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Oceanic Turbulence: Big Bangs or Continuous Creation?

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In a discussion of the turbulence characteristics of patches of 'microstructure' in the ocean, the hypothesis advocated by Gibson (1982), that the patches are produced by very rare but extremely powerful turbulence-generating events which usually have 'fossilized' before their observation, is contrasted with the hypothesis of a turbulence field driven at the time and scale at which it is observed. In this 'continuous creation' notion, by no means original here, the driving energy is converted to turbulence kinetic energy in such a way that the observed overturning thickness scale L_T is linearly related to the length scale $(\varepsilon/N^3)^{1/2}$, where ε is the kinetic energy dissipation rate and N is the buoyancy frequency. (This relationship does not hold in boundary layers, where another length scale, the distance from the boundary, is imposed.) If the time scale of the largest vertical eddies is N^{-1}, the parameters of turbulence and its effects can be estimated by the measurement of N and L_T. For example, the kinetic energy dissipation rate would be proportional to $L_T{}^2N^3$, and the vertical eddy diffusivity would be proportional to $L_T{}^2N$. Careful attention must be paid to the sampling process and its assumptions. 'Fossilized' regions are expected, in the sense that these regions had previously been the sites of turbulence stronger than that present at the time of observation, but consideration of the fossilization process is not necessary for the interpretation of microstructure data.

Gibson [1982], in his discussion of several recent papers concerning oceanic 'microstructure' [*Gregg*, 1980; *Caldwell et al.*, 1980], finds that 'inconsistencies arise when the measured microstructure is interpreted as though it were actively turbulent at the time of observation.' Further, he finds that the 'inconsistencies are resolved and the observations are explained by assuming the microstructure patches are no longer actively turbulent (except possibly at the smallest scales) but are fossil remnants of previous turbulence; that is, fossil turbulence.' Gibson defines fossil turbulence as 'remnant fluctuations in various hydrophysical fields which persist after the fluid has ceased to be turbulent at the scale of the fluctuation.'

The extent to which oceanic microstructure can be interpreted in the terms found appropriate to turbulence in the laboratory has been debated for some years now, with each investigator taking his own point of view. Some have felt that there is not necessarily any connection between the two because of the complication of the oceanic flow by stratification and internal waves and because of the differences between the known energy sources of the laboratory and the unknown, but presumably quite different, turbulence-producing energy sources of the ocean. Others have believed that the laboratory results are directly applicable to oceanic situations. At times a dispute like that over the characterization of microstructure patches as 'active' or 'fossil' turbulence may seem purely semantic. Not so. The implications for the observation and understanding of the basic physics of the oceans are quite different depending on which view is taken. If the patches are produced by active turbulence, with the local production of turbulent kinetic energy and temperature variance nearly equal to the dissipation of these quantities, the sources can be pinpointed by the dissipation rates observed by microstructure instruments. If, on the other hand, these patches are fossil and their observable structure no longer reflects the forcing energy input, the energy must be provided by rare, but extremely powerful, events, and we can only guess at the nature of these events. Without more information we could not estimate average values of fluxes or dissipation rates.

In the following the utility of the fossil metaphor will be discussed, and the inconsistencies that *Gibson* [1982] finds in the interpretations of *Gregg* [1980] and *Caldwell et al.* [1980] will be considered in detail. Then an explication of Gibson's conceptual framework will be given, with the objective of making it a bit more accessible, and the application of his ideas to oceanic data will be discussed. A contrasting view, 'continuous creation,' is then presented with a graphic representation, and some newly available evidence is discussed.

THE FOSSIL METAPHOR

A fossil is defined (*Webster's New Collegiate Dictionary*) as 'any impression, or trace, of an animal or plant of past geological ages which has been preserved in the earth's crust.' Fossil carries with it the connotation of the motionlessness and inertness of the long dead. But here, in the context of oceanic turbulence, the term is connected with a live, active creature: 'fossil turbulence will generally be moving at all scales and may be actively turbulent at scales smaller than the fossil turbulence fluctuation of interest.' Museum attendance would certainly decrease if the dinosaurs were given to moving at any scale. If Gibson's conceptual framework turns out to be useful in discussing oceanic turbulence, let us hope a more helpful metaphor can be found.

THE DISCREPANCY IN THE WORK OF *GREGG* [1980]

In a study of microstructure patches in the main thermocline, *Gregg* [1980] exhibits a few sections of fine resolution temperature profiles which contain patches with inverted temperature gradient (Gregg's Figures 17 (3-m-thick patch), 18 (2.5 m), and 20 (7.5 m)). Gregg interprets them as actively

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turbulent overturning patches. Gibson suggests that the dissipation rate in such patches should be related to the patch thickness L_P as $\varepsilon = 3(L_P{}^2)N^3$. This formula as applied by Gibson yields remarkably large estimates of the dissipation rate. Gibson then asks whether such large dissipations were observed. Unfortunately, the only information in Gregg's paper which might be related to ε is the vertical separation, L_{sep}, between zero crossings of the vertical temperature gradient. By numerical experiment, Gibson relates L_{sep} to the Batchelor scale as $L_{sep} = 9L_B$, where $L_B = (\nu D^2/\varepsilon)^{1/4}$. Taking as an example the thickest overturn, the 7.5-m one in Gregg's Figure 20, Gibson calculates for it an ε of $3 \times 750^2 \times 0.005^3 = 0.2$ cm^2 s^{-3}, on the basis of which he expects L_{sep} to be 0.17 cm. But Gregg found values of L_{sep} no smaller than 1.5 cm, in Gibson's view an inconsistency sufficiently serious to bring the interpretation of the patches as active turbulence into question.

