Ocean Bi–Polar Seesaw and Climate: Southern Versus Northern Meltwater Impacts

Dan Seidov, Bernd J. Haupt, Eric J. Barron, and Mark Maslin

EMS Environment Institute, Pennsylvania State University, University Park, Pennsylvania

Model simulations targeting the ocean circulation response to changes in surface salinity in the high latitudes of both Northern and Southern Hemispheres demonstrate that meltwater impacts in one hemisphere may lead to a strengthening of the thermohaline conveyor driven by the source in the opposite hemisphere. This, in turn, leads to significant changes in poleward heat transport. Further, meltwater events caused largely by sea ice melting can lead to deep–sea warming and thermal expansion of abyssal water, that in turn can cause a substantial sea level change even without a major ice sheet melting. Experiments with a glacial ocean circulation regime prone to northern and southern meltwater events imply that glacial cycles may have been influenced by both northern and southern deepwater sources. Importantly, the experiments suggest that the southern source can be a more powerful modulator of the meridional deep–ocean conveyor that the northern source, which challenges our current vision of the North Atlantic Deep Water as an ultimate driver of deep–ocean circulation. Our experiments show that the southern impact can overpower northern ones. Even in the experiment in which the amplitude of the perturbation in the North Atlantic was as high as -3 psu, and the amplitude in the Southern Ocean was only -1 psu, the deep–water regime was qualitatively the same as in the pure Southern Ocean scenario, with somewhat less deep-ocean warming, yet still global and substantial.

INTRODUCTION

One of the most noticeable attributes of present–day climate is its hemispheric asymmetry, with a warm ocean surface in the northern North Atlantic, a moderately cool northern North Pacific and a much colder Southern Ocean (e.g., *Weyl* [1968]; see also a discussion in *Weaver et al.* [1999]). This prominent hemispheric asymmetry of the thermal state of the ocean surface is determined by three major factors: (i) the freshwater regime of the subtropical Atlantic with evaporated water transferred to the Pacific Ocean over the Panama Isthmus; (ii) the North Atlantic ocean geometry that facilitates delivery of this warm and saline subtropical water far to the north to form deep convection that drives the global thermohaline circulation (THC), also known as a thermohaline ocean conveyor; and (iii) the Southern Ocean circumcurrent system that causes extreme cooling of the surface water, resulting in the formation of the densest deep–ocean water around Antarctica. Nonlinear interactions of these three factors result in an overturning regime prone to instability and rebounds. The idea of a so–called bi–polar ocean seesaw [*Broecker*, 1998; *Stocker*, 1998; *Broecker*, 2000] (see also a discussion in this volume [*Stocker et al.*, 2001]) serves to explain these rebounds. The bi–polar seesaw is an oscillating meridional overturning regime driven by two deepwater sources – the North Atlantic Deep Water (NADW) in the north, and Antarctic Bottom Water

The Oceans and Rapid Climate Change: Past, Present, and Future
Geophysical Monograph 126

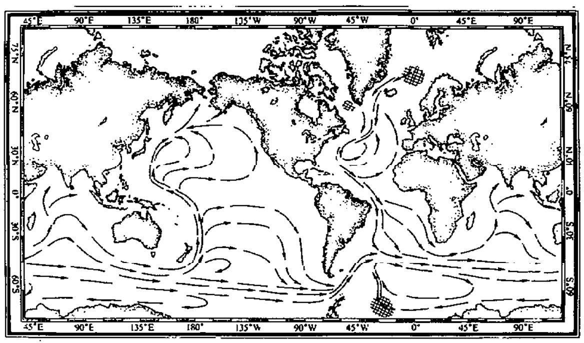

Figure 1. Stommel–Arons scheme of the deep ocean currents.

(AABW) in the south. *Broecker* [2001] argues that the thermohaline circulation flip–flops may be responsible for the 1500 year cycles in ice-rafted debris in the northern North Atlantic, which has been found by *Bond et al.* [1997]. Model results by *Ganopolski and Rahmstorf* [2001] indicate that the climate dynamics in this range of cyclicity is essentially nonlinear, and some caveats are needed in the discussion. However, although it is yet not clear whether the flip-flops might have been somehow linked to major glacial rebounds and whether they follow the same periodicity, as the cycles found by *Bond et al.* [1997], there is a little doubt that the flip-flops can have a substantial impact on the long-term climate dynamics.

The results of numerical experiments presented here are intended not only to better understand the bi–polar seesaw and its role in the ocean climate dynamics on the millennium time–scale, but also to assess the role of the Southern Ocean in driving the THC. This work is an extension of the efforts of *Seidov et al.* [2001b]. The major new element of this cited work and the results described here is the focus on the Southern Ocean response to different meltwater scenarios. Additions to the experiments presented in *Seidov et al.* [2001b] include new meltwater scenarios using present-day sea surface climatology and new scenarios for the post-glacial meltwater episodes.

This paper is structured as follows: The section "Meltwater events and THC" which follows gives an (incomplete) overview of the meltwater episodes that can be used as prototype for numerical simulations targeting the THC changes induced by such episodes, as well as some ideas that inspired our work. This section also contains an overview of numerical modeling efforts by many researchers that provide a starting point for our study. In the section "Scenarios of meltwater events" we describe how the basic knowledge of meltwater episodes in the past are translated into specific boundary conditions employed in the numerical experiments. The section entitled "Numerical experiments" deals with the specifics of the completed experiments including the numerical model and data utilized, and explains how to interpret the boundary conditions and the model output. In the next section "Present-day experiments", the results of the experiments with perturbations of the present-day surface climatology are discussed and illustrated (including sealevel change caused by thermal expansion). The results in the post-glacial meltwater scenarios are analyzed in the section "LGM experiments". These latter two sections have subsections representing the northern meltwater impacts, and various scenarios of southern impacts. The concluding section, "Discussion and Conclusions", contains our interpretation of the results as a whole, with some caveats added to emphasize the limitations associated with the preliminary character of our conclusions, and which serve to direct future efforts with more complete models.

1. MELTWATER EVENTS AND THC

During the last two decades, many researchers have contributed substantially to understanding the conveyor dynamics on decadal, centennial and millennial time scales (many studies are cited below). However, the role of the deepwater source in the Southern Ocean is still unclear, especially if compared to the far better understood role of the North Atlantic Deep Water formation. The goal of our work is to investigate further this issue of the competitive role of these two deepwater sources. To narrow our goals we focus on the high-latitudinal freshwater impacts as one of the most potent modulators of the THC dynamics.

The concept of meridional deep–ocean circulation controlled by high–latitudinal deepwater sources in two hemispheres is the foundation of the Stommel–Arons theory [*Stommel et al.*, 1958; *Stommel and Arons*, 1960]. Figure 1 reviews their theory emphasizing that the deep ocean equatorward flows are western boundary currents driven by deepwater sources in the high latitudes of the two hemispheres. The present–day deepwater regime largely depends on the strong NADW source, which was not necessarily always as today. High–latitudinal cooling causes convection and isopycnal outcropping in winter seasons. The strength of the deep western boundary currents (and therefore THC as a whole; see Figure 1) decisively depends on the volume of water either descending along these outcropping density surfaces, or mixing in the "convective chimneys", or both. The geologic record indicates that these volumes might have varied widely in the past climates, which in turn might have affected the climate on the centennial to millennial time scale. For example, these variations might have had a substantial impact on the glacial cycles of the late Quaternary (e.g., *Sarnthein et al.* [1995]; *Broecker* [1998]; *Broecker* [2000]).

A new twist of the Stommel–Arons vision of meridional overturning is given by the above–mentioned idea of the bi–polar seesaw [*Broecker*, 1998; *Broecker*, 2000] combined with the data from two ice cores, one in Greenland and one in Antarctica (see also *Blunier et al.* [1998] and *Stocker* [1998]). These two cores suggest [*Blunier et al.*, 1998] that some of the millennial glacial cycles of the Pleistocene in the Northern Hemisphere might have been out of phase with and preceded by those in the Southern Hemisphere.

Broecker et al. [1999] and *Broecker* [2000] argue that the deepwater production in the Southern Ocean has reduced from 15 Sv (1 Sv = 10^6 m^3/s) to 5 Sv over the last 800 years. They also suggest that the Little Ice Age (approximately 1350 – 1850 AD) might have been caused by far stronger deep ocean ventilation in the Southern Ocean. One of the reasons for the speed up of the Southern Ocean ventilation could be an increase in Atlantic Ocean salinity [*Broecker et al.*, 1999], whereas the slowdown could be caused by reduced surface salinity associated with warming and sea ice, icebergs, or ice sheet melting in the Southern Ocean after the Little Ice Age [*Broecker*, 2000]. *Seidov and Maslin* [2001] argue that the "heat piracy" of the North Atlantic might have been replaced by heat piracy of the Southern Hemisphere. Heat piracy means that the cross–equatorial heat transport is non–zero and can cause warming of one hemisphere by cooling another. Changing the sign of the cross–equatorial heat transport causes the hemispheres to trade places. For example, positive northward cross–equatorial oceanic heat transport, characteristic of the present–day Atlantic overturning regime, is thought to be replaced by southward cross–equatorial heat transport, characteristic of meltwater events in the North Atlantic [*Seidov and Maslin*, 2001]. Thus, meltwater events might have caused cooling of the North Atlantic and might have contributed to further cooling of the northern hemisphere because of the conveyor reduction or reversal.

