ON THE WIND-DRIVEN OCEAN CIRCULATION

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ABSTRACT

Streamlines of oceanic mass transport are derived from solutions to a vertically integrated vorticity equation which relates planetary vorticity, lateral stress curl, and the curl of the stress exerted by the winds on the sea surface. These solutions account for many of the gross features of the general ocean circulation, and some of its details, on the basis of the observed mean annual winds.

The solution for zonal winds (section 3) gives the main gyres of the ocean circulation. The northern and southern boundaries of these gyres are the west wind drift, the equatorial currents, and equatorial countercurrent. They are determined by the westerly winds, the trades, and the doldrums, respectively. For each gyre the solution gives the following observed features (from west to east): a concentrated current (e.g., the Gulf Stream), a countercurrent, boundary vortices (the Sargasso Sea), and a steady compensating drift. Using mean Atlantic zonal winds, the solution yields a transport for the Gulf Stream of 36 million metric tons per second, compared to 74 million as derived from oceanographic observations. The discrepancy can probably be ascribed, at least in part, to an underestimate of the wind stress at low wind speeds (Beaufort 4 and less) as derived from the relationship now generally accepted.

The solution for meridional winds (section 5) accounts for the main features of the current system off California. For a circular wind system (section 8) the solution gives a wind-span vortex which is displaced westward in relation to the wind system, in agreement with observations in the Northeast Pacific high-pressure area.

Based on these three solutions, a general nomenclature of ocean currents is introduced (section 9), applicable to all oceans regardless of hemisphere, and suggestive of the meteorologic features to which the currents are so closely related. In the light of this general system, the circulations of the northern and southern hemispheres, and of the North Atlantic and North Pacific are compared (section 10). Rossby's jet-stream theory of the Gulf Stream, and Maury's theory of thermohaline circulation are discussed, and it is concluded that the circulation in the upper layers of the oceans are the result chiefly of the stresses exerted by the winds.

1. Introduction

This study is a resumption of Ekman's (1923; 1932) intensive effort to account theoretically for the main features in the general ocean circulation. The temptation to reopen this problem is provided by three recent developments: (a) Rossby's (1936) introduction of lateral stresses associated with the horizontal exchange of large eddies; (b) Sverdrup's (1947) computation of the equatorial countercurrents in a baroclinic ocean from the known wind stress; and (c) Stommel's (1948) explanation of the westward intensification of a wind-induced circulation in terms of the variation of the Coriolis parameter with latitude.

In Ekman's and Stommel's model the ocean is assumed homogeneous, a case in which the currents extend to the very bottom. Not only is this in contrast with observations, according to which the bulk of the water transport in the main ocean currents takes place in the upper thousand meters, but it also leads to mathematical complications which rendered Ekman's analysis very difficult, and forced Stommel to resort to a rather arbitrary frictional force along the bottom.

To avoid these difficulties, we retain Sverdrup's integrated mass transport as the dependent variable. This device makes it possible to examine the more general case of a baroclinic ocean without having to specify the nature of the vertical distributions of density and current. In recognition of the evidence that currents essentially vanish at great depths, we shall depend on lateral friction for the dissipative forces. From Stommel we retain the rectangular boundaries, although we extend the basin to both sides of the equator and deal with the observed wind distribution rather than a simple sinusoidal distribution.

2. The differential equation of mass transport

Let the $x$-axis point eastward along the equator, the $y$-axis northward, and the $z$-axis vertically upwards from a level surface (assumed plane) just beneath the sea surface (unit vectors $\textbf{i}$, $\textbf{j}$, $\textbf{k}$). For the present, the nonlinear inertial terms are considered negligible, an assumption which will be examined in section 10. The equations of steady horizontal motion for a unit
\[ \nabla \rho + f k \times (\rho \nabla H) - \frac{\partial}{\partial x} \left( A_H \frac{\partial \nabla H}{\partial x} \right) - \frac{\partial}{\partial y} \left( A_H \frac{\partial \nabla H}{\partial y} \right) - \frac{\partial}{\partial z} \left( A_H \frac{\partial \nabla H}{\partial z} \right) = 0, \quad (1) \]

where \( \nabla H = i \partial / \partial x + j \partial / \partial y \), \( \rho \) is pressure, \( \rho \) is density, \( \nabla H = i \nu_x + j \nu_y \) is the two-dimensional (horizontal) velocity vector, and \( f = 2 \omega \sin \phi \) is twice the vertical component at latitude \( \phi \) of the earth's angular velocity \( \omega \). Observations have shown that \( A' \), the dynamic eddy viscosity associated with vertical shear, is different from \( A_H \), the eddy viscosity pertaining to horizontal shear, the two coefficients bearing a ratio rather larger than that of the width of the ocean currents to their depth. We shall assume \( A' \) horizontally isotropic, neglecting variations that might be related to differences between zonal and meridional motions of large horizontal eddies on a rotating earth.

The equations of motion will be integrated from some constant depth, \( -h \), beneath which the motion and horizontal pressure gradients essentially vanish (Sverdrup, 1947), to the surface, \( z = z_0(x, y) \). Let

\[ P = \int_{-h}^{0} \rho \, dz, \quad M_H = \int_{-h}^{0} \rho \nabla H \, dz, \quad (2a, b) \]

designate the integrated pressure and mass transport. Then

\[ \int_{-h}^{0} \nabla \rho \, dz = \nabla H P - \rho a \nabla H z_0, \]

where \( \rho_a \) is the atmospheric pressure and \( \nabla H z_0 \) is the gradient of the sea surface. The second term is negligible for the present investigation.

The integration of the horizontal shearing forces is more cumbersome. The simplest form is \( A_H \nabla H \rho M_H \), where \( A_H \) is an effective kinematic eddy viscosity which is assumed independent of \( x \) and \( y \). There is no evidence to indicate that the complicated expression given by a complete integral with a variable \( A_H = A' / \rho \) would be a more accurate description of the forces. Finally

\[ \int_{-h}^{0} \frac{\partial}{\partial z} \left( A_H \frac{\partial \nabla H}{\partial z} \right) \, dz = \left[ A_H \frac{\partial \nabla H}{\partial z} \right]_0 - \left[ A_H \frac{\partial \nabla H}{\partial z} \right]_0 = \tau, \quad (3) \]

where \( \tau \) is the stress applied by the wind on the sea surface. In most ocean areas it is adequate to select a depth \( h \) of 1000–2000 meters, which is comfortably above the sea bottom. The foregoing equation could be used even when the current extends to the bottom, provided \( \tau \) would then designate surface stress minus bottom stress.

