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WIND MIXING CURRENTS

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ABSTRACT

A wind system may create an ocean current by differential mixing in a two layer ocean; such a current may be imposed on other currents due to the wind. In one situation studied, such differential mixing produced an average transport of water about 10 to 20% of the transport due to wind stress.

WIND ACTION ON THE OCEAN

Sverdrup, et al. (1942: 489–503) have summarized the processes through which wind causes currents in the ocean:

a) currents directly driven by wind stress;
b) mass transport of water by wind waves;
c) currents caused by redistribution of density by wind transport. In addition, wind also causes mixing between the cold thermocline waters and the warm surface layer of the ocean. The following discussion shows that, through such mixing action, the wind may cause an additional current (in an ocean with a stable density stratification).

1 Contribution No. 35, Department of Oceanography, Agricultural and Mechanical College of Texas.
2 Reference to this process was suggested by R. S. Arthur.
Figure 1. The warming at 44°N 41°W during the period 18 September to 28 September 1949 occurred when the winds were Beaufort Force 4. Between 28 September and 8 October the column cooled by a large amount and the winds were Beaufort Force 6½.
It is assumed that a strong wind mixes more cold water upward into the warm mixed layer than does a weak wind. Observations tend to verify this. Consider the ship station at 44° N and 41° W for the period 18 September to 8 October 1949 as shown in Fig. 1. From 18 to 28 September the average of the Beaufort wind forces was 4. The depth of the then-mixed layer remained almost constant and the temperature of the surface layer increased. These occurrences suggest little or no mixing across the thermocline. In contrast, from 28 September to 8 October the average of the Beaufort wind forces was $5\frac{1}{2}$. During this period the mixed layer cooled almost 5° F and deepened about 110 feet, suggesting the upward mixing of a large amount of cold water into the surface layer from below the thermocline.

Consider that a strong wind with associated mixing across the thermocline persists in one section of the ocean while a weaker wind which causes little or no mixing occurs nearby. The colder water introduced into the surface layer under the stronger wind will alter the density distribution in the surface layer. In response to this new density distribution, a current will occur within the upper ocean layers.

EVALUATING THE WIND MIXING CURRENT

Assume that the ocean is a two-layer system which has initially constant densities in the upper and lower layers and no currents. Strong winds over a portion of the ocean will then cause mixing across the thermocline (boundary) so that a region of relatively colder water will be formed in the surface layer of the ocean. Neglecting the other wind effects, such a horizontal variation in the density distribution of the ocean surface layer will lead to a current. It will be assumed that, after transient effects have disappeared, such a current will be in geostrophic balance so that

$$u = -\frac{1}{\rho' f} \frac{\partial p}{\partial y},$$

$$v = \frac{1}{\rho' f} \frac{\partial p}{\partial x},$$

where $u$ and $v$ are the ocean velocity components in the $x$- and $y$-directions respectively, $\rho'$ the ocean density at a given point $(x,y)$ in the surface layer, $f$ the Coriolis parameter (vertical component of the earth's vorticity), and $p$ the pressure.

The hydrostatic equation is

$$p(z) = p_0 + g\rho'z,$$
Figure 2. Variations in the mixing cause horizontal variations in the density which lead to a current that increases with depth. The current at the bottom of the mixed layer must balance the slope of the thermocline if there is no current in the lower layer.
Figure 3. An adaptation of an illustration in Sverdrup "The Oceans" showing a current constant in the horizontal and with depth in a layer of constant density. (Between A and C)
where \( p_0 \) is atmospheric pressure at the ocean surface (assumed constant here), \( g \) gravity, and \( z \) the depth (measured positive downward). We use this equation to obtain:

\[
\begin{align*}
  \frac{u}{\rho' f} & = - \frac{gz}{\rho' g} \frac{\partial \rho'}{\partial y}, \\
  \frac{v}{\rho' f} & = \frac{gz}{\rho' g} \frac{\partial \rho'}{\partial x}.
\end{align*}
\]