Millennium time–scale climatic variations in recent geological history may also give a clue to possible future changes. The most obvious future example is a "green–house" climate that would be accompanied by major ice melting in either, or in both hemispheres. One of the most important implications of recent modeling and proxy data analyses is that ice melting may cause cooling of the hemisphere where the melting occurs, and warming of the opposite hemisphere (e.g., *Schiller et al.* [1997]; *Broecker* [2000]). Therefore our study targets both past and future THC change caused by meltwater (or equivalent thermal changes that would cause similar de–densification of the sea surface) events in either, or both hemispheres. Moreover, we question whether melting of the ice around Antarctica, or the ice in the Antarctic Circumpolar Current (ACC) is the crucial factor. We also explore sea level change caused by ocean density restructuring in the course of meltwater impacts.

Substantial evidence exists for variations in THC and in freshwater fluxes that can modify the character of deepwater currents. For example, the thermohaline meridional circulation of the present–day type was reduced during glacial periods due to alterations in atmospheric circulation as well as due to input of freshwater from melting icebergs in the North Atlantic (e.g., *Duplessy et al.* [1988, 1991]; *Sarnthein et al.* [1994, 1995]; *Seidov et al.* [1996]), and collapsed during meltwater episodes (e.g., *Manabe and Stouffer* [1988, 1995]; *Maslin et al.* [1995]; *Rahmstorf* [1995a]; *Rosell-Melé et al.* [1997]; *Zahn et al.* [1997]). However, further complicating the problem, *Wang and Mysak* [2000] show in their climate model that the THC could have been stronger prior to massive iceberg melting, especially during major ice sheet build-up (~114 ka BP).

A combination of paleoceanographic proxy data and ocean general circulation models can help to assess the circulation effects and origins of the quasi–periodic ice rafting pulses called Heinrich events (e.g., *Heinrich* [1988]; *Bond et al.* [1992]; *Andrews* [1998]; *Bradley* [1999]). Heinrich events occur every 7 to 13 k.y. and have duration of between 100 and 500 yr [*Dowdeswell et al.*, 1995a] and are thought to be an essential element of millennial-scale climate variability (see a review in this volume [*Maslin et al.*, 2001]). The ice–rafted debris found in deep–sea sediment during the Heinrich events may have originated from either the Laurentide or the European ice sheets (e.g., *Grousset et al.* [1993]; *Robinson et al.* [1995]; *Gwiazda et al.* [1996]; *Rasmussen et al.* [1997]; see Figure 2). At present it is debated whether the Heinrich events are caused by internal ice–sheet dynamics (e.g., *MacAyeal* [1992]), external climate changes [*Broecker*, 1994a; *Hulbe*, 1997], or ice sheet – THC interactions [*Wang and Mysak*, 2001]. The above-mentioned flip-flops of THC in the Atlantic Ocean might have been an important element in climatic changes caused by these meltwater events.

Substantial concern has been recently expressed about the stability of the West Antarctic Ice Sheet (WAIS) (see review ref. *Oppenheimer* [1998]). Uncertainties about Antarctic ice sheet mass balance and its contribution to global sea level rise is a major issue of debate [*Vaughan et al.*, 1999] even without a large–scale collapse of the WAIS. Hence, the potential for changes in freshwater fluxes or salinity variations to influence the Southern Ocean is also clearly evident. Moreover, *Schmittner and Stocker* [1999] give another reason for dilution of the sea surface during global warming, invoked by an increased equator–to–pole freshwater transport in a warmer atmosphere. The increased poleward moisture flux results in increased precipitation in the high latitudes, which, in concert with cryosphere melting due to a

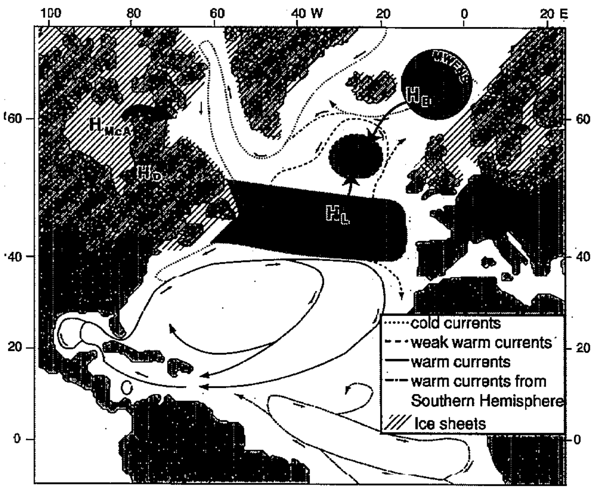

Figure 2. Reconstruction of the last glacial maximum circulation based on paleoceanographic proxy data (from *Seidov and Maslin*, 1999). Two possible source regions of icebergs, Laurentide ice sheet (H_L) and the Barents shelf (HB), are indicated; (MWENS = meltwater event in Nordic Seas).

warmer atmosphere, could enhance the meltwater impact on the ocean circulation.

Although sea surface salinity controls both NADW and AABW, the source and character of the meltwater in the Southern Hemisphere is noticeably different from the Northern Hemisphere. The meltwater episodes of the Pleistocene are usually associated with meltdown of iceberg flotillas that surged from the major glacial ice sheets or shelves (Laurentide ice sheet, or Barents ice shelf; see, e.g., *Sarnthein et al.* [1994]; *Dowdeswell et al.* [1995a]; *Maslin et al.* [1995]) in deglaciation cycles. As many believe, the Nordic Seas were at least seasonally ice–free even during the LGM [*Sarnthein et al.*, 1995; *Seidov and Maslin*, 1996]. In any case, cryosphere resources in the north are smaller than those surrounding Antarctica (e.g., *Cronin* [1999]). Although much less is known about meltwater events in the south, the shear volume of sea ice suggests that the Southern Hemisphere has a far greater potential.

Early studies [*Toggweiler and Samuels*, 1980] and more recent modeling results from *Goosse and Fichefet* [1999] confirm the importance of brine rejection during the formation of the sea ice, or equivalently, of increasing sea surface salinity caused by trapping the freshwater part of any freezing water volume in the sea ice. Moreover, AABW is not the only water mass that may be affected by diluting the surface water. Sea surface salinity controls both AABW and Antarctic Intermediate Water [*England*, 1992; *Stocker et al.*, 1992], which makes the southern impact via surface water freezing/melting cycles even more plausible. Brine rejection alone could have a substantial impact over the global thermohaline circulation, if it decreases due to reduced freezing around Antarctica caused by global warming or some other factors.

The key is that the northern North Atlantic and Nordic Seas as well as the Southern Ocean are marginally stable, and therefore are prone to destabilizing effects of even moderate de–densification of the sea surface, either due to its warming, or dilution. However, the brine rejection, ice melting, and water freezing all happen without a sea surface temperature rise because of the latent heat of melting, i.e. surface water temperature stays at the freezing/melting point as the phase change continues. (Only the near–surface temperature of water in direct contact with ice will not rise; the water below the ice can be warmed by warmer and saltier underlying water).

New research [*Seidov et al.*, 2001] substantially extends a preliminary study [*Seidov and Maslin*, 2001] by adding new northern and southern meltwater simulations. *Seidov and Maslin* [2001], following a large number of authors (e.g., *Gordon* [1986]; *Manabe and Stouffer* [1988]; *Broecker and Denton* [1989]; *Broecker* [1991]; *Gordon et al.* [1992]; *Maier-Reimer et al.* [1993]; *Broecker* [1994b]; *Weaver and Hughes* [1994]; *Manabe and Stouffer* [1995]; *Rahmstorf* [1995a]; *Sarnthein et al.* [1995]; *Schmitz* [1995]; *Weaver* [1995]; *Seidov and Haupt* [1997, 1999a]), believed that the ocean conveyor is most affected by the NADW production. *Seidov and Maslin* [2001], following *Broecker* [1998, 2000], debate the role of the southern source, but it was not directly modeled. *Seidov et al.* [2001] simulated both the southern and northern meltwater events and indicate that the southern source can be far more powerful in controlling the ocean thermohaline circulation than was previously thought. This new vision is in agreement with the arguments of [*Broecker*, 2000]. The work by *Seidov et al.* [2001] is also in line with model study of *Goosse and Fichefet* [1999], although substantial additional simulations and new views on the implications for sea level change are added. These recent works challenge the current paradigm of the global thermohaline conveyor being controlled largely by NADW formation (e.g., *Gordon et al.* [1992]).

Seidov et al. [2001] emphasize the role of the Southern Ocean and favor the southern source as having a more powerful influence in future changes if the global warming continues. This is in line with a growing modeling effort designed to investigate the climatic role of the Southern Hemisphere [*Goodman*, 1998; *Stoessel et al.*, 1998; *Hirschi et al.*, 1999; *Scott et al.*, 1999; *Wang et al.*, 1999a, 1999b; *Kamenkovich and Goodman*, 2001]. These studies describe the potential importance of feedbacks between northern and southern sources of deepwater. They suggest that freshwater forcing in the Southern Hemisphere may influence the

NADW formation and examine the importance of brine release rates on AABW formation.

