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Southern Ocean upwelling and eddies: sensitivity of the global overturning to the surface density range

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ABSTRACT

A simple interhemispheric ocean model is used to examine the sensitivity of water sinking in the northern hemisphere to the equator-to-pole density contrast. The model assumes that the sinking is compensated by upwelling in both the low latitude ocean and the Southern Ocean. We compare two vertical mixing schemes: one with a fixed vertical diffusivity and another with fixed mixing energy. The latter case implies that the vertical diffusivity depends on the simulated oceanic circulation. It is shown that when Southern Ocean upwelling is controlled *only* by northward Ekman transport, the rate of deep water formation has an *opposite* dependence on the equator-to-pole density contrast between the two vertical mixing schemes. However, when Southern Ocean upwelling is controlled by both Ekman transport *and* strong enough eddy-induced transport across the Antarctic Circumpolar Current, the two mixing schemes give qualitatively *similar* dependence: the rate of water sinking increases with the equator-to-pole density contrast, regardless of whether the diffusivity or the mixing energy is held fixed. It is suggested that the ACC eddies and vertical mixing jointly control the response of the overturning circulation to changes in the equator-to-pole density contrast.

1. Introduction

Differential heating and evaporation between the low and high latitudes creates an equator-to-pole density contrast, which is an important factor in controlling the strength of the meridional overturning circulation (MOC) in the ocean. The MOC in turn plays an important role in transporting heat from tropical to polar regions. Different branches of the MOC are believed to be controlled by different dynamics, and it is a nontrivial problem to try and understand how the strength of the MOC depends on the equator-to-pole density contrast. The answer may depend on the assumptions made about the circulation. Two appear to be of particular importance: (1) the assumption about the parametrisation of diapycnal mixing in the low latitude upwelling branch of the MOC; (2) the assumption about the mechanisms controlling deep water upwelling in the Southern Ocean.

Item (1) has been addressed recently by Nilsson and Walin (2001), whereas item (2) has been discussed by Gnanadesikan (1999). It is our aim here to combine these studies in an attempt to obtain additional insight as to how Southern Ocean dynamics, jointly with low latitude diapycnal mixing, can influence the relation between the rate of deep water formation in the northern hemisphere (NH) and the equator-to-pole water density contrast.

As in the single hemisphere case analyzed recently by Nilsson and Walin (2001), we consider two different representations of vertical mixing. One of them simply assumes that the vertical diffusivity is fixed, i.e. it does not depend on the ocean stratification and hence on the circulation. This is a widely used assumption in ocean general circulation models (GCMs). However, decoupling vertical diffusivity from stratification can be difficult to justify, particularly when a GCM is used to simulate switches between considerably different

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states of the MOC (for example, switches between active and inactive modes of deep water formation in the North Atlantic). The second mixing scheme we analyze assumes that the power available for mixing across isopycnals in the low-latitude ocean, rather than the coefficient of vertical diffusivity, is fixed (Kato and Phillips, 1969; Munk and Wunsch, 1998; Huang, 1999; Nilsson and Walin, 2001). This assumption leads to a dependence of the vertical diffusivity on the density contrast, coupling the diffusivity to the simulated oceanic circulation.

We first briefly outline a single-hemisphere case in which all deep water formed in high latitudes upwells at low latitudes. Then, the Southern Ocean upwelling is taken into consideration, using an interhemispheric model proposed recently and tested against a GCM by Gnanadesikan (1999).

2. Southern Ocean excluded

Conceptually, if the Southern Ocean is not considered, the MOC can be represented by two major branches: (1) sinking of surface water to depth in high northern latitudes (denoted T_n); (2) upwelling of the deep water through the low latitude pycnocline (denoted T_u).

These two branches are connected by a poleward water transport in the upper ocean and by an equatorward transport in the deep ocean to form a loop of MOC (Fig.1a). This highly simplified picture of the MOC is often used in box models of planetary scale, driven by a poleward density contrast $\Delta\rho$. Using the thermal wind balance $u_z = g\rho_y (f\rho_0)^{-1}$ and continuity (e.g. Lineikin, 1955; Robinson and Stommel, 1959; Bryan and Cox, 1967; Welander, 1986; see also Park and Bryan, 2000), the rate of water sinking in high latitudes scales as:

$$T_n = g\Delta\rho D^2 (f\rho_0)^{-1}, \tag{1}$$

where $\Delta\rho$ measures the equator-to-pole density contrast; $\Delta\rho$ is also the density difference between the deep ocean water and the light upper ocean water at low latitudes, D is the depth of the pycnocline, g is gravity, f is the Coriolis parameter and ρ_0 is a constant reference water density.

