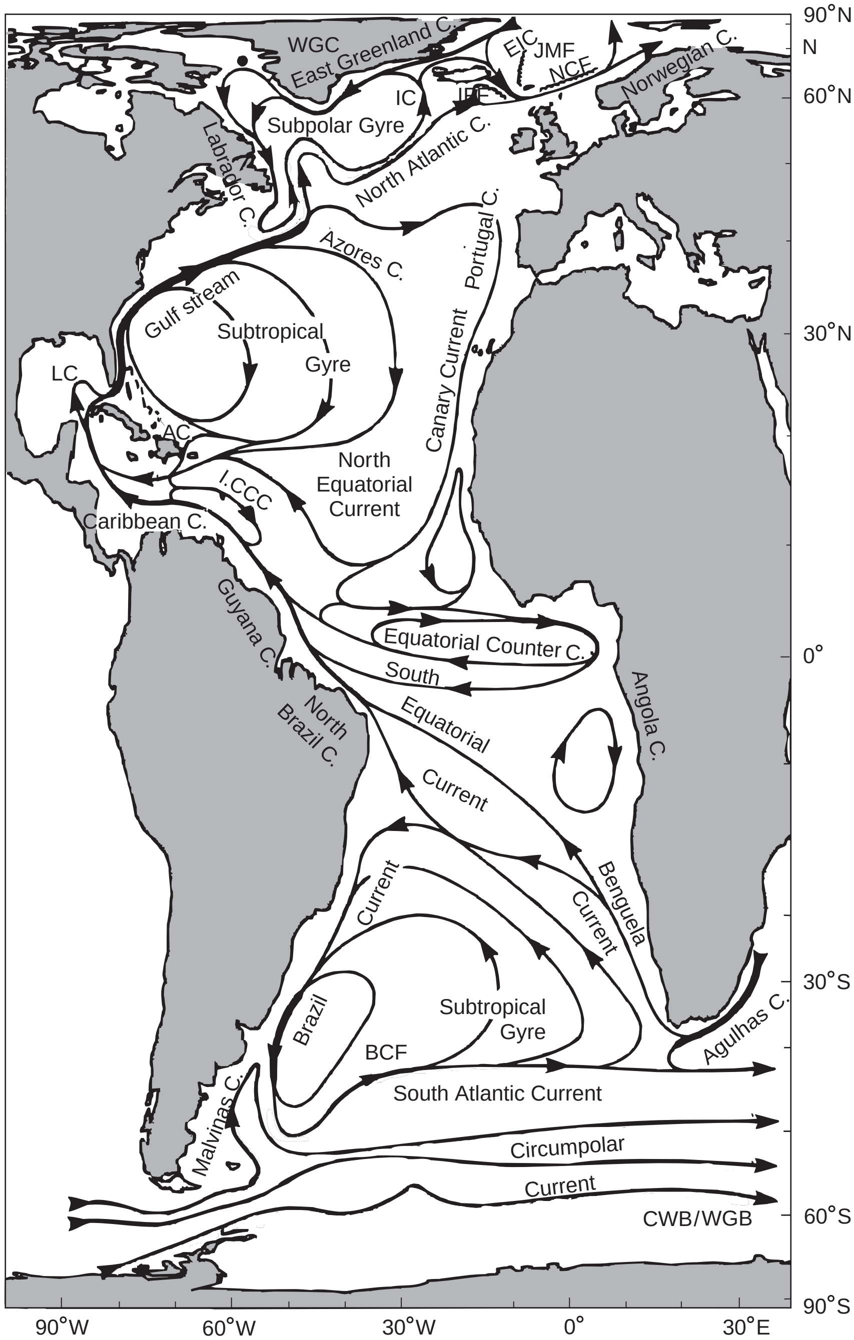


**Figure 1** Schematic of large-scale surface currents in the Atlantic Ocean. (From Tomczak and Godfrey (1994) *Regional Oceanography: An Introduction*.)