Gibson's choice of the 'overturn' in Figure 20 is unfortunate. On close inspection it appears likely that it does not represent an overturn at all. The *Thorpe* [1977] displacement plot of Figure 20 indicates that most of the water in the middle of the 'overturn' has not been displaced very far vertically and that the overturning involves only an apparent interchange between a parcel at depth 882.5–884.0 m and a few warm spots near 890 m. The sharp edges of the upper parcel give it an appearance typical of horizontal intrusions of water, in this case presumably slightly less saline than the water below it. The billowlike feature at the top of the parcel suggests that shear is associated with it. If the Thorpe plot were recalculated with the assumption that this parcel is an intrusive feature of smaller density than the water below, not participating in the overturning, then the warm spots near 890 m would be seen to interchange only locally, Thorpe displacements of 7.5 m would not be found, and the 'eddy' would be interpreted as a well-mixed region of near-adiabatic temperature gradient with some bits of colder water being entrained from below. In the ocean, Thorpe displacement calculations using temperature alone can be quite deceptive because an intrusion cannot be certainly identified. On many oceanic profiles these calculations must be run using either potential density or both potential temperature and salinity in the reordering scheme.

Gibson's argument should not be dismissed just because of an unfortunate choice of example. In Gregg's Figure 17 a 3-m-thick patch with a more eddylike signature (a complicated structure of apparent interchanges at many scales) on the Thorpe plot can be seen. Whether the temperature gradient is superadiabatic cannot be determined from the plot; an adiabatic gradient would slope only 0.3 mm off vertical in the vertical extent of this plot. Even so this patch might easily be in a state of near overturn: the signature would depend on the precise time the eddy was observed. To perform a calculation similar to Gibson's, we (1) estimate the N appropriate to the stratification against which the eddy is straining, taking 875.5 to 879 m as a generous estimate of the portion of stratification seen by the eddy, (2) calculate the temperature gradient as the difference between the end points divided by their separation, and (3) allow for the (larger) effect of salinity by assuming that the mean TS relation obtains. Then N turns out to be 0.0024 s^{-1}. Recently, *Dillon* [1982*a*] has found that ε is best expressed in these terms as $\varepsilon \sim (L_T)^2N^3$, where L_T is the rms Thorpe scale [*Thorpe*, 1977] discussed later in this paper. (The numerical coefficient is near 1 in the mean for the patches Dillon examined except those close to the surface, but the scatter is very large and the applicability of the form itself should not be assumed until many more situations have been examined.) Using this relation with $L_T \sim 100$ cm and $N = 0.0024$ s^{-1} yields 0.00014 cm^2 s^{-3} for ε and 1.06 cm for L_{sep}. Given the factors that increase Gregg's estimates of L_{sep} (lack of correction for the dynamic response of the thermistor and 1-cm-long cosine-taper smoothing), the discrepancy between Gregg's minimum L_{sep} of 1.5 cm and the calculated value of 1.06 cm is probably just about what would be expected. It is certainly not a discrepancy which should make necessary a major revision in our concept of oceanic turbulence. (Even if we did the calculation precisely in Gibson's manner, no such discrepancy could be found, because of Gregg's treatment of his data. The discrepancy could have been as well refuted by considering the resolution in Gregg's processed data [*Dillon*, 1982*b*].)

The differences between Gibson's calculation of ε and the one above are as follows:

1. A value of N appropriate to the overturning eddy is taken, rather than an oceanic mean (in this case Gibson used the often-quoted oceanic mean value 0.005 for N even though the 400-m mean value given by Gregg is only about half that). The estimate used above is close to the mean from 800 to 1200 m computed by Gregg. The choice of N is very important because it enters as the cube in the formula for ε.
2. The rms Thorpe displacement is used rather than the patch size. In processing oceanic data we find that after one structure has superimposed itself on another and so on, the resulting profile is far more complicated than one simple overturn in a constant gradient. The maximum Thorpe displacement is difficult to define in many instances, and the rms L_T has some statistical robustness to it, so we use the rms L_T.
3. The use of L_T and the dropping of the factor of 3 in Gibson's calculation amount to reducing the constant factor by something like a factor of 10. This change seems justified on the basis of the data available [*Dillon*, 1982*a*]. Whether this sort of calculation has general utility in estimating ε remains to be seen.

THE DISCREPANCY IN THE WORK OF *CALDWELL ET AL.* [1980]

According to Gibson, an 'internal inconsistency exists in the Caldwell et al. proposal that stratified turbulent temperature should be scaled using the dissipation rate $\varepsilon' = 0.42 DCN^2$ . . . if it is assumed that the individual patches are turbulent with dissipation rate ε'.' Gibson points out that '*Caldwell et al.* [1980] explicitly left the physical interpretation of their observed correlation . . . as an open question.' Indeed we did, and in fact we have not since used this expression to compute ε, although our software computes it in the course of data processing. As mentioned above, it does turn out to correlate fairly well with ε, except very near a boundary, for Cox number C less than 7500 ($C = 3Cx$), but probably only because the three most important terms in the turbulence kinetic energy equation are usually of the same order, with production of turbulent kinetic energy being roughly balanced by the rate of dissipation plus the rate of increase of potential energy. In other words, the efficiency of the typical stirring process is neither 100% nor 1%, but rather lies in between. *Oakey*'s [1982, Figure 14] histogram shows this nicely. A more detailed discussion of the observed balance of terms in this equation is given by *Dillon* [1982*b*]. The discrepancy arises, for Gibson, when the value of ε at the transition from laminar flow to turbulence in a stratified fluid, ε_T, is compared with $13DCN^2$. (The 0.42 is dropped now because the mean seems to be nearer 1 than 0.42.) A value for ε_T of $30\nu N^2$ is suggested by Gibson for reasons discussed below. If this value were correct, the Cox number at the transition would be $C_T = \varepsilon_T/(\nu N^2) = (30\nu N^2)/(DN^2) = 30\,Pr$, where Pr is the Prandtl number, ν/D. Gibson finds this inconsistent with the 'usual assumption that temperature microstructure with C values higher than about 10 is actively turbulent.' This is surely not an 'internal' inconsistency; no mention was made by Caldwell et al. of this 'usual assumption,' nor did we contend that the structure observed was isotopic turbulence. The 'usual assumption' may come from *Gregg*'s [1980] conclusion that 'Cox numbers of 10 and more imply the existence of zero crossings in the gradient data,' reached by considering his experience in examining oceanic profiles. The sort of nonoverturning structures he has seen have had Cox number no higher than 10; therefore a larger C implies some overturning. Gregg attached no dynamical significance to this observation.

