The Combined Effect of Tidally and Eddy-Driven Diapycnal Mixing on the Large-Scale Ocean Circulation

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ABSTRACT

Several recent studies have shown that ocean western boundaries are the primary regions of eddy energy dissipation. Globally, the eddy energy sinks have been estimated to integrate to about 0.2 TW. This is a sizable fraction of the tidal energy dissipation in the deep oceanic interior, estimated at about 1.0 TW and contributing to diapycnal mixing. The authors conduct sensitivity experiments with an ocean general circulation model assuming that the eddy energy is scattered into high-wavenumber vertical modes, resulting in energy dissipation and locally enhanced diapycnal mixing. When only the tidal energy maintains diapycnal mixing, the overturning circulation, and stratification in the deep ocean are too weak. With the addition of the eddy dissipation, the deep-ocean thermal structure becomes closer to that observed and the overturning circulation and stratification in the abyss become stronger. Furthermore, the mixing associated with the eddy dissipation can, on its own, drive a relatively strong overturning. The stratification and overturning in the deep ocean are sensitive to the vertical structure of diapycnal mixing. When most of this energy dissipates within 300 m above the bottom, the abyssal overturning and stratification are too weak. Allowing for the dissipation to penetrate higher in the water column, such as suggested by recent observations, results in stronger stratification and meridional circulation. Zonal circulation is also affected. In particular, the Drake Passage transport becomes closer to its observational estimates with the increase in the vertical scale for turbulence above topography. Consistent with some theoretical models, the Drake Passage transport increases with the increase in the mean upper-ocean diffusivity.

1. Introduction

Turbulent mixing in a stratified ocean requires mechanical energy. Two primary sources of such energy are due to winds and tides (Munk and Wunsch 1998; Wunsch and Ferrari 2004). In global ocean models, the associated turbulence has to be parameterized. Several recent studies, based on such models, have explored the impact of tidal energy dissipation in the deep ocean on the simulated large-scale circulation and stratification (e.g., Simmons et al. 2004; Saenko and Merryfield 2005; Jayne 2009). Here, we further elaborate on this topic, focusing on the combined effect of the tidal energy and the energy associated with mesoscale eddy field in the ocean (which essentially comes from the winds).

Eddies can be generated by a number of mechanisms, such as baroclinic–barotropic instability of mean flows, localized topographic steering, and direct wind–buoyancy perturbations. Those eddies with horizontal scales close to the first baroclinic Rossby radius have their energy about equally partitioned between kinetic and potential forms (Gill 1982). At steady state, the sources and sinks of the eddy energy have to balance. The processes that remove eddy energy remain unclear, but the candidates...
include the ocean bottom drag, air–sea interaction, and the transfer of eddy energy to the mean flow. It is also likely that a substantial fraction of the eddy energy gets scattered into high-wavenumber internal waves; the breaking of which results in enhanced diapycnal mixing (e.g., Tandon and Garrett 1996; Naveira Garabato et al. 2004; Marshall and Naveira Garabato 2008; Arbic et al. 2010). The potential energy generated by the mixing could then contribute to sustaining the large-scale overturning circulation in the ocean (Huang 1999). Recent estimates (Zhai et al. 2010; Xu et al. 2011) suggest that western boundary regions and the Antarctic Circumpolar Current (ACC) are not only the major eddy source areas but also the places where much of the eddy energy dissipates (Fig. 1). Globally integrated, the rate of eddy energy generation/dissipation has been estimated independently at about 0.2 TW by Zhai et al. (2010) and by Xu et al. (2011).

Some early models of deep-ocean circulation were based on the assumption of uniform upwelling (e.g., Stommel and Arons 1960). In the real ocean, turbulent mixing is far from uniform (e.g., Polzin et al. 1997; Kunze et al. 2006). Microstructure measurements (e.g., Gregg 1987) and tracer release experiments (e.g., Ledwell et al. 1993) have revealed that, in most of the upper ocean, below the mixed layer, a representative value for diapycnal diffusivity is \( \mathcal{O}(10^{-2}) \) m\(^2\) s\(^{-1}\). However, such diffusivity values would be an order of magnitude too small to maintain the observed stratification below the pycnocline (Munk 1966). Therefore, it has been argued that much stronger mixing must be confined to a small number of concentrated areas, including along the boundaries, from which the water masses are exported into the ocean interior (Munk and Wunsch 1998).