With the understanding that all operators and vectors have only horizontal components, the subscript \( H \) can be dropped, and the integrated equations of motion become

\[ \nabla P + f k \times M = \tau - A' \nabla M = 0. \quad (4) \]

The vertical integration enables us, therefore, to avoid any assumptions regarding the variation of \( A' \) with depth, and to postpone an examination of the density and velocity structure in depth.

The surface kinematic boundary condition requires \( w_z \) to vanish for steady state; furthermore it has been assumed that \( w_- = 0 \). Therefore \( \nabla M = 0 \), or

\[ M = k \times \nabla \psi, \quad (5) \]

where \( \psi \) is the mass-transport stream function (dimension: mass/time). Substituting \( \psi \) into (4) and performing the curl operation to eliminate \( P \) gives the differential equation of mass transport

\[ (A' \nabla \psi - \beta \partial / \partial x) \psi = - \text{curl} \, \tau, \quad (6) \]

where \( \text{curl} \, \tau \) is the vertical component of the wind-stress curl, \( \nabla \psi = \partial / \partial x + 2 \beta / \partial x \partial \psi^2 + \partial / \partial y \) is the biharmonic operator, and \( \beta = df / dy \) is the rate of change northward of twice the vertical component of the earth's angular velocity.

Equation (6) is an integrated version of the vorticity equation. Ekman calls \( \beta \partial \psi / \partial x = \beta \psi \), the planetary vorticity. Accordingly, the vorticity equation (6) expresses a balance between lateral stress curl, planetary vorticity and wind-stress curl.

For boundary conditions we choose

\[ \psi_{b=0} = 0, \quad (\partial \psi / \partial x)_{b=0} = 0, \quad (7a, b) \]

where \( \nu \) is normal to the boundary. The first equation states that the boundary itself is a streamline (whose value is arbitrarily selected as zero); the second equation states that no slippage takes place against the boundary.

The problem is to solve (6) subject to boundary conditions (7). Hidaka (1949) has given the formal solution employing the transformation into a variational problem. There is a remarkable analogy between this problem and a problem arising in the theory of elasticity (Rayleigh, 1893; Love, 1944, p. 464). The deflection \( \psi \) of a plate of uniform flexural rigidity \( A \), clamped at its boundaries and subjected to a load \(- \text{curl} \, \tau \), is governed by

\[ A \nabla \psi = - \text{curl} \, \tau, \quad (8) \]

subject to boundary conditions (7). Equation (6) reduces to this form when the planetary vorticity is negligible, a condition that might arise in high latitudes (\( \beta = 0 \)) and for quasigeostrophic flow (\( M \approx 0 \)). The analogy suggests experimental methods for finding streamlines under complex wind stress and boundary conditions.
3. Solution for zonal winds

A rectangular ocean is assumed, extending from $x = 0$ to $x = r$, and from $y = -s$ to $y = +s$. One may think that such an arbitrary choice of ocean boundaries would make it impossible to deal with actual conditions. Yet the proper solution for the main features in the ocean circulation can be expected to be rather insensitive to the choice of boundaries, because these main features are similar in all oceans (table 3) although the boundaries are not. The boundary conditions (7) become

$$\psi = 0, \quad \partial \psi / \partial x = 0, \quad \text{for} \quad x = 0, \quad x = r; \quad (9)$$

$$\psi = 0, \quad \partial \psi / \partial y = 0, \quad \text{for} \quad y = -s, \quad y = s. \quad (10)$$

For a zonal wind circulation ($\tau_y = 0$) the stress on the sea surface in the interval $-s < y < +s$ can be written as a sum of terms like

$$\tau_x = a \cos ny + b \sin ny + c$$

where $n = j\omega / s$, $j = 1, 2, \cdots$ etc. The vertical component of the wind-stress curl is then a sum of terms like

$$\text{curl}_{\tau} = -\partial \tau_x / \partial y = n(a \sin ny - b \cos ny), \quad (11)$$

and the particular integral of (6) becomes

$$\psi_{pn} = -n^{-2} A^{-1}(a \sin ny - b \cos ny)$$

$$= -n^{-2} A^{-1} \text{curl}_{\tau}. \quad (12)$$

Let $k = \sqrt{\beta / A}$ designate the Coriolis-friction wave number, assumed constant. The solution to the homogeneous equation

$$(\nabla^4 - k^4 \partial^4 / \partial x^4) \psi_{hn} = 0, \quad (13)$$

can be written as a product of $(a \sin ny - b \cos ny)$ and a function of $x$. For later convenience we choose the form

$$\psi_{hn} = \psi_{pn}(r A n^{-4} \beta^{-1} X_n - 1), \quad (14)$$

where $X$ is a function of $x$ to be determined below. The complete integral equals

$$\psi_x = \psi_{pn} + \psi_{hn} = r A n^{-4} X_n \beta^{-1} \psi_{pn},$$

or

$$\psi_x = -r X_n \beta^{-1} \text{curl}_{\tau}. \quad (15)$$

![Fig. 1. Roots $\alpha' + i\alpha''$ to characteristic equation (16). The dashed lines give the asymptotic solutions for $\gamma \ll 1$ which have been applied to the large scale features ($\gamma < 0.25$) in the ocean circulation.](image)

It follows from (12), (13) and (14) that the function $X(x)$ must obey the differential equation

$$X'''' - 2n^2 X'' - k^4 X' + n^4 X = \beta / rA = k^4 / r,$$

which yields

$$\gamma_n^4 k^4 X = 1 + \frac{1}{2}(p_1 + i p_2) e^{ikx} + \frac{1}{2}(p_1 - i p_2) e^{-ikx} + p_4 e^{ikx} + p_4 e^{-ikx},$$

where $\gamma = n/k$ is the essential nondimensional number in this investigation, and where $\alpha_1, \alpha_2, \alpha_3, \alpha_4$ are the roots of the characteristic equation

$$(a^2 - \gamma_n^2)^2 = \alpha. \quad (16)$$

This equation has two complex roots, $\alpha_1' + i\alpha_1''$, $\alpha_2' + i\alpha_2''$, and two real roots, $\alpha_1', \alpha_1'$ (fig. 1). For small values of $\gamma$,

$$\alpha_1 = -\frac{1}{2}(1 + i \sqrt{3}), \quad \alpha_2 = \gamma_n^4,$$

$$\alpha_3 = -\frac{1}{2}(1 - i \sqrt{3}), \quad \alpha_4 = 1, \quad (17)$$

as shown by the dotted lines in fig. 1. The boundary conditions (9) give

$$-p_1 \approx p_2 \sqrt{3} \approx 1 + p_3 \approx p_4 \approx \gamma_n^4 k, \quad (18)$$

provided

$$\gamma_n^4 \ll 1, \quad e^{-kr} \ll 1, \quad e^{-\gamma_n^4 k} \approx 1 - \gamma_n^4 k. \quad (19)$$

