Alternative Interpretations for Microstructure Patches in the Thermocline

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ABSTRACT

Two interpretations of microstructure patches measured in the main thermocline of the North Pacific by Gregg (1980) are questioned. He concludes that the observed microstructure is not fossil-temperature turbulence and that the observed Cox numbers imply vertical diffusivities by turbulent mixing which are about 10^{-2} times smaller than canonical values of order 10^{-4} m^2 s^{-1}. The first conclusion that the microstructure is not fossil depends on an unnecessary assumption that in order to be fossil the microstructure must not be moving: actually, fossil-turbulence microstructure must always have internal wave and laminar restratification motions. The second conclusion depends on the first. The pattern of Thorpe-displacement scales for the patches containing zero temperature gradients implies they were produced by turbulence with vertical scales as large as several meters. However, the large distances separating the zero-gradient points implies that the microstructure is fossil at all scales at the time of observation. Because the microstructure is fossil it follows that the observed Cox numbers C are small compared to their previous values C_0 when the microstructure was actively turbulent at all scales. Model calculations give $C_0 \approx 0.2 L_T^2 N/D$ values in the patches as large as 4×10^5 compared to observed C values less than 10^3, where L_T is the maximum Thorpe displacement. The data sample is apparently too small to include representative active turbulent regions because such regions are so intermittent in time and patchy in space. Turbulent vertical diffusivity estimates corrected for undersampling are well within the canonical range of values.

1. Introduction

Gregg (1980) has discussed possible interpretations of small-scale temperature fluctuations measured in vertical profiles through two depth ranges, 200–400 m and 800–1200 m, at 28°N, 155°W. A wealth of evidence about possible mixing processes in the main thermocline of the central North Pacific is provided by these data. The relative importance of salt fingering and Kelvin-Helmholz wave breaking were estimated by examination of the microstructure signals for signatures of different mixing mechanisms. The frequency and distribution of zero-vertical-temperature-gradient points were used to identify sites of active turbulence. As in Gregg (1977), low values of the mean Cox number for three expeditions to the same location were interpreted as indicating a low level of turbulent vertical heat flux compared to classical values estimated from mean temperatures and currents.

The purpose of the present paper is to question two interpretations of the microstructure data proposed by Gregg (1977, 1980). The first interpretation questioned is that the observed microstructure is turbulent and cannot be fossil turbulence. Gregg (1980) shows that about 25% of the data records considered consist of patches of zero-temperature-gradient activity. The distances separating the zero crossings were measured, and from the implied diffusional time scales Gregg concludes that the patches are actively turbulent and not fossil scars of previous turbulence. This conclusion will be examined in Section 2. It is shown that isotropic turbulence is necessary to produce zero gradient points, but that zero gradient points can persist after the turbulence has ceased. Rates of strain indicated by the measured separation distances are too small for the microstructure to be turbulent except at small scales or not at all.

From the low values of the bulk Cox numbers, and since overturning was observed in a small fraction of the vertical profiles, Gregg (1980, p. 941) concludes that the vertical heat flux due to the turbulent events must be quite low compared to the fluxes implied by mean property distributions. This is the second interpretation questioned. The problem of estimating average dissipation rates and diffusivities from microstructure measurements is examined in Section 3. Using the Osborn-Cox (1972) model $K = DC$, where K is the vertical diffusivity, D the molecular diffusivity and $C = \overline{(\nabla\theta^2)}/(\nabla\bar{\theta})^2$ is the Cox number of a scalar property θ such as temperature, a mean C value of about 700 is required for $K = 10^{-4}$ m^2 s^{-1}. Gregg (1977) reports C values of 2 (September 1971), 10 (June 1973) and 59 (February 1974) from the 28°N, 155°W site in the main ther-

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mocline and below. Gregg (1977) states that either the canonical values of K may be too large, or that a large variety of rings, quasi-geostrophic eddies and intrusions may provide mechanisms for vertical transport other than small-scale turbulence. However, it is not made clear how these horizontal motions enhance vertical transport other than by providing additional sources of small-scale turbulence.

Gregg (1980) shows that the microstructure activity is quite intermittent in the vertical direction, so that the most probable value of observed Cox number, the mode, is much less than the mean. Cox numbers may be even more patchy in the horizontal direction and if so, the C values from each layer in a single vertical profile may underestimate the layer mean. The possibility is discussed in Section 3 that a combination of factors may tend to cause underestimates of the space-time average Cox number due to undersampling. Patchiness in space causes an underestimate of the spatial average from a short data record. Intermittency in time of the spatial average causes an underestimate of the space-time average when only a few times are sampled. Gibson (1981) discusses the problem of sampling oceanic turbulence in strongly stratified layers and concludes that even towed-body dissipation measurements may have underestimated the mean dissipation rates due to undersampling.

Factors as large as 12–350 are needed to resolve the discrepancy between the Gregg (1977) values of C = 2–60 and the required value of about 700. Such factors are found to be consistent with the available evidence. Using the fossil turbulence model of Gibson (1980) and the Gregg (1980) measurements of zero-gradient-point separation lengths and Thorpe overturning scales, it is shown in Section 3 that the observed microstructure patches are fossil turbulence in an advanced state of decay and that previous Cox numbers were orders of magnitude larger when the microstructure was actively turbulent.

Finally, Section 4 is devoted to a discussion of definitions. An important limitation to progress in understanding stratified turbulence is the lack of general agreement over precise definitions of terms, particularly the terms "turbulence" and "fossil turbulence." Gregg (1980) uses the term "turbulence" to include classes of flow affected by buoyancy, viscosity and molecular diffusion. Based on the observations of Nasmyth (1970) and his resulting spectral evolution model for fossil turbulence, Gregg (1980) assumes the velocity field in fossil turbulence must be identically zero. Alternative definitions are proposed, based on universal similarity hypotheses for turbulent velocity and scalar fields extended to stratified fluids by Gibson (1980), which are consistent with the Gregg (1980) observations.

