INTENSITY OF THE VERTICAL TRANSFER OF WATERS IN THE OCEAN

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Abstract

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K. A. CHEKOTILLO

INTENSITY OF THE VERTICAL TRANSFER OF WATERS IN THE OCEAN

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In connection with the need to solve the problem of the possibility of burying products of radioactive decay in the depths of the ocean, interest has recently increased considerably in the vertical circulation within the mass of ocean waters. In the present work an attempt is made to give a quantitative estimate of the intensity of vertical water transfer between the sea surface and the near-bottom layers in a real ocean basin.

Let us consider a quasi-stationary circulation in an ocean of considerable depth. P. S. Lineikin ([1]) showed that, in the case when the vertical density gradients in the upper layer of the sea are small and when the main attention is devoted to motion at great depths, the effect of inertia may be neglected in the balance of forces acting on a water particle. Therefore, to describe currents in the ocean we adopt the following system of equations:

$$f\rho v + \frac{\partial}{\partial z}T_{zx} - \frac{\partial p}{\partial x} + A_l \Delta u = 0; \tag{1}$$

$$f\rho u - \frac{\partial}{\partial z}T_{zy} + \frac{\partial p}{\partial y} - A_l \Delta v = 0; \tag{2}$$

$$g\rho = \frac{\partial p}{\partial z}; \tag{3}$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}, \tag{4}$$

where the axes of the rectangular coordinate system x, y, z are directed respectively eastward, northward, and vertically downward; $f = 2\omega \sin\varphi$ is the Coriolis parameter; φ is geographic latitude; T_{zx}, T_{zy} are the components of the vertical shear stress in the directions x and y, respectively; A_l is the dynamic coefficient of horizontal turbulent exchange of momentum; Δ is the Laplace operator. The remaining symbols are generally accepted.

Let us estimate the order of magnitude of the terms in equations (1), (2). Since the importance of the first three terms of the equations is not in doubt, we shall estimate the order of magnitude of the last terms. Following approximately the method indicated by L. M. Fomin ([3]), we write for the x-axis

$$A_l \Delta u \approx \frac{A_l}{l^2} u,$$

where l is the horizontal scale of the quasi-stationary oceanic circulation. Denoting A_l/l^2 by f' and assuming that the coefficient of horizontal exchange is described by the formula

$$A_l = k l^{4/3},$$

we obtain:

$$A_l \Delta u \approx f' u = 2.15 \cdot 10^{-7} u \quad \text{(CGS)},$$

where the value of l has been taken as the minimum for the conditions of established oceanic–

currents (100 km) and where the value of k, according to R. V. Ozmidov ([2]), is taken to be 0.01 CGS.

It is useful to compare the last terms of (1) and (2) with the first terms. Since it is natural to assume that the order of magnitude of the components u and v is one and the same, and that $\rho = 1$, we have

$$\frac{A_l \Delta u}{f \rho v} = \frac{f'}{f}.$$

For $1^\circ > \varphi > 0$, $f'/f = 0.1$.
For $\varphi > 8^\circ$, $f'/f < 0.01$.

Thus, everywhere except in a narrow band immediately adjacent to the equator, the influence of horizontal turbulent exchange may be neglected. Taking this circumstance into account, we solve the system (1)–(4). The boundary conditions are as follows:

$$w_{-\zeta} = u_{-\zeta} \frac{\partial \zeta}{\partial x} + v_{-\zeta} \frac{\partial \zeta}{\partial y},$$

$$T_{z|-\zeta} = T_a, \qquad p_{-\zeta} = p_a,$$

where ζ is the elevation of the disturbed sea surface above the plane of the origin of coordinates; T_a is the tangential wind stress at the sea surface; p_a is

Fig. 1

Figure 1: Fig. 1

the atmospheric pressure. Using the well-known relations obtained by V. Ekman for drift current, after certain simplifications we find for the vertical component of velocity:

Fig. 1. Mean multiyear distribution of the time of vertical water transport in the northwestern part of the Pacific Ocean in August. Regions of water rise are hatched. Time is given in years. The sign ∞ refers to places where the mean vertical velocity is equal to zero

$$w = \frac{1}{\bar{\rho} f} \left[(e^{-\eta z} \cos \eta z - 1) \left(\operatorname{rot} T_a + \frac{\beta}{f} T_{ax} \right) - \right.$$

$$\left. - e^{-\eta z} \sin \eta z \left(\operatorname{div} T_a - \frac{\beta}{f} T_{ay} \right) + \frac{\beta}{f} \left(v_\gamma z - \frac{1}{f} \int_0^z \frac{\partial Q}{\partial x} \, dz \right) \right], \tag{5}$$

where $\beta = \partial f / \partial y$, v_γ is the meridional component of the gradient-convection current at the sea surface; $\eta = \pi / D$ (D is the depth of friction); $\bar{\rho}$ is the mean value of density in the interval $(-\zeta, z)$; Q is the so-called dynamic depth.

The time of water transport through the ocean thickness can now be found from the formula:

$$t = \frac{H - h_H}{\bar{w}}, \tag{6}$$

where $\bar{w}$ is the mean value of the vertical velocity within the limits from the sea surface to the bottom friction layer; h_H is the thickness of the bottom friction layer, amounting to several tens of meters; H is the depth of the sea. Since the ocean depth, which we took from the bathymetric chart, is determined with low accuracy, it is inadvisable in the preceding relation to introduce a correction for the thickness of the bottom friction layer.

Using formulas (5) and (6), the mean multiyear values of the time of vertical transport of water particles through the entire thickness of the ocean were calculated. As the value-

As the object for the calculations, the northwestern part of the Pacific Ocean was chosen; computations were carried out for August (Fig. 1). To determine the quantities T_a and D entering into (5), the wind was used, calculated from the atmospheric-pressure field [(5)]. The transition from wind speed to its tangential stress was made by the well-known formula of G. Neumann. The values of v_x were found on the basis of data on total currents given in [(4)]. The speed of the drift current at the sea surface, which is needed to determine D and v_y, was calculated by the well-tested formula:

$$(U)_0 = \frac{1.27 \cdot 10^{-2} V_0}{\sqrt{\sin \varphi}},$$

where V_0 is the wind speed at the sea surface. To determine the quantities Q, data on the density field were taken from the "Catalogue of Deep-Water Observations of the Pacific Ocean," compiled at the Institute of Oceanology of the Academy of Sciences of the USSR under the direction of V. A. Burkov.

As follows from Fig. 1, the characteristic value of the time of water transport from the sea surface to the bottom in the ocean ranges from 1 year to 10 years. Regions where the transport-time values exceed 10 years occupy comparatively small areas. It cannot, of course, be asserted that the calculated values of the transport time are entirely reliable. There is no doubt, however, that the orders of magnitude of the sought quantities have been determined correctly.

Institute of Oceanology
Academy of Sciences of the USSR

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CITED LITERATURE

[1] P. S. Lineikin, *Fundamental Problems of the Dynamic Theory of the Baroclinic Layer of the Sea*, L., 1957. [2] R. V. Ozmidov, DAN, **126**, No. 1 (1959). [3] L. M. Fomin, Tr. Institute of Oceanology, **66** (1963). [4] *Atlas of Surface Currents Northwestern Pacific Ocean*, Hydrographic Office US Navy, Washington, 1950. [5] *US Navy Marine Climatic Atlas of the World*, **2**, North Pacific Ocean, Direction of the Chief of Naval Operations, Washington, 1956.

Note: Figure translations are in progress. See original paper for figures.