Subject to these approximations,

$$X = -K e^{-ikx} \cos \left( \frac{\sqrt{3}}{2} kx + \frac{\sqrt{3}}{2kr} - \frac{\pi}{6} \right) + 1 - \frac{1}{kr} \left[ kx - e^{-kr} - 1 \right]$$

$$\frac{X'}{k} = K e^{-ikx} \sin \left( \frac{\sqrt{3}}{2} kx + \frac{\sqrt{3}}{2kr} \right) - \frac{1}{kr} \left[ 1 - e^{-kr} \right], \quad (20)$$

where $K = 2 / \sqrt{3} - \sqrt{3} / kr$. The meaning of the horizontal braces will be discussed later.

The important thing to notice is that, to the present degree of approximation, $X$ does not depend on $n$; i.e., that the response of the current to the zonal wind stress is without distortion in the $y$-direction for
\( \gamma \ll 1 \). Equation (15) can therefore be written
\[
\psi = -rX\vartheta^{-1}\text{curl}, \tau,  \tag{22}
\]
and the stream function computed directly from the observed (zonal) stress without expanding the stress in a Fourier series, provided such a series converges so rapidly that for all significant terms conditions (19) are fulfilled. Whether this is the case can usually be told by inspection. Anticipating the numerical values \( k = 0.016 \text{ km}^{-1}, r = 6000 \text{ km}, \) conditions (19) for 10 per cent accuracy in \( \psi \) give \( \gamma < 0.25 \), corresponding to a minimum zonal wave length \( 2\pi/n \) of about 1500 km. For the mean annual stress distribution (fig. 2) the shortest wave length of the important N-S variations is the distance between the northern and southern trades, 4000 km. The approximations leading up to (22) appear to be valid for a study of the general ocean circulation as it is related to the general atmospheric circulation.

For stationary compact cyclones and sharp fronts with overtones of less than 1500 km wave length (\( \gamma > 0.25 \)), it is necessary to compute separately the components \( \psi \), for each component wind stress \( \tau \), according to equation (15), and to sum. The roots \( \alpha \) can be read off fig. 1 and the \( p \)'s determined by the original boundary conditions (9).

There remain the conditions (10) at the northern and southern boundaries. We note (fig. 9) that the winds are weak at 60°-70° latitudes, and we shall satisfy (10) by assuming curl, \( \tau \) and its derivative to approach zero at these boundaries. It must be admitted that this easy way of satisfying boundary conditions is aided by the present lack of adequate wind data in high latitudes.

To the extent to which these conditions on curl, \( \tau \) are not satisfied there must be deviations from the present solution, but we may expect these deviations to be restricted to relatively narrow zones adjacent to the northern and southern boundaries for the following reasons: The general circulations in the Atlantic and Pacific Oceans are in a very rough sense symmetrical with respect to the equator (except for an asymmetry related to the asymmetry of the wind circulation), whereas the boundaries are not. To the south all oceans open into the Antarctic Sea, to the north they are practically closed. Even the Indian Ocean, with its northern boundary only 1000-2000 km north of the equator, has about the same circulation in its central and southern parts as the Atlantic and Pacific.

It is none the less possible that important features have escaped us by the present incomplete treatment,
and we hope to be able to give a complete numerical solution (for actual boundaries) at a later time. The similar but much simpler problem of a rectangular plate clamped at its edges and subject to a uniform load (equations 8, 9, 10 with \( \nabla \times \mathbf{v} \) constant) has received considerable attention, yet no complete theoretical solution has been found, and recourse has had to be made to rather elaborate numerical methods (Love, 1944, pp. 493–497). We note, however, that the solution in the vicinity of the transverse axis (corresponding to low and middle latitudes) for a plate which is more than twice as long as broad is practically the same as if the length were infinite.

4. Zonal wind currents

The wind stress has been computed for the North Pacific and the North and Central Atlantic for each month and 5-degree quadrangle. The source material consisted of the wind roses on the U. S. Hydrographic Office Pilot Charts of the North Pacific Ocean, the summary of Marine Data Cards of the U. S. Weather Bureau, and the Monthly Meteorological Charts of the Atlantic Ocean prepared by the Meteorological Office, Air Ministry, London. The variability of the wind and the hydrodynamic character of the sea surface has been taken into account in the manner described by Reid (1948). Fig. 2 shows the mass transport streamlines \( \psi(x, y) \) in a rectangular ocean computed from (22) for the mean annual zonal winds over the Pacific. The topography of the free surface and of other isobaric surfaces is related to the \( \psi \)-topography, the relationship depending however on the vertical density structure. For the equatorial Pacific, Reid (1948b) has successfully used a simple model of vertical density distribution consisting of a homogeneous layer of density \( \rho_0 \) and thickness \( h' \) above an exponential layer of density \( \rho_{-\infty} - (\rho_{-\infty} - \rho_0) \exp(1 + z/h') \).

He finds

\[
P + \text{constant} = \frac{\rho_0^2}{\rho_{-\infty} - \rho_0} z_0^2
\]

\[
= (5/2) g (\rho_{-\infty} - \rho_0) h' = f \Psi
\]

for the elevation \( z_0 \) of the surface above, and the depression of \( h' \) beneath their mean level.

The major ocean gyres.—The zonal wind system divides the ocean circulation into a number of gyres, each bounded by latitudes \( \varphi_i \) for which \( M_\varphi = 0 \), or by (22),

\[
\nabla \times \mathbf{v} = 0 \quad \text{at} \quad \varphi = \varphi_i.
\]

The major dividing lines between gyres correspond to the latitudes of maximum west winds, of the northerly and southerly trades, and of the doldrums (fig. 2). These compare favorably with those determined from oceanographic observations.

The latitudinal axis of each gyre may be defined by \( M_\varphi = 0 \), or

\[
\frac{d}{dy} \left( \frac{\nabla \times \mathbf{v}}{\beta} \right) = 0 \approx \frac{d^2 \tau_x}{dy^2} \quad \text{at} \quad \varphi = \varphi_i.
\]

The Atlantic Sargasso Sea is associated with the inflection point in the mean wind stress curve between the westerly winds and the northeasterly trades.\(^2\)

The inflection points between the doldrums and the northern and southern trades determine the boundaries of the equatorial countercurrent.

