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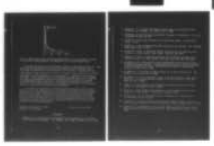

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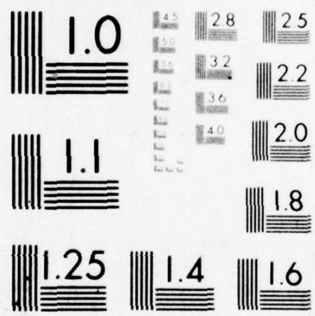

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(Nekotoryye dannyye o krupnomasshtabnykh hkharakteristikakh poly gorisontal'nykh komponent skorosti v okeane)

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SOME DATA ON LARGE-SCALE FIELD CHARACTERISTICS OF HORIZONTAL VELOCITY COMPONENTS IN THE OCEAN

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*A series of statistical field characteristics of horizontal velocity /1708
components in the ocean are calculated by using the method of analytical filtering of the investigated functions in different filter passbands. It it shown that the spectral energy density function of velocity fluctuations is in good agreement with the 5/3 power law up to 24-h fluctuation periods. The values of a series of terms in the equation of motion of ocean waters are also estimated.

At the present time, one of the important problems of naval hydrodynamics is the determination of the role of turbulent friction in liquid motion in the coastal and open parts of the ocean. This problem may be solved in two different ways. The first consists in the direct calculation of the derivatives of turbulent stresses $-\overline{\rho u_i' u_j'}$ in the Reynolds equations, which describe the turbulent flow studied:

$$\rho\left(\frac{\partial \bar{u}_i}{\partial t}+\bar{u}_j\frac{\partial \bar{u}_i}{\partial x_j}\right)=F_i-\frac{\partial \bar{p}}{\partial x_i}+\frac{\partial(-\overline{\rho u_i' u_j'})}{\partial x_j},\quad \frac{\partial \bar{u}_j}{\partial x_j}=0, \tag{1}$$

where $i, j = 1, 2, 3$ (the repeating index denotes summation); x_i are Cartesian coordinates; $\bar{u}_i$ are the averaged velocity components; u_i' are the fluctuational or turbulent velocity components, equal to the deviations of the corresponding instantaneous (nonaveraged) velocity components from their averaged values; F_i are the components of external forces; p is the pressure; ρ is the density of the liquid; and t is the time. The upper bar in the Reynolds equations denotes a given type of averaging (for example, time averaging) with a definite scale (period). Terms describing the action of molecular viscosity forces were omitted from Eqs. (1), since they are negligibly small.

The second method of studying the role of turbulent friction in the sea involves the introduction of certain hypotheses relating turbulent stresses in the Reynolds equations to the field of averaged velocities in the flow. The various coefficients thus introduced cannot be determined theoretically, but can be estimated experimentally in the turbulent current studied. However, in most cases these coefficients are not universal constants, but are themselves substantially dependent on the characteristics of the flow velocity field and many other factors. In addition, the hypotheses themselves are frequently inadequately substantiated and their applicability may be called into question. Therefore, the first method of studying turbulent friction in the sea, not involving the introduction of any hypotheses or coefficients, but based on a direct calculation of the corresponding terms in the /1709
Reynolds equations, appears to be preferable.

*Numbers in the right margin indicate pagination in the original text.

A major objective in this case is not only to determine the values of turbulent stresses and their derivatives, but, most importantly, to study their dependence on the averaging scale T_0, used in resolving the velocity field into the averaged (regular) and fluctuational (turbulent) components. The special importance of this question for sea conditions is due to the presence in sea turbulence of eddies of the most diverse dimensions; this must inevitably result in a substantial dependence of the averaged quantities and fluctuation velocities (and hence, Reynolds stresses) on the scale T_0 used in a given problem.

To calculate the forces of turbulent friction from experimental data, it is necessary to have a long series of observations of current velocities at a series of ocean points sufficiently close to each other. Unfortunately, no such data are available to the investigators at the present time. However, a sufficient number of day-to-day recordings of horizontal velocity components [illegible] = u and u_2 = v at individual points of the ocean are now available, which make it possible to calculate the Reynolds stresses $-\rho\overline{u'v'} = -\rho\overline{v'u'}$, $-\rho\overline{u'^2}$, $-\rho\overline{v'^2}$ and to study the dependence of these stresses on the averaging period T_0.

