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The 8.2 kyr BP event simulated by a global atmosphere–sea-ice–ocean model

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Abstract. Seven freshwater perturbation experiments were performed with a global atmosphere–sea-ice–ocean model to study the mechanism behind the 8.2 kyr BP Holocene cooling event. These experiments differed in initial state and duration of the applied freshwater pulse, while the amount of freshwater was kept constant (4.67x10^{14} m^3). One of the scenarios, with freshwater added to the Labrador Sea at a rate of 0.75 Sv during 20 years, resulted in weakening of the North Atlantic thermohaline circulation during 320 years and surface cooling varying from 1 to 5°C over adjacent continents. This result is consistent with proxy data, suggesting that a meltwater-induced weakening of the thermohaline circulation caused the event. Moreover, our results indicate that the time-scale of the meltwater release and the initial state are important, as both have a strong effect on the magnitude and duration of the produced model response.

1. Introduction

The most distinct Holocene climatic event – with a duration of 300 to 400 years – occurred around 8.2 kyr BP (thousand calendar years before present) [*Alley et al.,* 1997]. This '8.2 kyr event' is clearly registered as a cooling in high resolution proxy data around the North Atlantic region, such as Greenland ice cores [*Alley et al.,* 1997], ocean sediments [*Bond et al.,* 1997], European lake cores [*von Grafenstein et al.,* 1998] and tree-ring records [*Klitgaard-Kristensen et al.,* 1998]. In addition, African lake level data indicate a strong reduction in precipitation [e.g., *Gasse,* 2000]. Although still under debate [see, e.g., *Hu et al.,* 1999], most researchers favor the hypothesis that the 8.2 kyr event has been caused by a perturbation of the North Atlantic thermohaline circulation (THC) by a catastrophic release of meltwater associated with the final stages of deglaciation of the Laurentide ice sheet. The estimates for the involved amount of freshwater range from 1.2x10^{14} m^3 [*Hu et al.,* 1999] to 5x10^{14} m^3 [*von Grafenstein et al.,* 1998]. The duration of the meltwater pulse is relatively uncertain, as estimates vary from 1 to 500 yr [e.g., *Klitgaard-Kristensen et al.,*1998; *Barber et al.,* 1999]. Our objective is to study the mechanism behind the 8.2 kyr event by analyzing transient freshwater perturbation experiments performed with a global atmosphere–sea-ice–ocean model. In the perturbation experiments, the duration of the meltwater pulse was varied, as considerable uncertainty on the involved time span exists. The total amount of freshwater released was fixed. The results are compared with available proxy data.

Numerous model studies have suggested that the THC is very sensitive to relatively small changes in the North Atlantic freshwater budget. For instance, experiments with coupled ocean–atmosphere models have revealed that freshwater releases to the North Atlantic may result in a temporary century-scale weakening of the THC [e.g., *Manabe and Stouffer,* 1997; *Fanning and Weaver,* 1997]. The resulting pattern of surface cooling shows resemblance to proxy evidence of past climatic events such as the Younger Dryas (~12.7-11.5 kyr BP) [e.g., *Manabe and Stouffer,* 1997; *Mikolajewicz et al.,* 1997]. However, so far, none of these simulations have been started from a climate in equilibrium with past boundary conditions (e.g., continental ice sheets, insolation, vegetation, atmospheric CO_2 concentration) and thus these experiments must be considered as sensitivity studies. Clearly, this hampers evaluation of model results with proxy data. In the simulations presented here, we aim to reproduce the 8.2 kyr event as precisely as possible and therefore, we have adjusted boundary conditions to the 8.5 kyr BP state.

