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Vertical heat transport in eddying ocean models

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[1] The effect of mesoscale eddies on the vertical heat transport of the ocean is examined using two eddy-resolving numerical models. The global heat transport by the mean flow and diffusion are both downwards and are balanced by an upward eddy heat flux. Mean and eddy advective heat fluxes dominate the subpolar regions, while diffusive flux is important primarily in the subtropics. In the subtropical abyss, the mean advective heat flux is balanced by a combination of eddy and diffusive fluxes and the classical Munk-type advective-diffusive heat balance must be modified. The Munk and Wunsch (1998) expression for the vertical turbulent diffusivity over-estimates the diffusivity by as much as a factor of four near the base of the main thermocline. This implies that the mixing required to close the meridional overturning circulation determined by Munk and Wunsch (1998) may be an over-estimate due to the neglect of mesoscale eddies. **Citation:** Wolfe, C. L., P. Cessi, J. L. McClean, and M. E. Maltrud (2008), Vertical heat transport in eddying ocean models, *Geophys. Res. Lett.*, *35*, L23605, doi:10.1029/2008GL036138.

1. Introduction

[2] In a steady-state ocean, the heat flux across any horizontal control surface is zero (neglecting the very small fluxes associated with geothermal heating and viscous dissipation). A critical question is how different dynamical processes contribute to this zero sum. Interest in this question started with *Munk* [1966], who estimated the abyssal turbulent diffusivity by balancing the turbulent diffusion of heat (and other tracers) against advection by the time-mean flow. The energy required to sustain abyssal turbulent dissipation consistent with the estimated diffusivity of 10^{-4} m^2 s^{-1} exceeds that available to the general circulation through wind work on geostrophic currents. This discrepancy lead *Munk* [1966] to conclude that the tides may be an important energy source driving the North Atlantic meridional overturning circulation (MOC). The *Munk* [1966] study has been refined by *Munk and Wunsch* [1998, hereinafter MW], but the essential assumptions and conclusions are unchanged.

[3] While there is some evidence that diffusivities below 3000 m may approach 10^{-4} m^2 s^{-1}, numerous measurements support the conclusion that diffusivities at the base of the main thermocline (between 1000 m and 3000 m depth) are smaller than that required by the MW balance by nearly a factor of ten [*Kunze et al.*, 2006]. The problem of determining how the upwelling branch MOC is maintained in the face of the apparently insufficient mixing has been termed the 'missing mixing' problem. Several solutions to this problem have been proposed: MW and others have suggested that mixing 'hot spots' near boundaries and rough topography may sufficiently numerous and vigorous to raise the horizontally averaged diffusivity to the required level. Another possibility is that the bulk of the deep water upwells adiabatically in the southern ocean, reducing the need for diabatic mixing in the subtropics [*Toggweiler and Samuels*, 1995]. Currently, the evidence for these solutions is not conclusive and it is worth exploring alternative mechanisms.

[4] One such alternative is that the sub-thermocline heat balance is altered by the presence of mesoscale eddies. This note examines this hypothesis by studying the global oceanic vertical heat budget with the effects of mesoscale eddies explicitly considered. The vertical heat budget in low-resolution ocean models has been considered by *Gregory* [2000] and *Gnanadesikan et al.* [2005], who found that globally integrated heat advected by the mean flow is downwards and is opposed by sub-gridscale processes (including parameterized convection and parametrized mesoscale eddies). A downward heat transport by the mean flow is inconsistent with a global advection-diffusion balance, but it remains unclear how resolved versus parametrized eddies alter the heat budget and how different processes at low and high latitudes combine to produce the global heat flux discussed by *Gregory* [2000] and *Gnanadesikan et al.* [2005].

[5] Since the vertical heat transport is virtually impossible to measure directly on a global scale, we examine the vertical heat flux using two eddy-resolving global circulation models: the MITgcm and POP models.

