The dependence of AABW transport in the Atlantic on vertical diffusivity

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Abstract. Simple theoretical arguments are employed to study the dependence of the volume and transport of the Antarctic Bottom Water (AABW) in the Atlantic Ocean on the vertical diffusivity. We have found that while the vertical extent of the AABW cell decreases with the intensification and deepening of the North Atlantic overturning cell, the transport of AABW into the Atlantic increases. The latter fact is explained by the increase in the deep meridional pressure gradient, which drives the flow. An estimate of the AABW transport is then derived from the density balance in the deep western boundary layer.

1. Introduction.

Antarctic Bottom Water (AABW) forms in the Weddell and Ross seas at the Antarctic coast, where the cold surface waters become salty due to salt rejection during the local winter and sink to the bottom of the Southern Ocean. This water then propagates northward, mixes with Circumpolar Deep Water and forms northward-moving tongue of very dense bottom water.

The transport of AABW is sensitive to a number of factors in a numerical model. England (1992), (1993) and Stocker, et al. (1992) demonstrated the importance of surface salinity at the Antarctic coast in setting density properties and ultimately the formation rate of this water mass. The relationship between the AABW transport and the strength of the NADW circulation is rather complex and has been addressed in a number of studies. Intensification of NADW formation generally leads to a reduction of AABW in the deep ocean (Cox, 1989), AABW recedes southward and is confined to greater depths. England (1993), on the other hand, suggests that the density of AABW in turn controls the inter-ocean exchange of NADW. A decrease in characteristic density of AABW allows more NADW to leave the Atlantic basin in his model. Also important is the mixing of temperature and salinity by mesoscale eddies, parameterized by the Laplacian diffusion in numerical models. In particular, decrease in the horizontal mixing intensity reduces the erosion of AABW during its sinking to the ocean floor, and enhances its density and northward transport (England, 1993).

The important role of the vertical diffusivity in the ocean circulation has been long known to the oceanographic community (Stommel, 1958; Munk, 1966). Numerical models show that water mass properties and their formation rates are very sensitive to the value of this parameter. The main known effects of increasing vertical diffusion are the deepening of the thermocline, the intensification of upwelling at low latitudes, and the enhancement of North Atlantic Deep Water (NADW) formation. In his sensitivity study, F. Bryan (1987) demonstrated a cube root dependence of both the thermocline depth and the strength of the North Atlantic (NA) overturning on the vertical diffusion coefficient, $k_v$. His model, however, only included the Northern Hemisphere and therefore excluded effects due to Southern Ocean processes.

In a recent study, A. Gnanadesikan (1999; G99 hereafter) proposed a simple theoretical model of the oceanic pycnocline, relating the pycnocline depth, the NA overturning $(T_O)$, and the Southern Ocean winds and eddies. He assumed that the oceanic pycnocline is maintained by the balance between low latitude heating and Southern Ocean freshening and upwelling of cold pycnocline waters; see also Stommel, 1958.

Our study focuses on the role of vertical diffusivity in setting the properties of northward transport of AABW into the Atlantic in our numerical model with simplified geometry. The term “AABW” is used rather loosely in our study with no bottom topography. In the real Atlantic, the corresponding northward flowing bottom water is likely to be a combination of AABW and Circumpolar Deep Water. However, since this water mass is the deepest and the densest in a model at mid-and low latitudes, our scaling analysis should remain valid in more realistic configuration. The total transport may, however, be overestimated due to the absence of topography, as suggested by Hovine and Fichefet (1994).

The changing vertical diffusivity in our study has a double effect on the bottom water transport: the diffusivity intensity controls the deep diffusion-induced upwelling in the low latitudes; and it also alters the NADW overturning which, in turn, affects the distribution of AABW in the Atlantic basin. In what follows we will estimate the effects of both of these processes.

