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Atlantic pycnocline theory scrutinized using a coupled climate model

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[1] Simulations with changed Southern Ocean wind-stress, oceanic vertical mixing, surface freshwater forcing and global warming confirm the basic equations of Gnanadesikan's (1999) theory for the Atlantic: one vertical scale, the pycnocline depth D, contributes inversely proportional to low-latitudinal upwelling and linearly to Southern Ocean eddy transport. The maximum Atlantic overturning is confirmed to be quadratic in D but is also shown to be linear in a meridional density difference $\Delta\rho$. Our simulations strongly suggest that the theory needs to be complemented by a dynamical equation for $\Delta\rho$ since changes in both D and $\Delta\rho$ are significant and mutually independent. While, under global warming D varies strongly and $\Delta\rho$ is practically constant, the situation is reversed for altered surface freshwater fluxes. Similarly, variations in the meridional length scale of Southern Ocean outcropping require a dynamical equation in order to capture the fundamental behavior of the Atlantic meridional overturning circulation. **Citation:** Levermann, A., and J. J. Fürst (2010), Atlantic pycnocline theory scrutinized using a coupled climate model, *Geophys. Res. Lett.*, *37*, L14602, doi:10.1029/2010GL044180.

1. Introduction

[2] The Atlantic meridional overturning circulation (AMOC) plays a fundamental role for global climate [*Vellinga and Wood*, 2002, 2007]. In its conceptual understanding, two qualitatively different approaches have been undertaken. On the one hand, *Stommel* [1961] stresses the importance of an oceanic meridional density difference $\Delta\rho$, that controls the North Atlantic Deep Water (NADW) formation and thereby the AMOC strength. In this context, *Rahmstorf* [1996] finds a linear relation between NADW formation and the meridional density difference in a global ocean circulation model. A scaling of this sinking process [*Bryan*, 1987] indicates that such a linear flow law would only hold, if the rate of northern sinking was independent of a vertical scale depth D, the pycnocline. Other simulations with ocean general circulation models (OGCM) confirm this linearity [e.g., *Hughes and Weaver*, 1994; *Levermann and Griesel*, 2004; *Griesel and Morales-Maqueda*, 2006] and therefore reinforce the view of a stable pycnocline depth. On the other hand, focusing on processes in the Southern Ocean (SO), *Gnanadesikan* [1999, hereafter G99] presented a framework, which directly links the strength of the overturning rate to the pycnocline depth D. This idealized model, which has been used as a paradigm for the meridional overturning circulation [*Kamenkovich and Sarachik*, 2004; *Marzeion and Drange*, 2006; *Johnson et al.*, 2007], solely allowed for variations in the universal pycnocline depth D, but kept the meridional density difference $\Delta\rho$ as an external parameter. Each of the two approaches emphasizes a different oceanic property whose variations are assumed to control the overturning strength. Here we investigate variations of both the density difference $\Delta\rho$ and the pycnocline depth D by use of a variety of experimental setups. We find that the idealized model of G99 is an appropriate description for the large-scale oceanic circulation, but has to be extended by the dynamics of $\Delta\rho$.

[3] G99 identifies four essential mechanisms controlling the magnitude of the overturning circulation. These processes are described through a vertical scale D, identified as the oceanic pycnocline depth. The upwelling in the SO m_W has its origin in predominant westerly winds and the zonal ocean band at the latitude of the Drake Passage. A scaling of this Drake Passage effect shows no direct dependence on stratification in the SO and is therefore parameterized to be independent of D. Several approaches scale the northward volume transport m_N, either employing boundary layer theory [*Gnanadesikan*, 1999] or geostrophic balance [*Bryan*, 1987; *Marotzke*, 1997; *Johnson and Marshall*, 2002] to describe the sinking rate or maximum overturning, which are used synonymously in this framework. In principle, these different studies agree on the obtained scaling for the overturning strength

$$m_N = C_N \cdot \Delta\rho \cdot D^2. \tag{1}$$

In G99, C_N is a constant which comprises geometry and boundary layer structure, and $\Delta\rho$ is a meridional density difference at or below the pycnocline depth D. *de Boer et al.* [2010] showed that the relation does not hold for a density difference in the upper ocean layers. The quadratic dependency on D arises from vertical integration and the representation of the meridional pressure difference as $D \cdot \Delta\rho$. The third process included in G99 is upwelling in low latitudes, m_U, described by a vertical advection-diffusion balance [*Munk and Wunsch*, 1998] which yields

$$m_U = C_U \cdot \frac{\kappa}{D}. \tag{2}$$

The constant C_U is associated with the effective upwelling area and κ denotes the average vertical diffusivity. Eddies contribute to the large scale advection of tracers and a parameterization from *Gent and McWilliams* [1990, here-

