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Estimating tidally driven mixing in the deep ocean

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[1] Using a parameterization for internal wave energy flux in a hydrodynamic model for the tides, we estimate the global distribution of tidal energy available for enhanced turbulent mixing. A relation for the diffusivity of vertical mixing is formulated for regions where internal tides dissipate their energy as turbulence. We assume that 30 ± 10% of the internal tide energy flux dissipates as turbulence near the site of generation, consistent with an estimate based on microstructure observations from a mid-ocean ridge site. Enhanced levels of mixing are modeled to decay away from topography, in a manner consistent with these observations. Parameterized diffusivities are shown to resemble observed abyssal mixing rates, with estimated uncertainties comparable to standard errors associated with budget and microstructure methods. *INDEX TERMS:* 4568 Oceanography: Physical: Turbulence, diffusion, and mixing processes; 4544 Oceanography: Physical: Internal and inertial waves; 4524 Oceanography: Physical: Fine structure and microstructure. **Citation:** St. Laurent, L. C., H. L. Simmons, and S. R. Jayne, Estimating tidally driven mixing in the deep ocean, *Geophys. Res. Lett.*, *29*(23), 2106, doi:10.1029/2002GL015633, 2002.

1. Introduction

[2] Recent studies [*Munk and Wunsch*, 1998; *Egbert and Ray*, 2001; *Jayne and St. Laurent*, 2001] have implicated the internal tides as a major source of mechanical energy for mixing, providing up to 1 TW of power to the deep ocean. While some studies suggest this amount of energy is sufficient for powering the mixing that closes the meridional overturning cell of the ocean's thermohaline circulation [*Webb and Suginohara*, 2001], other studies suggest that internal tides provide only half of the needed energy [*Munk and Wunsch*, 1998]. This highlights the need for studies examining the internal tide's contribution to mixing.

[3] Turbulent mixing levels in the ocean are quantified as a vertical or diapycnal diffusivity (k_v). Measurements of ocean turbulence [*Gregg*, 1987] and released tracers [*Ledwell et al.*, 1998] have indicated that vast regions of the ocean interior are mixing at a diffusivity level of $k_v = 0.1 \times 10^{-4}$ m^2 s^{-1}. This is regarded as the background level of turbulent mixing in the ocean. Mixing levels near sites of internal tide generation, however, are known to exceed this background value. Microstructure observations have provided direct evidence of internal tide driven turbulence along the continental slopes [*Lien and Gregg*, 2001; *Moum et al.*, 2002], open-ocean seamounts [*Lueck and Mudge*, 1997; *Kunze and Toole*, 1997] and abyssal topography [*Polzin et al.*, 1997; *Ledwell et al.*, 2000; *St. Laurent et al.*, 2001].

2. Parameterization

[4] In the present study, we formulate a scheme for estimating enhanced turbulent mixing rates in regions where internal tides are generated. We build on a parameterization of internal tide energy flux used with a hydrodynamic model for the tides [*Jayne and St. Laurent*, 2001]. *Jayne and St. Laurent* [2001] estimated the energy flux as

$$E \simeq \frac{1}{2}\rho N_b \kappa h^2 u^2. \left[\mathrm{Wm}^{-2}\right] \qquad (1)$$

Here ρ is the reference density of seawater, N_b is the buoyancy frequency along the seafloor, and (κ, h) are the wavenumber and amplitude scales for the topographic roughness. The squared barotropic tidal speed u^2 is determined through solution of the Laplace tidal equations. Further details of the barotropic model and the tidal simulation are given by *Jayne and St. Laurent* [2001].

[5] The energy flux estimates from (1) provide a constraint on the amount of energy available for the enhancement of turbulence near sites of internal tide generation. However, not all of the generated energy flux contributes to local mixing, since some energy radiates away from the generation sites as low-mode waves. We specifically seek the "local dissipation efficiency" (q), quantifying the fraction of energy likely to dissipate locally by turbulent processes.

[6] The mechanisms that aid in dissipating internal tides are discussed by *St. Laurent and Garrett* [2002]. They suggest that enhanced turbulence levels near internal tide generation sites are attributable to the dissipation of high-mode waves. For the mid-ocean ridge topography they examine, they suggest that less than 1/3 of the generated energy flux is dissipated as turbulence close to the generation site. They conclude that most internal tide energy flux is carried by waves with low-mode baroclinic structures. These waves are only weakly influenced by dissipative mechanisms, allowing them to radiate over basin scales.