Ocean general circulation models (GCMs) have proven to be a useful tool for addressing the impact of the heterogeneous nature of the small-scale turbulence on the large-scale circulation (e.g., Simmons et al. 2004; Saenko and Merryfield 2005; Saenko 2006; Jayne 2009). Here one such GCM is employed to evaluate the potential impact of the eddy energy sink on diapycnal mixing and large-scale ocean circulation, separately and in combination with the tidal energy dissipation. In section 2, we present some basic arguments suggesting a link between the kinetic energy dissipation in the oceanic interior and large-scale overturning circulation in the abyss. We also provide a brief review of the literature on the role of vertical diffusivity in maintaining zonal circulation in the Southern Ocean. The model, the mixing scheme, and the design of numerical experiments are described in section 3. This is followed by a presentation of main results in section 4 and conclusions in section 5.

2. Links between small-scale mixing and circulation

a. Overturning circulation

In a stratified ocean at steady state, the link between the dissipation of turbulent kinetic energy and upwelling in the abyss can be described by a buoyancy budget of the form

\[
\mathbf{u} \cdot \mathbf{V} b + w b_z = (K_v N^2)_z + Q_b, \tag{1}
\]

where \( \mathbf{V} \) is the two-dimensional gradient operator; \( b = -g \rho/\rho_0 \) is the buoyancy, where \( g, \rho, \) and \( \rho_0 \) are the acceleration due to gravity, ocean potential density, and reference density, respectively; \( K_v \) is the vertical diffusivity due to small-scale turbulent mixing; \( N^2 = b_z \) is the squared buoyancy frequency; and \( Q_b \) represents buoyancy...
sources and sinks. The velocity field \([\mathbf{u}(u, v, w)]\) is assumed to satisfy \(\nabla \cdot \mathbf{u} + \omega = 0\), and it represents the flows with scales larger than those associated with the small-scale turbulence (represented by the first term on the right-hand side). Assuming there are no significant buoyancy fluxes at the bottom and that \(Q\) is small, one can integrate Eq. (1) globally from the bottom to some horizontal surface in the deep ocean (Gnanadesikan et al. 2005) to obtain

\[
\langle wb \rangle = (K_v N^2),
\]

where \(\langle \cdot \rangle\) denotes horizontal integration. This balance implies that the potential energy created by the small-scale mixing. According to Borowski et al. (2002); that is, the transport of the ACC increases with the increase of mesoscale eddies, which tend to flatten it. According to Munday et al. (2011), in a realistic situation (with wind stress applied to the ocean) the transport of the ACC increases with the increase of global-mean vertical diffusivity at the base of pycnocline, provided that the diffusivity is relatively large. We shall test this theory too. In some other theoretical models the transport of the ACC has been related to the local diapycnal mixing in the Southern Ocean (e.g., Karsten and Marshall 2002; Borowski et al. 2002). We shall show that such a connection may also hold as long as the mean diffusivity is relatively large.

\[
\Psi \approx \frac{\chi}{f_{\text{ACC}}},
\]

Although much of \(\chi\) is likely maintained by the wind stress (e.g., Saenko 2009), a portion of \(\chi\) could be supplied by diapycnal mixing. Munday et al. (2011) used a (two layer) version of Eq. (6) to propose a conceptual model of the ACC. Their model is based on competition between diapycnal mixing, which tends to deepen the pycnocline north of ACC, and mesoscale eddies, which tend to flatten it. According to Munday et al. (2011), in a realistic situation (with wind stress applied to the ocean) the transport of the ACC increases with the increase of global-mean vertical diffusivity at the base of pycnocline, provided that the diffusivity is relatively large. We shall test this theory too. In some other theoretical models the transport of the ACC has been related to the local diapycnal mixing in the Southern Ocean (e.g., Karsten and Marshall 2002; Borowski et al. 2002). We shall show that such a connection may also hold as long as the mean diffusivity is relatively large.