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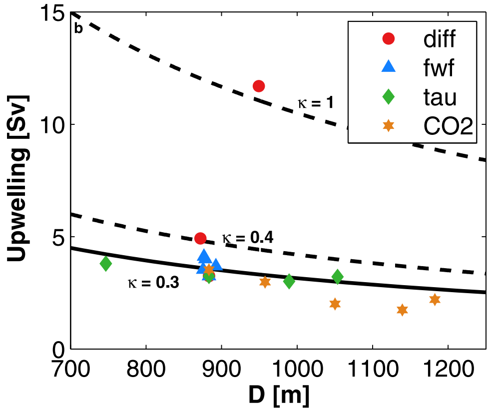

Figure 1. Low-latitudinal upwelling m_U is inversely proportional to D. Symbols represent different experimental groups with large variations in D. Experiments with standard vertical diffusivity ($0.3 \cdot 10^{-4}$ m^2/s) follow G99's scaling with D. Variation in vertical diffusivity (red dots) confirm the predicted scaling with κ (dashed lines).

after GM90] provides an additional equation for eddy-induced tracer transport m_E linear in D

$$m_E = C_E \cdot \frac{D}{L_y}. \quad (3)$$

The proportionality constant C_E comprises flow geometry and a thickness diffusion coefficient. Potential energy for the eddy induced transport is provided by outcropping isopycnals in the SO whose slope D/L_y is determined by its meridional extent L_y. While G99 applied his scaling analysis to the entire ocean keeping the meridional density gradient fixed, *Schewe and Levermann* [2010] showed that the overturning of each basin scales with a corresponding density difference along the western boundaries of the Atlantic, Pacific and Indian oceans, respectively. Since current observations do not support significant overturning in other basins we restrict our analysis to the Atlantic.

2. Model and Experiments

[4] In order to investigate the processes included in the G99 model, experiments were carried out with changed SO wind stress, vertical diffusivity, anomalous surface freshwater fluxes and global warming scenarios. All results are based on simulation with the model of intermediate complexity CLIMBER-3α [*Montoya et al.*, 2005]. The oceanic component (based on the MOM-3 GFDL code) has a horizontal resolution of 3.75° and 24 variably spaced vertical levels. The influence of baroclinic eddies on tracer transport is included via the parameterization of GM90 with a constant thickness diffusivity of 250 m^2/s.

[5] The first group of experiments (denoted "CO_2") comprises five equilibrium simulations with 1, 2, 4, 8 and 16 times the pre-industrial CO_2 concentration of 280ppm (described by *Levermann et al.* [2007]). Another set of steady state experiments ("fwf") investigates the influence of anomalous freshwater fluxes to the North Atlantic between 30°N–50°N, analog to *Levermann et al.* [2005] with 0.1, 0.2, 0.3 and 0.35 Sv. The third set of experiments ("tau") varied the zonal wind stress in the SO by multiplication with a factor α = 0.5, 1.0, 1.5 and 2 in the latitudinal band between 30°S and 71.25°S as first suggested by *Toggweiler and Samuels* [1995] and used to investigate the relation between density differences and Drake Passage effect in CLIMBER-3α by *Schewe and Levermann* [2010]. For the last set of equilibrium runs ("diff"), the vertical background diffusivity κ is varied from 0.3, 0.4 to $1.0 \cdot 10^{-4}$ m^2/s [*Mignot et al.*, 2006]. All simulations were integrated for at least 5000 years with constant boundary conditions to allow quasi-equilibration of the deep ocean.

[6] Most experiments are closely linked to one of the transport mechanisms of G99. Changing κ has an immediate effect on low-latitudinal upwelling m_U, while the fwf-experiments alter the meridional density gradient in the North Atlantic. The tau-experiments change the magnitude of the Ekman transport in the SO, which also influences the local eddy activity [*Hallberg and Gnanadesikan*, 2006]. In contrast to these experiments, the CO_2-experiments cause a global warming which affects the ocean in many aspects. Thus, this set of experiments is suitable to study the theory in a comprehensive way.

3. Low-Latitudinal Upwelling and SO Eddies

[7] According to G99, low-latitudinal upwelling m_U is inversely proportional and eddy-induced transport m_E directly proportional to the pycnocline depth D. Following the definition of *Gnanadesikan et al.* [2007] (with horizontal integration between 80°W–0° and 20°S–20°N), the experiments performed with CLIMBER-3α span an interval of $D \in$ [750 m, 1250 m] (Figure 1). At this depth the vertical resolution of the model is about 200 *m*, which means that three model layers are covered. Since D is determined from the *e*-fold depth of an exponential fit to the vertical density profile, we obtain continuous values of D which are influenced by the whole column. For this reason the variations in the pycnocline depth D are significant in the considered experiments, despite the model's coarse resolution.