For the upwelling flow, a simple advective–diffusive balance $wT_z = k_v T_{zz}$ is normally employed (e.g. Munk, 1966), which yields a rate of deep water transport through the low-latitude pycnocline as follows:

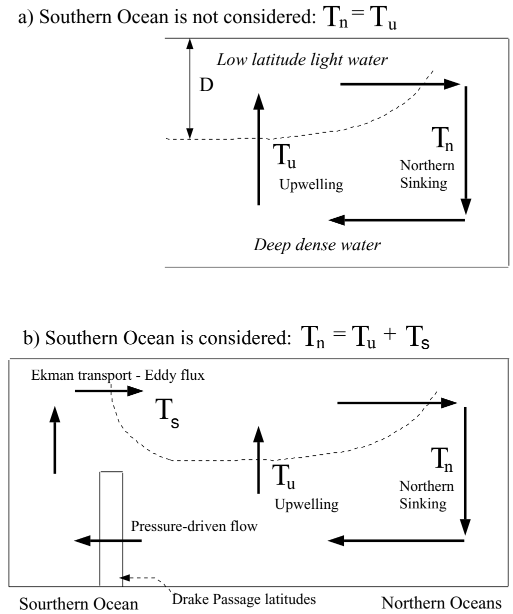

Fig. 1. Schematic representation of the flows in the model without (a) and with (b) the inclusion of Southern Ocean dynamics.

$$T_u = \frac{Ak_v}{D}, \tag{2}$$

where A is the low-latitude ocean area and k_v is a coefficient of vertical diffusivity. Alternatively, one may assume that most of the vertical mixing is localized along the lateral boundaries, in which case A represents the total area of such confined regions and k_v is accordingly adjusted to keep the same rate of upwelling (Marotzke, 1997). In this study we will refer to T_u as low-latitude upwelling.

Next, from the balance $T_n = T_u$ one can derive a dependence of the MOC on the equator-to-pole density contrast of the form:

$$T_n = T_u = (Ak_v)^{2/3} \left(\frac{g\Delta\rho}{f\rho_0} \right)^{1/3}. \tag{3}$$

Equation (3) implies that the strength of the MOC is *directly* proportional to $\Delta\rho$ to the power $1/3$.

So far we assumed that the coefficient of vertical diffusivity k_v *does not* depend on the simulated oceanic stratification. We next find out how the relation between the MOC and $\Delta\rho$ changes if we assume that k_v depends on the stratification. One of the assumptions one can make (put forward originally by Kato and Phillips (1969) is that the external energy needed to lift up the dense deep water against gravity through the low-latitude pycnocline is fixed. The estimates of external energy available from the potential sources are discussed by Huang (1999) and Munk and Wunsch (1998). An expression for the external energy (termed mixing energy) is then given by:

$$\varepsilon = \rho_0 \int k_v N^2 \, dz, \tag{4}$$

where $N^2 = -(g/\rho_0)\rho_z$. Assuming that k_v is vertically uniform yields a coefficient of vertical diffusivity as a function of $\Delta\rho$ (e.g. Nilsson and Walin, 2001):

$$k_v = \varepsilon/(g\Delta\rho), \tag{5}$$

which implies that k_v increases (decreases) when the density difference between the deep water and the light thermocline water decreases (increases). In other words, unlike in most GCMs, the vertical diffusivity is allowed to depend on the simulated oceanic circulation when the system moves from one steady state to another.

Combining eq. (5) with eq. (3) leads to another dependence of MOC on the equator-to-pole density difference:

$$T_n = T_u = (A\varepsilon)^{2/3}\left(\frac{1}{g\Delta\rho f\rho_0}\right)^{1/3}. \tag{6}$$

In contrast to eq. (3), relation (6) states that the strength of the MOC is *inversely* proportional to $\Delta\rho$ to the power 1/3. Similar dependence of the overturning circulation on $\Delta\rho$ was derived and discussed by Nilsson and Walin (2001). In fact, they consider a more general case in which the MOC is proportional to $\Delta\rho^\alpha$ so that α is either positive or negative. Nilsson and Walin (2001) refer to the latter case as the freshwater-boosted regime of circulation. The inverse dependence of MOC on $\Delta\rho$ in the case of negative α implies that gradual freshening in high latitudes increases the strength of MOC by reducing $\Delta\rho$.