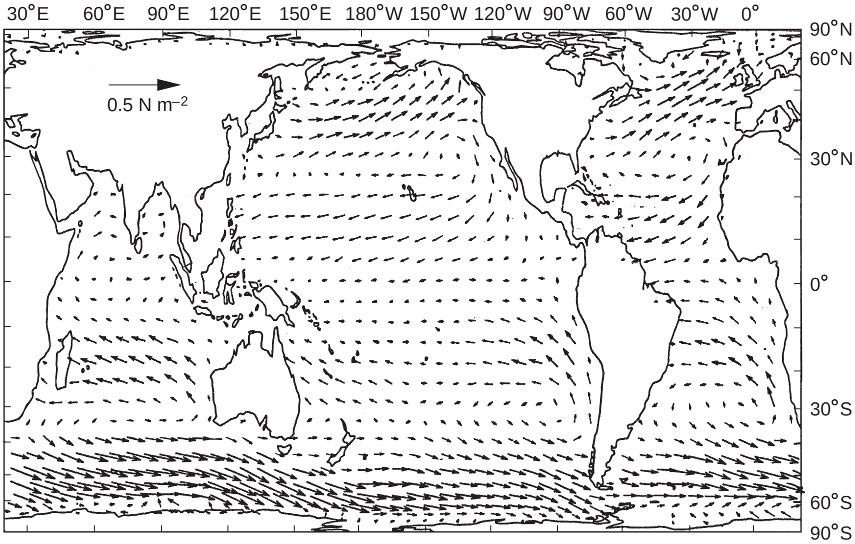


**Figure 2** Global mean surface wind stress, which is related to wind eqn [1]. (From Tomczak and Godfrey (1994) *Regional Oceanography: An Introduction*.)

currents that we refer to as the surface/wind-driven circulation.

The first step in estimating currents from density involves computation of dynamic height using the principle of isostasy. Isostasy describes, for example, the pressure field in a glass of ice water. An ice cube represents a region where water is slightly less dense than its surroundings. Buoyancy forces elevate the

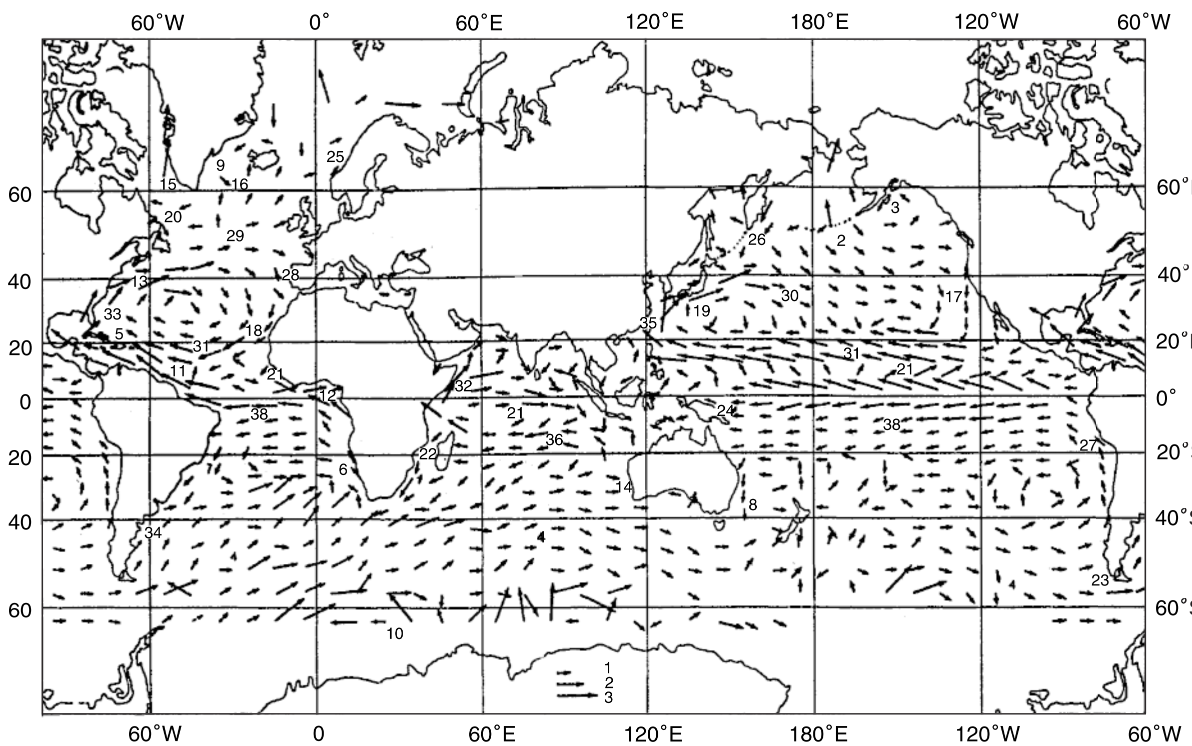


**Figure 3** Surface currents inferred from ship-drift measurements. To simplify the presentation, there are three vector sizes in this figure indicated by the scale vectors at the bottom of the figure. A vector the size of vector 1 corresponds to flow speeds in the range 0–10 cm $s^{-1}$, vector 2 is 10–20 cm $s^{-1}$, vector 3 is more than 30 cm $s^{-1}$. While some of the values have questionable reliability, the vectors show the general patterns large-scale circulation at the ocean surface. From Stidd CK (1974) Ship Drift Components: Means and Standard Deviations, SIO Reference Series 74-33 as appearing in Burkov VA (1980) *General Circulation of the World Ocean*. Gidrometeoizdat Publishers, Leningrad, published for the Division of Ocean Sciences, National Science Foundation, Washington, DC, by Amerind Publishing Co. Pvt. Ltd., New Delhi. 1993.