2. Zero gradient points and the existence of turbulence

The patches of zero-crossing temperature gradients are taken to be regions of active turbulent mixing by Gregg (1980) on the grounds that negative temperature gradients probably represent static instabilities which will generate vertical velocities and "hence turbulence" (p. 927). However, it is not clear that such restratification velocities will necessarily be turbulent and not laminar or internal wave motions. It is also not clear whether such vertical velocities will be significant compared to the ambient velocity field, especially when the density fluctuations associated with the zero-gradient points are small. Because a fluid moves in the vertical does not mean the motion is turbulence. Because zero-gradient points must be caused by turbulence does not mean the fluid must continue to be turbulent as long as the zero-gradient points exist.

Gibson (1968) has shown that zero-gradient points of dynamically passive scalar fields such as temperature are a characteristic signature of turbulence which cannot be produced by laminar flows. Random, overturning motions of turbulence are necessary to distort the planar isothermal surfaces so that molecular diffusion can pinch off isolated hot and cold spots and cause the isothermal surfaces to become multiply connected. The production, evolution and decay of zero-gradient points is the central physical mechanism of small-scale turbulent mixing according to the Gibson (1968) theory.

The zero-gradient-point generation and decay process is illustrated schematically in Fig. 1, which shows the response of a uniform-gradient temperature field to a transient test eddy. The velocity $\mathbf{v}^\theta$ of an isothermal surface consists of the fluid velocity $\mathbf{v}$ and a diffusion velocity $D\nabla^2\theta\nabla\theta/(\nabla\theta)^2$, i.e.,

$$\mathbf{v}^\theta = \mathbf{v} - D\nabla^2\theta\nabla\theta/(\nabla\theta)^2, \qquad (1)$$

as shown by Gibson (1968). When $\nabla^2\theta = 0$ the isotherms move with the fluid, but, when the isotherms are distorted, diffusion velocities develop in a direction to return them to the $\nabla^2\theta = 0$ configuration. Estimating from Eq. (1) using $|\nabla\theta| \approx \Delta\theta/L$, etc., in order for the eddy in Fig. 1 to overcome such diffusional relaxation, its velocity difference must exceed a value of order D/L, where D is the molecular diffusivity of θ and L is the size of the eddy. The test eddy velocity field of Fig. 1 is non-zero for a time $\sim L^2/D$, just long enough to produce zero-gradient points, which then evolve only by diffusion. In a turbulent flow, all eddies must be larger than $(\nu/\gamma)^{1/2}$, which is larger than our test eddy by a factor of $(\nu/D)^{1/2}$, and will produce more complicated structures than shown in Fig. 1 by continued straining. The hot and cold spots can undergo secondary split-

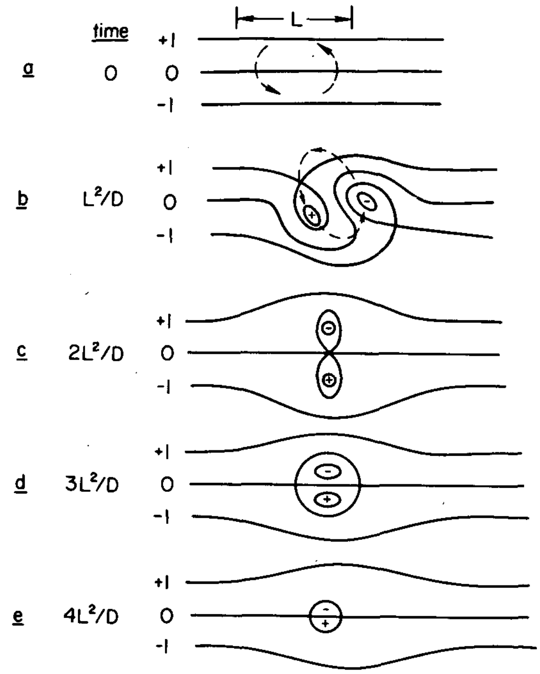


FIG. 1. Generation of zero-gradient configurations by a small transient eddy. (a) An eddy on scale L acts on a uniform temperature gradient region for a time L^2/D and then ceases. The eddy velocity must be greater than D/L to cause large distortions of the isothermal surfaces (note: actual turbulent eddies occur at larger scales and are more persistent than our test eddy, producing more complex structures). (b) Distorted isothermal surfaces become multiply connected. A hot spot and a cold spot form with corresponding hot and cold saddle points. (c) The saddle points merge momentarily to form a double saddle point. (d) The double saddle point forms a saddle line and a doublet. The geometry of the doublet is stabilized by molecular diffusion as its size and strength decrease. (e) The doublet vanishes by molecular diffusion.

ting, as shown in Gibson (1968), but no different quasi-stable zero-gradient configurations will be produced.

Assuming $V > D/L$ in Fig. 1, the eddy will pull hot fluid down and cold fluid up, distorting the isotherms so that zero-gradient points can form. When a hot spot forms, a hot saddle point is formed, and when a cold spot forms, a cold saddle point must also appear. The different zero gradient configurations can be distinguished by the signs of the eigenvalues of the tensor $\partial^2\theta/\partial x_i \partial x_j$ at the point. Hot spots correspond to $(-, -, -)$, cold spots to $(+, +, +)$, hot saddle points to $(-, -, +)$ and cold saddle points to $(+, +, -)$. Zero-gradient configurations such as volumes $(0, 0, 0)$, hot lines $(-, -, 0)$ and cold lines $(+, +, 0)$ are unstable.

As the overturn proceeds the two saddle points may merge to form a double saddle point, with $\theta_{,ij}$ eigenvalues $(0, 0, 0)$. The double saddle point is unstable and will immediately form a ringlike saddle line, with $\theta_{,ij}$ eigenvalues $(0, +, -)$. The $\theta = 0$ isothermal surface bifurcates into two lobes, forming a doublet. The doublet formation process shown in Fig. 1 illustrates all five zero-gradient configurations which can persist for finite periods of time; that is, hot spots, cold spots, hot and cold saddle points, and saddle lines.

Having been formed, the doublet configuration has a diffusional stability which opposes any convective motions which will tend to split it up. This can be shown by considering the direction of the hot and cold spot zero-gradient point velocities using the expression derived in Gibson (1968), i.e.,

$$\mathbf{v}' \approx \mathbf{v} - 3D\frac{\nabla(\nabla^2\theta)}{\nabla^2\theta}, \qquad (2)$$

where $\mathbf{v}'$ is the velocity of a hot or cold spot. The direction of the diffusional velocity is toward a configuration of maximum symmetry, with no skewness in the temperature distributions near the hot and cold points. Thus the doublet configuration will tend to preserve itself until the hot and cold spots annihilate each other, as shown in Fig. 1. Doublets not only look like soap bubbles, they act like soap bubbles. Molecular diffusion tends to preserve their geometry against convective distortion just as surface tension preserves the geometry of a two-lobed soap bubble.

