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Sensitivity of water mass transformation and heat transport to subgridscale mixing in coarse-resolution ocean models

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[1] This paper considers the impact of the parameterization of subgridscale mixing on ocean heat transport in coarse-resolution ocean models of the type used in coupled climate models. Increasing the vertical diffusion increases poleward heat transport in both hemispheres. Increasing lateral diffusion associated with transient eddies increases poleward heat transport in the southern hemisphere while decreasing it in the northern hemisphere. The results are interpreted in the context of a simple analytical model. *INDEX TERMS:* 4532 Oceanography: Physical: General circulation; 4568 Turbulence, diffusion, and mixing processes; 4203 Oceanography: General: Analytical modeling; 4255 Numerical modeling; 4279 Upwelling and convergences. **Citation:** Gnanadesikan, A., B. L. Samuels, and R. D. Slater, Sensitivity of water mass transformation and heat transport to subgridscale mixing in coarse-resolution ocean models, *Geophys. Res. Lett.*, *30*(18), 1967, doi:10.1029/2003GL018036, 2003.

1. Introduction

[2] The oceans play an important role in determining the transport of heat within the climate system. The poleward transport of heat plays an important role in maintaining climate stability. Recent work by one of our colleagues suggests that without this heat transport the globe would freeze over, [*Winton*, 2003]. The representation of subgridscale processes has long been felt to play an important role in the magnitude of heat transport. *Bryan* [1987] demonstrated that the transport of heat was strongly affected by the coefficient of vertical diffusivity, K_v. Increasing K_v was shown to stir warm tropical waters down into the ocean interior, resulting in a larger pressure difference between high and low latitudes and a stronger heat transport. This work was further verified by *Park and Bryan* [2000] in the context of both isopycnal and level-coordinate ocean circulation models.

[3] However, in recent years it has become clear that the large-scale overturning may be mechanically driven by Southern Ocean winds and eddies as well as by tropical heating and mixing [*Toggweiler and Samuels*, 1998; *Gnanadesikan*, 1999]. This leads to a picture such as that in Figure 1. In the northern hemisphere, light water is converted to dense water at a rate T_n. Following *Bryan* [1987] and *Park and Bryan* [2000] this flux can be scaled as the reduced gravity between light and dense waters $g' = g\Delta\rho/\rho$, the pycnocline depth D and the Coriolis parameter f

$$T_n = Cg'D^2/f = g'D^2/\epsilon \quad (1)$$

where C is a constant that parameterizes the effects of friction and basin geometry. Essentially, this parameterization assumes some proportionality between the east-west geostrophic flux of light water and the north-south overturning flux.

[4] In order for the ocean to be in steady state, T_n must be counterbalanced by an equivalent conversion of dense water to light water. In *Bryan* [1987] it was assumed that this conversion occurred in the tropics, with a flux $T_u = T_n$

$$T_u = K_v A/D \quad (2)$$

where K_v is the vertical diffusion coefficient and A is the area of the tropics m^2. If $T_n = T_u$, the pycnocline depth D will scale as $K_v^{1/3}A^{1/3}$ and the overturning as $K_v^{2/3}A^{2/3}$.

[5] An alternative location for watermass conversion is the Southern Ocean. *Gnanadesikan* [1999] proposed that the magnitude of the Southern Ocean watermass conversion flux T_s was controlled by a balance between Southern Ocean winds and eddies.

$$T_s = \frac{\tau_s L_x}{\rho f} - \frac{A_I D L_x}{L_y} = T_{ek} - T_{eddy} \quad (3)$$

where τ_s is the wind stress within Drake Passage latitudes, L_x is the length along a latitude circle in m, ρ is density, f the Coriolis parameter, A_I the lateral diffusion coefficient for layer thickness [*Gent and McWilliams*, 1990], and L_y is the north-south scale in m over which the pycnocline shallows in the Southern Hemisphere. The northward Ekman flux (T_{ek}) driven by the wind stresses can either be supplied by deep upwelling or by eddy fluxes (T_{eddy}) of light water from lower latitudes. Increasing A_I favors the latter pathway.