surface of the ice cube above the surface of the surrounding fluid. Similarly, a region in the ocean with less dense water than its surroundings will have a slightly elevated sea surface. In the ocean, this result involves the tacit assumption that currents in the abyssal ocean are weak relative to those nearer the surface, as is usually the case. The ocean's surface topography implied by the density field is referred to as dynamic height. **Figure 4** shows dynamic height computed from density between 2000 and 0 m depth. The variations of a meter or more are large enough to account for the pressure gradients that force the large-scale gyres.

The connection between pressure and large-scale currents involves the principle of 'geostrophy'. Geostrophic currents arise from a balance of the forces involving pressure gradient forces and Coriolis accelerations. This balance is a consequence of the large horizontal scales of the flow combined with the rotation of the earth. If the earth were not rotating, the sea surface elevations would accelerate horizontal flows down the pressure gradient, as occurs with smaller-scale motions such as surface gravity waves. With large-scale geostrophic flows, however, the Coriolis effect gives rise to currents that flow perpendicular to the pressure gradient, as indicated by the arrows in **Figure 4**.

Geostrophic currents such as those in **Figure 4** provide evidence of the surface/wind-driven circulation. This point requires some explanation, since the geostrophic flows are associated with large-scale pressure-gradient forces in the top kilometer of the ocean. As discussed below, winds directly drive motions in a relatively thin layer at the ocean surface known as the surface mixed layer. But these directly wind-driven flows give rise to other large-scale flows and, in turn, to the large-scale pressure gradients that can be estimated with dynamic height. Thus, it is accurate to refer to the large-scale geostrophic surface circulation as the wind-driven circulation because the pressure gradients would not exist without the wind.

## Wind-Driven Surface Layer

### Surface Mixed Layer

The surface mixed layer is loosely defined as a part of the water column near the surface where observed temperature and salinity fields are vertically uniform. In practice this layer extends from the ocean surface to a depth where stratification in temperature or density exceeds some threshold value. Typically, the underlying water is more strongly stratified. The mixed-layer depth often undergoes large diurnal and seasonal variations, varying between 0 and 100 m. However, the surface mixed layer rarely occupies more than 1% of the total water column.

Winds provide the primary source of mechanical forcing for the motions that homogenize water properties within the mixed layer. Mixing can also result from destabilizing effects of cooling and evaporation near the surface, as these can temporarily

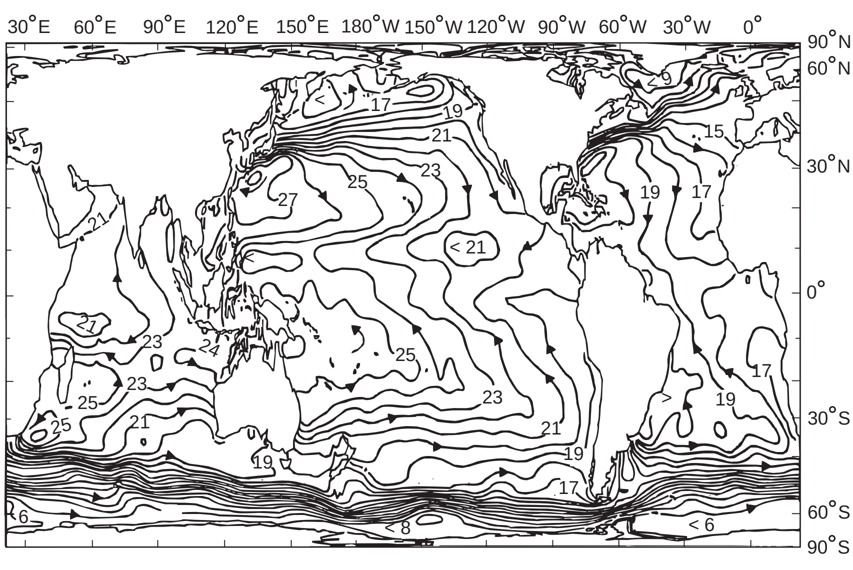


**Figure 4** Dynamic height ($m^2\ s^{-2}$) computed using the density field between 0 m and 2000 m depth, and assuming that the pressure field at 2000 m has no horizontal variation. Dynamic height is roughly proportional to the sea-surface height, in meters, multiplied by 10. (Based on Levitus (1982).)

give rise to localized regions where the surface water overlies less dense water. Restratification of a stable water column involves solar heating from above, reduced salinity from precipitation, and other more subtle processes.

