Sensitivity of the Ocean State to Lee Wave–Driven Mixing

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ABSTRACT

Diapycnal mixing plays a key role in maintaining the ocean stratification and the meridional overturning circulation (MOC). In the ocean interior, it is mainly sustained by breaking internal waves. Two important classes of internal waves are internal tides and lee waves, generated by barotropic tides and geostrophic flows interacting with rough topography, respectively. Currently, regarding internal wave–driven mixing, most climate models only explicitly parameterize the local dissipation of internal tides. In this study, the authors explore the combined effects of internal tide– and lee wave–driven mixing on the ocean state. A series of sensitivity experiments using the Geophysical Fluid Dynamics Laboratory CM2G ocean–ice–atmosphere coupled model are performed, including a parameterization of lee wave–driven mixing using a recent estimate for the global map of energy conversion into lee waves, in addition to the tidal mixing parameterization. It is shown that, although the global energy input in the deep ocean into lee waves (0.2 TW; where 1 TW = 10¹² W) is small compared to that into internal tides (1.4 TW), lee wave–driven mixing makes a significant impact on the ocean state, notably on the ocean thermal structure and stratification, as well as on the MOC. The vertically integrated circulation is also impacted in the Southern Ocean, which accounts for half of the lee wave energy flux. Finally, it is shown that the different spatial distribution of the internal tide and lee wave energy input impacts the sensitivity described in this study. These results suggest that lee wave–driven mixing should be parameterized in climate models, preferably using more physically based parameterizations that allow the internal lee wave–driven mixing to evolve in a changing ocean.

1. Introduction

The breaking of internal waves represents the main source of diapycnal mixing in the ocean interior (Garrett and Kunze 2007). Diapycnal mixing in turn plays a key role in maintaining the ocean stratification and the meridional overturning circulation (MOC): the convective creation of dense surface water that occurs in a few locations of the global ocean (e.g., Southern Ocean, Nordic seas, and Labrador Sea) is balanced by the upwelling of deep water driven by both turbulent mixing in the ocean interior and the winds in the Southern Ocean (e.g., Marshall and Speer 2012; Talley 2013). Through an advection–diffusion balance, the upwelling driven by turbulent mixing is associated with a downward heat flux, which is responsible for the stable and finite stratification observed in the deep ocean.
Munk and Wunsch (1998) estimated that 2 TW (1 TW = 10^{12} W) of energy needs to be dissipated to maintain the observed MOC. Tides and winds input energy in the ocean on large scales and part of this energy dissipates through breaking internal waves. In the deep ocean, two important classes of energetic internal waves are internal tides and lee waves. Both are generated by the interaction of the bottom flow (due to barotropic tides in the former case and geostrophic motions in the latter case) with rough topography in the stratified ocean.

The energy lost by the barotropic tide field in the deep ocean to internal tides has been estimated to be 1 TW (Egbert and Ray 2000; Jayne and St. Laurent 2001), which represents about a third of the work done by tides in the global ocean (Munk and Wunsch 1998).

The wind provides $O(1)$ TW of power into geostrophic flows, predominantly in the Southern Ocean (Wunsch 1998). This wind power input contributes to drive all the major ocean currents, such as the western boundary currents, the equatorial current system, and the Antarctic Circumpolar Current (ACC) in the Southern Ocean. Much of this energy is subsequently converted into geostrophic eddies through baroclinic instability.

Several processes can dissipate eddy energy in the ocean (Wunsch and Ferrari 2004; Ferrari and Wunsch 2009). Among them, bottom drag provides a direct mechanical damping of eddies (Arbic and Flierl 2004), with an estimated 0.2–0.8 TW of energy dissipation (Sen et al. 2008). Zhai et al. (2010) showed that the western boundary of the oceanic basins acts as a “graveyard” for the westward-propagating ocean eddies, dissipating 0.1–0.3 TW poleward of 10° in latitude. Over recent years, evidence has accumulated that a significant fraction of the energy in the geostrophic eddy field is dissipated through the generation of lee waves over small-scale topographic roughness and their subsequent breaking (e.g., Naveira Garabato et al. 2004; Nikurashin and Ferrari 2010; Nikurashin et al. 2013). Recently, independent estimates of the global energy input into internal lee waves from geostrophic motions based on linear theory (Bell 1975a) have been provided by Nikurashin and Ferrari (2011) and Scott et al. (2011). These authors predict consistent global energy conversions in the range of 0.2–0.4 TW, and highlight the Southern Ocean as the area of strongest lee wave generation in the World Ocean. The prominent role of the Southern Ocean for lee wave generation can be explained by the concurrent presence of rough topography and vigorous geostrophic motions. The ACC is indeed a strong and highly barotropic geostrophic current, with significant bottom velocities, and is associated with intense mesoscale eddy activity (e.g., Morrow et al. 1994; Meredith and Hogg 2006). Observational studies reveal that intense turbulent diapycnal mixing and energy dissipation are common throughout areas of rough bottom topography under the ACC, such as the Kerguelen Plateau (Polzin and Firing 1997; Waterman et al. 2013), Drake Passage, and Scotia Sea (Heywood et al. 2002; Naveira Garabato et al. 2004; Sloyan 2005; St. Laurent et al. 2012). The collocation of enhanced energy dissipation with the ACC fronts and weak tidal flows in those regions are suggestive of a subinertial energy source for the enhanced turbulence, indicating that the high rates of mixing and dissipation are likely associated with an internal lee wave mechanism.

The diapycnal mixing from the breaking internal waves occurs on scales too small for global ocean models to explicitly resolve and has to be parameterized [see Jayne (2009) for a review of the different diapycnal mixing parameterizations used over time in climate models]. In the most recent ocean general circulation models (OGCMs) or climate models, only the local dissipation of internal tides is parameterized, often using the semiempirical scheme formulated by St. Laurent et al. (2002) (e.g., Simmons et al. 2004; Saenko and Merryfield 2005; Jayne 2009). In the present study, we focus on the combined effect of tidal- and eddy-driven diapycnal mixing on the ocean state, with the eddy energy sink being provided here by the generation of internal lee waves (Nikurashin and Ferrari 2011).

We conduct a series of sensitivity experiments (described in section 2) to study how the addition of lee wave energy dissipation and its spatial distribution impacts the ocean state in a climate model (section 3). We show that sensitivity of the ocean state in our model is not only due to the increase in energy flux into internal waves but can also be substantially explained by the different spatial distribution of the internal tide and lee wave energy input. Therefore, these results suggest that lee wave–driven mixing specifically impacts the ocean state and should be parameterized in climate models.

