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Stability of the Atlantic Meridional Overturning Circulation in the Last Glacial Maximum climate

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[1] The stability of the glacial Atlantic Meridional Overturning Circulation is examined using a coupled model of intermediate complexity. Two slightly different climatic states are generated. One has a southward overturning freshwater transport at the southern border of the Atlantic basin, the other a northward transport. Pulse experiments with varying magnitude always result in a collapsed circulation in case of a southward transport, while the overturning recovers in case of a northward transport. In the latter case recovery is due to a positive salinity-overturning feedback, which strengthens the remnant circulation cell that exists in the 'collapsed' state. This is amplified by advection by wind-driven currents and a southward ITCZ shift. The glacial circulation is more easily perturbed than the modern and restoring timescales are considerably longer, matching the duration of Heinrich events. **Citation:** Weber, S. L., and S. S. Drijfhout (2007), Stability of the Atlantic Meridional Overturning Circulation in the Last Glacial Maximum climate, *Geophys. Res. Lett.*, *34*, L22706, doi:10.1029/2007GL031437.

1. Introduction

[2] Meltwater pulses are thought to have weakened or suppressed the Atlantic Meridional Overturning Circulation (AMOC) at regular time intervals during the last glacial. For example, during Heinrich event H1 the AMOC was nearly eliminated during ca. 2000 yrs, to recover sharply afterwards [*McManus et al.*, 2004]. The associated meltwater pulse is thought to have been relatively modest and of much shorter duration. Apparently the glacial AMOC could be perturbed quite easily for extended time periods.

[3] The stability of the AMOC is a key issue in climate research studies for the glacial climate as well as the present and future climate. Here several related issues play a role. First, there is the issue of mono- or bistability: is the collapsed AMOC an equilibrium solution of the system (bistability) or is the shut-down only temporary (monostability)? Second, there is the question of how far away the system is from a (temporary or permanent) collapse as indicated by the amount of freshwater needed to induce a collapsed state. And third, there is the rate of recovery (if any) once the enhanced freshwater input is ended.

[4] The existence of a stable collapsed AMOC was studied by *de Vries and Weber* [2005] (hereinafter referred to as VW05) for present-day conditions in a series of sensitivity experiments with a coupled model of intermediate complexity. They found that the direction of the freshwater transport by the overturning circulation at the southern boundary of the Atlantic ocean plays a key role. A stable collapsed state exists only in case of a southward transport. This idea, which was first put forward by *Rahmstorf* [1996] in the context of a box model, was subsequently confirmed by *Marsh et al.* [2007] and *Dijkstra* [2007]. Coupled GCMs usually have a northward overturning freshwater transport [*Weber et al.*, 2007]. This might explain the monostable behavior found for coupled GCMs in pulse experiments [*Vellinga et al.*, 2002; *Stouffer et al.*, 2006].

[5] In the present paper we revisit the issue of diagnosing the stability of the AMOC, but now for the glacial climate. We generate two glacial states which have comparable climates, but differ in the direction of the overturning freshwater transport at the southern boundary of the Atlantic basin. It will be shown that these two states differ in their stability characteristics. We examine the differences in the AMOC response to freshwater pulses and the mechanisms that control these different responses.

2. Experimental Design

[6] The model used is the global coupled model ECBilt/CLIO, which was applied earlier by VW05 for Pre-Industrial (PI) conditions. The standard model version included a constant freshwater flux reduction of about 0.2 Sv in the Atlantic area to compensate for excessive precipitation. The maximum in sea surface salinity (SSS) near the South American continent was not very pronounced in the standard run, in contrast to observations. VW05 varied the longitudinal distribution of the southern SSS in four sensitivity experiments by (partly) replacing the basin-scale freshwater flux correction by local corrections over the southern gyre. This modification affects the azonal component M_{az} of the oceanic freshwater transport across the southern boundary, causing it to increase. In turn, the overturning freshwater transport M_{ov} decreases and even changes sign. States with positive (northward) M_{ov} were found to be monostable, those with negative (southward) M_{ov} bistable. Here we use two cases as initial conditions for the spin-up towards a glacial climate. One case has a local freshwater flux reduction of 0.15 Sv over the west of the Atlantic 17–33°S belt, the other a local reduction of 0.2 Sv and an additional west-east dipole correction of ±0.25 Sv (cases A and D in VW05, respectively). The same flux corrections are used in the glacial experiments.