The Gibson Fossil Turbulence Framework

On what is the relation $\varepsilon_T = 30\nu N^2$ based? From what characteristic of turbulence does Gibson believe a critical value of the Cox number for the laminar-turbulence transition arises? Relations involving a proportionality between ε_T and νN^2 can be obtained in many ways. Gibson gives several:

1. He calculates a gradient Richardson number in terms of ε and sets it equal to 1/4. This argument is unconvincing because it involves setting the mean square shear equal to the square of the mean shear.
2. Measurements by *Stillinger* [1981] in stratified grid turbulence indicate that $\varepsilon = 23\nu N^2$ when the vertical flux, considered here to be an index of turbulence, becomes zero.

A simpler argument seems clearer, if no more quantitative. One essential for fully developed turbulence is a separation of scales between the productive scale, that of the overturning eddy, and the dissipative. If the Kolmogoroff scale L_K is taken as the dissipative scale and $L_R = (\varepsilon/N^3)^{1/2}$ is related to the overturning scale, a little algebra shows that the ratio of overturning scale to dissipation scale can be represented in terms of ε, N, and ν as

$$L_R/L_K = (\varepsilon/\nu N^2)^{3/4}$$

If the condition for transition to turbulence in a stratified fluid is taken as a certain value of this ratio, then the dissipation at transition is

$$\varepsilon_T = (L_R/L_K)_T^{4/3}\nu N^2$$

Experimentation is required to establish a value for $(L_R/L_K)_T$ and to find out if it has a universal value. If, for example, its value were 10, then

$$\varepsilon_T = 21.5\nu N^2$$

The above is typical of many 'derivations' in turbulence. It yields only a plausible hypothesis to test. It does seem likely though that for transition to turbulence an ε of some multiple of νN^2 is required. In Gibson's paper, constants equivalent to $[(L_R/L_K)_T]^{3/4}$ take on several values in the neighborhood of 10–13.

Is a certain value of the Cox number necessary and sufficient for the existence of turbulent flow? Gibson believes so, basing his belief on reasoning given just after his equation (3): 'If a volume of size h contains just one zero-gradient point surrounded by fluid with uniform temperature gradient, the Cox number may be estimated by assuming half the fluid volume has twice the surrounding gradient and the rest nearly zero gradient.' Why? 'This gives a Cox number of 2, which may be considered an estimate of the Cox number to be expected for turbulence at transition for a scalar field of $Pr = 1$.' The arithmetic is correct, but no clue is given as to why this calculation is related to the transition to isotropic turbulence. Just a bit later an argument is given for the dependence of C_T, the Cox number at the transition to turbulence, on Prandtl number. (Gibson apparently used the symbol '$\sim$' to denote 'proportional to' rather than 'approximately equal to.' We shall not.) This argument starts from

$$C_T = \langle(\delta T/\delta L)^2\rangle/T_z^{\,2}$$

where δT is a typical temperature anomaly, δL is the thickness of the typical temperature gradient, and T_z is the mean vertical temperature gradient. Apparently it is assumed that the overturn is completely filled with these gradients, which are the largest possible. Taking δT proportional to $L_K T_z$ and δL proportional to $L_B = (\nu D^2/\varepsilon)^{1/4}$,

$$C_T \propto [(L_K T_z)^2/L_B^{\,2}]/T_z^{\,2} \propto Pr$$

Gibson then reasons that if C_T is proportional to Pr and if $C_T = 2$ when Pr is equal to 1 (see preceding), then in general C_T must be $2\,Pr$. If the above is accepted, then the quantity $2\,Pr$ is the value of the Cox number for an eddy barely capable of overturning. But the premise that $C_T = 2$ for $Pr = 1$ is not justified, certainly not as a criterion for transition to isotropic turbulence.

What about the discrepancy? There is no discrepancy between Gregg's observation that Cox numbers above 10 always accompany overturns and a value of the Cox number of several hundred for which eddies have sufficient separation of scales for transition to isotropic turbulence. However, the lack of provable discrepancies in some published interpretations of a limited number of microstructure profiles does not discredit the application of the notion of fossilization to the ocean.

