Yes, We Have No Abyssal Mixing

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Abstract. A number of indirect means for inferring turbulent mixing have been developed over the past two decades. In particular, a combination of internal wave/wave interaction theory and observations has suggested a relationship between the turbulence dissipation rate $\varepsilon$ and levels of the vertical wavenumber spectra for shear and strain. Here, we apply a parameterization based on finescale strain variance $<\xi_z^2>$ to hydrography collected during several WOCE and ACCE cruises in the Southern Ocean, eastern S. Pacific, eastern polar gyre of the N. Atlantic and tropical Indian. Inferred average eddy diffusivities $K$ are consistently $O(0.1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1})$.

1. Introduction

Turbulent mixing is important for water-mass modification, maintaining the ocean’s stratification and driving the global meridional thermohaline circulation (Munk and Wunsch 1998; Wunsch and Ferrari, 2003; Ganachaud and Wunsch, 2000). Microstructure measurements (Gregg, 1987) and tracer-release experiments (Ledwell et al., 1993; 2000) have established that mixing in the stratified midlatitude ocean interior away from boundaries is weak $O(0.1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1})$. This has led to speculation that elevated mixing confined to boundaries drives the ocean thermohaline circulation. Weak turbulence is found over flat abyssal basins and gentle smooth slopes (Toole et al., 1994; Kunze and Sanford, 1996; Polzin et al., 1997). However, over rough topography such as seamounts (Nabatov and Ozmidov 1988; Kunze and Toole, 1997; Lueck and Mudge, 1997), ridges (Polzin et al., 1997; Althaus et al., 2003; Klymak et al., 2002), canyons (St. Laurent et al., 2001; Carter and Gregg, 2002) and in hydraulically controlled passages between deep basins (Roemmich et al., 1996; Polzin et al., 1996; Ferron et al., 1998), turbulent mixing has been inferred to be 100-1000 times ocean interior values.

Most estimates to date have been based on specialized shear and temperature microstructure data with $O(1 \text{ cm})$ resolution (e.g., Oakey 1982). Such measurements are usually made in localized experiments focussed on dynamical processes.

A number of indirect methods are now available for estimating turbulent dissipation rates and mixing more broadly. These range from $O(1 \text{ m})$ outer turbulence scales to $O(10 \text{ m})$ finescale internal waves. At outer turbulence scales, Thorpe overturn scales (Dillon, 1982; Galbraith and Kelley, 1996) represent potential energy available for turbulence. Similarly, unstable Richardson numbers (Kunze et al., 1990; Peters et al., 1995; Polzin, 1996) represent the kinetic energy in internal wave shear available for turbulence production through shear instability.

Nonlinear internal wave/wave interaction theories (McComas and Müller, 1981; Henyey et al., 1986) have led to parameterizations for the turbulent dissipation rate $\varepsilon$ depending on internal wave shear and strain levels. These are based on the rate of transfer of energy through the internal wave vertical wavenumber spectrum toward small scales and turbulence production. In this paper, we will apply one of these parameterizations to full-depth CTD profiles from WOCE and ACCE in selected regions. The parameterizations were originally formulated for dissipation rate $\varepsilon$. Here, we will express the scalings for diapycnal eddy diffusivities $K = \gamma \varepsilon N^2$ (Osborn, 1980) assuming a ‘mixing efficiency’ $\gamma = 0.2$.

2. Finescale Parameterizations

The scalings were first validated against oceanic data by Gregg (1989) who found that a semi-empirical scaling based on 10-m first-difference shear

$$K = K_0 \frac{<V_x^2>^2}{<\xi_z^2>^2}, \quad (1)$$

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reproduced observed average profiles to within a factor of two for four midlatitude data sets, where $K_0 = 0.05 \times 10^{-4}$ m$^2$ s$^{-1}$ and $K_0 = 0.05 \times 10^{-4}$ m$^2$ s$^{-1}$ is from the Garrett and Munk model spectrum (Gregg and Kunze, 1991) treated in the same way as the data. After Gargett (1990) pointed out that 10-m first-difference shear would underestimate the shear spectral level $E$ when it was high since $E_{k_z} = \text{constant}$ (rolloff wavelength $\lambda_c = 2\pi/k_c \sim 10$ m for GM levels), a vertical wavenumber spectral method was developed (Gregg and Kunze, 1991; Kunze et al., 1992).