The western currents.—Along any fixed latitude, \( \psi \) varies with \( X \) only, and the curves \( X(x) \) and \( X'(x) \) in fig. 3 can be interpreted as the north–south component of transport, and transport per unit length, drawn to an arbitrary scale. For definiteness, consider sections at latitudes \( \varphi_i \) through the center of gyres.

When \( X \) and \( X' \) are computed from (20) and (21), it is found that the equations fall naturally into three parts, each of which dominates in a given sector. At

\(^2\) The dependence \( \tau(y) \) of the width of the ocean basin on latitude somewhat modifies this conclusion when applied to non-rectangular ocean basins (Munk and Carrier, 1950).

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![Fig. 3. Plot of equations (20) and (21), giving the west–east variation in transport (~X) and transport velocity (~X') from the western shore (x = 0) to the eastern shore (r – x = 0). The scale for the central and eastern solutions is exaggerated relative to the scale for the western solution.](image-url)
the western edge of the ocean \( x \ll r \), and

\[
X_W = -\frac{2}{\sqrt{3}} e^{-x/k} \cos \left( \frac{\sqrt{3}}{2} kx - \frac{\pi}{6} \right) + 1, \quad (20W)
\]

\[
\frac{X'_W}{k} = \frac{2}{\sqrt{3}} e^{-x/k} \sin \frac{\sqrt{3}}{2} kx, \quad (21W)
\]

representing slightly underdamped oscillations of wave length

\[
L_W = \frac{4\pi}{\sqrt{3}k} = \frac{4\pi}{\sqrt{3}k} A \cdot \beta.
\]

### Table 1. Extrema of \( X_W \) and \( X'_W \).

<table>
<thead>
<tr>
<th>( x/L_W )</th>
<th>( X_W )</th>
<th>( X'_W/k )</th>
<th>( X'_W/k )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( 1/6 )</td>
<td>( 1/6 )</td>
<td>( 4/6 )</td>
<td>( 1 )</td>
</tr>
<tr>
<td>( 0.45 )</td>
<td>( 1.17 )</td>
<td>( 1.09 )</td>
<td>( 0.97 )</td>
</tr>
<tr>
<td>( 0.55 )</td>
<td>( 0.00 )</td>
<td>( -0.99 )</td>
<td>( 0.00 )</td>
</tr>
</tbody>
</table>

A review of oceanographic observations suggests 200 km and 250 km as typical values for the widths of the western currents and countercurrents, respectively. The corresponding values of \( A \) are \( 3.3 \times 10^7 \) and \( 6.5 \times 10^7 \) cm² sec⁻¹ (equation 24), indicating perhaps an increase of \( A \) with distance from shore as might be expected on the basis of Prandtl’s mixing-length theory. We shall set \( A = 5 \times 10^7 \) cm² sec⁻¹. This compares favorably with the values \( 4 \times 10^7 \) and \( 7 \times 10^7 \) cm² sec⁻¹ for the region of the Atlantic equatorial countercurrent, the former determined by Montgomery (1939) from the diffusion of salt, the latter by Montgomery and Palmén (1940) from dynamic principles. One may expect, however, that the value of \( A \) varies with time and location, and that for any fixed time and location it has different values, depending on the dimensions of the phenomenon under investigation. Without any further information we shall be obliged to consider the above value as constant.

The total transports of the western current and countercurrent are found by substituting values for \( X_W \) from table 1 into (22):

\[
\psi_{\text{oc}} = -1.17 \beta \phi \psi, \quad (26)
\]

The resulting expressions are independent of \( A \), and the transports can be computed with a relatively high degree of accuracy, being subject largely to uncertainties of the same order as those in computing wind stress. A comparison is made in table 2 between the transport values of some western currents determined from oceanographic observations, and those computed...
from the zonal wind stress according to (26). The effect of the meridional wind component, which supports the zonal winds, is introduced by a meridional wind factor (see section 7)

$$-\text{curl}_x \tau (d\tau_z/dy)^{-1}.$$  \hspace{1cm} (27)

The approximation involved in permitting the "constants" $\beta$ and $r$ to assume different values at different latitudes implies that these vary slowly compared to $\tau$ and $\psi$. However, not even a rapid change of $r$ introduces an appreciable error into (26); the only essential modification is a widening and corresponding weakening of the western current, determined by the inclination of the western coastline relative to the assumed north–south direction (Munk and Carrier, 1950).

The computed transport values differ from the observed values by as much as a factor of two, a discrepancy not surprising if one considers that, among other uncertainties, wind and current data are not for the same year, nor necessarily for the same time of year. On examining the possible sources of errors, one is led to ascribe some of the discrepancy to an underestimate of wind stress for low wind speeds. We shall return to this question in section 10.

**The western boundary vortices (Sargasso Sea).**—Beyond the countercurrent, the next term in (20) and (21) can no longer be neglected. It pivots the asymptotic solution $X = 1$ to $X = 1 - x/r$, and shifts $X'/k = 0$ to $X'/k = -1/kr$ (fig. 2, bottom). The resulting streamlines represent an exponentially decaying vortex trail centered along the axis $\phi_a$ (fig. 5). Only the two western vortices need be considered. Accordingly the Sargasso Sea (or any of the other dynamically analogous areas) is a region of relatively sluggish circulation containing a moderately strong anticyclonic vortex at the western side and a weak anticyclonic vortex near its center. The two vortices represent two humps in the sea surface which are separated by a trough with a saddle point in the vicinity of Bermuda. Reid's model (equation 23) gives 10 cm and 2 cm for the peak elevations of the western and eastern humps above the saddle point.

Felber (1934) has analyzed the extensive observational material based on surface drifts of vessels, and his streamlines show a saddle point southwest of Bermuda. A chart of the topography of the sea surface prepared by Defant (1941) on the basis of the *Meteor* observations and preceding observations, does indeed show two humps. Recent detailed observations from closely spaced lowerings of bathythermographs, though inconclusive, are not inconsistent with the above picture.

**Central ocean drift.**—Away from both boundaries, the equations

$$X_C = 1 - x/r, \quad X'_C/k = -1/kr, \quad (20C, 21C)$$

$$M_y = \partial \psi/\partial x = \beta^{-1} \text{curl}_x \tau$$  \hspace{1cm} (28)

give a broad constant drift that compensates for the swift, shallow western currents. Equation (28) was derived by Sverdrup (1947). By dropping the lateral stress term this equation could have been written directly from (6) without the restriction to zonal winds. The validity of the theory in mid-ocean has been adequately confirmed by the successful application of (28) to the equatorial currents of the eastern Pacific (Sverdrup, 1947; Reid, 1948a, b).

