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A simulation of the Atlantic meridional circulation during Heinrich event 4 using reconstructed sea surface temperatures and salinities

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Abstract. With a simplified two-dimensionnal Atlantic Ocean model, we investigate the response of the deep ocean circulation to reconstructed sea surface salinity and temperature used as surface boundary conditions at two time slices: just before and during Heinrich event 4. In contrast to previous studies, we do not make any assumption about freshwater input. Our model results suggest that the recorded estimations of surface hydrological changes during an Heinrich event are able to induce a drastic slowdown, or even a collapse, of the overall Atlantic thermohaline circulation. This is, to some extent, consistent with records of the deep sea ventilation at these times.

1. Introduction

During the last glacial period, the climate of the Earth has undergone many abrupt changes. The first evidence of such rapid changes was found in the isotopic records of past precipitations above Greenland [*Dansgaard et al.*, 1982, 1993]. These climatic excursions, with typical durations between 1000 and 3000 years, are named Dansgaard-Oeschger events. In deep sea marine sediments from the North Atlantic Ocean, six rapid events have been also discovered during the last glacial period [*Bond et al.*, 1992; *Heinrich*, 1988]. These so-called Heinrich events are characterized by an abrupt increase in the ice-rafted debris (lithic material larger than 150 μm) found in the sediment and correspond to massive iceberg discharges in the North Atlantic between 40°N and 55°N. These Heinrich events occur every 7 to 10 kyr during glacial periods, and their duration is of the order of 1000 years [*François and Bacon*, 1994]. The interpretation of these Dansgaard-Oeschger and Heinrich events as large-scale and large-amplitude climatic changes came with the study of high-resolution marine records of sea surface conditions [*Bond et al.*, 1993; *Cortijo et al.*, 1995]. A remarkable correlation was indeed found between sea surface temperature proxies and each isotopic excursion found in the Greenland records, or Dansgaard-Oeschger event. The Heinrich events are associated with the coldest time periods, just preceding the largest increases in temperature.

Unlike the glacial-interglacial changes, these rapid climatic shifts are not associated with variations in the orbital parameters of the Earth. *Broecker et al.* [1985] were among the first to promote the idea that the ocean thermohaline circulation might have played a crucial role. Many numerical experiments have since shown that this hypothesis was sound and that the ocean circulation can indeed switch abruptly between different equilibria [*Bryan*, 1986; *Rahmstorf*, 1995; *Stocker and Wright*, 1991] or can even experience self-sustained oscillations [*Quon and Ghil*, 1995; *Sakai and Peltier*, 1996; *Winton and Sarachik*, 1993]. Numerical models of ice sheets have also demonstrated the possibility of ice sheet oscillations [*Oerlemans*, 1983], and such a mechanism was suggested [*MacAyeal*, 1993] to explain the iceberg discharges associated with Heinrich events. The first tentative modeling experiments coupling the ice sheet dynamics with the ocean-atmosphere system have since been performed [*Paillard*, 1995; *Paillard and Labeyrie*, 1994]. If the mechanism of the Dansgaard-Oeschger oscillations is still a matter of debate, there is now a wide acceptance that the iceberg discharge during Heinrich events induces a slowdown or even a complete disruption of the North Atlantic thermohaline circulation, thus greatly reducing the northward oceanic heat transport and cooling the high northern latitudes.

To address this question, many numerical experiments have been performed with different ocean models, assuming some kind of large additional freshwater input at high northern latitudes in the North Atlantic Ocean. If these experiments have clearly demonstrated the sensitivity of the thermohaline circulation to freshwater forcing, none of them can be taken as even close to what really happened during a Heinrich event. Indeed, most of these experiments were started from a present-day ocean circulation state. However, during glacial times, when Heinrich events occurred, the oceanic deep circulation was different from the present one. Furthermore, these experiments assumed very different freshwater forcings, ranging in intensity from 0.01 Sv to 1 Sv, in duration from 10 years to thousands of years, and in geographic locations from Labrador Sea to some North Atlantic latitude band [*Maier-Raimer and Mikolajewicz*, 1989; *Manabe and Stouffer*, 1995; *Stocker and Wright*, 1996; *Mikolajewicz et al.*, 1997]. Besides, these experiments were all performed with oceanic mixed boundary conditions or simplified coupled ocean-atmosphere models. This has led to criticism and skepticism toward these results [*Zhang et al.*, 1993], since such model configurations are far less stable than the more classical oceanic restoring boundary conditions.

In order to better represent what really happened during an Heinrich event, and since we are not interested in further studying the stability properties of the ocean, we used an

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ocean model with restoring boundary conditions. In addition to being much more stable, this also allows us to avoid making any assumption about the freshwater input intensity or location. For this purpose, we need reconstructions of the sea surface temperature (SST) and salinity (SSS) fields during some particular Heinrich event. Such a reconstruction was made for Heinrich event 4 (HE 4) [*Cortijo et al.*, 1997]. This event occurred at about 35 ka. It was chosen for such a reconstruction because it is well marked in all North Atlantic marine sediment cores, which is not the case for HE 3. In contrast to HE 1 and HE 2, it occurs at a time of nearly constant ice volume and nearly constant insolation forcing: The climatic variations associated with HE 4 are therefore not linked to the Milankovitch theory and glaciation cycles but result solely from internal (ice sheet or oceanic) variability. The reconstructed temperatures during the event are 0°C to 4°C cooler than they were just before the event, while the salinities are similarly lowered by 0‰ to 2 or 3‰. The HE 4 sea surface reconstruction extends from 40°N to 60°N in the North Atlantic, which encompasses the area where the iceberg discharge occurs. The largest variations are observed in the iceberg melting latitudinal band, between 40°N and 50°N, while sites above 50°N did not experience significant temperature and salinity changes. Most importantly, this reconstruction reveals a coherent spatial pattern for SST and SSS changes and gives us confidence to use them as an input of an oceanic model.

