

Implications of a new eddy parameterization for ocean models

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Abstract. A new parameterization of eddies in a coarse resolution ocean model yields deep ocean salinities and temperatures that are significantly closer to observations than with previous parameterizations. This is achieved because dense water is able to flow over sills and into the deep ocean without being diluted with the surrounding water. In addition, the depth to which surface-induced tracers penetrate in the Southern Ocean is now realistic. This depth is overestimated by previous ocean models, including those used to estimate global warming. Adding this new eddy parameterization to coupled atmosphere-ocean models is expected to yield greater global warming.

Introduction

Mesoscale eddies with horizontal scales of 10 to 100 kilometres occur throughout the ocean. They correspond dynamically to the much larger weather systems in the atmosphere with horizontal scales of several thousand kilometres. Because the ocean's eddies are so much smaller, it is prohibitively expensive to resolve them explicitly within ocean models used for climate studies. Hence the effects of these mesoscale eddies will need to be parameterized for the foreseeable future.

Eddies act both to diffuse tracers and to alter the mean advection. A heuristic scheme for including eddy-induced advection in eddyless Cartesian models (Gent and McWilliams (1990), Gent et al (1995)) has been shown to have several beneficial effects on the ability of models to simulate the present ocean state (Danabasoglu et al (1994), Hirst and McDougall (1996)). Here we explain that this scheme is actually a simple parameterization of the eddy-induced advection obtained from Temporal-Residual-Mean theory (McDougall and McIntosh (1996)).

The Temporal-Residual-Mean (TRM) Velocity

By examining the conservation equations for the mean density and density variance, it has recently been shown (McDougall and McIntosh (1996)) that tracers should be advected in an eddyless steady z -coordinate model with the Temporal Residual Mean (TRM) circulation, having horizontal and vertical velocities

$$\bar{\mathbf{V}}^\# = \bar{\mathbf{V}} + \Psi_z \quad \text{and} \quad \bar{w}^\# = \bar{w} - \nabla_H \cdot \Psi \quad (1)$$

where $\bar{\mathbf{V}}$ and \bar{w} are the simple Eulerian averaged velocities (temporal averages at a fixed point in space), the subscript z denotes vertical differentiation while ∇_H is the horizontal gradient operator. The extra vector streamfunction of the TRM circulation is given by

$$\Psi = -\frac{\bar{\nabla}'\gamma'}{\bar{\gamma}_z} + \frac{\bar{\nabla}_z \frac{1}{2}(\gamma')^2}{(\bar{\gamma}_z)^2} + \frac{\bar{\nabla}}{\bar{\gamma}_z} \left(\frac{\frac{1}{2}(\gamma')^2}{\bar{\gamma}_z} \right)_z, \quad (2)$$

where γ is neutral density, which is the oceanographic density variable that correctly accounts for the complicated compressible nature of seawater:— it is defined in Jackett and McDougall (1996) and it corresponds to potential temperature in the atmosphere. The overbar in (2) represents a simple Eulerian average while the primed quantities are deviations from this average. For a steady mean flow without small-scale vertical (diabatic) mixing processes, the total effect of mesoscale eddies is incorporated into the mean density equation in the simple conservation statement $\bar{\mathbf{V}}^\# \cdot \nabla_H \bar{\gamma} + \bar{w}^\# \bar{\gamma}_z = 0$. This simple equation demonstrates the meaning of the term “residual” in the temporal-residual-mean name for this velocity vector. The TRM velocity is the residual left after adding the Eulerian-mean velocity and the velocity derived from the streamfunction of equation (2). It is this residual motion that is adiabatic and it is this residual velocity that advects tracers in the ocean.

If the third term in the streamfunction Eq. 2 is ignored, then the lateral velocity $\bar{\mathbf{V}}^\#$ is a Taylor series approximation to the thickness-weighted lateral velocity of fluid between two adjacent density surfaces. The latter is the sum of the mean velocity along such a surface, $\bar{\mathbf{V}}^N$, plus the “bolus” velocity (Rhines (1982), McDougall (1991)), $\bar{\mathbf{V}}^B$, due to the temporal correlation between the lateral velocity and the thickness between a pair of density surfaces. The third term in Eq. 2 is needed because the variables that are advected in a z -coordinate model are not the same as in a density-layered model.

