Stratification-dependent Mixing May Increase Sensitivity of Atlantic Overturning to Global Warming

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Abstract

We use the Earth System Model of Intermediate Complexity CLIMBER-3α to investigate the effect of stratification-dependent mixing on the stability of the Atlantic Meridional Overturning Circulation under an idealized CO_2 increase scenario. The vertical diffusivity of the ocean is parameterized as $\varkappa \sim N^{-\alpha}$, where N is the local buoyancy frequency. For all parameter values $0 \le \alpha \le 3$, we find the AMOC to decrease in response to increasing CO_2 concentrations. However, the decrease of the AMOC is significantly stronger for $\alpha \ge \alpha_{cr} \approx 1.5$. Depending on the value of α, two separate model states develop that remain different even after the CO_2 concentrations are stabilized. This is explained by a halt of dense water formation in the subpolar gyre, which is caused by a positive feedback between stratification and mixing anomalies.

1 Model Description & Experiments

The model CLIMBER-3α is a coupled climate model of intermediate complexity. It combines a 3-d, 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 coarse resolution of $7.5° \times 22.5°$, assuming a universal vertical structure of temperature and humidity.

1.1 Mixing Parameterization

The background vertical diffusivity of the ocean $\varkappa$ is calculated depending on the buoyancy frequency N as

$$\varkappa = \varkappa_0 \left(\frac{N}{N_0}\right)^{-\alpha}$$

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).

1.2 Experiments

The model is initialized by equilibriums runs of $\sim$2000 yr. Then, the concentration of atmospheric CO_2 is increased with a rate of 1% yr^{-1} until it reaches the four-fold of the preindustrial value (1120 ppm) after 140 yr. Afterwards, the CO_2 concentration is stabilized (fig. 2).

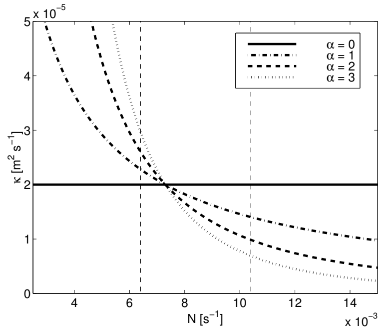

Figure 1: Background diffusivity $\varkappa$ depending on buoyancy frequency N for values of α between $\alpha = 0$ ($\varkappa$ = const.) and $\alpha = 3$ (steepest curve). $N_0 = 0.0073\ s^{-1}$, $\varkappa_0 = 0.2\ cm^2 s^{-1}$. The vertical dashed lines indicate the range of N found in the unperturbed experiments around pycnocline depth (i.e. between 450 and 800 m) in the global ocean.

2 Results

When the increase of the CO_2 concentration is started, the AMOC weakens. A recovery sets in around the time of stabilization of the CO_2 concentration (fig. 2). Two different model states remain when the AMOC approaches equilibrium again after $\sim$1000 yr. They are separated by a critical value $1 < \alpha_{cr} < 1.5$.

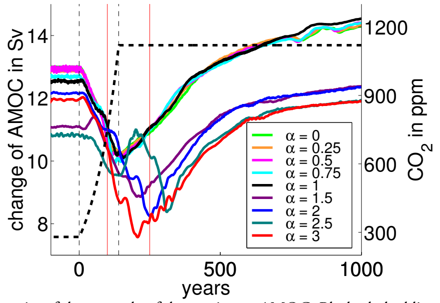

Figure 2: Timeseries of the anomaly of the maximum AMOC. Black, dashed line: CO_2 concentration. The vertical lines indicate the beginning and end of the CO_2 increase (black, dashed) and the time at which the values in figures 3 and 4 are taken (red).

2.1 Effects on Subpolar Gyre Strength

As the CO_2 concentration increases, the warming and strengthened hydrological cycle of the atmosphere increases the buoyancy flux into the North Atlantic Ocean. The decreased surface density implies higher stratification near the surface, which limits vertical diffusion. The higher the value of α, the stronger is the effect on vertical diffusion.

Figure 3 illustrates that during the transient, freshwater is more confined to the surface in the supercritical $\alpha = 2$ case than in the subcritical $\alpha = 1$ case. This leads to a stronger light surface anomaly, and denser subsurface waters (100 - 300 m depth) north of and at the latitude of the Greenland-Scotland Ridge (GSR), i.e. at the northern rim of the SPG.

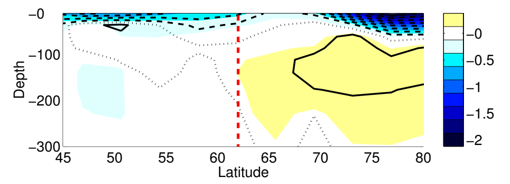

Figure 3: Shading: Change in the zonally averaged Atlantic density deviation from the equilibrium ($\Delta\rho$) caused by an increase of α, i. e. $\Delta\rho(\alpha = 2) - \Delta\rho(\alpha = 1)$ in kg m^{-3}, taken at $t = 100$ yr. Contours: Change in the zonally averaged Atlantic salinity deviation from the equilibrium (ΔS) caused by an increase of α, i. e. $\Delta S(\alpha = 2) - \Delta S(\alpha = 1)$ at the same time; dotted is zero contour, dashed is negative, solid is positive. Contour interval 0.3 PSU. The dashed red line indicates the approximate position of the GSR.

This dense anomaly reduces the density gradient across the SPG, and weakens the SPG by triggering the feedbacks described in Levermann and Born (2007): The northward heat transport gets weakened, leading to reduced heat loss over the SPG. At the same time, reduced northward salt transport leads to a freshening of the SPG.