[6] At steady state

$$T_n = T_u + T_s \quad (4a)$$

$$\frac{g'D^2}{\epsilon} = \frac{K_v A}{D} + \frac{\tau_s L_x}{\rho f} - \frac{A_I D L_x}{L_y} \quad (4b)$$

yielding a cubic equation in D. We will rely upon the insights gained from the solution of this equation to interpret our solutions.

[7] *Klinger et al.* [2003] recently verified this scaling using idealized isopycnal and level coordinate circulation models. They noted that a key nondimensional parameter

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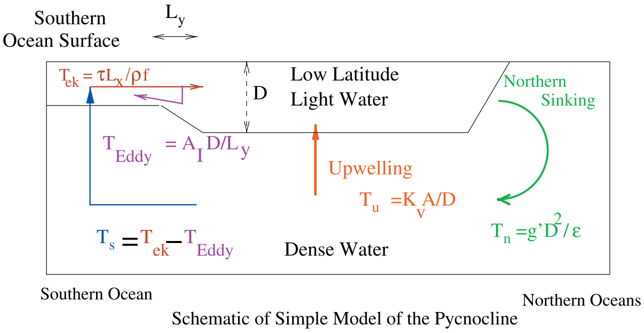

Figure 1. Schematic of thermohaline circulation, after *Gnanadesikan* [1999].

emerging from this equation is the ratio of the Ekman flux to the overturning when $\tau_s = A_I = 0$ (i.e. the diffusive scaling of *Bryan* [1987]). If we let $A = L_x * L_t$

$$E = \frac{\tau_s L_x}{\rho f} \left(\frac{K_v^2 A^2 g'}{\epsilon} \right)^{-1/3} = \frac{\tau_x L_x^{1/3} \epsilon^{1/3}}{K_v^{2/3} L_t^{2/3} g'^{1/3}} \qquad (5)$$

They argue that for reasonable parameter values E is of order 1, so that "globally averaged K_v may need to be on the low side of observational estimates for the meridional overturning to be dominated by southern wind." This note compares these results with those found in a model with more realistic geometry.

[8] One implication of (4b) is that it should be possible to obtain the same T_n (and by implication the same pycnocline depth D) while varying K_v and A_I so as to keep the sum of the first and third terms on the right-hand side of (4b) constant. It is thus theoretically possible to have relatively similar density structures (at least in the global average) with different subgridscale mixing. However, when K_v and A_I are both large, most of the transformation of dense water to light water will occur in the tropics, while when they are both small, most of this transformation will occur in the Southern Ocean. In *Gnanadesikan et al.* [2002], we verified this basic picture. In the present note, we extend this work to look at oceanic heat transport.

2. Experiments

[9] The basic suite of experiments is described in *Gnanadesikan et al.* [2002]. It comprises four runs at a nominal resolution of 4 degrees consisting of a matrix of high and low vertical diffusion and lateral diffusion coefficients. The low values of 0.15 cm^2/s for vertical diffusion and 1000 m^2/s were chosen based on the North Atlantic Tracer Release Experiment [*Ledwell et al.*, 1998]. In order to explore the impact of higher values of lateral diffusion, such as might be associated with higher sea-surface height variability in the Southern Ocean, we also used a value of lateral diffusion of 2000 m^2/s. This results in a decrease on the right-hand side of (4b). We estimated that increasing the pycnocline diffusion to 0.6 cm^2/s would compensate for this decrease, allowing the pycnocline depth and T_n to remain constant. The resulting four runs are denoted by *vertical diffusion* + *lateral diffusion* so that KVHI + AIHI has both vertical and lateral diffusion high, and KVLO + AILO has them both low. Wind stresses for these runs were given by *Hellerman and Rosenstein* [1983] and topography was the "realistic" topography of the GFDL R15 coupled general circulation model. A key feature of this topography is that Drake Passage is quite wide, with its northern edge at 48.9S (as opposed to 55S in the real world). Applied heat and salt fluxes from *da Silva et al.* [1994] were used, but with an additional "flux correction" applied by restoring the temperatures and salinities back to the dataset of *Levitus and Boyer* [1994].

