Atlantic meridional overturning and climate response to Arctic Ocean freshening

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[1] The global deglaciation event that followed Last Glacial Maximum was punctuated by a sequence of rapid climate changes. This began with the Laurentide Ice Sheet derived calving event called Heinrich Event 1, continued with the Bølling-Allerød warming that occurred synchronously with meltwater pulse 1a, and was in turn followed by the Younger-Dryas abrupt return to cold conditions. Although it has long been believed that these events are consequences of the response of Atlantic meridional overturning circulation to freshwater inputs, it has always been assumed that these inputs were applied directly to the Atlantic Ocean itself. We address the question of how the Atlantic meridional overturning circulation would have responded to the recently hypothesized addition of freshwater into the Arctic Ocean at the time of onset of the Younger-Dryas, and thereby demonstrate that this response is essentially identical to the response to North Atlantic freshening. **Citation:** Peltier, W. R., G. Vettoretti, and M. Stastna (2006), Atlantic meridional overturning and climate response to Arctic Ocean freshening, *Geophys. Res. Lett.*, *33*, L06713, doi:10.1029/2005GL025251.

1. Introduction

[2] Interest in the possibility of rapid climate change occurring as a consequence of the ongoing warming of the atmosphere due to increasing greenhouse gas concentrations [*Stocker and Schmittner*, 1997] derives from the recognition that extremely rapid changes of climate have occurred in the past [*Dansgaard et al.*, 1993]. These previous examples of abrupt climate change have occurred under the colder than modern conditions characteristic of a glacial climate regime rather than the warmer than modern conditions that will apparently characterize our future. Evidence for the occurrence of millennial timescale variability of climate under glacial conditions is abundant and includes oxygen isotopic records from the Greenland Summit ice cores [*Johnsen et al.*, 1992] as well as similar records from deep sea sedimentary cores [*de Abreu et al.*, 2003] and records of relative sea level history based on coral and mangrove sequences from the tropics [*Fairbanks*, 1989; *Hannebuth et al.*, 2000]. However, the mechanisms responsible for such rapid shifts in climate regime remain inadequately studied, and until these mechanisms can be unambiguously identified we will be in no position to understand the probability that they might operate under future conditions.

[3] Because the time interval between Last Glacial Maximum at 21,000 years before present (BP) and the onset of the modern Holocene interval at 10,000 years BP includes such a complex sequence of rapid climate changes, it is an interval that is (perhaps uniquely) worthy of study. This is especially the case as several of the events in this interval have been associated with radically different mechanisms. For example the onset of the rapid Bølling-Allerød (B-A) warming is suggested to have been caused by a recovery of the strength of the Atlantic Meridional Overturning Circulation (MOC), due either to freshwater forcing (FWF) of the southern ocean caused by a sudden melt-back of Antarctic ice [*Clark et al.*, 2002], or to an automatic recovery of the strength of the Atlantic overturning circulation following a shutdown caused by the meltwater delivered to the surface of the Atlantic by Heinrich Event 1 [*Peltier*, 2005]. Similarly, the onset of the Younger Dryas (Y-D) cold reversal is suggested to have been a response to freshening of either the North Atlantic caused by a switch in the drainage of glacial Lake Agassiz, from the south to the east [*Broecker et al.*, 1989] or by a switch from the south to the north [*Tarasov and Peltier*, 2005]. No adequate discussion of the mechanism responsible for the rapid climate change that marked the cessation of the Y-D has yet been proposed.

[4] That each of the events in this sequence is intimately connected to variations in the strength of the Atlantic MOC seems clearly established on the basis of a recently published time series for the Pa/Th kinematic tracer of the strength of the Atlantic MOC [*McManus et al.*, 2004]. In order to investigate the relative merits of the competing hypotheses as to the mechanism responsible for the rapid climate changes that occurred subsequent to LGM, we have employed a modern coupled atmosphere-ocean sea ice-land surface processes model (the NCAR CSM1.4 Model [see *Boville and Gent*, 1998; *Otto-Bliesner and Brady*, 2001]) to simulate the response of the climate system to a range of freshwater forcing strengths applied to the ocean surface in two regions of interest, respectively, the North Atlantic in the region between 50°N and 70°N latitude and the Arctic in the region occupied by the Beaufort Gyre which is located adjacent to the modern outlet of the McKenzie River in northern Canada (Figure 1). Our choice of the former region is connected with it having been chosen as the region of focus for an ongoing intercomparison project within which a suite of modern models have been inter-compared in terms of their response to an applied surface freshwater anomaly [*Stouffer et al.*, 2006]. This choice was motivated in part by the fact that it is this region of the Atlantic Ocean where the surface was consistently freshened during each of the Heinrich Events that occurred during Marine Oxygen