[7] Direct measurements of turbulent dissipation rates (ϵ) above Mid-Atlantic Ridge topography have been described by *St. Laurent et al.* [2001]. These data were collected over two 4-week sample periods in a region between 12°–22°W longitude and 20°–25°S latitude above fracture zone topography in the Brazil Basin. The average dissipation rate, vertically integrated to a height of 2000 m above the topography, was 1.2 ± 0.2 mW m^{-2}. If this dissipated energy is assumed to be provided by the internal tide, it can be compared to estimates of generated internal tide

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energy flux. *St. Laurent and Garrett* [2002] considered the internal tide energy flux over the region between 8° –20°W longitude and 22° –32°S latitude, and found 4 ± 1 mW m^{-2} as the spatially averaged generated power. Taken together, these dissipation and generation estimates suggest an average local dissipation efficiency of $q = 0.3 \pm 0.1$.

[8] Estimates of radiated and generated energy flux have also been documented for the Hawaiian Ridge. *Ray and Mitchum* [1997] have examined TOPEX/POSEIDON altimetry data for the sea-surface elevation signal associated with the mode-1 and mode-2 internal tides radiated from Hawaii. They found that the signal of low-mode radiation was apparent to distances over 1000 km away from Ridge. *Ray and Mitchum* [1997] estimate that mode-1 accounts for 15 GW. Additionally, *Egbert and Ray* [2001] used TOPEX/POSEIDON altimetry data to estimate the total barotropic to baroclinic tidal conversion. They estimate a total baroclinic conversion of 20 GW at the Hawaiian Ridge. These estimates imply that at least 75% of the generated energy flux is radiated to the far field, leaving $q \leq 0.25$ for the dissipation efficiency. Smaller dissipation efficiencies are likely since other low modes are also radiated to the far field. Indeed, results from the Hawaiian Ocean Mixing Experiment suggest $q \simeq 0.05$ (Eric Kunze, personal communication, 2002).

[9] Motivated by the dissipation and generation estimates from the abyssal Brazil Basin, we propose an initial estimate of the dissipation efficiency of $q = 0.3 \pm 0.1$. We have limited information on which to assess the variability of this parameter for different internal tide generation regions. However, we expect this to be appropriate for the mid-ocean ridge regions similar to the Brazil Basin. For isolated ridges such as Hawaii, $q = 0.3$ may be considered an upper estimate, consistent with the altimetry derived energy levels.

[10] Our model for the turbulent dissipation rate follows as $\epsilon \simeq (q/\rho)E(x, y)F(z)$, where $E(x, y)$ is the time-averaged map of generated internal tide energy flux, and $F(z)$ is the function for the vertical structure of the dissipation, chosen to satisfy energy conservation within an integrated vertical column, $\int_{-H}^{0} F(z)dz = 1$. The parameterization for the turbulent diffusivity follows from the *Osborn* [1980] relation for the mechanical energy budget of turbulence,

$$k_v \simeq \frac{\Gamma q E(x,y) F(z)}{\rho N^2}. \quad [\mathrm{m^2\,s^{-1}}] \tag{2}$$

Here, Γ is related to the mixing efficiency of turbulence, and $\Gamma = 0.2$ is generally used [*Osborn*, 1980]. The diffusivity given in (2) is meant to quantify only the enhanced levels of mixing supported by the dissipation of internal tide energy. A background diffusivity of 0.1×10^{-4} m^2 s^{-1} is still assumed to be sustained in all regions, and must be added to the estimate from (2). Contributions to the diffusivity from other sources of enhanced mixing, such as turbulence in critical layers and frictional boundary layers, could also be added to (2).