3. The model and mixing scheme

A parameterization of small-scale mixing in an ocean GCM has to ensure that the power that maintains such mixing against gravity does not exceed the power that may be available locally from different sources (e.g., Osborn 1980). One such parameterization has been proposed by St. Laurent et al. (2002) for the case where the mechanical energy that supports small-scale turbulence comes from dissipation of internal tides in the deep ocean.

Here we follow these ideas. Vertical diffusivity in our ocean GCM is determined by a combined effect of tidal and eddy energies

\[
K_v = K_b + \Gamma \epsilon_1 (x, y, z) + \epsilon_2 (x, y, z),
\]

where

\[
\epsilon_1 (x, y, z) = q_1 E_{\text{tidal}} (x, y) F_1 (z) (\text{W kg}^{-1})
\]

and

\[
\epsilon_2 (x, y, z) = q_2 E_{\text{eddy}} (x, y) F_2 (z) (\text{W kg}^{-1})
\]
In Eqs. (7)–(9), $K_b = 10^{-5} \text{ m}^2 \text{ s}^{-1}$ is the background diffusivity that is to account for the observational evidence that in most of the upper ocean and away from rough topography the diffusivity is relatively weak (e.g., Gregg 1987; Ledwell et al. 1993; Kunze et al. 2006); $E_{\text{tidal}}(x, y)$ and $E_{\text{eddy}}(x, y)$ are the estimated energy fluxes provided by internal tides (Jayne and St. Laurent 2001) and mesoscale eddies (Zhai et al. 2010; Fig. 1), respectively; and $q_1$ and $q_2$ are the fractions of the respective energies that dissipate locally. Similar to Saenko and Merryfield (2005), we prohibit $K_y$ in Eq. (7) from exceeding $20 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$, which is about 2 times the global-mean near-bottom diffusivity estimated by Ganachaud and Wunsch (2000). In the model sensitivity experiments discussed in section 4, we consider the cases where $q_1 = \frac{1}{3}$ (St. Laurent and Garrett 2002) and $q_2 = 1$. St. Laurent et al. (2002) argue that 0.3 ± 0.1 value
for the fraction of the energy that dissipates locally is reasonable for the Brazil Basin. However, this may not necessarily hold for any ocean basin (e.g., Althaus et al. 2003). In our sensitivity experiments the \( q_2 = 1 \) case corresponds to an extreme situation wherein (i) all of the eddy energy sinks (that remain after accounting for \( \Gamma = 0.2 \) mixing efficiency) support the diapycnal mixing in the ocean interior and (ii) they do so locally. This will be compared with another extreme case where \( q_2 = 0 \). Whenever \( q_1 \neq 1 \) and/or \( q_2 \neq 1 \), an implicit assumption is that the rest of the associated energy (1 – \( q_1 \) and 1 – \( q_2 \)) could contribute to sustaining the background internal wave field. The vertical structure functions \( F_1 \) and \( F_2 \) satisfy \( \int_{-H}^{0} F_1(z) \, dz = \int_{-H}^{0} F_2(z) \, dz = 1 \). For simplicity, we assume that \( F_1(z) = F_2(z) = F(z) \). The latter has the form that ensures exponential decay of the mixing energy from the bottom \( H(x, y) \) upward (e.g., St. Laurent et al. 2002),

\[
F(z) = \frac{e^{-\frac{(H+z)}{h}}}{h(1 - e^{-\frac{H}{h}})},
\]

where \( h \) is the vertical decay scale for turbulence, which is varied in the sensitivity experiments between 300 and 1000 m: that is, roughly within its uncertainty ranges (St. Laurent et al. 2002).