[7] Forcings and boundary conditions for the Last Glacial Maximum (LGM) were imposed following the protocol of the Paleoclimate Modeling Intercomparison Project (PMIP). They are detailed on the PMIP2 Web site (http://www-lsce.cea.fr/pmip2). Each simulation was run for a few millennia, where the LGM runs were initialised from the

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Table 1. The Terms of the Atlantic Freshwater Budget[a]

	E_{net}	M_{az}	M_{ov}	M_{dif}	M_{BS}	Δ
run N, PI	0.32	0.12	0.09	0.04	0.07	0.18
run N, LGM	0.30	0.18	0.09	0.03	0.0	0.10
run S, PI	0.33	0.38	−0.11	0.01	0.05	0.09
run S, LGM	0.30	0.32	−0.03	0.01	0.0	0.05

[a]Terms in Sv. Net evaporation E_{net}, the azonal, overturning and diffusive transports M_{az}, M_{ov} and M_{dif} at the southern boundary, and transport through Bering Strait M_{BS} in the PI and LGM experiments with northward (N) and southward (S) overturning freshwater transport. All terms are averages over the last 100 yr of the experiments. The last column gives the freshwater forcing Δ (in Sv) needed to cause a collapse in the hysteresis experiment (see section 3.3).

corresponding PI control states. There are only minor differences in the LGM climate in the two experiments. In both runs, the AMOC strength increases during the LGM. This is probably due to the very low density of Antarctic Bottom Water in the control climate of ECBilt/CLIO, which favors an intensified glacial AMOC [*Weber et al.*, 2007]. The location of North Atlantic Deep Water (NADW) formation and the sea-ice margin clearly shift southward during the LGM.

[8] The changes in the freshwater budget are given in Table 1. There is a small reduction in net evaporation E_{net} and a cessation of the inflow of freshwater through Bering Strait due to the lowering of sea level during the LGM. (Here E_{net} is defined as evaporation minus precipitation and run-off, plus brine rejection, minus sea-ice melt.) Changes in the oceanic transport terms are somewhat larger, with M_{ov} remaining northward during the LGM in one case and southward in the other. The two experiments are denoted as run N (northward transport) and run S (southward transport).

[9] In the present experiments M_{ov} increases in the LGM climate or remains comparable to the modern value. Similar results are found in a range of climate models that exhibit varying AMOC responses to LGM boundary conditions [*Weber et al.*, 2007]. This intercomparison study showed that the M_{ov} response is primarily determined by salinity changes during the LGM, irrespective of the response in overturning strength. All models show a more saline flow at depth during the LGM, while most models also show fresher thermocline and intermediate waters in the southern Atlantic. The latter response might mimic a weakened Agulhas leakage under glacial climate, as suggested by proxy data [*Peeters et al.*, 2004].

3. Results

[10] Here we examine the different aspects of 'stability' by a number of sensitivity experiments, which all branch off from the two equilibrated LGM states. First we apply freshwater pulses over the Atlantic 50–70°N belt during 100 yr, with constant amplitudes of 0.1, 0.2 and 0.4 Sv. Then we compute part of the hysteresis diagram by applying a slowly increasing freshwater flux to the same latitude belt.

3.1. Response to Imposed Meltwater Pulses

[11] The AMOC collapses within 100 yr in response to the imposed freshwater pulses in all six experiments. Figure 1 gives the results for a 0.4 Sv pulse. Case N shows a gradual restoration after another 800 yr, while case S remains collapsed. Results are similar for the smaller-amplitude pulses, but the timescale of recovery in case N depends on the amplitude of the pulse (it is 500 yr for the 0.2 Sv pulse, while recovery starts within a few years for the 0.1 Sv pulse). Case S does not recover when these experiments are continued for another millennium. We thus conclude that the sign of the overturning freshwater transport determines the AMOC response to a freshwater pulse in the LGM climate too, similar to our earlier results for the PI climate. For $M_{ov} > 0$ the AMOC recovers and for $M_{ov} < 0$ it remains collapsed, although the differences in climate for both cases seem marginal at first sight. In order to understand these different responses, we have to consider the large-scale salinity advection feedback associated with the direction of the overturning freshwater transport.