2. The Models

[6] The MITgcm is the Massachusetts Institute of Technology General Circulation Model [e.g., *Hill et al.*, 1999], configured in a flat-bottomed, equatorially centered, rectangular 2400 km × 9800 km domain with a zonally periodic channel which extends to the bottom (at 2400 m) and occupies the southernmost 1200 km. The horizontal grid is 5.4 km, so that mesoscale eddies are resolved in this idealized domain. The wind and thermal forcing are steady and buoyancy is a linear function of temperature only. The domain, forcing, and numerics are identical to those used by *Wolfe and Cessi* [2008] in their 'cold pole' configuration, except that the diffusivity has been reduced to $\kappa_v = 4.9 \times 10^{-5}$ m^2 s^{-1}. This model was integrated for 275 years, by which time it had achieved near statistical equilibrium. The results shown below are averaged over the last 15.2 years of the simulation. Significantly longer or shorter averaging periods do not lead to significant changes in the results.

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[7] The POP model is the 0.1° Parallel Ocean Program eddy-permitting global simulation described by *Maltrud and McClean* [2005]. While POP uses a nonlinear equation of state, we focus on the heat budget rather than the buoyancy budget for consistency with the MITgcm results and because temperature is closer to equilibrium than salinity. The background vertical diffusivity is 10^{-5} m^2 s^{-1} above 2000 m depth and increases smoothly to 10^{-4} m^2 s^{-1} below 3000 m. POP uses the KPP vertical mixing scheme [*Large et al.*, 1994] which parameterizes several processes—such as turbulent boundary mixing, shear instability, and convection—that may alter the local diffusivity from its background value, but these effects are generally small away from a thin surface layer. The results below are averages over model simulation years 1999–2001. Due to drift and time-dependence in the forcing the POP model is not in statistical equilibrium below the main thermocline. However, given the close similarity between the MITgcm and POP results, we believe it unlikely that the qualitative features of the results (in particular, the signs of the heat flux terms) would change if POP were closer to equilibrium.

3. Heat Transport

[8] Following *Paparella and Young* [2002], we integrate the time-mean heat budget from the bottom $-H$ to a reference level z and over an arbitrary horizontal subdomain $\mathcal{A}$ to find

$$\int_{-H}^{z}\int_{\mathcal{A}}\frac{\Delta\mathcal{H}}{\Delta\tau}\,dA\,dz' + \int_{\mathcal{A}}\overline{w\mathcal{H}}\,dA + \int_{-H}^{z}\oint_{\partial\mathcal{A}}\overline{\boldsymbol{u}_H\mathcal{H}}\cdot\hat{\mathbf{n}}\,d\sigma\,dz' = 0, \quad (1)$$

where $\mathcal{H} = \rho_o c_p \theta$ is the specific heat content, θ is the potential temperature, and $\overline{\cdot}$ is the temporal average over the time interval $\Delta\tau$. dA and $d\sigma$ are the differential area and length of $\mathcal{A}$ and its boundary $\partial\mathcal{A}$, respectively. The boundary condition $\boldsymbol{u}\mathcal{H}\cdot\nabla H = 0$ has been used, so the third term is nonzero only above topography.

[9] The heat flux can be decomposed into a sum of fluxes resulting from different physical processes:

$$\overline{\boldsymbol{u}\mathcal{H}} = \overline{\boldsymbol{u}}\overline{\mathcal{H}} + \overline{\boldsymbol{u}'\mathcal{H}'} + \overline{\boldsymbol{u}_t\mathcal{H}_t} + \overline{\boldsymbol{u}_c\mathcal{H}_c}, \quad (2)$$

where the terms on the right-hand-side represent the heat flux due to the time-mean flow, $\overline{u}$, the deviation from the time average, $\boldsymbol{u}'$ (i.e., eddies), parameterized turbulent diffusion, $\boldsymbol{u}_t$, and convection, $\boldsymbol{u}_c$. In POP, the convection term also includes contributions from the other processes parameterized by the KPP scheme.

