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Simulation of the cold climate event 8200 years ago by meltwater outburst from Lake Agassiz

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[1] The cold climate anomaly about 8200 years ago is investigated with CLIMBER-2, a coupled atmosphere-ocean-biosphere model of intermediate complexity. This climate model simulates a cooling of about 3.6 K over the North Atlantic induced by a meltwater pulse from Lake Agassiz routed through the Hudson strait. The meltwater pulse is assumed to have a volume of 1.6×10^{14} m^3 and a period of discharge of 2 years on the basis of glaciological modeling of the decay of the Laurentide Ice Sheet (LIS). We present a possible mechanism which can explain the centennial duration of the 8.2 ka cold event. The mechanism is related to the existence of an additional equilibrium climate state with reduced North Atlantic Deep Water (NADW) formation and a southward shift of the NADW formation area. Hints at the additional climate state were obtained from the largely varying duration of the pulse-induced cold episode in response to overlaid random freshwater fluctuations in Monte Carlo simulations. The model equilibrium state was attained by releasing a weak multicentury freshwater flux through the St. Lawrence pathway completed by the meltwater pulse. The existence of such a climate mode appears essential for reproducing climate anomalies in close agreement with paleoclimatic reconstructions of the 8.2 ka event. The results furthermore suggest that the temporal evolution of the cold event was partly a matter of chance. *INDEX TERMS:* 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 4255 Oceanography: General: Numerical modeling; 4532 Oceanography: Physical: General circulation; 4215 Oceanography: General: Climate and interannual variability (3309); 9325 Information Related to Geographic Region: Atlantic Ocean; *KEYWORDS:* coupled climate model, thermohaline circulation, meltwater pulse

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1. Introduction

[2] In the early Holocene about 10 kyr ago the northern summer solar irradiance attained a maximum and the global mean temperature finally reached, after an erratic transition from the Last Glacial Maximum (LGM), a value typical for interglacial conditions. Then around 8.2 kyr ago, records of relative abundance of the oxygen isotope $\delta^{18}O_{ice}$ from Greenland ice cores give evidence of an abrupt cooling event which disappeared within roughly two centuries. Although that temperature excursion constitutes only a fraction of the temperature excursions connected with the Dansgaard-Oeschger events and the Younger Dryas, the so-called 8.2 ka event represents an outstanding negative temperature anomaly during the otherwise stable Holocene period [*Dansgaard et al.*, 1993; *Johnsen et al.*, 1997; *Jouzel et al.*, 1997]. Records of $\delta^{18}O_{ice}$ from different Greenland ice cores, as from the Greenland Ice Core Project (GRIP and NorthGRIP) and the Greenland Ice Sheet Project 2 (GISP2) are in close agreement, aside from the inherent natural fluctuations (Figure 1). During the 8.2 ka event $\delta^{18}O_{ice}$ drops on average by 1.5‰ which corresponds to a surface air temperature drop of 3–6 K depending on the transformation method employed [*Johnsen et al.*, 1995; *Cuffey and Clow*, 1997; *Dahl-Jensen et al.*, 1998; *Johnsen et al.*, 2001]. Measurements of the relative abundance of the atmospheric nitrogen isotope $\delta^{15}N$ in the GRIP ice core give as the best estimate a cooling of 7.4 K in Greenland [*Leuenberger et al.*, 1999].

[3] The reduced air temperature during the 8.2 ka event is linked with drier conditions and stronger winds over the North Atlantic, colder and fresher conditions in the North Atlantic, drier monsoon regions and intensified North Atlantic trade winds [*Alley et al.*, 1997]. More recent paleoclimatic data from various locations in the Northern Hemisphere (NH) provide additional information on associated climate anomalies. These proxy data comprise marine sediments in the northern North Atlantic [*Klitgaard-Kristensen et al.*, 1998; *Bianchi and McCave*, 1999], lake sediments, tree ring and speleothem data in Europe [*von Grafenstein et al.*, 1998; *Klitgaard-Kristensen et al.*, 1998; *Tinner and Lotter*, 2001; *Korhola et al.*, 2002; *Baldini et al.*, 2002; *Magny et al.*, 2003], lake sediments and pollen data in North America [*Yu and Eicher*, 1998; *Hu et al.*, 1999; *Dean et al.*, 2002; *Shuman et al.*, 2002; *Spooner et al.*, 2002], ice

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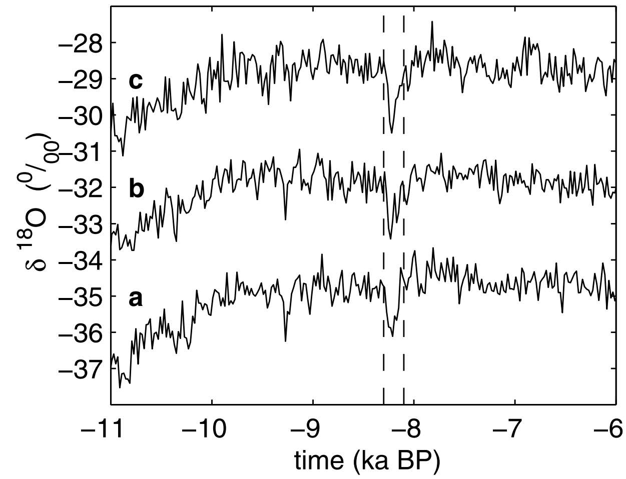

Figure 1. Time series of $\delta^{18}O_{ice}$ from Greenland ice cores (a) NorthGRIP, (b) GRIP, and (c)GISP2. The later two series are offset by 3‰ each. The drop in $\delta^{18}O_{ice}$ around 8.2 ka BP is marked by dashed lines at 8.3 and 8.1 ka BP.

core data from China [*Thompson et al.*, 1997], lake levels and stalagmite data in Africa [*Gasse*, 2000; *Neff et al.*, 2001; *Thompson et al.*, 2002; *Arz et al.*, 2003] and marine sediments in the tropical North Atlantic [*deMenocal et al.*, 2000]. Unlike the proxy data from the NH, ice core data from Bolivia [*Thompson et al.*, 1998] and from Antarctica [*Masson et al.*, 2000] show no clear signal of the 8.2 ka event.

[4] It is generally suggested that potential triggers for the outstanding 8.2 ka event could be either freshwater fluxes to the North Atlantic or changes in insolation. Insolation changes from solar activity contribute to climate anomalies and are of importance for inducing the 8.2 ka cold event [*van Geel et al.*, 1999; *Bond et al.*, 2001; *Muscheler et al.*, 2003]. However, several climate modeling studies indicate that an abrupt temperature drop centered over the North Atlantic as large as reconstructed for the 8.2 ka event, cannot be simply induced by changes in insolation, as was investigated, for instance, for the reduced insolation at the Maunder Minimum [e.g., *Cubasch et al.*, 1997; *Ganopolski et al.*, 2001; *Bauer et al.*, 2003]. However, model simulations by *Goosse et al.* [2002] with a coupled atmosphere-ocean-sea-ice model suggest that an abrupt cooling event can occur after a reduction in insolation through positive feedback effects. From their sensitivity experiments followed that cold events occur with a delay of 30–1150 years after the abrupt reduction in insolation which makes the begin of the cold event unpredictable. Therefore freshwater fluxes from the glacial ice sheets are considered as the major trigger as they attenuate the North Atlantic deep water (NADW) formation and thereby the northward heat transport [*von Grafenstein et al.*, 1999; *Barber et al.*, 1999; *Alley et al.*, 2003]. Model studies all agree that North Atlantic freshwater forcing is effective in influencing the climate [*Clark et al.*, 2002; *Rahmstorf*, 2002]. However, the response of the climate system is highly sensitive to the properties of freshwater forcing, such as its rate, its total volume and the forcing region. Furthermore, model studies suggest that the sensitivity of the climate system to freshwater forcing differs for different climate states, e.g., for glacial or interglacial states [*Ganopolski and Rahmstorf*, 2001a], and moreover different climate models may respond differently to the same forcing.

[5] A number of simulations have been performed to reproduce the Younger Dryas, which represents a relapse from interstadial to stadial conditions starting at about 12.9 ka BP and ceasing abruptly at 11.6 ka BP [e.g., *Maier-Reimer and Mikolajewicz*, 1989; *Wright and Stocker*, 1993; *Fanning and Weaver*, 1997; *Manabe and Stouffer*, 1997]. In these simulations constant freshwater fluxes were applied to the North Atlantic over multicentennial periods. The applied freshwater fluxes were based on reconstructions of the meltwater drainage from the Laurentide Ice Sheet (LIS) which amounts to a few tenths of one Sv (1 Sv = 10^6 m^3 s^{-1}) in the long-term mean.

[6] *Manabe and Stouffer* [1995] investigated abrupt climate changes with a coupled atmosphere-ocean general circulation model (GCM) using a pulse-like freshwater forcing in the North Atlantic. The climatic change induced by a 10-yearlong pulse of 1 Sv (equivalent to 3×10^{14} m^3) applied in the deep convection area, 50–70°N, was much stronger than applying the same pulse farther south at 20–50°N. The response in the former case was an abrupt weakening of the Atlantic overturning circulation from 19 to 7 Sv and a strong cooling over the North Atlantic. After only two decades the overturning circulation was restored to 15 Sv through enhanced northward advection of warm and saline surface waters. However, then a second weakening to 8 Sv occurred, partly caused by a decreased density of the surface water and an increased density of the subsurface water in northern regions which hindered deep convection. From the second weakening the system recovered slowly, within about one century, to its initial state.

