Stratification-dependent mixing may increase sensitivity of a wind-driven Atlantic overturning to surface freshwater flux

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[1] Stratification-dependent mixing is employed in a coupled climate model of intermediate complexity with a 3-dimensional ocean component. Oceanic vertical diffusivity is calculated as $\kappa \sim N^{-\alpha}$, where N is the local buoyancy frequency. The sensitivity of the Atlantic meridional overturning circulation (AMOC) to freshwater forcing is tested for exponents $0 \leq \alpha \leq 2$ by first slowly increasing, then decreasing the freshwater forcing over the North Atlantic, keeping the model close to equilibrium. The surface fresh anomaly imposed between 20°N and 50°N in the Atlantic reaches the deep ocean by vertical diffusion, and by AMOC advection via the northern convection sites. The fresh anomaly leads to enhanced stratification and thereby reduces vertical mixing stronger for higher values of α. Consequently, the freshwater anomaly reaches the northern deep water formation regions less diluted, and reduces the AMOC more strongly compared to lower values of α. Our findings indicate that modeled changes in the AMOC depend critically on the details of the mixing parameterization employed in the model.

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1. Introduction

[2] Much of the oceanic heat transport in the Atlantic (estimated to about 1 PW (1 PW = 10^{15} W) at its maximum by *Ganachaud and Wunsch* [2000] and *Talley et al.* [2003]) can be connected to the Atlantic Meridional Overturning Circulation (AMOC). Past climate shifts in the North Atlantic region are often attributed to changes in the AMOC [e.g., *Keigwin and Lehman*, 1994; *Ganopolski and Rahmstorf*, 2001; *Rahmstorf*, 2002; *Clark et al.*, 2002]. To better understand those past climate changes, and a likely future weakening of the AMOC [*Gregory et al.*, 2005], it is necessary to better understand the mechanisms of the AMOC response to changes in the forcing fields [*Zickfeld et al.*, 2007]. As a response to slowly varying freshwater forcing in the North Atlantic, a hysteresis behavior of the AMOC has been found in numerous zonally averaged and three-dimensional models [e.g., *Ganopolski et al.*, 2001; *Schmittner and Weaver*, 2001; *Rahmstorf et al.*, 2005], and has been the focus of many studies because of its significance for estimating the response of the AMOC to varying atmospheric forcing [e.g., *Manabe and Stouffer*, 1995, 1997; *Stouffer et al.*, 2006]. Together with wind, and fluxes of heat and freshwater, vertical mixing is one important factor in shaping the AMOC [*Kuhlbrodt et al.*, 2007], and *Prange et al.* [2003] showed that the AMOC hysteresis behavior may disappear under weak vertical mixing in a three-dimensional model. Investigating further the influence of different parameterizations of vertical mixing, stratification-dependent mixing has been employed in a variety of idealized experimental set-ups, ranging from two-layer models of a hemispheric AMOC [*Nilsson and Walin*, 2001] to global coupled models [*Marzeion et al.*, 2007], to test the response of the AMOC to freshwater forcing. Hemispheric models of the AMOC forced purely by temperature gradients at the surface, and assuming advective-diffusive upwelling through the pycnocline to be the only closing branch of the AMOC produced an enhanced AMOC as a response to decreased meridional surface density gradients if the dependence of mixing on stratification was strong. This was caused by the advection of the light surface anomaly along the southward flowing branch of the AMOC into the ocean interior, leading to reduced stratification, and hence increased mixing and upwelling through the pycnocline [*Nilsson and Walin*, 2001; *Nilsson et al.*, 2003]. *Marzeion et al.* [2007, 2009] used a model with the AMOC mainly driven by southern ocean winds [*Schewe and Levermann*, 2009]. They found the AMOC to be weakened when subjected to freshwater and idealized CO_2 forcing, even if the dependence of mixing on stratification was strong. Moreover, the sensitivity of the AMOC to changes in the forcing was found to be significantly increased under stratification-dependent mixing compared to constant mixing.

[3] Here, we study the influence of stratification-dependent mixing on the AMOC under a gradually changing freshwater forcing, keeping the model close to its equilibrium state at all times. In the light of future climate change this might be a more relevant scenario than rapid changes in surface buoyancy forcing.