Finding that (1) underestimated turbulence in the high-frequency internal wave field at the edge of the Yermak Plateau, Wijesekera et al. (1993) reported that an alternative scaling based on finescale strain $\xi_z$ of similar form

$$K = K_0 \frac{\langle \xi_z^2 \rangle}{\langle \xi_z^2 \rangle}$$

(2)

better explained their data. Mauritzen et al. (2002) used the same strain-based approach to argue for enhanced mixing over the Mid-Atlantic Ridge.

Stronger turbulence is anticipated in wave fields of higher-than-GM frequency-content. Henyey (1991) pointed out that the rate at which energy is transferred to small scales depends on the internal wave field’s aspect ratio $k_h/k_z$ which is related to the frequency-content $\omega$ and hence the shear/strain variance ratio $R_\omega = \langle V_z^2 \rangle / (N^2 \langle \xi_z^2 \rangle)$ ($= 3$ for GM model). Subsequent parameterizations have been formulated in terms of both shear and strain variances. Below, we express these in terms of strain variances and the shear/strain ratio. Polzin et al. (1995) found that

$$K = K_1 \frac{\langle \xi_z^2 \rangle}{\langle \xi_z^2 \rangle} f_1(R_\omega)$$

(3)

exhibited better skill than (1), collapsing observations to within a factor of two for data sets with lower- and higher-than-GM frequency-content, where $K_1 = 0.07 \times 10^{-4}$ m$^2$ s$^{-1}$ and

$$f_1(R_\omega) = \frac{R_\omega + 1}{\sqrt{\frac{R_\omega[1-R_\omega + \sqrt{(1-R_\omega)^2 + 8R_\omega^2f^2/N^2}]}{GM R_\omega[1-GM R_\omega + \sqrt{(1-GM R_\omega)^2 + 8GM R_\omega^2f^2/N^2}]}}$$

Based on ray-tracing numerical simulations, Sun and Kunze (1999) derived a similar relation

$$K = K_2 \frac{\langle \xi_z^2 \rangle}{\langle \xi_z^2 \rangle} g(R_\omega)$$

(4)

where $K_2 = 0.0012 \times 10^{-4}$ m$^2$ s$^{-1}$,

$$g(R_\omega) = \frac{(R_\omega + 1)^2}{g(R_\omega) + f_1(R_\omega)}$$

and $r_1 = 10$. Most recently, Gregg et al. (2003) have synthesized the work of Henyey (1991) and Polzin et al. (1995) to explain the observed reduction of turbulence at the equator, separating dependence on the shear/strain ratio $R_\omega$ and $f/N$ into two different functions

$$K = K_0 \frac{\langle \xi_z^2 \rangle}{\langle \xi_z^2 \rangle} h(R_\omega) j(f/N)$$

(5)

where

$$h(R_\omega) = \frac{3R_\omega (R_\omega + 1)}{4GM R_\omega^2} \sqrt{\frac{2}{R_\omega - 1}}$$

and

$$j(f/N) = \frac{f_30 farcosh(N/f)}{f_30 farcosh(N_0/f_30)}$$

where $f_{30} = f(30^\circ)$ and $N_0 = 5.2 \times 10^{-3}$ rad s$^{-1}$.

In the following very preliminary results, we present parameterized eddy diffusivities $K_1$ (5) based on CTD strain assuming a shear/strain variance ratio of either $R_\omega = 3$ (GM value) or the semidiurnal value ($\omega f + f^2$) which is less (greater) than 3 equatorward (poleward) of 45º. Modifications for higher and lower shear/strain ratio $R_\omega$ (corresponding to lower and higher frequency-content respectively) can be deduced from Fig. 1.
Figure 1. Functional dependences on shear/strain ratio $R_{\omega}$ in the Polzin et al. (1995), Sun and Kunze (1999), and Gregg et al. (2003) parameterizations. The Sun and Kunze dependence $g(R_{\omega})$ has been rescaled to give 1 for a shear/strain ratio of 3 for the purposes of comparison. Midlatitude Coriolis frequency $f(30^\circ)$ and abyssal buoyancy frequency $N = 10^{-3}$ rad s$^{-1}$ have been used for functions $f_1$ and $g$ depending on these quantities. The Polzin et al. and Gregg et al. functional dependences on shear/strain ratio $R_{\omega}$ are indistinguishable. For higher shear/strain variance ratios (excess near-inertial shear), as often found in the ocean, the three functions are similar for $R_{\omega} < 10$, implying enhancement by as much as a factor of five in this range.