Suppose we consider an isolated temperature doublet in a fluid which is not moving. The gravitational instability will be balanced by the diffusional stability when the gravitational velocity scale $v_g \sim LN'$ is comparable to or less than the diffusion velocity $v_d \sim D/L$ which will develop if the doublet is distorted. Thus the doublet will be diffusionally stable to gravitational distortion when

$$L \leqslant (D/N')^{1/2}, \qquad (3)$$

where L is the size of the doublet and N' is the Väisälä frequency corresponding to the temperature difference ΔT between the temperature of the hot and cold points. That is,

$$N' = \left(\frac{g\alpha\Delta T}{\rho L}\right)^{1/2}, \qquad (4)$$

where α is the coefficient of thermal expansion, ρ the fluid density and g the acceleration of gravity.

In the initial stages, N' in (3) and (4) must be less than the ambient

$$N = \left(\frac{g\alpha}{\rho}\frac{\partial T}{\partial z}\right)^{1/2}$$

because ΔT must be less than its initial value $L(\partial T/\partial z)$. The scale of the largest doublet produced by turbulence will be the Batchelor scale $(D/\gamma)^{1/2}$, where γ is the rate of strain of the fluid. The smallest rate of strain in a stratified turbulent flow is about $5.5N$ according to Gibson (1980). Therefore, the largest doublet in a stratified microstructure region produced by turbulence should be no larger than $(D/$

$N)^{1/2}$. But since $L < (D/N)^{1/2}$ and $(D/N)^{1/2} < (D/N')^{1/2}$ then $L < (D/N')^{1/2}$, which satisfies the condition of Eq. (3) for a doublet to be diffusionally stable with respect to gravitational forces. Similar arguments show that molecular diffusion will also prevent gravitational distortions of the vertical alignment of a hot spot with its saddle point during their ultimate annihilation in the return of a turbulent microstructure region to a uniform vertical gradient. The process is illustrated in Fig. 2. Thus we conclude that zero-gradient configurations produced by turbulence in a stratified fluid will tend to be stabilized by molecular diffusion at all stages of their existence, and will not produce turbulence even though gravitational instabilities (temperature inversions) exist.

Turbulent eddies will produce more complex zero-gradient structures than the prototypes shown in Figs. 1 and 2, with secondary splitting of maximum points. Large-scale dispersion may cause N' to exceed N by bringing hot and cold spots from widely separated layers together. Resulting gravitationally driven velocities will be larger than either diffusion velocities or the ambient turbulence when $N' > \gamma$. However, the local restratification velocities will reduce the length scales of the zero-gradient points so that the Reynolds number of the flow remains small and the flow nonturbulent. The Reynolds number is $v_g L/\nu = N'L^2/\nu = D/\nu < 1$, where L^2 will be D/N' when $N' > \gamma$, and γ is the rate of strain characteristic of the microstructure patch. The Rayleigh number of the flow $\mathrm{Ra} = g\alpha L^4(\delta T/L)/\nu D\rho$ [a factor of g/ρ is missing from the expression in Gregg (1980, p. 924)] can be written as $(N'L^2/D)(N'L^2/\nu) = \mathrm{PeRe} = \mathrm{PrRe}^2$, where $N'L^2/D$ is the Peclet number. When Ra exceeds a critical value Ra_{cr} for a heated boundary layer a cellular flow pattern results, but it is difficult to say what the relevance of the Rayleigh number is in a microstructure field since gravity-driven cellular flows will always be present because of the zero-gradient points, even when $\mathrm{Ra} < \mathrm{Ra}_{cr}$, as shown in Fig. 2. Even when N' is large, Re will still be small and the resulting flow nonturbulent: locally, the Rayleigh number should approach a universal constant of order Pr^{-1} since $\mathrm{Ra} \approx \mathrm{Re}^2\mathrm{Pr}$ and $\mathrm{Re} \approx \mathrm{Pr}^{-1}$.

Observations of zero vertical gradients of temperature (and density) in the ocean are therefore not an indication of a resultant turbulent velocity field, as assumed by Gregg (1980). Low-Reynolds-number cellular velocity fields will result from the presence of zero-gradient points in a gravitational field, but the velocity is not turbulence in the usual sense of the word. Density inversions are a necessary but not a sufficient condition for active turbulence in a stratified fluid.

Gregg (1980) presents measurements of the vertical scale separating points of zero vertical temperature gradient. The diffusional time constant of temperature fluctuations on the measured scales is shown to be less than the Väisälä period $2\pi N^{-1}$, which Gregg takes as evidence that the microstructure regions must be actively turbulent. Length scales as large as 5 cm and as small as 1.5 cm were observed in layers with $N \approx 5 \times 10^{-3}$ rad s^{-1}.

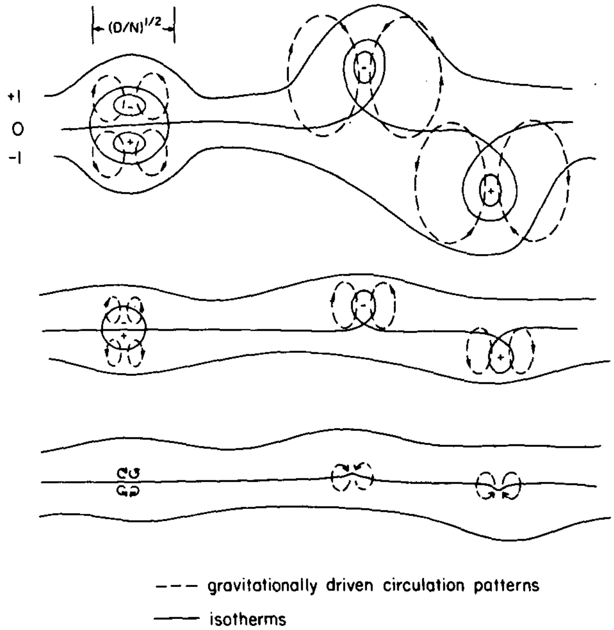


FIG. 2. Annihilation of zero-gradient configurations by diffusion in a gravitational field and induced flow patterns. (a) A doublet, a cold spot and a hot spot are shown in a gravitational field. All have density inversions, but their geometry is stabilized by molecular diffusion. The gravity driven viscous flow patterns shown are not turbulent in the usual sense. (b) The size and strength of the zero-gradient features and the resulting flows decrease with time due to molecular diffusion and viscosity. (c) All zero gradients annihilate, leaving only large-scale distortions in singly connected isothermal surfaces.