[10] A final pair of experiments is presented here that evaluates the robustness of the heat transport response to lateral diffusion. These simulations were made using the new higher-resolution ocean component of the GFDL coupled model, referred to as OM2. The model has a resolution of 2 degrees in the east-west direction, and an average resolution of 1 degree in the north south direction. There are 50 vertical levels (so that the total number of grid points in OM2 is about 16 times that found in the coarser models). It uses the explicit mixed layer scheme of *Large et al.* [1994] and has a recoded version of the isopycnal mixing scheme of *Griffies et al.* [1998] and *Griffies* [1998]. The surface winds in this model are given by the NCEP reanalysis and fluxes are given from the atmospheric component of the GFDL coupled model, forced according to the Atmospheric Model Intercomparison Project (AMIP) protocol. Two runs are made with this model, one with a lateral diffusion coefficient of 1000 m^2/s (referred to as OM2HI) the other with a lower value of 600 m^2/s (referred to as OM2LO). These models are only run out for 440 years, so that the circulation is not, strictly speaking, at equilibrium.

3. Results

[11] Subgridscale parameterization has a major impact on the heat transport. This can be seen in Figure 2a, which compares KVLO + AILO and KVHI + AIHI. The export of heat between 20S and 20N is 2.6 PW in the KVLO + AILO case and 3.6 PW in the KVHI + AIHI case. The value of 3.1 estimated by *Trenberth and Caron* [2001] lies in between these two. The models line up reasonably well with the atmospheric reanalysis in the Northern Hemisphere, but bracket the observations in the Southern Hemisphere. The impact of changing only lateral or vertical diffusion is shown in Figure 2b. Vertical diffusion clearly has the larger impact, increasing the poleward heat transport by 0.2 PW in the Northern Hemisphere and 0.8 PW in the Southern Hemisphere. Lateral diffusion has a smaller impact, increasing poleward heat transport in the Southern Hemisphere by 0.4 PW, but *reducing* it in the Northern Hemisphere.

[12] In order to link these change to the theory of *Gnanadesikan* [1999], we examine the northward transport of "light" waters in the model suite (Figure 2c). A value of σ_t = 27.4 is chosen to separate the light and dense waters, since it separates Antarctic Intermediate Waters from North Atlantic and North Pacific Deep Waters. When K_v is small, the dominant pathway for watermass conversion is the Southern Ocean, and the northward transport of lighter waters is relatively constant in low latitudes (black and red lines in Figure 2c). When K_v is large, the northward transport of water increases throughout the tropics (blue and green lines in Figure 2c) implying strong net water mass conversion there.