While each of the three components of Ψ lead to velocity components of roughly the same magnitude (about one millimetre per second), the second and third terms produce velocity components that tend to be along the contours of tracers on density surfaces. Hence these two components make smaller contributions to the tracer budgets than the first term in (2) and will be ignored in what follows. By making the common assumption that unresolved baroclinic eddies transport density horizontally down the mean horizontal gradient of density, that is, $\bar{\nabla}'\gamma' = -\kappa \nabla_H \bar{\gamma}$, the first term of (2) becomes the same vector streamfunction that is advocated by Gent and

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McWilliams (1990) and Gent et al (1995). We note that because $\bar{\nabla}^N$ is not equal to $\bar{\nabla}$, this extra advection does not parameterize bolus advection alone, nor is it equal to the down-gradient flux of mean thickness between adjacent density surfaces as is assumed by Gent and McWilliams (1990) and Gent et al (1995).

In the absence of a better parameterization of the streamfunction, (2), we follow previous authors in using the Gent and McWilliams (1990) suggestion, but we note that the interpretation of the total velocity vector and of its individual components is different to that of Gent and McWilliams (1990) and Gent et al (1995). Also, McDougall and McIntosh (1996) have shown that the model's tracer variables are not the Eulerian averaged variables, nor are they the thickness-weighted variables as was recently advocated by de Szoeke and Bennett (1993), but rather they are the tracer values averaged along the appropriate density surfaces. This curious combination of Eulerian (to obtain $\bar{\gamma}$) and quasi-Lagrangian averaging (to obtain the averaged

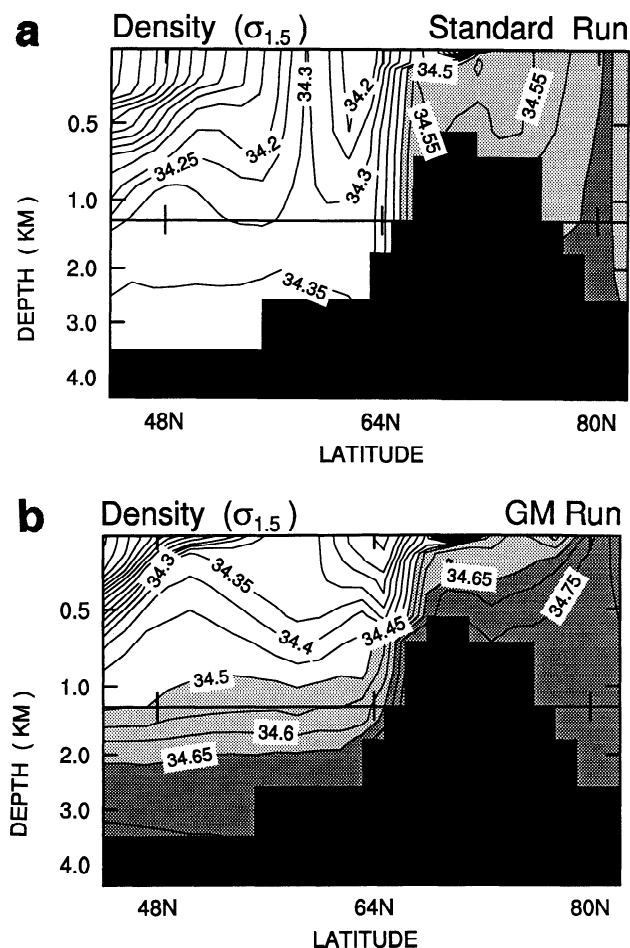


Figure 1. A vertical section of potential density (referenced to 1500 db) through the Denmark Strait following approximately the path of dense water down the continental slope. The standard model that includes the unphysical horizontal diffusion process is shown in (a) while the new model containing the TRM parameterization of Gent and McWilliams (1990) and without horizontal diffusion is illustrated in (b). The shading of potential density contours is the same in both figures.