[7] First simulations with a coupled atmosphere-ocean-sea-ice model to reproduce the 8.2 ka cold event were performed by *Renssen et al.* [2001, 2002]. Following *von Grafenstein et al.* [1998], they used a total freshwater discharge of 4.7×10^{14} m^3, and tested the climate response to pulse durations of 10, 20 and 50 years. An expansion of the forcing duration to 500 years was ruled out to induce a significant reduction in the overturning stream function [*Renssen et al.*, 2001, 2002]. The decadal freshwater forcing simulations resulted in a considerable weakening of NADW formation and a southward shift of the deep convection area. Large annual-to-decadal fluctuations occurred around a mode of reduced meridional overturning circulation for many centuries before the unperturbed state was restored. Ensemble simulations, started from different initial conditions of the early Holocene climate, indicated that the duration of the cold phase can exceed the pulse duration by a factor of up to 100. The main conclusion from these model simulations was that the duration of the cold phase is governed by the high-frequency (annual-to-decadal) climate variability and hence is unpredictable.

[8] Glaciological modeling on the drainage of the meltwater from the LIS decay improved continually [*Licciardi et al.*, 1998; *Marshall and Clarke*, 1999; *Clark et al.*, 2001; *Leverington et al.*, 2002; *Teller et al.*, 2002; *Clarke et al.*,

2004]. The analyses substantiate that the retreat of the LIS together with geographical circumstances lead to a concourse of meltwater in a lake system, termed shortly Lake Agassiz. Final remains of the LIS formed a northern dam which around 8.2 ka BP suddenly burst open, releasing a meltwater pulse of 1.6×10^{14} m^3 through the Hudson Bay into the North Atlantic [*Leverington et al.*, 2002; *Teller et al.*, 2002]. This water volume corresponds to 5.2 Sv if released in one year, which is three times less than used by *Renssen et al.* [2001, 2002] and two times less than used by *Manabe and Stouffer* [1995].

[9] In the following, CLIMBER-2 simulations are presented using a freshwater forcing adherent to the meltwater pulse released from Lake Agassiz. Section 2 describes transient simulations testing the sensitivity of the model to the freshwater pulse with additional random freshwater perturbations, and with baseline freshwater fluxes released through the St. Lawrence River. The simulations provide support for a weakly stable climate mode. Its stability properties are analyzed in section 3. The existence of such a climate mode assists in simulating a centennial cold event in reasonable agreement with paleoclimate data for the 8.2 ka event, as shown in section 4. Section 5 summarizes the main conclusions.

2. Transient Climate Simulations

2.1. Description of Climate Model and Boundary Conditions

[10] The climate model employed here is CLIMBER-2, version 3. CLIMBER-2 is a coupled climate model of intermediate complexity consisting of submodels for the atmosphere, the ocean with sea ice, and the vegetation. The submodels are interactively coupled through fluxes of heat and water, and momentum is transferred from the atmosphere to the ocean. CLIMBER-2 is designed for multimillennial climate simulations and therefore describes only the large-scale patterns of the dynamics on a coarse spatial resolution with daily time steps. The orography and the bathymetry are schematically resolved.

[11] The atmosphere model predicts the distributions of temperature, humidity and velocity components using a statistical-dynamical approach. The horizontal resolution is 10° in latitude and about 51° in longitude. The dynamical properties are determined on 10 vertical levels for the troposphere and the stratosphere. The radiation scheme for the short-wave and the long-wave radiation fluxes accounts for stratus and cumulus clouds, water vapor, carbon dioxide and aerosols. The radiative fluxes are calculated on 16 vertical levels. The ocean model computes the distributions of temperature, salinity, and meridional and vertical velocity components. The oceanic properties are determined as zonal means for the Atlantic, the Indian Ocean and the Pacific Ocean. In the southern circumpolar belt, which links the three ocean basins, the zonal mean of the zonal velocity is also computed. The ocean has a resolution of 2.5° in latitude and 20 layers in the vertical. The sea-ice distribution is computed by a thermodynamical sea-ice model including a parameterized advection scheme. Land surface processes are derived from the surface fluxes between land and atmosphere, soil moisture, and snow cover. The dynamical vegetation model has the same horizontal resolution as the atmosphere model. It calculates for each grid cell fractions of coverage of two vegetation types (grassland and forest) using annual mean precipitation and temperature above freezing. Further details on the climate model and its validation are given in the work of *Petoukhov et al.* [2000], *Ganopolski et al.* [2001] and *Brovkin et al.* [2002].

[12] The following transient simulations for the 8.2 ka event are started from an equilibrium state adapted to the boundary conditions for 9 ka BP. These boundary conditions are the solar irradiance according to the orbital parameters of eccentricity, obliquity and precession [*Berger*, 1978], the atmospheric CO_2 concentration of 261 ppm [*Raynaud et al.*, 2000], and a remnant Laurentide ice sheet on the North American continent [*Marshall and Clarke*, 1999]. The combined effect of the boundary conditions produces for 9 ka BP about the same global and hemispherical temperatures in the annual mean as for preindustrial conditions with CO_2 concentration of 280 ppm. The reason for this is that the annual mean effects on the total energy from lowering the atmospheric CO_2 concentration balance the effects from introducing a fraction of glacial coverage on the North American continent. (The neglect of the remnant LIS resulted in a larger annual mean temperature for 9 than for 0 ka BP.) With the given boundary conditions also the global annual mean precipitation of 2.8 mm/day and the NADW formation of 20 Sv are about the same for 9 ka BP and for preindustrial conditions. However, at 9 ka BP the top-of-the-atmosphere solar irradiance at 65°N was 42 Wm^{-2} larger for June and was 3 Wm^{-2} smaller for December than at 0 ka BP [*Berger*, 1978]. This implies a stronger seasonal temperature cycle in the early Holocene leading to a stronger monsoon circulation. Furthermore, the surface air temperature in the North Atlantic sector (60–80°N) is about 0.6 K warmer at 9 than at 0 ka BP, which results mainly from the about 10% smaller NH sea-ice area at 9 ka BP. The annual mean climate characteristics obtained from multimillennial equilibrium simulations for 9 and 0 ka BP are summarized in Table 1.

2.2. Freshwater Pulse and Noise

[13] *Teller et al.* [2002] and *Leverington et al.* [2002], who employed a glaciological model for the decay of the LIS, suggest a pulse-like drainage of meltwater from the Lake Agassiz around 8.2 ka BP. The estimated pulse volume is 1.6×10^{14} m^3, which corresponds to 5.2 Sv if released in one year. Applying the same volume in two years to the Atlantic between 50 and 70°N in CLIMBER-2, results in a rapid decrease in the surface air temperature over the North Atlantic sector (60–80°N) by 3.6 K. This cold phase lasts only two decades (Figure 2, dotted line). The thereby enhanced meridional temperature gradient induces a rapid response, involving a warming over the North Atlantic. This warming is largely caused by northward advection of heat and by a reduction in the sea-ice area in the North Atlantic. This warming for a few years is followed by a secondary cool phase in response to the lagged restoration of the deep convection, which is similar to the response described by *Manabe and Stouffer* [1995] and by *Renssen et*

Table 1. Characteristics of Holocene Equilibrium States[a]

Parameter	Value	Initial	Intermediate	Present
		Boundary Conditions		
Insolation	year BP	9000	8200	0
CO_2	ppm	261	261	280
		Climate Variables		
T	deg C	13.9	13.9	13.9
T_{NH}	deg C	14.3	13.8	14.3
T_{SH}	deg C	13.6	14.0	13.6
T_{NA}	deg C	−5.0	−8.2	−5.6
PRC	mm/day	2.8	2.7	2.7
A_{NH}	10^6 km^2	8.3	10.4	9.6
A_{SH}	10^6 km^2	13.3	10.3	13.6
NADW	Sv	19.9	13.6	20.1
AABW	Sv	−17.0	−20.5	−16.6

[a]The characteristics of the initial and present (preindustrial) climate states from equilibrium simulations with CLIMBER-2 are compared with the characteristics of the intermediate (INT) equilibrium state connected with the 8.2 ka event. The boundary conditions for these simulations are the insolation depending on orbital parameters, atmospheric CO_2 concentration and a remnant glaciated area in North America applied in the initial and the intermediate state. The characteristic climate variables are the annual mean surface air temperature: T, global; T_{NH}, Northern Hemisphere; T_{SH}, Southern Hemisphere; T_{NA}, North Atlantic sector (60–80°N); the precipitation: PRC, global; the sea-ice area: A_{NH}, NH; A_{SH}, SH; and the ocean deep water formation: NADW, North Atlantic Deep Water; and AABW, Antarctic Bottom Water.

al. [2002]. The total duration of the cold phase is about 70 years which is much shorter than inferred from the paleoclimatic data for the 8.2 ka event. Thus CLIMBER-2 simulations corroborate previous studies with coupled atmosphere-ocean models, showing that a short-term freshwater pulse alone is insufficient to produce a 200-yearlong climate anomaly.

[14] So far, any short-term variability in the runoff associated with the melting of the LIS has been ignored. The variability of the meltwater flux routed into the North Atlantic is poorly known and may be considered to range between 0.01 and 0.1 Sv. The relative broad range of the freshwater fluctuations results from the inherent internal variability of glaciological and atmospherical processes. Hence we simulate the potential impact from a randomly varying freshwater flux by adding white noise of different standard deviation (σ) to the surface freshwater fluxes computed by the model. The model freshwater fluxes consist of precipitation, evaporation, river runoff, sea-ice melting and sea-ice formation. The impact of additional freshwater fluctuations is tested through ensembles of Monte Carlo simulations. Each Monte Carlo ensemble is obtained with a different σ, and each simulation of an ensemble is obtained with a different seed supplied to the random number generator. Thereby each simulation of an ensemble is driven by a different realization of random numbers. The sampling interval of the noise is one year. The white noise flux is continuously applied in the same latitudinal belt (50–70°N) where the freshwater pulse is released.