2. Numerical results.

We carried out a series of experiments in a GCM (MOM 1.0) within an idealized global geometry under restoring boundary conditions for both temperature and salinity. The model is forced by zonally uniform wind stress from Helleman and Rosenstein (1983). The model's domain, forcing profiles, and mixing parameters are identical to those described in Goodman (1998), with the exception of the vertical diffusivity. In Goodman (1998) a vertically varying $k_v$.
was employed, whereas here, it is constant with depth. We therefore do not account for the fact, that vertical mixing is distributed unevenly in the World Ocean with higher values concentrated near rough topography (Polzin et al., 1997, etc.). The described here experiments are identical to one another except for the value of \( k_z \), which varies from \( 10^{-6} \text{ m}^2\text{s}^{-1} \) to \( 2 \times 10^{-2} \text{ m}^2\text{s}^{-1} \). A horizontal diffusion coefficient is \( 10^3 \text{ m}^2\text{s}^{-1} \).

As \( k_z \) increases, we observe both deepening and strengthening of the NA overturning cell (figs. 1a,c and 2). The values of \( T_n \) from the series of GCM experiments versus the corresponding values of \( k_z \) are shown in fig. 3a. As one can see, the curve \( T_n + B k_z^{2/3} \) provides a very good fit to the experimental points. The 2/3-power dependence on the vertical diffusivity contradicts findings by F. Bryan (1987), who reports a 1/3-power dependence in his model with a single cell, but is in general agreement with theoretical scaling (Marotzke, 1997; G99).

The lower meridional circulation cell in the Atlantic, which exists due to the northward propagation of AABW, responds differently to increasing diffusivity. The northward transport of AABW increases, while the thickness (2\( H_a \) in fig. 2) of the cell decreases. Since the northward flow is confined to a western boundary current of invariant width, set by the momentum viscosity (see the next section), the decreasing thickness also implies a reduced volume of AABW present at any moment. An increasing transport \( T_n \) along with the decreasing vertical extent of the cell thus may seem counterintuitive. In what follows, we provide an explanation for this phenomenon.

3. Dynamical Interpretation of the Results.

We start from deriving a relation between transport in the NADW cell and its thickness. We closely follow the argument presented in G99 to relate the transport of NA cell \( T_n \) and its thickness \( H \). The reader is referred to G99 for the details of the derivation. Boundary layer theory implies that a balance exists between the horizontal diffusion of momentum and the meridional pressure gradient. This allows us to estimate the meridional velocity in a boundary layer of the width \( L_m = (A_v/\beta)^{1/2} \):

\[
\frac{A_h}{L_m^2} \frac{V}{L_m} \sim \frac{1}{\rho_0} \left[ \frac{\partial \rho}{\partial y} \right] \quad \rightarrow \\
V \sim L_m^2 \frac{1}{\rho_h \rho_0} \left[ \frac{\partial \rho}{\partial y} \right] = \frac{1}{\beta L_m \rho_0} \frac{g H \Delta \rho}{\ell_u} \quad \rightarrow \\
T_n = V \times H \times L_m = C \frac{g \Delta \rho}{\rho_0 \beta k_z} H^2
\]

where \( C \) is the constant that incorporates the effects of

Figure 1. Left column: Meridional overturning streamfunction (Sv) in the model Atlantic. Right column: potential density (referenced to 4000m) zonally averaged from 180E to 169W to in the bottom 2000m of the Atlantic basin. Two values of vertical diffusivity \( k_z \): (a,b) \( 10^{-6} \text{ m}^2\text{sec}^{-1} \), and (c,d) \( 2 \times 10^{-2} \text{ m}^2\text{sec}^{-1} \).
geometry and boundary layer structure. \( \frac{\partial p}{\partial y} \) is scaled as \( gH\Delta p/L_m \), where \( L_m \) is the meridional extent of the density gradient in the upper layers and \( \Delta p \) is the meridional density contrast in the upper layers.

Note that according to (1), transport \( T_a \) depends on both density contrast \( \Delta p \) and the cell thickness \( H \). In this study we assume that \( \Delta p \) is fixed, mainly because of the unchanged restoring boundary conditions. The transport of the NADW cell thus increases only due to its deepening and there is a direct relationship between cell’s thickness \( H \) and \( T_a \). Other scenarios are possible, in which the transport of a cell intensifies mainly due to the increase in \( \Delta p \), as for example in the AABW cell in this study (see below). Another example is a single-hemisphere thermohaline overturning, with a cell reaching the bottom. The thickness of the cell is thus fixed; its transport can however change due to changes in the meridional density gradients in the upper half of the cell.