2. Numerical simulations

a. Model

We performed a series of climate simulations using the Geophysical Fluid Dynamics Laboratory (GFDL) CM2G ocean–ice–atmosphere coupled model used for the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5) suite (Dunne et al. 2012). The ocean model is the Generalized Ocean Layer Dynamics (GOLD) isopycnal model (Hallberg and Adcroft 2009). The zonal resolution is 1°. The meridional resolution is 1° in the midlatitudes from 20° to 60°, and it
increases toward $\frac{1}{2}^\circ$ poleward of $60^\circ$ and toward $\frac{3}{4}^\circ$ equatorward of $20^\circ$. There are 63 vertical layers: 59 in the ocean interior, 2 for the refined bulk mixed layer, and 2 additional buffer layers for smooth water mass exchange with the isopycnal interior (Hallberg 2003). The isopycnal coordinates allow the interior ocean to be handled more naturally, and diapycnal mixing is explicitly parameterized, without spurious diapycnal mixing arising from advection as in $z$-coordinate models (Griffies et al. 2000; Ilicak et al. 2012). Thus, the isopycnal coordinates provide a favorable framework to study the effects of changing the diapycnal mixing. Along-isopycnal tracer mixing occurs via a Laplacian diffusion operator, while the horizontal viscosity uses a combination of Laplacian and biharmonic operators. The CM2G simulations shown here were run for 1000 years using year 1990 concentrations of carbon dioxide and other radiatively active gases and aerosols.

b. Diapycnal mixing parameterizations

Melet et al. (2013) show that the vertical distribution of the dissipation of internal wave energy matters for the ocean state and should evolve in time and space with the ocean state. Regarding internal tides, a physically based vertical profile of their energy dissipation has been formulated by Polzin (2009). However, we are currently lacking a corresponding theory for the vertical structure of lee wave energy dissipation. In the present study, we do not intend to examine the most realistic case for the internal waves’ energy dissipation, but rather to examine whether the addition of lee wave energy dissipation impacts the ocean state in a climate model in the first place. Therefore, for simplicity, the turbulent dissipation rates of both internal tide and lee wave energy, $\epsilon_T$ and $\epsilon_L$, are parameterized following the same semi-empirical scheme proposed by St. Laurent et al. (2002) for internal tide–driven local mixing:

$$\epsilon_T = \frac{q_T E_T(x,y)}{\rho_0} \frac{e^{-z_T / H_T}}{z_T (1 - e^{-H_T / H_T})}$$

and (1)

$$\epsilon_L = \frac{q_L E_L(x,y)}{\rho_0} \frac{e^{-z_L / H_L}}{z_L (1 - e^{-H_L / H_L})},$$

where $\rho_0$ is the reference density for seawater (set here to $1035 \text{ kg m}^{-3}$); $q_T$ and $q_L$ are the fractions of the generated internal tide and lee wave energy flux that are dissipated locally (i.e., near their generation sites), respectively; $z$ is the height above the bottom; $z_T$ and $z_L$ are the fixed vertical decay scales for the dissipation of internal tides and lee waves, respectively; and $H$ is the total depth of the ocean. The vertical structure of dissipation in the St. Laurent et al. (2002) formulation corresponds to an exponential function that satisfies energy conservation within an integrated vertical column (i.e., its integral over depth equals one). The energy that is not dissipated locally is simply disregarded in this formulation and is accounted for in our model in the ad hoc background mixing.

In (1), $E_T$ is the energy flux per unit area out of the barotropic tide (Fig. 1a). The $E_T$ is diagnosed online using the scaling relationship of Jayne and St. Laurent (2001):

$$E_T(x,y,t) = \frac{1}{2} \rho_0 N_b \kappa h^2 \langle U^2 \rangle.$$  (3)

Here, $N_b$ is the buoyancy frequency along the seafloor, $\langle U^2 \rangle$ is the barotropic tide variance [barotropic tidal velocities are computed with eight major tidal components by the Oregon State tidal model (Egbert and Erofeeva 2002)], and ($\kappa$, $h$) are the wavenumber and amplitude scales for the topographic roughness. The topographic roughness $h^2$ is computed as the root-mean-square of height deviations from a least-squared fit of the topography to a plane over a grid box using the Smith and Sandwell (1997) bathymetric dataset. The term $\kappa$ is a characteristic scale of topography, set here to $2\pi$ (10 km)$^{-1}$ following Jayne and St. Laurent (2001). This equation [(3)] is consistent with internal tide generation based on linear wave theory (Bell 1975b). Therefore, it is formally valid for subcritical topography only (where the slope of the topography is smaller than the slope of radiated tidal frequency beams of internal waves). This is not really problematic because internal tide generation is dominated by subcritical topography in the ocean (Nycander 2005). Over supercritical topography (where the slope of the topography is larger than the slope of radiated tidal frequency beams of internal waves), this scale relationship tends to overestimate the energy flux into internal tides (e.g., Nycander 2006). In our simulations, the globally integrated energy flux into internal tides is 1.6–1.7 TW, of which 1.4 TW occurs in the ocean deeper than 1000 m (Fig. 1a).

In (2), $E_L$ is the energy flux per unit area that goes into lee waves. In this study, we used the static estimate made by Nikurashin and Ferrari (2011) for $E_L$ (Fig. 1b), by linearly interpolating their estimate onto our ocean model grid. No extrapolation was done in regions of missing values. The generation rate of lee waves used in our simulation globally integrates to 0.2 TW.

When comparing the energy flux into internal tides and lee waves, two main differences appear. First, the globally integrated energy flux into internal tides in the deep ocean (1.4 TW) is stronger than that into lee waves (0.2 TW). Second, their spatial distributions are strikingly different (Fig. 1). While lee waves are mostly generated in regions...
of both elevated near-bottom kinetic energy and rough topography (such as the ACC), internal tides are mostly generated over continental slopes, midocean ridges, and island arcs (e.g., Nycander 2005). As a result, the Southern Ocean plays a very different role in the generation of lee waves and internal tides: while it accounts for half of the lee wave total energy flux, it is only responsible for \( \frac{17}{2} \% \) of the internal tide generation (Figs. 1c,d).