The detailed mechanisms by which wind generates small-scale motions (i.e., motions on scales smaller than the mixed-layer depth, such as breaking waves) are quite complex and incompletely understood. Nevertheless, the effect of wind on the surface mixed layer is commonly parameterized through a stress $\tau_w$ on the ocean surface. Wind stress has units of force per unit area. The standard empirical relation has the form of eqn [1].

$$\tau_w = \rho_{air} C_d u^2, \qquad [1]$$

where $\rho_a$ is the density of air, $C_d \approx 10^{-3}$ is a drag coefficient that may depend on wind speed and atmospheric stability, and $u$ is the wind speed 10 m above the sea surface. Ten meters is the standard height that commercial ships use to mount their anemometers, and ship reports still account for most of the direct measurements of wind over the ocean.

## Ekman Dynamics

The small-scale motions that mix temperature and salinity also mix momentum. As a result, the momentum of the wind is efficiently transmitted throughout the mixed layer, thereby accelerating horizontal currents. The resulting motions have large horizontal length scales comparable to those of the wind systems that drive them. In 1905, V.W. Ekman developed a model revealing the influence of the earth's rotation on such large-scale flows. His dynamical model presumed that the force associated with a divergence in the vertical momentum flux is balanced by Coriolis accelerations associated with the horizontal flows. The vertical momentum flux in this wind-driven Ekman layer is the result of turbulent mixing processes that Ekman parametrized using Fick's law. Thus, the vertical turbulent flow of momentum is made proportional to the vertical gradient of the large-scale horizontal velocity. This parametrization involves an uncertain constant of proportionality called the vertical eddy viscosity $A_v$. Ekman's model predicts horizontal currents that simultaneously decrease and rotate with depth. Within this so-called Ekman spiral, currents decrease away from the surface with a vertical scale $D$ known as the Ekman depth (eqn [2]).

$$D = (2A_v/f)^{1/2} \qquad [2]$$

This relation provides our first introduction to the Coriolis parameter $f = 2\Omega \sin\theta$, where $\theta$ is latitude, which appears here because of Coriolis accelerations in the Ekman dynamics. The angular velocity of the earth is a vector of magnitude $\Omega = 2\pi\,\text{day}$ that is parallel to the earth's axis of rotation. The Coriolis parameter equals twice the magnitude of the vector component that is parallel to the local vertical. The vertical component is the only component that creates horizontal Coriolis accelerations with horizontal flow. This dependence on the local vertical and the sphericity of the earth explain the sin $\theta$ factor in the formula for $f$. Thus, for any given flow, Coriolis accelerations are strongest at the poles, negligible at the equator, and smoothly varying in between. As discussed below, this geometric detail has profound implications for the surface/wind-driven circulation.

Consider a typical mixed-layer depth of 30 m at mid-latitudes, where $f \approx 10^{-4}\,\text{s}^{-1}$. By relating these two quantities to an Ekman depth $D$, one deduces a vertical eddy viscosity $A_v \approx 0.05$ m$^2$ s$^{-1}$. This value is many orders of magnitude larger than the kinematic viscosity of water, $\nu \approx 10^{-6}$ m$^2$ s$^{-1}$. The large value of $A_v$ indicates the efficiency of turbulent mixing compared with molecular diffusion. But $A_v$ arises from the use of Fick's law to parametrize the turbulence, and Fick's law is an oversimplified model for turbulence. In fact, details of the wind-mixed layer that depend heavily on $A_v$, such as spiraling velocities, are rarely observed.

There is, nevertheless, one very important and robust conclusion from Ekman theory. The net mass transport (the Ekman transport) within the mixed layer, i.e., the vertical integral of the horizontal flow, has magnitude

$$U_{\text{Ekman}} = \tau/(\rho f) \qquad [3]$$

where $\rho$ is the density of water. This result does not depend on $A_v$. Thus, while the details and vertical extent of the Ekman flow depend on the complexities of mixing, the net Ekman transport does not. Furthermore, Ekman theory predicts that the net transport $U_{\text{Ekman}}$ is directed 90° to the right of the wind in the northern hemisphere and 90° to the left of the wind in the southern hemisphere. This result is quite contrary to what one would find if the earth were not rotating.