2. Model and Experiments

[4] We use the global coupled climate model CLIMBER-3α [*Montoya et al.*, 2005], which combines a coarse-resolution statistical-dynamical atmosphere model [*Petoukhov et al.*, 2000] with a 3-dimensional ocean general circulation model based on the GFDL MOM-3 code and a dynamic and thermodynamic sea-ice model [*Fichefet and Maqueda*, 1997]. The ocean has 24 variably spaced vertical levels, and a horizontal resolution of 3.75° × 3.75°. A second-order moment tracer

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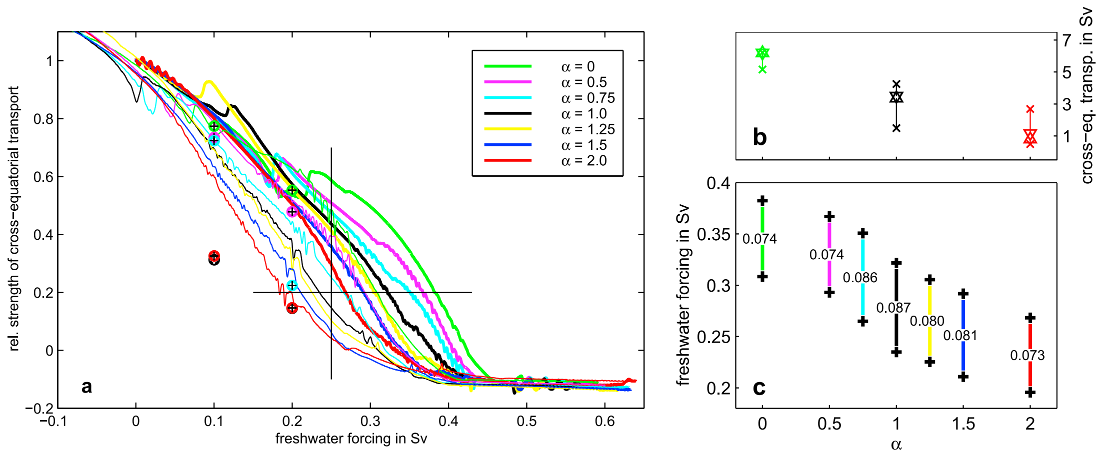

Figure 1. (a) Relative strength of the cross-equatorial volume transport as a function of the freshwater forcing, thick lines: increasing freshwater forcing, thin lines: decreasing freshwater forcing. Circles with black crosses: reference values from the experiments with abruptly increased freshwater forcing from *Marzeion et al.* [2007]. Black lines indicate the freshwater forcing and overturning strength at which the values in Figures 1b and 1c, and Figures 2 and 3 were taken. (b) Cross-equatorial transport as a function of α at 0.25 Sv of freshwater forcing. Crosses indicate the transport during the transient increase and decrease of the freshwater forcing. Downward (upward) pointing triangles indicate the transport after 2000 yrs of constant freshwater forcing, branching off the increasing (decreasing) branch of the freshwater forcing. (c) Strength of the freshwater forcing at which the cross-equatorial transport is 20% of the unperturbed state, as a function of α; upper values for the branch of increasing freshwater forcing, lower values for the branch of decreasing freshwater forcing; the numbers indicate the difference between the two values.

advection scheme [*Prather*, 1986] is employed to reduce numerical diffusion [*Hofmann and Morales Maqueda*, 2006]. In the atmosphere, constant lapse rates of temperature and humidity are assumed, and the horizontal resolution is 7.5° in latitude and 22.5° in longitude. Due to the coarse atmospheric grid, a climatological wind stress based on the NCEP/NCAR reanalysis is applied to the ocean surface. Heat and freshwater fluxes between ocean and atmosphere are computed on the ocean grid, without any flux adjustment.