To conclude the overview of THC response to meltwater events, a sketch in Figure 3 illustrates the idea of the bipolar seesaw (see also *Broecker* [2000]) in the light of recent numerical experiments by *Seidov et al.* [2001] and *Seidov and Maslin* [2001]. The upper panel (Figure 3a) represents the present–day, or interglacial mode of the conveyor, whereas two other panels show the northern meltwater mode with AABW dominance (Figure 3b) and the warm–house mode of the conveyor with complete NADW dominance (Figure 3c) in the Atlantic Ocean. The cross–equatorial heat transport follows the conveyor rebounds. The role of the cross–equatorial heat transport seems to be a very powerful climate feedback element and throughout the text we revisit its possible role in climate change caused by meltwater events. However, although our experiments give some clue to how the flip-flops of the conveyor and corresponding flip-flops of the cross-equatorial heat transport can affect the climate, a coupled cryosphere-ocean-atmosphere model is the only alternative that can fully address the climate system dynamics. Therefore, we consider the results and conclusions outlined in the next sections as a preliminary estimate of a possible behavior of THC if southern meltwater events occur, and of the differences between northern and southern meltwater events based on the same model.

2. SCENARIOS OF MELTWATER EVENTS

The numerical experiments discussed in the next sections describe idealized meltwater impacts that model either southern, northern, or combined southern and northern meltwater events. Scenarios of changes of the sea surface climatology are simple and straightforward. There are two large groups of sensitivity experiments that simulate de-densification of sea surface in the high–latitudes: (1) scenarios of possible future meltwater events that might happen due to major climate change (e.g., due to ongoing global warming, or yet unknown large–scale climate fluctuations on century time scales), and (2) idealized scenarios of meltwater events derived from the glacial cycles of the Pleistocene, with the LGM as a model for a glacial state. Two control runs in each group have been completed. Two major present–day sub–groups and one LGM sub–group are shown in Figure 4 (see detailed description of the experiments below and in Tables 1 and 2).

In the scenarios of possible future change of the THC, the upper layer temperature and salinity are restored to the present–day annual mean sea surface temperature (SST) and sea surface salinity (SSS) [*Levitus and Boyer*, 1994; *Levitus et al.*, 1994]. The *Hellerman and Rosenstein* [1983] annual mean wind stress is used in present–day simulations.

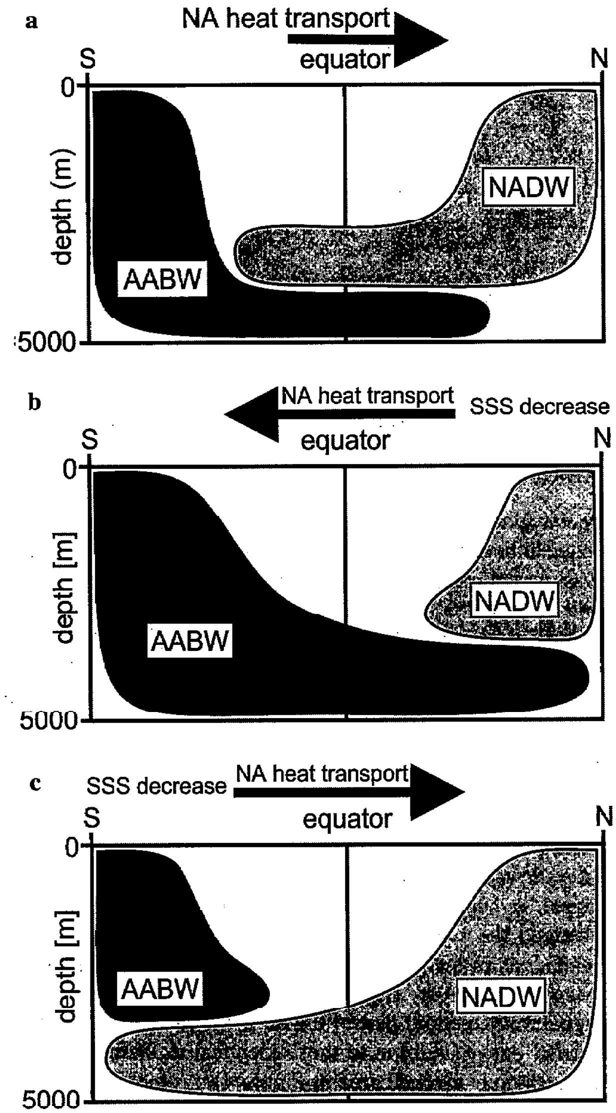

Figure 3. Schemes of water mass layering and overturning structure: (a) present–day; (b) present–day northern meltwater event; (c) present–day southern meltwater event. Direction of cross–equatorial oceanic heat transport is shown by arrows above each scheme.

In idealized scenarios of impacts of the meltwater events of the Pleistocene, basically falling in to either Heinrich Events, or Dansgaard–Oeschger Events categories (see above), the upper layer temperature was restored to reconstructed annual mean sea surface temperature (SST) during the Last Glacial Maximum (LGM). The LGM SST are modified *CLIMAP* [1981] SST, which are systematically reduced in the tropics to reflect new results indicating a

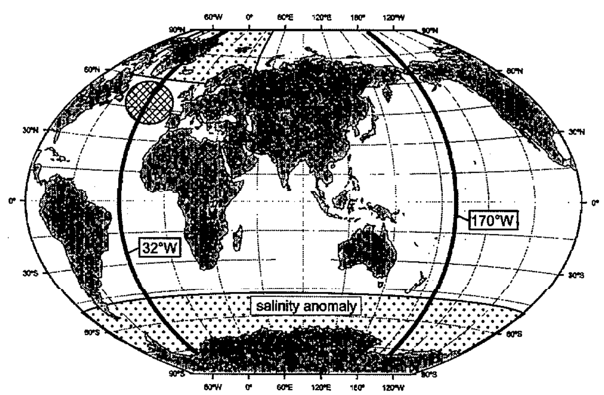

Figure 4. Idealized meltwater events. Present–day and LGM NA and SO meltwater events (dotted) and LGM NNA meltwater event (double–hatched).

cooler glacial tropical sea surface. The CLIMAP SST were reduced between 20°S to 20°N by 4°C and this perturbation was zeroed exponentially poleward between 20°S and 60°S and 20°N and 60°N.

The annual LGM see surface salinity (SSS) was constructed as described in *Seidov et al.* [1996] and *Seidov and Haupt* [1997]. Surface salinity in meltwater scenarios is the present-day and LGM SSS modified in key areas as shown in Figure 4 and Tables 1 and 2.

Wind stress fields were extracted from the output of the T42 Hamburg atmospheric circulation model, which was forced by present–day and glacial sea–surface climatologies [*Lorenz et al.*, 1996]. In the Southern Ocean, a low–salinity signal is superimposed on the LGM SSS, retaining unchanged the LGM SST and wind stress, in the very same manner as in the present–day runs (see above). In the North Atlantic, however, experiments that model a low–salinity signal in the central part of the northern North Atlantic were carried out in addition to those identical to the present–day runs (double–hatched area in Figure 4). As the runs with the low–salinity impact confined in the Nordic Seas simulate Dansgaard–Oeschger Events (e.g., *Oeschger* [1984], *Dowdeswell et al.* [1995b]; see also discussion in *Seidov and Maslin* [1999]), the scenarios with the southward shifted low–salinity signal simulate impacts of the Heinrich Events. Heinrich events are thought to be caused by melting of iceberg armadas from the Laurentide Ice Sheet [*Ruddiman and McIntyre*, 1981] (and possibly from the Barents Shelf, e.g., *Sarnthein et al.* [1995]). The essential difference from the Dansgaard–Oeschger events is that meltwater in Heinrich events caps the southward–shifted convection and therefore are associated with complete shut–off or even reversal of the Atlantic thermohaline conveyor.

The rebound of ocean conveyor depends crucially on the SSS anomalies. The perturbations we apply in our idealized scenarios do not exceed those found in paleoreconstructions (e.g., *Duplessy et al.* [1991]; *Sarnthein et al.* [1995]). However, there may be significant uncertainties in the SSS reconstructions. To test the effect of these uncertainties on the global circulation, *Seidov and Haupt* [1999a] have performed a number of numerical experiments in which the SSS in the meltwater pools was altered by as much as 1 psu. In none of these runs did the SSS modification prevent the capping of convection and the depression of the conveyor (see also *Seidov and Maslin* [1999]). Thus, despite the uncertainties, the SSS reconstructions by *Duplessy et al.* [1991] and *Sarnthein et al.* [1994] give a very solid foundation for numerical simulations based on these data [*Seidov et al.*, 1996]. The thermohaline circulation collapse due to freshwater impact in *Seidov et al.* [1996] compares well with the paleoreconstructions in *Sarnthein et al.* [1995]. It has been confirmed that the results of the regional NA model by *Seidov et al.* [1996] are still valid in global circulation experiments [*Seidov and Haupt*, 1997, 1999b; *Seidov and Maslin*, 1999]. Moreover, it has been shown in these studies that the meltwater signal in the Nordic Seas during the MWE is so strong that the circulation response is a robust feature in all numerical experiments. These results from previous realistic and idealized simulations, as well as the implications based on proxy analyses, provide more confidence in the results of the idealized study presented here.

Table 1. Amplitudes of sea surface salinity anomalies (in psu).