The Ekman transport describes the net horizontal motion in a thin surface mixed layer. Implications for flows in the interior of the ocean depend on the large-scale patterns in the wind stress, as shown in **Figure 2**. In particular, westward wind stress near the equator results in poleward Ekman mass transport and eastward wind stress at higher latitudes drives equatorward Ekman transport. This pattern results in a convergence of fluid that gives rise to an elevated sea surface at the center of the clockwise wind system. Ultimately, this horizontally convergent Ekman transport has only one direction in which to go – down. The resultant downward motion at the base of the mixed layer, called Ekman pumping, occurs in all mid-latitude ocean basins. Likewise, at higher latitudes, counterclockwise wind systems cause horizontally divergent Ekman transport, a depressed sea surface, and an upward motion known as Ekman suction.

In the classic wind-driven ocean circulation models discussed below, vertical Ekman flows drive the horizontal geostrophic flow. In fact, the net effect of all the complex motions in the mixed layer is often reduced to a simple prescription of the vertical Ekman-pumping velocity $W_{\mathrm{Ekman}}$ (eqn [4]).

$$W_{\mathrm{Ekman}} = \mathrm{curl}\,(\tau_w/\rho f) \qquad [4]$$

where $\mathrm{curl}(\tau_w/\rho f)$ represents the curl of the surface wind-stress vector divided by $\rho f$. Thus, it is not simply the magnitude of the wind stress that determines the Ekman pumping velocities, but its spatial distribution. The Ekman-pumping velocity is often applied as a boundary condition at the sea surface associated with a negligibly thin mixed layer.

On average, Ekman-pumping speeds rarely exceed 1 $\mu$m s$^{-1}$. Nevertheless, such minuscule vertical velocities give rise to the most massive current systems in the ocean. This remarkable fact reflects the enormous constraint that the earth's rotation plays in large-scale ocean dynamics.

## Large-scale Dynamics

The directly wind-driven flows within the Ekman layer occupy only a small fraction of the total water column. In fact, the impact of the wind extends considerably deeper. The connection between the minute Ekman-pumping velocities and the tremendous horizontal flows associated with the surface/wind-driven circulation involves a balance of forces that is very different from that in the surface mixed layer.

Far from continental boundaries, and below the surface mixed layer, the basin-scale circulation varies on length scales measured in thousands of kilometers. The time-averaged horizontal velocities sometimes exceed 1 m s$^{-1}$, but they more generally vary between 1 and 10 cm s$^{-1}$. With these scales, flows are plausibly geostrophic. Furthermore, when the density is uniform, geostrophic flows exhibit no vertical variation. Rather, they behave like a horizontal continuum of vertical columns of fluid. It is reasonable to approximate the region between the mixed layer and the thermocline as a region of constant density. In this region the geostrophic columns of fluid span many hundreds of meters. The earliest models of the ocean circulation obtained remarkable predictive skills by assuming that columnar geostrophic flow extends from the top to the bottom of the ocean.

Downward Ekman-pumping velocities, as small as they may be, effectively compress the fluid columns. Under the influence of downward Ekman pumping, fluid columns below the mixed layer compress vertically and expand horizontally, as if they were conserving their total volume. Likewise, Ekman suction causes water columns to stretch vertically and contract horizontally. Because of the earth's rotation, the effect of Ekman pumping and suction on large-scale motions is related to the principle of angular momentum conservation in classical mechanics. For example, a water column that undergoes the stretching effect of Ekman suction is not unlike a rotating figure skater who draws in her arms, thereby decreasing her moment of inertia and rotating more rapidly. (Note: The analogy is incomplete because Ekman pumping is a consequence of external forcing, whereas the spinning skater is unforced. Nevertheless, the comparison is physically relevant.) Ekman pumping is then similar to a skater extending her arms, which causes a reduction of rate of spin.

The connection between water-column stretching and horizontal flow in the ocean involves one additional subtle point: Water columns on the earth rotate by virtue of their location on the earth's surface. Vertical fluid columns that appear stationary in the earth's frame of reference are actually rotating at a rate that is proportional to the vertical component of the earth's angular velocity. The absolute rotation rate is the sum of the earth's rotation plus the rotation relative to the earth. More precisely, the fluid's absolute vorticity includes a contribution from the planetary vorticity, which has magnitude equal to the Coriolis parameter $f$, plus a contribution from its relative vorticity.

Relative vorticity is measured from a frame of reference that is fixed on the earth's surface. The earth itself rotates at a rate of one revolution per day. In comparison, large-scale ocean currents progress around an ocean basin at average speeds much less than 1 m s$^{-1}$, so the period of rotation associated with a complete circuit can be several years. Thus, the relative rotation rate of large-scale ocean currents is negligible compared with the rotation rate of the earth itself. For this reason, it is a very good approximation to neglect the relative vorticity and equate the vorticity of the large-scale circulation with $f$. The critically important result is that fluid columns change their vorticity by changing their latitude. Just as the skater who extends her arms starts to rotate more slowly, a water column undergoing the compression of Ekman pumping will travel toward the equator where the magnitude of $f$ is smaller.