Though a more global data set would be needed to perform precise three-dimensional oceanic modeling experiments, the available data can still be used with a simplified ocean model in order to obtain the main characteristics of the ocean circulation at this time. The simplified two-dimensional ocean model that we are using will also permit a larger number of experiments in order to assess the robustness of our results.

2. Model Description and Control Experiment

The Atlantic and Southern Oceans have a prevalent role. The relative amount of deep water formed in the south and in the north determines the characteristics of the deep waters in the rest of the world ocean. More importantly, the heat transported northward in the Atlantic is directly linked to the intensity of North Atlantic Deep Water (NADW). Though the global oceanic circulation may have been entirely different in the remote past, there is no evidence in the late Quaternary that deep water formation occurred anywhere but in the North Atlantic and around Antarctica, just as nowadays. Furthermore, the reconstructions of past temperatures and salinities, during the Last Glacial Maximum as well as during HE 4, are concentrated in the Atlantic basin. It seems therefore natural to concentrate our modeling only on the Atlantic sector of the global ocean. Our simplified two-dimensional ocean model is a sector extending from 80°S to 90°N and 70° wide in longitude. The bottom is flat at 5000 m depth.

Two-dimensional ocean models have been used recently to investigate the stability of the thermohaline circulation [*Hovine and Fichefet*, 1994; *Marotzke et al.*, 1988; *Sakai and Peltier*, 1997; *Stocker et al.*, 1992; *Thual and McWilliams*, 1992; *Winton*, 1997; *Wright and Stocker*, 1992] and also to reconstruct the thermohaline circulation during the last glacial maximum [*Fichefet et al.*, 1994; *Ganopolski et al.*, 1998]. Though they differ in many respects, these previous experiments have proven the usefulness of two-dimensional models to represent the large-scale features of the oceanic thermohaline circulation. The model used here is very similar to the *Winton* [1997] model, where the meridional velocity is related to the meridional pressure gradient using a Rayleigh damping as a parameterization for the friction on meridional velocity. This is, to first order, equivalent to the more sophisticated Wright and Stocker approach [*Stocker et al.*, 1992; *Wright and Stocker*, 1992]. The Southern Ocean needs special attention. Since there are no east or west boundaries in this part of the ocean, at least in the top 2000 or 3000 m, there can be no zonal pressure gradient, and therefore the geostrophic meridional velocity must vanish. In the Wright and Stocker approach, this is taken into account by setting their "closure" parameter to zero in the top 3000 m between 55°S and 65°S. Equivalently, in our case, we simply set the Rayleigh coefficient r to infinity in this same part of the ocean. For convection, we use an enhanced vertical diffusion instead of a more classical convective adjustment scheme. This improves notably the numerical behavior of the model [*Schmidt and Mysak*, 1996].

We impose restoring surface boundary conditions for temperature and salinity using the relaxation time constants τ_T and τ_S and the observed surface temperatures and salinities. The numerical implementation uses standard integration techniques. The model has 12 vertical layers down to 5000 m, the top layer being 100 m deep. There are 21 latitude bands. The model equations are given in the appendix, and the numerical values for the parameters are listed in Table 1.

In order to assess the ability of the model to reproduce the main features of the thermohaline circulation, we first performed an experiment with present-day surface boundary conditions. We used the *Levitus* [1982] temperature and salinity fields, averaged zonally over the Atlantic Ocean, as the prescribed surface forcing for our control experiment. A crucial choice for such a two-dimensional model is the definition of the averaging area in the deep water formation regions. Indeed, if we choose to average zonally the Arctic Ocean surface conditions, the resulting prescribed salinities in the high northern latitudes of the model will be much too low, because of the low salinities in the Bering Strait area of the Arctic Basin. In such a situation, the model cannot form deep waters in the North Atlantic. It makes much more sense, from an oceanographic point of view, to use only the Norwegian Sea sector of the Arctic Basin as our prescribed boundary

Table 1. Numerical Values of the Model Parameters

Parameter	Value	Unit
N_{min}^2	10^{-7}	s^{-2}
τ_C	30	days
r	10^{-4}	s^{-1}
κ_V	10^{-4}	$m^2.s^{-1}$
κ_H	100	$m^2.s^{-1}$
τ_T	30	days
τ_S	100	days