[15] The Monte Carlo simulations show that freshwater noise superposing the freshwater pulse has a minor influence on the amplitude of the temperature drop, but can influence the duration of the cold phase considerably. This is demonstrated by time series of surface air temperature anomalies for the North Atlantic sector from 31 Monte Carlo simulations with σ = 0.05 Sv (Figure 3a). After the freshwater pulse, the temperature drop ΔT_{NA} remains close to 3.6 K, while the duration of the cold phase varies between 15 and 150 years. The associated rate of NADW formation weakens rapidly by about 40%, and as long as the surface air temperature is reduced, NADW formation is about 30% weaker than initially (Figure 3b). Two series of ΔT_{NA} from the Monte Carlo ensemble with an extended cold period are compared with the series from the noise-free simulation and the temperature anomalies reconstructed from GISP2 data using the conversion scheme of *Cuffey and Clow* [1997, Figure 2]. The simulated temperature anomalies obtained with the freshwater noise correspond more closely to the reconstructed anomalies in terms of duration of the cold event than those without noise. However, the simulations subject to noise still underestimate the temperature variability.

2.3. Freshwater Pulse, Noise, and Baseline Freshwater Flux

[16] Several lines of evidence indicate enhanced runoff through the St. Lawrence River prior to the 8.2 ka event, originating from the ongoing melting of the LIS or from the overflow of proglacial lakes. Estimates of such a meltwater runoff range up to 0.15 Sv [*Clark et al.*, 2001]. To mimic that runoff, a continuous baseline freshwater flux is added to the North Atlantic at 40–50°N for 1000 years. Sensitivity experiments show that a baseline flux larger than 0.1 Sv together with the freshwater pulse (without noise) leads to a collapse of the overturning circulation. Such a flux brings

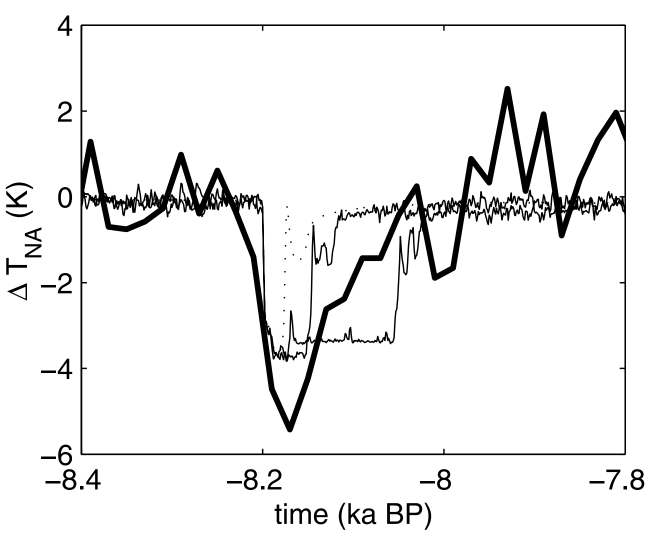

Figure 2. Time series of surface air temperature anomalies of North Atlantic sector (60–80°N), ΔT_{NA}, from three simulations, and GISP2 $\delta^{18}O_{ice}$ data (thick line) converted according to *Cuffey and Clow* [1997]. The simulation with only a freshwater pulse (thin dotted line) produces a cooling by 3.6 K lasting about 20 years, while two other simulations (thin continuous lines) chosen with additional freshwater noise of σ = 0.05 Sv produce similar cooling but lasting up to 150 years.

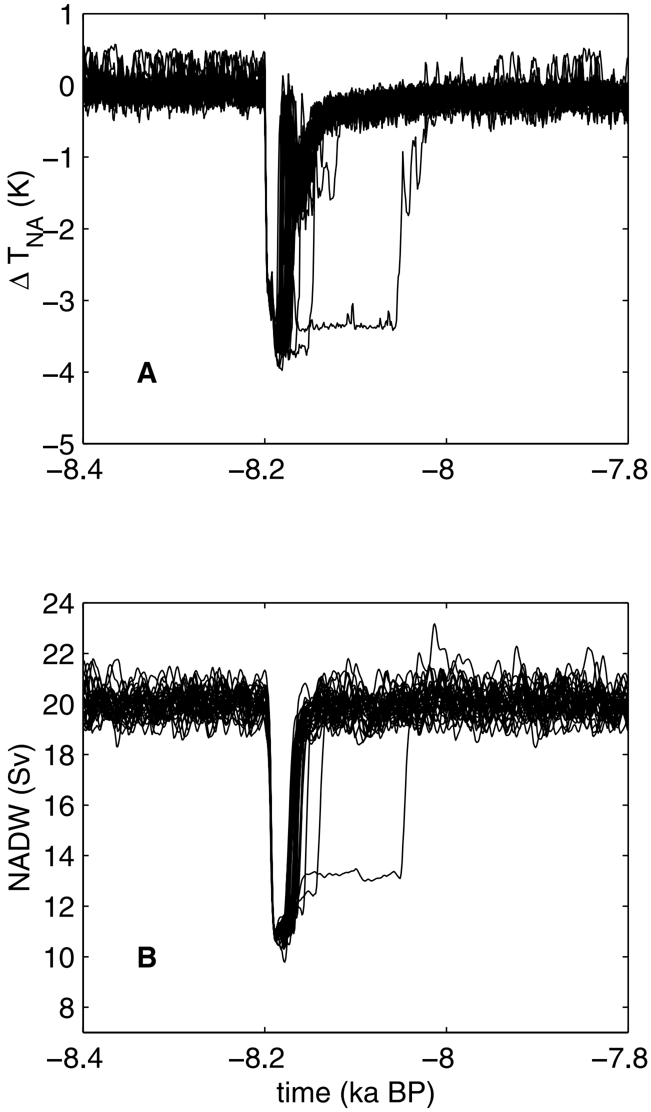

Figure 3. Time series from an ensemble of 31 Monte Carlo simulations with freshwater pulse at 8.2 ka BP and freshwater flux fluctuations of σ = 0.05 Sv applied in Atlantic 50–70°N. (a) The surface air temperature response ΔT_{NA} in K. (b) Maximum of the North Atlantic meridional stream function (NADW) in Sv.

the overturning circulation toward its bifurcation transition, beyond which the meridional overturning circulation cannot be sustained [*Rahmstorf*, 1995a; *Ganopolski and Rahmstorf*, 2001a]. Prior to the collapse, the intensified meridional temperature gradient induced through the northern cooling after the freshwater pulse causes a temporary strengthening of the meridional circulation and thereby an enhanced northward transport of saline water. After the collapse of NADW formation, the temperature over the North Atlantic sector drops by more than 5 K.

[17] Simulations with the additional baseline freshwater flux applied in 40–50°N show a minor additional cooling over the North Atlantic and a minor additional weakening of the overturning circulation compared to the simulations without the baseline freshwater flux (see Figures 3 and 4). Figure 4 shows the time series from 31 Monte Carlo simulations obtained with the same realizations of the freshwater noise as in Figure 3 but with the additional baseline freshwater flux of 0.06 Sv. Though the baseline freshwater flux has in most cases little effect on the temporal evolution of the cold phase, in a few cases the evolution of the cold phase is more variable and can be prolonged by up to 200 years.

[18] The impact of the freshwater pulse and the baseline freshwater flux obviously becomes difficult to predict when freshwater noise is added to the North Atlantic. Figure 5 selects four simulations of NADW formation obtained with the two-year freshwater pulse of 2.6 Sv at 50–70°N, a baseline freshwater flux of 0.03 Sv at 40–50°N until 8 ka BP, and with freshwater noise. The four simulations differ in σ of either σ = 0.05 or 0.07 Sv, and in perturbation area of

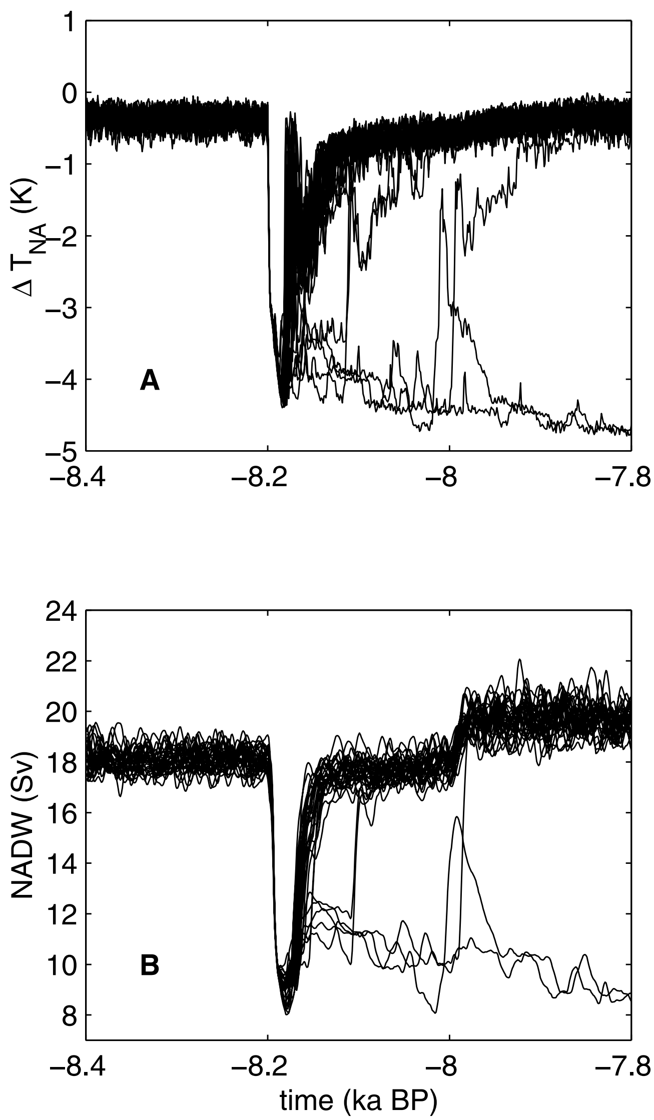

Figure 4. Same as Figure 3 but with additional baseline freshwater flux of 0.06 Sv until 8 ka BP applied in Atlantic 40–50°N.