[10] Insertion of (2) into (1) gives

$$\mathcal{T} + \mathcal{M} + \mathcal{E} + \mathcal{D} + \mathcal{C} = 0, \quad (3)$$

where the tendency $\mathcal{T}$ is the first term in (1) and heat transport by the mean flow through the boundary of the domain is

$$\mathcal{M} \equiv \int_{\mathcal{A}}\overline{w}\overline{\mathcal{H}}\,dA + \int_{-H}^{z}\oint_{\partial\mathcal{A}}\overline{\boldsymbol{u}}_H\overline{\mathcal{H}}\cdot\hat{\mathbf{n}}\,d\sigma\,dz. \quad (4)$$

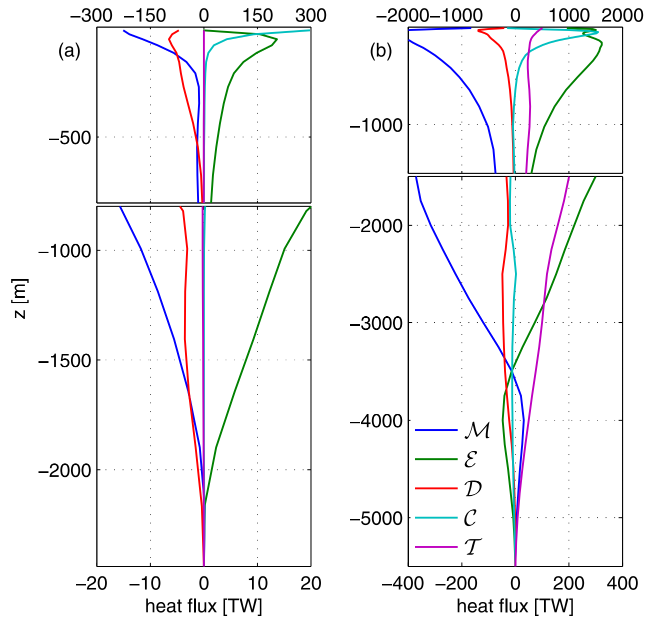

Figure 1. Global vertical heat flux due to mean advection $\mathcal{M}$, eddy advection $\mathcal{E}$, diffusion $\mathcal{D}$, and convection $\mathcal{C}$ for the (a) MITgcm and (b) POP models. Also shown is the tendency term $\mathcal{T}$. The scale has been expanded below the double horizontal line to show detail at depth.

The second term on the rhs of (4) vanishes if $\mathcal{A}$ is the whole domain. Expressions for $\mathcal{E}$ and $\mathcal{C}$ are defined in a manner similar to (4). If $\mathcal{A}$ is the global domain, then $\mathcal{M}$, $\mathcal{E}$, and $\mathcal{C}$ represent the vertical heat flux due to the mean flow, eddy flow, and convection, respectively, and $\mathcal{D} \equiv -\int_{\mathcal{A}}\kappa_v\overline{\mathcal{H}}_z\,dA$ parametrizes turbulent diffusion; otherwise, $\mathcal{A}$ is a subdomain and the terms in (3) represent the volume integrated heat flux divergence due to the same processes.

[11] When $\mathcal{A}$ is the global domain, both $\mathcal{M}$ and $\mathcal{D}$ are negative at all levels in the MITgcm, representing a downward flux (Figure 1a). Thus the mean circulation, $\overline{\boldsymbol{u}}$, is a net source of potential energy. To balance the heat budget, the eddy fluxes $\mathcal{E}$ are large and positive throughout the water column. Though POP shows significant thermal drift at depth, it has the same pattern of negative $\mathcal{M}$ balanced primarily by $\mathcal{E}$ above 3500 m depth—approximately the sill depth of the Antarctic Circumpolar Current (ACC). $\mathcal{C}$ is negligible below a thin surface layer in both models.

[12] We can further probe the heat budget by partitioning the vertical heat flux into contributions from the southern subpolar (SSP) region, the subtropics/tropics, and the northern subpolar (NSP) region. In the MITgcm, the SSP and NSP regions are defined as the regions of Ekman suction and the subtropics/tropics are the remainder of the domain. Following MW, we define the subtropics/tropics in POP as the region from 40°S to 48°N. The SSP region is defined as everything south of 40°S and the NSP region is taken to be between the latitudes 48°N and 70°N. The Arctic Ocean makes a negligible contribution to the global heat budget and is excluded.

