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The role of stratification-dependent mixing for the stability of the Atlantic MOC in a coupled global climate model

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Motivation

By adding potential energy to the water column, diapycnal mixing is thought to be one of the driving mechanisms of the meridional overturning circulation (MOC). Today, most state-of-the-art ocean general circulation models calculate the diapycnal background diffusivity independently of the stratification. Results from conceptual models (Nilsson and Walin, 2001), layer models (Marzeion and Drange, 2006), and idealized GCMs (Nilsson and Broström, 2003) suggest that a more physical parameterization could lead to fundamental changes of the stability properties of the MOC with respect to freshwater perturbations. Specifically, increased vertical mixing as a result of weakened stratification could lead to an increase of the MOC as a response to freshwater forcing rather than a decrease.

We show that while the suggested feedback itself may exist, freshwater forcing will not necessarily result in a weaker stratified ocean, so that it is to be doubted that the feedback will have a major impact on the stability of the MOC.

Model Description

The model CLIMBER3α is a coupled climate model of intermediate complexity. It combines a 3d, 24 layer ocean general circulation model based on the GFDL MOM3 code with a statistical-dynamical atmosphere (POTSDAM-2) and a dynamic and thermodynamic sea-ice module. The horizontal resolution of the ocean is $3.75° \times 3.75°$, while the atmosphere uses a resolution of $7.5° \times 22.5°$, assuming a universal vertical structure of temperature and humidity.

Mixing Parameterization

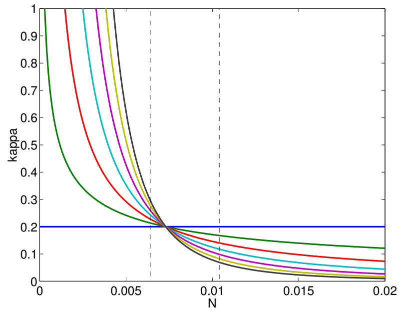


Figure 1: Background diffusivity κ depending on buoyancy frequncy N for values of α between $\alpha = 0$ ($\kappa =$ const.) and $\alpha = 3$ (steepest curve). $N_0 = 0.0073\ \mathrm{s}^{-1}$, $\kappa_0 = 0.2\ \mathrm{cm}^2\mathrm{s}^{-1}$. The black dashed lines indicate the range of N found in the unperturbed experiments around pycnocline depth (i.e. between 450 and 800 m).

The background vertical diffusivity of the ocean κ is calculated depending on the buoyancy frequency N as

$$\kappa = \kappa_0 \left(\frac{N}{N_0}\right)^{-\alpha}$$

where N_0 is chosen to represent the typical value of N in the pycnocline to assure compatibility between the different experiments. The parameter α controls the coupling strength between the stratification and vertical mixing (see fig. 1).

Experiments & Results

Three groups of experiments were conducted: first the model was run to steady state using $\alpha = 1$, $\alpha = 2$, and $\alpha = 3$. Then, a freshwater flux of 0.1 Sv (called sflx01 experiments hereafter) and 0.2 Sv (called sflx02 experiments hereafter) was imposed to the Atlantic Ocean between 20° N and 50° N.

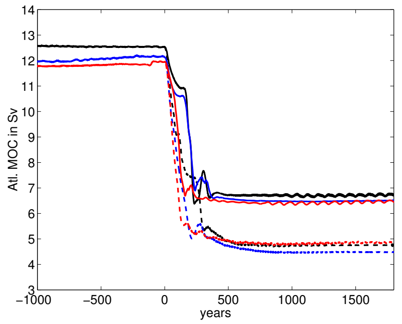
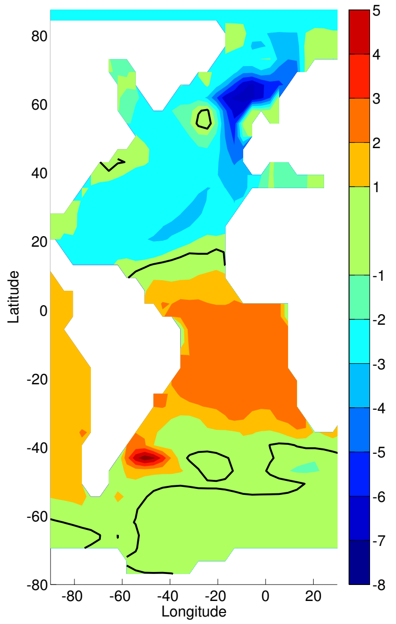
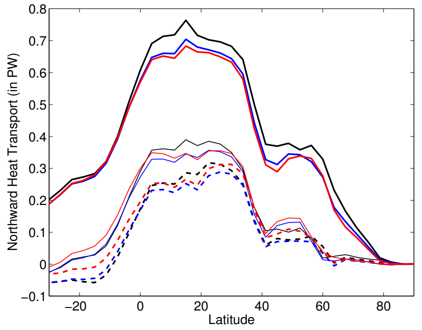


Figure 2: Timeseries of the maximum of the MOC, using $\kappa_0 = 0.2\ \mathrm{cm}^2\mathrm{s}^{-1}$. Freshwater hosing begins at the time $t = 0$. Black lines $\alpha = 1$, blue lines $\alpha = 2$, red lines $\alpha = 3$. Solid lines: unperturbed and sflx01 experiments, dashed lines sflx02 experiments.