3. Data and Methods

CTD profiles are available spanning the world ocean thanks to WOCE. We examine roughly 1000 CTD profiles from four regions where there were accompanying lowered ADCP velocity profiles: the Southern Ocean north of Wilkes Land; the eastern S. Pacific, the eastern polar gyre of the N. Atlantic and the tropical Indian. Because of as yet unresolved difficulties with interpreting the lowered ADCP data, only the CTD data will be presented here.

The CTD profiles have 1- or 2-m vertical resolution and uncertainties of ~ 0.0003°C for temperature and 0.0003 psu for salinity, consistent with WOCE guidelines -- except in some of the tropical Indian data where tests of Falmouth Scientific sensors produced higher noise levels.

Strain is estimated both from buoyancy frequency $\xi_z = (N^2 - \ddot{\nu}^2)/\ddot{\nu}^2$ and from potential temperature $\theta \xi_z = (\theta_z - \ddot{\theta}_z)/\ddot{\theta}_z$. The buoyancy frequency estimate is sensitive to salinity noise when stratification is weak, while the temperature-inferred strain can be contaminated by finescale water-mass variability (thermohaline intrusions or interleaving). Both can contain contributions from smallscale subinertial vortices (Polzin et al. 2003). To infer finescale strain variance levels for turbulence parameterization (5), the profiles were first broken into half-overlapping 256-m segments starting from the bottom. The mean stratification $(\bar{N}^2, \ddot{\theta}_z)$ is based on quadratic fits to the individual profile segments. The segments are windowed at both ends with a 10% sin$^2$ taper before fast Fourier-transforming.

Vertical wavenumber spectra for strain $S(\xi_z)(k_z)$ were typically flat at intermediate resolved wavenumbers, becoming slightly red at the highest resolved wavenumbers, consistent with the GM model. At the lowest resolved wavenumbers ($k_z > 150$ m), there was considerable variability that appeared to be associated with depth variability in the background stratification, particularly in the upper pycnocline and near the bottom. To eliminate the effect of variable background stratification on the strain variance estimates, only high wavenumbers were included.

Strain variance levels $<\xi_z^2>$ for (5) were obtained by integrating the spectra from a minimum vertical wavenumber (min $k_z$) corresponding to $\lambda_z = 150$ m out to both

$$<\xi_z^2> = \int_{\text{min } k_z}^{\text{max } k_z} S(\xi_z)(k_z)dk_z = 0.05 \text{ and } 0.2 \quad (6)$$

GM model strain variances were computed over the same wavenumber bands

$$\text{GM }<\xi_z^2> = \frac{\pi E_0 b j_1}{2} \int_{\text{min } k_z}^{\text{max } k_z} k_z^2 dk_z \left( k_z + k_{z*} \right)^2$$

(Gregg and Kunze, 1991). The smaller of the four strain ratios $<\xi_z^2>/\text{GM}<\xi_z^2>$ (buoyancy frequency and potential temperature times two integration limits) was chosen to be plugged into finescale parameterization (5). Shear/strain variance ratios $R_{\omega}$ corresponding to the GM model ($= 3$) and a semi-diurnal frequency $f_1$ are assumed (Fig. 1). For $R_{\omega} = 3$, parameterization (5) returns the same value as parameterization (2) multiplied by the $j(f_1 N)$. Midlatitude parameterizations (3) and (4)
do not vanish as \( f/N \rightarrow 0 \) so will overestimate dissipation rates and eddy diffusivities on the equator.