**Eastern coastal strip.**—The eastern solution

$$X_E = 1 - \frac{x}{r} - \frac{1}{kr} \left[ 1 - e^{-k(r-x)} \right], \quad (20E)$$

$$X'_E/k = \frac{1}{kr} \left[ 1 - e^{-k(r-x)} \right] \quad (21E)$$

**TABLE 2.** The mass transport of some western currents determined from the wind stress (equation 26) and from oceanographic observations.

<table>
<thead>
<tr>
<th>Current</th>
<th>Lat.</th>
<th>$10^4 \beta$ cm$^{-1}$ sec$^{-1}$</th>
<th>$r$ km</th>
<th>$10^4 \frac{d^2 X}{dy^2}$ g sec$^{-2}$</th>
<th>Merid. wind factor</th>
<th>Wind stress g sec$^{-1}$</th>
<th>Oceanogr. g sec$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gulf Stream</td>
<td>35°N</td>
<td>1.9</td>
<td>6,500</td>
<td>70</td>
<td>1.30</td>
<td>36</td>
<td>74 (55)$^4$</td>
</tr>
<tr>
<td>Kuroshio Ext.</td>
<td>35°N</td>
<td>1.9</td>
<td>10,000</td>
<td>50</td>
<td>1.25</td>
<td>39</td>
<td>65$^5$</td>
</tr>
<tr>
<td>Oyashio C.</td>
<td>50°N</td>
<td>1.5</td>
<td>5,500</td>
<td>-13</td>
<td>-6.5</td>
<td>-5</td>
<td>-7$^6$</td>
</tr>
<tr>
<td>Brazil C.</td>
<td>20°S</td>
<td>2.2</td>
<td>5,500</td>
<td>-20</td>
<td>-5.8</td>
<td>-5 to -10$^7$</td>
<td></td>
</tr>
</tbody>
</table>

$^4$ Sverdrup et al. (1942), pp. 605–761.
$^5$ Adjusted for a supposed southward motion of $19 \times 10^4$ g of slope water per second. Sverdrup et al. (1942), p. 679.
$^6$ For August (Uda, 1938).
represents an exponential slippage zone of width, say, $\pi/k$ (fig. 3). For $A = 5 \times 10^3 \text{ cm}^2 \text{ sec}^{-1}$ this corresponds to roughly 200 km. This feature is likely to be obscured by meridional wind currents (section 6).

The west-east asymmetry.—The westward intensification of ocean currents has been explained by Stommel (1948) as related to the planetary vorticity. The asymmetry may be expressed as either of the ratios

$$\frac{M_y(WC)}{M_y(C)} = -0.55kr; \quad \frac{r}{x(WC)} = \frac{\sqrt{3}}{2\pi} kr$$

of the maximum western current to the central drift, or of the width of the ocean to that of the western current. The asymmetry increases with $r$, decreases with $A$ and $\varphi$. For the Atlantic we found $kr \approx 100$.

But the asymmetry is not only one of narrow swift currents versus broad slow currents. The countercurrent and opposing vortices are part of the western dynamic pattern for which there is no counterpart on the eastern side of the oceans. The equivalent mathematical statement is that the real parts of the two complex roots $\alpha_1$ and $\alpha_2$ of the characteristic equations (16) are always negative, whereas the real roots $\alpha_3$ and $\alpha_4$ are positive. The leftward trend in fig. 1 of $\alpha_1'$ and $\alpha_2'$ with increasing $\gamma$ shows that the asymmetry becomes even more pronounced for large zonal wind wave-numbers.

These remarks can be interpreted in terms of the differential equation of mass transport (equation 6):

$$\text{Lateral stress curl + planetary vorticity + wind stress curl} = 0.$$  

\[ \text{western solution} \quad \text{central solution} \]

In the central and eastern ocean areas the planetary vorticity and wind stress curl have opposite signs, and balance is achieved in which the lateral stress plays a negligible part (except in the unimportant eastern coastal zone). Along the western boundary the planetary vorticity and wind stress curl have the same sign, and the lateral stress curl balances both planetary vorticity and wind stress curl. It can be verified that in this region the wind stress curl is numerically unimportant, although it is of course the primary cause of the circulation. The above remarks apply not only to both hemispheres, but also to the cyclonic as well as the anticyclonic gyres. The only way in which the intense currents and the boundary vortices could occur on the eastern side of the oceans would be for the earth to rotate in the opposite sense.

\[ \text{Dynamically it is perhaps more significant to consider the} \]
\[ \text{concentration of virtually all the oceans' vorticity against their} \]
\[ \text{western shores.} \]

5. Meridional wind solution

Superimposed on the eastern drift we find in various oceans rather well developed currents along the eastern boundary, the Benguela, California, Peru and Alaska currents being the outstanding examples. These currents are seasonal and vary with the winds. Since they are not contained in our solution for a zonal wind circulation we must assume that they depend on the vorticity provided by the meridional winds.

The procedure to be followed is similar to the one in section (3). The meridional wind stress in the interval $0 < x < r$ will be written as a sum of terms $\tau_{ym} = a \cos mx + b \sin mx + c$, where $m = j\pi/r, j = 1, 2, \ldots$ etc. Neglecting the effects of northern and southern boundaries, $\psi = \psi(x)$, and equation (6) reduces to

$$\left( A \frac{d^4}{dx^4} - \beta \frac{d}{dx} \right) \psi_m = -\text{curl}_m \tau = -\frac{d\tau_{ym}}{dx}. \quad (6M)$$

The general solution for the $m^{th}$ component of $\psi$ is

$$\psi_m = \frac{\alpha_m^2 d\tau_{ym}}{m \frac{dx}{dx}}$$

$$\frac{k^4A(1 + \delta_m^2)}{\delta_m^4}$$

$$+ e^{-ikx} \left[ q_1 \cos \left( \frac{\sqrt{3}}{2} kx - \theta \right) + q_2 + q_4 e^{-k(r-x)} \right]. \quad (15M)$$

where $\delta_m = m/k$ is the ratio between the meridional wind stress wave number $m$ and the Coriolis-friction wave number $k = \sqrt[4]{\beta/A}$, and the $q$'s are constants to be determined from the four boundary conditions (9).