[12] In case N the 'on' state of the AMOC transports freshwater into the Atlantic basin. The initial response to a shut-down is a salinification of the Atlantic, starting near the southern boundary and gradually spreading northward (Figure 1, bottom two panels). Eventually, this positive anomaly will counteract the negative anomaly that resulted from the imposed freshwater pulse and which led to the shut-down of NADW formation. Because of the gradual salinification in the northern Atlantic, convection will

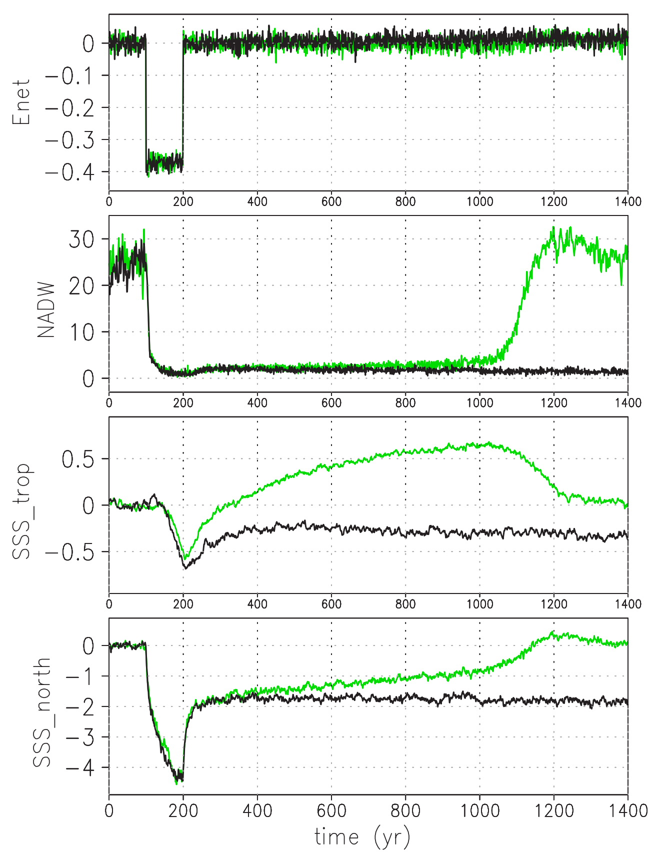

Figure 1. Response to a freshwater pulse of 0.4 Sv (during yrs 101–200) in net evaporation over the Atlantic basin, the AMOC maximum (north of 45°N) and SSS in the tropical (20°S–20°N) and northern (north of 45°N) Atlantic ocean in runs N (green) and S (black). E_{net} (in Sv) and SSS (in PSU) are given as anomalies wrt to the average over the first 100 yr.