4. Results

4.1 Southern Ocean

The Southern Ocean data are from two meridional cruise tracks north of Wilkes Land (Fig. 2). One track ran south along 95ºE, then shifting west to cross the eastern edge of Kerguelen Plateau to continue to Antarctica, then north to the southwest corner of Australia along 115ºE during December 1994 (I08). The other ran south of New Zealand and back during January 1996 (P14). The average eddy diffusivity \( \langle K \rangle \) (Fig. 3) is about \( 0.1 \times 10^{-4} \) m\(^2\) s\(^{-1}\) throughout most of the water column (\( z \)). Smaller values of 0.02-0.03 \( \times 10^{-4} \) m\(^2\) s\(^{-1}\) are found in deep basins below 4500 m depth (neutral density \( \sigma_n > 50 \)) where there are few measurements. The diffusivities decrease weakly away from the bottom (\( h \)). Slightly higher diffusivities are predicted for a semidiurnal than a GM shear/strain ratio because semidiurnal shear/strain variance ratios exceed 3 for most of the profiles. Vertically integrated inferred dissipation rates \( \int_{-h}^{0} \varepsilon \, dz \) (Fig. 2) are everywhere less than the deep-ocean average surface tidal dissipation of 2 mW m\(^{-2}\) inferred by Egbert and Ray (2001) but bracket the GM dissipation \( \int_{-h}^{0} 5K_{GM} N^2(z) \, dz = 0.4 \) mW m\(^{-2}\) where \( K_{GM} = 0.07 \times 10^{-4} \) m\(^2\) s\(^{-1}\).

4.2 East Pacific Rise

Data over the East Pacific Rise in the eastern equatorial and S. Pacific is from two meridional cruise legs (Fig. 4). The east track along 90ºW that crosses the equator east of the Galapagos was collected during February 1993 (P19) while the west track along 103 ºW that extends over 30-70ºS was collected during February 1994 (P18). Average diffusivities \( K \) are about \( 0.03 \times 10^{-4} \) m\(^2\) s\(^{-1}\) in the water column (Fig. 5). For semidiurnal shear/strain ratios, higher values
are found in deep basins below 4000 m (neutral density greater than 45), which are mostly at high latitude. Vertically integrated dissipation rates (Fig. 4) are again everywhere less the the Egbert and Ray average deep-ocean surface tidal dissipation but comparable to the GM dissipation. Note higher dissipations over smallscale topographic features ESE of the Galapagos and around 26°S, 90°W.

**Figure 4.** As in Fig. 2 but in the eastern S. Pacific over the East Pacific Rise. Note the high dissipation at 26°S, 90°W over a narrow ridge.

### 4.3 Eastern Polar Gyre of the N. Atlantic

Data in the eastern polar gyre of the N. Atlantic was collected on three separate cruises during November 1996, May 1997 and October 1997 sampling roughly the same track (Fig. 6, A24). In depth (z) space, average diffusivities are several times $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$ in the upper 2000 m (Fig. 7) as found by Polzin et al. (2002), decreasing to $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$ at depth. In height above bottom (h) space, the diffusivities decrease away from the bottom from several times $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$ to $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$. Vertically integrated dissipations (Fig. 6) are noticeably higher over the Mid-Atlantic Ridge between 40 and 45°N than the neighboring basins, but less obviously so further north.

**Figure 5.** As in Fig. 3 but in the S. Pacific over the East Pacific Rise. Deeper waters are found at higher latitudes.

**Figure 6.** As in Fig. 2 but in the eastern polar gyre of the N. Atlantic. Three separate cruises are overplotted. High vertically integrated dissipations are found over the Mid-Atlantic Ridge at 40-45°N.
High dissipation is also inferred on the continental slope southeast of Greenland near the Denmark Strait Outflow plume.

Figure 7. As in Fig. 3 but for the eastern side of the N. Atlantic polar gyre.

4.4 Tropical Indian

Data from the tropical Indian Ocean comes from 3 legs: I01W along the Gulf of Aden and across the mouth of the Arabian Sea during September 1995, I01E across the Bay of Bengal during October 1995 and I10 across the mouth of the Timor Sea from Java-Sumatra to northwest Australia during November 1995 (Fig. 8). Diffusivities are $0.03 \times 10^{-4}$ m$^2$ s$^{-1}$ through most of the water column with slightly higher values of $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$ around 4500-m depth ($z$) and neutral density $\sigma_n = 50$ (Fig. 9). Diffusivities $K$ decrease from $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$ at the bottom to $0.03 \times 10^{-4}$ m$^2$ s$^{-1}$ by 1500-mab (meters above bottom) in height above bottom ($h$) coordinates. The high vertically integrated dissipations found on the western side of the Arabian Sea (Fig. 8) are misleading. They are due to deep mixed-layers and extremely sharp pycnoclines in the uppermost bins, possibly products of the preceding Monsoon, and are not due to internal wave strain. This is an instance of a failure of the scheme outlined in section 3 to isolate the internal wave strain from the background. High dissipations near the Andaman Island ridge on the eastern side of the Bay of Bengal (Fig. 8) are associated with elevated signals near the bottom over rough topography.