Even though the above simple scaling analysis is constrained by a number of assumptions, the reversal of the dependence of the MOC on $\Delta\rho$ from direct in eq. (3) to inverse in eq. (6) is fundamental. We next take into consideration Southern Ocean dynamics, assuming that the upwelling occurs not only at low latitudes but also in the Southern Ocean.

3. Southern Ocean included

Following Gnanadesikan (1999), we assume that for the upper Southern Ocean at Drake Passage latitudes, the principal components of flow are the northward Ekman transport driven by the winds and a return flow associated with meso-scale eddies. The residual of these two flows gives a net water transport to the north, which is compensated below the Drake Passage sill depth by a southward, pressure-driven flow (Fig. 1b). In order to close the Southern Ocean overturning loop, the same amount of water must upwell south of 60°S.

Under these assumptions, the rate of Southern Ocean upwelling is determined by the windstress τ and the eddy-induced transport across the Antarctic Circumpolar Current (ACC). Using the Gent and McWilliams (1990) parametrization to represent the latter yields the following expression for the net deep water upwelling in the Southern Ocean, T_s (Gnanadesikan, 1999):

$$T_s = \left(\frac{\tau}{f\rho_0} - \frac{A_I D}{L_y}\right) L_x, \tag{7}$$

where A_I is the eddy diffusion coefficient and L_x is the circumference of the Earth at Drake Passage latitudes.

Now, instead of $T_n = T_u$, we have a new balance $T_n = T_u + T_s$, which implies that a fraction of the water downwelled in NH can upwell in the Southern Ocean. Using eqs. (1), (2) and (7) to represent, respectively, T_n, T_u and T_s yields a cubic equation for D in a form similar to that of Gnanadesikan (1999):

$$\frac{C_1 g\Delta\rho}{f\rho_0} D^3 + \frac{A_I L_x}{L_y} D^2 - \frac{\tau L_x}{\rho_0 f} D - k_v A = 0 \tag{8}$$

where a constant C_1 is introduced to be consistent with the original model of Gnanadesikan (1999). To obtain a corresponding value for C_1, we equate a ratio of C_1/f in our model to the ratio of $C/(\beta L_y)$ in the Gnanadesikan (1999) model so that $C_1 = fC/(\beta L_y)$, where β is the north–south gradient of the Coriolis parameter. Using Gnanadesikan's values, such as $C = 0.16$, $L_y = 1500$ km, $\beta = 2 \times 10^{-11}$ m^{-1}s^{-1} and the a value for f of 1.2×10^{-4} s^{-1}, we obtain $C_1 = 0.64$. The area of the low-latitude upwelling A is set to 2.5×10^{14} m^2, the circumference of the Earth at Drake Passage latitudes is $L_x = 25\,000$ km, the constant reference density is $\rho_0 = 1000$ kg m^{-3}. The magnitude of the windstress τ and the eddy diffusion coefficient A_I are varied.

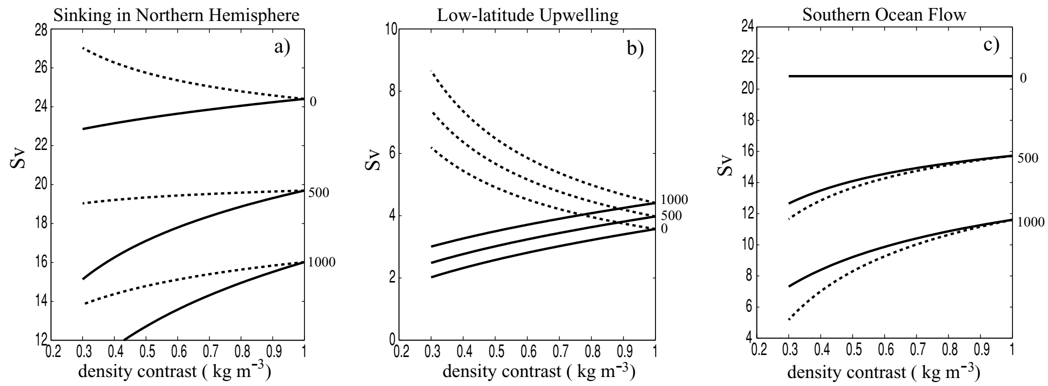

Fig. 2. Dependence of northern hemisphere sinking (a), low-latitude upwelling (b) and the Southern Ocean upwelling (c) on the density contrast parameter for the cases of constant vertical diffusivity (solid) and constant energy of mixing (dashed). The different curves represent different levels of eddy activity in the Southern Ocean as indicated by the eddy diffusion coefficient A_I, given on the right axis in $m^2\ s^{-1}$. The windstress is 0.1 Pa.