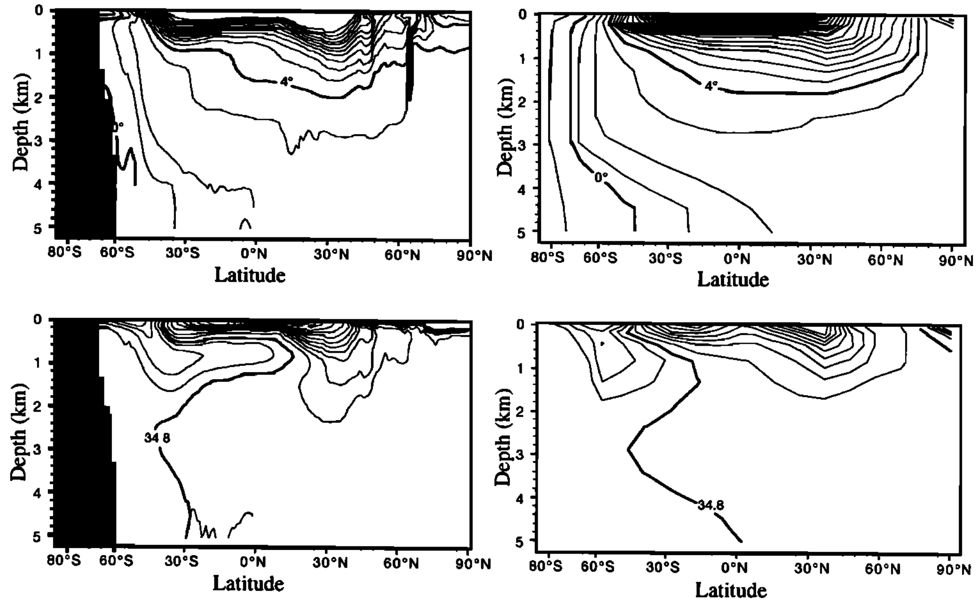

Figure 1. (top) temperature and (bottom) salinity annual mean fields in the model for the control run (on the right) compared to the *Levitus* [1982] climatology (on the left). Contours are every 1°C for temperature and every 0.2‰ for salinity.

conditions. In this area (north of 60°N), we therefore used only the 60°W-60°E sector of the ocean.

Similarly, Antarctic Bottom Waters (AABW) are formed in extremely narrow regions, on the east side of the Antarctic peninsula in the west part of the Weddell Sea, and also in the west part of the Ross Sea. Using a global average of temperatures and salinities all around Antarctica completely overrepresents regions were no deep water formation is possible. We therefore made our zonal averaging in only a 10° longitude band around the deep water formation areas. It must be emphasized that these somewhat arbitrary choices are crucial in determining the exact balance between NADW and AABW. This misrepresentation of the deep water formation regions is one of the major limitation of two-dimensional oceanic models.

Using the wind stress from the *Hellerman and Rosenstein* [1983] climatology, we have integrated the model from rest for at least 10,000 years, up to the equilibrium. The seasonal cycle is explicitly resolved, which is crucial for a good representation of the high-latitude convection areas. The resulting annual mean temperature and salinity fields are compared to the climatology in Figure 1. We note the good agreement between the modeled SST and SSS fields and the *Levitus* [1982] climatology. The annual mean overall meridional streamfunction is shown in Figure 2a: The NADW and AABW strengths are also in reasonable agreement with current estimates [*Macdonald and Wunsch*, 1996], as are their respective depths and extensions. The low-latitude wind-driven cells, in the upper thermocline, are superimposed on this general thermohaline circulation. The "no east-west boundary region" in the Southern Ocean is also driven by the upper wind stress, and the induced overturning contributes significantly to the interhemispheric water mass exchange [*Toggweiler and Samuels*, 1993].

3. Past SST and SSS Reconstruction Accuracy

Since we will investigate in the following the response of this model to paleoclimatic forcings, it is important to examine the accuracy of the available data. SST are obtained usually from faunal assemblages, and the standard statistical estimates give an error of about 1° to 2°C [*Prell*, 1985]. The sea surface salinities (SSS) are estimated using the isotopic composition of planktic foraminifera. A careful error analysis shows that the main sources of uncertainty for these SSS estimates are the SST values and, to a smaller extent, the slope of the δ^{18}O-salinity regression [*Cortijo et al.*, 1997; *Schmidt*, 1999]. In our study, we use two extreme cases for the δ^{18}O-salinity slope, and the errors arise mainly from the SST errors. Indeed, a 2°C error on SST may eventually lead to a 1‰ error on SSS, which is very important. However, it is then crucial to understand how this may affect density, since the thermohaline circulation is driven by density gradients, not directly by temperature or salinity gradients. The error analysis must therefore go beyond a simple examination of SST and SSS errors. In particular, these errors are not independent and cannot be added like random errors. On the contrary, they mostly cancel out when computing density. Indeed, the two statistically independant estimates are the δ^{18}O of calcite and the temperature. If we use a δ^{18}O-S slope of 2, we get the following differential:

$$d\delta^{18}O_{CaCO3} \approx d\delta^{18}O_{H2O} - 0.25\, dT \approx 0.5\, dS - 0.25\, dT.$$

The differential of density is given by $d\rho = \beta\, dS - \alpha\, dT$, with α and β being the thermal and haline expansion coefficients. This gives

$$d\rho \approx 2\,\beta\, d\delta^{18}O_{CaCO3} + (0.5\,\beta - \alpha)\, dT.$$

With $\beta \approx 0.8$ (‰-density/‰-salinity), $\alpha \approx 0.3$ (‰-density/°C) for warm waters, and $\alpha \approx 0.1$ (‰/°C) for cold waters, we get $\partial\rho/\partial T \approx 0.5\,\beta - \alpha \approx 0.1$‰/°C for warm waters and $\partial\rho/\partial T$