[13] In both POP and the MITgcm, the largest contribution to the advective heat flux comes from the SSP region (Figures 2a and 2d). Here, $\mathcal{M}$ is strongly negative (downward) and almost entirely balanced by $\mathcal{E}$; $\mathcal{D}$ is

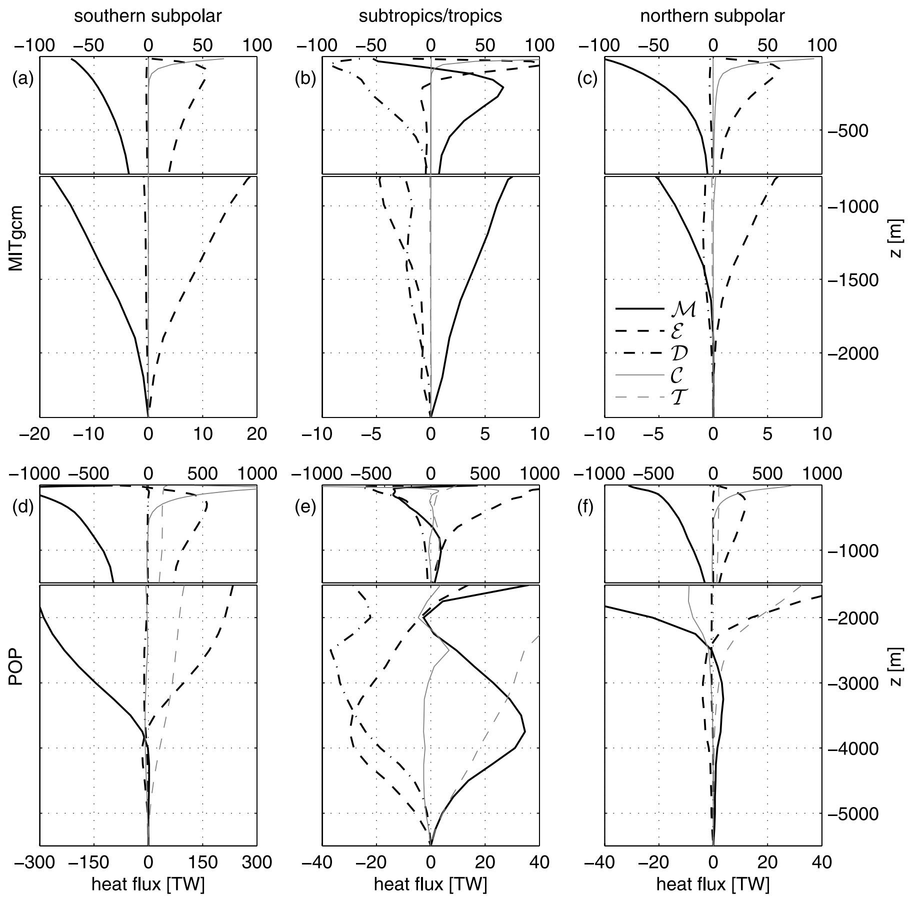

Figure 2. As in Figure 1, but showing the regionally integrated heat flux divergence. The columns show the heat budget in the southern subpolar region, the subtropics (and tropics), and the northern subpolar region, respectively, while the rows show the results for the MITgcm and POP models, respectively. Note that the range of the abscissa is larger at depth for the southern subpolar gyre than for the other two basins.

negligible. The SSP region thus satisfies the balance $\mathcal{M} + \mathcal{E} \approx 0$ advanced in recent theories of the southern ocean stratification [e.g., *Karsten et al.*, 2002; *Cessi et al.*, 2006; *Hallberg and Gnanadesikan*, 2006].

[14] $\mathcal{M} < 0$ in the SSP region is a consequence of the geometry and wind forcing of the southern ocean. The reentrant geometry of the channel region makes geostrophically balanced meridional flow impossible. Water is driven equatorward at the surface by the Ekman circulation and returned poleward in the bottom boundary layer (MITgcm) or below the level of the topography (POP). This circulation tilts the isopycnals toward the vertical, creating available potential energy which is released by baroclinic eddies. In the MITgcm, both the channel and the region where $\mathcal{M} < 0$ extend to the bottom of the domain. POP has realistic topography and the channel is unblocked only in the upper 3500 m of the water column. Below this depth, a mean, geostrophically balanced meridional flow is possible and $\mathcal{M}$ and $\mathcal{E}$ reverse sign.