Equation (8) is solved analytically, using a range of values for $\Delta\rho$ and considering only physically meaningful solutions (which, for example, do not produce negative depth of the pycnocline). Two cases are analyzed, with either a constant vertical diffusivity of $k_v = 10^{-5}\ m^2\ s^{-1}$ or a constant energy of mixing ε. In the latter case, expression (5) is substituted into eq. (8). ε is avaluated from eq. (5) assuming $k_v = 10^{-5}\ m^2\ s^{-1}$ and $\Delta\rho = 1\ kg\ m^{-3}$, for consistency with the case of constant vertical diffusivity. Having solved eq. (8) for D, the three flow components (T_n, T_u and T_s) are readily found.

Figure 2 shows the dependence of the three flow components on the density contrast, using three different values for A_I. As long as the eddy-induced transport across the ACC is zero (i.e. $A_I = 0$), the Southern Ocean upwelling is controlled only by winds and for a constant windstress (set in this solution to 0.1 Pa) the Southern Ocean upwelling is constant (Fig. 2c). In this case the two vertical mixing schemes give an *opposite* dependence of the water sinking in NH on the density contrast (Fig. 2a), similar to the single-hemisphere model of Nilsson and Walin (2001) discussed in the previous section. However, activating the eddy-induced transport across the ACC by increasing A_I above a critical value (see below) results in qualitatively the *same* dependence of the NH sinking on $\Delta\rho$, regardless of whether the vertical diffusivity or the energy available for mixing is held fixed. For both vertical mixing schemes, the rate of NH deep water formation increases with $\Delta\rho$ (Fig. 2a). However, deep water formation in the fixed energy case has weaker sensitivity to the density contrast than in the fixed diffusivity case.

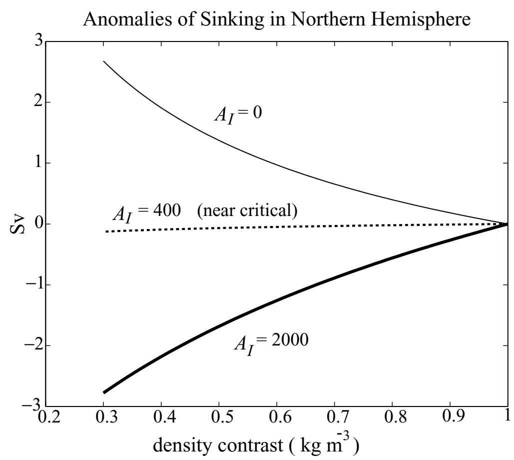

Fig. 3. Dependence of northern hemisphere sinking anomaly (relative to the corresponding sinking at $\Delta\rho = 1\ kg\ m^{-3}$) on the density contrast parameter for the case of constant energy of mixing, using three different values of eddy exchange coefficient (given in $m^2\ s^{-1}$): $A_I = 0$ (thin solid), $A_I = 2000$ (thick solid) and $A_I = 400$ (dashed). The windstress is 0.1 Pa.

As follows from Fig. 2a and further illustrated in Fig. 3, for the given set of parameters the critical value for A_I is below 500 $m^2\ s^{-1}$ (near 400 $m^2\ s^{-1}$, see Fig. 3). However, this critical value is within the range of available estimates for A_I, which is between 100 and 3000 $m^2\ s^{-1}$ [see Gnanadesikan (1999) and references therein]. This result suggests that there may

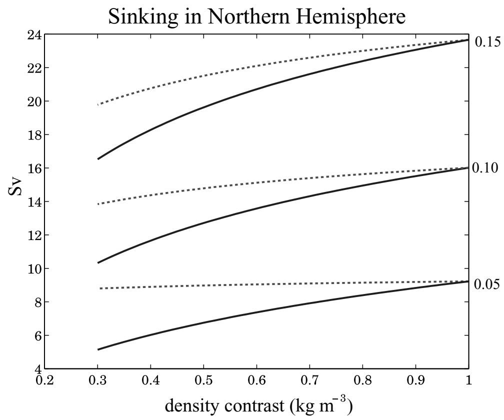

Fig. 4. Dependence of northern hemisphere sinking on the density contrast parameter for the constant vertical diffusivity (solid) and constant energy of mixing (dashed) under different magnitudes of southern hemisphere windstress. The windstress magnitudes are given on the right axis in Pa.

exist a threshold for the eddy-induced transport across the ACC above (below), which the northern sinking increases (decreases) with $\Delta\rho$. It highlights the role of ACC eddies in controlling the sensitivity of deep water formation in the NH to the poleward density contrast.