[5] The vertical diffusivity κ in the ocean interior is parameterized as

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

[*Gargett and Holloway*, 1984; *Gregg*, 1989] where N is the local buoyancy frequency, $N_0 = 7.3 \cdot 10^{-3}\ \mathrm{s}^{-1}$ is a typical value of N around the depth of maximum AMOC in the model, and $\kappa_0 = 0.2 \cdot 10^{-4}\ \mathrm{m}^2\ \mathrm{s}^{-1}$ sets the vertical diffusivity at this depth. This parameterization implies that for $\alpha > 0$, mixing is enhanced for reduced stratification, and the parameter α is controlling the sensitivity of the mixing to changes in stratification. In the experiments presented here, the parameter space between $\alpha = 0$ (constant diffusivity) and $\alpha = 2$ (stronger sensitivity to stratification, constant rate of energy conversion by mixing, see *Nilsson et al.* [2003] and *Marzeion et al.* [2007] for a more detailed discussion of significance of the parameter values) is explored. This range of α includes the value of α_{cr} as given by *Marzeion et al.* [2007, 2009], and implies that the vertical diffusivity for $\alpha = 2$ varies between $0.1 \cdot 10^{-4} \leq \kappa \leq 0.3 \cdot 10^{-4}\ \mathrm{m}^2\ \mathrm{s}^{-1}$ in the upwelling branch of the AMOC (i.e., between 450 and 800 m depth).

[6] Starting from equilibrium runs of ~3000 yr for $\alpha = 0.0$; 0.5; 0.75; 1.0; 1.25; 1.5; and 2.0, anomalous freshwater forcing is applied to the North Atlantic between 20°N and 50°N. The freshwater forcing is increased at a rate of 0.1 Sv/1000 yr, keeping the model close to, but not in, an equilibrium state, until the freshwater forcing reaches a magnitude of ~0.6 Sv. Afterwards, the freshwater forcing is decreased again at a rate of −0.1 Sv/1000 yr. The freshwater forcing is compensated for by a corresponding negative freshwater flux anomaly in the North Pacific between 30°N to 60°N and 150°E to 120°W.

3. Results

[7] Figure 1a shows the cross-equatorial volume transport of the AMOC, normalized by the respective unperturbed transport for each value of α, as a function of the anomalous freshwater forcing applied to the North Atlantic. The maximum value of the unperturbed AMOC stream function varies between 13 Sv for $\alpha = 0$ and 12 Sv for $\alpha = 2$ (not shown). Note that the difference in AMOC strength between the increasing and decreasing branches of the freshwater forcing does not correspond to a hysteresis behavior of the AMOC as found in many other models. Control experiments branching off of the increasing and decreasing branch of the varying freshwater forcing experiments were conducted by keeping the forcing constant at 0.25 Sv of freshwater for $\alpha = 0$, 1, and 2. The results show (see Figure 1b) that the difference between increasing and decreasing branches is merely the result of the slow response of the AMOC to the varying forcing. After 2000 yr of constant freshwater forcing, the difference between the control experiments disappears. Similarly, the intermittent increases of the AMOC stream function during the increasing phase of the freshwater forcing are related to changes in the location of the main convection sites