Some of the above assumptions and parameters are likely to be only crude representations of the turbulent mixing in the ocean. Our motivation, however, is not so much to examine the most realistic case but to conduct a set of sensitivity experiments. In these, we aim to make a step toward estimating the potential role of the tidal and eddy dissipation energies, separately and in combination, in maintaining the oceanic stratification and large-scale overturning circulation. Examples of the vertical diffusivity maps at the middepth ocean due to the tidal energy dissipation and eddy energy dissipation, obtained using the above assumptions and climatological distribution of \( N^2 \), are shown in Fig. 2. As expected, the largest values of \( K_v \) are found above the major topographic features and along the western boundaries. Note, in particular, that in some weakly stratified ocean regions, such as in the subpolar Atlantic, the vertical diffusivity due to the eddy energy dissipation can be locally up to an order of magnitude larger than its background value (Fig. 2, bottom), even though the eddy energy sink in this region is relatively weak (Fig. 1).

The ocean model employed is a version of the National Center for Atmospheric Research (NCAR) Community Ocean Model (Gent et al. 1998), which was developed from version 1 of the Geophysical Fluid Dynamics Laboratory Modular Ocean Model. This is the same model as in Saenko and Merryfield (2005), but with the horizontal resolution of \( 1.41^\circ \times 0.94^\circ \) (longitude × latitude) and 33 vertical levels. The model employs the eddy transport parameterization of Gent and McWilliams (1990), with coefficients of thickness diffusivity and isopycnal diffusivity set to \( 10^3 \) m\(^2\) s\(^{-1}\). Wind stress forcing consists of daily averages from a 30-yr run of the third-generation Canadian Centre for Climate Modelling and Analysis (CCCma) atmospheric general circulation model (AGCM3) with climatological sea surface temperatures (SSTs). Surface heat and freshwater fluxes consist of daily averages from the same CCCma AGCM3 run, together with restoring to (linearly interpolated in time) monthly climatological fields of SST and sea surface salinity (SSS) from the Polar Science Center Hydrographic Climatology (PHC) 2.0 (Steele et al. 2001) with restoring time scales of 30 days for temperature and 180 days for salinity. This climatology is a merged product that combines data from the World Ocean Atlas and the Arctic Ocean Atlas. We present results from five sensitivity experiments, all run for 5000 yr (i.e., to near equilibrium). These differ by the energy sources that sustain vertical mixing in the ocean and by the value set for the vertical decay scale \( h \) for turbulence (Table 1).

4. Results

a. Impact of eddy energy

The simulated thermal structure in the upper \( \sim 2 \) km of the ocean is close to that observed (this is illustrated in Fig. 3, top, for the “Tidal” experiment; see Table 1). It is largely maintained by the wind stress and its curl (Saenko 2009), as can be expected based on the theories of the ventilated thermocline (Luyten et al. 1983). The deep ocean, however, is too cold in the Tidal model (Fig. 3, top). Based on the arguments presented in Munk and Wunsch (1998), one reason for this could be related to insufficient mechanical energy for sustaining diapycnal mixing in the abyss. According to these arguments, in the presence of cold water sources in polar regions the deep ocean is expected to be filled with these waters if the mixing of heat from the upper ocean is not strong enough (e.g., Saenko and Merryfield 2005). Consistent with these arguments, the abyssal ocean becomes warmer when more energy is supplied to sustain the mixing in the “Tidal + Eddy” experiment (Fig. 3, top). Furthermore, the different geographic distributions between the tidal and eddy energy sources are an important factor in maintaining the thermal structure in the abyss. To illustrate this, we conducted an additional sensitivity experiment (“Tidal*”) setting \( q_1 = 0.5 \) and \( q_2 = 0 \) in Eqs. (7)–(9). This provides about the same amount of energy for the mixing as in the Tidal + Eddy
model. As might be expected, the integrated meridional overturning circulation (MOC) in the Tidal* experiment (not shown) is largely the same as in the Tidal + Eddy case (see next paragraph). However, in the Tidal* experiment, the largest warming is less widespread and is mostly confined to the middepth ocean (Fig. 3, bottom).