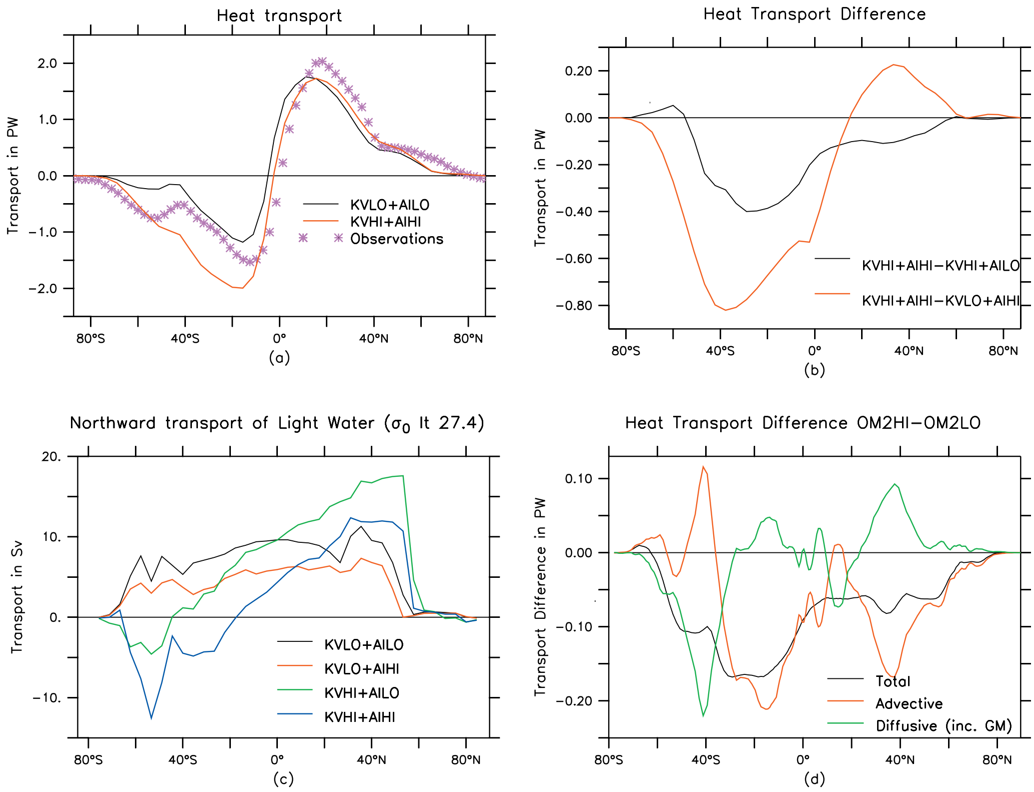

Figure 2. Dependence of heat transport and storage on subgridscale parameterizations. (a) Meridional heat transport, KVLO + AILO (solid), KVHI + AIHI(red) and observations (stars). (b) Change in meridional heat transport due to changing A_I (black) and K_v (red) in the 4 degree model. (c) Northward transport of light ($\sigma_0 < 27.4$) water in 4 degree simulations. Increase towards the right means dense water is being transformed to light water. (d) Change in heat transport caused by changing A_I in 2 by 1 degree model, total (solid), advective component (red), and diffusive component (green).

[13] If we take the maximum northward flow as T_n, the range in northward flow as T_u and the residual as T_s we can examine the connection between heat transport and our simple theory. The change in poleward heat transport in the northern hemisphere ΔH_n and southern hemisphere ΔH_s are shown Table 1. ΔH_s scales as the change in T_s, assuming a temperature difference of 13 degrees. $\Delta\ H_n$, however does not scale as neatly, implying that there are changes in the temperature distribution associated with T_n. Such changes are beyond the scope of this simple theory.

[14] We can compare the modeled water mass transformations with Equation (4b). In the real world, the $\sigma_0 = 27.4$ is found at a depth of about 1000m. Assuming that about 17 Sv of water is transformed across this surface $g'/\epsilon = 17$. We can take $A = 2.4 \times 10^{14}$ m^2, $L_x = 26{,}000$ km and $L_y^s = 1200$ km. The baseline diffusive transport associated with these parameters is 6.0 Sv for the KVLO runs. T_{ek} at the Southern tip of Drake Passage in the model is 28.7 Sv. This is about midway between the 19.5 Sv found at the true latitude of Drake Passage using the Hellermann winds and 40 Sv using the ECMWF winds [*Trenberth et al.*, 1989]. Using these numbers we can solve (4b) for T_s, T_n and T_u (Table 1). The correlation coefficient between the predicted and modeled transports is 0.9. Given that all the above parameters were selected from observations, this is remarkably good agreement.

[15] Analysis of the scaling of the various terms shows that the worst predictions are for T_s. Even when the effective T_{ek} and L_x/L_y in (3) are fit from the model output, the RMS error is 3.2 Sv. If a quadratic rather than a linear dependence on D is used for the fit, however, the RMSE drops to 0.5 Sv. This indicates that to some extent the *geometry* of the flow depends on the subgridscale parameters. Similarly, there is some evidence from the output that the effective area over which upwelling occurs (A in Equation (2)) also increases as K_v increases, and decreases as A_I increases. This would account for the drop in the diffusive upwelling between AILO and AIHI cases, despite the fact that D decreases as predicted by the theory. We do not at present have a theory to explain these relatively subtle dependencies of the flow geometry on subgridscale physics.