[11] In the estimates we present here, our selection of the vertical structure function $F(z)$ in (2) is motivated by turbulence observations from the abyssal ocean [*St. Laurent et al.*, 2001] and the continental slope [*Moum et al.*, 2002]. These observations suggest that the dissipation rate can be roughly modeled as an exponential function that decays away from the topography. Specifically, we use a dissipation profile of the form $\epsilon(z) = \epsilon_0 \exp(-(H+z)/\zeta)$, where ζ is the vertical decay scale, and ϵ_0 is the maximum dissipation rate at the bottom, $z = -H$. This gives

$$F(z) = \frac{e^{-(H+z)/\zeta}}{\zeta(1 - e^{-H/\zeta})}. \tag{3}$$

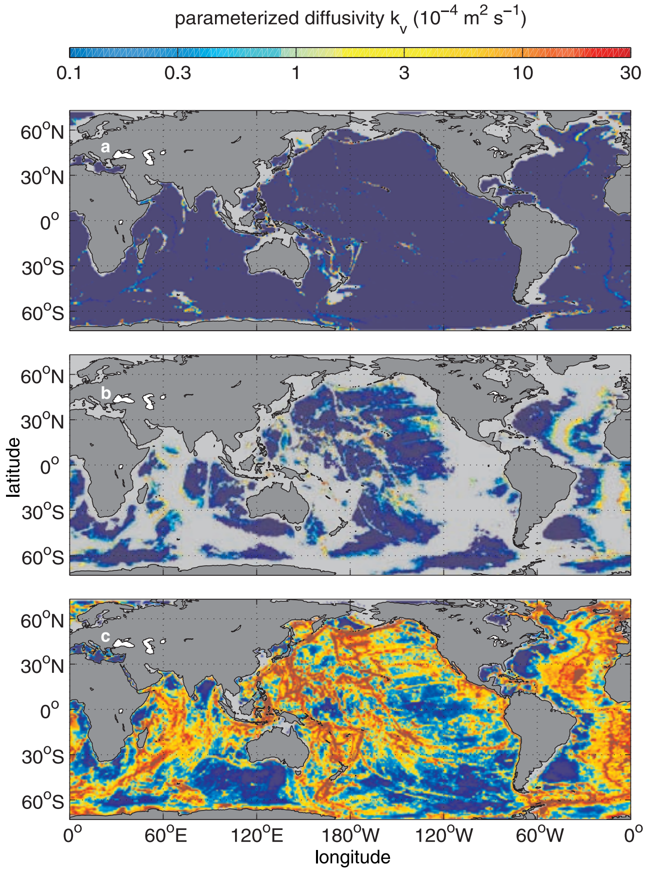

Figure 1. Estimates of turbulent diffusivity along (a) the 1000 m depth level, (b) the 4000 m depth level, and (c) the bottom boundary of the ocean. At all levels, a background diffusivity of 0.1×10^{-4} m^2 s^{-1} was added to the parameterized estimates. Seafloor topography shallower than the given depth level is shaded gray in (a) and (b). In (c), gray shading is used for seafloor regions shallower than 100 m.

3. Results

[12] We have calculated the three dimensional distribution of k_v using estimates of $E(x, y)$ from numerical simulations of the tides [*Jayne and St. Laurent*, 2001], and N^2 calculated from a climatology of oceanic salinity [*Levitus et al.*, 1994] and temperature [*Levitus and Boyer*, 1994]. In (2), we take $q = 0.3$ and $\Gamma = 0.2$. The vertical decay scale in (3) was taken as $\zeta = 500$ m, which is an upper estimate based on turbulence observations [*St. Laurent et al.*, 2001] Estimates from (2) were made for regions where the ocean was deeper than 100 m, which generally excludes the continental shelves. The bathymetry data of *Smith and Sandwell* [1997], version 8.2, were used.

[13] Maps of k_v reveal significant spatial variations (Figure 1). At the 1000 m depth level (Figure 1a), the influence of enhanced mixing due to internal tides is limited to a small number of regions. Sites with diffusivities exceeding 1 ×