Exp.	NA	SO	WED	ANT	ACC
CRPDC (#1)	–	–	–	–	–
#2	-2.0	–	–	–	–
#3	–	-1.0	–	–	–
#4	-3.0	-1.0	–	–	–
#5	–	–	-3.0	–	–
#6	–	–	–	-1.0	–
#7	–	–	–	–	-1.0

CRPDC – Control Case (annual mean present–day sea surface climatology); NA – North Atlantic; SO – Southern Ocean; WED – the Weddell Sea; ANT– Antarctica coastline; ACC– ACC bound signal.
Salinity anomalies are added to the present–day annual mean sea surface salinity in the bands between 60°N and 80°N in NA, and/or 50°S and the coast of the Antarctica (SO). The modified salinity was merged using a cosine filter to the unchanged field within two latitudinal grid points (8°). In the SO the anomalies are circumglobal. The WED low salinity is confined to the Weddell Sea only. The ANT signal is in the band of 4° thickness around Antarctica; ACC is the signal in the band between around 50°S, approximate position of the ACC axes.

Table 2. Meridional overturning in the Atlantic Ocean (north of 30°S) in Sv (1 Sv = 10^6 m^3/s).

Exp.	NADW production	Convection depth in NA (km)	NADW outflow at 30°S	AABW inflow at 30°S
#1 (CRPDC)	16	3–4	10	6
#2 (NA-2 psu)	10	2	4	4
#3 (SO-1 psu)	25	bottom (> 4 km)	20	0
#4 (NA-3 +SO-1)	20	bottom (> 4 km)	14	4
#5 (WED-3 psu)	15	bottom (> 4 km)	10	4
#6 (ANT-1 psu)	20	bottom (> 4 km)	12	4
#7 (ACC-1 psu)	20	bottom (> 4 km)	12	4

3. NUMERICAL EXPERIMENTS

All experiments are completed using the GFDL MOM version 2, a well documented and extensively utilized ocean circulation model [*Pacanowski*, 1996]. A newer version 3 is now the current version. However, to our knowledge, no changes that are critical to a scenario-type coarse resolution modeling have been reported so far. A rather coarse resolution of 6°x4° with 12 levels is employed, as is appropriate for a pilot comparison between several low–salinity regimes in which the focal point is the large–scale thermohaline circulation. This resolution is comparable with other coarse resolution studies [*Toggweiler et al.*, 1989; *Weaver et al.*, 1994; *Manabe and Stouffer*, 1995; *Rahmstorf*, 1995b; *Seidov and Haupt*, 1999a; *Seidov and Maslin*, 1999] addressing similar problems, and has proven to be sufficient for studying the response of ocean meridional overturning to freshwater signals. For example, *Seidov and Haupt* [1999a] demonstrate that water transports, convection depths, and inter–basin water exchanges are reasonably well–simulated in a study with a similar spatial resolution using similar boundary conditions.

The world ocean is bounded by Antarctica in the south and 80°N in the north. Barents shelf area is included, and the eastern boundary in the North Atlantic Ocean sector is at 40°E. The Mediterranean Sea and the Arctic region are excluded.

In our experiments we use isopycnal mixing [*Gent and McWilliams*, 1990] as it is implemented in *MOM-2* [1996]. Inclusion of isopycnal mixing is crucial because, as it has been shown (see discussion in *McWilliams* [1998]), ocean circulation models without isopycnal mixing suffer from exaggerated convection, especially in the Southern Ocean region. *Hirst et al.* [2000] indicate that using of Gent-McWilliams mixing scheme substantially improved performance of a coupled ocean atmosphere model as well.

A great number of idealized numerical experiments have been completed in our study of northern and southern meltwater events. There is no way to arrange them all in one observable table. However, they can be classified into two major groups, with 5 subsets in the first group comprising the present–day runs, and 3 subgroups in the second group, comprising the LGM runs.

First subsets in each group consists of just one experiment, that is, the control run: The control run in the present–day climate (PDC) scenario (CRPDC), and the control run in the LGM scenario (CRLGM). Other runs in any group represent different meltwater scenarios. In the experiments, either the present–day or LGM sea–surface climatology are used for the surface boundary conditions (see above). Every run is compared to the corresponding control runs within its main groups, i.e., either to CRPDC, or CRLGM. All runs are 2000 model years long, with 5–fold acceleration in the deep layers (in MOM, this means that the deep ocean is effectively run for 10000 years). A complete steady state is reached in all numerical experiments (with global temperature and salinity trends in the end of experiments smaller than 10^{-4} °C and 10^{-5} psu per century respectively).

Within the PDC main group of 7 experiments, there are following subsets (Tables 1 and 2): (1) Three experiments (Exp. 2-4) that target the northern versus southern meltwater impacts; this is the core of our study; (2) two experiments (Exp. 5 and 6) that quantify the difference between local sources in the Southern Ocean around Antarctica (for example, the role of the Weddell Sea, Ross Sea and the Antarctica continental shelf; and (3) one experiment (Exp. 7) exploring whether melting of the ice around Antarctica, or the ice in the Antarctic Circumpolar Current (ACC) is the crucial factor. Exp. 1 is the control run, CRPDC (see above), with unchanged present-day climatology.

In the LGM group there are three major sub–groups: (1) LGM runs with southern and northern meltwater impacts that have the same character as in the subgroups 1 of the present–day runs, and (2) experiments with the northern impact shifted southward, as in *Seidov and Maslin* [1999, 2001].

All steady states of THC in sensitivity runs with perturbed surface salinity are compared to the steady states in the control runs. Solving this system simulates evolution of the ocean from an initial state to a steady state of the ocean currents and thermohaline structure that are fully adjusted to the sea–surface boundary conditions.

In the cold climates of the Pleistocene low salinity signals in the high latitudes are mainly due to melting of sea ice or icebergs, with poleward water vapor transport from the tropics and THC feedbacks being important but as a secondary factor (e.g., *Bradley* [1999]. High salinity signals are due to brine rejection process that accompanies seawater freezing (e.g., *Gill* [1982]). In warm ice–free climates, poleward water vapor transport or river run–off can be the only cause of a low salinity signal, whereas increased evaporation (unlikely in high latitudes) might be a cause of an increased surface salinity elsewhere. However, in our approach, temperature changes could be modeled by changing salinity, as our concern is sea surface density rather than temperature or salinity specifically. A useful rule–of–thumb is that the same density increases can be achieved by either increase of salinity by approximately 1 psu, or by decrease of temperature by about -5°C [*Pond and Pickard*, 1986]. Hence, meltwater is a convenient means by which to control density changes that actually drive the convection.

Regarding the strength of the low–salinity signal, even 1 psu is a rather moderate estimate of the possible dilution of sea surface water during a southern post-LGM meltwater event. For instance, *Goosse and Fichefet* [1999] argue that even the reduction of brine rejection alone can cause a 1 psu decrease in sea surface salinity. *Duplessy et al.* [1996] show that the low salinity anomalies in the Southern Ocean could be up to -1.8 psu during the LGM. Thus, much stronger anomalies might be expected in a southern meltwater episode. *Labeyrie et al.* [1986] argue that the periphery of the Antarctic ice sheet was eroded during some of the glacial cycles of Pleistocene, and that only 10,000 km^3 of meltwater could have reduced the sea surface $\delta^{18}O$ by 1 ‰ (which translates to approximately 2 psu in sea surface salinity [*Duplessy et al.*, 1996]). *Anderson and Andrews* [1999] revisited the problem of the late Quaternary Antarctic meltdown, and argue that significant deglaciation of the Weddell Sea continental shelf could have taken place prior to the last glaciation. *Birchfield and Broecker* [1990] point out that a relatively small freshwater flux converted to a low–salinity signal will hamper the conveyor operation. For instance, they show that a freshwater flux of 0.1 to 0.3 Sv in the North Atlantic can cause 0.3 to almost 1 psu reduction of salinity in 1000 years. The 0.3 Sv flux during a thousand years would convert to 10,000 km^3 a year, a value that is only about 4 times greater than the present–day annual meltwater production in Antarctica of about 2,500 km^3/year (e.g., *Vaughan et al.* [1999]).

In contrast to many studies aimed at sensitivity of the circulation to freshwater fluxes (e.g., *Maier-Reimer et al.* [1993; *Weaver and Hughes* [1994]; *Rahmstorf* [1995a]; *Manabe and Stouffer* [1997]) we use a restoring boundary condition on salinity (e.g., *Bryan* [1987]). Normally, a restoring boundary condition on salinity is considered as inferior to a flux formulation. A compromise could be the use of implied freshwater flax (e.g., see a discussion in *Weaver* [1999] and *Ezer* [2001] in this volume). However, in a numerical model, it is difficult to control a salinity change to match observations using a flux formulation. In order to assess compatibility issue, we have calculated apparent freshwater volumes that would be needed to dilute the surface layer to achieve a respective salinity change in the imposition domain. These volumes can be thought of as virtual freshwater volumes that would have to be added to a thin surface layer, where the SSS is specified, to dilute the water in this layer to a prescribed SSS reduction. (In the runs with increased SSS, this would be the freshwater to be removed to achieve the respective increase in SSS caused by brine rejection).

Following *Manabe and Stouffer* [1995], we estimate the rates by which these virtual freshwater fluxes would have to be added within 10 years. The total amount of freshwater added at those rates is fairly realistic. For instance, to dilute a 10 m layer of water with salinity of 35 psu by 1 psu in 10 years in one of the Southern Ocean experiments (Exp. 8 in Table 1), a freshwater (or equivalent sea ice) layer of about only 0.3 m thickness would be needed. The freshwater fluxes to maintain the low salinity signals employed in this study are comparable to those in *Manabe and Stouffer* [1995]. For example, to maintain – 1 psu anomaly in the SO experiment (Table 1), a freshwater flux of 0.06 Sv is needed.