The rate of MOC differs between the Tidal, Eddy, and Tidal + Eddy experiments but mostly below ~2-km depth (Fig. 4). This is expected because most of the energy sustaining the small-scale mixing in the model is constrained to dissipate below 2-km depth by the vertical profile function. In the Tidal model, about 8 Sv is transported upward across 3.5-km depth and north of about 25°S (Fig. 4a). This is less than the observational estimates presented by Talley et al. (2003). She estimated that in the Pacific the bottom water upwelling is about 10 Sv, with a comparable contribution in the Indian Ocean. The eddy energy, on its own, can drive as much as 5 Sv of the bottom water flux across 3.5 km (Fig. 4b). However, when combined with the tidal energy, the effect is nonlinear with respect to the overturning circulation (Fig. 4c). The maximum (in fact the minimum, as plotted in Fig. 4) overturning in the abyss increases in the Tidal + Eddy model compared to those in the Tidal and Eddy models by about 1.2 and 4.2 Sv, respectively. The increase is consistent with the arguments presented in section 2. The transport associated with the outflow of North Atlantic Deep Water (NADW) is roughly the same in all three models.

The observed stratification increases upward. In the zonal mean, the only region where relatively large values of $N^2$ penetrate to the abyssal ocean is roughly between 30° and 50°S (Fig. 5a). This general structure of the observed stratification is captured by the Tidal model (Fig. 5b). The ratio between the simulated and observed $N^2$ reveals that between 1- and 3-km depth the stratification simulated by the Tidal model is too strong north of 40°S, whereas below 3 km it is too weak (Fig. 5c). The latter could be indicative of insufficient mixing energy and hence too weak vertical mixing in the abyss (Saenko and Merryfield 2005). [On the other hand, too strong mixing could also lead to unrealistic values for the vertical scale of stratification. Assuming the so-called advective–diffusive balance holds (Munk and Wunsch 1998; Wunsch and Ferrari 2004), this scale is $K_v/w$, where $w$ is the vertical velocity. For $w = 10^{-7}$ m s$^{-1}$ (a realistic value), the value of $K_v = 10^{-3}$ m$^2$ s$^{-1}$ gives 10 km for the vertical scale of stratification.]

Aside from the diapycnal mixing, there are likely many other physical processes that are represented only approximately by the model. As for the mixing, the available estimates suggest that it would require some 2–3 TW of mechanical energy to maintain the observed stratification in the deep ocean (Munk and Wunsch 1998; St. Laurent and Simmons 2006). It is therefore not expected

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**Table 1. List of main numerical experiments.**

<table>
<thead>
<tr>
<th>Expt</th>
<th>Energy source(s)</th>
<th>Vertical decay scale (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tidal</td>
<td>Tidal</td>
<td>500</td>
</tr>
<tr>
<td>Eddy</td>
<td>Eddy</td>
<td>500</td>
</tr>
<tr>
<td>Tidal + Eddy</td>
<td>Tidal and eddy</td>
<td>500</td>
</tr>
<tr>
<td>Tidal + Eddy1000</td>
<td>Tidal and eddy</td>
<td>1000</td>
</tr>
<tr>
<td>Tidal + Eddy300</td>
<td>Tidal and eddy</td>
<td>300</td>
</tr>
</tbody>
</table>
that adding \(-0.2\) TW of the eddy energy to \(-1\) TW of the tidal energy dissipation in the deep ocean would correct for the too weak simulated stratification in the abyss. Nevertheless, below about 4-km depth the Tidal + Eddy model does tend to simulate a stronger stratification in the ocean north of the equator and south of 55\(^\circ\)S (Fig. 5d). In the next subsection, we shall illustrate that, given the same amount of energy to support the diapycnal mixing (i.e., tidal plus eddy), the abyssal stratification can be quite sensitive to the vertical decay scale for turbulence above the topography.