[15] The wind forcing in the NSP region is similar to that in the SSP region, but meridional boundaries limit the penetration of the $\mathcal{M} < 0$ region to approximately 1500 and 2000 m depth in the MITgcm and POP models, respectively (Figures 2c and 2f). Below this depth, $\mathcal{M}$ is positive and the diffusive fluxes become more important.

[16] Diffusive fluxes in the NSP and SSP regions are small and the global structure of $\mathcal{D}$ is determined by the subtropics (Figures 2b and 2e). Consistent with *Munk* [1966], $\mathcal{M} > 0$ below the subtropical thermocline. Unlike the subpolar regions, a MW-type balance is possible in the subtropics, though it is modified by baroclinic eddies which cancel a large part of the mean advective flow since $\mathcal{E} < 0$. The eddy heat flux is thus thermally indirect in the deep subtropics and acts as a source rather than a sink of available potential energy (APE). Similar upgradient eddy heat fluxes have been observed in the Gulf Stream [*Wunsch*, 1999].

4. Estimating the Turbulent Diffusivity, κ_v

[17] The MW balance is obtained by assuming that the circulation is statistically steady ($\mathcal{T} = 0$) and that, when averaged over the subtropics and tropics, (3) is approximated by $\mathcal{M} + \mathcal{D} \approx 0$. The goal is an estimate, $\hat{\kappa}_v$, of the true turbulent diffusivity, κ_v. To this end, MW assume that the isotherms are approximately flat throughout the subtropics and that $\overline{w}$ is spatially uncorrelated with $\overline{\theta}$. Under these assumptions, the averaged heat budget reduces to

$$\left(A\langle\overline{\theta}_z\rangle\right)_z\langle\overline{w}\rangle \approx \left(\hat{\kappa}_v A\langle\overline{\theta}_z\rangle\right)_z \qquad (5)$$

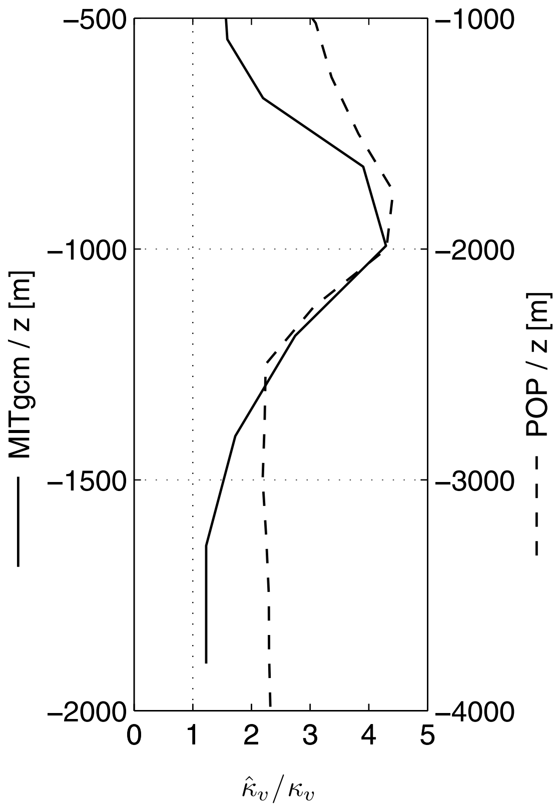

Figure 3. The estimated diffusivity $\hat{\kappa}_v$ relative to the background diffusivity κ_v in the subtropics as a function of depth z for the MITgcm (solid, left ordinate) and POP (dashed, right ordinate) models.