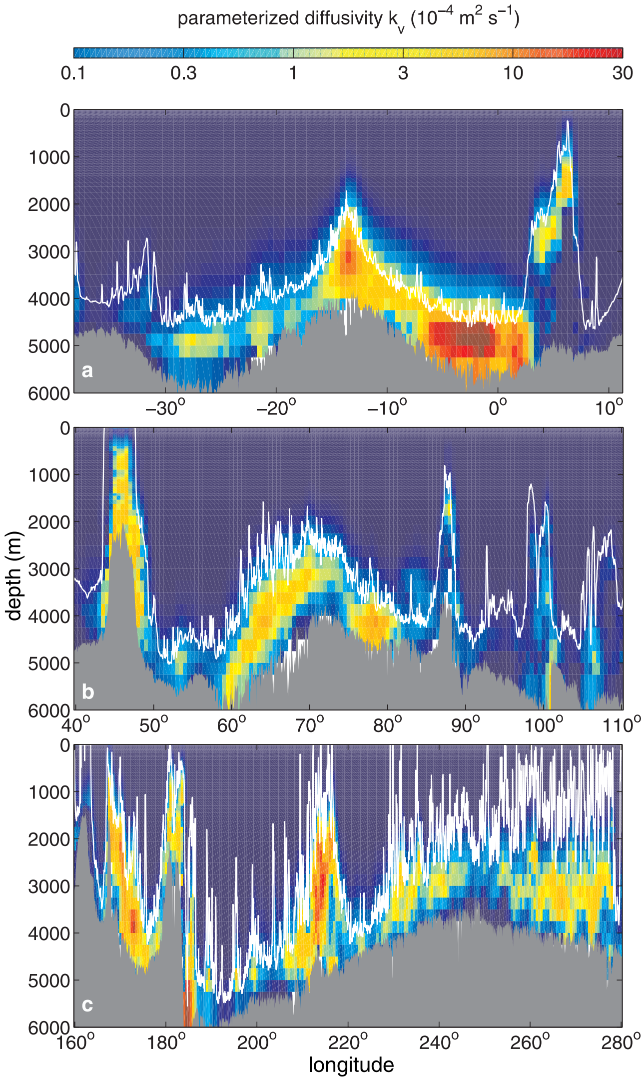

Figure 2. Diffusivity sections averaged between 24°–28°S for the (a) Atlantic, (b) Indian, and (c) Pacific Basins. A background diffusivity of 0.1×10^{-4} m^2 s^{-1} has been added to the parameterized estimates. For each section, the envelope defined by deepest (shaded) and shallowest (white line) bathymetry in the 4° latitude band is indicated.

10^{-4} m^2 s^{-1} comprise only 2% of the total area at this level. These include the Hawaiian Ridge and the Kerguelen Plateau, where maximum estimates of diffusivity are (10–30) × (10^{-4} m^2 s^{-1}. At the 4000 m depth level (Figure 1b), 13% of the map's area shows diffusivities exceeding 1×10^{-4} m^2 s^{-1}. Regions of enhanced mixing include the flanks of the major mid-ocean ridges in the Atlantic and Indian Oceans. The maximum diffusivities estimated from (2) always occur along the seafloor, and these are shown in Figure 1c. Regions with diffusivities exceeding 1×10^{-4} m^2 s^{-1} comprise 60% of the map's area. Diffusivities exceed 10×10^{-4} m^2 s^{-1} over 16% of the map's area, and 30×10^{-4} m^2 s^{-1} over 6%.

[14] Sections of k_v also reveal significant spatial variations, especially with respect to spatial variations in bottom roughness. A zonal section of diffusivity, averaged in the latitude band between 24°S and 28°S, is shown in Figure 2 for the Atlantic (Figure 2a), Indian (Figure 2b), and Pacific (Figure 2c) oceans. In each panel of the figure, the envelope defined by the deepest and shallowest bathymetry for the 4° latitude band is shown.

[15] In the South Atlantic (Figure 2a), the Mid-Atlantic Ridge (15°W), and the Walvis Ridge (5°E) are sites where elevated levels of diffusivity reach the mid-depth ocean. At abyssal depths, diffusivity estimates in the Angola Basin exceed those in the Brazil Basin. This signal is tied to differences in abyssal stratification, with weaker stratification occurring in the Angola Basin.

[16] In the Indian Ocean (Figure 2b), enhanced levels of diffusivity are estimated along the southern edge of Madagascar (45°E), and along the Central Indian Ridge (70°E). Diffusivity levels are notably weak along rough topography in the eastern Indian Ocean, including the regions along the Ninety East Ridge and the East Indian Ridge (100°E). Despite the presence of topographic roughness in these regions, tidal forcing is weak, resulting in low estimates of internal tide energy flux.

[17] Across the South Pacific Ocean (Figure 2c), elevated mixing rates at shallow depths are associated with the bathymetry of the South Fiji Basin (170°E) and the seamounts of French Polynesia (215°E). Elevated mixing levels are also found along the Sala Y Gomez Ridge (230°–280°E).