3.1 Present–day Experiments

Table 1 includes the control run and examples from the first of the two major groups (see also Figure 4): Exp.1 is the CRPDC. Exp. 2 (NA) has a low–salinity perturbation in the high–latitudinal North Atlantic; Exp. 3 (SO) has a low–salinity signal in the Southern Ocean, Exp. 4 (NA and SO combined) has low–salinity signals in both these two regions, Exp. 5 (WED) has a low–salinity signal confined to the Weddell Sea only (we have also run an experiment with low–salinity signals in the Ross Sea; as the results do not substantially differ from those in WED, these runs are not present here), Exp. 6 (ANT) has only the area within one grid step thickness around Antarctica (repeating the shoreline) affected by meltwater, and Exp.7 (ACC) has a freshwater band between 40°S and 50°S, i.e. in a water band approximating the ACC. The experiments shown in Table 1 are the runs that have been done with varying the amplitude of the salinity perturbations.

The low–salinity signal is applied as a negative salinity anomaly with different amplitudes, from 0.5 to 3 psu in the North Atlantic, from 0.2 to 1 psu in the Southern Ocean, and from 1 to 3 in the combined cases. The low salinity in the Weddell Sea is lower than in CRPDC, with maximum differences from 0.5 to 3 psu. In the experiments exploring the relative importance of Antarctic coastal meltwater impacts versus ACC impacts the low–salinity signals are 1 psu.

We discuss the results of experiments in the same sequence as they are shown in Table 1. Six major sub–groups of events are selected, as listed above, in Exp. 2 through 7. However, only three major cases of NA, SO and the combined NA+SO cases (see notations in Table 1) are illustrated, with others only briefly discussed in the text.

North Atlantic events. An important characteristic of the thermohaline circulation is the meridional overturning streamfunction. Notably, the most important is the overturning in the Atlantic Ocean because this ocean is the most active element of the THC. Moreover, the largest component of meridional heat transport is determined by the meridional overturning.

The results of the experiments simulating low–salinity impact in the North Atlantic conform to what is already well known from previous work (see "Meltwater events and THC" section above). The conveyor is weaker and shallower and convection is shifted southward. For comparison with all sensitivity tests, the overturning in the CRPDC is shown in Figure 5a and Figure 6a. Figure 5a depicts overturning in the Atlantic Ocean, whereas Figure 6a shows the World Ocean. Figure 7a illustrates present–day model temperature sections in the Atlantic and Figure 8a illustrates the section in the Pacific Oceans in CRPDC, along meridians 32°E and 170°W respectively. Figure 9 depicts northward oceanic heat transport in the Atlantic Ocean, whereas Figure 10 shows global northward oceanic heat transport. Similarly, each of the above fields is given in Figures 5–10, part b, for a low–salinity signal in the North Atlantic. Figures 5–10, part c, showing the results with a low salinity signal in the Southern Ocean, represents southern impacts.

Northward cross–equatorial heat transport in the North Atlantic, which is a characteristic of the present–day climate, is dramatically reduced in the scenario with a strong northern freshwater impact (Figures 9 and 10). This result conforms to *Manabe and Stouffer* [1995, 1997] results showing the possibility of a northern cold episode following a northern meltwater event. However, the impact of freshwater on the conveyor depends crucially on the location of the freshwater lid. In our simulations even an excessive northern low–salinity signal of -2 psu did not cause a complete termination

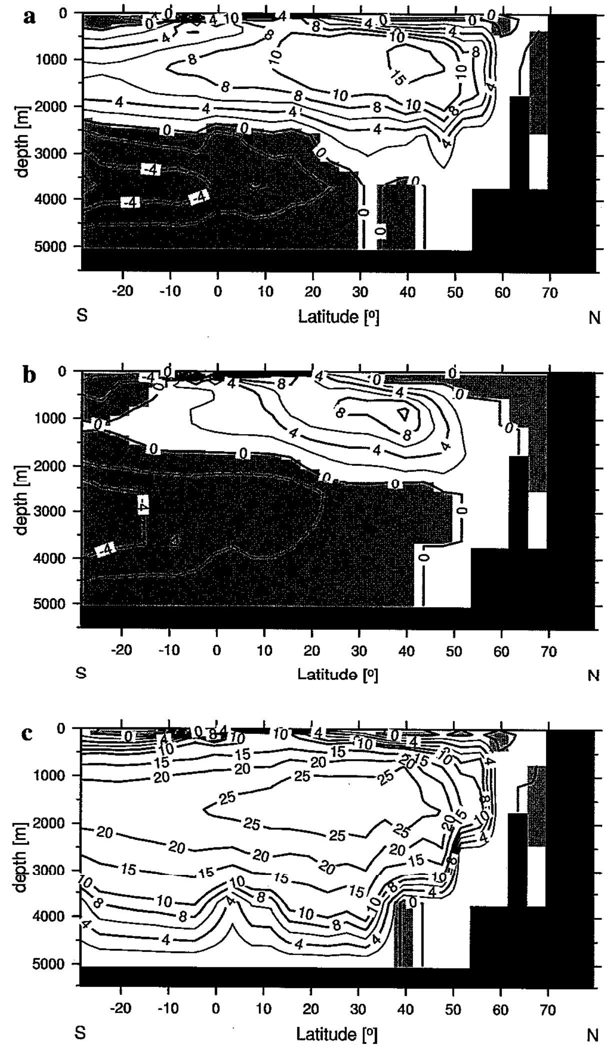

Figure 5. Meridional overturning in the Atlantic Ocean; (a) CRPDC; (b) NA event; (c) SO event (see Table 1). Streamfunction is shown in Sv (1 Sv = 10^6 m^3/s).

of the conveyor (in contrast to the runs in *Seidov and Haupt* [1997] and *Seidov and Maslin* [1999], where more southward spread of meltwater caused complete cessation or even reversal of the conveyor). Importantly, during meltwater events in the North Atlantic, only the North Atlantic is affected by southward cross–equatorial heat transport, whereas during meltwater events in the Southern Ocean the whole Northern Hemisphere gains heat at expense of heat loss in the Southern Hemisphere (Figure 9).

Temperature differences between this low–salinity scenario (Exp. 2) and the control case CRPDC (Exp. 1) at 3000

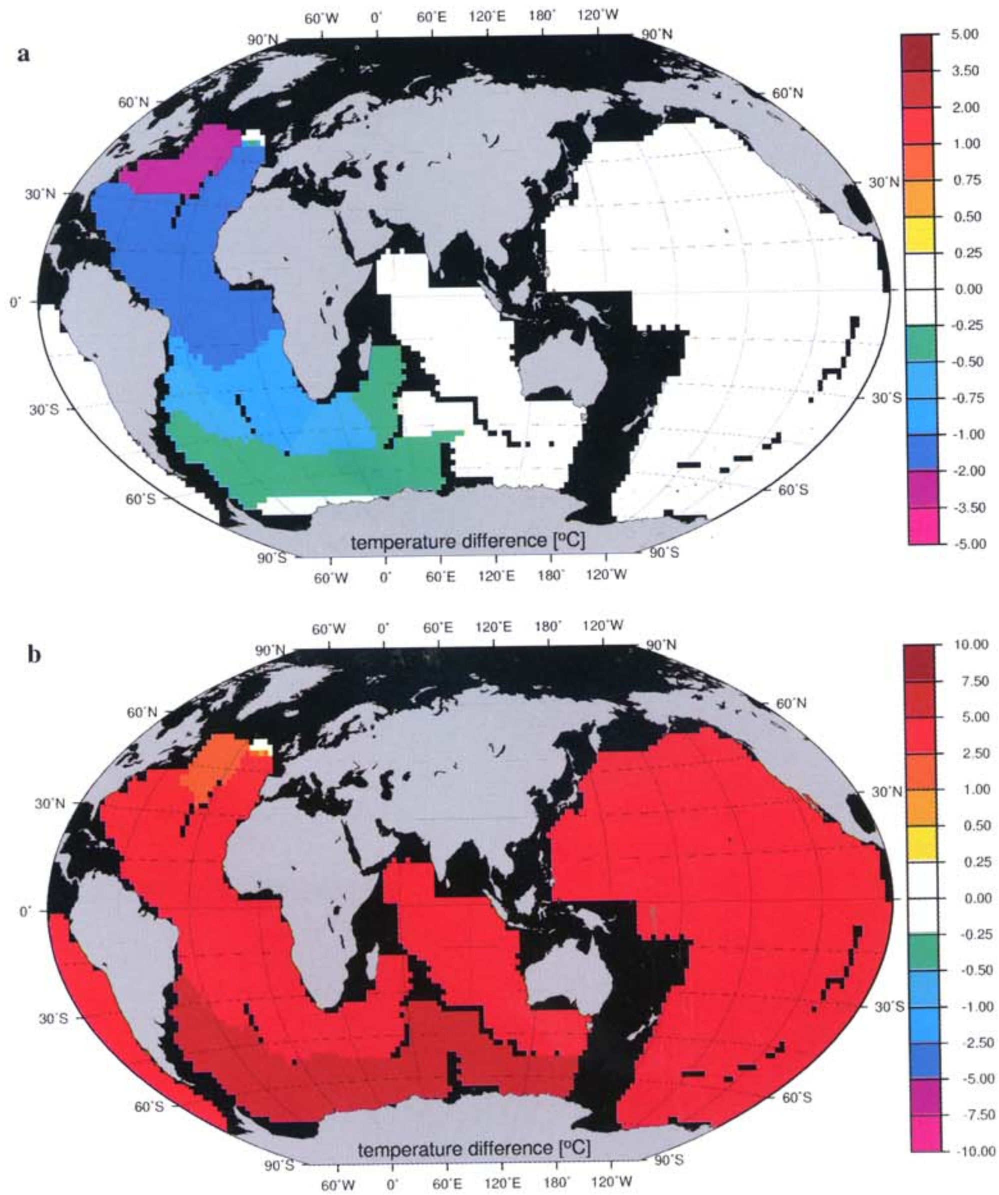

Plate 1. Temperature differences at 3000 m depth between (a) NA and CRPDC experiments and (b) SO and CRPDC experiments (see Table 1).