Clearly a relationship exists between the zero-gradient separation distance and the diffusive cutoff wavenumber of the temperature gradient spectrum. To explore this relationship a turbulent mixing signal was synthesized on a computer by adding six sine waves with amplitudes corresponding to the Batchelor scalar gradient spectrum $k\exp(-2k^2)$. The mean zero-gradient separation distance L was computed for 100 cycles of the lowest frequency. Wavenumbers chosen were 0.1, 0.18, 0.31, 0.58, 0.81 and 1.2. It was found that the zero-crossing wavenumber π/L was 16% greater than the wavenumber of the spectral peak. For the universal scalar gradient spectrum the peak occurs at about 0.3 $(\gamma/D)^{1/2}$. Therefore the zero-crossing separation L measured by Gregg (1980) can be used to estimate the rate of strain of the microstructure since π/L should be about 0.4 $(\gamma/D)^{1/2}$.

With $k = 2\pi/\lambda$ and $\lambda = 10$ cm, we find $\gamma \approx 3.5 \times 10^{-3}$ rad s^{-1}, which is about eight times less than the minimum rate of strain $\gamma = 5.5N = 2.75 \times 10^{-2}$ rad s^{-1} required for the microstructure to be turbulent according to the criterion of Gibson (1980). Thus the microstructure region should be classified as fossil temperature turbulence at all scales. The smallest zero-gradient separation distance reported by Gregg (1980) was 1.5 cm, corresponding to a rate of strain of 3.9×10^{-2} rad s^{-1}. This is slightly greater than $5.5N$, suggesting that the microstructure is associated with nearly saturated internal wave motions, but with little turbulence. Therefore we conclude that most of the microstructure observed by Gregg (1980) is nonturbulent, although the presence of large numbers of zero-gradient points indicates that it has been turbulent in the past.

It might be questioned whether the thermistor used by Gregg (1977) had adequate frequency response to accurately resolve zero-gradient separation lengths in the range 1.5–5 cm. About 20 temperature gradient spectra are presented for the depth range 200–1200 m in Figs. 9b–9d of Gregg (1977), apparently representing all of the data considered in Gregg (1980). All the spectra seem to exhibit spectral peaks, which suggests that the true separation distances between zero-gradient points has been resolved in Gregg (1980). It should be noted that even if the patches were all actively turbulent and the thermistor frequency response prevented resolution of the true dissipation scale, the Cox numbers of the microstructure patches would still be underestimated by the same large factors as shown by Fig. 3 in the following section, but for a different reason than suggested by the present paper.

In the next section we show that the range of possible turbulent wavelengths λ is $1.2L_R \geqslant \lambda \geqslant 15L_K$, where L_R is the Ozmidov length $(\epsilon/N^3)^{1/2}$, L_K the Kolmogoroff length $(\nu^3/\epsilon)^{1/4}$, and ϵ the rate of viscous dissipation of kinetic energy per unit mass. Since $\gamma = (\epsilon/\nu)^{1/2}$ we find $\epsilon = 1.8 \times 10^{-9}$ m^2 s^{-3}, $1.2L_R = 14.5$ cm, and the viscous wavelength $15L_K = 8.4$ cm in the microstructure regions with 1.5 cm zero-crossing scales and $N = 5 \times 10^{-3}$ rad s^{-1}. Thus the scales of the turbulent motions are about 10–100 times smaller than the scales of the original turbulence, since overturning scales in the microstructure regions measured by Gregg (1980) using the method of Thorpe (1977) are 1–8 m, indicating previous vertical turbulent overturning on scales at least this large.

3. Estimates of space-time average dissipation rates and vertical diffusivities

A primary goal of microstructure investigations in the ocean has been to provide estimates of the mean viscous dissipation rate of kinetic energy ϵ, the mean diffusive dissipation rate of temperature variance χ, and to infer vertical diffusivities K from these quantities as a function of depth, location and time. Progress has been slow because the data have been difficult to acquire and the fluid mechanics of stratified turbulence is so poorly understood that the available data have been difficult to interpret.

As mentioned previously, using most oceanic measurements of Cox number with the Osborn-Cox (1972) model for the vertical diffusivity

$$K = DC \tag{5}$$

gives values of K which are very low compared to classical values estimated from mean profiles of temperature or salinity and mean currents. Sverdrup *et al.* (1942) state that K in the ocean is usually in the range $(3–90) \times 10^{-4}$ m^2 s^{-1}. Gregg (1977, 1980) assumes that his vertical profile average C values are representative of the space-time average C for the region sampled, giving K values 3–4 orders of magnitude less using (5). However, this assumption seems questionable in view of the evidence discussed in Section 2 that the microstructure regions observed were fossil rather than actively turbulent, and the general observation that C, χ and ϵ are extremely patchy in space, both in the vertical and horizontal directions.

Because none of the microstructure observed was actively turbulent at the large observed overturning scales, the space-time average C value to be used in (5) must be larger than the space average measured values of 2–59. One measure of the magnitude of previous turbulence activity is the overturning scale of Thorpe (1977) computed by Gregg (1980). The Thorpe displacement ζ is the distance a particle at z must be moved to reorder a vertical temperature profile with overturns into a monotonic profile. Therefore the maximum Thorpe displacement $\zeta_{max} \equiv L_T$ provides a lower bound for the maximum scale of vertical overturning, and a means of estimating the intensity of previous turbulent mixing activity.