In ocean models, the diapycnal turbulent mixing is represented by diapycnal diffusivity \( K_d \). Spatially and temporally varying turbulent diffusivities are inferred from the dissipation using the Osborn (1980) model

\[
K_d = \frac{\Gamma \epsilon}{N^2},
\]

where \( N \) is the buoyancy frequency, and \( \Gamma \) is related to the mixing efficiency of turbulence and is generally set to \( \Gamma = 0.2 \) [see, e.g., Oakey (1982) and St. Laurent and Schmitt (1999) for justification of this choice]. However, to deal with very weak stratification where vertical buoyancy
Table 1. List of the numerical simulations and characteristics. The last column shows the global internal wave (IW) energy dissipation (TW), which corresponds to the integral over the World Ocean of the internal wave energy dissipation.

<table>
<thead>
<tr>
<th>Expt</th>
<th>Energy source</th>
<th>$z_T$ (m)</th>
<th>$q_T$</th>
<th>$z_L$ (m)</th>
<th>$q_L$</th>
<th>IW global energy dissipation (TW)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tidal</td>
<td>Tidal</td>
<td>300</td>
<td>$1/3$</td>
<td>—</td>
<td>—</td>
<td>0.5</td>
</tr>
<tr>
<td>Tidal_Lee</td>
<td>Tidal + lee</td>
<td>300</td>
<td>$1/3$</td>
<td>300</td>
<td>1</td>
<td>0.7</td>
</tr>
<tr>
<td>Tidal_scaled</td>
<td>Tidal (×1.45)</td>
<td>300</td>
<td>$1/3$</td>
<td>—</td>
<td>—</td>
<td>0.7</td>
</tr>
<tr>
<td>Tidal_Lee_Z900</td>
<td>Tidal + lee</td>
<td>300</td>
<td>$1/3$</td>
<td>900</td>
<td>1</td>
<td>0.7</td>
</tr>
<tr>
<td>Tidal_Lee_q03</td>
<td>Tidal + lee</td>
<td>300</td>
<td>$1/3$</td>
<td>300</td>
<td>$1/3$</td>
<td>0.6</td>
</tr>
</tbody>
</table>

Fluxes cannot be sustained and $\Gamma$ should be much smaller than 0.2, we replaced $\Gamma = 0.2$ in our model by

$$\Gamma = 0.2 \frac{N^2}{N^2 + \Omega^2},$$

where $\Omega$ is the angular velocity of Earth, as in Melet et al. (2013).

In addition to mixing due to the local dissipation of high-mode internal tides and lee waves, different parameterizations account for other diapycnal mixing mechanisms in the model. Shear-driven mixing is accounted for by a Richardson number–dependent parameterization (Jackson et al. 2008). The bottom boundary layer is mixed using the parameterization of Legg et al. (2006), and geothermal heating is included based on Adcroft et al. (2001). Finally, a background diapycnal diffusivity is added to account for other sources of mixing that are not specifically parameterized, such as the dissipation of radiating low-mode internal tides or near-inertial waves. This background diapycnal diffusivity is prescribed using a latitude-dependent profile taking a minimum value of $2 \times 10^{-6} \text{ m}^2 \text{s}^{-1}$ at the equator and a value of $2 \times 10^{-5} \text{ m}^2 \text{s}^{-1}$ at $30^\circ \text{N/S}$ (Harrison and Hallberg 2008). Finally, an additional ad hoc diffusivity is added based on the dissipation scheme proposed by Gargett and Holloway (1984):

$$\epsilon_G = 1.0 \times 10^{-7} \text{ W m}^{-3} + 6.0 \times 10^{-4} \text{ J m}^{-3} \times N,$$

where $N$ is the buoyancy frequency. A ceiling of $2.5 \times 10^{-6} \text{ W m}^{-3}$ is imposed on the dissipation of energy in (6). This last piece $\epsilon_G$ was introduced late in the development of CM2G to reduce some strong abyssal cold biases (Dunne et al. 2012); ideally, it would ultimately be supplanted by more physically consistent schemes, like the lee wave parameterization described here.

c. Sensitivity experiments

To study the sensitivity of the ocean state to lee wave–driven mixing, three main simulations and two additional sensitivity simulations were performed (Table 1). Each simulation is 1000 years long. The largely unknown fraction of internal wave energy that dissipates locally ($q_T$ and $q_L$) and vertical decay scales ($z_T$ and $z_L$) in the St. Laurent et al. (2002) formulation need to be prescribed for each simulation.

In the first main experiment, only internal tide–driven mixing is parameterized (experiment Tidal). In the St. Laurent et al. (2002) scheme, the turbulent dissipation rate is chosen to decay exponentially away from the topography with a fixed decay scale to match observations. A vertical decay scale in the range 300–500 m has been shown to fit observations from the Brazil Basin and Hawaiian Ridge (St. Laurent and Nash 2004). In standard CM2G simulations (Dunne et al. 2012), the decay scale for internal tide energy dissipation is set to 300 m to produce an overall better solution. Here, we retain this value for the decay scale of internal tides. To match observations from the Brazil Basin, St. Laurent et al. (2002) assumed a uniform and fixed value of $q_T = \frac{5}{3} q_L$; we retain this value in the following for internal tides. With this choice of parameters, internal tides provide 0.5 TW (the global integral of $q_T E_T$) of energy dissipation to support diapycnal mixing in the deep ocean.

In the second main experiment, the lee wave–driven mixing parameterization is added to that of the internal tide (experiment Tidal_Lee). Linear theory of internal waves predicts that lee waves radiate from topographic features with smaller wavelengths than those leading to internal tide generation (Bell 1975b). As a result, lee waves tend to have larger vertical wavenumbers than internal tides and are more prone to dissipate locally. Lee waves are also not likely to propagate far from their generation site and may be trapped inside the flow in which they were generated. Indeed, lee waves are radiated with the intrinsic frequency $\kappa_L U_f$, where $\kappa_L$ is the wave wavenumber fixed by topography, and $U_f$ is the mean flow velocity that can vary both in the vertical and horizontal. For lee waves to propagate, their intrinsic frequency must be greater than the Coriolis frequency $f$ and smaller than the buoyancy frequency $N$. During their propagation, when the Coriolis frequency changes or the mean current velocity decreases such that the
intrinsically frequency, lee waves encounter critical layers in the vertical or turning points in the horizontal. While mean flows are observed to decrease with height above bottom in some locations (e.g., Waterman et al. 2013), the equivalent barotropic velocity structure typical of the Southern Ocean suggests that the critical layers in the vertical are not common. On the other hand, lee waves are primarily generated by fronts and eddies of the ACC with horizontal scale of \( O(10–100\text{ km}) \) and hence likely encounter turning points and become trapped in the horizontal until they dissipate. While this dissipation may not be near the generation site, it likely takes place within \( O(10–100\text{ km}) \) from the generation site, determined by the horizontal scale of fronts and eddies. This horizontal scale is comparable to the grid scale of the climate model used in this study and hence the lee wave dissipation can be regarded as local. Therefore, we consider here for the main experiment, Tidal_Lee, the extreme case where all the energy going into internal lee waves is dissipated locally and supports diapycnal mixing by setting \( q_L = 1 \).