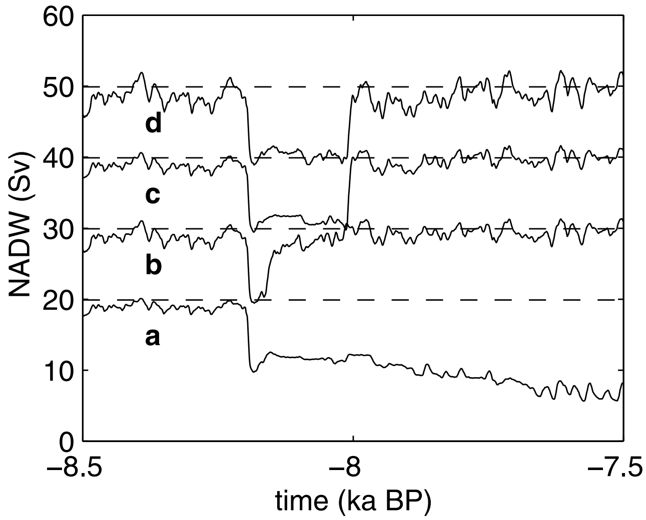

Figure 5. Time series of NADW in Sv obtained with freshwater pulse at 8.2 ka BP, baseline freshwater flux of 0.03 Sv until 8 ka BP, and freshwater fluctuations differing in σ and perturbation area. (a and b) σ = 0.05 Sv and (c and d) σ = 0.07 Sv; perturbation area is 50–70°N in Figures 5a and 5c and 40–70°N in Figures 5b and 5d, while seed for random number generator is unchanged. Time series are offset by 10 Sv and dashed lines mark reference value of initial state.

either 50–70°N or 40–70°N, while the seed supplied to the number generator is unchanged. The time series in Figure 5 suggest that freshwater noise applied in the North Atlantic may either promote (Figure 5a) or impede (Figures 5b–5d) the collapse of NADW formation. Sometimes simulations with freshwater fluctuations show a climate response extending over two centuries after the two-year freshwater pulse. This was interpreted as a hint at the existence of an intermediate stable mode of the meridional overturning circulation in the North Atlantic. The properties of that mode, found in CLIMBER-2, are discussed next.

3. Intermediate Climate Mode

3.1. Climate Characteristics

[19] Commonly, reference is made to two basic climate modes, which are described as the "ON" mode and the "OFF" mode of the meridional overturning circulation in the North Atlantic. A typical circulation pattern of the ON mode is presented by the equilibrium circulation for 9 ka BP in Figure 6a, while the OFF mode is reached after the collapse of the NADW formation (not shown). In addition to the common two circulation modes, a circulation mode intermediate to the ON and OFF mode is obtained by an transiently enhanced freshwater flux into the North Atlantic. This mode is hereafter called the "INT" mode.

[20] The INT mode can be reached from the equilibrium state at 9 ka BP by adding a freshwater flux of 0.04 Sv for 800 years to the Atlantic between 40 and 50°N, followed by a two-yearlong freshwater pulse of 2.6 Sv at 50–70°N. Then every transient forcing is switched off (i.e., no additional freshwater forcing and fixed orbital parameters) and the system is integrated for 5000 years to reach full equilibrium. The INT equilibrium state exhibits about the same global annual mean temperature and precipitation as obtained for the early Holocene at 9 ka BP and for the preindustrial climate state (Table 1). However, the INT state has a lower NH temperature and a higher Southern Hemisphere (SH) temperature than the other two climate states. Hence the INT state includes a larger northern sea-ice area, a smaller southern sea-ice area, and the NADW formation is considerably reduced while the Antarctic Bottom Water (AABW) formation is increased.

[21] The weaker North Atlantic overturning circulation of the INT state is associated with a shift of the area of deep convection from latitudes north of 60°N to latitudes south of 50°N (Figure 6). This implies that the overturning circula-

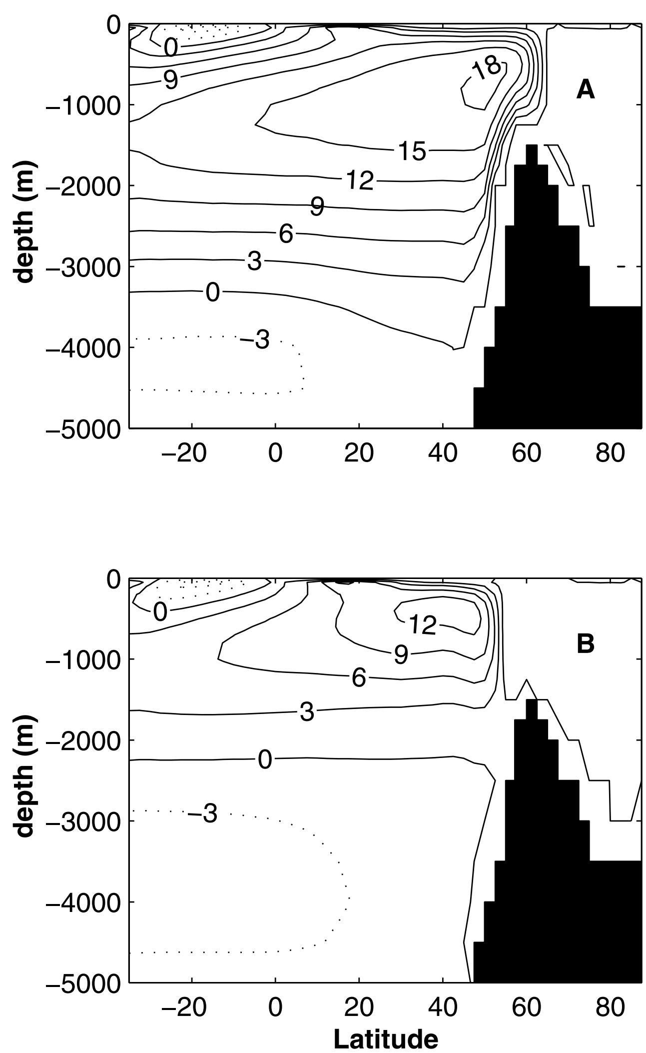

Figure 6. Atlantic meridional stream function in Sv of equilibrium ON state and INT state adapted to boundary conditions (Table 1) for (a) 9 ka BP and (b) 8.2 ka BP, respectively. Isolines are in steps of 3 Sv with continuous lines for clockwise circulation and dotted lines for anticlockwise circulation.

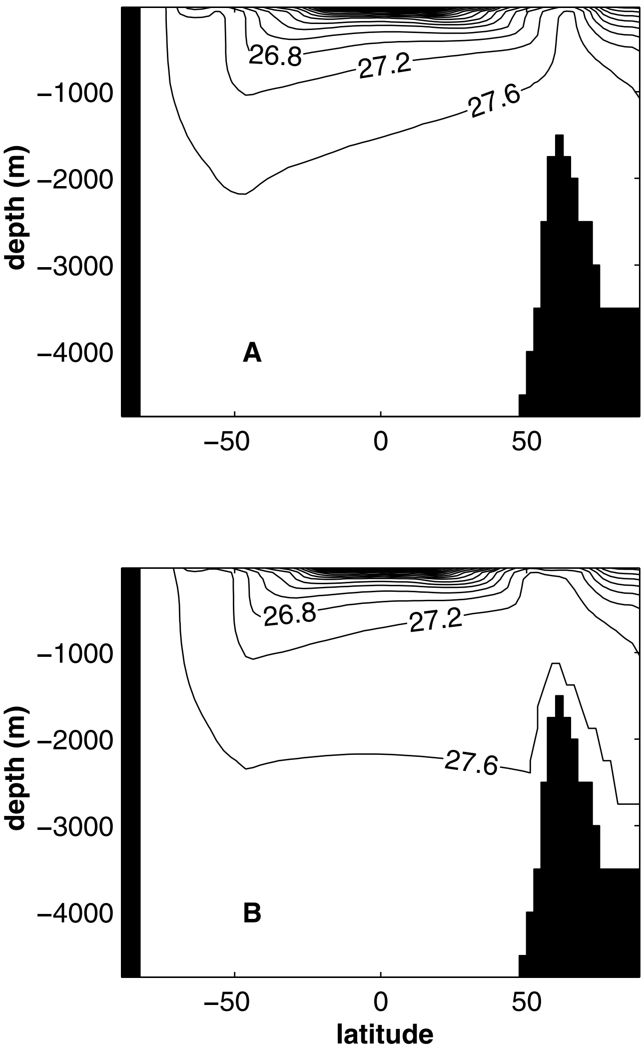

Figure 7. Same as Figure 6 but for potential density in kg m^{-3} above 10^3kg m^{-3} with isolines in steps of 0.4 kg m^{-3}.

tion cell is shallower and that NADW extends less deeply while AABW extends further northward. The ocean circulation changes are connected with changes in the distribution of temperature and salinity involving different density distributions between the initial ON state and the INT state (Figure 7). The southward shift of deep convection is evident from the southward shift of the outcropping of the subsurface isopycnals. The shallowing of the overturning cell is related to a more stable density stratification of the water column enclosing the outcropping region. The weakening of NADW formation contributes to the cooling in the North Atlantic through a 20% decrease in the maximum of the northward heat transport. This leads to a warming in the SH and a decrease of the southern sea-ice area in reaction of the bipolar seesaw effect [*Crowley*, 1992]. The spatial circulation pattern of the INT state resembles the circulation pattern of the "cold" mode obtained for LGM conditions [*Ganopolski and Rahmstorf*, 2001a], though the NADW formation is weaker in the INT mode than in the glacial mode.