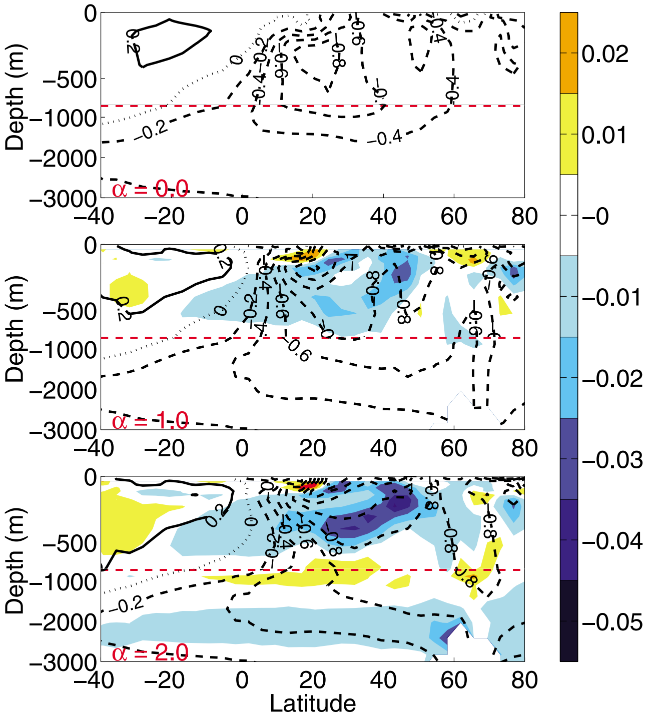

Figure 2. Shading: zonally averaged Atlantic vertical diffusivity anomalies at 0.25 Sv freshwater forcing of the increasing branch in cm^2 s^{-1}. Contours: zonally averaged Atlantic salinity anomalies at the freshwater forcing, solid: positive, dotted: zero contour, dashed: negative. Contour spacing is 0.2 psu. Note that the vertical axis above 700 m depth (red dashed line) is stretched; 700 m is approximately the level of no motion. (top) $\alpha = 0$, (middle) $\alpha = 1$, and (bottom) $\alpha = 2$.

in the model, and thus an artifact of the model not being in true equilibrium during the increase of the freshwater forcing.

[8] The higher the value of α, the weaker is the AMOC, both on the increasing and decreasing branch of the freshwater forcing. This becomes more evident in Figure 1c: While the distance of the increasing and decreasing branches remains very similar for all values of α (indicating that the response time of the AMOC does not vary with α), the freshwater forcing required to weaken the cross-equatorial transport to 20% of the unperturbed value decreases with increasing α. Unlike in the results of *Marzeion et al.* [2007] where the freshwater forcing was increased abruptly and not slowly as here, the sensitivity of the AMOC increases not abruptly, but gradually, with increasing α.

[9] This behavior can be explained as the result of a shift between advective and diffusive pathways of the transport of the freshwater anomaly away from the forcing region, into the lower branch of the Atlantic overturning (Figure 2): The fresh anomaly induced at the surface by the anomalous freshwater forcing reaches the deep interior of the Atlantic Ocean both by advection along with the AMOC, and by vertical downward diffusion. The freshening of the surface water in the forcing region leads to a light anomaly, which increases stratification, and thus reduces the vertical diffusivity at the base of the fresh (and light) anomaly. The higher the value of α, the greater is the sensitivity of the vertical diffusivity to changes in stratification. Figure 2 shows that a negative diffusivity anomaly can be found underneath the region of the freshwater forcing (20°N to 50°N) in approximately 300–500 m depth. This anomaly is stronger for higher values of α.

[10] This negative anomaly of the diffusivity limits the exchange between the implied fresh surface anomaly and the saltier subsurface. Instead, a stronger fresh anomaly is carried northward along with the AMOC. Therefore, with increasing α the fresh anomaly in the high northern latitudes and in the deep return flow of the AMOC get stronger, while a minimum of the fresh anomaly establishes near the level of no motion (contours in Figure 2). The freshening of the high latitude North Atlantic leads to a strong light anomaly (Figure 3a), and a reduced meridional density gradient north