Gibson's Activity Index

A fossil is defined as a structure of a certain scale which had been active Kolmogorovian turbulence in the past but now is significantly affected by buoyancy at that scale. How is a fossil to be recognized? By its rate of strain, γ. Gibson hypothesizes a specific rate of strain, γ_0, at which fossilization takes place. When the source of a distrubance is removed, γ decreases until it reaches γ_0. Then fossilization sets in. Motions with $\gamma > \gamma_0$ are considered active; motions with $\gamma < \gamma_0$ are in the process of fossilization with the scales significantly affected by buoyancy becoming smaller and smaller. How are γ and γ_0 to be calculated? γ could be estimated in several ways. For this paper, Gibson chooses to calculate it from the observed peak wave number k_p of temperature gradient spectra:

$$\gamma = (\varepsilon/\nu)^{1/2} = [\nu D^2(k_p/0.3)^4/\nu]^{1/2} = D(k_p/0.3)^2$$

Calculation of γ_0 is more difficult. If Gibson is right, that turbulence-generating events are extremely rare, it is virtually unobservable. He gives several arguments which relate γ_0 to C_0, the Cox number at the point of fossilization, but C_0 is no more known or observable than γ_0. The accuracy of the estimate of γ_0 is important for evaluation of these ideas. Only by comparison of observed values of γ with γ_0 can a believer in the importance of fossils determine whether a given microstructure patch is actively turbulent or fossil. It is not clear whether γ_0 and C_0 are universal or whether they take on different values in different situations. In the following arguments, δT is the rms temperature anomaly, τ is a $1/e$ time for the destruction of the temperature anomalies by molecular diffusion, L is the thickness of the eddy, and χ is the rate of destruction of temperature variance:

1. The following premises are required: (1) for any eddy, $\chi \sim (\delta T)^2/\tau$ [*Gibson*, 1968]; (2) at the point of fossilization, $\tau \propto N^{-1}$; (3) for any eddy, $\delta T \sim LT_z$; and (4) at the point of fossilization, $L = (\varepsilon/N^3)^{1/2}$ (several arguments given in text). Combining these, we find that χ_0, the value of χ at the point of fossilization, is

$$\chi_0 \propto \varepsilon_0 T_z^2/N^3$$

But χ is defined as $2D\langle(\nabla T)^2\rangle$, so

$$\varepsilon_0 \propto DN^2C_0$$

and, since $\gamma = (\varepsilon/\nu)^{1/2}$,

$$\gamma_0 \propto (C_0/Pr)^{1/2}N$$

To find the constant of proportionality, Gibson makes use of his previously derived result for the laminar-turbulent boundary by considering C and ε at the intersection of the two boundaries, arriving at

$$\varepsilon_0 = [1/2 \times (L_R/L_K)_T{}^{4/3}]DN^2C_0$$

$$\gamma_0 = [1/2 \times (L_R/L_K)_T{}^{4/3}]^{1/2}(C_0/Pr)^{1/2}N$$

All this has only produced a relation between γ_0 and C_0.

2. A similar result is obtained by equating the value of the 'low wave number "fine structure" temperature gradient spectrum observed by *Gregg* [1977] for high Cox number microstructure' with the (universal) form of the temperature gradient spectrum accompanying the inertial subrange of velocity spectra at the wave number $2\pi(\varepsilon/N^3)^{-1/2}$. (The only observations, even tentative, of such a subrange in the ocean have been in surface or bottom boundary layers [*Gargett et al.*, 1979; *Newberger and Caldwell*, 1981].)

Unfortunately, this relation between γ_0 and C_0 is no help in determining γ_0, and it is not clear that either is determinable even in principle. For lack of any other way to proceed, Gibson substitutes the observed Cox number for C_0. (It would seem that he might as well have calculated γ_0 from the observed ε.) He then computes an activity index:

$$A_T = \frac{\gamma}{\gamma_0'} = \frac{\gamma}{[\frac{1}{2}(L_R/L_K)_T{}^{4/3}]^{1/2}(C/Pr)^{1/2}N}$$

in which $[\frac{1}{2}(L_R/L_K)_T{}^{4/3}]^{1/2}$ is taken as 3.6. Gibson believes that the observed value, C, will always be less than C_0 and therefore that the substitution of C for C_0 cannot produce a computed value of A_T smaller than the true value, so that in deciding whether a given patch is active or not, the only error in categorization that could be caused would be judging patches at the point of fossilization to be active when they are in reality fossil. The plot that Gibson presents has as its axes $[D(k_p/0.3)^2]/[3.6(C/Pr)^{1/2}N]$ versus C. For an A_T of 0.1, a typical value for the points on this plot [*Gibson*, 1982, Figure 2], the transition to turbulence apparently occurs at a Cox number greater than 1000.

Gibson's view, then, is that nearly all microstructure patches observed in the ocean to date were in a fossilized state at the time of their observation. They are remnants of an earlier, more powerful stirring and at that time would have been found in a state of isotropic turbulence, little affected by the stratification even at the largest stirring scales. He imagines the stirring in the ocean to be accomplished by very rare, but exceedingly powerful, stirring followed at any given location by long periods of fossilization.

CONTINUOUS CREATION

A contrasting notion, which might be called a 'continuous creation' hypothesis as opposed to Gibson's 'big bangs' hypothesis, is that fluctuations of kinetic energy and temperature are constantly being created on many scales at various rates. A patch of water responds to an increase in driving by increasing its stirring, perhaps entraining neighboring pieces of water. It may even create a microscale water mass, with a slightly different T/S relation. The patch's density may have so changed in relation to the water beside it that it slides horizontally, producing a small 'intrusion' into its neighbors. If the rate of production of turbulence decreases in the patch, its stirring will slow, a more stable gradient may appear in it, and it may lose water to neighboring patches of increasing turbulence. But throughout its history it remains roughly in equilibrium with the stratification it feels, its thickness being maintained near $(\varepsilon/N^3)^{1/2}$. Another, still rather vague, way of stating this is to say that the largest eddies are always in near equilibrium with the stratification. These rather anthropomorphized scenarios are oversimplified and do not apply to much of the water column where the energy input is insufficient to overturn the water at all, much less put it into a state of entraining turbulence. In other parts of the water column small, fitful overturns are incapable of producing the turbulent patches that are so prominent, if rare, in typical profiles.