At the present time, most recordings of current velocities in the sea are obtained by means of BPV-2 type printing meters. By virtue of their structural characteristics, BPV-2 current meters placed in the sea record instantaneous directions of the horizontal current velocity vector at 10-30 min intervals, and moduli of this vector averaged over [illegible] min. It is clear that such instruments are not suitable for studying the "fine" structure of the velocity field and high-frequency fluctuations in the spectrum of sea turbulence. However, the extremely large horizontal dimensions of the seas and oceans cause the appreciable large-scale character of horizontal processes. Therefore, in determining the large-scale characteristics of the field of horizontal velocity components in the ocean, a long series of observations (on the order of many days) is required, and short-period fluctuations, which make a negligible contribution to the total energy of horizontal turbulence, may be neglected altogether. Therefore, the selection of BPV-2 should be considered fully adequate for the study of large fluctuations of the velocity field in the ocean, which contain a significant fraction of the energy of ocean turbulence.

Major disadvantages of many velocity recordings by BPV-2 instruments are the appreciable errors introduced into the instrument readings by their mobility on buoy-type anchor devices used in the measurement of sea currents. However, as evidenced by experience with such buoy devices, appreciable errors are observed only in the readings of instruments placed in the upper sea levels (approximately down to 100 m). Therefore, we selected for our calculations the data of an instrument which operated at a depth of 200 m, where the errors introduced by the motions of the buoy on the wave may be considered absent.

Data of hydrological current meters for studying large-scale horizontal turbulence in the sea were used for the first time by Shtokman back in 1939-40. In 1941, Shtokman[1] calculated the Reynolds stresses $-\rho\overline{u'v'}$, $-\rho\overline{u'^2}$ and $-\rho\overline{v'^2}$ for sea turbulence. This was done by using data of day-to-day current measurements with the Ekman-Mertz current meter in the Caspian Sea. The measurements were made every 2 hours and lasted 23 days. The fluctuation velocities u' and v' were determined by Shtokman by subtracting the velocities $\overline{u}$ and $\overline{v}$, averaged over the entire observation period, from the "instantaneous" velocity values recorded by the current meter. Later, similar studies in the Black Sea were made by Khlopov[2] and Gezentsvey.[3] /17

The latter, in calculating $\overline{u'^2}$ and $\overline{v'^2}$, successively used as the "instantaneous" velocity values those recorded by the current meter and those averaged over periods of 1 h and 6 h. The fluctuation velocities were obtained by subtracting the values of the velocity $\bar{u}$ and $\bar{v}$, averaged over the entire observation period (14 days), from these "instantaneous" values. This method enabled Gezentsvey to reveal the dependence of $\overline{u'^2}$ and $\overline{v'^2}$ on the averaging period of "instantaneous" velocities. However, the constancy of the basic averaging scale T_0, used in the calculation of $\bar{u}$ and $\bar{v}$, did not permit her to relate $\overline{u'^2}$ and $\overline{v'^2}$ to this scale. As was noted above, it is precisely this dependence which should be of interest to an investigator.

The Reynolds stresses for sea turbulence for a constant averaging scale T_0 were also calculated by Stommel[4] from velocity measurements in the Straits and by Ichiye,[5] who used for this purpose photographs of the velocity field made with the aid of an electromagnetic current meter (EMCM) in the region of the Kuroshio.

The quick-response equipment built in the last few years has made it possible to determine Reynolds stresses caused by high-frequency fluctuations of the velocity field in the ocean. Studies have been made along these lines by Bowden and Fairbairn,[6,] Kolesnikov et al.,[8-10] Grant, Stewart and Moilliet,[11,12] and Bowden and Howe.[13] In these authors' experiments, the duration of the current velocity recordings usually did not exceed 5 to 10 min, and the fluctuation velocities were calculated relative to the velocity components averaged over such a time interval. The majority of these studies also showed that for such scales of the phenomenon, the statistical characteristics of sea turbulence obey the relationships established in the theory of locally isotropic turbulence by A. N. Kolmogorov.

In studying the dependence of turbulent Reynolds stresses in the sea on the averaging scale T_0, it is natural to proceed as follows. Let the recording of any given component of the current velocity at some point of the sea be described by the function u(t), which may be considered continuous and differentiable. Such a function may be represented by a superposition of simple harmonic oscillations with different amplitudes and periods T_n (or angular frequencies $\omega_0 = 2\pi/T_n$). If by virtue of certain consideration we choose the time scale of the averaging equal to T_0, we thereby refer all components with periods greater than T_0 to "regular" changes in averaged velocity $\bar{u}(t)$, and higher-frequency velocity changes should in this case be treated as turbulent fluctuations u'(t).