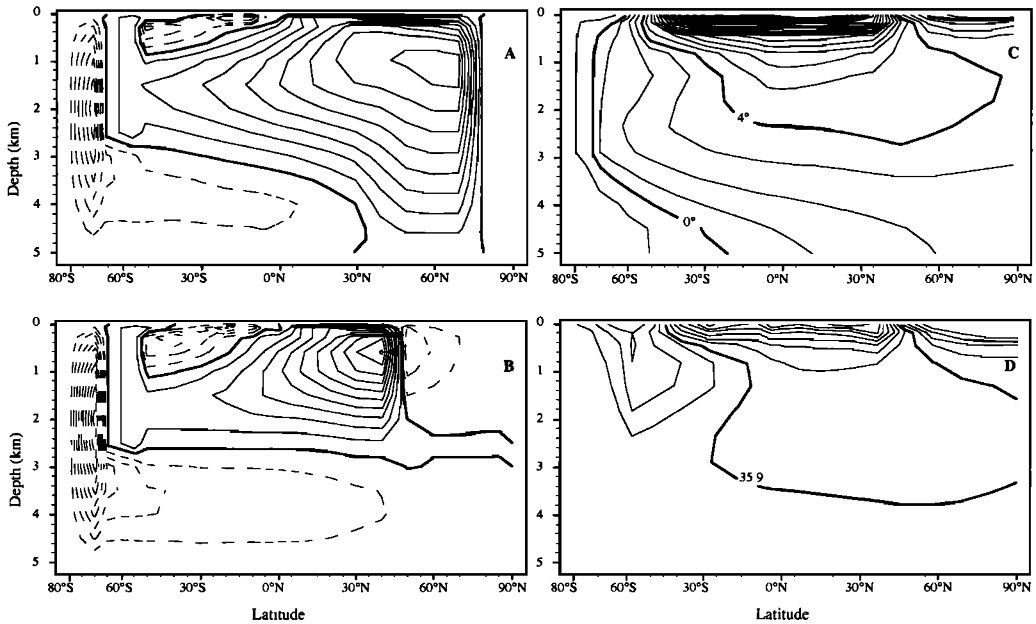

Figure 2. (a) Stream function obtained in the control run (contours every 2 Sv). The maximum of the North Atlantic Deep Water (NADW) overturning is 16.9 Sv at about 60°N, while the maximum of the Antarctic Bottom Water (AABW) is 13.2 Sv at 70°S. (b) The Last Glacial Maximum (LGM) experiment. The maximum of the NADW (or Glacial North Atlantic Intermediate Water (GNAIW)) overturning is 20.2 Sv at about 40°N, while the maximum of the AABW is 20.8 Sv always at 70°S. (c) Temperature and (d) salinity annual mean fields in the model for the LGM run (contours every 1°C and 0.2‰, respectively).

≈ 0.3‰/°C for cold waters. This means that for a 2°C error on the SST estimates, the corresponding density error will be between ≈ 0.2‰ and ≈ 0.6‰. This is considerably different from the independent accumulation of a 1‰ error on salinity with a 2°C error on temperature. In particular, if *CLIMAP* [1981] is wrong at low latitudes, this has only a minor effect on the estimation of the density of seawater, since it depends almost entirely on the estimation $\delta^{18}O$ of calcite alone, as also noticed by *Lynch-Stieglitz et al.* [1999]. If we use a $\delta^{18}O$-salinity slope of 1, the situation is even more favorable. Indeed, we get $\partial\rho/\partial T = 0.25\ \beta - \alpha \approx -0.1$‰/°C for warm waters and $\partial\rho/\partial T \approx +0.1$‰/°C for cold waters. For a 2°C error on the SST estimates, the corresponding density error will always remain below 0.2‰.

If the errors on temperature and salinity reconstructions may be quite large, up to 2°C and up to 1‰, respectively, in some cases, this translates into much smaller errors for the density of seawater, between 0.2‰ and 0.6‰ in the worst case. Since the thermohaline circulation results from large-scale density gradients, and since the spatial coverage provided by the paleodata shows coherent large-scale anomalies in SST and SSS, we believe it does make sense to use paleotemperature and paleosalinity reconstructions as input of an ocean model to estimate the impact of sea surface changes on the deep ocean circulation.

4. The Last Glacial Maximum Experiment

In order to investigate the role of temperature and salinity changes on the oceanic circulation during the Last Glacial Maximum (LGM), we performed an experiment where the SST and SSS fields are relaxed to their LGM values, as estimated by *CLIMAP* [1981] for the temperatures and by *Duplessy et al.* [1991] and *Wang et al.* [1995] for the salinities. In the Southern Ocean, where no salinity reconstruction is yet available, we used a SSS anomaly equal to the global shift of +1.1‰, resulting from the about 120 m overall sea level lowering. Since the overall surface wind patterns remain the same during the LGM and since the influence of the wind stress is mostly limited to the upper thermocline ocean, we think that, in contrast to SST and SSS changes, wind stress changes will not affect significantly the general thermohaline structure. We are therefore using the present-day wind stress climatology as a surface boundary condition for the glacial ocean. The model is then integrated for several thousand years, up to the equilibrium. The resulting oceanic deep circulation is presented in Figure 2b, while the corresponding temperatures and salinities in the ocean interior are shown in Figures 2c and 2d. The circulation changes in this experiment are essentially located in the northern Atlantic. The deep water formation occurs at lower latitudes, around 40°-50°N instead of 60°-70°N in the control run. The resulting NADW flow is also much shallower and remains above 3000 m, and therefore owes the new name of Glacial North Atlantic Intermediate Waters (GNAIW) [*Duplessy et al.*, 1988]. The Atlantic deep waters are therefore coming mainly from the south, even though the AABW flow remains almost unchanged at about 2 to 3 Sv. The ocean bottom waters cool by about 1°C as a result of the greater extent of the cold AABW contribution. For the same reason, the bottom waters are also fresher by about 0.1‰, in addition to the global +1.1‰ shift. Conversely, since the North Atlantic waters are sinking at lower latitudes, the temperatures around 2000 m are slightly warmer in the LGM experiment than during the control run.