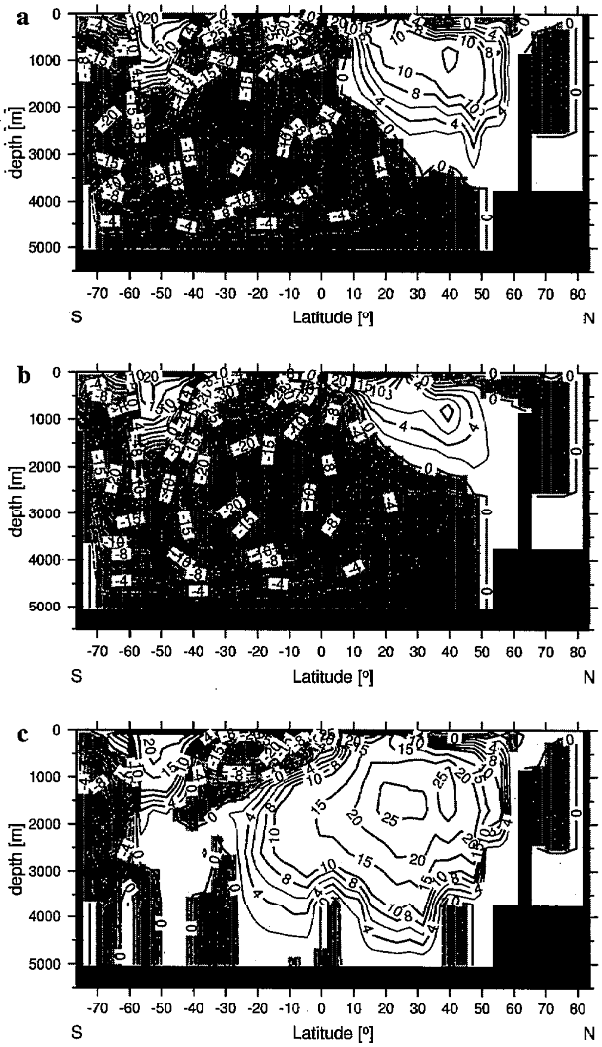

Figure 6. As in Figure 5 for the World Ocean.

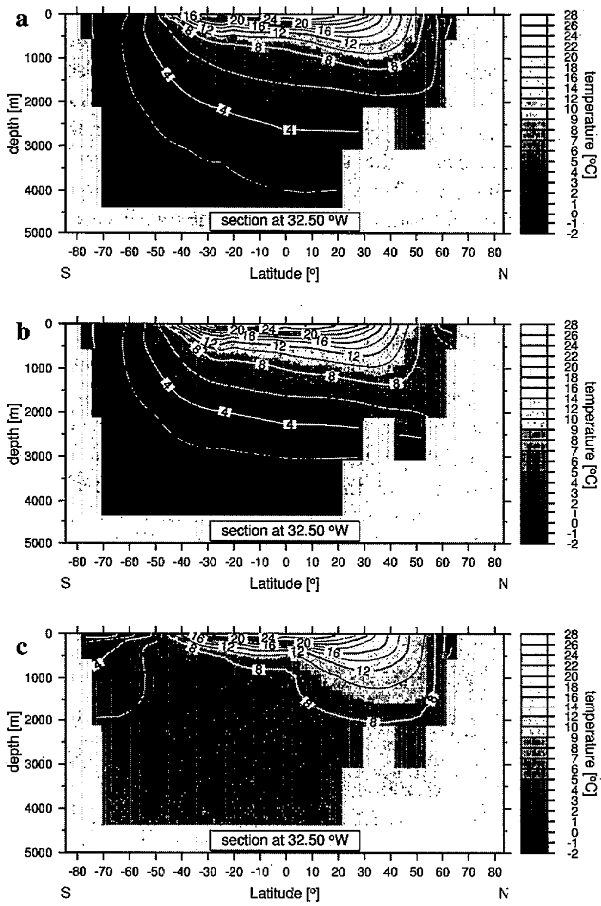

Figure 7. Temperature sections in the Atlantic Ocean at 32°W: (a) CRPDC; (b) NA event; (c) SO event.

m depth are shown in the color plate in Plate 1a. Northern events indicate cooling in high latitudes of the Atlantic Ocean. This occurs because the reduced NADW production led to a shallow conveyor, with deep water characterized by cooler and fresher water than today in these latitudes. The reduced NADW outflow, coincident with reduced replacement water crossing the equator in the North Atlantic, has an evident imprint on the oceanic heat transport.

Southern Ocean events. In contrast to the predictable and understandable results of the northern low-salinity impact, the results of the Southern Ocean surface freshening are less intuitive. Two aspects are particularly noteworthy. First, the circulation changes driven by the low-salinity signal were much stronger, and second, they led to a very strong warming of the deep ocean. Figures 5–10, part c, shows results of sea surface de-densification in the Southern Ocean in Exp. 3. Increased overturning and northward heat transport are the signatures of the southern low-salinity impacts. Temperature sections in the Atlantic Ocean (Figure 7c) and in the western Pacific Ocean (Figure 8c) and temperature differences between the SO and CRPDC cases at 3000 m depth (Plate 1b) show the dramatic worldwide warming caused by the retreat of AABW and increase of NADW.

It has long been recognized that the increase of NADW production can cause cooling of the upper waters of the Southern Atlantic as the poleward heat flux increases (e.g., *Manabe and Stouffer* [1988]; *Crowley* [1992]). However, we emphasize that the deep ocean thermal trend in the southern meltwater impact scenario can be of opposite sign to those in the upper layers. In contrast to the North Atlantic meltwater scenario, there is a substantial warming of the deep ocean everywhere. The warming takes place over the entire deep ocean and its maximum shifts to the southern edges. This deep–sea warming is caused not only by a substantially increased (by 40 to 60 %) NADW production, but also largely because the meridional overturning takes over the entire deep ocean, pushing away the lessened AABW. In the North Atlantic scenario, the meltwater impact on the conveyor caused thermal effects only in the deep Atlantic Ocean, whereas in the Southern Ocean, the meltwater scenario impact is global. The increased NADW outflow in the deep layers leads to increased, compensating northward surface water flow, which might further increase NADW production until the atmosphere warms up to reduce the cooling of the sea surface and subsequently reduce deep convection.

North Atlantic versus Southern Ocean. Although in Exp. 4 (Table 1; not shown in figures) the amplitude in the North Atlantic (-3 psu) perturbation was three times the Southern Ocean (-1 psu) event, the deep–water regime is qualitatively similar to the experiment with a Southern Ocean–only perturbation. There is somewhat less deep–ocean warming, but

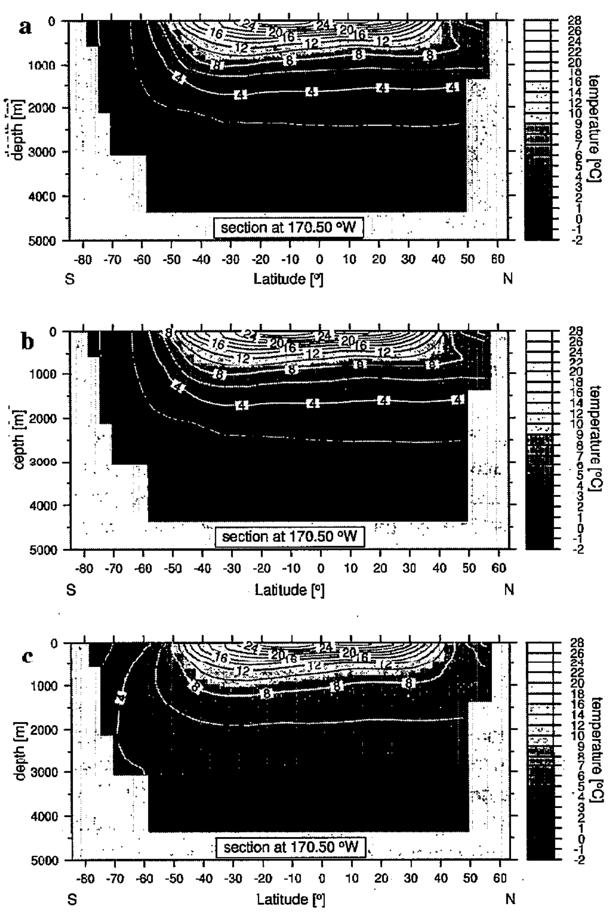

Figure 8. As in Figure 7 for the Pacific Ocean at 170°W.