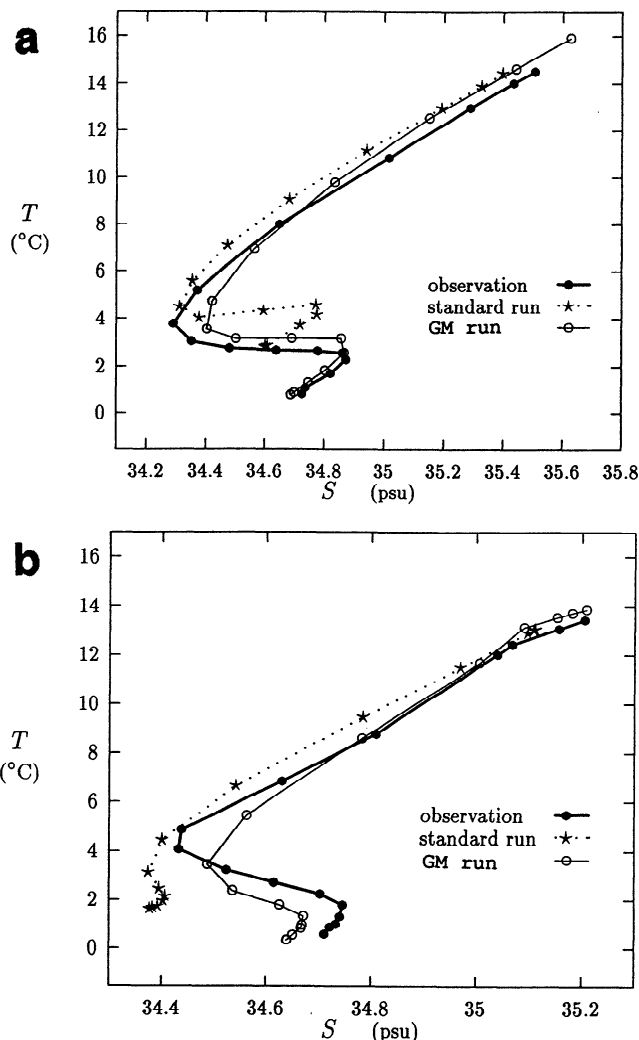


Figure 2. Salinity-potential temperature diagrams from the South Atlantic (a; at 35°S , 25°W) and from the South Indian (b; at 35°S , 90°E). Run 1 contains the standard model physics (including the unphysical horizontal diffusion) while run 2 includes the Gent and McWilliams parameterization of the temporal-residual-mean.

tracer value) allows such models to more faithfully reproduce water masses than has been realised to date.

Ocean Model Results

The global ocean model that we have run is based on the Bryan-Cox-Semtner code (Cox (1984)). It has 21 levels in the vertical and is run at a horizontal resolution of 1.6° latitude and 2.8° longitude. The bottom topography is as accurate as is consistent with a model of this resolution (for details see Hirst and McDougall, 1996). The presence of this topography is an important way in which the present model differs from that of Danabasoglu et al (1994) where the minimum ocean depth was set to be 2500m. Both model runs discussed here have a quasi-horizontal diffusivity directed along density surfaces (Redi (1982)) of $10^3 \text{ m}^2 \text{ s}^{-1}$. In the standard model run we need a horizontal diffusivity of $0.7 \times 10^3 \text{ m}^2 \text{ s}^{-1}$ to maintain numerical stability. The new model with the TRM parameterization needs no artificial

horizontal diffusivity and the TRM diffusivity is $10^3 \text{ m}^2 \text{ s}^{-1}$. The surface boundary conditions on salinity and temperature are the same in both cases, involving relaxation to observed values with a time scale of 30 days.

One of the major benefits of the TRM advection scheme is the ability to eliminate the unwanted horizontal diffusion. This benefit can be seen in the potential density section (Figure 1) which passes through the Denmark Strait and approximately follows the path of the densest water down the continental slope. As the dense water attempts to flow southwards over the sill in the standard ocean model (Figure 1a) it suffers dilution with the ambient fluid and the overflow water is reduced in density by more than 0.2 kg m^{-3} . With the TRM parameterization of Gent and McWilliams (1990) (Figure 1b) the horizontal diffusion process is absent and the dense overflow water at the Denmark Strait proceeds with very little dilution, arriving in the deep ocean at much the same density as is observed. The change in density on falling down the continental shelf is one quarter of that in the standard model.

