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Some Speculations on Oceanic Diffusion Diagrams

Akira Okubo

113 Macaulay Hall

The Johns Hopkins University

Baltimore, Maryland 21218 USA

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Abstract

Oceanic diffusion diagrams are examined on the basis of the similarity theory of turbulence.

Some problems remain to be answered only by future studies.

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Introduction

Because of the complexity involved in the process of oceanic diffusion, hardly a single theory can explain or interpret the entire pattern of diffusion. Progress in oceanic diffusion still depends heavily on an experimental approach carried out in the field. Many efforts have been made in experimental studies on oceanic diffusion especially since Pritchard and Carpenter (1960) developed a new field technique of continuous measurement of the fluorescent concentration (Joseph, Sendner and Weidemann, 1964; Carter and Okubo, 1965; Karabashev and Ozmidov, 1965; Foxworthy, Tibby and Barsom, 1966; Ito, Fukuda and Tanigawa, 1966; van Dam and Davids, 1966; Han and Yoon, 1970; Talbot, 1970; Kullenberg, 1969, 1971, 1972, and others).

Summarizing available data of patch diffusion from field experiments Okubo (1971) constructed two kinds of diffusion diagrams; one showing horizontal variance (σ_{rc}^2) versus diffusion time (t) and the other showing apparent diffusivity (K_a) versus the scale of diffusion (ℓ). Here we reproduce the results in Figs. 1 and 2, where the "old data" represent diffusion data prior to 1961 and the "new data" represent those after 1961. Schuert's data are derived from his subsurface (300 m depth) experiment in the sea off Hawaii (Schuert, 1970). One concludes that, irrespective of the various macroscopic oceanographical conditions, the variance increases with time at a power between 2 and 3 and, equivalently the apparent diffusivity increases with the scale of diffusion, indicating non-Fickian diffusion. Simply because many of the diffusion data provide information only on the equivalent radius, the horizontal variance given in Fig. 1 is calculated after a symmetrical treatment of the original data (Joseph and Sendner, 1958). A more appropriate measure of variance would be defined on the actual distribution, usually elongated. However, the general feature of the oceanic diffusion shown in Figs. 1 and 2 would remain essentially unchanged as suggested by Okubo (1968) and Kullenberg (1972).

Observed results of oceanic diffusion have often been compared with predictions of the similarity theory of turbulence, particularly for the inertial subrange characterized by the rate of energy dissipation, ϵ , (Kolmogorov, 1941). In the present paper the author first follows this line of thought to interpret the oceanic diffusion diagrams and then brings up some problems to stimulate further studies in the matter.

Interpretation based on the similarity theory of turbulence

The oceanic diffusion of a patch of substance is a problem in relative diffusion by turbulence (Batchelor, 1950). With the aid of dimensional arguments, one finds a regime of relative diffusion where the initial size of the patch is infinitesimal. This regime is characterized by

$$\sigma_{rc}^2 = c_1 \epsilon t^3 \quad (1)$$

where c_1 is a numerical constant. From (1) one can derive the relation between¹ the apparent diffusivity and the scale of diffusion (the 4/3 power law or Richardson's law of diffusion):

$$K_a = c_2 \epsilon^{1/3} \ell^{4/3} \quad (2)$$

where c_2 is the numerical constant.

The basic concept involved in the laws (1) and (2) is that the eddies responsible for the horizontal spread of substance are locally isotropic and homogeneous, lying in the "inertial subrange." The properties of these eddies thus depend only on the rate of energy dissipation which must balance the rate of energy transfer through the range if a statistical equilibrium is established.¹ Note that neither homogeneity nor isotropy holds for oceanic turbulence over the scales in which we are interested. At best the isotropy may be only in a horizontal plane.

In Figs. 1 and 2 we attempt to fit the third power law to the variance and the $4/3$ power law to the diffusivity. The results from the similarity theory of turbulence appears to apply to the diagrams locally for some time and length scales. All that is required is that ϵ varies with the time and length scales with which we are concerned.

One may notice a clear jump in line fitting which occurs at a time of a half day to a day or at a length scale of 1 km or so. Interestingly enough, this gap almost coincides in its time and length scales with a gap in the energy spectrum of oceanic turbulence proposed by Ozmidov (1965).