Ozmidov (1965) has proposed the following estimate of K from ϵ and N:

$$K = 0.1(\epsilon/N^2). \tag{6}$$

Eq. (6) is similar to an expression derived by Weinstock (1978) except that Weinstock uses a coefficient of 0.8 rather than 0.1.

By comparison of buoyancy, inertial and viscous forces on an overturning eddy, Gibson (1980) shows that the criterion for the existence of turbulence in a stratified fluid is

$$L_R > L > L_K, \tag{7}$$

where L is the vertical scale of the eddy. From the universal turbulent velocity spectrum and a computation similar to that for the scalar spectrum, the wavelength of the smallest overturning eddy is about $15L_K$. From a critical Richardson number criterion, Gibson (1980) infers that in order for turbulence to exist in a stratified fluid $\gamma \geqslant 5.5N$. Combining these criteria and (7) gives a minimum overturning wavelength in stratified fluids of about $1.2L_R$. Therefore a more precise criterion for the existence of turbulence in a stratified fluid is

$$1.2L_R \geqslant \lambda \geqslant 15L_K\,, \tag{8a}$$

where λ is the wavelength of possible overturning turbulence motions.

Stillinger (1981) has recently provided experimental evidence to compare with the theoretical constants of Eq. (8a) as well as other aspects of the Gibson (1980) fossil turbulence model. Using a constant temperature heated x-film anemometer, velocity components and dissipation rates were measured in a classical grid turbulence flow field. Various N values were provided by salt stratification. Density fluctuations were measured with a small conductivity probe. The vertical density flux $\overline{\rho' w'}$ was measured, where primes indicate fluctuations of the density ρ and vertical velocity w. Taking $\overline{\rho' w'} = 0$ as the criterion for non-turbulence Stillinger found $\gamma \geqslant 4.9N$ as the criterion for active turbulence, which is in close agreement with $\gamma \geqslant 5.5N$ derived by Gibson (1980). Stillinger (1981) also measured an overturning scale $\lambda' = 2\rho'_{\rm rms}(\partial\bar{\rho}/\partial z)^{-1}$, similar to the Thorpe overturning scale L_T discussed in the following. Departures of λ' from the nonstratified downstream growth law were observed at about $1.4L_R$ for three values of N. Therefore in terms of λ' the criterion for active turbulence indicated by Stillinger (1981) is

$$1.4L_R \geqslant \lambda' \geqslant 15.4L_K\,, \tag{8b}$$

which is in good agreement with (8a).

From (8a) and (8b) it appears that $\lambda \approx \lambda'$ for active turbulence unaffected by buoyancy. The Thorpe scale L_T is probably about half the overturning wavelength at fossilization and will decrease slowly with time thereafter. Stillinger (1981) found decreasing λ' values after the beginning of fossilization ($\lambda' = 1.4L_{R_0}$) and even after all turbulence had ceased ($1.4L_R < 15.4L_K$), suggesting that the internal wave motions of fossil vorticity turbulence continue to mix the fossil salinity turbulence (since $\rho'_{\rm rms}$ decreases) as predicted by the Gibson (1980) theory. Restratification density fluxes $\overline{\rho' w'}$ were zero or only slightly negative after the turbulence had been damped, a property of fossil turbulence also predicted by the Gibson (1980) fossil turbulence model.

If we identify $1.2L_R/2$ as the maximum Thorpe displacement L_T, we find from (6) that

$$K = 0.29L_T^2N, \tag{9a}$$

where

$$\epsilon_0 = 2.8L_T^2N^3. \tag{9b}$$

Gregg (1980) gives values of L_T as large as 7.5 m within his most active patch 30 m thick [shown in Fig. 15b of Gregg (1980) and in Fig. 3d of this paper]. This implies a local K value of about 800×10^{-4} m^2 s^{-1}, a factor of 10–1000 times larger than the canonical estimates. From (9b), ϵ_0 is about 2×10^{-5} m^2 s^{-3}, representing the dissipation rate necessary to produce the largest overturns which persist. The patch occupies about 30 m out of a total record length of 3063 m from the 22 records below 200 m depth obtained in the *Aries* 9, *Tasaday* 1 and *Tasaday* 17 expeditions to 20°N, 155°W, or about 1% of the data. If an average of only about 0.1% of the North Pacific thermocline were occupied by active turbulent patches with this stength, the space-time average K due to turbulence would be within the range of canonical values.

Clearly the patch is fossil-temperature turbulence at the largest scales. If the patch were actively turbulent with $\epsilon = 2 \times 10^{-5}$ m^2 s^{-3}, the zero gradient separation distance L, given by

$$L = \frac{\pi}{0.4}\left(\frac{D}{\gamma}\right)^{1/2} = \frac{\pi}{0.4}\left(\frac{D^2\nu}{\epsilon}\right)^{1/4}, \tag{10}$$