The decrease in \( H_a \) is then explained by the downward expansion of the upper overturning cell in the North Atlantic. Indeed, the sum of \( H \) and \( H_a \) (see fig. 2) tends to remain constant in the model (figs. 1a,c) and AABW thus occupies the volume left available by the upper overturning cell. We can determine \( H \) from the equation (1) and \( H_a \) as equal to \( 5000 \text{m}^2/\text{m} \). We use constant values of \( L_m = 1,500 \text{km} \) and \( \Delta p = 0.1 \text{kg m}^{-1} \). The resulting values of \( H \) and \( H_a \) agree well with those from the numerical experiments (fig. 3b).

We now estimate the transport of the AABW cell. As in the upper ocean, we assume that the northward flow of AABW is concentrated within the boundary current of width \( L_m \) and is driven by the meridional gradient in pressure. Since the meridional flow changes direction at \( z = -2H-H_a \), the horizontal pressure gradients have to be negligible at that depth. Then, in order to estimate the pressure gradient we only need to account for the density contrast within the AABW layer. The equation 1 now takes the following form:

\[
T_a = C_a \frac{g\Delta p_a}{\rho_0 \beta a} H_a^2
\]

where \( L_m \) is the meridional length scale of the AABW and \( \Delta p_a \) is the meridional density contrast. The constant \( C_a \) depends on geometry of the boundary current. As we know, the thickness \( H_a \) of the AABW cell decreases in response to intensifying NA overturning. According to (2), the decrease in \( H_a \) will act to slow AABW transport. On the other hand, the intensified diffusion and the increased downward advection of lighter waters by NADW cell will decrease density in the low latitudes (figs. 1b,d; \( \rho_L - \Delta \rho_a \) in fig.2). The lightening of deep low-latitude water enhances its density difference \( \Delta p_a \) with dense bottom water entering Atlantic from the Southern Ocean (see figs. 1b,d). Increasing \( \Delta p_a \) will act to increase \( T_a \).

Our next step is to estimate \( \Delta \rho_a \). For simplicity, we assume here a linear dependence of density on temperature and salinity; density then satisfies the same equations as temperature and salinity. Consider then density balance in the AABW region in the Atlantic (fig.2). The water with density \( \rho_a \) enters from the south at the rate \( T_a \); a part of it \( (T_a) \) then upwells through the upper surface \( A_o \) of the boundary current, the remainder leaves in the north with density \( \rho_n - \Delta \rho_a \). The resulting convergence of the horizontal advective density flux is balanced by the upwelling of denser waters and density loss through the vertical diffusion:

\[
T_a \rho_a - (T_a - T_a) \rho_a - \Delta \rho_a = k_a A_o \frac{\delta \rho}{H_a} + T_a \rho_a \rightarrow
\]

\[
\Delta \rho_a (T_a - T_a) = k_a A_o \frac{\delta \rho}{H_a}
\]

where \( k_a \) is vertical diffusivity and \( \delta \rho \) is the vertical density contrast. Note that horizontal diffusion is neglected in (3) mainly because of its smallness. In addition, diazocic diffusion associated with lateral diffusion should in principle be even smaller since tracers are widely believed to mix mainly along isopycnic surfaces.
subject of a further study. As in the NADW cell, the transport of the AABW is also proportional to the square of layer thickness and a meridional density gradient, but it intensifies for different reasons. While the increase in the \( T_n \) can be attributed to increasing thickness with unchanged density gradient, the intensification of the AABW transport \( T_n \) is explained by the strong increase in the meridional density contrast between dense water flowing from the Southern Ocean and lighter water at low latitudes. The decrease in the low-latitude density is described through the density balance in the AABW layer. The convergence of the horizontal density flux is balanced by the density loss through the vertical diffusion and upwelling. The diffusion flux increases rapidly with increasing diffusivity both due to the increase in \( k_\nu \) and increase in the vertical stratification. The latter effect is caused by the deepening NADW cell, which pushes isopycnal surfaces closer together.

Based on our results, we expect the rate of AABW formation to be largely determined by the mixing processes and density distribution in the deep ocean. In contrast, the volume of AABW present at any time is more likely controlled by the amount of space left available for this water mass by the upper overturning cell and the bathymetry of the ocean bottom.

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