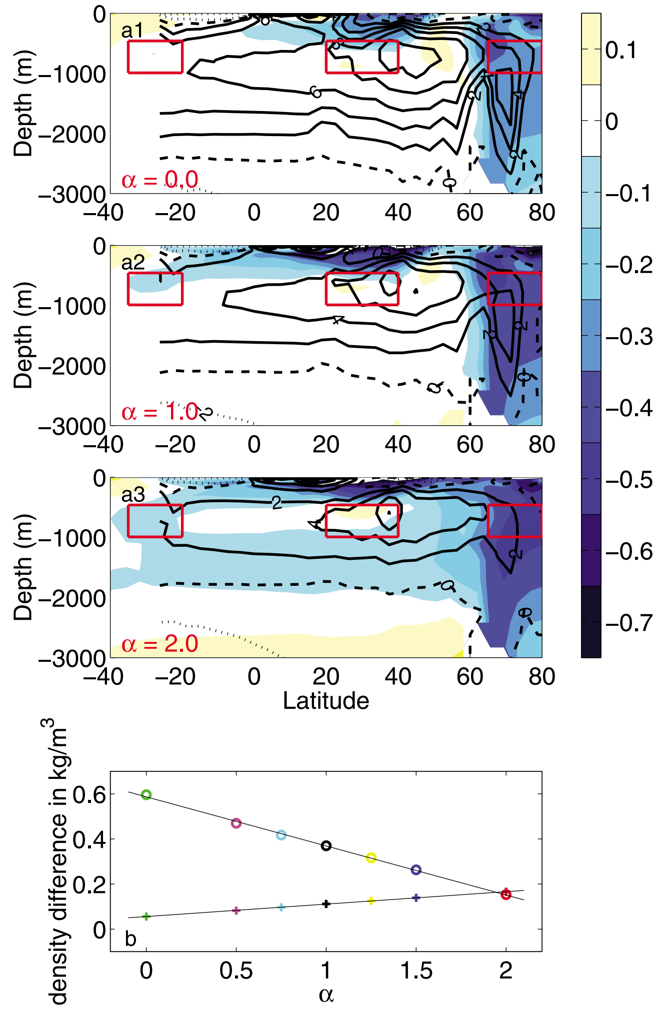

Figure 3. (a) Shading: zonally averaged Atlantic density anomalies at 0.25 Sv freshwater forcing of the increasing branch in kg m^{-3}. Contours: Atlantic overturning stream function at the same freshwater forcing, solid: positive, dashed: zero contour, dotted: negative. Contour spacing is 2 Sv. Red boxes indicate the areas between which the density difference in Figure 3b were calculated. (b) density difference in the Atlantic between the areas 65°N to 80°N and 20°N to 40°N (circles), and between 20°N to 40°N and 20°S to 35°S (crosses), averaged between 460 m and 1000 m depth (red boxes in Figure 1a), as a function of α, at 0.25 Sv freshwater forcing. Colors as in Figure 1a.

of the maximum of the AMOC around the level of no motion (Figure 3b).

[11] Since there is a direct connection between this northern meridional density gradient in the ocean interior and the northern part of the AMOC [*Levermann and Griesel*, 2004; *Schewe and Levermann*, 2009], the smaller density gradient leads to a weakening of the AMOC. At the same time, the meridional density gradient south of the maximum of the AMOC remains unchanged, indicating that changes to diffusive upwelling through the pycnocline play a lesser role. It were changes related to this part of the AMOC that were responsible for the "freshwater-boosted" regime found by *Nilsson and Walin* [2001]. The weakened AMOC then contributes itself to the fresh anomaly in the high latitudes by slowing down the advective removal of the freshwater.

4. Discussion and Conclusions

[12] Our results illustrate the importance of the history of the freshwater forcing for the AMOC even when discussing the equilibrium state of the ocean: Using the same model, *Marzeion et al.* [2007] conducted experiments with increased freshwater forcing in the same region as applied here. However, in their case, the freshwater forcing was not increased gradually, but directly switched from 0 to 0.1 (and 0.2) Sv. For comparison, their results are shown in Figure 1a (circles with black crosses). The strong dependence of the sensitivity of the AMOC subjected to freshwater forcing on the value of α in their case originated from a feedback between a fresh anomaly close to the surface, increased stratification, and reduced downward mixing of freshwater in the regions of dense water formation for "supercritical" values of α. In the experiments presented in this study, the gradual increase of the freshwater forcing ensures that the model operates in the "subcritical" mode at all times: The build-up of a sufficiently strong vertical salinity gradient in the regions of dense water formation, which was responsible for the increased sensitivity of the "supercritical" experiments of *Marzeion et al.* [2007], is inhibited by the continuing downward mixing of freshwater during convection events (compare Figure 3a with *Marzeion et al.* [2007, Figure 5]). The difference in the results thus confirm the mechanism proposed by *Marzeion et al.* [2007].