4. Discussion

[18] The meridionally averaged sections of diffusivity (Figure 2) show that the most enhanced levels of mixing occur within the envelope of rough topography characterizing the section. This is consistent with observations of turbulence above rough topography (e.g., Figure 2 of *St. Laurent et al.* [2001]).

[19] We have examined the sensitivity of (2) to the efficiency parameters, and to the assumed decay scale for the vertical structure (3). Uncertainty terms for a spatially averaged diffusivity estimate $\overline{k_v}$ were propagated using the relation

$$\frac{\delta\overline{k_v}}{\overline{k_v}} = \left(\left(\frac{\delta\Gamma}{\Gamma}\right)^2 + \left(\frac{\delta q}{q}\right)^2 + \left(\frac{\delta\zeta}{\zeta}\right)^2 + \left(\frac{\delta\overline{N^2}}{\overline{N^2}}\right)^2 \right)^{1/2}. \tag{4}$$

Here, $\delta\overline{N^2}$ is the standard deviation of the spatially varying stratification. The additional uncertainty parameters were taken as $\Gamma \pm \delta\Gamma = 0.2 \pm 0.04$, $q \pm \delta q = 0.3 \pm 0.1$, and $\zeta \pm \delta\zeta = 500 \pm 200$ m. The standard error value (68% confidence interval) for the mixing efficiency is based on direct estimates of Γ from oceanic microstructure [*St. Laurent and Schmitt*, 1999]. This uncertainty analysis was applied to estimates of mixing rates in the abyssal Brazil Basin, where comprehensive estimates of mixing rates have been made

Table 1. Comparison of Parameterized and Budget-Derived Diffusivities for Various Neutral Density Classes in the Brazil Basin

bounding surface γ (kg m^{-3})	inverse budget	microstructure $\overline{K_v}$ (10^{-4} m^2 s^{-1})	parameterization
28.133	2.9 ± 2.3	1.2 ± 0.6	1.4 ± 0.8
28.160	2.0 ± 1.3	1.2 ± 0.6	1.7 ± 1.0
28.205	1.8 ± 0.8	1.8 ± 1.1	1.7 ± 1.0
28.270	1.8 ± 0.8	3.1 ± 1.8	2.1 ± 1.2

Inverse budget and microstructure derived estimates of $\overline{K_v}$ are described by *Morris et al.* [2001]. Parameterized estimates are derived from (2) and (4).

[*Polzin et al.*, 1997; *Morris et al.*, 2001]. Average diffusivities were calculated for a series of Brazil Basin control volumes, defined by a neutral surface above and the topography below. These neutral density classes correspond to $\gamma \geq 28.133, \gamma \geq 28.16$, $\gamma \geq 28.205$, and $\gamma \geq 28.27$ kg m^{-3}, and are equivalent to the control volumes used by *Morris et al.* [2001]. Table 1 shows a comparison of parameterized and previous diffusivity estimates. The parameterized estimates compare well with previous estimates. It follows that our assumed local dissipation efficiency of $q \simeq 0.3$ is reasonable, at least in the Brazil Basin. Parameterized diffusivity estimates would significantly exceed previous estimates there if $q = 1$ were used. Additionally, we note that our estimates of uncertainty for the parameterization are comparable to the uncertainties of previous estimates.

[20] We have proposed a parameterization for enhanced mixing rates that occur in regions of internal tide generation. Tidal energy flux is partitioned into a fraction q that dissipates above the generation region, and a fraction $1-q$ that radiates away as low-mode waves. If $q \simeq 0.3$ is applicable as a global average for the local dissipation efficiency of baroclinic tides, then roughly 0.3 TW of power is directly consumed in abyssal waters by diapycnal mixing according to this mixing scheme. This is within the *Webb and Suginohara* [2001] estimate of 0.14 TW to 0.35 TW of dissipation for closing the mass budget of North Atlantic Deep Water. However, the remaining 0.7 TW of baroclinic tidal energy produced in the deep ocean must dissipate somewhere.

[21] The diffusivity parameterization presented here is regarded as a preliminary formulation. This mixing scheme is intended for application in OGCMs, particularly those used at coarse resolution to study the long time scales associated with the thermohaline circulation. A preliminary implementation of the parameterization in such a model is discussed by *Simmons et al.* [2002], whose initial results suggest that the equilibrium behavior of OGCM simulation is significantly improved by this new mixing scheme. Thus, it is hoped that the current parameterization serves as a useful tool in studies of long term climate.

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