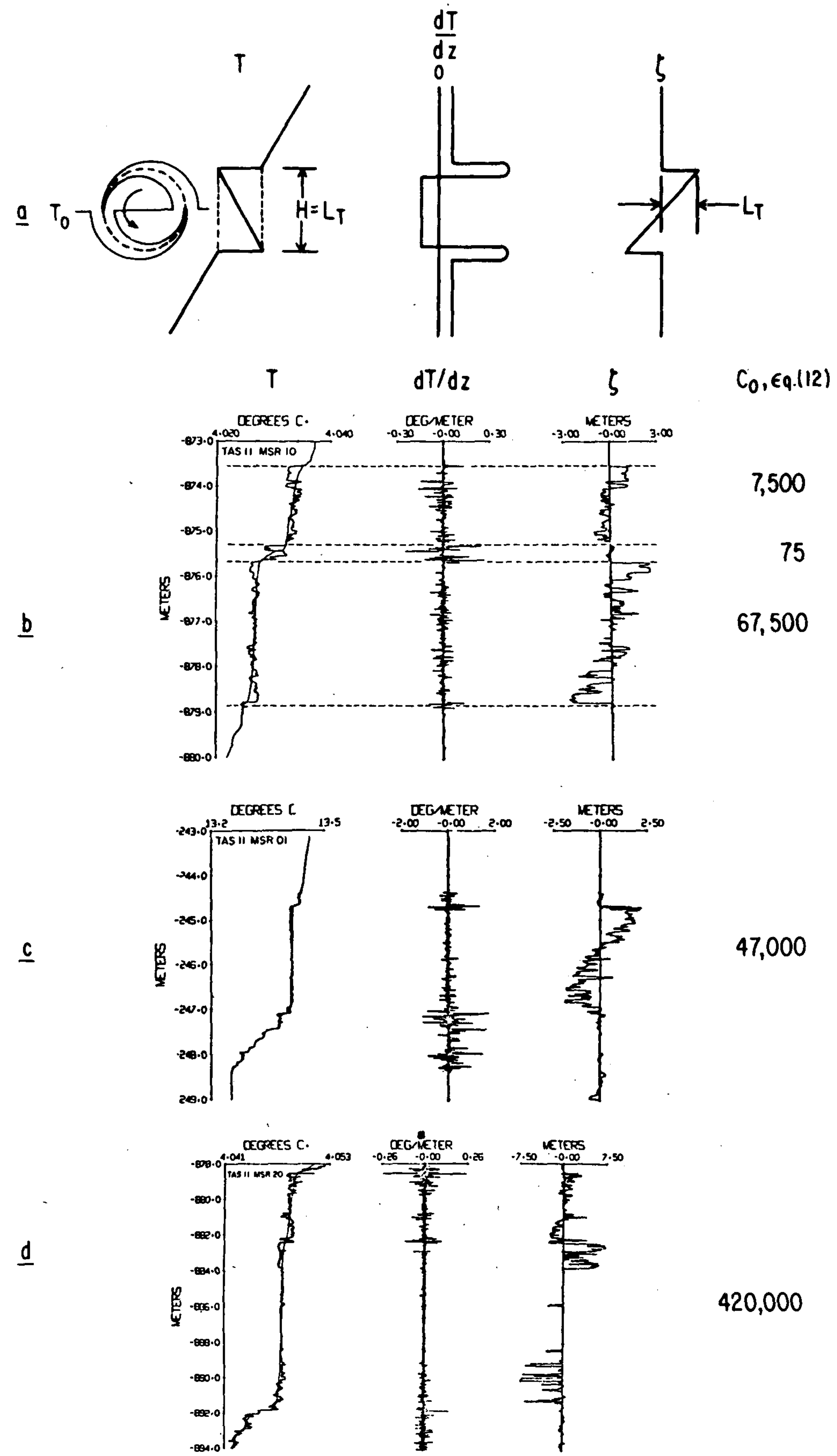
is only 0.1 cm, much smaller than could be measured with the MSR temperature sensors and smaller than the observed separations of 1.5–5 cm. The rate of strain $\gamma = (\epsilon/\nu)^{1/2}$ would be about 800 times larger than N.

The persistence time of such a fossil-temperature turbulence feature is not clear, but it may be very long. An upper bound is the diffusion time scale of a temperature feature of scale L_T of L_T^2/D, which is 12 years for the $L_T = 7.5$ m patch. The actual persistence time of the fossil will be much less because ambient shears and internal straining due to wave motions and gravity will enhance dispersion of the patch and the mixing of the microstructure. The fossil temperature turbulence lifetime presumably ends when the last zero-gradient point produced by the initial active turbulence event is annihilated by molecular diffusion. As shown previously, the effect of gravity on zero-gradient point evolution diminishes as their N' values decrease. Gregg's (1980) result that typically 20–25% of his profiles are occupied by centimeter-scale inversions (p. 941) suggests a long persistence time, since active turbulence must occupy a much smaller volume fraction of the fluid.

We may estimate the Cox number which existed at the time when the patch was actively turbulent using an expression from Gibson (1980), i.e.,

$$\epsilon_0 = 13DC_0N^2, \tag{11}$$

where the subscripts indicate that the turbulence is at the point of fossilization; that is, the inertial forces of the largest eddies are equal to the buoyancy forces imposed by the ambient stratification N. Substituting



T
dT/dz
0
ζ
a
T0
H=LT
LT
T
dT/dz
ζ
C0, Eq.(12)
b
DEGREES C.
4.020
4.040
DEG/METER
-0.30
-0.00
0.30
METERS
-3.00
-0.00
3.00
TAS II MSR 10
METERS
-873.0
-874.0
-875.0
-876.0
-877.0
-878.0
-879.0
-880.0
7,500
75
67,500
c
DEGREES C
13.2
13.5
DEG/METER
-2.00
-0.00
2.00
METERS
-2.50
-0.00
2.50
TAS II MSR 01
METERS
-243.0
-244.0
-245.0
-246.0
-247.0
-248.0
-249.0
47,000
d
DEGREES C.
4.041
4.053
DEG/METER
-0.26
-0.00
0.26
METERS
-7.50
-0.00
7.50
TAS II MSR 20
METERS
-878.0
-880.0
-882.0
-884.0
-886.0
-888.0
-890.0
-892.0
-894.0
420,000

$L_{R_0} = (\epsilon_0/N^3)^{1/2}$ and $L_T = (1.2/2)L_{R_0}$ in (11) and rearranging gives

$$C_0 = 0.21 L_T^2 N/D = 0.21(L_T/L_{BF})^2, \qquad (12)$$

where $L_{BF} = (D/N)^{1/2}$ is the characteristic microscale of fossil scalar turbulence with diffusivity D, defined by Gibson (1980). The Cox number for the patch with $L_T = 7.5$ m is about 420 000 from (12).

It is interesting to note that combining (11) and (5) gives

$$K = 0.08 \frac{\epsilon_0}{N^2}, \qquad (13)$$

which is remarkably similar to the Ozmidov expression in Eq. (6). Gibson (1980) defines a turbulence activity parameter A_T, which is related to the dissipation rate ϵ of the small-scale turbulence and the dissipation rate ϵ_0 at fossilization by

$$\epsilon = \epsilon_0 A_T^2. \qquad (14)$$

Combining (14) and (13) shows that Ozmidov's expression should be valid under conditions of active turbulence near fossilization, where $A_T \approx 1$. The large coefficient of 0.8 proposed by Weinstock (1978) implies $A_T \approx 0.3$; that is, turbulence which is slightly affected by stratification.