Table 1. Numerical Values for the Changes in Poleward Heat Transport in the Southern ΔH_s and Northern ΔH_n Hemispheres Relative to KVLO + AILO, Net Transformation of Water Relative to the $\sigma_0 = 27.4$ Surface and the Mean Depth of That Surface

	KVLO + AILO	KVHI + AILO	KVLO + AIHI	KVHI + AIHI
ΔH_s	0	0.61	0.20	0.96
ΔH_n	0	0.28	−0.05	0.18
T_n	11.3	16.9	7.3	11.9
	13.9	*19.7*	*7.1*	*11.6*
T_s	7.6	−3.1	4.2	−7.6
	9.9	*6.3*	*1.6*	*−5.8*
T_u	3.7	20.0	3.1	19.5
	3.9	*13.4*	*5.5*	*17.4*
D	886	1270	697	1020
	903	*1076*	*649*	*827*

Italics show the predictions made using Equation (4b) with the constants as listed in the text.

[16] Even with these caveats, the theory provides an explanation for the changes in poleward heat transport with K_v and A_I. Increasing K_v increases D as more dense water is converted to light water in the tropics. Increasing D increases T_{eddy} and T_n and thus boosts poleward heat transport in both hemispheres. Increasing A_I decreases D as less dense water is converted in the Southern Ocean, resulting in a decrease in T_n and thus northern hemisphere heat transport. However $A_I * D$ increases so that T_{eddy} (and thus poleward heat transport) increases in the Southern Hemisphere. Our simple theory (given parameters taken from observations) does not account for the fact that the Southern Hemisphere response to changing K_v is about twice the Northern Hemisphere response (Table 1). We suspect that the fact that T_{eddy} has a quadratic dependence on D in the GCMs (rather than the linear dependence in the scaling) is at the root of the problem.

[17] There are two interesting contrasts between these results and those of *Klinger et al.* [2003]. The first is that at the lower values of diffusion the flow is clearly dominated by the Southern Ocean transformation pathway, even though the diffusion coefficient is relatively high. Why is this? Examination of Equation (5) shows two reasons. The first is that we use a $g'/\epsilon = 17$ based on observations of the circulation, while *Klinger et al.* [2003] use a value of 500. Analysis of *Park and Bryan* [2000] show that their equivalent value would be around 25. A higher value of g'/ϵ will result in an overestimate of the importance of the diffusive circulation, so that *Klinger et al.* [2003] would find E = 1.5 for the low value of K_v, whereas our value is 4.7.

[18] Additionally, our response tends to be concentrated in the Southern Hemisphere, while *Klinger et al.* [2003] see the largest response in the north. We believe this difference to be due to the difference in model geometry. In the *Gnanadesikan* [1999] scaling, there is no transient eddy flux in the Northern Hemisphere. Since the pycnocline shallows only over a small region of the Atlantic, the equivalent of L_x in Equation (2) is only a few thousand km, and the equivalent T_{eddy} is much smaller than for the Southern Ocean and is neglected. This cannot be done in a sector model. When a northern T_{eddy} is included in (4b) using the same L_x as for the Southern Hemisphere, T_n becomes much larger than T_s and shows little sensitivity to A_I.

[19] The response of the heat transport to changes in A_I is also found in a much higher-resolution model with more realistic physics (Figure 2d). The similarity is encouraging, as it indicates that the insights derived from lower resolution models can be used to guide development of higher resolution coupled models. As seen in Figure 2d, reducing A_I reduces both the subgridscale eddy transport (green line) in the Southern Ocean, where the isopycnals slope most steeply and the the advective transport of heat (red line) out of the tropics.

4. Conclusions

[20] As ocean circulation models are increasingly used to predict future behavior of the physical climate and biogeochemical cycles, it is more and more important to understand why such models differ from each other and how these differences manifest themselves. This paper suggests a framework within which the impact of changing subgridscale parameterizations on heat transport can be understood.

[21] Differing heat transport in ocean circulation models can produce significant differences when these models are coupled to atmospheric models. These differences are larger than might be expected from runs (such as those in the present paper) in which the effective atmospheric temperature is held fixed. Preliminary simulations with the GFDL coupled model show that SST differences between GCMs forced with different wind stresses may be amplified by a factor of 3 in coupled simulations, as changes in cloudiness affect the surface air temperature and radiation balance.

[22] An issue which we have not covered here is the impact of changing wind stress on the heat transport. In a paper currently in preparation, we explore this issue, and demonstrate that both tropical and Southern Ocean winds play a critical role in heat transport.

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