The pattern of a shallower and southward shifted NADW (or GNAIW) flow is consistent with $\delta^{13}C$ and $\Delta^{14}C$ measurements on the shells of benthic foraminifera living at this time. Indeed, deep waters in the Atlantic Ocean are systematically depleted in ^{14}C and enriched in ^{13}C during the LGM, suggesting a reduced deep ventilation. However, the intermediate waters, above 2500 m, exhibit the opposite changes [*Duplessy et al.*, 1988], which indicates a possibly better than present intermediate water ventilation. Previous experiments with a two-dimensionnal (2-D) ocean model forced by reconstructed SST and SSS for the LGM [*Fichefet et al.*, 1994] have shown the same general southward shift of the deep water formation area and the same shallower sinking of these waters. However, in contrast to our results, the NADW (or GNAIW) strength was significantly weaker in the LGM run than in the control run. This probably results from slight differences in model configuration near the deep water formation areas. In particular, their top level is only 50 m deep instead of 100 m here. We believe 100 m is more representative of what the foraminiferal assemblages are able to record in terms of anomalous SST and SSS.

A coupled ocean-atmosphere simulation has also been performed for this time period [*Ganopolski et al.*, 1998], where a 2-D ocean was coupled to a simplified atmospheric circulation model. In this simulation where LGM SST and SSS fields are not used as input, the corresponding NADW strength is very similar to the control run. This result is quite comparable to our results. The corresponding depth and location of the deep water formation are also very similar to our findings. In contrast, another LGM simulation, using a 3-D ocean model coupled to an energy and moisture balance model [*Weaver et al.*, 1998], exhibits a much reduced and much deeper NADW formation, without significant changes in the location of the deep water formation area, together with an almost complete wanishing of AABW, in contradiction to paleoceanographic data. However, in contrast to the Ganopolski et al. model, water vapor is diffused, not advected, in the atmosphere. This may lead to quite unrealistic precipitation patterns and probably has some strong influence on the resulting thermohaline circulation.

Both the comparison of the control run with the climatology and the quite reasonable results obtained for the LGM experiment make us confident that our simplified ocean model is suitable for simulating the state of the Atlantic Ocean before and during a Heinrich event, using the available SST and SSS data.

5. The Heinrich Event 4 Experiments

The reconstructed SST and SSS anomalies [*Cortijo et al.*, 1997] are shown in Figure 3. The methodology used to obtain these fields is strictly similar to the one used for the LGM. In particular, it is possible to reconstruct both winter and summer SSTs and thus to have an idea of the seasonal cycle for temperature. However, the reconstructed SSS anomaly fields are assumed to be annual mean values, and the salinity seasonal cycle remains therefore identical to the present-day one. This may represent a significant limitation of the method, since the high-latitude convection is mainly seasonal and depends strongly on the surface salinity fields. Nevertheless,

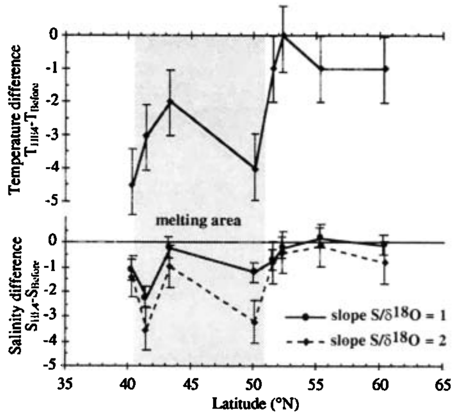

Figure 3. Summary of the Heinrich event 4 (HE 4) surface hydrological perturbation [from *Cortijo et al.*, 1997]. The differences in summer sea surface temperature and surface salinity during HE 4 and before the event are plotted here to emphasize the cooling (between 3° and 4°C) and the freshening (between 1 and 3‰, depending on the choice of $\delta^{18}O$-S slope) in the iceberg melting area.

for the HE 4 experiments, the major limitation comes from the very restricted area where such surface hydrological reconstructions are available. Indeed, the model needs a full SST and SSS coverage of the Atlantic Ocean as boundary conditions, whereas these fields are available only in the 40°N-60°N latitude band for time slices around HE 4. A major difficulty in extending this data coverage southward concerns the precise dating and correlation of paleoclimatic proxies accross the whole Atlantic Ocean. If such a correlation is possible in the vicinity of the iceberg melting area, with the clear identification of the ice-rafted debris defining Heinrich layers, the problem is much more difficult for more southern sediment cores.

We decided to use the available data together with the LGM boundary conditions in order to run the model. The following HE 4 experiments are therefore perturbation experiments, around the LGM state, to changes in the hydrological properties of the surface ocean near the iceberg melting area. The data are indeed available precisely in this area, where the icebergs associated with the Heinrich events are melting and depositing their debris to the sediment, but this latitude band is also the area where the LGM North Atlantic Deep Water (or GNAIW) forms. We can therefore experiment, with our simplified model, how the SST and SSS changes in this critical area affect the overall thermohaline circulation pattern.

We have performed three experiments. The first one uses the reconstructed SST and SSS just before the Heinrich event. The situation during the event corresponds to the second experiment, assuming that the $\delta^{18}O$-salinity relationship is 1:1, which assumes that the surface salinity decrease is due only to the meltwater from the ice sheets, in the same fashion as the global glacial-interglacial transitions [*Fairbanks*, 1989]. The third experiment is also for the HE 4 event, assuming that the $\delta^{18}O$-salinity relationship is 1:2. This is the present-day relationship observed for surface ocean waters, which results from the evaporation and precipitation processes [*Craig and Gordon*, 1965]. For each of these experiments, the model is integrated for 3000 years, up to the equilibrium. It is worth