In the central and eastern ocean areas ($e^{-ikx} \ll 1$) for a relatively broad meridional wind system ($\delta \ll 1$) equation (15M) gives

$$\psi(x) = \beta^{-1} \left[ \tau_y \tau_y - (\tau_y)^2 \right]$$

$$+ k^{-1}(\text{curl}_m \tau), (1 - e^{-k(r-x)}) \right], \quad (29)$$

$$M_y(x) = \beta^{-1} \left[ (\text{curl}_m \tau)_y - (\text{curl}_m \tau)_x e^{-k(r-x)} \right], \quad (30)$$

independent of the wave-number $m$. Beyond the eastern coastal strip, (30) reduces again to Sverdrup's equation (28).

6. The California current system

Off the coast of California, perhaps as a consequence of the coastal mountain range, the usual situation is for the summer winds south of Cape Mendocino (the Cape Horn of California) to blow from NNW and to reach their maximum speed some distance from shore (fig. 6). Along this wind axis the curl of the wind stress changes sign, and, neglecting the very small last term in (30), the direction of transport changes sign also.
The existence of this biologically important countercurrent is confirmed by an oceanographic section off Monterey occupied by the *E. W. Scripps* on 11–14 June 1939 (fig. 6). The width of the area during this cruise appeared to have been considerably narrower than would be consistent with average June winds. The total transport of the countercurrent according to (29) is about $6 \times 10^{11} \text{ g sec}^{-1}$; according to the oceanographic data, $5 \times 10^{11} \text{ g sec}^{-1}$.

The boundary between the California Current and countercurrent represents a trough in the sea surface (inset, fig. 6) which corresponds to the axis of the northerly winds. Off the coast of Mexico, both the trough and wind axis are located further offshore. A second trough associated with an eddy south of Pt. Conception appears to be related to a second maximum of the northerly winds extending from Pt. Conception in a line with San Nicolas Island.

The foregoing application of the meridional solution suffers from the limited extent of the wind system and the lack of adequate meteorologic and oceanographic observations. All that can be said is that the average wind distribution off Monterey is consistent with the existence of a south-flowing current offshore and a countercurrent inshore, and that the computed transports of these currents compare favorably with the measured transports. It should be noted that this current-countercurrent system is a consequence of the local wind stress curl and dynamically altogether different from the current-countercurrent system along the western side of the oceans, which has been interpreted as a boundary phenomenon.

7. **General solution**

Consider now the general field of wind stress

$$
\tau = \tau_x(x, y) + j \tau_y(x, y)
$$

associated with large-scale atmospheric circulation, \(\gamma \ll 1, \delta \ll 1\). Outside the western and eastern boundary regions, (6) then has the integral

$$
\psi_C = -rX_C\beta^{-1} \text{curl}_r \tau,
$$

where \(X_C\) is given by (20C), and

$$
\text{curl}_r \tau = \frac{1}{r - x} \int_x^r \text{curl}_x \tau \, dx
$$

represents a *running* average along a segment of latitude between \(x\) and the eastern boundary. If variations of the wind stress curl with \(x\) in the western and eastern boundary regions are so gradual that \(\psi\) fluctuates essentially with \(X(x)\), then

$$
\psi = -rX_C\beta^{-1} \text{curl}_r \tau
$$

is a suitable approximation for the entire ocean area. The above condition requires \(m \ll k\), which is true by hypothesis. It is in the nature of the (running) average curl of \(\tau\) that it should become the less affected by local fluctuations in the wind field the larger the value of \(r - x\). This must be at least a partial explanation for the known persistence of zonal currents in the western portion of all oceans.

8. **The wind-spun vortex**

For a stationary, anticyclonic, circular wind system

$$
\tau_x = 0, \quad \tau_y = -\Gamma gpe^{-r^2},
$$

\(\Gamma\) and \(p\) are constants.
reaching a maximum stress $\Gamma/\sqrt{2} \xi$ at a distance from
the center $p = \frac{1}{2} \sqrt{2} q^{-1}$ and zero stress at the center
and at great distances, the stream function in the central ocean area (equation 20C, 31, 32) becomes

$$
\psi = -\frac{1}{\beta} \int \left( \nabla \times \tau \right) d\beta
= -\frac{1}{\sqrt{\pi}} \Gamma \beta^{-1} \left[ \left[ 1 - 2(qy)^2 \right] \left[ 1 - I \right] + \frac{1}{2} I' \right],
$$

(34)

where

$$
I(q\rho) = \frac{2}{\sqrt{\pi}} \int e^{-(q\rho)^2} d(q\rho)
$$

is the probability integral and $I'' = d^2 I/d(q\rho)^2$. It is
assumed that the wind system essentially vanishes some distance from shore.

![Fig. 7. Mass transport stream lines $\psi/\Gamma$ (nondimensional)
resulting from the circular wind system shown above.](image)

The mass-transport streamlines (fig. 7) reveal an
anticyclonic vortex centered at $-1/q, 0$, and a trough
with a saddle point at $1/q, 0$. The center of the oceanic vortex is thus displaced west from the center of the
atmospheric vortex by 12 times the radius of the maximum stress circle. The net anticyclonic circulation
in the vortex is $(2.0 + .228)\Gamma/\beta = 5.2 \tau_{max}/\beta$.

Meteorologic and oceanic observations in the eastern Pacific definitely show a westward displacement of the
oceanic anticyclone relative to the Pacific high-pressure area with which it is associated. The magnitudes
of the displacement ($\approx 1000$ km) and the circulation
($\approx 20 \times 10^{12}$ g sec$^{-1}$) are about 60 per cent of what
one might suppose on the basis of the foregoing theory. However, the lack of radial symmetry of the wind sys-
tem because of its poor development in the western portion prevents a more quantitative application of the
theory. The oceanographic observations are inadequate to either confirm or deny the existence of the
saddle point.

It should be noticed that $\psi$ vanishes at $x = +\infty$
but not at $x = -\infty$. This is the result of having
neglected the lateral stress vorticity in dealing with
the first-order equation (28) which admits only one
boundary condition. For large anticyclones ($q/k \ll 1$),
fig. 7 should be a good mid-ocean approximation, and
one may expect a tail of predominately zonal currents
west of the wind system, feeding water into the system
north of the wind axis and expelling it south of the
wind axis (vice versa for cyclones), with relatively
weak compensation currents farther to the north and
south. The effect of lateral stress will be to shorten the
tail in proportion to the compactness of the wind sys-
tem. For a wind system consisting of alternate
cyclonic and anticyclonic cells over the entire globe,
Goldsbrough (1935) finds a current system of similar
cellular structure but displaced westward relative to the
wind cells.