The continuous creation view is supported by a recent study of microstructure patches in the mixed layer and upper seasonal thermocline at Ocean Station P and in a reservoir [*Dillon*, 1982*a*]. Dillon finds that χN^{-1} is much larger than $(\delta T)^2$, that is, that in these patches the destruction of temperature anomalies by molecular diffusion is so strong that temperature fluctuations would disappear in a time much smaller than N^{-1} if the production of these anomalies by the straining velocity field ceased. This quite surprising result means that if stirring at the scale of a patch were suddenly to stop, the temperature anomalies in it would be smoothed, on average, before they fall back to their equilibrium positions in a stable density profile. Some corroboration for this comes from *Gregg*'s [1980] observation that 'virtually all of the negative gradients are less than 5 cm thick . . . thus all of the maximum ages are much less than the local stability period.' Dillon's calculation is strong evidence for a relatively continuous production of turbulence in the ocean.

Dillon [1982*a*] also finds that L_T is approximately equal to

$(\varepsilon/N^3)^{1/2}$, that is, that the rms Thorpe displacement is approximately equal to the buoyancy length scale, if the appropriate value of N, that felt by the patch, is used in the calculation rather than some large-scale mean value. This is evidence that the thickness of an eddy is determined by its working against stratification, while a rough balance between production and dissipation of turbulence is maintained. In Gibson's language, patches are usually to be found at the onset of fossilization.

What about the laminar-turbulent transition? Some evidence supports the view that the smallest Cox number at which overturns are to be found is about 20, as Gibson says. *Dillon* [1982*a*] finds very few overturns in parts of the water column where the Cox number is less than 20. There is also evidence that fully turbulent flow exhibiting a full Batchelor spectrum including the k^{+1} subrange is found only in regions with Cox numbers in the thousands [*Dillon and Caldwell*, 1980]. If these results are taken as dynamically significant, they might indicate that a transition from nonturbulent flow to marginally overturning turbulence is found when the Cox number is near 20, but that isotropic turbulence with a ratio between production and dissipation scales sufficient to produce a Batchelor spectrum occurs only when the Cox number is much larger, perhaps in the hundreds or even thousands. (Caution must be exercised in attributing dynamic significance to the Cox number because after all in the region of an overturn C must have a value significantly larger than 1, and also C is really just a ratio between the high-frequency signal of turbulence and low-frequency features which may be regarded as noise interfering with our view of the turbulence. So the values of C required to observe various phenomena may reflect only the level of signal to noise needed.)

TURBULENCE PARAMETERS

To define the condition of a patch of water suspected of being in a turbulent condition, the following parameters are often used:

- L_p estimate of the vertical extent of the patch;
- L_T measure of the mean vertical distance individual parcels of water in the patch have been displaced in the vertical (the Thorpe scale [*Thorpe*, 1977]);
- L_m time-averaged estimate of the vertical displacements of parcels observed at a fixed depth;
- N local buoyancy frequency, understood here as a measure of the stratification against which the overturning eddies must strain, or in a slightly different view as the reciprocal of the time in which displaced parcels would be returned to their positions of static stability by buoyancy;
- ε kinetic energy dissipation rate; and

a measure of the rate at which the displaced parcels are exchanging heat with their immediate environments, thereby losing their (thermal) identity as displaced when their heat excess or defect has made its contribution to the vertical heat transport. Either the Cox number or χ can be used.

Some precision in the definition of these quantities is required for a sensible discussion:

The 'patch size' L_p is usually determined by eyeballing vertical temperature gradient profiles for the extent of 'active' regions, regions in which the temperature gradient seems to remain nonzero while switching back and forth from positive to negative. This L_p is not at all the same as the more precisely defined L_T, which is the rms of the displacements of the water parcels from their positions in a reordered stable profile. Quite commonly L_p is much larger than L_T; that is, the overturns in a patch often do not extend all the way from top to bottom. We might say that a patch of more or less continuous turbulence may be composed of a number of overturning eddies. (We might suspect such a patch of being a 'fossil,' in the sense that it might have been created by an overturn of vertical extent L_p, but at the time of observation have only enough energy available to it to sustain smaller eddies. The same patch, however, might have been created by the several overlapping eddies that are observed.)

The scale L_m commonly is calculated from a time series of, for example, the deviations T' of the temperature (better the density) at a fixed depth from the mean T at that depth, together with an estimate of the mean vertical gradient at that depth, T_z, as

$$L_m = \langle (T')^2 \rangle^{1/2}/T_z$$

This estimate is as good as can be done with data from moored or constant-depth-towed sensors. It is similar to the displacements often calculated from sequences of CTD profiles for use in internal wave studies. It is quite different from L_T in that it lumps internal wave displacements with displacements caused by turbulence, whereas displacements due to internal waves are not directly included in the calculation of L_T. In the absence of overturns, L_m is an estimate of the mean internal wave displacement, while L_T is zero. In the absence of internal waves the two scales may be similar. (Internal wave straining of a profile with turbulent overturns may have a minor effect on L_T.)

The way in which N is defined is also important. We might consider two definitions of N, an internal one calculated from the mean density gradient inside an overturning eddy which yields the best estimate of the period of vertical oscillation of parcels of water inside the eddy or an external one calculated over a vertical region including the edges of the patch which better estimates the stratification against which the entraining turbulence of the patch is working. These definitions, which seem fairly clear for an idealized eddy, become difficult and sometimes impossible to apply to a real profile, which presents a confusing picture of overlapping overturns. The best choice seems to be an N calculated over a region somewhat larger than the patch, using the mean density gradient of the reordered, stable Thorpe profile. Use of the reordered profile removes much of the dependence of the calculated value of N on the phase the overturn happens to have when the instrument passes through it. In no event is a value of N calculated over a vertical region many times thicker than the patch relevant to the dynamics of that patch, and of course oceanic averages such as the commonly quoted 0.005 s^{-1} are relevant only in making crude estimates.