2. Model and experimental design

We performed our experiments with the ECBILT-CLIO three-dimensional global atmosphere–sea-ice–ocean model [*Goosse et al.,* 2000]. The atmospheric part is version 2 of ECBILT, a spectral T21, three-level, quasi-geostrophic model described in detail by *Opsteegh et al.* [1998]. This model includes a full hydrological cycle and simple parameterizations of diabatic heating processes. As an extension to the quasi-geostrophic equations, an estimate of the neglected terms in the vorticity and thermodynamic equations is incorporated as a temporally and spatially varying forcing. The sea-ice–ocean component is the CLIO model, consisting of a primitive-equation, free-surface ocean general circulation model (OGCM) coupled to a comprehensive thermodynamic–dynamic sea-ice model [*Goosse and Fichefet,* 1999]. The OGCM contains a detailed formulation of boundary layer mixing [*Goosse et al.,* 1999] and a parameterization of density-driven downslope flows [*Campin and Goosse,* 1999]. The horizontal resolution of CLIO is 3° by 3°, and there are 20 unequally spaced levels along the vertical. Under present-day forcing conditions, ECBILT-CLIO systematically overestimates precipitation over the North Atlantic and Arctic Oceans. To overcome associated problems (e.g., a too slow THC), precipitation is artificially reduced over the Atlantic and Arctic Oceans, after which the excess water is added to the Pacific Ocean. With this sole flux correction, the model produces a control climate

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that is generally comparable to that of coupled GCMs [*Goosse et al.*, 2000].

As a first step, the model was run with 8.5 kyr BP boundary conditions until it reached a quasi-equilibrium state (i.e., after 550 years). The following changes in boundary conditions were applied compared to a modern climate simulation. Insolation was adjusted according to *Berger* [1978] and atmospheric concentrations of greenhouse gases were lowered following *Raynaud* [2000] (i.e., CO_2 from 353 to 261 ppmv, CH_4 from 1720 to 650 ppbv and N_2O from 310 to 270 ppbv). Moreover, the surface albedo was changed to account for global vegetation changes [*Adams*, 1997] and the altitude of some grid cells was increased at the location of the Laurentide ice sheet [*Peltier*, 1994]. For the latter adjustment, we assumed an ice-sheet configuration in which the Hudson Bay was already deglaciated. The 8.5 kyr BP quasi-equilibrium state is characterised by increased seasonality compared to the modern steady simulated climate over most continents, most notably over Eurasia and Northern Africa. Compared to a modern control climate, temperatures are generally higher during boreal summer (up to 5°C difference), and lower during boreal winter (up to 2°C difference). Moreover, compared to today, precipitation increases over Northern Africa and parts of Asia due to a strengthened summer monsoon. These changes reflect mainly the influence of the different insolation and are in general agreement with proxy records for the early Holocene. The strength of the THC is similar in the modern and 8.5 kyr BP states, as in both cases, North Atlantic Deep Water is exported from the Atlantic basin (at 20°S) at a mean rate of 15 Sv (1 Sv = 10^6 m^3s^{-1}).

Second, the 8.5 kyr BP quasi-equilibrium model state was perturbed (at t=550) in four experiments by adding a fixed amount of $4.67x10^{14}m^3$ freshwater to the Labrador Sea at the following linear rates: a) 1.5 Sv during 10 years, b) 0.75 Sv during 20 years, c) 0.3 Sv during 50 years and d) 0.03 Sv during 500 years. The quantity of $4.67x10^{14}m^3$ freshwater lies in the higher range of the published estimates based on proxy data [e.g., *von Grafenstein et al.*, 1998].