Adopting a value for the eddy diffusion which is closer to the middle of the above range, such as $A_I = 1000$ m^2 s^{-1} and increasing τ, brings the two mixing cases more closely together in terms of the sensitivity of deep water formation to $\Delta\rho$ (Fig. 4). That is, the sensitivity of NH deep water formation to the equator-to-pole density difference for both vertical mixing schemes is larger for stronger Southern Ocean winds. However, the two mixing schemes have different dependence of low-latitude upwelling on $\Delta\rho$ (Fig. 2b), which does not change qualitatively with the introduction of Southern Ocean upwelling.

4. Discussion and conclusions

Nilsson and Walin (2001) employed a two-layer single-hemisphere model to analyze the impact of freshwater forcing on the thermohaline circulation. They compared two different schemes for diapycnal mixing, which assume either fixed diapycnal diffusivity or fixed mixing energy. The model approach used in Nilsson and Walin (2001), as in other classical scalings, assumes that the water sinking in high latitudes is fully balanced by diapycnal flow in low latitudes, i.e. without involving deep water upwelling in the Southern Ocean. Nilsson and Walin (2001) showed that in their model, a fixed diapycnal diffusivity leads to a thermohaline circulation intensity which increases with the equator-to-pole density contrast, whereas the fixed mixing energy case leads to a circulation which decreases with the density contrast. The latter case in their model leads to a circulation for which a positive freshwater forcing in high latitudes acts to strengthen the MOC (as a booster in the terminology of Nilsson and Walin, 2001).

Here we re-examined this conclusion, using a model similar to that used by Nilsson and Walin (2001) but extended to include both hemispheres. Our motivation is that the assumption of no Southern Ocean upwelling, adopted by Nilsson and Walin (2001), is difficult to justify. Moreover, it has been argued (e.g. Webb and Suginohara, 2001) that more than half of the North Atlantic Deep Water (NADW) is brought up to the surface in the Southern Ocean. Thus, we allow for a fraction of deep water to upwell in the Southern Ocean. The latter is controlled by the northward Ekman transport at Drake Passage latitudes and the return eddy-induced flow across the ACC. We show that the case when Southern Ocean upwelling is controlled *only* by Ekman transport is similar to the single hemisphere case analyzed by Nilsson and Walin (2001). That is, the fixed diffusivity and fixed mixing energy schemes have an *opposite* dependence of deep water formation on the equator-to-pole density contrast. However, when Southern Ocean upwelling is controlled by both Ekman transport *and* strong enough eddy-induced transport across the ACC, the two mixing schemes have qualitatively *similar* dependence of the deep water formation on the equator-to-pole density contrast. That is, the rate of sinking in the NH increases with the equator-to-pole density contrast, regardless of whether the diffusivity or the mixing energy is held fixed. In other words, taking into consideration the deep water upwelling in the Southern Ocean and the dynamical effect of eddies in the ACC can turn the freshwater boosted regime of the overturning circulation, described by Nilsson and Walin (2001), into a freshwater impeded regime. However, by fixing the external mixing energy, the rate of deep water formation becomes less sensitive to the equator-to-pole density contrast compared to the fixed vertical diffusivity case. Hence, the ACC eddies and vertical mixing jointly control the response of the overturning circulation to changes in the equator-to-pole density contrast.

Simple models may not always correctly capture oceanic behavior quantitatively; however, an analysis of their qualitative solutions can be quite useful. Our results suggest that parametrizations employed in GCMs to represent the Southern Ocean eddies *and* vertical mixing can be crucial when simulating climate transitions between different regimes of overturning circulation. This should be kept in mind, for example, when investigating the impact of freshwater forcing on NADW formation. Also, our results have important implications for simulations of future climates, where often the equator-to-pole density difference reduces in response to increasing greenhouse gases. It is likely that in such simulations the climate response is overestimated when vertical diffusivity is decoupled from the simulated oceanic circulation. Finally, we have shown that the classical scaling laws which do not take into consideration both a dependence of vertical mixing on oceanic circulation and Southern Ocean dynamics are too simplified and should be interpreted accordingly.

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