Using idealized numerical simulations of the Drake Passage region with a spatially uniform mean flow and Coriolis frequency (therefore not representing the trapping of lee waves discussed above), Nikurashin and Ferrari (2010) reported that up to 50% of the lee wave radiated energy is dissipated locally. However, in the region north of Kerguelen Plateau, where the topographic roughness is smaller than in Drake Passage and hence the lee wave generation process is more linear, observations suggest that the fraction of local energy dissipation can be as small as 0.1 (Waterman et al. 2013). Therefore, we performed a sensitivity experiment to the value of \( q_L \), using the same fraction of local dissipation as for internal tides, \( q_L = \frac{1}{3} \) (experiment Tidal_Lee_q03).

Regarding the vertical decay scale of lee waves’ energy dissipation, we use the same value as for internal tides in our main experiment Tidal_Lee. However, Sloyan (2005) and Naveira Garabato et al. (2004) show that in the Southern Ocean, observed mixing rates exceeding background values by more than an order of magnitude are common at depths greater than 500–1000 m and can extend throughout the entire water column at the Antarctic Polar Front and Subantarctic Front, where the ACC is stronger. Both studies suggest that the enhanced mixing rates are associated with the dissipation of lee waves. Therefore, we also performed a sensitivity experiment to the value of \( z_L \), using a longer decay scale for lee wave energy dissipation than for internal tides, with \( z_L = 900\text{ m} \) (experiment Tidal_Lee_Z900).

To assess the importance of the spatial distribution of tidal and lee wave energy compared to the importance of the total energy input into internal waves over the World Ocean, a third main simulation is performed. In this experiment, called Tidal_scaled, energy is input only at internal tide generation sites and has been rescaled by a factor \( R \) so that the global energy input is the same as in Tidal_Lee (where both internal tide and lee wave energies were injected according to their own spatial distribution). The scaling factor \( R \) of the internal tide energy input is

\[
R = \frac{q_T \int E_T \, dx \, dy + q_L \int E_L \, dx \, dy}{q_T \int E_T \, dx \, dy} = 1.45, \tag{7}
\]

where the double integral is a longitude–latitude integral over the World Ocean, and with the lee wave energy input (0.24 TW) being entirely dissipated locally (\( q_L = 1 \)) and a third (\( q_T = \frac{1}{3} \)) of the internal tide energy input (1.64 TW) being dissipated locally.

In the next section, impacts of lee wave–driven mixing on the ocean state are first assessed based on the Tidal, Tidal_Lee and Tidal_scaled simulations. The sensitivity of the results to the choice of lee wave energy dissipation parameters (experiments Tidal_Lee_q03 and Tidal_Lee_Z900) are discussed in section 3f.

3. Impacts of lee wave–driven mixing on the ocean state

After 1000 years of simulations, the deep ocean state is not yet fully equilibrated, but the drift has slowed significantly (the slope of the linear regression of the global-mean temperature below 3000-m depth for the last century of the Tidal simulation is less than 0.01°C century\(^{-1}\)). The drift of the model over time can be explained by the prescription of the year 1990 concentrations of atmospheric carbon dioxide and radiatively active aerosols and gases in our ocean–ice–atmosphere coupled simulations. Indeed, because the observed state of the ocean in 1990 that is used for the model initial conditions is not in balance with the 1990 radiative forcing, the prescription of this radiative forcing leads to a committed warming of the ocean over time. [By contrast, simulations with the same model that use preindustrial forcing in a spinup to quasi equilibrium as in the simulations used in this study, but then followed by historical forcing, are actually slightly colder than observed (Hallberg et al. 2013).] In the following, results are presented for the last century of the 1000-year simulations, but are similar to preceding centuries in the simulations.
a. Impacts on diapycnal diffusivities and stratification

As shown by (1), (2), and (4), diapycnal diffusivity can change when adding lee wave dissipation due to both the additional energy input $E_L$ and subsequent changes in stratification. We first examine the changes only due to the additional source of energy, assuming that the stratification is unchanged between the simulations. To do so, we consider the changes in diapycnal diffusivities after only 1 year of simulation. We verified that changes in stratification were negligible between the different simulations after this short interval.

The globally averaged vertical profiles of the buoyancy fluxes due to the internal wave–driven mixing parameterizations (1) and (2) for the three main simulations are shown in Fig. 2a. Note that because of the no-flux bottom boundary condition, handled through (5), the buoyancy flux induced by the internal wave–driven mixing first increases over the bottommost cell of the model and then decreases above in agreement with the exponential decay in the St. Laurent et al. (2002) formulation (Fig. 2a). As expected from (1), (2), and (4), adding the lee wave energy dissipation to the tidal dissipation leads to an increase of the buoyancy flux (Fig. 2a), which is mostly confined to the bottommost 2 km (Fig. 2b). When the spatial distribution of the lee wave energy flux is not taken into account (experiment Tidal_scaled), the increase in diapycnal diffusivities, and hence buoyancy fluxes, in the ocean interior is 45%. This is consistent with the value of the scaling factor $R$ of the energy flux in (4), because the energy flux has the same spatial distribution in the Tidal and Tidal_scaled simulations. However, lee wave and internal tide generation sites have a different spatial distribution, with energy being input in shallower regions for internal tides than for lee waves (Figs. 1, 3).

Because of the imposed energy conservation in mixing parameterizations (1) and (2), the same internal wave energy flux at the bottom results in higher rates of the energy dissipation throughout the water column in shallower regions than in deeper ones. The amplitude of the normalization factor of the vertical profile of the dissipation in (1) and (2) indeed depends on the local ocean depth and is larger in shallower regions. When globally averaging the profiles of energy dissipation or buoyancy flux in height above the bottom as done in Fig. 2, this leads to a globally stronger buoyancy flux near the ocean bottom in the Tidal_scaled simulation than in the Tidal_Lee simulation.