This is a measure of the wind mixing current at any depth \( z \). The mixing process has not affected any pressure below the surface layer. Hence there is no current below the density discontinuity. Margules formulae\(^4\) for the slope of density discontinuities on a rotating earth, together with (4) and (5), give the relationships

\[
\begin{align*}
  u (H) & = - \frac{gH}{\rho' f} \frac{\partial \rho'}{\partial y} = - \gamma \frac{\partial H}{f \partial y}, \\
  v (H) & = \frac{gH}{\rho' f} \frac{\partial \rho'}{\partial x} = \gamma \frac{\partial H}{f \partial x},
\end{align*}
\]

where \( \gamma = g \frac{\rho - \rho'}{\rho'} \) and \( H (x, y) \) is the depth of the interface. These expressions assert that the current in the mixed water at the interface is balanced geostrophically by the slope of the interface. We can show that this slope, required for geostrophic balance, is a natural result of the mixing process.

For the mixing process assumed here, \( \rho' \) changes such that

\[
\Delta \rho' = \frac{\Delta H}{H} (\rho - \rho').
\]

Equation (8) tells us that, since space changes exist because of mixing,

\[
\frac{\partial \rho'}{\partial x} = (\rho - \rho') \frac{\partial H}{H} \frac{\partial x}{\partial x}.
\]

\(^1\) By using the depth of water from a horizontal plane as a measure of the pressure, we are assuming that the mixing process does not affect the height of the horizontal surface. Dr. R. S. Arthur has pointed out that 100 m of water at salinity 35 °/oo, with the upper 50 m at 20° C and the lower 50 m at 10° C, mixes to a length 100 m less 1 cm, so that the mixing process does not leave the surface horizontal. Mr. R. O. Reid has kindly computed the current resulting from the slope and has found that it is 10% or less of the corresponding wind mixing current.

\(^4\) See any standard reference in meteorology.

\(^5\) \( \Delta \) is any small change or the differential.
This is essentially the same as equation (7), which is an independent relation between two expressions for the velocity. Thus the slope of the interface that results from mixing balances the geostrophic current created by mixing.
Some of the features of such a wind mixing current are shown in Fig. 2. It is assumed that a steady wind which is uniform over the region is blowing over the right-hand portion, while a calm exists over the left portion. These conditions have prevailed for some time so that transient effects are no longer present. The density is constant in the vertical above and below the thermocline transition zone, although it varies in the x-direction. A geostrophic current would be directed into the figure and would vary from zero at the top surface to a maximum at the deepest part of the transition zone. Fig. 3 shows the pressure and density distribution for a geostrophically balanced current in a layer of constant density. This can be compared with Fig. 2 to show how a wind mixing current differs from familiar types of currents that must occur in the vicinity of sloping thermoclines. Fig. 2 of this paper differs from one by Sverdrup, et al. (1942: 44, fig. 106) in that Fig. 2 shows a horizontal variation in density above the transition zone, while no such density transition appears in Fig. 3, which is adapted from Sverdrup, et al.

**ORDER OF MAGNITUDE OF THE WIND MIXING CURRENT**

Since the wind mixing current varies linearly with depth, in the model above, the transport of water is

\[
T_y = \frac{1}{2} g H^2 \frac{\partial \rho'}{\partial y} = \frac{1}{2} H v(H), \quad (10)
\]

\[
T_x = -\frac{1}{2} g H^2 \frac{\partial \rho'}{\partial x} = \frac{1}{2} H u(H). \quad (11)
\]

The data in Figs. 1 and 4 show that there was more mixing across the thermocline at Station "D" than at Station "C" (Lat. 53N Long. 35W) during the period 28 September to 8 October 1949. The average wind speed at Station "C" was force 4 for the period. If only one-half of the variation at "D" was due to mixing, then the average water transport due to mixing current between "D" and "C" was 0.4 ft²/sec. If all of the winds between "C" and "D" had been in the same direction, the transport due to wind stress during this period could have been about 27 ft²/sec (Sverdrup, et al., 1942: 498). Hence it would appear that the wind mixing current is small compared to possible wind stress currents. However, taking into account the direction of the wind, the resultant stress transport for this period was 2 to 5 ft²/sec. Thus the wind mixing current was 10 to 20% of the net transport by the wind stress for this 10-day period.

*This is probably a maximum estimate, since each wind is assumed to give the maximum possible transport.*
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