b. Sensitivity to vertical decay scale

As noted in section 3, some observational studies on turbulence above rough topography provide evidence for its sharp decay above the bottom (see, e.g., St. Laurent et al. 2002 and references therein). This is one of the reasons why some of the related modeling studies employ Eq. (10) with \(h = 500\) m to represent the vertical structure of the kinetic energy dissipation that maintains the associated diapycnal mixing (e.g., Simmons et al. 2004; Saenko and Merryfield 2005; Jayne 2009). However, results from some other observational studies suggest a larger decay scale for the turbulence above the bottom (see Decloedt and Luther 2010, and references therein). In addition, examples have been presented where the vertical structure of the dissipation may vary considerably within the same study region and may even increase upward (e.g., Naveira Garabato et al. 2004). The estimates by Kunze et al. (2006) suggest a rather uniform structure for the mean vertical profile of the dissipation as a function of height above the bottom.

Motivated in part by this uncertainty in the vertical structure of the kinetic energy dissipation, we discuss here two additional sensitivity experiments, “Tidal + Eddy1000” and “Tidal + Eddy300.” For simplicity, these are identical to the Tidal + Eddy experiment, except the vertical decay scale for turbulence above the bottom \(h\) in Eq. (10) is either increased to 1000 m or decreased to 300 m from its 500-m value in the Tidal + Eddy model (see Table 1) \((h\) is most likely a function of location). The corresponding global-mean simulated diffusivity at \(-1000\)-m depth then varies between \(0.38 \times 10^{-5}\) m\(^2\) s\(^{-1}\) and \(0.55 \times 10^{-5}\) m\(^2\) s\(^{-1}\) when \(h\) is set to, 300 and 1000 m, respectively. These values of mean diffusivity are within the observational estimates. In particular, the global-mean diffusivity estimates by Decloedt and Luther (2010) at 1000-m depth range from about \(0.15 \times 10^{-5}\) m\(^2\) s\(^{-1}\) to \(0.6 \times 10^{-5}\) m\(^2\) s\(^{-1}\). The two global-mean estimates by Kunze et al. (2006) at this depth are \(0.15 \times 10^{-5}\) m\(^2\) s\(^{-1}\) and \(0.4 \times 10^{-5}\) m\(^2\) s\(^{-1}\). In the deep ocean, the global-mean diffusivity in our model experiments is within the estimates of Decloedt and Luther (2010),
ranging from about $10^{-4}$ m$^2$ s$^{-1}$ at 3000-m depth to $10^{-3}$ m$^2$ s$^{-1}$ at 5000-m depth. Such a range is also consistent with estimates from inverse models (Ganachaud and Wunsch 2000; see also Lumpkin and Speer 2003). In contrast, the estimates of Kunze et al. (2006) for the mean diffusivity in the abyss are $10^{-4}$ m$^2$ s$^{-1}$, although they do acknowledge that the lowered ADCP (LADCP)–CTD data they use do not account for tidal dissipation and that their sampling may miss major hotspots of enhanced mixing.

The changes in the decay scale for turbulence affect the MOC in the abyssal ocean (Fig. 6). Compared to the Tidal + Eddy case, the rate of anticlockwise overturning below about 2-km depth increases in response to the larger $h$ by about 1.2 Sv and decreases in the case of the smaller $h$ by about 6.1 Sv. In other words, the rate of the large-scale overturning can be changed by changing vertical distribution of the energy dissipation: that is, without changing the total amount of mechanical energy maintaining the turbulent mixing. An explanation for this can be provided by assuming locally the vertical advective–diffusive balance of buoyancy; that is,