In order to separate, in a given curve, the components with periods greater than T_0 from other oscillations, it is obviously necessary to "pass" the function u(t) through a certain type of filter. In the ideal case, low-frequency oscillations (with $\omega < \omega_0 = 2\pi/T_0$) should pass through such a filter without any attenuation (the amplitude ratio of the transmitted to the arriving signal A/A_0 should be equal to unity), and the amplitudes of harmonics with frequencies greater than ω_0 should fall to zero. The frequency characteristic of such an "ideal" filter is shown in Fig. 1 by a broken line. The function u(t) can be filtered either by means of electronic devices (the function in this case is given by a suitable electric signal), or by analytical means. Analytically, the filtering (or smoothing) operation may be represented as follows: /171

$$\bar{u}(t) = \int_{-\infty}^{\infty} u(t-\tau) K(\tau) d\tau, \qquad (2)$$

where τ is the integration variable, and the function $K(\tau)$ is called the smoothing core.

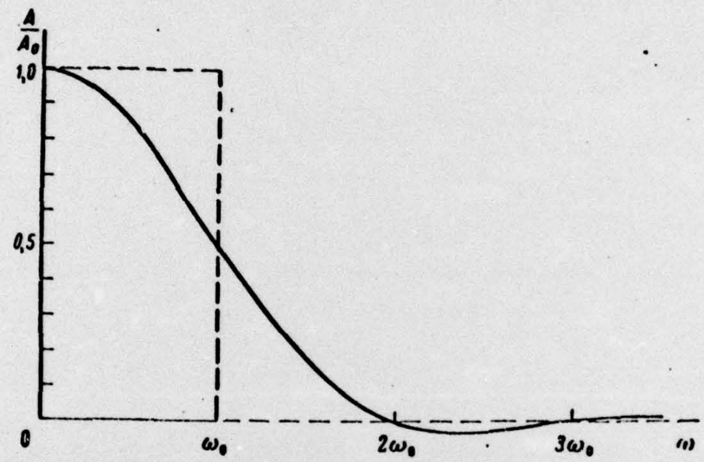

Fig. 1. Spectral characteristics of smoothing cores (3) (broken line) and (4) (solid line).

For an "ideal" smoothing with a rectangular frequency (spectral) characteristic, shown in Fig. 1 by a broken line, the function $K(\tau)$ is

$$K(\tau) = \frac{\sin \omega_0 \tau}{\pi \tau}, \qquad (3)$$

where $\omega_0 = 2\pi/T_0$.

Indeed, if such a core is made to act on a simple harmonic oscillation $\cos \omega t$, it is readily seen that this oscillation remains unchanged when $\omega < \omega_0$ and will be completely "extinguished" (the integral in Eq. (2) will be zero) when $\omega > \omega_0$. When $\omega = \omega_0$, the integral in Eq. (2) will take on the value of 1/2.

In a practical application of the smoothing operation to experimental data, the use of an ideal filter is still difficult in the majority of cases. Indeed, function (3) damps out relatively weakly with increasing τ, and therefore the integration in formula (2) must be performed over a very large interval of change in τ. Therefore, instead of core (3), it is necessary to use other functions different from zero only on a finite segment of the τ axis. In this case, the spectral smoothing characteristic deteriorates to some extent, and becomes different from the "ideal" one. On the advice of Professor W. H. Munk (USA), we used a smoothing core of the fol- /1
lowing form, very convenient for practical applications:

$$K(\tau) = \begin{cases} \dfrac{1 + \cos \omega_0 \tau}{T_0} & -\dfrac{\pi}{\omega_0} \leqslant \tau \leqslant \dfrac{\pi}{\omega_0}, \\ 0 & |\tau| > \dfrac{\pi}{\omega_0}. \end{cases} \qquad (4)$$

The spectral characteristic of the smoothing core 4 is shown by a solid line in Fig. 1. It is obvious that this characteristic differs appreciably from the "ideal" one. When core 4 is used, components with periods greater than T_0 are also partly attenuated, and higher-frequency components do not damp out completely. However, as is evident from the graph, oscillations with frequencies greater than $2\omega_0$ will be practically absent in the smoothed function, since the characteristic curve for $\omega > 2\omega_0$, oscillating about the axis of abscissas, approaches it rapidly with increasing ω, and even the first maximum has an absolute value of less than 0.03. At the point $\omega = \omega_0$, the spectral characteristic of core 4, like that of an "ideal" core, has a value of 1/2.