The ocean state, and notably its stratification, will adjust to the changes in diapycnal diffusivity and buoyancy fluxes due to the addition of lee wave energy dissipation. Changes in stratification would in turn impact the diapycnal diffusivity (4). After 1000 years of simulation, the diapycnal diffusivities are larger and the stratification weaker in the ocean interior below 1000 m due to the addition of lee wave energy dissipation (Figs. 4b and 5a). More precisely, the zonal global mean of the squared buoyancy frequency is decreased by up to 30% (Fig. 4b), and diapycnal diffusivities are particularly increased in the deep ocean between 40° and 60°S, and in the tropics, in agreement with the latitudinal distribution of the energy flux into lee waves (Fig. 1). Indeed, the spatial distribution of the lee wave energy dissipation partly explains the impacts on diapycnal diffusivity and stratification: when the increase in energy flux into the internal waves is taken into account but not its spatial distribution (experiment Tidal_scaled), the impacts are weaker (cf. Figs. 4b,c and 5b,c).

b. Impacts on the temperature field

The temporal evolution of temperature and salinity is impacted by diapycnal diffusivities. With the diapycnal diffusivity term in the primitive equations being expressed according to a Fickian diffusion formalism and under the assumption of a linear equation of state and identical diffusivities of heat and salinity, the temporal evolution of the density $\rho$ due to internal tide– and lee wave–driven diapycnal diffusivities $K_{d,T,L}$ is given by

$$\frac{\partial \rho}{\partial t} = \frac{\partial}{\partial z} \left[ K_{d,T,L} \frac{\partial \rho}{\partial z} \right] = -\frac{\rho_0}{g} \frac{\partial}{\partial z} [\Gamma (\epsilon_T + \epsilon_L)],$$

where $g$ is the gravity acceleration. Although the model uses a fully nonlinear equation of state, (8) is useful for understanding the model’s results.

Because of the no-flux bottom boundary condition, $\Gamma (\epsilon_T + \epsilon_L)$ converges in a bottom boundary layer and then diverges in the water column above, in agreement with the exponential decay of the dissipation. Therefore, the addition of lee wave dissipation (2) should lead over time to lighter water at the bottom of the ocean and denser water above.

Integrations of the model indeed show that differences of temperature between the Tidal_Lee and Tidal simulations are positive in the deep ocean and negative above (Fig. 6a). After 1000 years of simulations, the temperature field has reached a quasi equilibrium in the 500–1500-m layer where the addition of lee wave energy dissipation leads to a cooling (except in the Atlantic Ocean, where the results are less robust in this depth range compared to other basins because of the larger internal variability in this basin; Fig. 7a). The corresponding differences of temperature between the Tidal_Lee and Tidal simulations are shown in Fig. 6c.
cooling occurs over almost the entire ocean. Over the last century of the simulations, the basin-averaged differences reach \(-0.14^\circ\mathrm{C}\) over the Pacific and Southern Oceans, \(-0.10^\circ\mathrm{C}\) over the Indian Ocean, and are very small over the Atlantic Ocean (although a cooling is also observed in the southern Atlantic).

In the deep ocean, the temperature is still drifting slowly after 1000 years of simulations, but the drift has weakened considerably and the differences of temperature have reached a quasi equilibrium (except in the Pacific Ocean; Fig. 7b). The spatial distribution of the differences of temperature in the deep ocean induced by the addition of lee wave dissipation is shown in Fig. 6e. Warming of the deep ocean occurs over almost the entire ocean. It is stronger in the Atlantic (basin average of \(0.37^\circ\mathrm{C}\) over the last century of simulation) and Indian (\(0.30^\circ\mathrm{C}\)) Oceans than in the Pacific (\(0.24^\circ\mathrm{C}\)) or Southern (\(0.20^\circ\mathrm{C}\)) Oceans.

The spatial distribution of the internal tide and lee wave energy dissipation partly explains the sensitivity described above. Indeed, when the lee wave energy is distributed over the internal tide generation sites (experiment Tidal_scaled), the amplitude of the induced differences of temperature is weaker than when the lee wave energy is input at lee wave generation sites (Fig. 6b). The spatial distribution of the differences of temperature is also altered (Figs. 6d,f). Notably, relative to the Tidal simulation, a warming of the Southern Ocean in the 500–1500-m layer is simulated in Tidal_scaled instead of the general cooling simulated in Tidal_Lee.

c. Impacts on the MOC

The meridional overturning circulation corresponds to the surface formation of dense water in the polar
oceans, their sinking to the deep ocean, and their returning pathways back to the surface ocean. Simple conceptual models have been formulated over the years to link the MOC and stratification in the abyssal ocean to diapycnal upwelling, eddies, and wind forcing (e.g., Stommel and Arons 1960; Pedlosky 1992; Munk and Wunsch 1998; Gnanadesikan 1999; Nikurashin and Vallis 2012). The controlling physical mechanisms of the returning pathways have been long debated (adiabatic pathways through wind-driven upwelling or diabatic...
pathways through diapycnal mixing) but have now come into clearer focus (Marshall and Speer 2012; Talley 2013).

The global ocean MOC diagnosed in density space from the Tidal simulation is shown in Fig. 8. In the upper ocean ($\sigma_2 < 1035$), the main circulation corresponds to the wind-driven subtropical cells (McCreary and Lu 1994). Below, in the 1035–1036.9-$\sigma_2$ density layer, the upper cell of the MOC is found with a maximum of 17 Sverdrups (Sv; $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$). This cell mainly

![Figure 5](image-url)
corresponds to the circulation of North Atlantic Deep Water (NADW) formed in the Nordic seas and Labrador Sea. Below the $\sigma_2 = 1036.9$ isopycnal surface and in the Southern Ocean, the lower cell of the MOC is found. This cell mainly corresponds to the convection of Antarctic Bottom Water (AABW) formed around Antarctica. The circulations of NADW and AABW are intertwined (Talley 2013). The return pathways of NADW and AABW to their formation regions involve both a diabatic lightening in the ocean interior mainly through diapycnal mixing to form Pacific and Indian Deep Waters (PDW and IDW), an adiabatic wind–driven upwelling of PDW, IDW, and NADW, and a lightening through surface buoyancy fluxes in the Southern Ocean (Marshall and Speer 2012; Talley 2013).