[13] It is further noteworthy that our results agree with the "freshwater-boosted" overturning described by *Nilsson and Walin* [2001] and *Nilsson et al.* [2003]. An increased diffusivity through the pycnocline is evident in our results (see Figure 2, bottom) as predicted by *Nilsson and Walin* [2001] and *Nilsson et al.* [2003]. However, since the AMOC in CLIMBER-3α is driven mainly by Southern Ocean winds [*Mignot et al.*, 2006; *Schewe and Levermann*, 2009], changes in the low latitude stratification, and hence upwelling, are of minor importance for the rate of the AMOC. The resulting changes in low latitude upwelling are very small compared to the effect of the reduced cross-equatorial, meridional density gradient.

[14] Unlike many other models [e.g., *Ganopolski et al.*, 2001; *Schmittner and Weaver*, 2001; *Rahmstorf et al.*, 2005], the model used here does not exhibit a hysteresis behavior when subjected to slowly varying freshwater forcing. This is likely related to shortcomings in the model's ability to reproduce the observed surface salinity distribution of the North Atlantic, which is discussed in detail by *Montoya et al.* [2005]. The mechanism explained in this study does however not depend on a hysteresis behavior of the model, and not even on the details of the North Atlantic circulation. While the results of *Marzeion et al.* [2009], using the same model set up forced with idealized CO_2 forcing, depend on the North Atlantic sub-polar gyre circulation changes and therefore potentially may be sensitive to changes in the salinity distribution, we do not expect the results presented here to be sensitive to the specific model. Nonetheless, a detailed comparison remains as an important next step for future work.

[15] In summary, we have shown that stratification-dependent mixing may increase the sensitivity of the AMOC to freshwater forcing in the North Atlantic Ocean. The reason is the weakening of the diffusive, and strengthening of the advective pathway in the propagation of the fresh anomaly, which leads to a reduced meridional density gradient in the northern deep return flow of the AMOC. Similarly, the experiments of *Marzeion et al.* [2007, 2009] have shown a greatly enhanced weakening of the AMOC in response to freshwater pulses and idealized CO_2 forcing when stratification dependent mixing was employed. Therefore, the sensitivity of the AMOC to changes in the forcing found in models employing constant vertical mixing may have to be understood as a lower limit when considering the complexity of the processes linking vertical mixing in the ocean interior to the AMOC.

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References

Clark, P. U., N. G. Pisias, T. F. Stocker, and A. J. Weaver (2002), The role of the thermohaline circulation in abrupt climate change, *Nature*, *415*, 863–869.

Fichefet, T., and M. Maqueda (1997), Sensitivity of a global sea ice model on the treatment of ice thermodynamics and dynamics, *J. Geophys. Res.*, *102*, 12,609–12,646.

Ganachaud, A., and C. Wunsch (2000), Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data, *Nature*, *408*, 453–456.

Ganopolski, A., and S. Rahmstorf (2001), Rapid changes of glacial climate simulated in a coupled climate model, *Nature*, *409*, 153–158.

Ganopolski, A., V. Petoukhov, S. Rahmstorf, V. Brovkin, M. Claussen, A. Eliseev, and C. Kubatzki (2001), CLIMBER-2: A climate system model of intermediate complexity. Part II: Model sensitivity, *Clim. Dyn.*, *17*, 735–751.

Gargett, A. E., and G. Holloway (1984), Dissipation and diffusion by internal wave breaking, *J. Mar. Res.*, *42*, 15–27.

Gregg, M. C. (1989), Scaling turbulent dissipation in the thermocline, *J. Geophys. Res.*, *94*, 9686–9698.

Gregory, J. M., et al. (2005), A model intercomparison of changes in the Atlantic thermohaline circulation in response to increasing atmospheric CO_2 concentration, *Geophys. Res. Lett.*, *32*, L12703, doi:10.1029/2005GL023209.

Hofmann, M., and M. A. Morales Maqueda (2006), Performance of a second-order moments advection scheme in an ocean general circulation model, *J. Geophys. Res.*, *111*, C05006, doi:10.1029/2005JC003279.

Keigwin, L. D., and S. J. Lehman (1994), Deep circulation change linked to Heinrich Event 1 and Younger Dryas in a mid-depth North Atlantic core, *Paleoceanography*, *9*, 185–194.

Kuhlbrodt, T., A. Griesel, M. Montoya, A. Levermann, M. Hofmann, and S. Rahmstorf (2007), On the driving processes of the Atlantic meridional overturning circulation, *Rev. Geophys.*, *45*, RG2001, doi:10.1029/2004RG000166.