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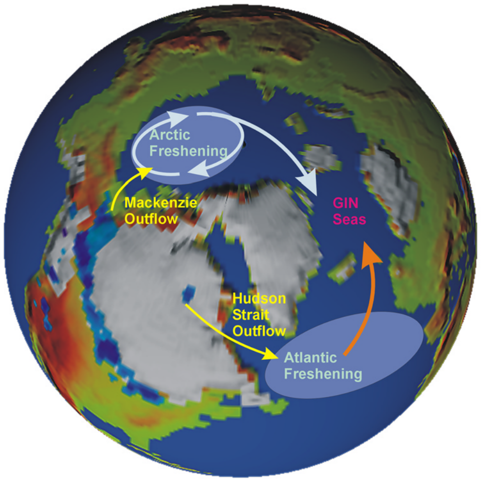

Figure 1. Geographical locations in which intense freshwater forcing was applied to the surface of the ocean and the distribution of land ice and pro-glacial lakes at the beginning of the Y-D. It will be noted that the Barents Sea ice sheet was absent by 12,800 years before present as it disintegrated during the MWP-1a event. Of further note is the fact that Bering Strait was dry land at this time implying that the only route whereby water added to the Arctic Basin could equilibrate with the global ocean was by flow through Fram Strait and thus into the Greenland-Iceland-Norwegian Seas where deepwater forms today.

Isotope Stage 3 (MOIS 3) as well as during Heinrich Event 1 early in the deglaciation process [*Hemming*, 2004]. The choice of the Beaufort Sea as a second region of FWF is motivated by the recently published hypothesis [*Tarasov and Peltier*, 2005] that it was this region that was recipient of intense FWF at the time of Y-D onset. There is also some indication that data extracted from deep sea sedimentary cores retrieved from the Eastern Arctic Basin reveal an equally plausible alternate source for Arctic FWF, one that could derive from proglacial lakes surrounding the Fennoscandian ice sheet, and which subsequently drained into the Laptev Sea [*Spielhagen et al.*, 2005]. Whether the freshwater source was from the Eastern or Western Arctic Basin, the resulting transport to the GIN seas through Fram Strait would be identical [*Aagaard and Carmack*, 1994].

2. Experimental Design

[5] The quality of CSM1.4 model predictions of the response to applied FWF has been established through participation in the ongoing Coupled Model Intercomparison Project. The results for CSM1.4 are fully documented in the current literature [*Stouffer et al.*, 2006]. The design of the new analyses we report here involves the integration of paired numerical experiments, the first member of the pair in each case being a modern control simulation. The second experiment in each set consists of an integration in which FWF is applied to the surface of the ocean at a fixed rate, either 1 Sv (1 Sv = 10^6 m ^{3} s^{-1}) or 0.3 Sv for a fixed period of 100 years, after which the