The impact of adding the lee wave energy dissipation to that of the internal tide on the MOC is shown in Fig. 8. Lee wave–driven mixing mostly impacts the lower cell of the MOC (Fig. 8d): differences are mainly found below the $1036.8-\sigma_2$ isopycnal surface, except in the Southern Ocean, south of $40^\circ$S. The lower cell of the MOC is lighter and stronger ($+2$ Sv, which corresponds to an increase in intensity of 10%) in the Tidal_Lee simulation than in the Tidal one. This is consistent with the lower cell of the MOC being primarily closed by ocean mixing: stronger abyssal diapycnal mixing leads to stronger rates of AABW density transformation and a stronger AABW circulation. In contrast, only very small changes can be noted for the upper cell of the MOC corresponding to NADW circulation, which is consistent with it being...
closed primarily at the surface through wind-driven upwelling.

When energy is only input at the internal tide generation sites, the differences in the MOC are less pronounced than when energy is input both at the internal tide and lee wave generation sites (Figs. 8d,e), especially in the Southern Ocean, although a lightening of the lower cell of the MOC still occurs. The different patterns of diapycnal diffusivity change (Fig. 5) between the three simulations, with enhanced diapycnal diffusivities in the Southern Ocean and Indo-Pacific tropical band when lee wave energy is input at the lee wave generation site compared to when it is redistributed over the internal tide generation sites explain these differences in the MOC.

It can also be noted that AABW reaches higher latitudes in the Northern Hemisphere in the Tidal_scaled simulation compared to the Tidal_Lee one. This can be due to the different distribution of the energy into internal waves between the two simulations, with higher mixing levels in the Southern Hemisphere in the Tidal_Lee simulation (Fig. 1), resulting in higher erosion of AABW.

d. Impacts on the circulation

To document the changes in circulation induced by the addition of lee wave–driven dissipation in the model, the barotropic streamfunction (BSF) was computed (Fig. 9). The main changes in the vertically integrated circulation are found in the Southern Ocean. The ACC is weakened between the Polar Front [PF; determined here as the position of the 2°C isotherm at 200-m depth following Orsi et al. (1995)] and the Subtropical Front [STF; determined here as the position of the 10°C isotherm at 150-m depth following Orsi et al. (1995)]. The zonally averaged BSF differences are at a maximum around 45°S, where they reach ~9 Sv (corresponding to a weakening of 7% of the ACC, which is robust with regard to the centennial variability of the ACC in our simulations). However, the positions of the PF and STF, and therefore of the ACC, are only slightly impacted (Fig. 9b). Differences in the BSF are significantly weaker, especially in the Southern Ocean, between the Tidal_scaled and Tidal simulations than between the Tidal_Lee and Tidal simulations (Figs. 9c,d). Therefore, the spatial distribution of the lee wave energy dissipation, with half
Fig. 8. Global MOC in density ($\sigma_2$) coordinates for years 901–1000 in the (a) Tidal, (b) Tidal_Lee, and (c) Tidal_scaled simulations. Differences of the global MOC between the (d) Tidal_Lee and Tidal simulations and (e) Tidal_scaled and Tidal simulations.
of the energy flux accounted for by the Southern Ocean, is mainly responsible for the changes in circulation observed in the Southern Ocean.

In the Southern Ocean, strong westerly winds drive an equatorward Ekman transport in the surface layers. The resulting deepening of the isopycnal surfaces drives the ACC through geostrophic balance. Changes in diapycnal diffusivities and stratification due to the addition of lee wave energy dissipation impact the density field, which in turn impacts the strength of the ACC. In our simulations, isopycnal slopes are steeper south of 50°S in Tidal than in Tidal_scaled and Tidal_Lee (Fig. 10). Therefore, stronger diapycnal diffusivities in the Southern Ocean (Fig. 5) lead to flatter isopycnal surfaces (Fig. 10) and weaker ACC transport in our simulations (Fig. 9). However, this result may not be robust because the model used in the present study does not explicitly resolve baroclinic eddies, and the parameterized eddy response in the ACC may not be as strong as the eddies in the real ocean should be (Hallberg and Gnanadesikan 2006).

In a series of sensitivity experiments, Jayne (2009) also found a relationship between the mean transport through the Drake Passage and the power dissipation in the abyssal ocean. He suggested that the strength of the ACC is related to the strength of the abyssal circulation. Gent et al. (2001) proposed a mechanism where the strength of the ACC is strongly related to the magnitude of the southward transport in the intermediate layer of the ocean at the latitude of the Drake Passage. They suggested that the meridional Ekman transport drives about 100 Sv of the Drake transport, and the global thermohaline circulation drives approximately 30 Sv. The latter transport is more directly impacted by the diapycnal diffusivity. In Jayne (2009), the Drake Passage transport decreases when the abyssal power dissipation increases. Our results tend to confirm this relationship.

e. Impacts on the ocean ventilation

To study the ocean ventilation in climate models, a tracer known as ideal age (Thiele and Sarmiento 1990) is diagnosed. This tracer is set to zero in the mixed layer and ages at a rate of 1 yr yr⁻¹ thereafter. At depth, ideal age is therefore small (young) in regions of deep convection and large (old) in quiescent regions such as the Pacific subtropical gyres (Figs. 11a,d). Ideal ages in climate models are a function of the duration of the simulation when the latter is shorter than several thousands of years. Indeed, the time scale of ventilation of the oldest water in the ocean is roughly 1500 years (e.g., England 1995). However, the sensitivity of ideal age to the addition of lee wave–driven mixing gives some insight to the impact of lee wave–driven mixing on the ventilation of water masses. The differences of ideal age between the Tidal_Lee and Tidal simulations show that lee wave–driven mixing impacts the ventilation of the deep ocean (Figs. 11b,e). In particular, ventilation is increased in the (tropical) Pacific Ocean, where lee wave–driven mixing leads to younger water, whereas older water is found in the Southern Ocean in the Tidal_Lee simulation. These differences in ventilation are consistent with the strengthening of the lower cell of the MOC, which implies more upwelling of deep young water in the Pacific and a stronger return flow to the Southern Ocean that tends to reduce the gradient of ideal age between the tropics and the Southern Ocean (Fig. 11).