Water properties are often examined on a salinity-potential temperature diagram; in Figure 2 we compare the observed water properties with those of the standard model and the model with the TRM parameterization. The North Atlantic Deep Water properties (the salinity maximum feature in Figure 2a) are reproduced much more faithfully with the new parameterization. The improvement in the Bottom Water found in the South Atlantic (Figure 2a) is primarily due to it being much cooler, whereas in the Indian (Figure 2b) (and also in the Pacific Ocean, not shown) the improvement is primarily due to the increase in salinity. It should be emphasised that to date no attempt has been made to tune the new model to observations. For example, it is

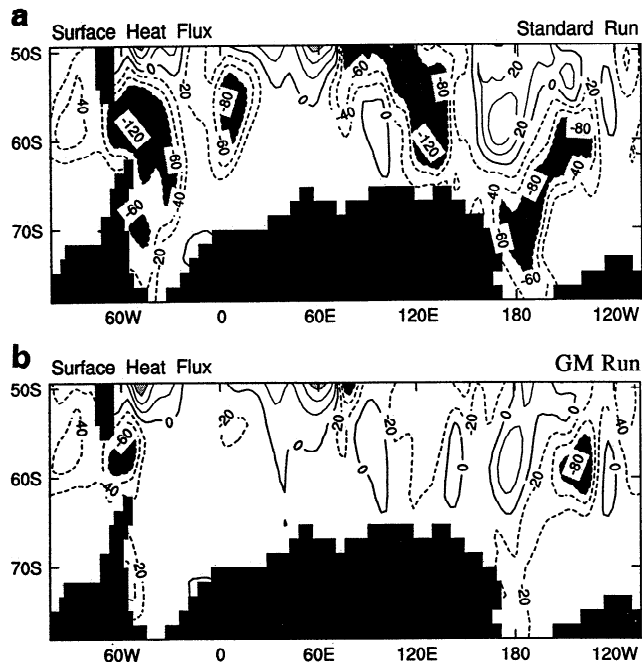


Figure 3. Air-sea heat flux (W m^{-2}) in the Southern Ocean region of the standard model (a) and with the Gent and McWilliams (1990) parameterization of the temporal-residual-mean (b). The mostly negative numbers are consistent with heat being fluxed out of the ocean to the atmosphere.