3.2. Stability Analysis

[22] The stability of the Atlantic overturning circulation is usually tested by analyzing the response of the zonal mean stream function to a slowly changing freshwater flux, such that the system can be considered to be continually in equilibrium. Thereby typical freshwater fluxes are obtained through which the overturning circulation bifurcates from the ON state to the OFF state and vice versa [*Stocker and Wright*, 1991; *Rahmstorf*, 1995a]. In order to find other possible stable modes, we use a very slow freshwater flux

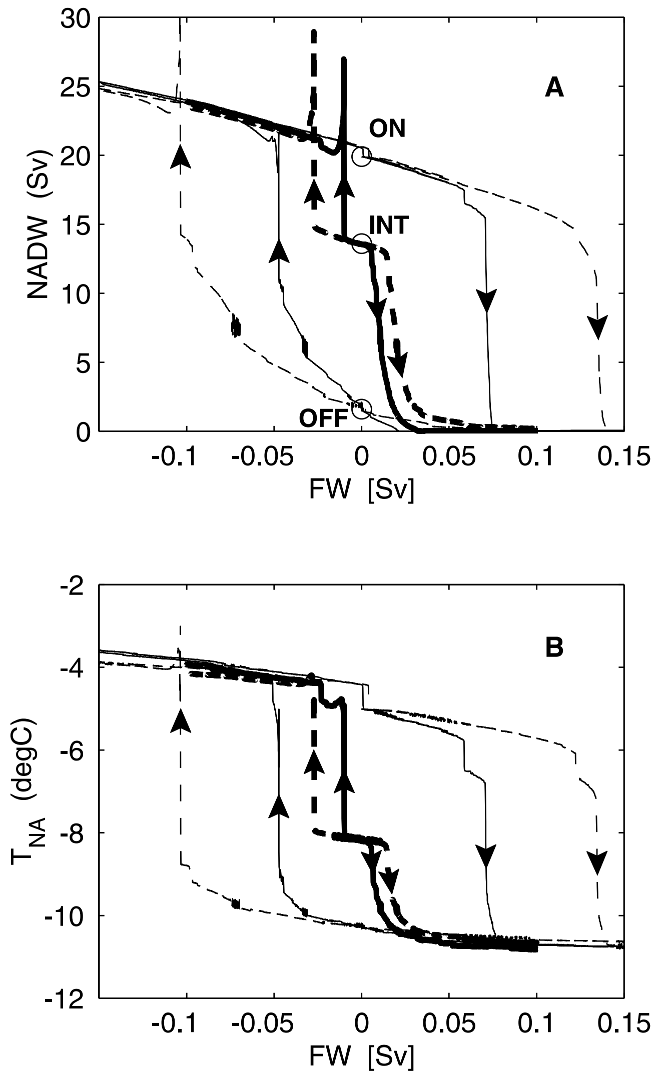

Figure 8. Stability diagrams showing (a) NADW formation in Sv and (b) T_{NA} in deg.C in response to freshwater flux changes in the North Atlantic (FW) using a rate of 0.01 Sv/kyr. The equilibrium states are marked by circles, where the INT state lies between the ON and OFF states on the hysteresis loops. The stability of the INT state is traced by thick lines starting from the INT state, while stabilities of ON and OFF states are traced by thin lines starting from the ON state. Results are shown for freshwater forcing applied at 50–70°N and 20–50°N by continuous and dashed lines, respectively.

rate of 0.01 Sv/kyr. Hysteresis experiments, started from the ON state at 9 ka BP, however, reproduced the hysteresis loop and its bifurcation characteristics as obtained for modern conditions *Ganopolski and Rahmstorf*, 2001a]. The typical freshwater flux needed to reach the bifurcation transition from the ON mode to the OFF mode is about 0.07 Sv and 0.13 Sv for freshwater added at 50–70°N and 20–50°N, respectively (Figure 8).

[23] In the stability diagram, the INT state lies intermediate to the ON and the OFF states (Figure 8a). The weaker NADW formation in the INT state is connected with a colder surface air temperature in the North Atlantic sector (Figure 8b). The cooling results from a reduced northward heat transport and from an increased NH sea-ice area which implies a stronger sea-ice albedo feedback. The INT state is attained by an abrupt freshwater forcing applied to the North Atlantic circulation, which was initialized with boundary conditions for 9 ka BP. Motivated by estimates of the LIS meltwater production, a weak freshwater flux of 0.04 Sv for 800 years is applied in the latitudinal belt 40–50°N, and then a two-year pulse of 1.6×10^{14} m^3 is released in the belt 50–70°N. The location of the INT state in the stability diagram is seen to differ from the secondary "on" state obtained with an ocean GCM by *Rahmstorf* [1995a], as no trace leads from the hysteresis loop to the INT state. More details on the potential for a reduced meridional overturning circulation in conjunction with multiple convection patterns are discussed in the work of *Rahmstorf* [1995b].

[24] The stability of the INT state is tested by applying either negative or positive freshwater fluxes with the same rate and in the same two regions as in the hysteresis experiments, but starting from the INT state. The stability of the INT state is weaker than those of the ON and the OFF states. Changes in the NADW formation rate of the INT state can be induced by considerable smaller freshwater fluxes than those needed to reach the bifurcation transitions of the hysteresis loop. The sensitivity of the INT state to freshwater forcing applied in the southern region 20–50°N is smaller than applied in 50–70°N. This is related first to the fact that a subtraction of freshwater in 50–70°N directly reinforces the northern deep water convection, and therefore the ON mode is restored more rapidly than by subtracting freshwater in 20–50°N. Second, an addition of freshwater in the northern region, where the net freshwater flux from precipitation, evaporation, runoff and sea-ice contributions has a local maximum [*Ganopolski and Rahmstorf*, 2001b], is more effective in impeding the NADW formation than adding freshwater in the southern region.

[25] The stability analysis of the INT mode by monotonously changing freshwater fluxes indicates an asymmetric

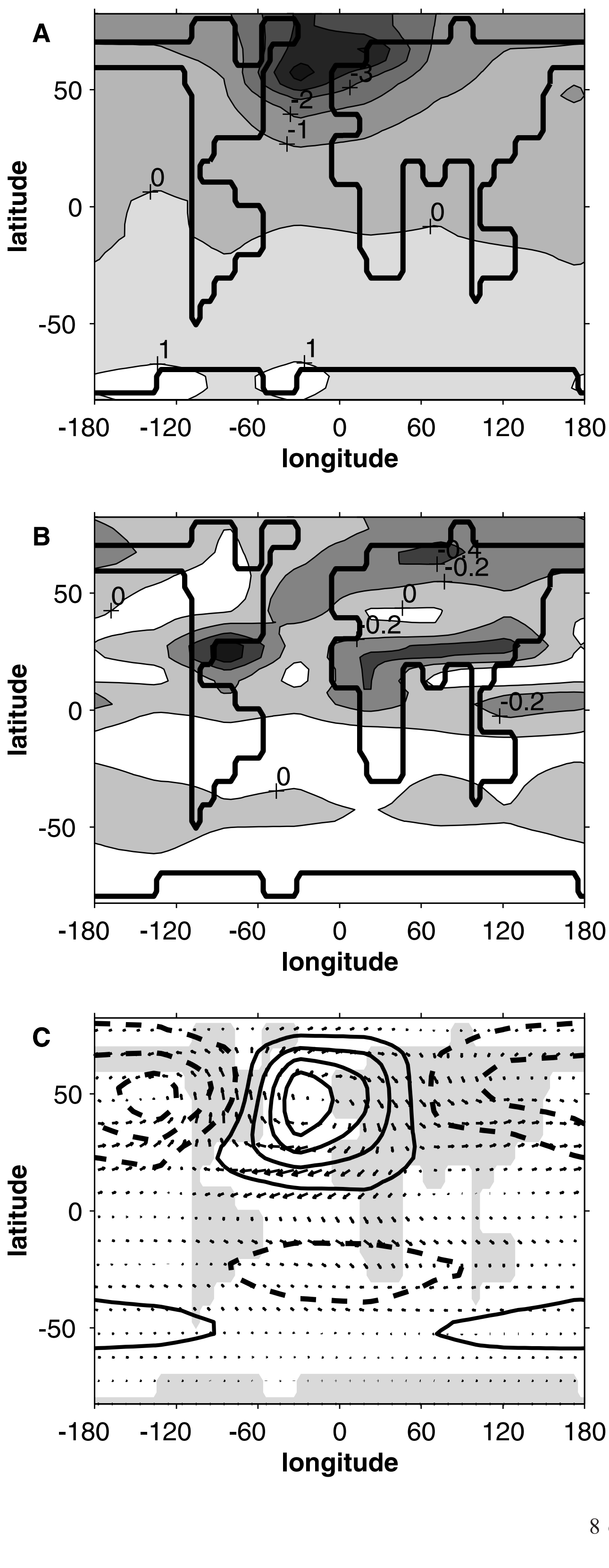

Figure 9. Decadal mean climate anomalies for the 8.2 ka event from a simulation with freshwater pulse of 1.6×10^{14} m^3 through Hudson Strait, overlaid by noise with σ = 0.05 Sv and baseline freshwater flux of 0.06 Sv through St. Lawrence relative to the "ON" state at 9 ka BP showing (a) surface air temperature anomalies in steps of 1 K, (b) summer (JJA) precipitation anomalies in steps of 0.2 mm/day, and (c) changes of surface wind vectors (arbitrarily scaled) together with sea level pressure anomalies in steps of 0.4 hPa with positive anomalies mainly over the North Atlantic. See color version of this figure at back of this issue.

response of the overturning circulation with respect to the freshwater forcing. The asymmetry is seen from the larger negative freshwater flux needed to initialize the ON mode than the positive freshwater flux needed to initialize the OFF mode. This suggests that random freshwater perturbations would more likely push the North Atlantic overturning circulation toward the OFF mode than to the ON mode. However, once the deviation from the INT mode begins, the ON mode is reached more rapidly than the OFF mode. The rapidity of recovery is also visible by the overshoot effect in the stability diagram.