f. Sensitivity to the lee wave energy dissipation parameters

The sensitivity of the previous results to our choice of \( q_L \) and \( z_L \) are shown in Fig. 12. Results are more sensitive to a reduction of \( q_L \) (and therefore of the energy dissipated by internal waves; Table 1) than to an increase of \( z_L \). When \( q_L \) is reduced from 1 to \( \frac{1}{3} \), the sensitivity of the ocean state to lee wave–driven mixing is qualitatively the same but is weaker. The global-mean abyssal temperature warming is 0.1°C when only a third of the lee wave energy is dissipated locally, compared to 0.3°C when the lee wave energy is entirely dissipated locally (Fig. 12a, black, red, and orange symbols). The ideal age of the deep tropical water and Drake Passage transport are also less reduced by lee wave–driven mixing when \( q_L \) is set to \( \frac{1}{3} \) instead of 1 (Figs. 12b,c). Finally, the strength of the abyssal MOC is also impacted by the choice of the parameter \( q_L \) (Fig. 12d). Note that the simulation using \( q_L \) equal to a third is not directly comparable to the Tidal_scaled simulation because the global energy dissipated by internal waves is different in these two simulations (Table 1).

Increasing the vertical decay scale for the lee wave energy dissipation from \( z_L = 300 \) m to \( z_L = 900 \) m only slightly changes the World Ocean abyssal temperature and deep tropical water ideal age (Figs. 12a,b; red and purple symbols). However, the large-scale circulation is sensitive to an increase of \( z_L \): the ACC transport is slightly less weakened (Fig. 12c; compare the black, red, and purple symbols), and the meridional overturning circulation of AABW within the Southern Ocean is stronger (Fig. 12d; triangles), while the basin-scale meridional overturning circulation of AABW is less enhanced (Fig. 12d; squares). Therefore, the sensitivity of the ocean state to the lee wave energy dissipation is qualitatively robust, but the amplitude of the impacts depends on the choice of lee wave energy dissipation parameters.
FIG. 9. Barotropic streamfunctions for years 901–1000 (referenced at the North Pole) for the (a) Tidal_Lee (red lines) and Tidal (black lines) simulations and their differences (colors) and for the (c) Tidal_scaled (blue lines) and Tidal (black lines) simulations and their differences (colors). Units are Sv and contours are every 20 Sv. The positions of the Polar Front and of the Subtropical Front are shown in (b) and (d) for the Tidal (black lines), Tidal_Lee (red lines), and Tidal_scaled (blue lines) simulations. The color bar is the same for all panels.
4. Discussion and conclusions

Diapycnal mixing plays a key role in maintaining the ocean stratification and meridional overturning circulation. In the ocean interior, diapycnal mixing is mainly sustained by breaking internal waves. Currently, most climate models only explicitly parameterize the local dissipation of internal tides, using the semiempirical scheme of St. Laurent et al. (2002). Although internal tides probably are the most energetic internal waves in the ocean interior, other classes of internal waves contribute to diapycnal mixing. Among them are lee waves, which are generated by the interaction of geostrophic motions with the topography in the stratified ocean. The global energy input into lee waves represents a sizable fraction of the energy input into the internal tides, and the spatial distribution of the energy flux into internal tides and lee waves is strikingly different: lee waves are mostly generated in the Southern Ocean where the energy conversion into internal tides is weak.

In this study, we investigated the impact on the ocean state of adding lee wave energy dissipation to that due to internal tides. To do so, we performed long-term (1000 year) simulations of the ocean–ice–atmosphere GFDL CM2G including a parameterization of the local dissipation of 1) internal tides only (experiment Tidal), 2) internal tides and lee waves (experiment Tidal_Lee), and 3) internal tides with the energy flux amplitude rescaled so that the global energy is the same as in the simulation accounting for internal tides and lee waves (experiment Tidal_scaled). The parameterization of lee wave–driven mixing is based on the St. Laurent et al. (2002) scheme, using a recently estimated global map of energy conversion into lee waves (Nikurashin and Ferrari 2011). By comparing the Tidal and Tidal_Lee simulations, we show that although the global energy input into lee waves is small compared to that into internal tides, lee wave–driven mixing makes a significant impact on the ocean state. The main response of the model to lee wave–driven mixing is a reduction of the stratification of the ocean by up to 30%. This is associated with a warming of the abyssal ocean by 0.2°C–0.3°C. The lower cell of the MOC is also impacted, with a lightening and a 10% increase of its strength. These results are consistent with previous studies of diffusivity sensitivity in ocean general circulation models (Bryan 1987; Simmons et al. 2004; Saenko and Merryfield 2005; Jayne 2009). The increase of the lower cell of the MOC due to the addition of lee wave–driven mixing is consistent with a partly diabatically driven return flow. A recent estimate of the water mass transformation by internal wave–driven mixing in the deep ocean reports that the lee wave–driven mixing can drive up to 10 Sv of the lower cell of the MOC (Nikurashin and Ferrari 2013). In our study, however, the magnitude of the MOC response to the lee wave–driven mixing is lower. Although the two studies address a similar question of the impact of lee waves on the large-scale ocean circulation, they are different and complement

![Fig. 10. Depth of the isopycnal surfaces (σ₀: kg m⁻³) zonally averaged over the Southern Ocean for years 901–1000 for the Tidal (black lines), Tidal_Lee (red lines), and Tidal_scaled (blue lines) simulations.](image)
each other. Nikurashin and Ferrari (2013) estimated the rate of the water mass transformation by the lee wave-driven mixing for the present state of the ocean using the stratification observed in the ocean. In their study, internal wave-driven mixing can only impact the MOC because the stratification is set to be static. In contrast, in our study we use a climate model to estimate the response of the ocean state, including its stratification, to the
FIG. 12. Sensitivity to the lee wave energy dissipation parameters. (a) Global ocean 4000–6000-m mean of temperature (°C) for the last century of the simulations. (b) Ideal age averaged between 5°N and 5°S and 1500–3500-m depth for the last century of the simulations. (c) Transport through Drake Passage averaged over the last 500 years of the simulations. (d) Triangles: Strength of the abyssal limb of the meridional overturning circulation (Sv) south of 55°S (corresponding to the recirculation of AABW in the Southern Ocean) as a function of the internal wave energy dissipation (TW) integrated over the ocean south of 40°S (defining here the northern boundary of the Southern Ocean). Squares: Strength of the abyssal limb of the meridional overturning circulation (Sv) north of 55°S (corresponding to the basin-scale recirculation of AABW) as a function of the internal wave energy dissipation (TW) integrated over the ocean north of 40°S.
addition of the lee wave–driven mixing. Our results show that the main response of the model to lee wave–driven mixing is not as much to increase the lower cell of the MOC as to warm the abyssal ocean and reduce its stratification. Moreover, in long simulations, stratification in climate models is known to drift away from the observed values. Hence, to compare model results to those in Nikurashin and Ferrari (2013), the model would need to be retuned to bring its stratification closer to the one observed in the ocean. In our study, the vertically integrated circulation is also slightly impacted by lee wave–driven mixing, notably in the Southern Ocean that accounts for half the lee wave energy flux. Finally, we show that the different spatial distribution of the internal tide and lee wave energy input is largely responsible for the sensitivity described in this study (by comparing the Tidal_scaled and Tidal_Lee simulations, where the total energy flux into internal waves is similar, but its spatial distribution is different). The importance of the patchiness and spatial distribution of internal wave–driven mixing highlighted in the present study is consistent with results from previous studies on the spatial distribution of internal tide–driven mixing (e.g., Simmons et al. 2004; Jayne 2009; Friedrich et al. 2011). Therefore, our modeling results suggest that lee wave–driven mixing, dominated by the Southern Ocean, impacts the ocean state in climate models and should be explicitly parameterized in such models.