Ozmidov suggested that the main energy input to oceanic turbulence was concentrated in certain restricted spectral bands and that the turbulence spectrum between these bands should obey the $5/3$ power law but with different values of ϵ . Since the energy supplied tends to be transferred from larger eddies to smaller eddies, the value of ϵ increases as we go from a region of the $5/3$ law to the adjacent region lying to the smaller-scale side of the spectrum. Fig. 3 shows schematically the Ozmidov spectrum where the energy input occurs at three main bands at a length scale of 10 m associated with wind waves, at 10 km associated with inertial and tidal fluctuations and at 1,000 km associated with the atmospheric pressure system for general circulation. Only the middle band is relevant to the discussion of our diffusion diagrams. Fig. 4 shows the $4/3$ power law fitted locally to two regions of the diagram (for details, see Okubo and Ozmidov, 1970)

From Fig. 1 we can compute the value of $c_1 \epsilon$ for the two regions separated by a time scale of a half day or so. The result gives $c_1 \epsilon = 2.5 \cdot 10^{-5} \text{ cm}^2 \text{ sec}^{-3}$ for the time range from 1 hr to a half day and $c_1 \epsilon = 5.4 \cdot 10^{-6} \text{ cm}^2 \text{ sec}^{-3}$ for the time range from a half day to several days. Similar estimates could be made from Figs. 2 and 4 for the length-scale range. Unfortunately nothing certain can be assigned to the numerical value of c_1 . We shall make a "retrospective" approach to determine the proper value of c_1 .

According to Ozmidov (1965), a good estimate for ϵ is $0 (10^{-3} \text{ cm}^2 \text{ sec}^{-3})$ in a length-scale range of 50 m to 5 km which nearly corresponds to a diffusion-time range of 1 hr to 1/2 day. Hence an appropriate value of c_1 would be 0.01 as far as the order of magnitude is concerned. We then estimate ϵ as $2.5 \cdot 10^{-3} \text{ cm}^2 \text{ sec}^{-3}$ in the time range from 1 hr to 1/2 day and $5.4 \cdot 10^{-4} \text{ cm}^2 \text{ sec}^{-3}$ in the range from 1/2 day to several days. The difference, $2.0 \cdot 10^{-3} \text{ cm}^2 \text{ sec}^{-3}$, between two values of ϵ would be accounted for by the rate of energy flux due to local external forces if the Ozmidov spectrum be accepted.

The fact that an apparent gap exists at 1/2 day or so leads to speculation that the external forces may be tidal. According to Munk and MacDonald

¹In the inertial subrange the similarity theory predicts that the turbulence energy spectrum $E(\kappa)$ is given by $E(\kappa) = A \epsilon^{2/3} \kappa^{-5/3}$, where κ is the wave number and A is a universal constant (- $5/3$ power law).

(1960), Jeffreys obtained $1.1 \cdot 10^{19}$ erg sec⁻¹ as the estimate of tidal dissipation in shallow water of all oceans. Assuming that this amount of dissipation is representative of the dissipation over continental shelves occupying 5.5 per cent of the earth's surface or $2.8 \cdot 10^{17}$ cm², we obtain the average rate of dissipation of tidal energy per unit mass of water to be $4.0 \cdot 10^{-3}$ cm² sec⁻³. The rough agreement between the two figures, i.e., $2.5 \cdot 10^{-3}$ versus $4.0 \cdot 10^{-3}$ might be purely accidental, however.

If this concept of a localized subrange is correct, similar behavior may also be found in data concerning the horizontal dispersion of discrete particles, e.g. surface floats, which react only to horizontal components of oceanic turbulence. However, the dispersion diagram for particles of this sort, prepared by Ichiye and Olson (1960) does not reveal the feature similar to the diffusion diagram, though the data for large-scale dispersion obtained by them from drift bottle studies appear to be of poor quality (Ichiye and Olson's diagram is reproduced in Fig. 5).

In order to see the difference between the patch diffusion and float dispersion, we plot in Fig. 6 the apparent horizontal diffusivity versus the scale of diffusion for two types of diffusion. To convert the float dispersion diagram (Fig. 5) to the $K_a - \ell$ diagram, we use the relations that $F = 3.03 K_a$ (Stommel, 1949) and $(\ell + \ell_o)/2 = \sigma_{rc} = 1/3 \ell$. A 4/3 power law is fitted to roughly indicate the upper boundary with surface floats and the lower boundary with large-scale patches. Note that data points for small-scale patches in the upper mixed layer lie in the vicinity of the upper boundary line, while those for small-scale patches in the coastal thermocline (Kullenberg, 1969) lie in the vicinity of the lower boundary line.