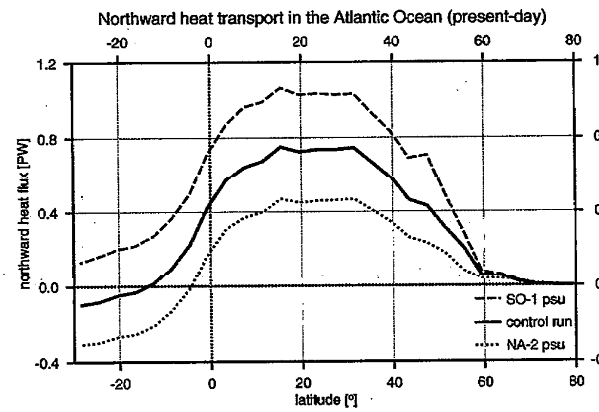

Figure 9. Northward heat transport (in PW; 1 PW = 10^{15} W) in the Atlantic Ocean in CRPDC (solid line), NA (dotted line) and SO (dash line) runs.

it remains global and substantial. Basically, the results of the runs with perturbations to two sources, in the North Atlantic and Southern Ocean, demonstrates a more powerful response to a meltwater event in the Southern Ocean than for those in the North Atlantic. However, much of this power stems from increased NADW production, adding to the evidence of the importance of the North Atlantic region. Most importantly, the problem of deep–ocean teleconnections is now seen from a very different angle.

Weddell Sea, Antarctica–bound, and ACC–bound scenarios.

Importantly, neither the Weddell Sea ice melting (Exp. 5 in Table 1), nor the Antarctica near–shore ice melting alone (Exp. 6), or the ACC–bound alone (Exp. 7) have impacts that are nearly as strong as those caused by the whole Southern Ocean events. Surprisingly, the Weddell Sea scenario did not give as noticeable a warming as was found in the whole Southern Ocean scenario. This contrasts with the belief that the Weddell Sea is the key point for the THC. Even with the amplitude of the freshwater signal in the Weddell Sea of -3 psu, the impact was far less than in the Southern Ocean scenario with only -0.2 psu. This model result implies that AABW formed around Antarctica may be more important for the conveyor dynamics than the major portion originating in the Weddell Sea. It is not clear, however, whether this would be the case in a coupled ocean–atmosphere model, with the low–salinity signal spreading from the Weddell Sea circumpolarly. In this case, the results would conform more to the SO cases, rather than to experiments with the Weddell Sea as the center of action. Our grid resolution is not sufficient to resolve the Weddell Sea in detail; therefore, we can indicate only that a local impact tied to this area would be far less powerful than a distributed impact of the whole Southern Ocean, or a large part of it.

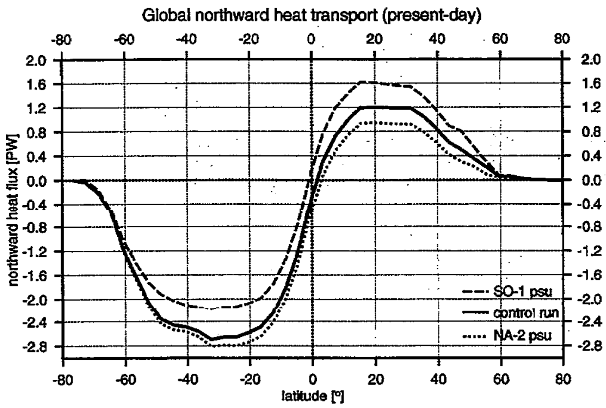

Figure 10. As in Figure 9 for the World Ocean.

In the case of a meltwater confined to either the Antarctica coastline, or the ACC bound scenario, far weaker impact is recorded in the numerical runs (Exp. 6 and 7 in Table 1).

Table 2 shows NADW production rates, the depths to which the convection reaches at the NADW convection sites, the meridional overturning at the critical latitude of 30°S, and the outflow of NADW and inflow of AABW. The balance of these two flows determines the state and intensity of the conveyor.

Implications for changes in sea level. Sea level rise caused by melting of major ice sheets is a central issue of global warming forecasts (see references in *Karl* [1993]; *Warrick et al.* [1993]; *Houghton* [1997]). However, there is also an indirect sea level effect of meltwater events caused by thermal restructuring of the world ocean. Therefore, the impact of de–densification of the sea surface is not limited to changes in oceanic circulation. As the deep ocean warms up, the sea elevation will change because of the thermal expansion of sea water. Historic hydrographic data suggest that thermal expansion of the ocean can contribute tens of centimeters to the observed sea level rise over the last century [*Godfrey and Love*, 1992]. Ocean circulation models predict ocean level rise caused by thermal expansion due to THC changes (e.g., *Church et al.* [1991]; *Weaver and Wiebe* [1999]; *Jackett et al.* [2000]; *Knutti and Stocker* [2000]). Some simulations (e.g., *Church et al.* [1991]) indicate that the thermal expansion of the ocean associated with a global warming of 3°C temperature rise by the year 2050 will result in up to a 30 cm sea level rise. On the other hand, cooling of large segments of the world ocean would compensate for the land ice sheet melting and reduce the sea level rise caused by such melting.

The difference of the sea elevation relative to the sea floor was calculated for each of the sensitivity experiments. Sea level change in Exp. 3 relative to CRPDC is shown in Figure 11. These differences in sea surface height are due to the differences in the 3–D density field caused by different T and S distribution in the world ocean. Significant sea level rise (up to 2–3 m in the SO case, Figure 11b) is evident. At the same time, the only region where sea level rise was significant in the NA case (Exp. 2; Figure 11a) is the Nordic Seas. Lowering of sea level was found in NA experiments (Figure 11a); interestingly, this lowering is far away from the northern source. Strong uprise of sea level in the SO or combined cases are largely due to deep–ocean warming. As the deep ocean is warmed up in response to the southern meltwater scenario, the water column expands and sea level rises. Notably, in many sensitive coastal areas the sea level rise can be over 1 m. Hence, it is possible that a meltwater episode, especially in the Southern Ocean (with a substantial global deep ocean warming and salinity redistribution), could strongly impact island nations and coastal regions through a noticeable sea–level change. Importantly, this sea–level rise could occur without significant melting of the ice sheets, including WAIS, which is considered the most vulnerable to climate change. If some melting of WAIS, or any other ice sheet were to happen, the effect would lead to even more dramatic changes than those shown in Figure 11.

Evolution of sea level in Exp. 2 and Exp. 3 relative to sea level in CRPDC is shown in Figure 12. The thermal response of the deep ocean and related sea level change was remarkably fast. In an integration of the model for 1000 years (tracer time without deep-ocean acceleration) with the low–salinity signal superimposed on the steady state of the control run, the first 50 % of total sea level change and warming of deep ocean was reached within first 150 years of the 2000 years of the spin up, with a much slower increase followed. Note, however, that sea level rise in Exp. 2 is almost an order of magnitude smaller than in Exp. 3. In fact, a negative global sea level change could have been expected in Exp. 2, with the deep ocean cooling. However, the deep ocean cooling in Exp. 2 is counterbalanced by substantially more warming of intermediate depths in the northern meltwater scenario (caused by shoaled NADW outflow). Therefore, there is a net rise, rather than lowering of the global sea level. Figure 12 implies that a southern meltwater episode can be a more dangerous environmental challenge than a northern meltwater episode, at least within the range considered for de–densification of the present–day sea surface.

3.2 LGM Experiments

The second large group of experiments are the runs with low salinity perturbations applied to the LGM, rather than to the present–day sea surface conditions. Experiments from this group are shown in Table 3, with all notations, except for NNA (northern North Atlantic) as in Table 1. LGM scenarios that we discuss in this paper are reduced to only four

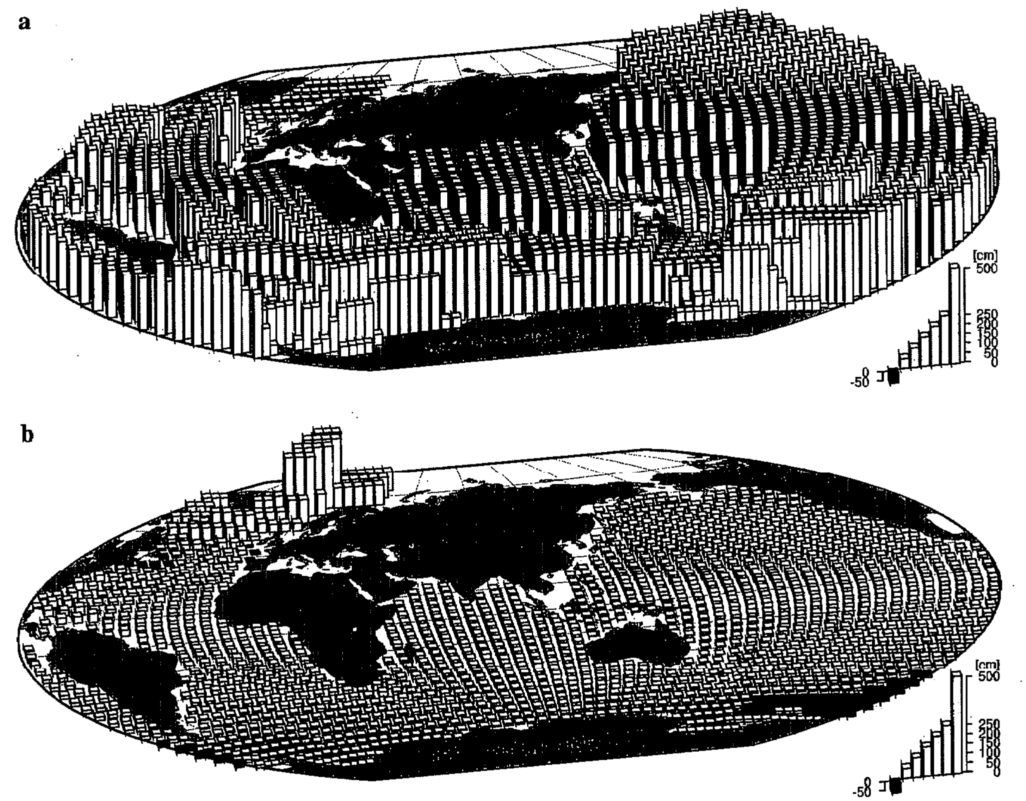

Figure 11. Sea level change in present–day NA and SO meltwater scenario (Exp. 2 and 3 in Table 1). Heights of the bars show the level change relative to the sea level in CRPDC run (Exp. 1 in Table 1).