3. Effect of freshwater perturbations and comparison with proxy data

In our experiments, the response of the THC to prescribed freshwater forcing was clearly dependent on the perturbation time-scale (see Fig. 1a). A pulse distributed over 500 years (rate is 0.03 Sv) produced only a slight response, as the maximum value of the overturning streamfunction in the Nordic Seas was reduced from 17 to 16 Sv. This slight response was expected, as 0.03 Sv is considerably smaller than the rate required to significantly weaken the (present-day) THC in other model studies [e.g., 0.06 Sv in *Rahmstorf*, 1995]. In contrast, the pulses distributed over 10, 20 and 50 yr resulted in a substantial, event-like weakening of the Nordic Seas meridional overturning (down to 3 Sv). Moreover, in the case of the 10 yr perturbation, the THC switched permanently (i.e., for at least 1000 years) into a different state that is characterized by a significantly reduced overturning (mean in Nordic Seas of 5 Sv) and high variability at a decadal time-scale (Fig. 1a). In the experiments with pulses of 50 and 20 yr duration, however, the model recovers to its pre-perturbed state of 17 Sv after 145 and 320 years, respectively. Consequently, the time-scale of the 20-year perturbation experiment corresponds to the duration of the 8.2 kyr BP event suggested by proxy data (i.e., 300 to 400 yr).

To investigate the effect of initial state, we repeated the 10, 20 and 50 yr freshwater perturbations to the same 8.5 kyr BP quasi-equilibrium state, but at a different moment in time (i.e., t=555 instead of t=550). In the repeated 10 and 50 yr perturbation experiments, the model showed responses similar to the ones described above (Fig 1b). However, in the 20 yr case started at t=555, the model's THC did not recover as in the 20 yr perturbation started at t=550 and remained for at least 600 years in a perturbed, weakened state similar to that of both 10 yr cases (compare Figs. 1a and 1b). Evidently, the 20 yr perturbations bring the model's THC close to a threshold between two states and then conditions at the time of the perturbation determine if the THC recovers or not. A

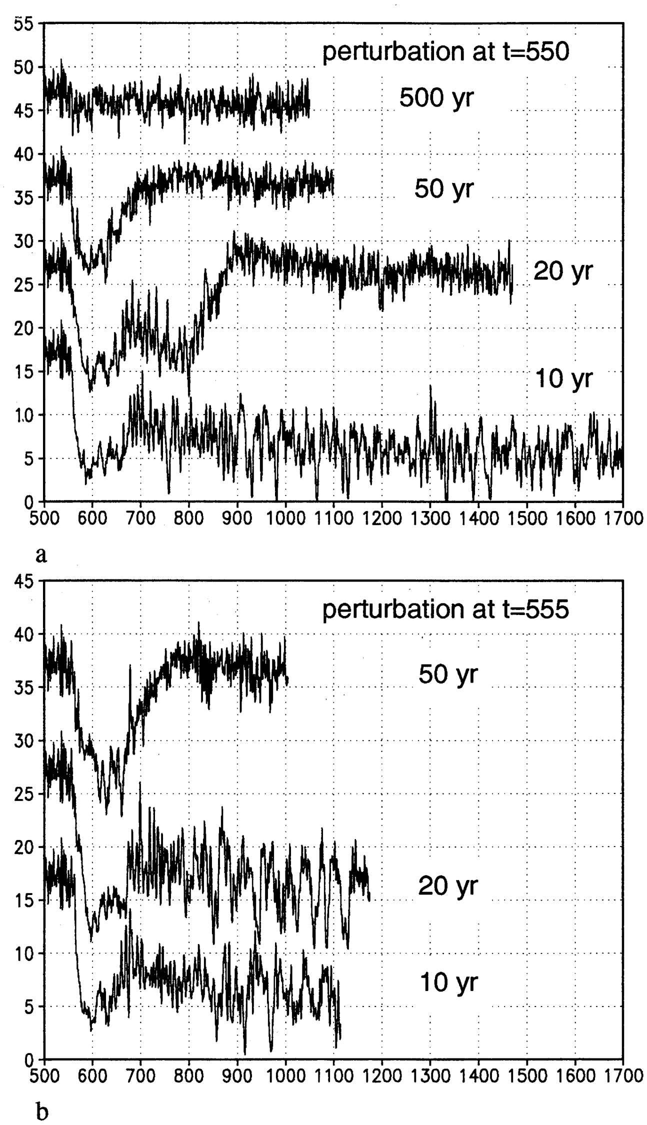

Figure 1a-b. Annual mean value of the maximum meridional overturning streamfunction (Sv) in the Nordic Seas for seven perturbation experiments plotted against time (years). The freshwater pulse ($4.67x10^{14}m^3$) is distributed over varying periods as signified in the figure and starts at t=550 (Fig. 1a) and t=555 (Fig. 1b). Note that, for convenience, the values of the 20 yr, 50 yr and 500 yr (only Fig. 1a) cases have been elevated by 10, 20 and 30 Sv, respectively.