Yet another interpretation of the diagrams is possible on the basis of the similarity theory. All the releases of substance in the diffusion experiments in question were made in the vicinity of the sea surface. A dye patch diffuses not only horizontally but also vertically. As a result, the center of mass of the dye patch tends to locate in the deeper layer as time goes on. From velocity measurements by moored current meters, Webster (1969) found that ϵ decreased monotonically with depth. Typical values were $3 \cdot 10^{-3}$ cm² sec⁻³ at 10 m and $6 \cdot 10^{-4}$ cm² sec⁻³ at 100 m. On the other hand, from hot-film anemometer records, Stewart and Grant (1962) determined ϵ in the surface layer of the sea. They too noticed a decrease in ϵ with depth between 2 and 15 m. At a depth of 2 m, the value of ϵ ranged from $5.2 \cdot 10^{-3}$ to $4.5 \cdot 10^{-2}$ cm² sec⁻³ and at a depth of 15 m, ϵ was $1.1 \cdot 10^{-3}$ cm² sec⁻³. Fig 7 summarizes some estimates of ϵ in the sea. This may be supplemented by a similar figure presented by Faller (1971).

These all indicate that ϵ decreases with depth. This, combined with the fact that the center of mass of the patch lowers with time, permits one to speculate that the effective value of ϵ of the patch decreases as time increases or equivalently as the scale of diffusion increases. This would result in the overall growth rate of the variance with time being slower than the third power and in the overall growth rate of apparent diffusivity with scale being smaller than the 4/3 power. As a matter of fact, a simple model assuming $\epsilon \propto z^{-1}$, z being the depth of water predicts that $\sigma_{rc}^2 \propto t^{2.5}$ and $K_a \propto \ell^{1.2}$. If this alternative speculation is correct, the above-mentioned difficulty in interpreting the difference between the diffusion

diagram of substance patch and that of surface floats may be resolved.

Diffusion in subsurface water

We have some evidence that the diffusion rate in deep water is in fact as low as that indicated by the small value of ϵ . Schuert (1970) has succeeded in a field experiment in measuring the turbulent diffusion of dye at a depth of 300 meters. He reveals that the values of the characteristic parameters of diffusion are smaller than those obtained from surface experiments. Data points calculated from Schuert's results are included in Figs. 1, 2, and 6. A typical value of ϵ reads 10^{-3} , 10^{-4} and $10^{-5}(\text{cm}^2 \text{ sec}^{-3})$ for the upper layer, 300 m depth, and 1,000 m depth, respectively (Fig. 7). Hence, from (1) the relative horizontal size, i.e., the standard deviation, of substance patch at the same diffusion time is estimated as σ_{rc} (upper layer): σ_{rc} (300 m depth): σ_{rc} (1,000 m depth) = 1 : 0.32 : 0.10. According to Fig. 1, the horizontal size of substance patch in the upper mixed layer reaches typically 2 km after 2 days. At the same diffusion time, the horizontal size of substance patch at 300 m is estimated to be 640 m. In Schuert's experiment at 300 m depth the standard deviation of the dye patch was 550 m after 2 days. The agreement is fairly good. If we extend this concept to the diffusion at 1,000 m depth, say, the characteristic size of dye patch would be only 200 m after 2 days. Recently T.E. Ewart, University of Washington, has attempted a deep water dye study at 1,000 m depth. A preliminary calculation suggests that the diffusion rate at 1,000 m depth might be smaller than that of Schuert's experiment at 300 m (Ewart, 1972).

Unlike the upper mixed layer, the turbulence structure in the deep waters may be "patchy." In particular the deep water is divided into a series of relatively uniform "layers" separated by thin interfacial regions or "sheets." The formation of this step-like structure remains unexplained. Generally speaking the sheets are stable, but internal waves travel on the sheets. The Kelvin-Helmholtz type instability can occur on the sheets associated with the shear from the internal waves. The instability grows to breakers which rapidly degenerate into a patch of turbulence (Woods and Wiley, 1972). This billow turbulence should primarily control the mixing of deep water. Recently Garrett and Munk (1972) have proposed an interesting mechanism of mixing due to internal wave shearing associated with the step-like structure. According to them, the instability on the sheets followed by the generation of turbulent patches can primarily control the mixing of deep water both in horizontal and in vertical directions. Since the mixing of this sort decreases with Brunt-Väisälä frequency, both horizontal and vertical diffusivities decrease with increasing depth. Garrett and Munk's theory could connect the mixing mechanism with the dissipation rate of turbulence. The work is still premature, but it is worthy to note that the dissipation rate does decrease with depth.