9. Classification of ocean currents

In writing this paper we have felt an increasing need
for a general nomenclature of ocean currents, appli-
cable in all oceans regardless of hemisphere, and sugges-
tive of the meteorologic features to which the cur-
cents are so closely related. A possible system is pro-
posed in fig. 8. The corresponding geographic names
for all oceans are summarized in table 3. The circula-
tion is divided into gyres, which are equivalent in scale
to the climatic belts. Each gyre contains one or
two vortices equivalent in scale to semipermanent
pressure centers. It is suggested that the term eddy
be reserved for random features associated with
turbulence.

In high (north or south) latitudes the cyclonic sub-
polar gyre corresponds to the region of the cyclonic
storms, the anticyclonic subtropical gyre to the anti-
cyclones. The equatorial currents and countercurrent
enclose two additional gyres (see fig. 2) correponding
to the trades and the doldrums, but these are so narrow
that they are best not regarded as gyres. A typical
gyre is composed of zonal currents to the north and
south, a strong persistent current and boundary vortex
on its western side, and a compensating drift in the
central and eastern portion, upon which a vari-
able eastern current and wind-spun vortex are
superimposed.

The foregoing classification has chemical and bio-
logical implications (see charts IV and VI, figs. 217,
216, and 214, Sverdrup et al., 1942). The subtropical
gyres, for example, enclose a relatively warm, haline
body of water, poor in phosphaes, relatively low in biological activity, and blue in color ('blue is the desert color of the sea'). The boundaries of the gyres are marked by remarkably pronounced horizontal gradients in all these variables. The western boundary vortices are biologically unique in that they represent the only regions in the open sea where freely floating organisms in the sunlit zone can be expected to remain for any length of time in a uniform environment. Of such regions, named halostases by biologists, the best known is the Sargasso Sea, named after its concentration of Sargassum, or Gulf Weed, a brown alga characteristic of subtropical regions. The six remaining halostases (table 3) are equivalent dynamically to the Sargasso Sea and it would not be surprising were they also distinguished by concentrations of freely floating organisms with narrow environmental tolerance.

10. Discussion and conclusions

The classification of ocean currents in table 3 is put forth with some hesitation, as it is almost certain to include some unwarranted oversimplification. Additional data on currents and wind stress are likely to lead to revisions. Particularly in the southern oceans the boundary conditions for the subpolar gyre differ radically from those we have considered, and the dynamics of the Antarctic Circumpolar Current is not properly understood. But even so, a preliminary table is helpful in interpreting the main features in the ocean circulation and, what is more important, to point out reasons for essential differences in different oceans.

For example, the Gulf Stream and Brazil Current occupy equivalent positions in the North and South Atlantic, yet the Gulf Stream has about ten times as much transport. According to (26) and table 2 this difference is due chiefly to the larger wind stress curl in the North Atlantic. Differences in wind stress curl can be ascribed to the northward displacement of the climatic equator relative to the geographical equator, which reduces the distance between the trades and the westerlies in the northern hemisphere to about 60 per cent of that in the southern hemisphere. The displacement of the trades in turn seems to be related to the unequal distribution of land and sea in the two hemi-
### Table 3. The general ocean circulation.

<table>
<thead>
<tr>
<th>Classification (see fig. 8)</th>
<th>North Atlantic</th>
<th>South Atlantic</th>
<th>North Pacific</th>
<th>South Pacific</th>
<th>North Indian</th>
<th>South Indian</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Polar Current</strong> (z)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E. Greenland (3)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Western C. (z)</td>
<td>Labrador (6) ?</td>
<td>Falkland</td>
<td>Oyashio</td>
<td></td>
<td>a</td>
<td>a</td>
</tr>
<tr>
<td>W. bdry vortex (z)</td>
<td>p ? (50°N, 50°W)</td>
<td>a ?</td>
<td></td>
<td>a</td>
<td>a</td>
<td>a</td>
</tr>
<tr>
<td>Eastern C. (m)</td>
<td>Norwegian (3)</td>
<td>a</td>
<td>Alaska</td>
<td>a</td>
<td>A</td>
<td>a</td>
</tr>
<tr>
<td>Wind-spun vortex (z, m)</td>
<td>a</td>
<td>a ?</td>
<td>Gulf of Alaska</td>
<td>a</td>
<td>S</td>
<td>a</td>
</tr>
<tr>
<td></td>
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<td></td>
</tr>
<tr>
<td><strong>West Wind Drift</strong> (z)</td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td></td>
<td>N. Atlantic (38)</td>
<td>West wind drift</td>
<td>N. Pacific (20)</td>
<td>West wind drift</td>
<td>West wind drift</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Gulf Stream (70)</td>
<td></td>
<td>Kuroshio Ext. (65)</td>
<td></td>
<td>A</td>
<td></td>
</tr>
<tr>
<td>Western C. (z)</td>
<td>Brazil (7)</td>
<td></td>
<td>Kuroshio (23)</td>
<td>E. Australia ?</td>
<td>Agulhas (25)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Florida (26)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>W. bdry vortex (z)</td>
<td></td>
<td>Sargasso Sea</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eastern C. (m)</td>
<td>Canary ?</td>
<td>Benguela (16)</td>
<td>California (15)</td>
<td>Peru (13)</td>
<td>p (v)</td>
<td></td>
</tr>
<tr>
<td>Wind-spun vortex (z, m)</td>
<td>Azores vortex</td>
<td>p (30°S, 5°W)</td>
<td>E. Pac. vortex</td>
<td>p (30°S, 100°W)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

a: absent; p: present, but unnamed; m: related to meridional winds; S: system; v: variable; w: weak; z: related to zonal winds.
Figures in parentheses give mass transport in $10^4$ g sec$^{-1}$ estimated from oceanographic observations.
spheres. One is tempted, then, to trace the preponderance of the great Gulf Stream and Kuroshio systems over their southern counterpart to the excess of water-covered area in the southern hemisphere.

Differences in circulation between the North Atlantic and North Pacific can be at least partly ascribed to the greater intensity and northward displacement of the Atlantic zonal winds, especially during winter (fig. 9). The circulation in the subpolar gyre of the North Atlantic is complicated by two related factors: both the northern land boundaries and the axis of the westerly winds are further north in the eastern than the central and western portion. This may be the reason why the western current of the subpolar gyre is broken up into two parts (Sverdrup et al., 1942, pp. 651-666): the East-Greenland current flowing at an average longitude of 30°W from 80°N to 60°N; and the Labrador current at 60°W, flowing from 60°N to 40°N.