detailed analysis of this effect is beyond the scope of this paper and will be studied further in future experiments.

In addition to the THC weakening, the perturbed state is characterized by a southward shift of the main convection site south of Svalbard (75°N) to near the Norwegian coast (67°N) (see Fig. 2a-b). It should be noted that, in contrast to modern observations, in a simulation of present-day climate no significant convection occurs in the Labrador Sea region. The THC weakening and southward shift in convection results in a substantial increase in sea-ice coverage compared to the 8.5 kyr BP quasi-equilibrium state, with most of the Nordic Seas and the Denmark Strait becoming perennially ice covered. In addition, the annual mean sea ice volume in the Northern Hemisphere increases from $16x10^3$ km^3 to $23x10^3$ km^3.

As expected, weakening of convection and expansion of the sea-ice cover cause a considerable cooling of the lower atmosphere over the Nordic Seas and adjacent landmasses (Fig. 2c). Over the Nordic Seas, the surface cooling amounts to 5 to 10°C compared to the 8.5 kyr BP quasi-equilibrium state, with a maximum temperature depression between Svalbard and northern Norway (i.e., the main convection site in Fig. 2a which is ice-covered in Fig. 2b). Over Greenland, a surface cooling of 2 to 5°C and a 30% decrease in precipitation are noted in the model results (Fig. 2c). These values are consistent with the 3 to 8°C annual mean cooling and large drop in accumulation reconstructed for the 8.2 kyr event in Greenland ice cores [*Alley et al.*, 1997; *von Grafenstein et al.*, 1999]. In addition, in northwestern Europe, the simulated surface temperature decrease is 1 to 2°C. Again, this is in agreement with proxy evidence, as a cooling of about 2°C have been estimated for the North Sea and Germany [e.g., *Klitgaard-Kristensen et al.*, 1998; *von Grafenstein et al.*, 1999]. Furthermore, in the eastern North Atlantic Ocean, the model produces progressive surface cooling towards the northeast, from 1°C at 45°N near the European continent to 2°C at 60°N. This result compares well with the study of *Bond et al.* [1997], who inferred that ocean surface cooling in the eastern North Atlantic probably did not exceed 2°C during the 8.2 kyr event. Over the western North Atlantic Ocean, the surface cooling in the model is less pronounced. This result is, without doubt, strongly influenced by the absence of convection in the Labrador Sea in the model. Over North Africa, a 10% decrease in annual mean precipitation is simulated (not shown), which is consistent with the noted sudden fall in lake levels throughout this region [e.g., *Gasse*, 2000]. In the Southern Hemisphere the simulated changes in surface temperature are minor. Over the South Atlantic Ocean a slight increase in sea surface temperature (less than 1°C) is noted, which is due to an accumulation of heat, caused by a considerable decrease in the northward heat flux in the Atlantic (from ~$0.28x10^{15}$ to $0.15x10^{15}$ W at 20°S).