The averaging required in the estimation of ε is fairly straightforward, if it can be calculated over the same region as N and L_T, but the meaning of estimated Cox numbers requires some discussion. One view of the significance of the Cox number is that it represents the amount of heat being given up or absorbed by temperature anomalies, parcels of water that have been displaced from their equilibrium positions in the water column, normalized by the molecular heat

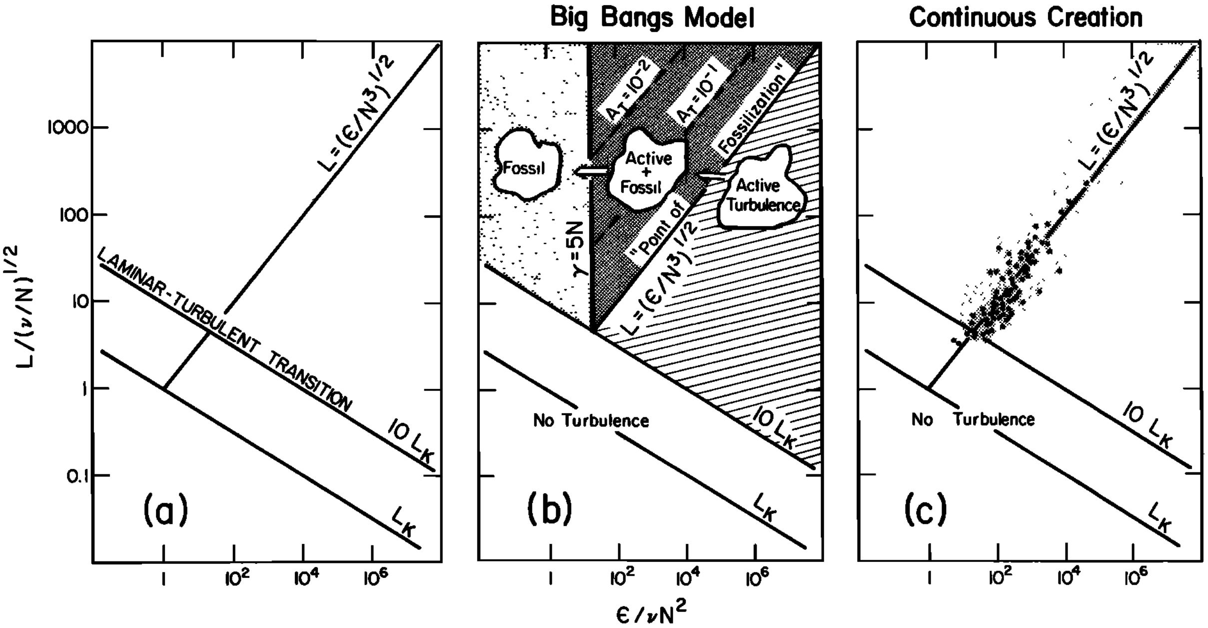

Fig. 1. Nondimensionalized vertical thickness scale versus nondimensionalized kinetic energy dissipation rate. The horizontal axis can also be considered as γ^2/N^2, where γ is the strain rate. Any microstructure patch may be plotted on these axes, but the interpretation of L varies depending on the view taken. In the 'big bangs' view, L is taken to be the patch size. In the 'continuous creation' view, L is defined as L_T, the rms Thorpe scale. Turbulent patches are expected only for scales larger than the Kolmogoroff dissipation scale by some factor, taken as 10 here. In Figure 1*b* the distribution and classification of patches according to the 'big bangs' hypothesis is shown. In Figure 1*c* the distributions expected in the 'continuous creation' view is shown, together with data from *Dillon* [1982*a*]. It should be noted that if L is taken as L_p, the 'patch size,' rather than L_T, the points would fall mostly in the 'active + fossil' region in Figure 1*b*, as expected by *Gibson* [1982]. In one sense the point at issue here is not the correctness but rather the usefulness of the different views.

flux through the region. In calculating the heat flux, the heat content of each parcel is counted as though it were still in its equilibrium position, but the heat gained or lost by the parcel is counted as deposited to (or absorbed from) its observed position. In other words the Cox number expresses the effect of mixing alone, not stirring. The closeness of the connection between χ, the rate of destruction of temperature anomalies, and the Cox number is clear; in the Osborn-Cox model it is the destruction of the temperature anomalies that transports the heat.

The calculation of heat transport from the Cox number involves an assumption that is often not explicit. The estimated heat flux depends on the selection of the region of interest in a crucial way. For example, if we are interested in the rate of heat transport through a particular patch, then we calculate the Cox number using the vertical temperature gradients within the patch. If we require an estimate over a large region, a seasonal thermocline for example, we use the quantities measured over it. In a given profile the thermocline may contain eddies at some depths, none at others. For meaningful heat flux calculations, the assumption must be made that the structure observed on each pass is one realization of a process that extends homogeneously throughout the thermocline (and that our sampling has been sufficient to determine the statistical parameters of the process). The anomalies observed on one pass may only carry heat for a fraction of a meter; in this interpretation we assume that at another time similar anomalies will carry the heat onward. If this is so, an eddy diffusivity for the thermocline may be meaningfully estimated as usual as

$$K = D_T \cdot C$$

Often, however, this assumption will not be justified and the eddy diffusivity estimate is not meaningful. For example, in a study of the bottom layer on the Oregon shelf [*Newberger and Caldwell*, 1981] we found Cox numbers in the tens of thousands in the bottom layer. Extending the computations throughout the water column would have produced a much smaller value, because the mean gradient would have been larger, but still a value indicating a large eddy diffusivity. But this value would have been erroneous because of the insulation provided to the bottom layer by its cap, a layer of extremely stable water just above it. Thus the Cox number calculation of the eddy diffusivity works only in a region that is vertically homogeneous in the statistical sense that an eddy is as likely to be found at one depth as another. The sampling required to establish this may be tedious indeed.