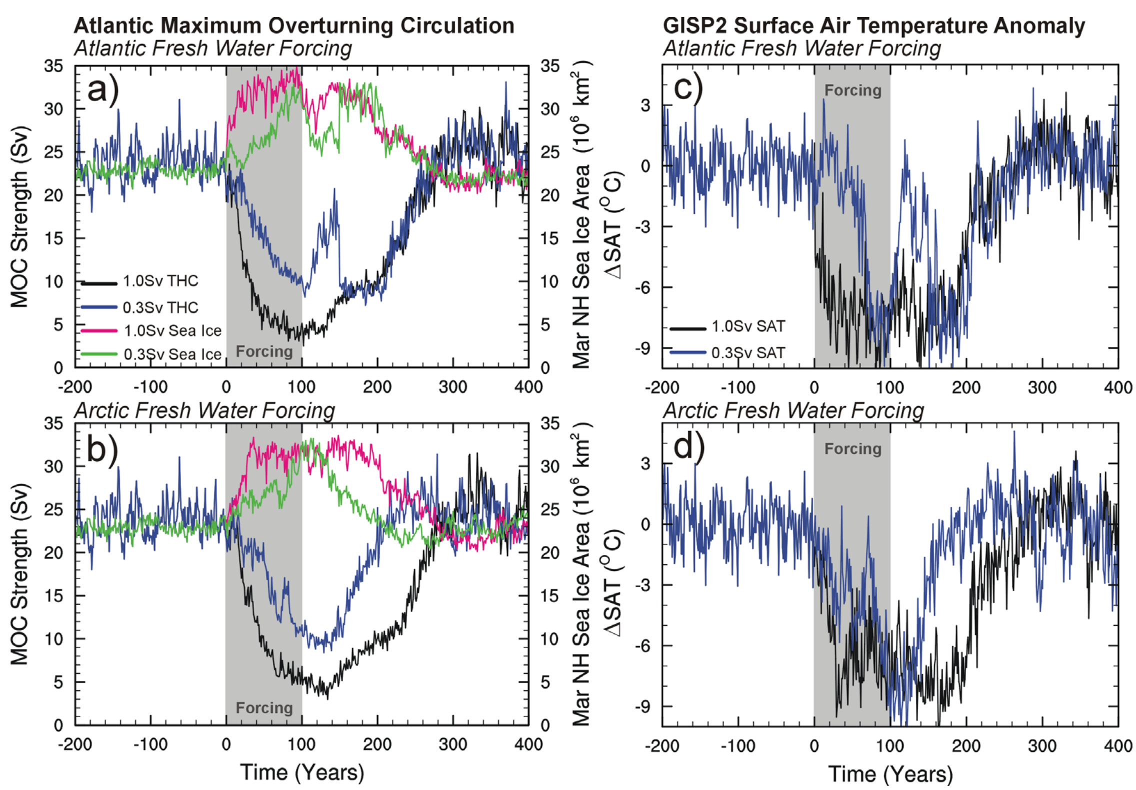

Figure 2. Climate system responses to ocean surface freshening. (a) and (b) show the evolving strength of the Atlantic MOC and northern hemisphere March sea ice area to the application of either 0.3 Sv or 1 Sv of freshening to either the surface of the Atlantic ocean (a) or the Arctic Ocean (b). (c) and (d) present the response to the same forcing scenarios in terms of the evolution of surface temperature annually averaged over the site of the Greenland Summit ice cores.

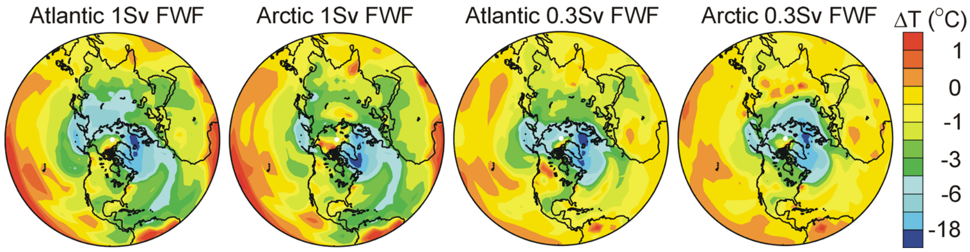

Figure 3. Northern hemisphere surface air temperature anomalies induced by freshwater forcing of either the North Atlantic or the Western Arctic Ocean. The anomalies are computed for the month of December (Experiment-Control). The results are shown for both the 0.3 Sv and the 1 Sv experiments. The anomalies are for the December average over the final 10 years of the interval of time during which freshwater forcing is applied.

forcing is set to zero and the integration continued. A 1 Sv freshwater pulse of 100 years duration is the equivalent of 8.7 meters of sea level rise. For comparison, the Greenland Ice Sheet is equivalent to 7 meters of sea level rise and MWP-1a resulted in 20 meters of sea level rise over 500 years. Since our primary focus in this work has been to understand the plausibility of the Arctic FWF hypothesis for Y-D onset, we expect that our use of a modern climate based control model will not be a significant cause for concern since this event occurred following the onset of B-A warmth at a time when the strength of the Atlantic MOC had already recovered to a strength near modern [*McManus et al.*, 2004]. It will nevertheless be important to test this assumption through further work. Of note is the fact that the Bering Strait is closed in this version of the CSM, as was the case during the Y-D.

3. Results

[6] Figures 2a and 2b present the primary results obtained concerning the evolution of the strength of the Atlantic MOC as a function of both the strength of the applied FWF (1 Sv or 0.3 Sv) and the region of application (Atlantic or Arctic). In each case the 100 year interval of forcing begins at the zero of time. Notable is the fact that, for forcing amplitude of 1 Sv, the response of the Atlantic MOC is essentially identical whether the forcing is applied to the Arctic Ocean over the Beaufort Gyre or over the North Atlantic Ocean itself. In both cases the Thermohaline Circulation (THC) collapses to almost zero strength and then recovers on a timescale of approximately 200 years following removal of the forcing. Although this timescale to recovery is too short to enable the model calculation to explain the 1000 year duration of the Y-D, it may be appropriately stretched simply by increasing the duration of the freshening event to that suggested by the recently published analysis of continental runoff [*Tarasov and Peltier*, 2005].