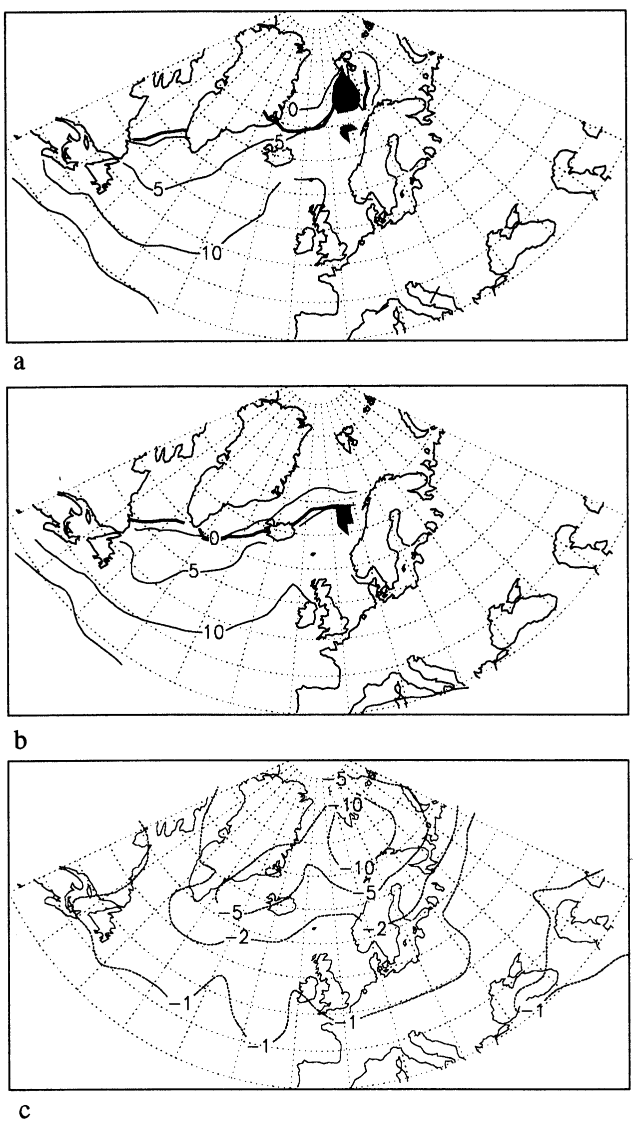

Figure 2a-c. 2a: Mean January sea surface temperatures (°C, thin lines), maximum sea-ice cover (thick line) and main convection sites (shaded) for the 8.5 kyr BP equilibrium state (years 500 to 550). 2b: As 2a, but for perturbed state (years 620 to 670 of 20 yr case). 2c: Annual mean surface temperature difference (°C) between two states shown in Figures 2a and 2b.

4. Concluding remarks

We performed seven freshwater perturbation experiments with a coupled atmosphere–sea-ice–ocean model to investigate the mechanism behind the Holocene cool event centered at 8.2 kyr BP. Our results show the following:

1. In our model, it is possible to reproduce the 8.2 kyr BP event with a freshwater perturbation. When applied over 20 years, the freshwater pulse produced in one case a model response (i.e., atmospheric cooling) that is consistent with available proxy data. This is true for the length of the simulated event (320 years) and the magnitude of the cooling around the North Atlantic basin (up to 10°C). Consequently, our results support the hypothesis that the 8.2 kyr event has been caused by a freshwater-induced THC weakening, associated with the final stages of deglaciation in North America.
2. THC weakening is sensitive to the time-scale of freshwater perturbations. In our simulations, only one scenario with a 20 yr pulse produced a model response in agreement with proxy data. A more gradual release (0.03 Sv during 500 yr) as proposed by *Klitgaard-Kristensen et al.* [1998] resulted in no significant THC weakening and surface cooling, whereas a sudden and strong freshwater pulse (10 yr) caused a permanent and significant THC weakening

during at least 1000 years. It should be noted that this response is probably model dependent.

3. When brought close to a threshold between two THC states, the model response depends also on conditions at the time of the perturbation. Two 20 yr perturbations applied at different times to the same quasi-equilibrium early Holocene model state, produced different responses. In one case, the model recovered after 320 years, whereas in another case, the THC remained in a weakened state. Further model experiments are required to study the effect of initial conditions.

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