A Graphic Representation

The possible states of a microstructure patch can be represented graphically by plotting its position on a coordinate system with axes $\varepsilon_{ND} = \varepsilon/(\nu N^2)$ and $L_{ND} = L/(\nu/N)^{1/2}$ (Figure 1*a*). The horizontal axis, the nondimensionalized kinetic energy dissipation rate, may be thought of as the square of the ratio of strain rate to buoyancy frequency, γ^2/N^2, as $L_R/(\nu/N)^{1/2}$, or as $(L_R/L_K)^{4/3}$. Here L_R is defined as $(\varepsilon/N^3)^{1/2}$ and L_K is the Kolmogoroff dissipation scale $(\nu^3/\varepsilon)^{1/4}$.

The vertical axis, L_{ND}, a nondimensionalized vertical length scale, has a different value and interpretation depending on the view taken with respect to fossil turbulence. In Gibson's 'big bangs' view, L is taken as L_p, an estimated patch size, and most patches are found in the region labeled 'active + fossil' (Figure 1*b*). In the 'big bangs' view a patch

TABLE 1. Characteristics of Equilibrium Eddy

Parameter	Value in Largest Eddy	Value in Cascade	Value in Smallest Eddies	Ratio
Thickness	L_T	$L_T/kL_T = 1/k$	$L_{ND}^{-1/2}(\nu/N)^{1/2}$	$L_{ND}^{3/2}$
Wave number	$1/L_T$	k	$L_{ND}^{1/2}(\nu/N)^{-1/2}$	$L_{ND}^{-3/2}$
Time scale	N^{-1}	$(kL_T)^{-2/3}N^{-1}$	$(L_{ND}N)^{-1}$	L_{ND}^{-1}
Velocity	NL_T	$(kL_T)^{-1/3}NL_T$	$L_{ND}NL_T$	L_{ND}
Strain rate	N	$(kL_T)^{2/3}N$	$L_{ND}N$	L_{ND}^{-1}
Dissipation	νN^2	$(kL_T)^{4/3}\nu N^2$	$L_{ND}^2\nu N^2 = L_T^2N^3$	L_{ND}^{-2}
$E(k)$	$N^2L_T^3$	$(kL_T)^{-5/3}N^2L^3$	$N^2L_T^3(L_{ND})^{-5/3}$	$L_{ND}^{5/3}$

is created in the region labeled 'active' on this plot and moves roughly to the left, beginning to fossilize as it passes the line labeled 'point of fossilization.' The patch then becomes more fossilized, the activity index A_T defining its degree of fossilization. When γ equals $5N$, the patch has become 'completely fossil.'

In the 'continuous creation' view the vertical scale L is defined more precisely as the Thorpe scale L_T. The admittance characteristic of the stratified environment is hypothesized to be such that energy is received most efficiently by motions with time scale N^{-1}. This seems almost obvious because N is the resonant frequency for vertical motions and the motion is most constrained in the vertical. The hypothesis is also made, as usual in turbulent flows, that the energy is fed into the largest eddies; that is, only the largest eddy is forced from outside. This scale is determined by the rate of energy input to it, together with the value of N. Straining of the velocity field in the eddy will generate smaller eddies, and the energy cascades to smaller and smaller eddies until the dissipative scale is reached, exactly as in nonstratified turbulent flows. The stratification affects that turbulence for the most part only through its effect on the largest eddy. The characteristics of the turbulence are assumed here to be determined exactly as they would be for a nonstratified eddy with the same time scale and the same energy input. In such a system the time required to establish or adjust the eddy structure to variations in the energy input is not much larger than N^{-1}, because the time scales of the smaller eddies are smaller than the time scale of the largest eddy and we assume, as usual, that an eddy loses its energy to smaller eddies within its overturning time. Therefore the entire eddy structure is in rough equilibrium with the forcing at frequency N; if the forcing changes, the eddy structure will adjust itself in a time roughly equal to N^{-1}.

With these hypotheses the characteristics of the turbulence can be written down just as in nonstratified turbulence. Here L_{ND} is defined in terms of the Thorpe scale, as $L_{ND} = L_T/(\nu/N)^{1/2}$ (the last column in Table 1 is the ratio of the value of each parameter at the productive scale to its value at the dissipative scale). A discussion of this scheme can be found in standard turbulence texts, *Tennekes and Lumley* [1972], for example.

From the dissipation in the smallest eddies given above, $L_T = (\epsilon/N^3)^{1/2}$ and L_{ND} can be found in terms of ϵ_{ND}, yielding $L_{ND} = \epsilon_{ND}^{1/2}$. Accepting this scheme, we expect to find the loci of turbulent patches along the $L_{ND} = \epsilon_{ND}^{1/2}$ line, as shown by the shading in Figure 1*c*, well above the line representing the nondimensionalized value of the Kolmogoroff scale L_K. And indeed we do find the patches there. The set of patches whose properties were determined by *Dillon* [1982*a*] lie on the plot exactly where they are expected. The scatter in the plot is partially caused by the sampling; relationships such as those in the table above are expected to hold only in the mean. It is expected also that some patches will not be precisely at equilibrium with energy input exactly equal to dissipation and that advective effects may be significant in a given case even though they would average out in the long term. Both of these effects will cause scatter on the plot.