As already noted, the initial data used in our calculations were the velocity recordings made by the BPV-2 instrument on a day-to-day buoy station operated in June-August 1958 in the Atlantic Ocean at a point with coordinates 55°15' N and 16°30' W and a depth of 3080 m. A current recorder registered the velocity vector modulus and its direction every 30 min. These data were used to calculate the horizontal velocity components u(t) and v(t), which were then subjected to a smoothing operation. The smoothing was carried out with core 4 using values of the parameter T_0 equal to 3, 6, 12 and 24 h. The smoothing technique is very simple. First, for a chosen parameter T_0, the values of function (4) are determined for each instant τ_n of velocity recording; these values are then multiplied by the corresponding values of $u(t_0 - \tau_n)$, and the products are summed. The sum thus calculated gives the value of $\overline{u}(t_0)$ for a specified value of time t_0. Then, the next instant $t_0 + \Delta t$, where Δt = 30 min, is taken as the initial point, and the value of $\overline{u}(t_0 + \Delta t)$ is similarly obtained. Continuing this operation, we eventually obtain a series of discrete values of the function $\overline{u}(t)$, averaged with a given smoothing period T_0. The values thus obtained for smoothed functions of horizontal velocity components are given in Figs. 2 and 3. In each case, the calculation was made for 200 consecutive instants t_n, i.e., the interval of change in argument t in these functions is 100 h.

The graphs of Figs. 2 and 3 clearly show that increasing the averaging period causes fluctuations of increasingly larger scale to disappear from the curves, and the curves themselves become increasingly "smoother."

Having found the values of averaged functions $\overline{u}(t)$ and $\overline{v}(t)$, one can also easily obtain the values of the corresponding fluctuation velocities u'(t) and v'(t). It is sufficient for this purpose to calculate the values of these averaged functions from the corresponding "instantaneous" values of u(t) and v(t). Squaring u'(t) and v'(t) (or multiplying them out) and averaging, we obtain the values of Reynolds stresses (we neglect the minus sign, and ρ = 1) for a given averaging scale T_0. The values of u'^2, v'^2 and $\overline{u'v'}$ calculated in this manner for T_0 = 3, 6, 12 and 24 h are given in the table. It is obvious from the data of this table that, as expected, a regular increase in the turbulent stresses u'^2 and $\overline{v'^2}$ is observed with increasing /17
averaging period T_0. At the same time, $\overline{u'v'}$ remains very small for all T_0. This fact, along with the approximate equality of $\overline{u'^2}$ and $\overline{v'^2}$ (when T_0 = 3, 6 and 12 h), attests to the isotropy of turbulent fluctuations of these dimensions in the investigated turbulent current in the ocean. This is also indicated by the tabulated data on the correlation factor between u' and v'. Indeed, this factor may be /171
assumed equal to zero within the accuracy of the calculation. The isotropy of