While we focused on lee wave–driven mixing in this study, lee waves could also impact the ocean state through drag extraction of momentum and vorticity from the general circulation. Indeed, lee wave drag has recently been identified as a significant player in the dynamics of extensive areas of the ocean, including the Antarctic Circumpolar Current (Naveira Garabato et al. 2013). Parameterizing this other lee wave–related process may also lead to significant changes in the deep ocean circulation and energy budget (Trossman et al. 2013).

The parameterization used in this study for internal wave–driven mixing corresponds to the commonly used St. Laurent et al. (2002) semiempirical scheme and is quite crude. In particular, we used fixed, uniform values of the fraction of local dissipation and of the vertical decay scale of the energy dissipation. However, additional experiments where these parameters are varied show that the impact of lee wave–driven mixing on the ocean state is sensitive to the values of the vertical decay scale and fraction of local lee wave energy dissipation. Physically, such parameters should be a function of the ocean state and evolve in time and space. Moreover, we used the same ad hoc formulation for the vertical distribution of internal tides’ and lee waves’ energy dissipation.

In reality, the vertical distribution of lee wave energy dissipation may depend on the vertical structure of their generating currents through the encountering of critical layers where the waves could no longer propagate vertically. Uncertainties also exist in the energy fluxes that go into internal tides and lee waves. Regarding internal tides, the global energy flux in our simulations is 1.7 TW, with 1.4 TW occurring in the ocean deeper than 1000 m. The generation of internal tides in the deep ocean might be too energetic in our simulations: using a barotropic tidal model, Jayne and St. Laurent (2001) estimated that 2.0 TW is scattered by the conversion of the barotropic tide into internal waves, with just 1.0 TW occurring in the ocean deeper than 1000 m. These values are in quantitative agreement with the observationally based estimate of Egbert and Ray (2000), who suggest that 1 ± 0.3 TW of tidal power is dissipated in the deep ocean. As for the energy flux into lee waves, we used the static estimate of Nikurashin and Ferrari (2011) in this study, which is consistent with the other available estimate of Scott et al. (2011) regarding the spatial distribution of the energy flux. However, the amplitude of the generation rate of lee waves integrates globally to 0.2 TW in the Nikurashin and Ferrari (2011) estimate, while it is almost twice as large in Scott et al. (2011). Moreover, both estimates might be biased high because they notably use modeled abyssal geostrophic flow that does not account for the steering of the bottom flow by the unresolved topography. Finally, the energy flux into lee waves should be diagnosed internally in the model, so that it evolves in space and time with the ocean state. Because of these uncertainties, our modeling conclusions should be regarded as qualitative rather than quantitative. However, the objective of this study was not to examine the most realistic case, but to conduct a series of sensitivity experiments to assess the impact of lee wave energy dissipation on the ocean state in a climate model as a first step.

Recently, Saenko et al. (2012) added a parameterization of diapycnal mixing due to eddies in an OGCM, using the same semiempirical scheme that is used here. In their study, the eddy-driven diapycnal mixing energy input corresponds to the map of eddy energy sink derived from Zhai et al. (2010), with most of the dissipation occurring along the western boundaries and partly in the Southern Ocean (Saenko et al. 2012, their Fig. 1). They show that combined together, eddy- and tidally induced mixing impact the thermal structure of the ocean, the deep stratification, and meridional overturning circulation. The energy sink amplitude integrates to around 0.2 TW globally for both internal lee waves and western boundary energy loss (Zhai et al. 2010), but the geographical distribution of the energy input into internal waves is strikingly different between these two studies.
In the lee wave case, the distribution of the energy flux is largest in the Southern Ocean, which accounts for half of the total energy flux, and is weak along the western boundaries. Our work is therefore complementary to theirs.

This work is one component of the ongoing Internal Wave-Driven Mixing Climate Process Team (CPT) collaboration. Ultimately, physically based parameterizations for the major processes contributing to ocean mixing through dissipation of internal waves will be added in climate models as part of this CPT collaboration to replace ad hoc mixing prescription [such as the Gargett and Holloway (1984) scheme] that are currently used and account for internal wave–driven mixing processes that are currently not explicitly parameterized. Processes targeted by this CPT include the dissipation of high- (Melet et al. 2013) and low-mode internal waves (Waterhouse et al. 2013, manuscript submitted to J. Phys. Oceanogr.) due to scattering at topography (Legg 2014), parametric subharmonic instability (PSI) at critical latitudes (MacKinnon and Winters 2005), nonlinear wave–wave interactions (Nikurashin and Legg 2011), but also dissipation of near-inertial waves (Jochum et al. 2013) and of internal waves at tall/steep topography (Legg and Klymak 2008; Klymak et al. 2010). For credible simulations of a changing climate, internal wave–driven mixing needs to be represented by physically based parameterizations in ocean models to allow mixing to evolve in space and time depending on the ocean state. Polzin (2009) has formulated a new parameterization of internal tide dissipation based on analytical solutions to a radiation balance equation, implemented in a climate model in Melet et al. (2013). The same approach is currently being used to infer a more physically based parameterization of lee wave energy dissipation (K. Polzin 2013, personal communication) that could be implemented in climate models.

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