$$w_b \approx (K_y N^2)_z \tag{11}$$

so that

$$w = \frac{(K_y N^2)_z}{N^2}. \tag{12}$$

If $K_y$ is given by Eq. (7), then by substituting (7) into (12) we obtain

$$w = K_y \frac{\partial (\ln N^2)}{\partial z} + \Gamma \frac{\epsilon}{N^2}. \tag{13}$$
where, in this case, $\epsilon = \epsilon_1 + \epsilon_2$ and $K_y$ and $\Gamma$ are positive constants. The first term on the right-hand side of Eq. (13) is positive definite, given the typically observed and modeled buoyancy distribution, and drives upwelling in the sense of enhancing the bottom cell. However, the second term on the right-hand side of Eq. (13) is negative definite and drives downwelling, because $\epsilon$ decays upward. Therefore, under the assumption that Eq. (11) holds, it is the competition between these two terms that determines the strength of the overturning circulation. Increasing the decay scale reduces $|\epsilon|$, and hence reduces the downwelling associated with the second term in Eq. (13), so that the overturning circulation strengthens (assuming the first term does not change much). Similarly, decreasing the decay scale weakens the near-bottom overturning (Fig. 6).

The Tidal + Eddy1000 model also simulates the strongest stratification below about 3 km depth. In contrast, when most of the energy that supports diapycnal mixing is constrained to dissipate within 300 m above the bottom, the simulated stratification in the abyss at steady state is the weakest (Fig. 7).

Increasing the decay scale for turbulence above the bottom increases the downward diffusion of heat. As a result, most of the deep ocean becomes warmer (Fig. 8). It is interesting to note that the largest warming is simulated along the western boundary in the deep Atlantic Ocean (Fig. 8b), even though the net rate of NADW overturning is not much affected by the change in the decay scale. To compensate for the stronger bottom water upwelling (Fig. 6), the sinking of the Antarctic Bottom Water becomes stronger with the increase in the decay scale for turbulence and the deep ocean near Antarctica becomes colder (Fig. 8).

Zonal circulation, particularly in the Southern Ocean, is also strongly affected. In our experiments, the net transport through the Drake Passage increases from 115.4 to 129.8 and 137.5 Sv when the scale for turbulence increases from 300 to 500 and 1000 m, respectively. For comparison, we note that Cunningham et al. (2003) estimated the baroclinic transport through Drake Passage at 136.7 $\pm$ 7.8 Sv (relative to the deepest common level for six hydrographic sections along the World Ocean Circulation Experiment line SR1b). A recent eddy-permitting Southern Ocean state estimate by Mazloff et al. (2010) gives 153 $\pm$ 5 Sv for the net Drake Passage volume transport.

As argued in section 2b, the distribution of $\chi/f$ is found to be a good proxy to the horizontal streamfunction in the Southern Ocean. The Tidal + Eddy1000 model estimate of $\chi/f$ is closer to the estimate of $\chi/f$ based on the observed ocean hydrography, as compared to the Tidal + Eddy300 and Tidal + Eddy model estimates of $\chi/f$ (Fig. 9). The Tidal + Eddy1000 model also simulates the largest values of JEBAR northeast of the passage (Fig. 9): that is, where the JEBAR effect maintains a deflection of the ACC to the north across the contour of $f/H$ and forms the Malvinas Current.

Fig. 6. Global meridional overturning circulation (Sv) below about 1-km depth in the (a) Tidal + Eddy1000 and (b) Tidal + Eddy300 experiments.
with the theory and numerical results presented by Munday et al. (2011). In particular, according to Munday et al. (2011), for fixed wind and eddy strength in the Southern Ocean, the ACC transport is insensitive to the diffusivity if the mixing-driven upwelling across the base of pycnocline is sufficiently small (relative to Ekman transport across the ACC). In our experiments, however, the diffusivity averaged over the whole ocean area at ~1-km depth essentially follows that averaged for the ocean south of 40°S (Fig. 10). As such, the changes in the ACC transport may as well be explained based on the Southern Ocean circulation theories where this transport has been related to the local vertical diffusivity (e.g., Karsten and Marshall 2002). We note that, consistent with Munday et al. (2011), it is found that in the presence of wind stress a relatively strong ACC transport can be maintained even in the limit of relatively low values of diapycnal diffusivity (Fig. 10). Finally, it should be pointed that Jayne (2009) also discussed the dependence of the ACC transport on the vertical diffusivity, but in a context different from that of Munday et al. (2011).