Figure 8. As in Fig. 2 but in the tropical Indian along the Gulf of Aden, across the mouths of the Arabian Sea, Bay of Bengal and Timor Sea. High vertically integrated dissipations are evident on the west side of the Arabian Sea.

5. Summary

The latest in a series of parameterizations for turbulent diapycnal eddy diffusivity $K$ (Gregg et al. 2003) was expressed in terms of finescale strain $\xi$ and applied to CTD profiles in four regions of the deep ocean assuming GM and semidiurnal shear/strain ratios. While evidence for high diffusivities ($K > 0.1 \times 10^{-4}$ m$^2$ s$^{-1}$) was found associated with rough topography, average profiles of eddy diffusivity were $0.1 \times 10^{-4}$ m$^2$ s$^{-1}$ or less in all four regions, much less than the canonical $10^{-4}$ m$^2$ s$^{-1}$ (Munk and Wunsch 1998). Vertically integrated dissipation rates were well below deep-ocean surface tidal dissipations inferred from satellite altimetry (Egbert and Ray 2001) but comparable to expected GM levels.
Figure 9. As in Fig. 3 but for the tropical Indian Ocean.

The low diffusivities found here are in sharp contrast to diffusivities inferred from bulk budgets (Munk and Wunsch 1998; Ganachaud and Wunsch 2000). We caution that there are many steps along the way in which error could have been introduced in our preliminary results:

- The strain variance $\langle \xi_z^2 \rangle$ could be underestimated. Vertical wavelengths larger than 150 m were excluded from the variance estimate on the grounds that these appeared to include large signals from the background stratification, particularly in the upper pycnocline and near the bottom. Moreover, of four different strain variances calculated, the one with the smallest GM-normalized ratio was always chosen. We have some confidence that the strains were not underestimated in that their high-wavenumber spectra resembled the GM model in most instances but this remains uncertain.

- The strain parameterization could be underestimating $K$ either because it is inherently flawed, or because the shear/strain ratio is considerably higher than the GM or semidiurnal prescription in much of the ocean. The parameterization has undergone rigorous testing in the open ocean by a number of investigators (Gregg 1989; Polzin et al. 1995; Gregg et al. 2003) and found to reproduce microstructure estimates to within a factor of two. One exception to this was in Monterey Canyon where the parameterizations underestimate turbulent dissipation by a factor of 30 (Kunze et al. 2002; Carter and Gregg 2002). This is a very different environment than the deep ocean. High shear/strain ratios $\sim O(10)$ are often observed so the latter possibility cannot be eliminated.

- The sampling included here may be inadequate to capture hotspots. Elevated turbulence has been associated with (i) generation by winds and tides and (ii) interactions with subinertial flow and topography. These processes are both spatially and temporally intermittent. At sites of internal wave generation, often only a small fraction of the energy is lost locally to turbulence. The bulk of the energy propagates away to be dissipated elsewhere, possibly not in the deep water sampled here. Wind generation is associated with winter storms (Alford 2001). Except for two of the N. Atlantic cruises, most of the data was collected during seasons of relative calm (rough weather was encountered during P14-15S). Finally, Bryden and Nurser (2003) have implicated hydraulically controlled passages between deep basins as the dominant sites of abyssal mixing. With the exception of a few of the N. Atlantic profiles, the data included here did not sample in such passages. The internal wave/wave interaction parameterizations here are unlikely to be applicable in such a strongly turbulent environment.

- Finally, we should recognize the possibility that the bulk budgets are overestimating mixing. St. Laurent et al. (2001) found that hydrographic inversions constrained by microstructure measurements of diapycnal mixing gave much smaller eddy diffusivities than inversions without such a constraint.

If we believe the strain-based parameterizations, a more complete census of the deep ocean with all WOCE hydrography is a clear next step, either to identify mixing hotspots or show where mixing is not. ARGO float data, which is less biased away from strong storms than shipboard measurements, should also be examined.
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