The physical mechanism that connects Ekman pumping and subsurface geostrophic flow was described mathematically by H.U. Sverdrup in 1947, and is summarized by the relation (eqn [5]).

$$\beta V = f(W_{\mathrm{Ekman}} - W_{\mathrm{deep}}) \qquad [5]$$

$V$ represents the meridional (positive northward) velocity integrated over the depth $H$ of the water column, $\beta = (1/R)\partial f/\partial\theta$ is the meridional variation in the Coriolis parameter, $R$ is the radius of the earth, $W_{\mathrm{Ekman}}$ is the Ekman pumping velocity, and $W_{\mathrm{deep}}$ is a vertical velocity at depth $H$. In this model, $V$ is the depth-integrated geostrophic velocity. While prescription of $W_{\mathrm{deep}}$ is discussed further below, the early models of the wind-driven circulation assumed that $W_{\mathrm{deep}} = 0$ at some depth well below the main thermocline. Thus, $V$ can be estimated using the Sverdrup relation with $W_{\mathrm{Ekman}}$ prescribed using eqn [1], eqn [4] and measurements of wind. $V$ computed in this way agrees remarkably well with $V$ computed from geostrophic flows estimated from the observed density field. The agreement is good everywhere except near the western boundaries of the ocean basins.

## Westward Intensification

While the Sverdrup relation provides guidance for the ocean interior, it cannot describe the basin-wide circulation. From a mathematical viewpoint, the Sverdrup relation is a purely local relation between wind stress and meridional flow, so it does not determine east–west flow within the basin. Moreover, the Sverdrup relation predicts that $V$ will be large only where $W_{\mathrm{Ekman}}$ is large. In fact, $V$ is observed to be largest in the intense western boundary currents found in each ocean basin, such as the Gulf Stream. This is problematic because $W_{\mathrm{Ekman}}$ fails to exhibit the westward intensification, or a change in sign. This means that the Sverdrup relation predicts weak western boundary flow in the wrong direction! Thus, the Sverdrup balance must break down, at least in certain regions.

It is common to presume that Sverdrup theory holds everywhere in the ocean interior except near the western ocean boundary (and northern and southern boundaries if the wind-stress curl does not vanish there). Then, $V$ can be integrated from east to west to determine the total transport required in a poleward western boundary current that returns the meridional Sverdrup transport back to its place of origin. This calculation comes close to predicting the observed transport in the poleward-traveling Gulf Stream at some locations. But models that predict the structure and location of the return flow require fundamentally different dynamics.

In 1948, H. Stommel developed a theory for the wind-driven circulation in which the ocean bottom exerts a frictional drag on the horizontal flow. In the ocean interior, the Stommel and Sverdrup dynamics are nearly indistinguishable, but bottom friction becomes important near the western boundary, allowing Stommel's model to predict a closed circulation for the entire ocean basin (**Figure 5**). The friction-dominated western boundary layer contains

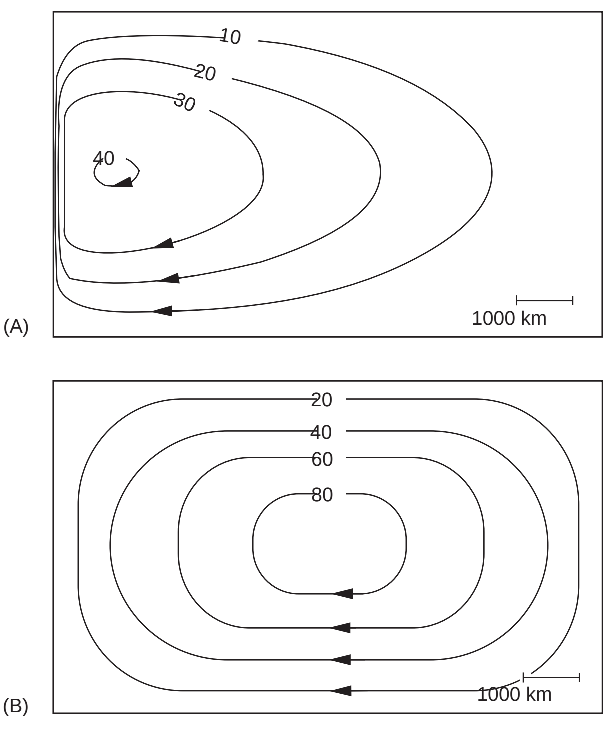


**Figure 5** Streamlines from Stommel's model indicating the total flow in an idealized flat-bottom subtropical gyre. The flow is everywhere parallel to the streamlines in the direction indicated by the arrows. Flow intensity is greatest where the streamlines are closest together. (A) An idealized subtropical gyre for a rotating earth in which the Coriolis parameter varies with latitude. (B) Streamlines for a 'uniformly' rotating earth, that is, for a Coriolis parameter that does not vary with latitude. (From Stommel (1948).)

the intense poleward analogue of the Gulf Stream. Stommel showed the remarkable fact that westward intensification of the wind-driven gyres is fundamentally linked to the latitudinal variation of the Coriolis parameter. That is, the Gulf Stream and its western-boundary analogues in all the ocean basins exist because of the sphericity of the rotating earth (**Figure 6**).