[8] The low latitudinal upwelling m_U is obtained as the difference between the maximum of the Atlantic stream function in the North Atlantic (20°N–70°N and 400–5000 m) and the SO outflow (maximum of stream function at 30°S) subtracting a small overturning cell in the Mediterranean which accounts for up to 3 Sv and is not meant to be included in the analysis of the Atlantic overturning. The experiments with changes in surface freshwater flux, in SO wind stress and global warming have a constant vertical diffusivity of $0.3 \cdot 10^{-4}$ m^2/s and confirm the inverse proportionality between m_U and D (Figure 1). The theoretical curve (solid line) is obtained with a constant effective horizontal upwelling area of 10^{14} m^2. By applying the same area, variations in κ are equally well captured by the theory (dashed curves). The fact that the effective area exceeds the surface area of the low-latitudinal Atlantic is likely due to spurious upwelling along the continental boundaries, which is a model artifact observed in coarse resolution models [*Yang*, 2003]. Vertical mixing in our simulations is furthermore enhanced along rough topography which increases the corresponding mixing coefficient along the coast up to values of $2 \cdot 10^{-4}$ m^2/s. These represent an offset of downward heat transport which results in an overestimation of the effective upwelling area.

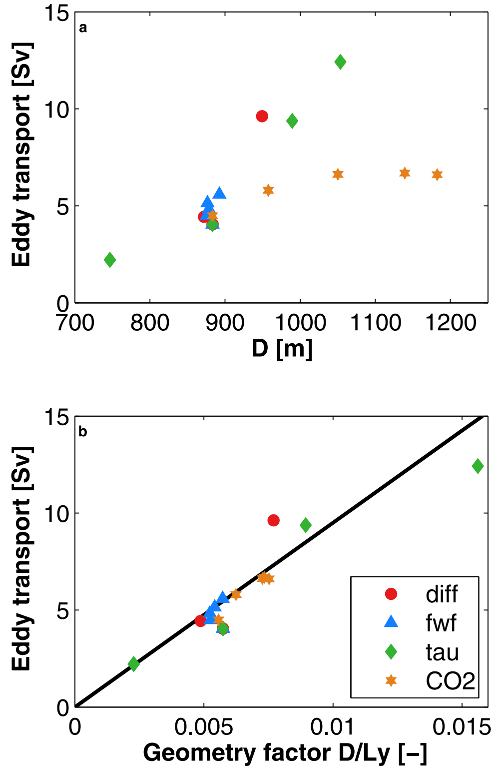

Figure 2. (top)The eddy-induced transport m_E from GM90 approximation as a function of D does not show linearity. (bottom) Correcting for the meridional extent of the outcropping L_y confirms the theoretical approach by G99.

[9] The eddy-induced return flow m_E defined here as the maximum of the negative eddy stream function in the SO (70°S–0°, 200–2000 m) does not exhibit a linear behavior in D (Figure 2, top) as could be expected from equation (3) for constant L_y and C_E. The largest discrepancy appears in the warming and SO wind experiments. This is due to variations in the meridional length scale for outcropping L_y. The experiments collapse onto a straight line (Figure 2, bottom), when the meridional extent of the outcropping region L_y is taken into account. In order to define L_y, we determine the density at 35°S at the depth D. This position is chosen to be just north of the outcropping region but close enough to assure that the corresponding isopycnal does indeed outcrop

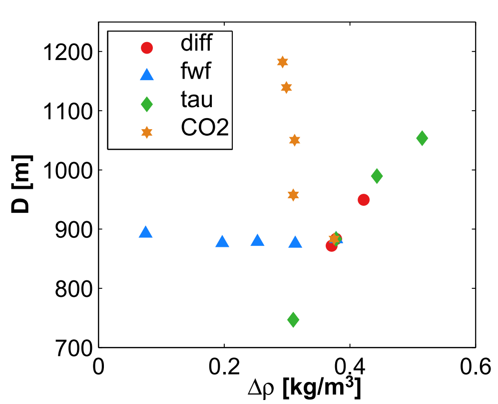

Figure 3. Experiments conducted with CLIMBER-3α exhibit strong variations in both D and $\Delta\rho$.

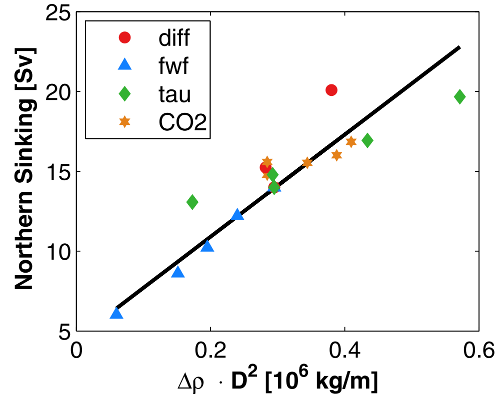

Figure 4. The northern sinking m_N is depicted as a function of the combined expression $\Delta\rho \cdot D^2$. Various equilibrium experiments in CLIMBER-3α confirm a linear relation, which is predicted by theory.

in the SO. L_y is then defined as the meridional distance between the crossing of the specific isopycnal at 450 m and 80 m. Values for L_y group around $1.6 \cdot 10^5$ m with small deviations, except for the wind and warming experiments. For example, a doubling of SO wind stress halves L_y, which can be explained by an increased Ekman transport from the south compressing the outcropping isopycnals [*Schewe and Levermann*, 2010].