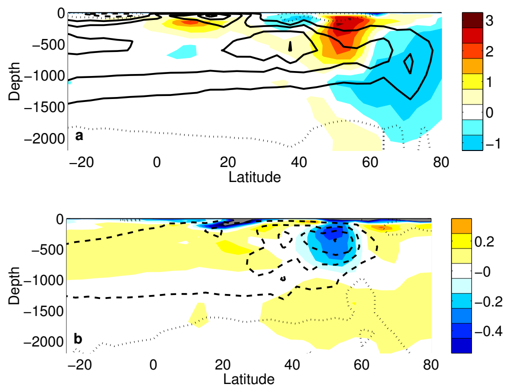

Figure 4 a: Shading: Change in the zonally averaged Atlantic temperature deviation from the equilibrium (ΔT) caused by an increase of α, i. e. $\Delta T(\alpha = 2) - \Delta T(\alpha = 1)$ in K, taken at $t = 250$ yr. Contours: AMOC streamfunction for $\alpha = 2$ at the same time; dotted is zero contour, solid is positive. Contour interval 2 Sv.
b: Shading: Change in the zonally averaged Atlantic density deviation from the equilibrium ($\Delta\rho$) caused by an increase of α, i. e. $\Delta\rho(\alpha = 2) - \Delta\rho(\alpha = 1)$ in kg m^{-3}, taken at $t = 250$ yr. Contours: Difference of the AMOC streamfunction Ψ caused by an increase in α, i. e. $\Psi(\alpha = 2) - \Psi(\alpha = 1)$ at the same time; dotted is zero contour, dashed is negative, solid positive. Contour interval 1 Sv.

Together, these two feedbacks decrease the density of the center of the SPG, further weakening it.
The increasing CO_2 concentrations therefore trigger a separation of the two SPG states described by Levermann and Born (2007) that persists after the stabilization of the CO_2 concentration (fig. 4). As a consequence, dense water formation in the center of the SPG is more reduced in the supercritical case than in the subcritical case (fig. 5).

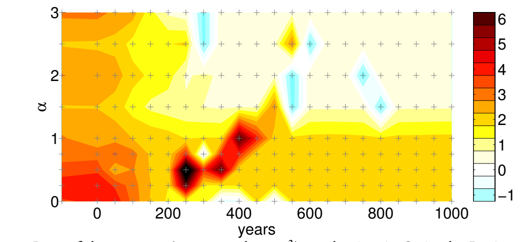

Figure 5: Rate of dense water ($\rho \ge 1028$ kg m^{-3}) production in Sv in the Irminger Sea, as a function of α and time. Grey crosses indicate positions of data points.

2.2 The Role of Sea Ice for the Recovery of the AMOC

During the transient, the ocean-atmosphere heat flux is getting reduced by the warmer atmosphere. This leads to decreased buoyancy loss, and subsequently decreased dense water formation. At the same time, the melting of sea ice leads to an increase in the area of the ocean surface that is exposed to the atmosphere (fig. 6).

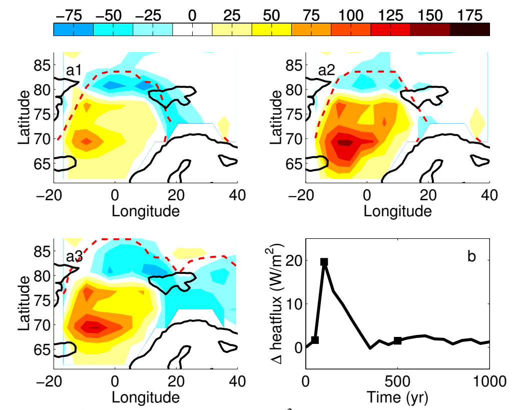

Figure 6 a: Maps of the heat flux anomaly (W m^{-2}) into the ocean for α=1, taken at 50 (a1), 100 (a2), and 500 (a3) years after the start of the CO_2 increase. The red dashed line indicates the largest annual extend of 80% sea ice coverage.
b: Timeseries of heat flux anomaly (W m^{-2}) into the ocean averaged over the area shown in panel a, squares indicate the times of the snapshots shown in panel a.

While the heat flux anomaly is positive in areas that were not subject to insulation from the atmosphere by sea ice in the pre-industrial equilibrium, the anomalies are negative in the area where the sea ice retreats under the influence of the atmospheric warming. Figure 6b shows that the heat flux anomalies integrated over the Nordic Seas disappear as the system approaches the new equilibrium under a higher CO_2 concentration due to this compensating effect. As a result, dense water formation and the AMOC north of the GSR recover.

3 Discussion & Conclusions

The weakening and recovery of the AMOC are only weakly affected by changes of the diffusivity in the pycnocline in low latitudes, that are caused by the mixing parameterization: The warm SAT anomaly due to the increased CO_2 concentration leads to a warming of the surface ocean. As this warm anomaly is penetrating the ocean, it increases the stratification at pycnocline depth. This leads to decreased diffusivities, with the magnitude of this effect depending on α. Subsequently, upwelling in low latitudes is weakened, affecting the strength of the AMOC. However, this effect is of second order: Following simple scaling, the observed decrease in low latitude vertical diffusivity accounts for $\sim$0.4 Sv weakening of the AMOC for $\alpha = 1$. This value increases only slightly to $\sim$0.5 Sv for $\alpha = 2$.
The model produces too high concentrations of sea ice in the Barents Sea under pre-industrial conditions. It is therefore possible that the model overestimate the change in sea ice cover, which would imply that also the recovery of the AMOC gets overestimated.

References

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