The wind stress pattern over the subpolar gyre is largely a reflection of the traveling cyclones, and one may expect a certain adjustment of the ocean circulation to the frequency and paths of these cyclones. The extent of such an adjustment depends on the time constants of the ocean circulation. Observational evidence indicates that the adjustment to individual storms is not a major one, yet that the main currents do reveal seasonal fluctuations. The time constants appear, therefore, to be of the order of weeks, perhaps months. Thus ocean currents might represent a convenient method of obtaining time averages of atmospheric circulation. For example, the intensity of the Gulf Stream (perhaps estimated from tide records) could be related to the mean wind vorticity in the region of subtropical anticyclones over the Atlantic.

The foregoing statement is somewhat at variance with the usual assertion that the Gulf Stream is caused by the trades. Rather it is related to the vector difference between the trades and the westerlies, or, strictly speaking, to the mean value of the wind stress curl at the latitude in question. It is at least conceivable that an increase in the velocity of the trades be confined to a narrow band of latitude, and that the wind stress curl and the Gulf Stream at say, 30°N remain unchanged. This is, however, unlikely for physical reasons as the mechanism responsible for the transfer of momentum from the westerlies to the trades would tend to distribute such an increase over the entire subtropical zone.

The validity of our statement that the intensity of the western current be proportional to the wind stress curl at the latitude in question depends on the approximations involved in neglecting the inertial terms in (1). Rossby (1936) has developed a wake-stream theory of the Gulf Stream in which the inertial terms play a predominant part. The theory has many attractive features. It explains the increase downstream in width and mass transport, the existence of an observed countercurrent on the western edge of the Gulf Stream, and a jet-like current profile which appears to be in accord with recent observations (Iselin and Fuglister, 1948).

In order to estimate the magnitude of the inertial terms it is necessary to specify the distribution of density and velocity with depth. Using Reid's exponential model for the density (23) and a similar law for the velocity:

$$v = v_0 \text{ for } -z \leq h', \quad v = v_0 e^{z/z_0} \text{ for } -z \geq h',$$

one obtains the expression

$$\frac{3}{8 \rho \beta h} \nabla \times (\mathbf{M} \cdot \nabla) \mathbf{M},$$

(35)

to be added to the left side of the differential equation of mass transport (6). The ratio

$$\frac{\sqrt{3}}{4} \frac{r_k^2 n \sigma_{\max}}{\rho \beta^2 h} e^{-z_0} \cos \left( \frac{\sqrt{3}}{2} k x + \frac{\pi}{6} \right),$$

(36)

of (35) to the planetary vorticity is small except along the inshore edge of the western current, where it may reach unity. It seems likely that some of our conclusions will be modified in the sense prescribed by Rossby.

This investigation has served to emphasize the fundamental importance of the wind stress curl rather than of the wind stress vector in determining the transport of ocean currents in meridionally bounded oceans. For example, the circulation shown in figs. 2 and 6 is in no ways modified by shifting the zero axis of the
wind stress curve. It so happens that in the subtropical zone the inflection point of the zonal wind stress curve falls close to where the wind stress changes sign, and it has been tacitly assumed in most textbooks that the change from a westward to an eastward current is related to a similar change in wind direction. This general misconception probably explains the difficulty in interpreting instances where the current runs against the wind, such as the California Countercurrent, the Point Barrow Current, and most important of all, the Equatorial Countercurrent. In the case of an irrotational wind field, such as a uniform west wind, over an ocean bounded on its western and eastern sides, it follows from (4) and (22) that \( \phi = 0, \nabla P = \tau, \) i.e., that there is no circulation and that the stress is balanced by the pressure distribution resulting from the piling up of water against the lee shore. With the important exception of the Antarctic Circumpolar Current, which has no major meridional boundaries, it can be stated that permanent ocean currents are related to the rotational component of the wind stress field over the ocean. These ocean currents would vanish were the wind stress field irrotational.

Finally we may examine the question of why the computed transports of the Gulf Stream and the Kuroshio current amount to only about one half the observed values (table 2). It does not seem reasonable that the oceanographic observations should be off by more than, say, 20 per cent, nor that the theoretical expression (26) for the transport should account for the discrepancy, since it is almost independent of the eddy viscosity and the shape of the ocean basin.

Maury and others have ascribed the North Atlantic circulation, in particular the Gulf Stream, to differential heating between equator and pole, to the freezing of ice, and to other processes that make up the thermohaline circulation. If we assume that the circulation were half wind-driven, half thermohaline, it would be a strange coincidence that the general pattern of the circulation, such as the boundaries of the gyres, should conform so closely to the general atmospheric circulation. Furthermore, Fuglister\(^9\) has found a high correlation between variations in the current with variations in the wind. It should also be noted that the thermohaline circulation insofar as it is related to the outflow of river water along the Atlantic seaboard would tend to reduce rather than to strengthen the Gulf Stream. It would seem therefore that the subtropical gyre, and probably also the subpolar gyre, are predominantly wind-driven.

Methods for computing wind stress from the observed wind speeds according to the equation

\[
\tau = C_{Dair} U^2
\]


are discussed by Reid (1948a) and in SIO Oceanographic Report 14.\(^10\) Underestimates of \( \tau \) may first of all result from underestimates in the wind speeds on the climatological charts from which the appropriate averages were taken. The preponderance of coastal stations, and the tendency of ships at sea to avoid regions of high wind, would lead to consistent errors, but these cannot account for more than a fraction of the discrepancies. The weakest link is the drag coefficient \( C_D \) which is based on measurements of Baltic storm tides, and a few other measurements (Sverdrup et al., 1942, pp. 489–500). In accordance with the views presently accepted we have assumed \( C_D = 0.026 \) at high wind speeds, \( C_D = 0.008 \) at low speeds, with the discontinuity occurring at Beaufort 4 (Munk, 1947).

In the trade-wind belt of the eastern Pacific, where the winds are predominantly Beaufort 4 and above, Sverdrup (1947) and Reid (1948a) have obtained satisfactory agreement between computed and observed transports. Supposing a value of .0026 were applicable at all wind speeds, then the effect of the southwesterly winds over the eastern Atlantic would make itself felt in a much higher meridional wind factor (table 2), and it can be demonstrated that the discrepancy between computed and observed transports could be largely accounted for. We are therefore led to propose a higher value of \( C_D \) at low wind speeds.\(^11\) The transports of the Gulf Stream and Kuroshio current are, after all, probably as good an indicator of the overall stress exerted by the winds on the ocean as any of the measurements on which the value of \( C_D \) is now based.

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REFERENCES


\(^{11}\) Neumann (1948) claims on the basis of his analysis of Baltic and North Sea tills that the drag coefficient actually increases with decreasing wind speed. His average value for \( C_D \) pertaining to Beaufort 3 is in excess of .0040.