where $\langle \cdot \rangle = A^{-1} \int_A (\cdot)\, dA$ and $A(z)$ is the horizontal area of the subdomain A at level z. This equation—which is equivalent to MW's equation (3.6) for vertically uniform A—is a linear ODE in the estimate $\hat{\kappa}_v$. Using the MW's boundary condition that $\hat{\kappa}'_v\ (z_0) = 0$ for some z_0, the expression for $\hat{\kappa}_v$ is

$$\hat{\kappa}(z) = \frac{A\theta_z}{(A\theta_z)_z}\bigg|_{z_0} \frac{A(z_0)\theta_z(z_0)}{A(z)\theta_z(z)} w(z_0) + \int_{z_0}^{z} \frac{A(z')\theta_z(z')}{A(z)\theta_z(z)} w(z')\, dz', \tag{6}$$

where the angle brackets and overbars have been suppressed for clarity. MW considered this to be a reasonable expression for $\hat{\kappa}_v$ for $z < z_1$, where z_1 is the bottom of the main subtropical thermocline ($z_1 \approx -1000$ m), and "above the deepest waters," which they took to begin at $z_0 \approx -4000$ m.

[18] POP has realistic bathymetry, so we follow MW and pick $z_0 = -4000$ m and $z_1 = -1000$ m. There is no unambiguous way to determine z_0 in the MITgcm model in a manner consistent with MW; we have chosen $z_0 = -2000$ m. The thermocline is shallower in the MITgcm than in POP, so we take $z_1 = -500$ m.

[19] The MW estimate $\hat{\kappa}_v$ agrees with the MITgcm κ_v at great depths and near 500 m depth (Figure 3a), but departs significantly from κ_v just below the base of the main thermocline ($\approx$1000 m) where $\mathcal{E}$ is large and negative and cancels with $\mathcal{M}$ (see Figure 2). The maximum values of the estimated diffusivity approach the canonical 10^{-4} m^2 s^{-1} of MW.

[20] In POP, $\kappa_v(z)$ below 2500 m is large (10^{-4} m^2 s^{-1}), so even though $\mathcal{E}$ is large and negative, $\mathcal{D}$ is of the same magnitude (Figure 2). Consequently, $\hat{\kappa}_v(z)$ is within a factor two of κ_v below 2500 m (Figure 3b). The error in $\hat{\kappa}_v(z)$ increases to a factor of four as $\kappa_v(z)$ decreases at mid-depth (near $z = -2000$ m) where the maximum of the MOC is located [see *Maltrud and McClean*, 2005, Figure 2]. At this depth, $\mathcal{M}$ and $\mathcal{E}$ drop to nearly zero and $\mathcal{D} + \mathcal{T} \approx 0$.

5. Summary and Conclusions

[21] Because $\mathcal{M}$ and $\mathcal{D}$ are both negative throughout most of the water column, the eddies are qualitatively important in determining the global mean flow: the heat budget could not be closed if the eddies were removed without either a qualitative change in the mean circulation or an additional vertical heat transport process. This result confirms that the conclusions of *Gregory* [2000] and *Gnanadesikan et al.* [2005] hold when eddies are resolved rather than parameterized and calls into question the validity of global advection-diffusion models sometimes used in climate studies [e.g., *Raper et al.*, 2001].

[22] The structure of $\mathcal{M}$ and $\mathcal{E}$ is dominated by the southern and, to a lesser extent, northern subpolar regions. The controlling influence of the southern subpolar region (i.e., the Southern Ocean) on the global circulation has been noted in earlier observational and modeling studies [e.g., *Toggweiler and Samuels*, 1995; *Doney et al.*, 1998].

[23] $\mathcal{D}$ is large primarily in the subtropical and tropical regions. In these areas, the abyssal heat budgets of the MITgcm and POP models depart significantly from an advective-diffusive balance due to resolved mesoscale eddies in the MITgcm and a combination of eddies and drift in POP. As a result, $\hat{\kappa}_v$ overestimates κ_v by as much as a factor of four. Such a large difference between $\hat{\kappa}_v$ and κ_v could have a significant impact on the estimates of mixing necessary to close the overturning circulation in the world ocean.

[24] That $\mathcal{E} < 0$ in the deep subtropics highlights an important difference between resolved eddies and eddies parameterized by the GM *Gent and McWilliams* [1990] scheme, since GM decreases APE both globally and locally by construction. Resolved eddies are constrained to decrease APE globally, but may increase APE locally. This is particularly true of eddies driven by barotropic instability processes (such as those occurring in western boundary currents and in the tropics) which are relatively insensitive to the buoyancy field. It is unclear how to formulate a realistic and stable eddy parameterization which would allow local upgradient fluxes of APE.