IMPLICATIONS FOR MEASUREMENTS

The indications are, then, that this 'continuous creation' hypothesis usefully describes oceanic turbulence. If so, then there is hope for the relatively simple measurement of oceanic turbulence and its effects by measuring L_T along with vertical gradients of T and S (N and $d\rho/dz$ need to be calculated from the gradients):

1. The kinetic energy dissipation rate ϵ is $L_T^2N^3$.
2. The vertical eddy diffusivity K is L_T^2N.
3. The vertical heat flux is $-\rho C_pL_T^2N\ dT/dz$.
4. The vertical salt flux is $-L_T^2N\ dS/dz$.
5. The vertical buoyancy flux is $-L_T^2N\ d\rho/dz$.

Of course, multiplicative constants of order 1 are expected in each case; but in the data we have examined to date, these constants are very close to 1 [Dillon, 1982a]. (Thorpe [1978] found $K/L_T^2N)^{-1}$ to be 0.1, but his estimates of K were less direct than Dillon's.) The estimation of L_T is really not easy in many cases. The starting point for its calculation, in regions where the T/S relation cannot be relied upon, is a density profile with a vertical resolution of a few centimeters and noise level less than the density difference between samples. In the ocean we are limited not only by the noise levels of our instruments but also in some cases by the uncertainty in our knowledge of the relationship which gives salinity as a function of temperature and conductivity, and even perhaps in some cases by uncertainty in the equation of state. For example, in a situation where N is 3 cph, the permissible noise level in 3-cm samples must be 0.1 ppm, at the limit of our instrument system. In situations such as freshwater lakes, where the density is a function only of the temperature, the estimation of L_T is not as difficult.

WHAT ABOUT FOSSILS?

In the 'continuous creation' view, fossils do exist, as places in the water column which have been more active in the past than at the time of observation. They can be recognized most easily in terms of the relation between the thickness scales. If LT_T is much smaller than L_p, the eyeball-estimated patch size, then probably L_T has decreased from a larger value, and so has ϵ. Gibson's activity index A_T is simply related to L_p and L_T, if we speculate that L_p is the 'fossil L_T,' that is, that during the generating event overturning took place throughout the patch. Then estimating γ as above $A_T = \gamma/\gamma_0 = L_T/L_p$; the activity index is equal to the ratio of the present L_T to its largest value in the past, assumed to be L_p. Such fossils will be found, but as long as the current crop of turbulence is in balance with its driving forces, and as long as our sampling plans take into account the possibility of occasionally encountering extremely energetic patches, we need pay them no attention in our calculations.

Of course, these questions are far from settled, but instruments are becoming available with which the sampling necessary to settle them can be obtained, at least in the upper ocean. The next few years should be quite interesting.

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REFERENCES

Caldwell, D. R., T. M. Dillon, J. M. Brubaker, P. A. Newberger, and C. A. Paulson, The scaling of vertical temperature gradient spectra, *J. Geophys. Res.*, *85*, 1917–1924, 1980.

Dillon, T. M., Vertical overturns: A comparison of the Thorpe and Ozmidov length scales, *J. Geophys. Res.*, *87*, 9601–9613, 1982*a*.

Dillon, T. M., The energetics of overturning structures: Implications for the theory of fossil turbulence, *J. Phys. Oceanogr.*, in press, 1982*b*.

Dillon, T. M., and D. R. Caldwell, The Batchelor spectrum and dissipation in the upper ocean, *J. Geophys. Res.*, *85*, 1910–1916, 1980.

Gargett, A. E., T. B. Sanford, and T. R. Osborn, Surface mixing layers in the Sargasso Sea, *J. Phys. Oceanogr.*, *9*, 1090–1111, 1979.

Gibson, C. H., Fine structure of scalar fields mixed by turbulence, *Phys. Fluids*, *11*, 2305–2327, 1968.

Gibson, C. H., On the scaling of vertical temperature gradient spectra, *J. Geophys. Res.*, *87*, 8031–8038, 1982.

Gregg, M. C., A comparison of finestructure spectra from the main thermocline, *J. Phys. Oceanogr.*, *7*, 33–40, 1977.

Gregg, M. C., The effects of bias error and system noise on parameters computed from *C*, *T*, *P* and *V* profiles, *J. Phys. Oceanogr.*, *10*, 1468–1492, 1979.

Gregg, M. C., Microscale patches in the thermocline, *J. Phys. Oceanogr.*, *10*, 915–943, 1980.

Newberger, P. A., and D. R. Caldwell, An inertial subrange in microstructure spectra, *J. Geophys. Res.*, *86*, 4265–4268, 1981.

Oakey, N. A., Determination of the rate of dissipation of turbulent energy from simultaneous temperature and velocity shear measurements, *J. Phys. Oceanogr.*, *12*, 256–271, 1982.

Stillinger, D. C., An experimental study of the transition of grid-generated turbulence to internal waves in a salt-stratified water channel, Ph.D. thesis, Univ. of Calif. at San Diego, La Jolla, 1981.

Tennekes, H., and J. L. Lumley, *A First Course in Turbulence*, MIT Press, Cambridge, Mass., 1972.

Thorpe, S. A., Turbulence and mixing in a Scottish loch, *Philos. Trans. R. Soc. London Ser. A*, *286*, 125–181, 1977.

Thorpe, S. A., The near-surface ocean mixing layer in stable heating conditions, *J. Geophys. Res.*, *83*, 2875–2885, 1978.

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