4. Northern Sinking

[10] The scaling for the northern sinking m_N (equation (1)) depends on variations in D and $\Delta\rho$. The meridional density difference $\Delta\rho$ is computed between regions along the Western Boundary of the American coast; (70°W–50°W, 35°N–45°N) and (45°W–20°E, 35°S–35°N) at a depth of 800 m. Figure 3 shows that both quantities are relevant for the volume transport and vary significantly across the different experiments. The CO_2 experiments, for example, result in strong variations in D, while the meridional density difference is relatively constant. The situation is reversed in the freshwater experiments, which exhibit almost no change in D, but strong variations in $\Delta\rho$. SO wind experiments yield variations in both quantities. Thus, the type of experiment is decisive for the respective role of D or $\Delta\rho$ and neither can be neglected.

[11] Defining northern sinking m_N as the maximum of the Atlantic stream function in the same region as above (subtracting the Mediterranean overturning cell as discussed above), reveals a scaling with $\Delta\rho \cdot D^2$ (Figure 4) for all experiments. Discrepancies can be attributed to possible variations in the coefficient C_N that comprises, for example, the geometry of the volume transport. We find a small m_N-offset of about 4 Sv which is mainly due to spurious upwelling at the boundary and over rough topography as discussed above and shown by *Mignot et al.* [2006]. Figure 4 is robust to changes in definition of $\Delta\rho$.

5. Conclusion and Discussion

[12] We investigate the applicability of the simple predictive theory of the oceanic pycnocline depth [*Gnanadesikan*, 1999] in a coupled climate model with comprehensive but

coarse oceanic component. A variety of model experiments confirm the universal role of the pycnocline depth D as defined by *Gnanadesikan et al.* [2007]. G99's expressions for low-latitudinal upwelling and northern sinking are applicable under variations of SO wind stress, vertical diffusivity and atmospheric CO_2 concentrations. As shown earlier [*Levermann and Griesel*, 2004] some variations in the Atlantic overturning are not captured by changes in D. For example, simulations with varying surface freshwater forcing change the meridional density gradient and the northern sinking but keep D constant.

[13] Our simulations thus suggest that the seemingly contradictory approaches of a dominant influence of the meridional density difference [*Stommel*, 1961; *Rahmstorf*, 1996] as opposed to the vertical density structure [*Gnanadesikan*, 1999] can be consolidated by implication of a dynamical equation for the meridional density difference in the North Atlantic as suggested by *Johnson et al.* [2007]. In contrast to G99, this restricts the scaling analysis to the Atlantic basin. *Schewe and Levermann* [2010] showed for the model used here that the overturning of each basin (Atlantic, Pacific and Indian Ocean) is proportional to the corresponding density difference along the western boundary of that particular basin. If that holds for the real ocean, it is not possible to capture the full range of possible behavior of the global overturning with only one density difference. As shown here the scaling however can be confirmed when applied to one basin only.

[14] The eddy return flow has a special role in this framework. Our simulations show that the behavior of the eddy return flow can only be consolidated with the simple parameterization when taking changes in the horizontal scale for the SO outcropping L_y into account. Its variations require a dynamical equation. One possibility could be the introduction of a meridional density difference in the SO $\Delta\rho_{so}$. The eddy-induced tracer transport m_E could then be parameterized as

$$m_E \propto D \cdot \Delta\rho_{so} \tag{4}$$

This equation is compatible with heat and salinity advection as that of *Johnson et al.* [2007] and represents the baroclinic instability in a similar fashion as *Gent and McWilliams* [1990]. Note further that while SO eddies have been reported to transport large amounts of heat and salinity [*Naveira-Garabato et al.*, 2007] it is not clear whether surface wind-driven upwelling, i.e., actual large scale water volume transport, can be compensated by a net eddy-induced downwelling. In other words G99's scaling is an analysis of the tracer transport. An additional set of equations is necessary for the momentum transport. The applicability of G99's theory to a situation without Atlantic overturning ("off-state") as in *Johnson et al.* [2007] is thus questionable.

[15] In summary, our simulations confirm the approach by *Gnanadesikan* [1999], but suggest a generalization by two additional dynamical equations for $\Delta\rho$ and L_y.

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