Friction is the key to closing the circulation cell. In Stommel's model the friction parametrization was chosen for its simplicity, but it is ultimately unrealistic. In 1950, W. Munk developed a similar flat-bottom model with lateral viscosity, an entirely different form of dissipation. Nevertheless, Munk's model produces an intense western boundary current for the same reasons as does the Stommel model.

The primary results from these frictional models are robust. Both theories deduce the zonal flow within the basin, and share the central conclusion that the return flow for the interior Sverdrup transport occurs in a meridional current near the western edge of the basin. This current is an example of a boundary layer, a narrow region governed by different physical balances from those dominating the larger domain. In the western boundary layer, fluid columns can change latitude because dissipation changes their vorticity, thereby counteracting the effects introduced by the Ekman pumping or suction. Both models show that this dissipative mechanism can only occur in an intense western boundary layer.

Dissipation in both models actually parametrizes many interesting smaller-scale phenomena. This is evident in Munk's model. The horizontal viscosity needed to produce a realistic Gulf Stream is many orders of magnitude larger than molecular viscosity, larger even than Ekman's vertical eddy viscosity, $A_v$. Modern theories show that these viscous parametrizations for ocean turbulence greatly oversimplify the effect of small-scale motions on the large-scale circulation. More importantly, the Stommel and Munk models neglect the fact that the ocean has variable depth and density stratification.

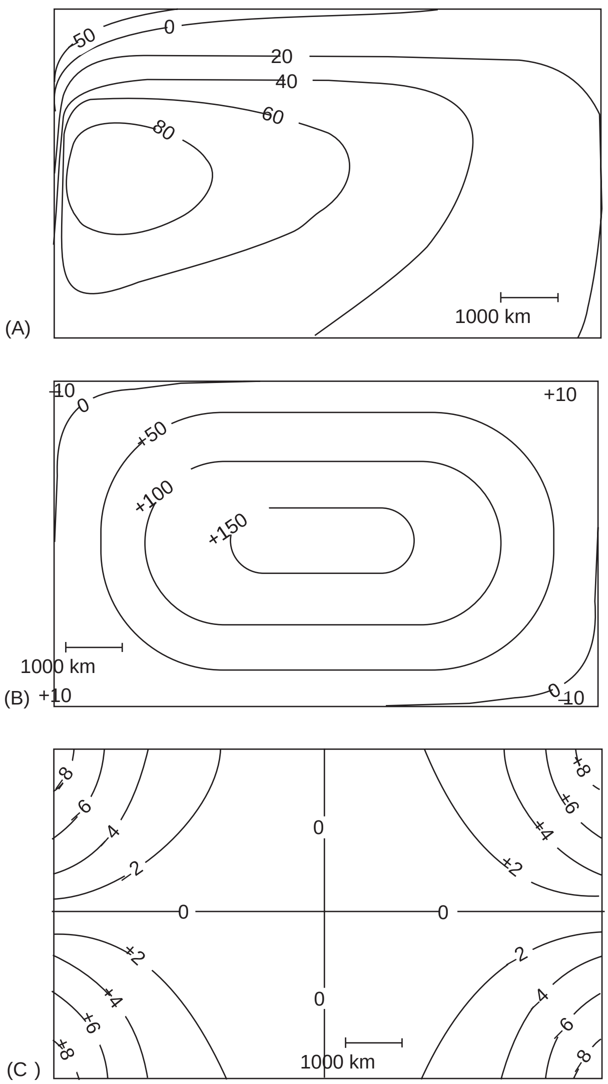


**Figure 6** Contours of sea-surface height from Stommel's model. (A) Sea-surface height consistent with the stream function of Figure 5. Features in this figure can be directly compared with the dynamic height computed from data in Figure 4. (B) Sea-surface height for Stommel's model after setting the Coriolis parameter to a constant. This effectively removes the geometrical factor associated with sphericity. Stommel referred to this as the case of a 'uniformly rotation ocean'. (C) The sea-surface height for the same wind distribution as in (A) and (B), but for a nonrotating ocean. (From Stommel (1948).)