An estimate of the age of a fossil patch of microstructure can be obtained from the ratio of observed to original Cox numbers C/C_0. Using the Gibson (1980) model, the Cox number decreases by a factor of about $(5.5N/\gamma_0)^{2/3}$ in the first Väisälä period during fossilization, but by factors of about $[1 - (N/\gamma_0)^{2/3}]$ for subsequent Väisälä periods. Thus in the $L_T = 7.5$ m patch, the Cox number should decrease to 14 600 in the first 21 min, and by factors of about 0.988 for each 21 min period thereafter until C = 8 when the patch was observed. This gives a possible fossil age of about 622 Väisälä periods or 9.1 days, assuming γ during the period was N and that somehow the patch is not dispersed by ambient shears.

Most other patches observed by Gregg (1980) have smaller L_T values, from 0.5 to 3 m. The corresponding range of C_0 values from (12) is from 7000 to 250 000, and the range of K values from (9a) is $(3.6\text{–}130) \times 10^{-4}$ m² s⁻¹.

A schematic pattern of T, dT/dz and ζ for an overturning event in a uniform temperature gradient is shown in Fig. 3a. A fluid cylinder of diameter H rotates 180° about a horizontal axis. Measured displacement scales in three microstructure patches presented by Gregg (1980) are shown for comparison in Figs. 3b–3d. Maximum $|\zeta|$ values L_T correspond to the patch thickness H. Note that the pattern shown in Fig. 3a differs from that presented by Gregg (1980, Fig. 11): Gregg's schematic profile depicts an *interchange* of two layers rather than an overturn, with $L_T = (1/2)H$ rather than $L_T = H$.

An actual turbulent overturn at high Reynolds number with internal mixing should produce a jagged version of the Z-shaped ζ profile of Fig. 3a, but the maximum ζ may not differ much from the original $L_T = H$ value with time if the mixing is uniform. Such jagged Z-shaped ζ profiles are shown in Figs. 3b, 3c and 3d, with patch height $H \approx L_T$. The same backward Z temperature profiles, with maximum dT/dz values at top and bottom, are also observed, strongly suggesting these microstructure patches were produced by overturning turbulence as proposed by Gregg. However, Cox numbers implied by the large vertical scales of the overturns using (12) are several orders of magnitude larger than the measured values, consistent with the previous evidence from the zero crossing scales that the microstructure is fossil turbulence in an advanced state of decay.

Recent computations by Dillon and Caldwell (personal communication) show an approximate equality between the rms Thorpe displacement $\zeta_{\rm rms}$ and $(L_R)_{\rm interior} = (\epsilon/N_i^3)^{1/2}$ for vertical temperature profiles through microstructure patches observed during MILE, where N_i is the Väisälä frequency measured within the interior of the patch. However, since $L_T \equiv \zeta_{\max} > \zeta_{\rm rms}$ and $L_R \equiv (\epsilon/N_e^3)^{1/2} \gg (\epsilon/N_i^3)^{1/2}$ if $N_i \ll N_e$, then L_T is still much larger than the buoyancy scale $(L_R)_{\rm external}$ based on the external Väisälä frequency N_e of the fluid averaged over a vertical scale larger than the patch, and such patches would still appear to be fossil-temperature turbulence. This is because actively turbulent fluid with eddy sizes comparable to the vertical scale of the microstructure patch must overcome buoyancy forces determined by the external stratification. The observed correlation $\zeta_{\rm rms} \approx (L_R)_{\rm interior}$ is very interesting, but seems to imply a correlation in the degree of fossilization for the observed patches rather than a lack of fossil turbulence.

Varying degrees of uncertainty should be attached to the preceding expressions. The criterion for active

FIG. 3. Ideal versus actual overturning patterns. (a) Schematic of temperature, temperature gradient and Thorpe displacement scale ζ profiles for a vertical overturn produced by the solid body 180° rotation of a fluid cylinder of diameter H about a horizontal axis. Note that $H = L_T$, the maximum $|\zeta|$. (b) Fig. 17, p. 234, Gregg (1980). The most active section of TAS 11 MSR 10. The light line of the T profile is the monotonic profile produced by the Thorpe algorithm. The T, dT/dz and ζ patterns suggest turbulent overturning on vertical scales up to 3 m, giving $C_0 = 68\,000$ from Eq. (12). (c) Fig. 18, p. 935, Gregg (1980), from TAS 11 MSR 01. $C_0 = 47\,000$. The salinity profile could permit salt fingering for this profile, but is stable for the profiles in 3b and 3d. (d) Fig. 20a, p. 936, Gregg (1980), from TAS 11 MSR 20. $C_0 = 420\,000$.

turbulence in terms of the rate of strain is about $\gamma \geqslant (5 \pm 1)N$, or in terms of the wavelength $(1.2 \pm 0.2)L_R \geqslant \lambda \geqslant (15 \pm 3)L_K$. The proportionality constant in $L_T \leqslant (0.6\text{–}1.2)L_{R_0}$ is poorly known and should be determined experimentally with high priority since it enters as the square in (9a), (9b) and (12). Thus $K = (0.07\text{–}0.29)L_T{}^2N$, $\epsilon_0 = (0.7\text{–}2.8)L_T{}^2N^3$ and $C_0 = (0.05\text{–}0.21)L_T{}^2N/D$. The zero-gradient separation length should be within about $\pm 20\%$ of the value given by (10). The uncertainty in (11) is about $\epsilon = (13 \pm 2)DC_0N^2$. The estimate of the fossil age for the 7.5 m overturn requires several assumptions, particularly an estimate of the rate of strain during the evolution of the microstructure. If the rate of strain is $5.5N$ rather than N as assumed, the estimate of fossil age decreases from 9 to 3 days.

4. Discussion of definitions

As indicated in the last section, the patches of multiple zero gradients were almost certainly produced by turbulence, but at a previous time since the fluid containing the microstructure is no longer actively turbulent. Therefore, the microstructure may be classified as fossil-temperature turbulence using the definition of Gibson (1980). The distances separating points of zero temperature gradient are assumed to reflect a local equilibrium between molecular diffusion and the local rate of straining of the fluid, as in Gibson (1968) and Batchelor (1959).