The mechanism of diffusion in the deep water may differ from that of the upper mixed layer. We are not even certain that the conventional theory of diffusion can be applied to deep water diffusion. For this to be achieved, we urgently need the simultaneous observations of dye diffusion and of the step-like structure in deep waters.

A problem similar to deep water diffusion should arise in mixing in stratified shallow waters. The step-like structure dominates also in stratified shallow wa-

Kullenberg's measurements of diffusion in a thermocline of coastal waters (Kullenberg, 1969, 1971) showed that the rate of diffusion was smaller by an order of magnitude than that of the upper mixed layer. On our diffusion diagrams his data points lie distinctly below those of the mixed layer (see Fig. 6).

Other possibilities

It is quite a surprise to us that the laws of diffusion derived from locally isotropic turbulence could be applied, possibly with a certain modification, to the results over a very wide range of scales, say from 10 m to 1,000 km. Even more surprising is the fact that the existence of the inertial subrange in the energy spectrum has not been verified for such a wide range of scales.

In this context one should notice other diffusion theories that deduce the same laws as predicted by the similarity theory of turbulence. Imahori and Hori (1951) applied a Fokker-Planck equation to turbulent diffusion as an extension of Brownian motion. A similar approach was used by Obukov (1959). Later Lin (1960a, b) developed a dynamical model of diffusion on the basis of a Langevin equation analogous to the Brownian motion theory. These models are characterized by a parameter of the assumed random-acceleration process and result in the diffusion laws (1) and (2) without direct appeal to Kolmogorov's theory. Lin's theory does not require the isotropy assumption. Krasnoff and Peskin (1971) examined Lin's theory to arrive at a conclusion that the restrictions on Lin's theory are not as severe as those of the similarity theory although some assumptions similar to the Kolmogorov theory are hidden in Lin's theory. However, the Langevin model will not reach its full potential until one accurately relates it to the Lagrangian form of the Navier-Stokes equation.

There are yet other possibilities. Even the dependence of our interpretation on the similarity theory of turbulence might not be appropriate if some other processes operate on the diffusing patch in the sea. It has become apparent that shear in the mean flow can play an important role in oceanic mixing (Bowden, 1970). The basic result of current shear is that the combination of the gradient of mean velocity and turbulent mixing in the same direction produces effective longitudinal diffusion. Thus, a uniform vertical shear in the horizontal current results in the longitudinal variance proportional to $\Omega K_z t^3$, where Ω denotes the vertical shear and K_z the vertical diffusivity. This implies that the similarity theory of turbulence is *not* the only theory which deduces the third power law of the variance. The shear model of diffusion does not require the similarity hypothesis of turbulence.

Recently some efforts have been made for understanding the mechanisms and processes which give rise to mixing in the upper boundary layers of the sea (Assaf, Gerard and Gordon, 1971; Faller, 1971). All these recognize the importance of Langmuir circulations (Langmuir, 1938) in the mixing. This problem also needs further investigation.

Finally, it is extremely important to study the variation of diffusion characteristics in connection with the environmental factors such as current shear, stratification etc. Kullenberg (1972) has recently made a serious attempt to relate the diffusion characteristics to environmental conditions. Theoretical predictions based on the shear diffusion model are found to compare well with the experimental results. This is quite a significant achievement in our understanding the diffusion process in the sea and is sufficient to encourage us to design similar studies in the future.

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Fig. 1. Variance vs. diffusion time: fit of the t^3 law locally.

Fig. 2. Apparent diffusivity vs. scale of diffusion: fit of the $l^{4/3}$ law locally.

Fig. 3. Ozmidov spectrum for oceanic turbulence.

Fig. 4. Fit of the $l^{4/3}$ power law based on the Ozmidov spectrum.

Fig. 5. Oceanic dispersion diagram after Ichiye and Olson (1960).

Fig. 6. A unified diagram of patch diffusion and float dispersion.

Fig. 7. Depth variation of the rate of turbulent energy dissipation per unit mass of seawater.