5. Discussion and conclusions

Using a simple advective–diffusive balance, Munk (1966) proposed that in order to maintain the observed stratification in the ocean the diapycnal diffusivity at the lower part of the pycnocline would have to be $O(10^{-4})$ m$^2$ s$^{-1}$. However, microstructure measurements (e.g., Gregg 1987) and tracer release experiments (e.g., Ledwell et al. 1993) have revealed an order of magnitude lower
diffusivity values, below the mixed layer and away from
topography. In an attempt to reconcile the estimated
by Munk diffusivity value with the observations, Munk
and Wunsch (1998) argued that a much stronger (com-
pared to the typically observed) mixing could be confined
to a small number of concentrated areas, including along
the boundaries.

Several recent studies have shown that ocean western
boundaries are the regions of significant sink of meso-
scale eddy energy (e.g., Zhai et al. 2010; Xu et al. 2011).
Here, we conducted a set of ocean general circulation
model experiments assuming that this energy is scat-
tered into high-wavenumber vertical modes resulting in
dissipation and enhanced diapycnal mixing. Globally,
the eddy energy sinks integrate to ~0.2 TW (Zhai et al.
2010; Xu et al. 2011), which is a sizable fraction of the
tidal energy dissipation in the deep ocean. The purpose
of the model experiments was to investigate the effect of
the eddy and tidal dissipation energies, in combination
and separately, on the ocean stratification and large-
scale circulation, through the impact of the dissipation
on diapycnal mixing. We show that, when only the tidal
energy maintains diapycnal mixing, the overturning
circulation and stratification in the deep ocean are too
weak. With the addition of the eddy dissipation, the
depth-ocean thermal structure becomes closer to that
observed and the overturning and stratification in the
abyss strengthen. Furthermore, the mixing associated
with the eddy dissipation can, on its own, drive a rela-
tively strong overturning in the abyss. However, when
combined with the tidal energy, the effect is nonlinear
with respect to the overturning circulation. We also note
that eddy dissipation is not estimated near the equator
because geostrophic balance was used in Zhai et al.
(2010). Including the missing mixing associated with
eddy dissipation near the equator could lead to an even
stronger overturning circulation.

Additional sensitivity experiments show that stratifi-
cation and overturning in the deep ocean are quite sen-
sitive to the vertical structure of diapycnal mixing. When
most of this energy is constrained to dissipate within
300 m above the bottom, the abyssal overturning and

![Fig. 9. The baroclinic potential energy scaled by the Coriolis parameter (\(\frac{\chi}{f}\); in Sv) in the Drake Passage region (a)–(c) in the models with different decay scale for turbulence \(h\) and (d) based on the observed PHC climatology (Steele et al. 2001). The dashed lines show the contours of \(fH\), where \(H(x, y)\) is the bottom topography resolved by the model. The dark gray areas indicate the regions with relatively large JEBAR effect: that is, where \(|J(\chi, H^{-1})| > 10^{-11}\) s\(^{-2}\). The baroclinic potential energy was computed using \(\sigma_2\) (potential density referenced to 2000-m depth), based on the model data or the observed climatology.]
stratification are too weak and the deep ocean is too cold. Allowing for the dissipation to penetrate higher into the water column, such as suggested by some recent observations, results in a stronger overturning and stratification in the deep ocean.

Zonal circulation in the Southern Ocean is also strongly affected by the variations in the vertical scale for turbulence. In particular, the simulated distribution of the baroclinic potential energy, which has been shown to be a good proxy for the vertically integrated transport, becomes progressively closer to the corresponding observational estimate when more mixing energy is constrained to dissipate higher above the bottom. In the model, the Drake Passage transport increases with the increase of the mean vertical diffusivity at the base of pycnocline, as predicted by some theories of the ACC (Munday et al. 2011).

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REFERENCES