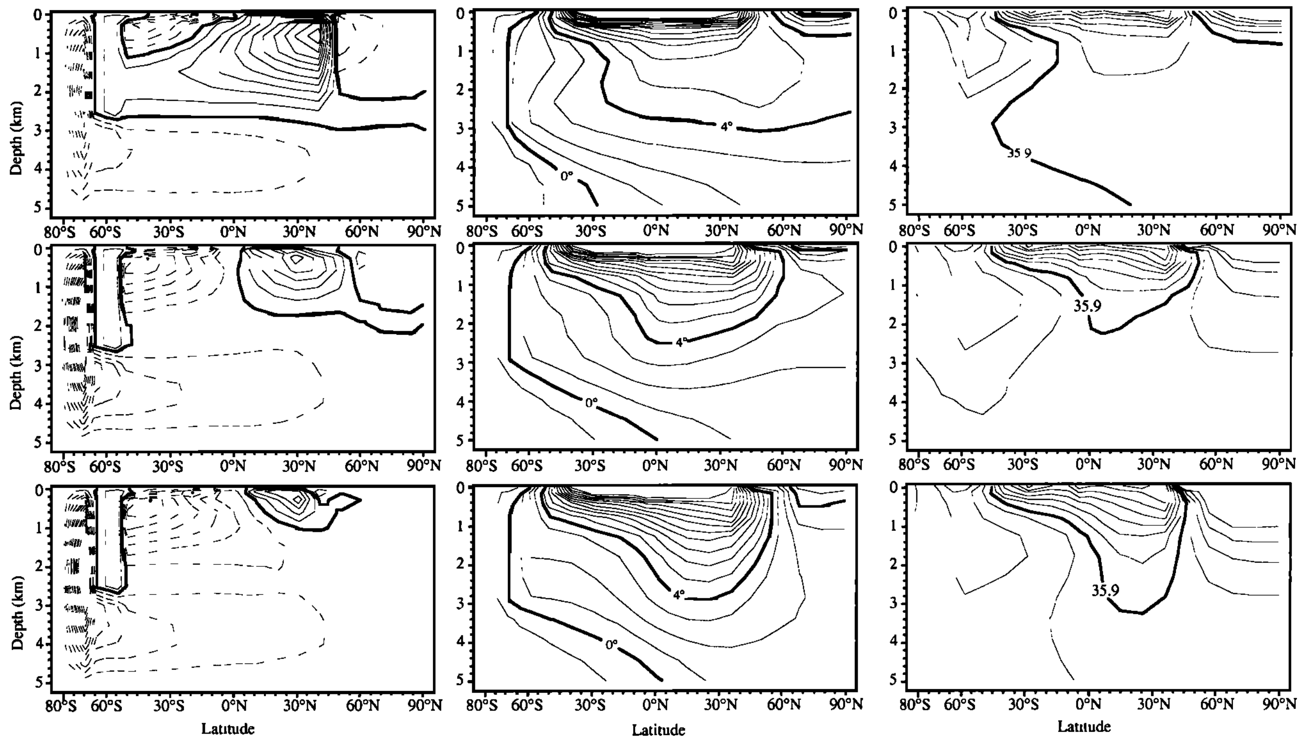

Figure 4. (top) Before HE 4. (middle) During HE 4, with δ^{18}O-S slope = 1. (bottom) During HE 4, with δ^{18}O-S slope = 2. (right) Stream functions (contours every 2 Sv). Maximum of clockwise flow: (top) 21 5 Sv, (middle) 11.2 Sv, (bottom) 9.0 Sv. Maximum of AABW: (top) 19.4 Sv, (middle) 31.6 Sv, (bottom) 33.3 Sv. (middle) Temperatures (contours every 1°C). (left) Salinities (contours every 0.2‰)

noticing that an Heinrich event lasts probably much less than 3000 years, and these experiments cannot pretend to reproduce exactly what happens during HE 4. Nevertheless, after a few hundred years of model integration, the North Atlantic circulation is very close to its equilibrium state. The results shown hereafter are therefore representative of a Heinrich event which would last at least several hundred years, which is a reasonable assumption [*François and Bacon*, 1994]. The data used for the SST and SSS are actually averages on a time period of the order of 1000 years. It is therefore quite sensible to look at the equilibrium response of the ocean to SST and SSS changes. These results are shown in Figure 4.

Since the reconstructed temperatures and salinities just before the event are only sligthly different from their LGM values, this first experiment is a small perturbation of the LGM state obtained before. The results are therefore not significantly different from the LGM ones. On the contrary, the SST and SSS are considerably lowered during the event, and the thermohaline circulation obtained by the model is notably affected. The reduction of North Atlantic Deep Water formation is such that there is, in our model, no more outflow of NADW from the North Atlantic to the south and, consequently, to other basins. Since the near-surface overturning is controlled in a large part by the wind stress, the results presented here suggest a nearly complete collapse of the circulation during Heinrich event 4. This collapse is even more evident when the 1:2 slope is used for salinity reconstructions. Since the deep waters are now fueled only by the southern flow, the deep North Atlantic is even cooler and fresher than it was in the LGM experiment.

6. Discussion

The collapse of the thermohaline circulation in the North Atlantic during Heinrich event 4 obtained in our numerical experiments is entirely consistent with the results of *Vidal et al.* [1997], which have demonstrated that the δ^{13}C of the benthic foraminifera, living in the deep sea at this time, was considerably depleted during each Heinrich event, thus suggesting a much reduced deep water ventilation. Such a reduction is also expected in order to account for the considerable cooling observed in ice core records at these times [*Dansgaard et al.*, 1993]. Indeed, as already mentioned in section 1, the Atlantic Ocean transports northward a considerable amount of heat. A reduction or a collapse of the thermohaline circulation is therefore probably the best explanation for the recorded coolings in Greenland. We have plotted, in Figure 5a, the Atlantic northward heat fluxes obtained from our experiments. As expected, the heat transport during HE 4 is much reduced. It is also interesting to note from Figure 5 that, though the LGM experiment overturning is stronger than the control one, the LGM oceanic heat transport is much smaller. This is due mainly to the more southerly location of deep water formation during LGM, which therefore considerably reduces the efficiency of meridional heat transport. This heat transport is somewhat larger before HE 4 than during LGM, which is also consistent with the higher temperatures recorded during marine isotopic stage 3. The interhemispheric heat transport is northward for the control, the LGM, and the "before HE 4" experiments, while it becomes southward for both HE 4 ones. This is consistent