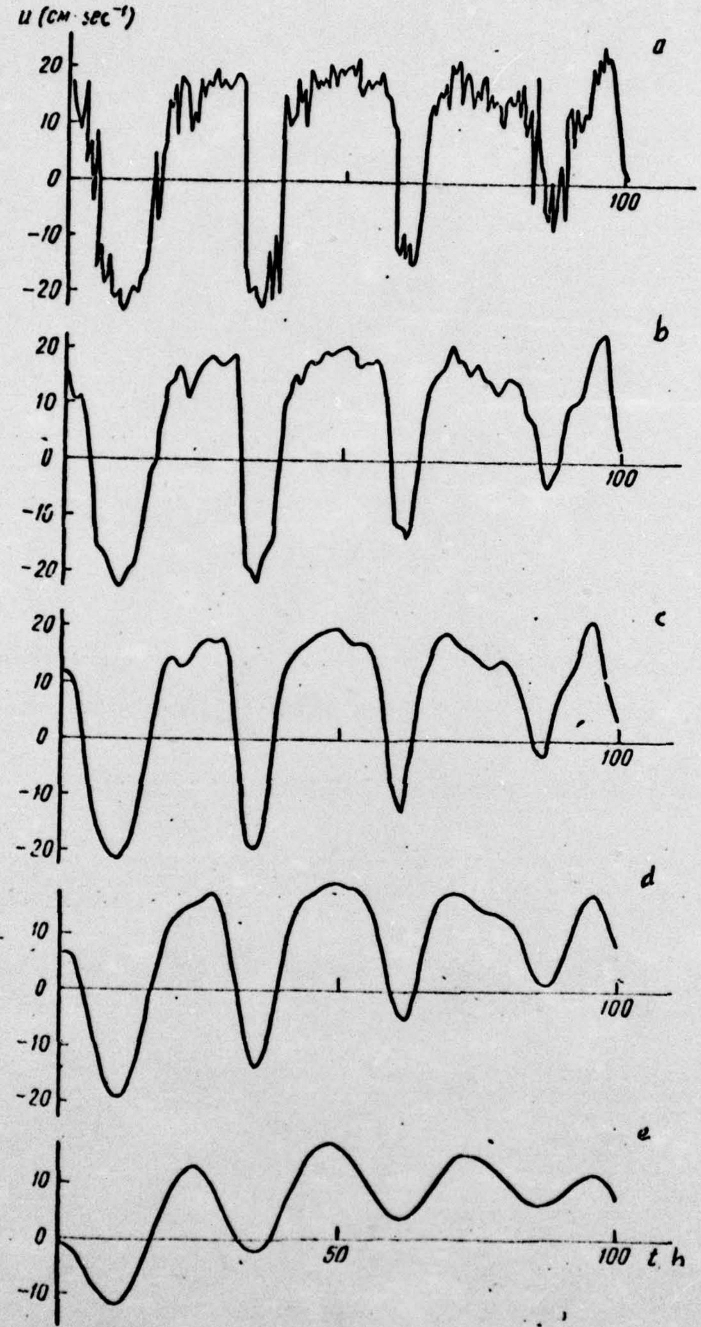

Fig. 2. The function u(t) for various averaging periods:
a - "instantaneous" values of the function u(t);
b - with averaging periods of 3 h;
c - 6 h; d - 12 h; e - 24 h.

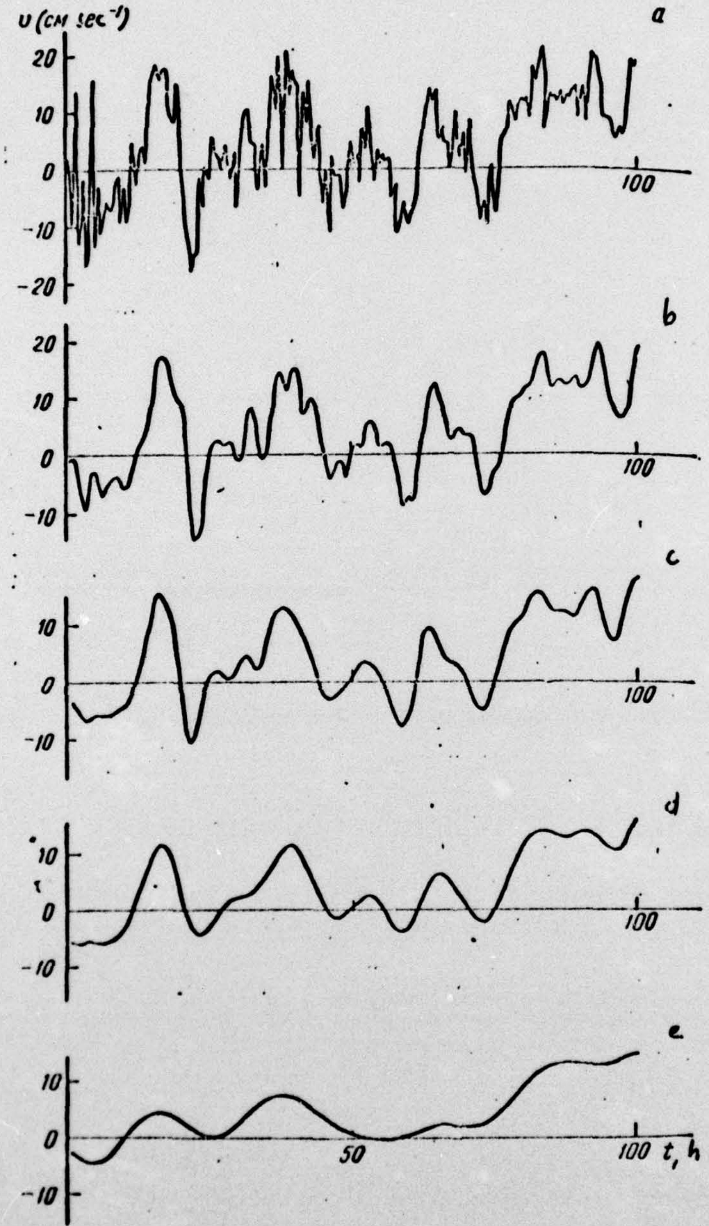

Fig. 3. The function v(t) for different averaging periods:
a - "instantaneous" values of the function v(t);
b - with averaging periods of 3 h;
c - 6 h; d - 12 h; e - 24 h.

Quantity	Averaging scale					
	T_0, h	0,5	3	6	12	24
	ω, 10^{-4} sec^{-1}	34,89	5,82	2,91	1,45	0,73
$\overline{u'^2}$	cm^{2}sec^{-2}	—	9,8	16,3	32,0	83,3
$\overline{v'^2}$	cm^{2}sec^{-2}	—	13,3	19,1	28,7	45,6
$\overline{u'v'}$	cm^{2}sec^{-2}	—	—0,9	0,1	0,8	3,7
$R=\dfrac{\overline{u'v'}}{\sqrt{\overline{u'^2}}\sqrt{\overline{v'^2}}}$	—	—	—0,08	0,01	0,03	0,06
E	cm^{2}sec^{-2}	—	23,1	35,4	60,7	128,9
$\sqrt{\overline{\bar{u}^2}}$	cm sec^{-1}	14,7	13,9	13,2	11,5	9,1
$\sqrt{\overline{\bar{v}^2}}$	cm sec^{-1}	8,2	7,4	7,0	6,1	5,2
$\sqrt{\overline{\left(\frac{\partial\bar{u}}{\partial t}\right)^2}}$	10^{-4} cm sec^{-2}	22,2	8,9	7,8	6,1	3,3
$\sqrt{\overline{\left(\frac{\partial\bar{v}}{\partial t}\right)^2}}$	10^{-4} cm sec^{-2}	26,7	8,9	5,6	3,3	1,1