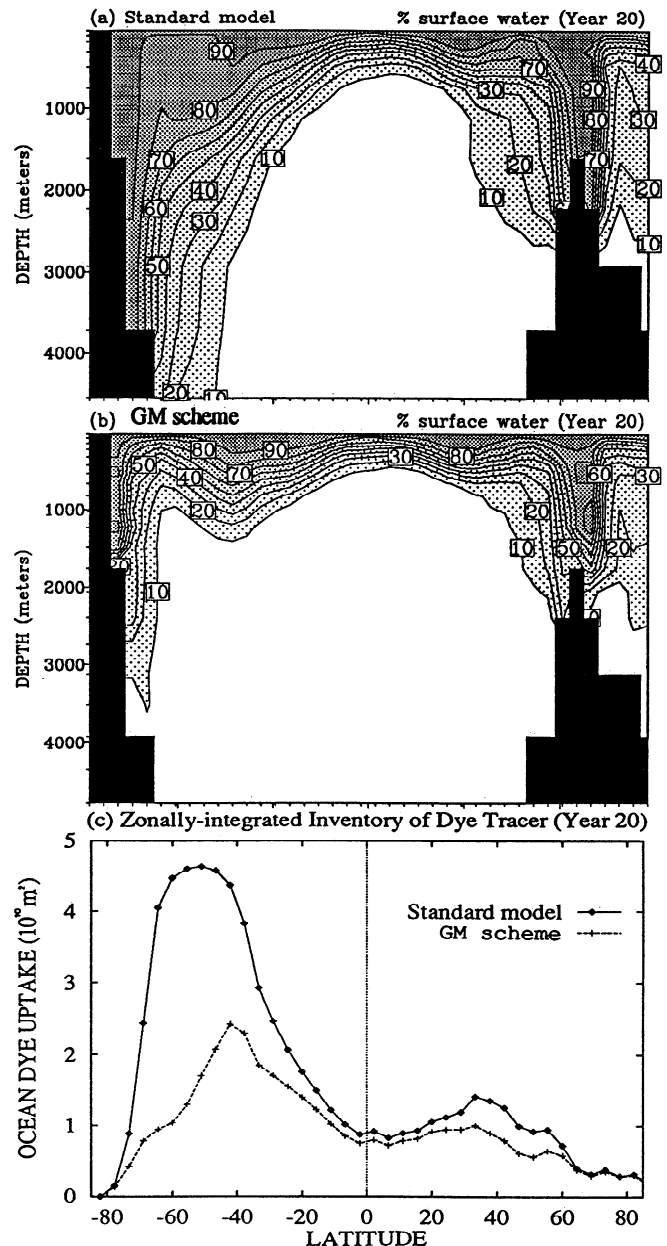


Figure 4. Zonally averaged concentration of a dye tracer that was put suddenly from zero to one at the surface twenty years earlier. Figure (a) is for the standard ocean model and (b) is for the model with the Gent and McWilliams (1990) parameterization. Figure (c) shows the volume integral of the dye concentration as a function of latitude, and the table shows the total volume integral of dye in each hemisphere.

known (England (1993), Hirst and Cai (1994)) that the properties of the Intermediate Water (at the location of the salinity minimum on Figure 2) can be changed by changing the diffusivity that acts along density surfaces.

In a steady state, the more realistic deep density field in the TRM model causes a more gravitationally stable thermocline and hence, for the same surface boundary conditions, the regions of deep convective activity are much reduced (Danabasoglu et al (1994), Hirst and McDougall (1996)). For similar reasons, the TRM model has an area-averaged air-sea heat flux in the Southern Ocean region that

is approximately half that in the standard run (Figure 3). This average air-sea heat flux is closer to the sparse climatological data that exist in this region (Jacobs et al (1985)). Just as importantly, Figure 3b has less extremes of heat flux. Because of these reductions in both the average and the extreme heat fluxes, we expect that when this type of ocean model is coupled to an atmospheric model, the flux corrections that are required in order to prevent climate drift will be reduced.

The unrealistically deep mixed layers that appear in the Southern Ocean of the standard ocean model imply that the "thermal flywheel" effect by which the ocean acts to delay the effects of Greenhouse-induced warming has been overestimated. Since this is the type of ocean model that has been used in the most recent predictions of climate change (Washington and Meehl (1989), Manabe et al (1991), Manabe and Stouffer (1994)) it is important to examine the amount by which this thermal flywheel has been overestimated. For this purpose we have run a very similar version of the model (with resolution and geometry chosen to match that of certain climate models, in particular, that of Manabe et al (1991)) and when a steady state was achieved, a passive dye of concentration equal to unity was imposed thereafter at the sea surface throughout the globe. The realistic behaviour of the Gent and McWilliams (1990) parameterization for certain measurable passive tracers has recently been confirmed (England (1996), Robitaille and Weaver (1995)).

After twenty years of integration, the zonally averaged concentrations for the two versions of the model was as shown in Figure 4. The standard model has mixed layer depths that are about a factor of two greater than observations and so this model substantially overestimates the rate at which the ocean can vertically move and mix the tracer imposed at the surface, particularly in the Southern Ocean (compare Figures 4a and 4b), with little hint of old Circumpolar Deep Water upwelling near 60°S (a known feature of the circulation in this region). The volume integral of the transient dye is shown in Figure 4c as a function of latitude. When integrated over each hemisphere, these volume integrals become the effective magnitude of the thermal flywheel in each hemisphere.

The global magnitude of this thermal flywheel is 77% larger in the standard model than in the model with the TRM parameterization. This means that for a given increase in surface heat flux under global warming, we expect the magnitude of warming averaged over the globe to be about 77% larger with the TRM parameterization than the present generation of coupled models predict. The situation is more extreme in the southern hemisphere where the thermal flywheel volume is overestimated by the standard model by a factor of two. These factors of a 77% increase in warming globally and 100% increase in the southern hemisphere only apply if the air-sea heat flux anomaly under global warming were the same both with and without the TRM parameterization. In practice the air-sea heat flux depends quite strongly on the ocean's role in transporting heat. Hence the above factors are only indicative of the influence of the TRM eddy parameterization and the real sensitivity to this parameterization is expected to be less than the factors of 77% and 100%, and will only be known once coupled model runs are performed with this new ocean physics.

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