## Topography, Stratification, and Nonlinearity

The simplified Stommel and Munk models describe the wind-driven circulation for a rectangular ocean that has uniform density, a flat bottom, and vertical side walls. It remains to put these idealized models in context for an ocean that has density stratification, mid-ocean ridges, and continental slopes and shelves. The flat-bottom constant-density models clearly oversimplify the ocean geometry. Were the mid-ocean ridges placed on land, they would stand as tall as the Rockies and the Alps. The assumption of constant density turns out to be an oversimplification of comparable proportions.

In flat-bottom models, deep currents are unimpeded by topographic obstructions. With realistic bathymetry, however, flow into regions of varying depth can lead to large vertical velocities. For rotating fluid columns, these vertical velocities affect vorticity. Computer models that add realistic bathymetry and Ekman pumping to the Stommel or Munk models show that such vertical velocities can substantially alter the horizontal flow pattern, so much so that the flows in the center of the ocean no longer resemble the observed surface circulation. Thus, in idealized constant-density models, realistic bathymetry eliminates the most remarkable similarities between the models and the ocean observations.

This conundrum can be reconciled in a model that has variable density. In a constant-density ocean, geostrophic fluid columns extend all the way to the bottom. This allows bottom topography to have an unrealistically strong influence on the flow. Density stratification reduces the vertical extent of columnar motion. Conceptually, a stratified ocean behaves almost like a series of distinct layers, each with variable thickness and constant density. For example, the main thermocline may be considered the interface between one continuum of fluid columns in a surface layer and a second continuum of fluid columns in an abyssal layer. Generalizations of the Stommel and Munk models have often treated the ocean as two distinct layers of fluid.

The main thermocline varies smoothly compared with the ocean bottom. This means that there are fewer obstructions to the columnar flow above the thermocline than below. In this sense, the thermocline effectively isolates the ocean bathymetry from the surface circulation. In fact, observed currents above the main thermocline tend to be stronger. While the Sverdrup theory applies to the top-to-bottom transport, stratification allows the flow to be surface intensified. Smaller abyssal velocities reduce the influence of bottom topography. Flat-bottom models describe a limiting case where the topographic effects are identically zero.

Without question, the vertical extent of the large-scale wind-driven circulation is linked to density stratification. Realistic models of the large-scale circulation must include thermodynamic processes that affect temperature, salinity, and density structure. For example, atmospheric processes change the heat and fresh water content of the surface mixed layer. Large-scale motions can result when the water column becomes unstable, with more dense water overlying less dense water. The resulting motion is often referred to as the thermohaline circulation, as distinct from the wind-driven circulation, but the conclusion to be drawn from the more realistic ocean-circulation models is that the thermohaline circulation and the wind-driven circulation are inextricably linked.

Additional factors come into play in the more comprehensive ocean models. For example, the persistent temperature and salinity structure of the ocean indicates that many large-scale features in the ocean have remained qualitatively unchanged for decades, perhaps even centuries. But there are no simple (linear) theories that predict the existence of the thermocline. The transport and mixing of density by ocean currents are inherently nonlinear effects. Other classes of nonlinearities inherent to fluid flow add other types of complexity. Such nonlinear effects account for Gulf Stream rings, mid-ocean eddies, and much of the distinctly nonsteady character of the ocean circulation. Ocean currents are remarkably variable. Variability on much shorter timescales of weeks and months, and length scales of tens and hundreds of kilometers, often dominates the larger-scale flows discussed here. Thus, it is not appropriate to think of the ocean circulation as a sluggish, linear, and steady. Instead, it is more appropriate to think of the ocean as a complex turbulent environment with its own analogues of unpredictable atmospheric weather systems and climate variability. Nevertheless, the simplified theories of steady circulation illustrate important mechanisms that govern the time-averaged flows.

In closing, two ocean regions deserve special mention: the equatorial ocean and the extreme southern ocean. Equatorial regions have substantially different dynamics compared with models discussed above because Coriolis accelerations are negligible on the equator, where $f = 0$. The wind-related processes that govern El Niño and the Southern Oscillation, for example, depend critically on this fact. The southern ocean distinguishes itself as the only region without a western (or eastern) continental boundary. This absence of boundaries produces a circulation characteristic of the atmosphere, with intense zonal flows that extend around the globe. They represent some of the most intense large-scale currents in the world, and derive much of their energy from the wind. So they too represent an important part of the surface/wind-driven circulation.

## See also

**Atlantic Ocean Equatorial Currents. Benguela Current. Brazil and Falklands (Malvinas) Currents. Canary and Portugal Currents. Current Systems in the Atlantic Ocean. Florida Current, Gulf Stream and Labrador Current. Surface, Gravity and Capillary Waves. Thermohaline Circulation.**

## Further Reading

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