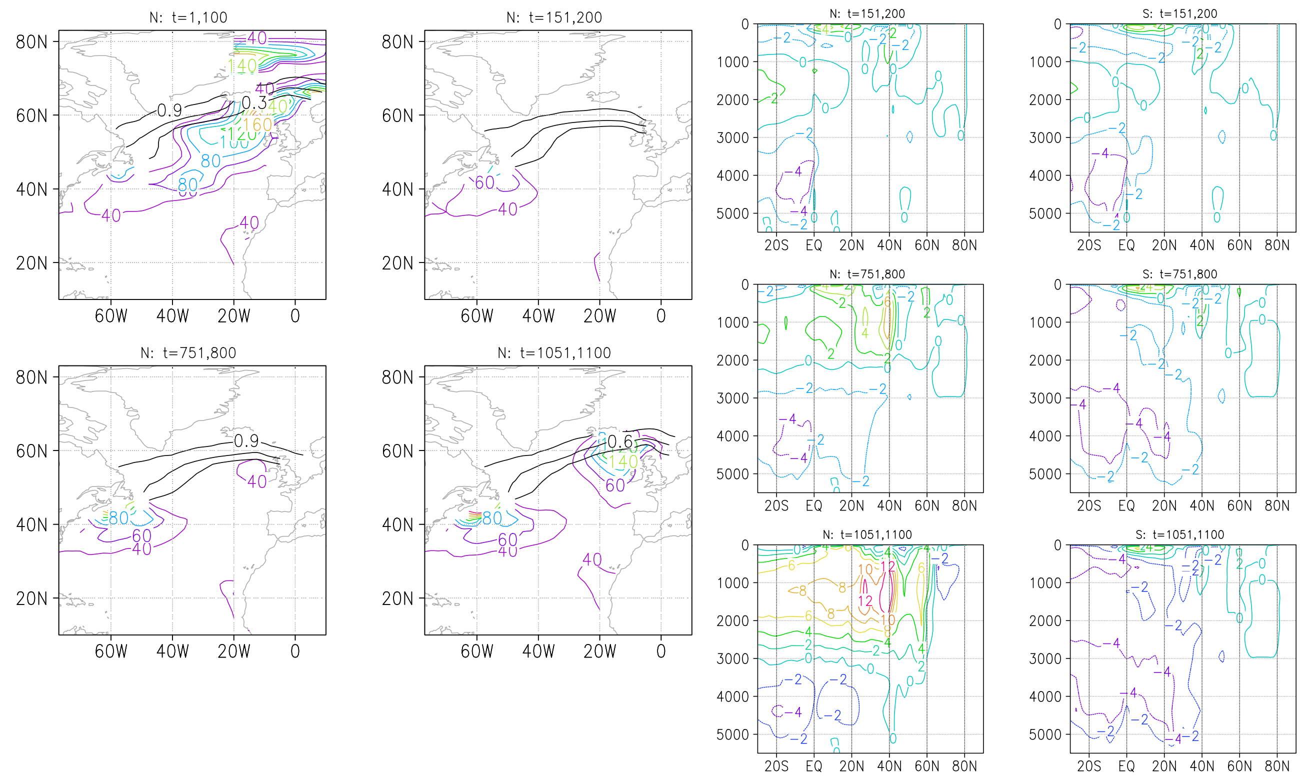

Figure 2. (left) The annual-mean convective layer depth *L* (in m; colored) and sea-ice fraction (0.3, 0.6, and 0.9 contours; in black) for experiment N for the control state and different periods during the 0.4 Sv pulse experiment. Here *L* is defined as the depth to which convective adjustment has been applied to fully mix the water column. (right) The Atlantic meridional overturning streamfunction (in Sv) for experiments N (left) and S (right). The averaging periods given at the top of each panel refer to the end of the pulse (yrs 151–200), the collapsed state (yrs 751–800) and the recovery phase or final collapsed state (yrs 1051–1100), see Figure 1.

resume. In case S the AMOC transports freshwater out of the Atlantic. This implies that the initial response to a shutdown is a freshening of the basin. This will maintain the northern negative salinity anomaly.

3.2. Mechanisms for Recovery

[13] During the LGM open-ocean convection occurs primarily south of Iceland, with a weak secondary site to the southwest of New Foundland (Figure 2). In addition, there is convection below sea ice in the GIN sea. Patterns are very similar in cases N and S. The imposed freshwater pulse causes a cessation of deep convection in the GIN Sea and south of Iceland, but the southern open-ocean site near New Foundland remains active. After the termination of the pulse surface salinities increase and deep-water formation intensifies over the southern site in case N, until the SSS anomaly becomes positive here. At this point, deep convection south of Iceland intensifies (Figure 2) and the system recovers fully. Several processes contribute to this.

[14] First, there is northward advection by a weak, shallow overturning cell (‘mini-AMOC’), which extends to depths of ca. 2 km and latitudes ca. 45°N in the Atlantic basin. The streamfunction is very similar in cases N and S directly after the collapse (Figure 2). The positive salinity anomaly in the southern Atlantic in case N, which extends from the surface to depths of ca. 1 km, is transported northward by the mini-AMOC. This constitutes a positive feedback on the strength of this shallow cell, which intensifies on a centennial timescale. In case S this feedback is absent and the mini-AMOC weakens with time.

[15] Second, the wind-driven currents transport surface waters northward along the American continent north of the equator. The model’s Gulf Stream separates at ca. 40°N, further advecting surface waters across the Atlantic basin toward the convection site south of Iceland. In case N these currents contribute toward the northward salinity advection, whereas in case S they mainly transport fresh water. The spatial pattern of SSS anomalies in the northern Atlantic clearly reflects the wind-driven transport (see Figure S1 of the auxiliary material).[1]

[16] Third, there is a southward shift of the Intertropical Convergence Zone (ITCZ) associated with the AMOC collapse. This causes a reduction in precipitation in the northern tropical Atlantic (AM, Figure S2) and a positive salinity anomaly (AM, Figure S1) which is also advected northward by wind-driven currents. This atmospheric feedback, which was first identified by *Vellinga et al.* [2002], was recently analysed in detail by *Krebs and Timmermann* [2007] using ECBilt/CLIO. They show that it accelerates the recovery of the AMOC considerably, implying a faster recovery timescale in models which contain this feedback as compared to ocean-only models or simplified coupled models which do not capture tropical air-sea interactions.