[26] The asymmetric response behavior brings us to the question of the balance between the effects related to the amplitude of the freshwater forcing and related to the response timescale. In an attempt to obtain an answer we performed ensembles of Monte Carlo simulations, applying white noise forcing to the North Atlantic circulation resting in the INT mode. The answer to the question strongly depends on the properties of the freshwater perturbations, such as the probability density distribution of the perturbations, and the area of perturbation. Noting that the set of simulations performed cannot meet the full complexity of the climate system, the findings can be summarized by three items:

[27] 1. The stable INT mode is resilient to white noise freshwater fluctuations with $\sigma = 0.01$ Sv, independent on the area of perturbation. If the freshwater perturbations are applied only in the southerly latitudes then the INT state is even retained for σ less or equal to 0.03 Sv.

[28] 2. The duration of the cold event can be prolonged if the freshwater perturbations are released only in the southerly latitudes. The thereby randomly reduced surface salinity of the northward transported water can hinder convection in the northern deep convection area through a stabilizing effect of the vertical density stratification.

[29] 3. The analysis of ensembles of Monte Carlo simulations with different σ suggests that the probability of transition from the INT mode to the ON mode is generally larger than the probability of transition from the INT mode to the OFF mode.

4. Simulated Climate Anomalies for the 8.2 ka Event

[30] The simulated climate anomalies through the meltwater outburst from Lake Agassiz to the North Atlantic are widespread. For demonstration, we choose a simulation from the Monte Carlo ensemble shown in Figure 4 which shows a two-century long cold phase. Spatial distributions of climate anomalies for 8 ka BP relative to the initial climate distributions at 9 ka BP are displayed as decadal means in Figure 9.

[31] The surface air temperature decrease is centered over the North Atlantic, reaching up to 5 K, and diminishes toward Europe, North America and equatorward (Figure 9a). In the south Atlantic and Pacific adjacent to Antarctica the temperature anomaly is reversed and exhibits a warming of 1 K. The precipitation anomalies are largest during the northern summer (JJA). In consequence of the differential cooling over land and ocean, the northern summer monsoon weakens and precipitation decreases over the African and American monsoon regions (Figure 9b). In zones northward and southward of the monsoon regions, precipitation increases partially and decreases again toward northern polar latitudes. The anomalies in the atmospheric circulation are mainly evident from the intensified trade winds in the tropical Atlantic (Figure 9c). Also the midlatitude westerlies in the NH intensify while the southern circumpolar westerlies attenuate slightly.

[32] Figure 10 shows time series from 8.4 to 7.8 ka BP of several climate variables from the same simulation as in Figure 9. The freshwater forcing composed of the two-year pulse, the baseline freshwater flux and the freshwater noise is displayed in Figure 10a. NADW formation decreases abruptly to 9 Sv after the release of the freshwater pulse, varies at about 11 Sv for about two centuries, and recovers rapidly to 20 Sv around 8 ka BP (Figure 10e). The annual-mean temperature over the North Atlantic sector (60–80°N), northwestern Europe (50–60°N) and North America (40–60°N) show a similar long cooling by 3.8 K (Figure 10b), 2.3 K (Figure 10c), and 0.8 K (Figure 10d), respectively. This cooling is in reasonable agreement with paleodata [e.g., *von Grafenstein et al.*, 1998]. At the same time the annual precipitation is reduced in the Greenland area (Figure 10f) in accordance with ice core data [*Alley et al.*, 1997]. The summer (JJA) precipitation in the African monsoon region (10–30°N), which is gradually decreasing during the Holocene, shows a drop during the cold phase (Figure 10g), similarly as indicated by *Gasse* [2000]. Figure 10h shows a slow decrease in the fraction of trees in North America (50–60°N) in agreement with pollen data [*Hu et al.*, 1999]. Since the climate model has a rather low resolution, the comparison of the simulation results in Figure 10 with paleodata is only of qualitative significance.

5. Conclusions

[33] The climate model CLIMBER-2 was used to test forcing mechanisms which can explain the observed climate anomalies about 8200 years ago. Among the proposed driving forces the release of meltwater from the Laurentide Ice Sheet to the North Atlantic is seen as the most effective driver. In particular, we suggest that the rapid release of the meltwater volume of 1.6×10^{14} m^3 from Lake Agassiz through the Hudson Strait could have triggered the 8.2 ka cold event. The meltwater flooded the northern North Atlantic and suppressed NADW formation in the Labrador Sea and the Nordic Seas. Thereafter, NADW formed south of 50°N at a rate 40% lower than initially. This southward shifted and shallower meridional overturning circulation is

Figure 10. Time series of regional climate variables from same simulation as for Figure 9, showing (a) the composite freshwater forcing, (b) the temperature over Atlantic sector 60–80°N, (c) the temperature over Europe 50–60°N, (d) the temperature over America 40–60°N, (e) the NADW formation rate, (f) the precipitation in Atlantic sector 60–80°N, (g) the precipitation in North African summer monsoon region (10–30°N), and (h) the tree fraction in America 50–60°N.

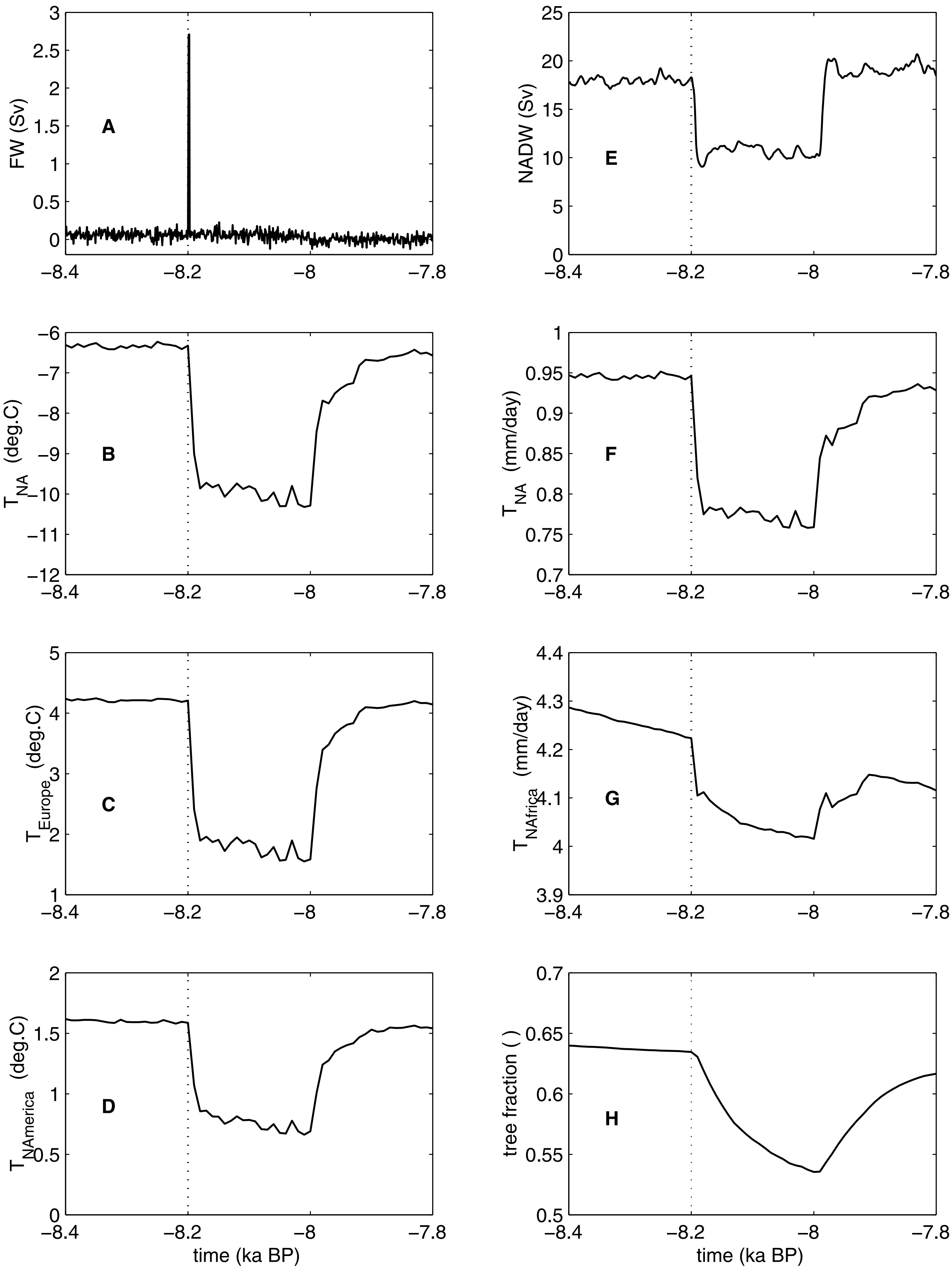
A
FW (Sv)
E
NADW (Sv)
B
T_{NA} (deg.C)
F
T_{NA} (mm/day)
C
T_{Europe} (deg.C)
G
$T_{NAfrica}$ (mm/day)
D
$T_{NAmerica}$ (deg.C)
H
tree fraction ( )
time (ka BP)
time (ka BP)

Figure 10

found to be in the neighborhood of a weakly stable climate mode, the INT mode. As long as this situation persists the large-scale climate in the NH is anomalously cold, mostly drier, and westerlies and trade winds are stronger.

[34] Hints at the existence of a weakly stable climate mode are obtained by the different response behavior of the climate model before and after the freshwater pulse in reaction to random freshwater fluctuations in the North Atlantic. The equilibrium state of the INT climate mode could be attained through a small (0.04 Sv) multicentury runoff of meltwater routed along the St. Lawrence pathway completed by the outburst flood routed through the Hudson Strait. The INT equilibrium state is found to be resilient with respect to random freshwater fluctuations which have amplitudes close to those of natural freshwater fluctuations in the North Atlantic connected with the internal climate variability. This follows from idealized Monte Carlo simulations with the coupled climate model. However, the Monte Carlo simulations also indicate that the stability of the INT state depends substantially on the area of the freshwater perturbation.