Levermann, A., and A. Griesel (2004), Solution of a model for the oceanic pycnocline depth: Scaling of overturning strength and meridional pressure difference, *Geophys. Res. Lett.*, *31*, L17302, doi:10.1029/2004GL020678.

Manabe, S., and R. Stouffer (1995), Simulation of abrupt climate change

induced by freshwater input to the North Atlantic Ocean, *Nature*, *378*, 165–167.
Manabe, S., and R. J. Stouffer (1997), Coupled ocean-atmosphere model response to freshwater input: Comparision to Younger Dryas event, *Paleoceanography*, *12*, 321–336.
Marzeion, B., A. Levermann, and J. Mignot (2007), The role of stratification-dependent mixing for the stability of the Atlantic overturning in a global climate model, *J. Phys. Oceanogr.*, *37*, 2672–2681.
Marzeion, B., A. Levermann, and J. Mignot (2009), Sensitivity of North Atlantic subpolar gyre and overturning to stratification-dependent mixing: Response to global warming, *Clim. Dyn.*, doi:10.1007/s0038200805214.
Mignot, J., A. Levermann, and A. Griesel (2006), A decomposition of the Atlantic meridional overturning circulation into physical components using its sensitivity to vertical diffusivity, *J. Phys. Oceanogr.*, *36*, 636–650.
Montoya, M., A. Griesel, A. Levermann, J. Mignot, M. Hoffmann, A. Ganopolski, and S. Rahmstorf (2005), The Earth system model of intermediate complexity CLIMBER-3α. Part I: Description and performance for present day conditions, *Clim. Dyn.*, *25*, 237–263.
Nilsson, J., and G. Walin (2001), Freshwater forcing as a booster of the thermohaline circulation, *Tellus, Ser. A*, *53*, 629–641.
Nilsson, J., G. Broström, and G. Walin (2003), The thermohaline circulation and vertical mixing: Does weaker density stratification give stronger overturning?, *J. Phys. Oceanogr.*, *33*, 2781–2795.
Petoukhov, V., A. Ganopolski, V. Brovkin, M. Claussen, A. Eliseev, C. Kubatzki, and S. Rahmstorf (2000), CLIMBER 2: A climate system model of intermediate complexity. Part I: Model description and performance for present climate, *Clim. Dyn.*, *16*, 1–17.
Prange, M., G. Lohmann, and A. Paul (2003), Influence of vertical mixing on the thermohaline hysteresis: Analysis of an OGCM, *J. Phys. Oceanogr.*, *33*, 1707–1721.
Prather, M. J. (1986), Numerical advection by conservation of second-order moments, *J. Geophys. Res.*, *91*, 6671–6681.
Rahmstorf, S. (2002), Ocean circulation and climate during the past 120,000 years, *Nature*, *419*, 207–214.
Rahmstorf, S., et al. (2005), Thermohaline circulation hysteresis: A model intercomparison, *Geophys. Res. Lett.*, *32*, L23605, doi:10.1029/2005GL023655.
Schewe, J., and A. Levermann (2009), The role of meridional density gradients for a wind-driven overturning circulation, *Clim. Dyn.*, doi:10.1007/s00382-009-0572-1.
Schmittner, A., and A. J. Weaver (2001), Dependence of multiple climate states on ocean mixing parameters, *Geophys. Res. Lett.*, *28*, 1027–1030.
Stouffer, R. J., et al. (2006), Investigating the causes of the response of the thermohaline circulation to past and future climate changes, *J. Clim.*, *19*, 1365–1387.
Talley, L. D., J. L. Reid, and P. E. Robbins (2003), Data-based meridional overturning streamfunctions for the global ocean, *J. Clim.*, *16*, 3213–3226.
Zickfeld, K., A. Levermann, M. G. Morgan, T. Kuhlbrodt, S. Rahmstorf, and D. W. Keith (2007), Present state and future fate of the Atlantic meridional overturning circulation as viewed by experts, *Clim. Change*, *82*, 235–265.

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