[1]Auxiliary materials are available in the HTML. doi:10.1029/2007GL031437.

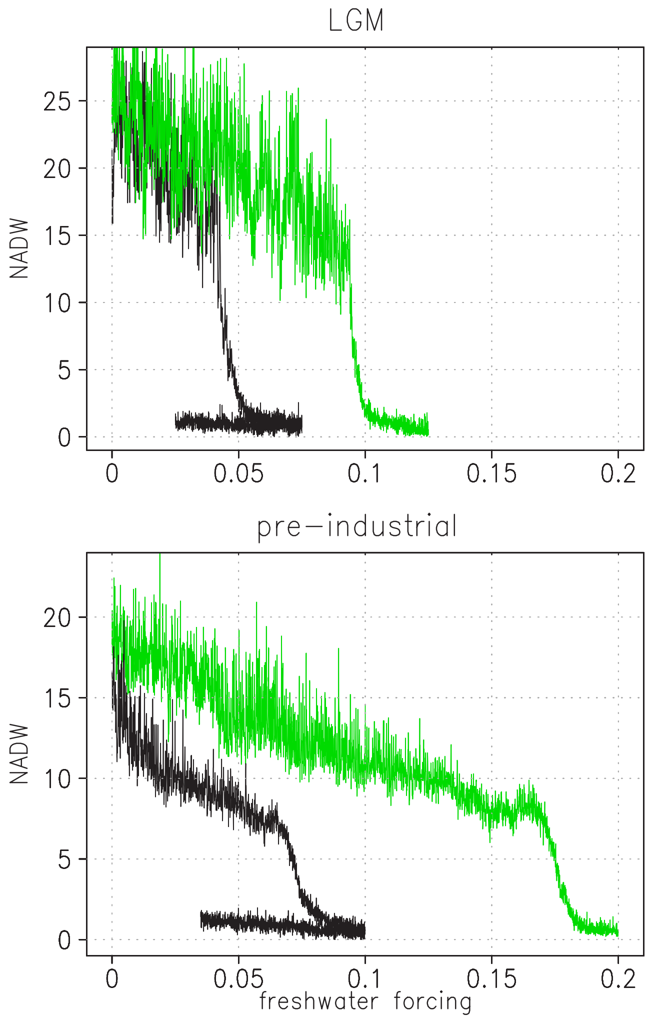

Figure 3. Partial hysteresis curves for case N (green) and case S (black) for (top) the LGM climate and (bottom) the PI climate. Shown is the AMOC maximum (north of 45°N) versus the freshwater forcing over the Atlantic 50–70°N latitude belt (both in Sv).

[17] The latter two processes operate in case N as well as case S, with similar strength. However, the salinity-overturning feedback is absent in case S as is evident from the pattern of SSS anomalies in the southern Atlantic. This prevents the initial spin-up of the weak overturning cell and the subsequent recovery of the AMOC.

3.3. Distance to the Point of Collapse

[18] Here we determine the amount of freshwater needed to cause a collapse by computing part of the hysteresis diagram. A rate of change of the freshwater forcing of 0.05 Sv/1000 yr is used [*Rahmstorf*, 1996]. This results in a basin-mean imbalance (taking 100-yr averages) in the Atlantic freshwater budget which is smaller than 0.02 Sv. A-priori we expect case S to be closer to the point of collapse than case N, based on their M_{ov} values. These place case N on the monostable branch further away from collapse than case S which is on the bistable branch (see Table 1). Computation of the hysteresis curves (Figure 3, top) confirms this conjecture. The amount of freshwater needed to attain a collapsed state (maximum overturning strength below 2 Sv) is 0.1 Sv in case N and 0.05 Sv in case S.