[35] It is conceivable that once the circulation pattern is shifted by the freshwater pulse in the neighborhood of the weakly stable INT mode, also natural climate variations of other origin, such as from fluctuations in insolation, are unable to push the system away from that mode and possibly can prolong the perturbed climate state. However, when the meltwater runoff eventually wanes, the simulations suggest that the initial circulation pattern is rapidly restored and the cold phase ends. The probability of a collapse of NADW formation is found to be rather low, provided the freshwater flux rates are reasonably constrained, because the timescale of restoration is much shorter than the timescale of collapse. Thus freshwater flux rates as derived from data on the demise of the LIS appear sufficient to produce a two-century long episode with a temperature drop of 3–6 K over Greenland.

[36] The existence of the weakly stable climate mode, intermediate to the common ON and OFF modes, is only partially surprising. First, the model simulations presented by *Renssen et al.* [2001, 2002] which are the only other climate model simulations dedicated to 8.2 ka cold event provided already hints at a moderate stable mode with a weaker NADW circulation. This weaker NADW circulation is also characterized by a southward shift of the main deep convection site and a more shallow circulation pattern. The duration of the weaker circulation pattern is governed by the high-frequency (annual-to-decadal) climate variability. This results from the nonlinear interactions in the atmosphere–ocean–sea-ice interface which causes the duration of the intermediate climate state to become unpredictable. Second, the CLIMBER-2 model shows for LGM conditions, when the global climate was considerably colder than at present, a North Atlantic circulation pattern similar to the INT mode. That "cold" mode [*Ganopolski and Rahmstorf*, 2001a] also exhibited deep convection south of 50°N, though the formation of NADW was more intense and extended deeper during the LGM than in the INT mode. Tests on the robustness of the INT mode show that the INT mode is not only attainable for early Holocene conditions, but also for climate conditions with present-day solar irradiance and preindustrial atmospheric CO_2 concentrations.

[37] The present climate model study demonstrates that knowledge of the temporal evolution of the meltwater flux rates and the meltwater pathways during the final stage of the Laurentide Ice Sheet is important for simulating the 8.2 ka cold event. The abruptly released meltwater volume [*Teller et al.*, 2002; *Leverington et al.*, 2002] is an essential ingredient in the present model simulation. The results obtained with the climate model offer a possible mechanism for producing large-scale climate anomalies, which are consistent with paleoclimate data from the NH for the 8.2 ka cold event. Short-term deviations between the simulated time series of climate variables and the reconstructed series are likely, because the climate model underestimates the natural climate variability, and at the same time the applied fluctuations in the runoff from the LIS melting are highly simplified in the present study. To obtain further confirmation for the occurrence of the outburst flood from Lake Agassiz other effects should be taken into account. The considered outburst flood, which corresponds to a global-mean sea level rise of 0.5 m, can be expected to be accompanied by abrupt flushing events and by the generation of huge gravity waves. Fast-propagating surface waves could have led to massive floods in coastal regions adjacent to the North Atlantic, for which further evidence should be available.

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References

Alley, R. B., P. A. Mayewski, T. Sowers, M. Stuiver, K. C. Taylor, and P. U. Clark (1997), Holocene climate instability: A prominent, widespread event 8200 yr ago, *Geology*, *25*(6), 483–486.

Alley, R. B., et al. (2003), Abrupt climate change, *Science*, *299*, 200–210.

Arz, H. W., F. Lamy, J. Pätzold, P. J. Müller, and M. Prins (2003), Mediterranean moisture source for early-Holocene humid period in the Red Sea, *Science*, *300*, 118–121.

Baldini, J. U. L., F. McDermott, and I. J. Fairchild (2002), Structure of the 8200-year cold event by speleothem trace element record, *Science*, *296*, 2203–2206.

Barber, D. C., et al. (1999), Forcing of the cold event of 8,200 years ago by a catastrophic drainage of Laurentide lakes, *Nature*, *400*, 344–348.

Bauer, E., M. Claussen, V. Brovkin, and A. Hünerbein (2003), Assessing climate forcings of the Earth system for the past millennium, *Geophys. Res. Lett.*, *30*(6), 1276, doi:10.1029/2002GL016639.

Berger, A. L. (1978), Long-term variations of daily insolation and Quaternary climate changes, *J. Atmos. Sci.*, *35*, 2362–2367.

Bianchi, G. G., and I. N. McCave (1999), Holocene periodicity in North Atlantic climate and deep-ocean flow south of Iceland, *Nature*, *397*, 515–517.

Bond, G., B. Kromer, J. Beer, R. Muscheler, M. N. Evans, W. Showers, S. Hoffmann,

R. Lotti-Bond, I. Hajdas, and G. Bonani (2001), Persistent solar influence on the North Atlantic climate during the Holocene, *Science*, *294*, 2130–2136.

Brovkin, V., J. Bendtsen, M. Claussen, A. Ganopolski, C. Kubatzki, V. Petoukhov, and A. Andreev (2002), Carbon cycle, vegetation and climate dynamics in the Holocene: Experiments with the CLIMBER-2 model, *Global Biogeochem. Cycles*, *16*(4), 1139, doi:10.1029/2001GB001662.

Clark, P. U., S. J. Marshall, G. K. C. Clarke, S. W. Hostetler, J. M. Licciardi, and J. T. Teller (2001), Freshwater forcing of abrupt climate change during the last glaciation, *Science*, *293*, 283–287.

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.

Clarke, G. K. C., D. W. Leverington, J. T. Teller, and A. S. Dyke (2004), Paleohydraulics of the last outburst flood from glacial Lake Agassiz and the 8200 BP cold event, *Quat. Sci. Rev.*, *23*, 389–407.

Crowley, T. J. (1992), North Atlantic deep water cools the Southern Hemisphere, *Paleoceanography*, *7*(4), 489–497.

Cubasch, U., R. Voss, G. C. Hegerl, J. Waszkewitz, and T. J. Crowley (1997), Simulation of the influence of solar radiation variations on the global climate with an ocean-atmosphere general circulation model, *Clim. Dyn.*, *13*, 757–767.

Cuffey, K. M., and G. D. Clow (1997), Temperature, accumulation, and ice sheet elevation in central Greenland through the last deglacial transition, *J. Geophys. Res.*, *102*(C12), 26,383–26,369.

Dahl-Jensen, D., K. Mosegaard, N. Gundestrup, G. D. Clow, S. J. Johnsen, A. W. Hansen, and N. Balling (1998), Past temperatures directly from the Greenland Ice Sheet, *Science*, *282*, 268–271.

Dansgaard, W., et al. (1993), Evidence for general instability of past climate from a 250-kyr ice-core record, *Nature*, *364*, 218–220.

Dean, W. E., R. M. Forester, and J. P. Bradbury (2002), Early Holocene change in atmospheric circulation in the Northern Great Plains: An upstream view of the 8.2 ka cold event, *Quat. Sci. Rev.*, *21*, 1763–1775.

deMenocal, P., J. Ortiz, T. Guilderson, and M. Sarnthein (2000), Coherent high- and low-latitude climate variability during the Holocene warm period, *Science*, *288*, 2199–2202.

Fanning, A., and A. J. Weaver (1997), Temporal-geographical meltwater influences on the North Atlantic conveyer: Implications for the Younger Dryas, *Paleoceanography*, *12*, 307–320.

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

Ganopolski, A., and S. Rahmstorf (2001b), Stability and variability of the thermohaline circulation in the past and future: A study with a coupled model of intermediate complexity, in *The Oceans and Rapid Climate Change: Past, Present, and Future*, *Geophys. Monogr. Ser.*, vol. 126, edited by D. Seidov, B. J. Haupt, and M. Maslinpp, pp. 261–275, AGU, Washington, D. C.

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

Gasse, F. (2000), Hydrological changes in the African tropics since the Last Glacial Maximum, *Quat. Sci. Rev.*, *19*, 189–211.

Goosse, H., H. Renssen, F. M. Selten, R. J. Haarsma, and J. D. Opsteegh (2002), Potential causes of abrupt climate events: A numerical study with a three-dimensional climate model, *Geophys. Res. Lett.*, *29*(18), 1860, doi:10.1029/2002GL014993.

Hu, F. S., D. Slawinski, H. E. Wright Jr., E. Ito, R. G. Johnson, K. R. Kelts, R. F. McEwan, and A. Boedigheimer (1999), Abrupt changes in North American climate during early Holocene times, *Nature*, *400*, 437–443.

Johnsen, S. J., D. Dahl-Jensen, W. Dansgaard, and N. Gundestrup (1995), Greenland palaeotemperatures derived from GRIP bore hole temperature and ice core isotope profiles, *Tellus, Ser. B*, *47*, 624–629.

Johnsen, S. J., et al. (1997), The $\delta^{18}O$ record along the Greenland Ice Core Project deep ice core and the problem of possible Eemian climatic instability, *J. Geophys. Res.*, *102*(C12), 26,397–26,411.

Johnsen, S. J., D. Dahl-Jensen, N. Gundestrup, J. P. Steffensen, H. B. Clausen, H. Miller, V. Masson-Delmotte, E. Sveinbjrnsdttir, and J. White (2001), Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: Camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP, *J. Quat. Sci.*, *16*, 299–307.

Jouzel, J., et al. (1997), Validity of the temperature reconstructions from water isotopes in ice cores, *J. Geophys. Res.*, *102*, 26,471–26,487.