[25] Despite the differences in models – our MITgcm configuration is idealized while the POP model has a realistic geometry – the general structure of the heat fluxes is the same in both models, indicating that the results are robust. Both models lend support to the main thrust of *Wunsch and Ferrari* [2004] and *Gnanadesikan et al.* [2005]: the wind is the main source of mechanical energy in the ocean. We would add that the wind largely controls the large-scale mean vertical heat flux as well.

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References

Cessi, P., W. R. Young, and J. A. Polton (2006), Control of large-scale heat transport by small-scale mixing, *J. Phys. Oceanogr.*, *36*, 1877–1894.

Doney, S. C., W. G. Large, and F. O. Bryan (1998), Surface ocean fluxes and water mass transformations in the coupled NCAR climate system model, *J. Clim.*, *11*, 1420–1441.

Gent, P. R., and J. C. McWilliams (1990), Isopycnal mixing in ocean circulation models, *J. Phys. Oceanogr.*, *20*, 150–155.

Gnanadesikan, A., R. D. Slater, P. S. Swathi, and G. K. Vallis (2005), The energetics of ocean heat transport, *J. Clim.*, *18*, 2604–2616.

Gregory, J. M. (2000), Vertical heat transports in the ocean and their effect on time-dependent climate change, *Clim. Dyn.*, *16*, 501–515.

Hallberg, R., and A. Gnanadesikan (2006), The role of eddies in determining the structure and response of the wind-driven Southern Hemisphere overturning: Results from the Modeling Eddies in the Southern Ocean (MESO) project, *J. Phys. Oceanogr.*, *36*, 2232–2252.

Hill, C., A. Adcroft, D. Jamous, and J. Marshall (1999), A strategy for terascale climate modeling, in *Proceedings of the Eighth ECMWF Workshop on the Use of Parallel Processors in Meteorology*, edited by W. Zwieflhofer and N. Kreitz, pp. 406–425, World Sci., Singapore.

Karsten, R. H., H. Jones, and J. Marshall (2002), The role of eddy transfer in setting the stratification and transport of a circumpolar current, *J. Phys. Oceanogr.*, *32*, 39–54.

Kunze, E., E. Firing, J. M. Hummon, T. K. Chereskin, and A. M. Thurnherr (2006), Global abyssal mixing inferred from lowered ADCP shear and CTD strain profiles, *J. Phys. Oceanogr.*, *36*, 1553–1576.

Large, W. G., J. C. McWilliams, and S. C. Doney (1994), Oceanic vertical mixing: A review and a model with a nonlocal boundary layer parameterization, *Rev. Geophys.*, *32*(4), 363–403.

Maltrud, M. E., and J. L. McClean (2005), An eddy resolving global 1/10° ocean model, *Ocean Modell.*, *8*, 31–54.

Munk, W. H. (1966), Abyssal recipes, *Deep Sea Res*, *13*, 707–730.

Munk, W., and C. Wunsch (1998), Abyssal recipes II: Energetics of tidal and wind mixing, *Deep Sea Res., Part I*, *45*, 1977–2010.

Paparella, F., and W. R. Young (2002), Horizontal convection is non-turbulent, *J. Fluid Mech.*, *466*, 205–214.

Raper, S. C. B., J. M. Gregory, and T. J. Osborn (2001), Use of an upwelling-diffusion energy balance climate model to simulate and diagnose A/OGCM results, *Clim. Dyn.*, *17*, 601–613.

Toggweiler, J. R., and B. Samuels (1995), Effect of Drake Passage on the global thermohaline circulation, *Deep Sea Res., Part I*, *42*, 477–500.

Wolfe, C. L., and P. Cessi (2008), Overturning in an eddy-resolving model: The effect of the pole-to-pole temperature gradient, *J. Phys. Oceanogr.*, in press.

Wunsch, C. (1999), Where do ocean eddy heat fluxes matter?, *J. Geophys. Res.*, *104*(C6), 13,235–13,249.

Wunsch, C., and R. Ferrari (2004), Vertical mixing, energy, and the general circulation of the oceans, *Annu. Rev. Fluid Mech.*, *36*, 281–314.

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