turbulent fluctuations breaks down to some extent only in the region of the 24-h averaging scale, where the values of $\overline{u'^2}$ and $\overline{v'^2}$ differ appreciably. This phenomenon is apparently due to a strongly manifested singularity in the region of the velocity fluctuation spectrum, i.e., ordered tidal oscillations of the diurnal period, which are distinctly different in the graphs of Figs. 2 and 3.

Data on the magnitude of $\overline{u'^2}$ and $\overline{v'^2}$ also make it possible to analyze the behavior of the sum of these quantities with increasing T_0. As can be readily seen, this sum is merely the total energy E (per 1/2 unit mass) of horizontal turbulent fluctuations with periods shorter than T_0. Increasing the averaging period from T_0' to T_0'' should give an addition to the turbulent energy, due to velocity fluctuations with periods located in the interval $\Delta T = T_0'' - T_0'$.* If the smoothing is performed with a whole set of smoothing parameters T_0, one can then, generally speaking, plot a graph of the function E(T) (or E(ω)). In other words, this method can be used to study the energy spectrum of large-scale turbulent fluctuations of the velocity field in the ocean. If however the increment of E(ω) due to turbulent fluctuations in the frequency interval Δω is divided by Δω, we obtain still another important characteristic of turbulence - the spectral energy density $f(\omega) = \Delta E(\omega)/\Delta\omega$. Obviously, in

*It should be emphasized that the statements made here will be strictly valid provided an "ideal" smoothing of the functions u(t) and v(t) is used. In our case, they hold only with a certain approximation. In addition, it should be kept in mind that a certain error is also introduced into the values of the function E as a result of the limited duration of the velocity component realizations used in the calculations.

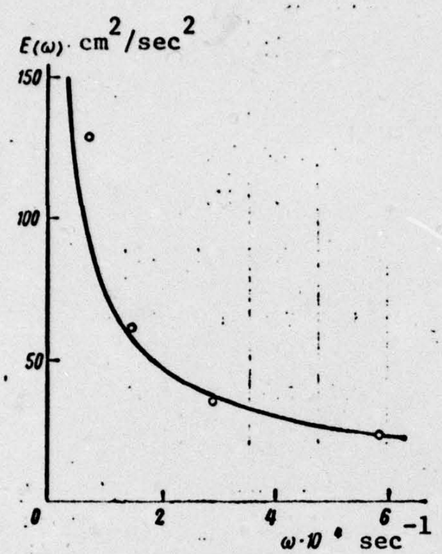

Fig. 4. Experimental values of the function E(ω) and theoretical curve of the 2/3 power law.

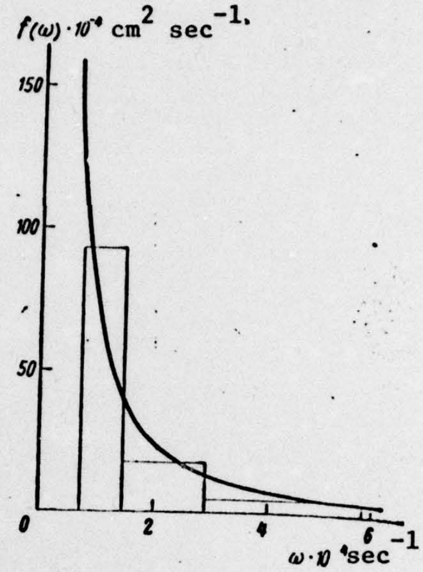

Fig. 5. Experimental values of the function f(ω) and theoretical curve of the 5/3 power law.

our case, only three values of the function f(ω) can be obtained, which were found /171 to be $4.2 \cdot 10^4$, $17.3 \cdot 10^4$, and $93.4 \cdot 10^4$ cm^2 sec^{-1}.* Since the Δω values used for calculating f(ω) are very large, it remains unclear to which ω one should refer the values of the function f(ω) that were obtained in this manner. For this reason, the graph of Fig. 5 does not show points of the function f(ω), but rather, rectangles with bases equal to Δω, and heights corresponding to the values of f(ω) for these intervals of change in the variable ω.