Klitgaard-Kristensen, D., H. P. Sejrup, H. Haflidason, S. Johnsen, and M. Spurk (1998), A regional 8200 cal. yr BP cooling event in northwest Europe, induced by final stages of the Laurentide ice-sheet deglaciation?, *J. Quat. Sci.*, *13*(2), 165–169.

Korhola, A., K. Vasko, H. T. T. Toivonen, and H. Olander (2002), Holocene temperature changes in northern Fennoscandia reconstructed from chironomids using Bayesian modelling, *Quat. Sci. Rev.*, *21*, 1841–1860.

Leuenberger, M. C., C. Lang, and J. Schwander (1999), Delta ^{15}N measurements as a calibration tool for the paleothermometer and gas-ice age differences: A case study for the 8200 B. P. event on GRIP ice, *J. Geophys. Res.*, *104*, 22,163–22,170.

Leverington, D. W., J. D. Mann, and J. T. Teller (2002), Changes in the bathymetry and volume of glacial Lake Agassiz between 9200 and 7700 ^{14}C yr B. P., *Quat. Res.*, *57*, 244–252.

Licciardi, J. M., P. U. Clark, J. W. Jenson, and D. R. MacAyeal (1998), Deglaciation of soft-bedded Laurentide Ice Sheet, *Quat. Sci. Rev.*, *17*, 427–448.

Magny, M., C. Begeot, J. Guiot, and O. Peyron (2003), Contrasting patterns of hydrological changes in Europe in response to Holocene climate cooling phases, *Quat. Sci. Rev.*, *22*, 1589–1596.

Maier-Reimer, E., and U. Mikolajewicz (1989), Experiments with an OGCM on the cause of the Younger Dryas, in *Oceanography*, edited by A. Ayala-Castanares, W. Wooster, and A. Yanez-Arancibia, pp. 87–100, UNAM Press, Mexico City.

Manabe, S., and R. J. 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: Comparison to Younger Dryas event, *Paleoceanography*, *12*(2), 321–336.

Marshall, S. J., and G. K. C. Clarke (1999), Modeling North American freshwater runoff through the last glacial cycle, *Quat. Res.*, *52*, 300–315.

Masson, V., et al. (2000), Holocene climate variability in Antarctica based on 11 ice-core isotopic records, *Quat. Res.*, *54*, 348–358.

Muscheler, R., J. Beer, and B. Kromer (2003), Long-term climate variations and solar effects, in *Solar Variability as an Input to the Earth's Environment*, edited by A. Wilson, pp. 305–316, Eur. Space Ag., Paris.

Neff, U., S. J. Burns, A. Mangini, M. Mudelsee, D. Fleitmann, and A. Matter (2001), Strong coherence between solar variability and the monsoon in Oman between 9 and 6 kyr ago, *Nature*, *411*, 290–293.

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.

Rahmstorf, S. (1995a), Bifurcations of the Atlantic thermohaline circulation in response to changes in the hydrological cycle, *Nature*, *378*, 145–149.

Rahmstorf, S. (1995b), Multiple convection patterns and thermohaline flow in an idealized OGCM, *J. Clim.*, *8*, 3028–3039.

Rahmstorf, S. (2002), Ocean circulation and climate during the past 12,0000 years, *Nature*, *419*, 207–214.

Raynaud, D., J.-M. Barnola, J. Chappellaz, T. Blunier, A. Indermühle, and B. Stauffer (2000), The ice record of greenhouse gases: A view in the context of future changes, *Quat. Sci. Rev.*, *19*, 9–17.

Renssen, H., H. Goosse, T. Fichefet, and J.-M. Campin (2001), The 8. 2 kyr BP event simulated by a global atmosphere-sea-ice-ocean model, *Geophys. Res. Lett.*, *28*(8), 1567–1570.

Renssen, H., H. Goosse, and T. Fichefet (2002), Modeling the effect of freshwater pulses on the early Holocene climate: The influence of high-frequency climate variability, *Paleocean ography*, *17*(2), 1020, doi:10.1029/2001PA000649.

Shuman, B., P. Bartlein, N. Logar, P. Newby, and T. Webb II (2002), Parallel climate and vegetation responses to the early Holocene collapse of the Laurentide Ice Sheet, *Quat. Sci. Rev.*, *21*, 1793–1805.

Spooner, I., M. S. V. Douglas, and L. Terrusi (2002), Multiproxy evidence of an early Holocene (8.2 kyr) climate oscillation in central Nova Scotia, Canada, *J. Quat. Sci.*, *17*(7), 639–645.

Stocker, T. F., and D. G. Wright (1991), Rapid transitions of the ocean's deep circulation induced by changes in surface water fluxes, *Nature*, *351*, 729–732.

Teller, J. T., D. W. Leverington, and J. D. Mann (2002), Freshwater outbursts to the oceans from glacial Lake Agassiz and their role in climate change during the last deglaciation, *Quat. Sci. Rev.*, *21*, 879–887.

Thompson, L. G., T. Yao, M. E. Davis, K. A. Henderson, E. Mosley-Thompson, P.-N. Lin, J. Beer, H.-A. Synal, J. Cole-Dai, and J. F. Bolzan (1997), Tropical climate instability: The last glacial cycle from a Qinghai-Tibetian ice core, *Science*, *276*, 1821–1825.

Thompson, L. G., et al. (1998), A 25,000-year tropical climate history from Bolivian ice cores, *Science*, *282*, 1858–1864.
Thompson, L. G., et al. (2002), Kilimanjaro ice core records: Evidence of Holocene climate change in tropical Africa, *Science*, *298*, 589–593.
Tinner, W., and A. F. Lotter (2001), Central European vegetation response to abrupt climate change at 8.2 ka, *Geology*, *29*(6), 551–554.
van Geel, B., O. M. Raspopov, H. Renssen, J. van der Plicht, V. A. Dergachev, and H. A. J. Meijer (1999), The role of solar forcing upon climate change, *Quat. Sci. Rev.*, *18*, 331–338.
von Grafenstein, U., H. Erlenkeuser, J. Mller, J. Jouzel, and S. Johnsen (1998), The cold event 8200 years ago documented in oxygen isotope records of precipitation in Europe and Greenland, *Clim. Dyn.*, *14*, 73–81.
von Grafenstein, U., H. Erlenkeuser, A. Brauer, J. Jouzel, and S. Johnsen (1999), A mid-European decadal isotope-climate record from 15,500 to 5000 years B. P., *Science*, *284*, 1654–1657.
Wright, D. G., and T. F. Stocker (1993), Younger Dryas experiments, in *Ice in the Climate System*, edited by W. R. Peltier, pp. 395–416, Springer-Verlag, New York.
Yu, Z., and U. Eicher (1998), Abrupt climate oscillations during the last deglaciation in central North America, *Science*, *282*, 2235–2238.

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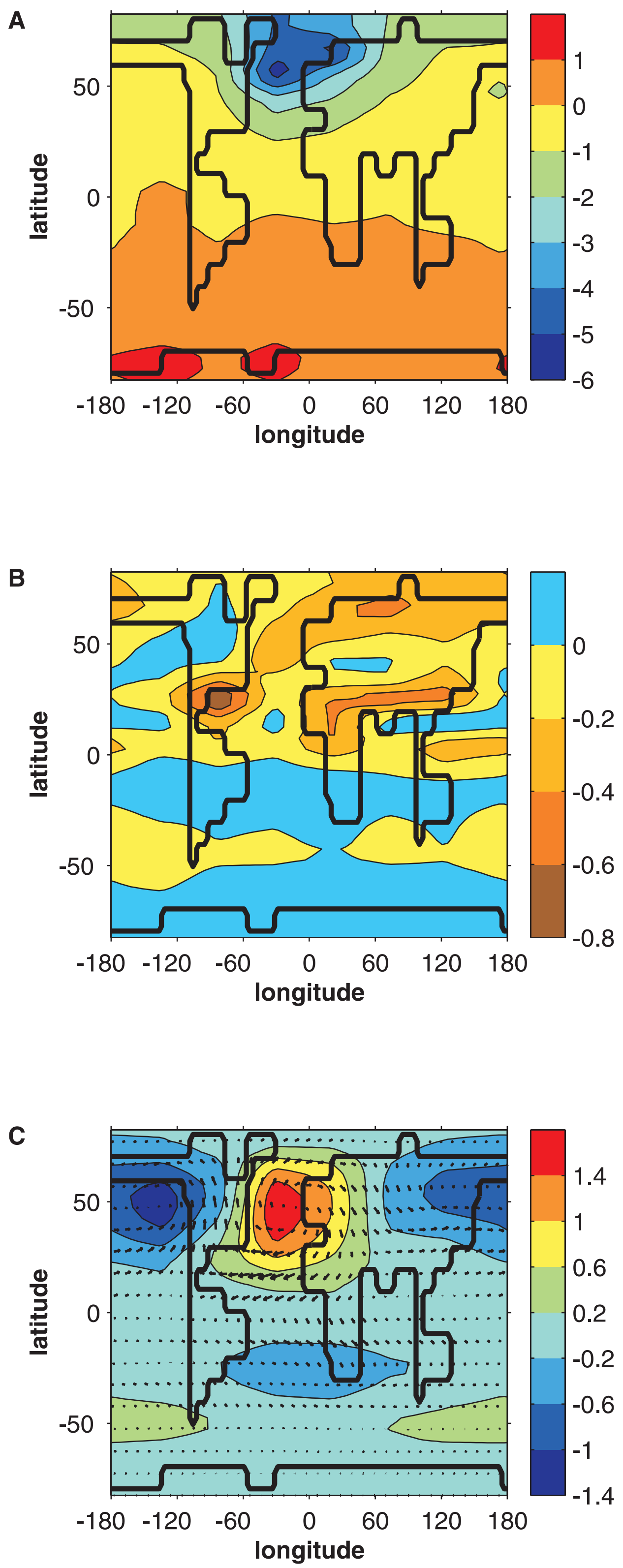

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