[19] When the computation of the hysteresis diagram is continued further on the 'off' branch, the basin-mean imbalance becomes larger (0.05–0.1 Sv). Apparently, lower rates of change and longer computation times are needed here in order to achieve a near-equilibrium state. For this reason we compute only part of the hysteresis diagram as shown in Figure 3, and examine the mono/bistability by the shorter pulse experiments described above.

4. Discussion: LGM Versus PI Results

[20] There are several striking differences between the present results for the glacial climate and those found earlier for the modern climate. Restoring timescales are considerably longer during the LGM, while AMOC recovery always sets in immediately after the termination of a pulse during the PI. Also, less freshwater is needed to cause a collapse during the LGM as summarised in Table 1.

[21] Longer recovery timescales for the LGM climate as compared to the PI climate were also found by *Bitz et al.* [2007], using the CCSM3 coupled model. They applied freshwater pulses which are equivalent to adding 0.16 Sv during 100 yr to LGM and PI states which have a northward overturning freshwater transport. Bitz et al. argue that the expansion of sea ice over convection sites in the northern Atlantic inhibits a fast AMOC recovery during the LGM. The sea-ice coverage reduces the atmospheric cooling which occurs in response to an AMOC shut-down, that makes subpolar waters colder and denser. A similar southward expansion of sea-ice over convection sites is found in ECBilt/CLIO after a collapse during the LGM (Figure 2), as well as a smaller cooling of glacial subpolar waters that are already very cold. The timescale of recovery is thus set by basin-scale salinity advection during the LGM, while it is set by surface heat loss during the PI.

[22] We note here that recovery after a freshwater pulse does not imply that a climatic state is monostable. It can be that longer and stronger pulses are needed to cross the boundary between the 'on' and 'off' modes [*Dijkstra*, 2007; VW05]. This seems especially true for the modern AMOC. The glacial AMOC is more easily perturbed and small amounts of freshwater are already sufficient to obtain the 'off' mode. This is also evident by comparing the hysteresis curves for the glacial and modern climate (Figures 3, top, and 3, bottom). The PI states are further away from the point of collapse than their LGM counterparts (consistent with results by *Ganopolski and Rahmstorf* [2001] and *Prange et al.* [2002], who found larger hysteresis widths during the PI than during the LGM). We hypothesize that this effect is due to the considerable cooling of subpolar waters, which sustains the PI overturning circulation under relatively large freshwater forcings. The northern Atlantic cools by 1 and 2°C along the 'on' branch of the PI hysteresis curve (surface temperatures north of 45°N; cases S and N), while it cools by only 0.3 and 0.5°C for the LGM background climate.

5. Summary and Conclusions

[23] We find that the direction of the overturning freshwater transport at the southern boundary of the Atlantic

basin acts as a control parameter on the stability of the glacial AMOC. The collapsed state is unstable for a northward transport, but stable for a southward transport. In the former case, recovery of the AMOC is associated with a basin-scale salinification which induces a strengthening of the remnant overturning cell (mini-AMOC). The effectiveness of recovery is related to northward advection of positive salinity anomalies, which are due to an AMOC collapse, both by the mini-AMOC and wind-driven currents. Positive salinity anomalies in the northern tropical Atlantic, that arise due to a southward shift of the ITCZ, also contribute to a recovery.

[24] Computation of the hysteresis diagrams shows that the overturning freshwater transport also gives some indication of the relative amounts of freshwater needed to obtain the (temporarily or permanently) collapsed state: larger amounts of freshwater are needed in the experiment where M_{ov} has the larger value. In the LGM climate the overturning freshwater transport remains comparable to the modern value or it increases, see Table 1. Naively, one might therefore expect that the LGM states are as far away from the point of collapse as the PI states or that they are even further away from it. However, this is not the case and the LGM states are more easily perturbed by meltwater pulses than the PI states. The value of M_{ov} clearly does not predict the relative stability of states with different background climates, although it seems to be a good diagnostic for different states with the same background climate.

[25] Recovery timescales after a meltwater pulse induced collapse are much longer for the glacial AMOC than for the modern one. This seems to be due to the larger sea-ice coverage over deep-water formation sites and the colder background climate of the LGM, which disable the thermal feedback associated with air-sea exchange that operates in the present climate. In the LGM climate recovery is completely determined by salinity advection. Recovery only sets in well after termination of an imposed freshwater pulse. Such a temporal pattern much better matches reconstructed Heinrich events than the rapid resumption usually simulated for the modern climate.

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