The values of the function E(ω) are shown in Fig. 4. As is evident from Figs. 4 and 5, the experimental values of the functions E(ω) and f(ω) are in good agreement with certain regular curves whose shape will be specified below.

The theory of locally isotropic turbulence offers expressions for the functions E and f which hold in the so-called inertial subinterval of turbulence scales. These expressions are

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*It should be noted that the functions E and f can obviously also be obtained in the usual manner, i.e., by a Fourier transformation of the correlation functions for the velocity components in the flow in question. However, our method of smoothing the functions u(t) and v(t) with different averaging periods T_0 has the advantage that it makes it possible to avoid nonstationary long-period fluctuations in the velocity field and in addition, along with the values of the functions L and [symbol illegible], one can also obtain data on the derivatives $\partial\bar{u}/\partial t$ and $\partial\bar{v}/\partial t$ for different T_0, so that one can in turn compare the orders of the terms in the Reynolds equations, as will be discussed in detail below.

$$E(k) = c_1 \varepsilon^{2/3} k^{-5/3}, \quad f(k) = c_2 \varepsilon^{2/3} k^{-5/3}, \tag{5}$$

where k is the wavenumber; ε is the dissipation rate of turbulent energy per unit mass, and c_1 and c_2 are dimensionless universal constants.

The observations whose data are used in this study were performed at a fixed point of space, and therefore in order to be able to apply formulas (5) to these /171 data, it is necessary somehow to replace the space variables in Eq. (5) by time variables. This substitution may be made with the aid of the hypothesis of "frozen turbulence." According to this hypothesis, turbulent eddies with a space wavenumber k during their motion through the observation point give rise at this point to velocity fluctuations with an angular frequency ω related to k by $k = \omega/V$, where V is the velocity of "transfer" of turbulent eddies by the mean flow. The "frozen turbulence" hypothesis is accurate in the case of turbulent flows in which the energy of turbulent fluctuations is much lower than the energy of mean motion. In our case, this condition is not fulfilled very well. Indeed, as can be easily seen from the tabulated data, the rms values of the flow velocities at high values of T_0 approach the mean flow velocities (in absolute value). This fact, along with the marked variability of the mean velocity with time, makes it necessary to apply the "frozen turbulence" hypothesis only very carefully to the data under consideration. However, we will use this hypothesis anyway, if only to get an idea of the possible form of the dependence of E and f on the frequency ω. Formulas (5) are thus transformed to

$$E(\omega) = a_1 \omega^{-5/3}, \quad f(\omega) = a_2 \omega^{-5/3}, \tag{6}$$

where

$$a_1 = c_1 \varepsilon^{2/3} V^{5/3}. \quad a_2 = c_2 \varepsilon^{2/3} V^{2/3}.$$

The graphs of functions (6) are shown in Figs. 4 and 5. The values of coefficients a_1 and a_2 were chosen as follows: $a_1 = 0.161$ cm^2 sec$^{-8/3}$, $a_2 = 0.153$ cm^2 sec$^{-8/3}$. As is evident from these graphs, the experimental points are in very good agreement with the theoretical expressions (6), with the exception of the values for $T_0 = 24$ h ($\omega = 0.73 \cdot 10^{-4}$ sec^{-1}), which are located slightly above the theoretical curves. This may again be explained by the presence, in this frequency region, of tidal currents giving a definite "spike" in the spectral functions E(ω) and f(ω). Obviously, one might suspect that such good quantitative agreement between experiment and theory is due to an accidental favorable combination of errors introduced by the perturbations and inaccuracies listed above. However, one can apparently state with a fair amount of confidence that the experimental data confirm the applicability of the expressions of the theory of locally isotropic turbulence to the investigated scale interval of horizontal turbulent formations in the sea. This conclusion should not be particularly surprising if one recalls that the horizontal dimensions of the ocean are of the order of thousands of kilometers, whereas the spatial dimensions of the investigated phenomena (calculated by multiplying the T_0 values by the corresponding mean current velocity) are only 1-10 km. For phenomena with such spatial dimensions, the hypothesis of local isotropy had also been confirmed previously by checking the applicability of the 4/3 power law to the coefficient of horizontal turbulent diffusion in the ocean.[14]

The above values of the coefficients a_1 and a_2 in formula (6) make it possible to estimate the order of magnitude of the turbulent energy dissipation rate in the sea, ε. Considering that the velocity V is of the order of 10 cm/sec, and the constants c_1 and c_2 are taken to be equal to unity, the order of magnitude of ε is found to be 10^{-2} cm^2 sec^{-3}, which is in good agreement with previous estimates of ε in the sea.[15]

The last two rows of the table give the rms values of the time derivatives of /17 of the velocity components $\bar{u}(t)$ and $\bar{v}(t)$ averaged with a different scale. These derivatives were calculated by dividing the different between consecutive values of $\bar{u}(t)$ and $\bar{v}(t)$ by the time interval separating these values. It follows from the table that the rms values of $\partial\bar{u}/\partial t$ and $\partial\bar{v}/\partial t$ are substantially dependent on the averaging scale, decreasing with T_0. This result illustrates once again how carefully one must make the estimate of the order of the terms in the equations of motion of seawaters. In making this estimate, it is always necessary to specify exactly the scale of the process to be described by a given system of equations of motion.