[7] Detailed analyses (not shown) demonstrate that there exists a slight phase lag (of approximately a decade) of the response to Arctic forcing compared to the response to Atlantic forcing. Notable also is the fact that there exists a strong oscillation of the strength of the MOC following its recovery to full strength. If the recovery phase of these simulations were taken to represent the end of the Y-D then this oscillation would provide a natural explanation for the previously identified pre-Boreal oscillation [*Fisher et al.*, 2002]. Inspection of the results for the cases in which the forcing is reduced to 0.3 Sv, close to the strength of the meltwater pulse recently inferred to have been responsible for Y-D onset [*Tarasov and Peltier*, 2005], reveals more significant differences in the response due to the application of FWF in the two different geographical regions. When this forcing is applied to the Atlantic, the MOC begins an aborted recovery immediately after the forcing is eliminated, but then finally recovers on a timescale identical to the 1 Sv case. For Arctic forcing, on the other hand, recovery begins and proceeds to completion immediately following the cessation of the FWF. In these more weakly forced cases, it is especially important to note that the 70% reduction in the strength of the forcing does not lead to a 70% reduction in the amplitude of the response but rather to a response that is only moderately reduced from the 1 Sv results. The adjustment to the applied forcing is therefore highly nonlinear. Also shown on Figures 2a and 2b are time series of the response of the area covered by sea ice to the reduction in intensity of the overturning circulations. The significant expansion in sea ice area constitutes a strong positive feedback on the surface cooling induced by the decrease in MOC strength.

[8] Figures 2c and 2d display the evolution of the surface temperature at the position of the Greenland Summit ice cores. Notable here is the fact that maximum depression of surface temperature at this important location is approximately 9°C, a result that is entirely consistent with recent inferences (10°C ± 4°C) based upon the analysis of gas isotope data from the GISP2 ice core across the Y-D event [*Gatchev and Severinghaus*, 2005]. These results therefore demonstrate that the plausibility of the Arctic forcing hypothesis [*Tarasov and Peltier*, 2005] for the Y-D cold reversal is strongly reinforced by the results of coupled climate simulation.

[9] The spatial distribution of the reduction in December surface air temperature caused by the freshwater forced decrease in the strength of the THC is displayed in Figure 3. Inter-comparison of the frames for Atlantic and Arctic "hosing" at a strength of either 1 Sv or 0.3 Sv will demonstrate that the spatial extents of the response for these two intensities of FWF are also extremely similar, with prominent cooling signals covering almost the entire northern hemisphere. Also evident, however, is a transition to warming in the equatorial region and some restriction of the region of strongest surface cooling to the region immediately surrounding the GIN seas when the amplitude of the applied freshwater anomaly is reduced from 1 Sv to 0.3 Sv. In all cases the centre of strongest cooling is immediately

over the GIN seas and adjacent land masses (Greenland and Fennoscandia).

4. Discussion

[10] Based upon the totality of the results reported herein we are in a position to interpret the complete sequence of rapid climate changes that occurred subsequent to Last Glacial Maximum. The first event in this sequence was Heinrich event 1, the net effect of which was to effectively arrest the Atlantic MOC. Our results suggest that recovery would have occurred spontaneously when the iceberg calving event ended. MWP-1a was triggered by the rapid onset of B-A warmth due to the recovery of the MOC to near modern strength. This event did not significantly perturb the THC because the freshwater influx into the ocean occurred hyperpycnally as suggested by *Tarasov and Peltier* [2005] and recently confirmed by *Aharon* [2006] through the analysis of samples from the floor of the Gulf of Mexico. Following a period of slow cooling that occurred following the influx of MWP-1a, the sudden cold reversal of the Y-D was apparently forced by a flood of freshwater into the Arctic Ocean, the evidence for which is especially clear in the Ba/Ca analysis of a western Arctic core by *Hall and Chan* [2004]; see also *Hillaire-Marcel et al.* [2004], *Andrews and Dunhill* [2004], *Poore et al.* [1999], and *Spielhagen et al.* [2005]. The final event in the sequence, the recovery of the THC following its reduction during the Y-D, occurred spontaneously once this Arctic derived freshwater flood ended.

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