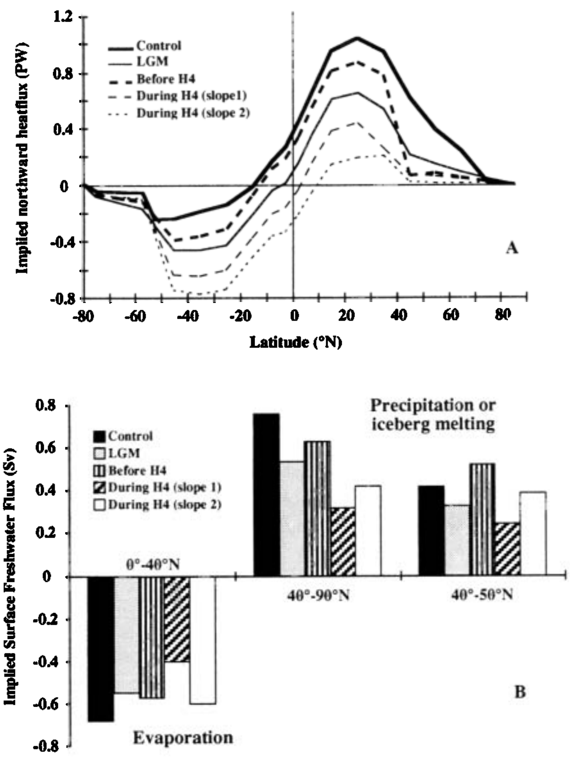

Figure 5. (a) The northward oceanic heat transport for the five numerical experiments. The maximum Northern Hemisphere values are, for the control run, 1.04 PW; for the LGM, 0.65 PW; before HE 4, 0.87 PW; during HE 4 (slope = 1), 0.43 PW; and during HE 4 (slope = 2), 0.20 PW. (b) The implied surface freshwater fluxes for different latitude bands: between 0° and 40°N, where evaporation dominates; between 40° and 90°N, where precipitation dominates; and the iceberg meltwater band between 40° and 50°N. In this last area, the freshwater fluxes required to maintain the equilibrium are not larger during HE 4 than they were before HE 4.

with the warming observed in Antarctica during the cold phases of the northern North Atlantic associated with Heinrich events [*Blunier et al.*, 1998].

We must nevertheless keep in mind that our HE 4 experiments are actually perturbation experiments around the LGM state of the Atlantic Ocean, since the coverage of SST and SSS reconstructions is limited to the high latitudes of the Atlantic. It may be possible that SST and SSS in the equatorial and south Atlantic are, before and during a Heinrich event, different from their LGM values. Nevertheless, if the deep water formation in the North Atlantic before a Heinrich event is in the 40°-50°N latitude band as is the case at LGM, then the decrease in salinity reconstructed from the $\delta^{18}O$ measurements during HE 4 is such that the collapse of the thermohaline circulation is probably inevitable, as suggested by our model.

With our ocean model, it is also quite straightforward to compute the surface freshwater fluxes required to maintain the equilibrium. These fluxes are plotted in Figure 5b for the low and high latitudes of the North Atlantic. We note that the evaporation at low latitudes and precipitation at high latitudes is reduced during the LGM compared to the control run, which is expected for a climate much cooler and therefore drier than today. More surprisingly, these required freshwater fluxes are also smaller during the Heinrich event (Figure 5b), which seems to contradict the considerable freshening recording by the SSS data and caused by the iceberg discharge. First, there is no reason to think that, during an Heinrich event, the meltwater flux at high latitude is compensated by evaporation at lower latitude. In other words, during a real Heinrich event, the water budget of the ocean is probably not equilibrated. The equilibrium freshwater fluxes shown in Figure 5b have therefore a quite limited relevance to the Heinrich event situation. Nevertheless, it is interesting to note that, when the Atlantic thermohaline circulation is greatly reduced, or even collapsed, the freshwater flux at high latitude required to maintain this state is quite low. This probably results from the hysteresis phenomena associated with the thermohaline multiple equilibria [*Rahmstorf*, 1995; *Stommel*, 1961]: Once the advective circulation is shut down by a freshwater input, it can stay down even when the freshwater input vanishes. Of course, the stability of the equilibrium states found in our study needs to be investigated with a more complex, coupled ocean-atmosphere model, and, ultimately, the understanding of the physics of Heinrich events requires a detailed modelling of the ocean, atmosphere, and ice sheet freshwater budgets.

7. Conclusion

We have performed numerical experiments, with a simplified two-dimensionnal ocean model, in order to better understand the physical consequences, on the deep ocean circulation, of the recorded North Atlantic surface cooling and freshening during a Heinrich event. In contrast to previous studies, we made no assumption about freshwater input. The main limitation of our experiments comes from the limited availability of detailed quantitative sea surface temperature and salinity reconstructions for such critical time periods. Assuming that the ocean state outside the North Atlantic is similar to the LGM situation, we have shown that the reconstructed SST and SSS for the HE 4 time period are strongly suggesting a nearly complete collapse of the North Atlantic thermohaline circulation. This result is consistent with records of the deep sea ventilation during Heinrich events, like the $\delta^{13}C$ of benthic foraminifera. Such a collapse has been predicted by many authors [*Manabe and Stouffer*, 1993; *Rahmstorf and Ganopolski*, 1998; *Stocker and Schmittner*, 1997] in the context of future, greenhouse-induced, global warming. It would therefore be of great value to better constrain the thresholds on thermal and haline forcing, beyond which such a collapse will inevitably occurs. A detailed reconstruction of the climatic conditions around a Heinrich event like HE 4 could be extremely helpful in assessing the odds of such an eventuality.