An analytical expression for the dependence of the rms value of $\partial\bar{u}/\partial t$ (or $\partial\bar{v}/\partial t$) on the averaging scale T_0 may be obtained from the following considerations. Obviously, $\partial\bar{u}/\partial t$ is determined mainly by the smallest-scale velocity oscillations existing in the function $\bar{u}(t)$ (i.e., oscillations with period T_0). However, according to the theory of locally isotropic turbulence, the amplitude of these fluctuations $\Delta\bar{u}$ is proportional to T_0 to the power 1/3. And since the characteristic time Δt for this fluctuation is again equal to T_0, we obtain for the rms value of the derivative $\partial\bar{u}/\partial t$

$$\sqrt{\overline{\left(\frac{\partial\bar{u}}{\partial t}\right)^2}} \approx \sqrt{\overline{\left(\frac{\Delta\bar{u}}{\Delta t}\right)^2}} = \gamma\frac{T^{1/3}}{T} = \gamma T^{-2/3}, \tag{7}$$

where γ is a proportionality coefficient.

Expression (7) may be obtained from the relation between the rms value of the derivative $\partial\bar{u}/\partial t$ averaged over the interval T and the structural function of the velocity field $D_{uu}(\tau)$. Indeed, we have

$$\sqrt{\overline{\left(\frac{\partial\bar{u}}{\partial t}\right)^2}} \approx \sqrt{\overline{\left(\frac{u(T)-u(0)}{T}\right)^2}} = \frac{\sqrt{D_{uu}(T)}}{T}.$$

If however we now use for the function $D_{uu}(T)$ the 2/3 power law known in the theory of locally isotropic turbulence, we immediately arrive at Eq. (7), obtained above.

A graph of the power law (7) with proportionality coefficient γ = 0.59 cm $sec^{-4/3}$ is given in Fig. 6. It is evident from this figure that the experimental points are again in good agreement with the theoretical curve, once more confirming the applicability of the theory of locally isotropic turbulence to horizontal processes of a given scale in the ocean.

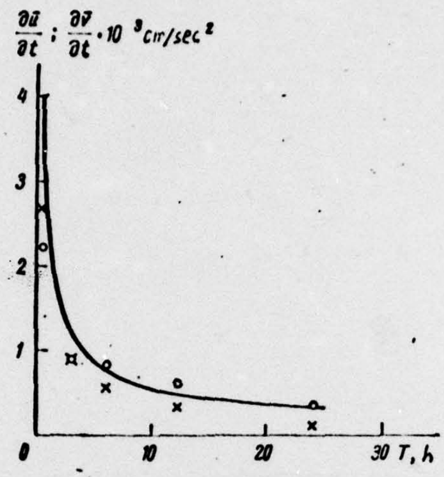

Fig. 6. Experimental values of the derivatives $\partial\bar{u}/\partial t$ (circles) and $\partial\bar{v}/\partial t$ (crosses) for different averaging periods, and theoretical curve of the 2/3 power law.

The values obtained for the nonstationary terms in the Reynolds equations of motion may be compared with the values of the components of the Coriolis force, which plays an important part in the dynamics of sea currents. Calculation of the components of Coriolis accelerations in our case leads to values ranging from $1.8 \cdot 10^{-3}$ to $0.6 \cdot 10^{-3}$ cm sec^{-2} (depending on the mean velocity). Comparison of these figures with values of nonstationary terms in the Reynolds equations leads to the conclusion that these terms surpass the components of the Coriolis force only for averaging scales (phenomena) shorter than 2-8 h. For larger-scale processes, terms with the Coriolis force become predominant in the equations of motion, and such processes may in a certain sense be considered quasi-stationary. /171

In conclusion, it should be noted that the numerical estimates given in this study for large-scale characteristics of the field of horizontal velocity components in the ocean are obviously typical only of the dynamic conditions existing during the observation period. Under different conditions, however, such estimates may of course change markedly. However, it may be stated with sufficient reason that the general character of the dependence of the investigated characteristics on the averaging scale T_0 is fairly universal and determined solely by the general characteristics of turbulent liquid motion.

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