The definition of fossil turbulence used in this paper and by Gibson (1980) is different than that used by Gregg (1980). Gregg (1980) states that the microstructure patches are "active" and not "decayed fossils" of previous events. This is claimed because the zero-crossing scale of 5 cm is less than the diffusion length in a Väisälä period "if the temperature fluctuations are no longer accompanied by velocity fluctuations over corresponding scales." "Decayed fossils" therefore apparently means a microstructure region produced by turbulence in which the velocity field has completely decayed and is identically zero. Clearly it is impossible for the velocity field to be zero in a density microstructure field with a nonzero gravitational force. Even if the density fluctuations were zero, the boundaries of all microstructure patches in the real ocean will be moved by internal waves and other motions; hence, "decayed fossils" cannot exist in the ocean by this definition. The fossil turbulence observed by Nasmyth (1970) and others using towed bodies and submarines refers to regions where turbulent velocity has been damped by stratification leaving only the ambient oceanic motions which are dramatically less. Ambient motions at microstructure scales are generally below the noise levels of the towed body velocity sensors, but were not assumed to be zero. Nasmyth (1970, p. 28) recognized the possibility of low-level turbulence below his threshold of detection in apparently fossil turbulence regions. He proposed (p. 35) that a progressive drooping would occur at the high-wavenumber end of the temperature spectrum as buoyancy damps turbulence to produce fossil turbulence, but like Gregg (1980) he did not recognize that the local strain rate of internal wave motions (etc.) would limit the droop to a new equilibrium wavenumber of order $(N/D)^{1/2}$, as described by Gibson (1980).

The definitions of turbulence used by Gibson (1980) and by Gregg (1980) also appear to be different. In this paper and in Gibson (1980) turbulence is defined as a random, isotropic, eddy-like motion in which inertial forces dominate both buoyancy and viscous forces according to a criterion such as (7) (or the equivalent Monin-Obukhov buoyancy flux criterion used in boundary layer flows). The probability laws describing such velocity fields should be universally similar when normalized by Kolmogoroff length and time scales, even in a stratified fluid. The terms "active," "isotropic," "3-D," and "overturning" turbulence are synonymous.

Gregg (1980, pp. 926, 927) refers to turbulence as "small scale random velocity fluctuations not necessarily having the 'universal' spectral forms." However, it is questionable whether any such nonuniversal velocity fluctuations exist when the conditions of (7) are satisfied. Kolmogoroff universal similarity has now been demonstrated in a large variety of laboratory, atmospheric and oceanic flows, with and without stratification, for a variety of statistical parameters including spectra. Significant departures from universal similarity have been sought by a large number of experimental studies, but have not been found.

According to the fossil turbulence theory of Gibson (1980) the buoyancy-dominated velocity field left after the turbulence has been damped by stratification is also classified as a form of fossil turbulence; that is, fossil vorticity turbulence. Although this motion is random, rotational and strongly nonlinear, it is qualitatively quite different from the overturning turbulence from which its energy derives and should not be classified as turbulence but as a form of internal wave. The smallest scales of the temperature microstructure will be maintained in a state of local equilibrium by the rate of strain of this internal wave field with length scale $(D/\gamma)^{1/2}$, where $\gamma < 5.5N$, just as the turbulence maintains a local equilibrium with length scale $(D/\gamma)^{1/2}$ with $\gamma > 5.5N$. As long as the zero-gradient points of the fossil-temperature turbulence persist, they should serve as an indicator of the local rate of strain (and hydrodynamic state) of the ambient velocity field in which the microstructure is embedded. It is not necessary that the velocity be turbulent. It is also not necessary that the small-scale velocity field be nonturbulent for the larger scale microstructure to be fossil turbulence. Fluctuations of any hydrophysical field produced by turbulence in a fluid which is no longer turbulent *at the scale of the fluctuation* is defined as a form of fossil turbulence.

5. Summary and conclusions

A reinterpretation of the microstructure data of Gregg (1977, 1980) indicates a dramatically higher level of turbulent mixing and diffusion. Rather than assuming the microstructure is actively turbulent because it contains density inversions, the scrambled temperature field is compared to the fossil-turbulence model of Gibson (1980). Overturning scales of the microstructure patches analyzed by Gregg (1980) are much too large, 0.5 to 7.5 m, to be produced by turbulence with the very low rates of strain indicated by the measured zero-gradient separation scales of 1.5–5 cm. Such separation scales indicate that the fluid motion is probably not turbulent at all, but is a combination of nearly saturated internal waves and a small-scale cellular motion associated with density inversions in zero-gradient point configurations driven by gravity. Zero-gradient points produced by turbulence in a stratified fluid are shown to reach a quasi-equilibrium state geometrically stabilized by molecular diffusion and viscosity, and will not produce turbulence by gravitational instability as assumed by Gregg (1980).

Since zero-gradient points are a characteristic signature of turbulence which cannot be produced by laminar flows or internal waves (and ignoring double diffusion instabilities) the patches of multiple zero crossings are interpreted as fossil-temperature turbulence. Using the measured Thorpe overturning scales and the Gibson (1980) results, the patch with largest over-turning scale observed by Gregg (1980) was found to have a previous Cox number of at least 4.2×10^5, several orders of magnitude larger than the Cox number measured at the time of observation. A possible age of this fossil patch was estimated to be about 9 days by a rather uncertain model.

Previous Cox numbers inferred from other patches are in the range $(2\text{–}70) \times 10^3$. Local vertical diffusivities for the patches range from $(4\text{–}130) \times 10^{-4}$ $m^2\ s^{-1}$. It is difficult to estimate the space-time average values without more information about the space-time statistics of the patches, but it seems clear that the measured microstructure is not inconsistent with classical K values as large as $10^{-4}\ m^2\ s^{-1}$ due to turbulent mixing.

It also seems clear that the ocean turbulence process has been severely undersampled. An adequate data set should contain some microstructure regions in their original state of active turbulence. However, from the estimated large ϵ_0 values, the Gregg (1977) thermistor frequency response would have been much too small to detect the larger Cox numbers, even if the measured patches were in their actively turbulent phase.

As concluded by Gregg (1980), much more needs to be known about the physical processes of stratified turbulence before its signature can be properly identified in the ocean. The possibility that information about previous mixing activity may be preserved by the fossil-turbulence structure is exciting since adequate data sets will probably not be achieved in most ocean regions for many years.

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