No significant difference in the response of the overturning was found for the different α (fig. 2). This was confirmed by repeating the set of experiments using $\kappa_0 = 0.1\ \mathrm{cm}^2\mathrm{s}^{-1}$. The decrease of the overturning, and the corresponding decrease in the oceanic northward heat transport (fig. 3) lead to cooling of the upper North Atlantic and warming of the upper South Atlantic, resulting in the well-know bipolar SST anomaly structure (fig. 4).





▲ Figure 3: Northward Atlantic Ocean heat transport. Colors like left, solid thick lines: unperturbed experiments, solid thin lines: sflx01 experiments, dashed lines: sflx02 experiments.

◀ Figure 4: Temperature anomaly of the upper 400 m (in K) of the $\alpha = 3$ sflx02 run at $t = 2000$ years. The black line is the 0 contour.

Temporal Evolution of Temperature and Salt

At the beginning of the freshwater hosing, the freshwater anomaly is advected northward and penetrates the deep ocean north of the ridges. From there, following the lower limb of the (by now weakened) overturning circulation, it gets advected south in a depth of $1000 - 1500$ m. In the South Atlantic, the heat anomaly in the upper 1000 m builds up quickly (fig. 5).

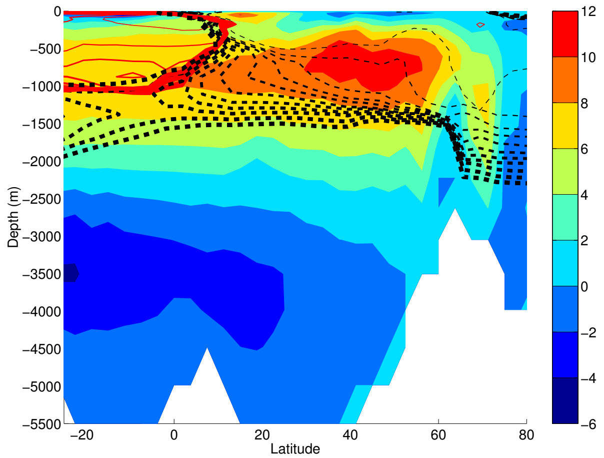


Figure 5, shading: Overturning streamfunction of the unperturbed $\alpha = 3$ experiment in Sv. Black lines: Position of the zonally averaged -0.2 PSU salinity anomaly in the corresponding sflx02 experiment, starting at $t = 150$ years with the thinnest line, linethickness increasing with 50 years timestepping, last (i.e. thickest) line shows the position at $t = 650$ years. Red lines: the same as black, but for the zonally averaged $+0.7$ K temperature anomaly.

Stratification and Diffusivity Changes

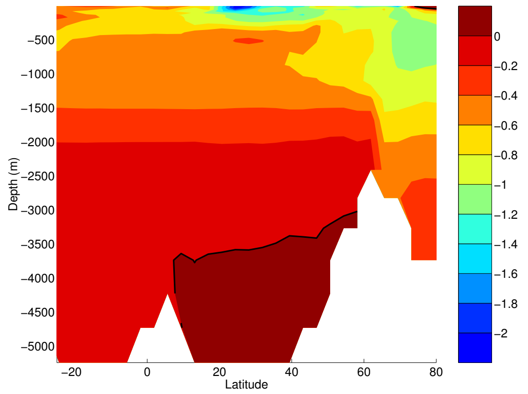


Figure 6: Zonally averaged density anomaly of the $\alpha = 3$, sflx02 experiment at $t = 2000$ years. Note that the downward gradient is close to 0 or positive nearly everywhere.

When approaching steady state, the positive temperature anomaly and the negative salinity anomaly form a common envelope, so that the net effect on the density stratification is rather small. In the entire deep ocean, stratification increases rather than decreases (fig. 6).

This means that also changes of the background diffusivity κ are small, and rather negative than positive at the pycnocline of ~ 500 m (fig. 7).

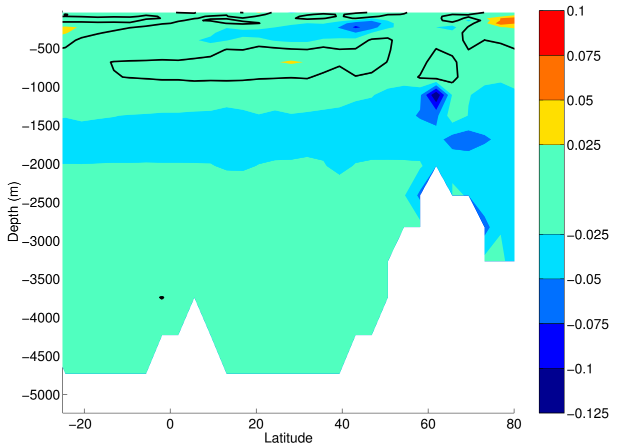


Figure 7: The zonally averaged anomaly of the background diffusivity κ of the $\alpha = 3$, sflx02 experiment at $t = 2000$ years. The black line is the 0 contour.

References:

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