Appendix: Model Equations

In the real ocean, the main dynamical balance is essentially the geostrophic one: The Coriolis force balances the pressure gradient one:

$$f_0 \sin\theta \; v = \frac{1}{\rho_0 R \cos\theta} \frac{\partial p}{\partial \lambda} + \textit{(higher - order terms)}, \qquad (1)$$

where f_0 is the Coriolis parameter, θ is the latitude, v is the meridional velocity, ρ_0 is the mean seawater density, R is the radius of the Earth, λ is the longitude, and p is he pressure. In a two-dimensional model, it is impossible to represent explicitly the zonal gradients. We need a new equation that relates the meridional velocity v to the merional pressure gradient $\partial p/\partial\theta$. Two main different approaches have been used so far. The first is neglecting the rotation of the Earth and replacing the geostrophic balance with a dissipative one [*Marotzke et al.*, 1988; *Sakai and Peltier*, 1997; *Thual and McWilliams*, 1992; *Winton*, 1997]. The alternative, introduced by *Wright and Stocker* [1991], is to assume that the mean zonal pressure gradient $\partial p/\partial\lambda$ is proportional to the meridional one, $\partial p/\partial\theta$. We are using the *Winton* [1997] parameterization, which is, to first order, equivalent to the more sophisticated Wright and Stocker approach. The model equations are:

$$v = \frac{-1}{r\rho_0 R}\frac{\partial p}{\partial\theta} + \frac{F_W}{f_0 \sin\theta}, \qquad (2)$$

where r is the ad hoc Rayleigh damping coefficient (in s^{-1}) and where F_W is a force applied only in the top level of the model, to account for the wind forcing:

$$F_W = \frac{\tau_W}{\rho_0 \delta_E}, \qquad (3)$$

in which τ_W is the zonal wind stress and δ_E is the thickness of the top layer of the model.

Since there are no east or west boundaries in the Southern Ocean, at least in the top 2000 or 3000 m, there can be no zonal pressure gradient, and therefore the geostrophic meridional velocity must vanish. This is taken into account by setting the Rayleigh coefficient r to infinity. The vertical velocity w is obtained from the mass balance equation, the pressure p is given by the hydrostatic equation, and the density ρ is computed as a function of T (in Celsius) and S (in permil) using a third-order polynomial approximation [*Cox et al.*, 1970]. We use the rigid lid approximation at the surface of the model.

The evolution equations for the temperature and salinity fields are

$$\frac{\partial T}{\partial t} + v\frac{1}{R}\frac{\partial T}{\partial\theta} + w\frac{\partial T}{\partial z} = \kappa_H \frac{1}{R^2\cos\theta}\frac{\partial}{\partial\theta}\left(\cos\theta\frac{\partial T}{\partial\theta}\right) + \left(c+\kappa_V\right)\frac{\partial^2 T}{\partial z^2} + F_T(T), \qquad (4a)$$

$$\frac{\partial S}{\partial t} + v\frac{1}{R}\frac{\partial S}{\partial\theta} + w\frac{\partial S}{\partial z} = \kappa_H \frac{1}{R^2\cos\theta}\frac{\partial}{\partial\theta}\left(\cos\theta\frac{\partial S}{\partial\theta}\right) + \left(c+\kappa_V\right)\frac{\partial^2 S}{\partial z^2} + F_S(S), \qquad (4b)$$

where $F_T(T)$ and $F_S(S)$ are the surface temperature and salinity forcings, null in the ocean interior and equal to the following at the surface:

$$F_T(T) = \frac{1}{\tau_T}\left(T - T_R\right), \quad F_S(S) = \frac{1}{\tau_S}\left(S - S_R\right), \qquad (5)$$

in which τ_T and τ_S are the relaxation time constants and T_R and S_R are the observed surface temperatures and salinities.

In equations (4a) and (4b), κ_H and κ_V are the horizontal and vertical eddy diffusivity. The variable c is the contribution of vertical convective mixing to vertical diffusivity. Following [*Schmidt and Mysak*, 1996], we use the formula below for c:

$$c + \kappa_V = \kappa_V \left(\frac{\kappa_C}{\kappa_V}\right)^{\frac{1}{2}\left(1-\tanh\left(\frac{N^2}{N_{min}^2}\right)\right)}, \qquad (6)$$

where N is the Brunt-Väisälä frequency:

$$N^2 = -\frac{g}{\rho_0}\frac{\partial\rho}{\partial z}. \qquad (7)$$

N_{min} is the marginaly stable value of N, and κ_C is the enhanced vertical diffusivity in the fully convective regime. This last parameter can also be written

$$\kappa_C = \frac{\Delta z_{max}^2}{\tau_C}, \qquad (8)$$

where Δz_{max} is the largest vertical grid spacing and τ_C is a characteristic time for vertical convection.

The numerical implementation uses standard integration techniques. The model has 12 vertical layers. The base of each layer lies at the following depths: 100, 300, 600, 1000, 1500, 2000, 2500, 3000, 3500, 4000, 4500, and 5000 m. There are 21 latitude bands positioned between the following latitudes: 80°S, 70°S, 65°S, 60°S, 55°S, 50°S, 40°S, 30°S, 20°S, 10°S, 5°S, 0°N, 5°N, 10°N, 20°N, 30°N, 40°N, 50°N, 60°N, 70°N, 80°N, and 90°N. The numerical values for the different parameters are given in Table 1.

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