runs (Table 3). These runs are: Exp. 1L (NA), which is the control run for LGM (CRLGM) with undisturbed LGM surface climatology as was described in the scenario section (LGM surface climatology was corrected to incorporate colder tropics; see above); setups of Exp. 2L and 4L are similar to analogues NO and SO present–day runs (Exp. 2 and 3 in Table 1). Exp. 3L is also similar to NA present–day run, except for the low–salinity signal is in the central part of the northern North Atlantic, south of Iceland, rather than in the Nordic Seas (see Figure 4). Exp. 5L combines the signals in Exp. 3L and Exp. 4L to produce an analogue to the present–day run in Exp. 4. We also refer to the results in *Seidov and Maslin* [1999, 2001].

Northern LGM events. The modeled LGM conveyor (Figure 13a), though weaker and shallower, is not as different from the modern conveyor as one might expect based on strong the surface condition changes relative to the present-day climatology. As in earlier results, the rate of overturning is up to half of the modern rate and the shape of the overturning resembles those of the present–day, though convection sites and NADW outflow was shifted southward (to reflect the shift of the convection sites) and is noticeably

Table 3. Amplitudes of sea surface salinity anomalies (in psu).

Exp.	NA	NNA	SO
CRLGM (#1L)	–	–	–
#2L	-2.0	–	–
#3L	–	-2.0	–
#4L	–	–	-1.0
#5L	–	-2.0	-1.0

CRLGM – Control Case (annual mean LGM sea surface climatology); NA – North Atlantic; NNA – (central) northern North Atlantic; SO – Southern Ocean
Salinity anomalies are added to the present–day annual mean sea surface salinity in the bands between 60°N and 80°N in NA, in the lens at about 50°N in NNA, and band between 50°S and the coast of the Antarctica (SO). The modified salinity was merged using a cosine filter to the unchanged field within two latitudinal grid points (8°). In the SO the anomalies are circumglobal.

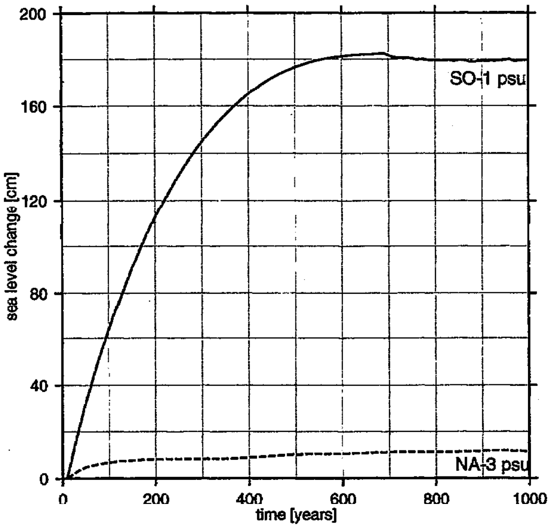

Figure 12. Evolution of the sea level change relative to the CRPDC sea level in time: solid line in SO event, and dash line in NA event.

shallower. This occurs despite the shift of NADW production to the middle of the northern North Atlantic and strongly reduced convection in the Nordic Seas (e.g., *Seidov and Haupt* [1999a]; *Seidov and Maslin* [2001]). This led *Seidov and Maslin* [2001] to conclude that convection in the northern North Atlantic is a key factor in maintaining the "normal" mode of the conveyor during glacial periods. However, it is important that although the conveyor operates in its "normal" or "conveyor-on" mode, there is very little cross–equatorial heat transport in the Atlantic Ocean (Figure 14) indicating an approximate balance between southern and northern heat budgets in this basin. Nevertheless, the global southward cross–equatorial heat transport does occur and could be instrumental in maintaining the glaciation in the north (see Figure 15).

In contrast, the meridional overturning stream function of northern meltwater events shows a reduction of the deep–water conveyor at Dansgaard – Oeschger (D-O) events (NA, Exp. 2L; Figure 13b), and a complete collapse at a Heinrich event (NA, Exp. 3L; Figure 13c). *Seidov and Maslin* [1999, 2001] indicate independence of the collapse event on the source of the meltwater based on different sources of meltwater capping convection in the central northern North Atlantic. In all scenarios of meltwater events the South Atlantic gains oceanic heat from the north due to a reversed conveyor and a change in the sign of the cross-equatorial heat transport in the Atlantic Ocean (see Figure 14 and a sketch in Figure 3b).

Southern LGM events. Stronger southern impacts, similar to those in the present–day runs, are registered in the LGM runs with low–salinity signals in the Southern Ocean. The overturning streamfunction in Exp. 4L confirms that a meltwater event in the Southern hemisphere can reverse the conveyor and heat piracy signs, and therefore reverse cooling to

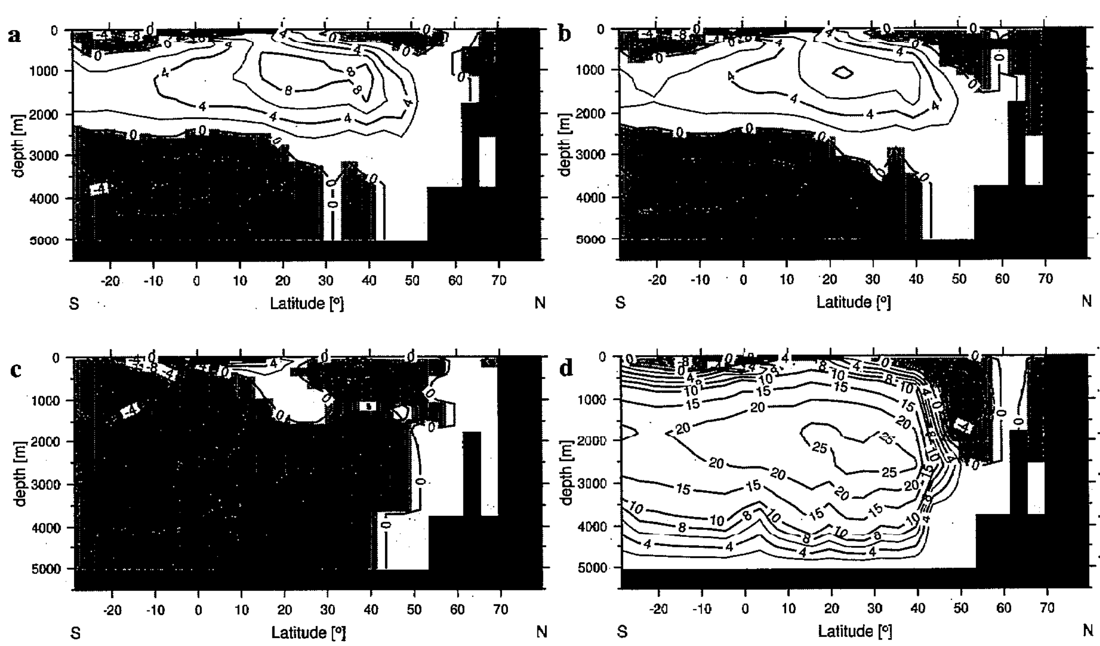

Figure 13. Meridional overturning in CRLGM (a), Exp. 2L (b), Exp. 3L (c), and Exp4L (d) (see Table 3).

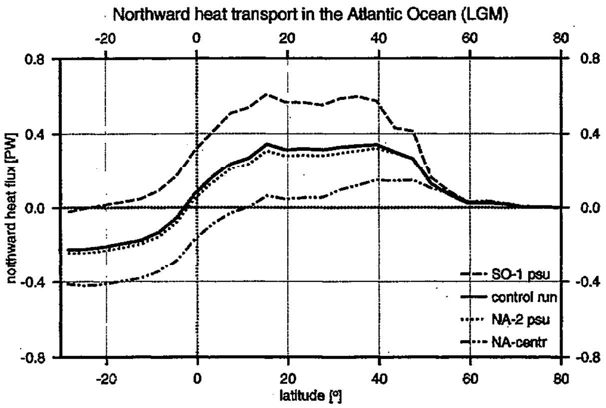

Figure 14. Northward heat transport in the Atlantic Ocean during LGM. As in Figure 9.

warming in the Northern Hemisphere (Figure 13d and Figure 14). Therefore, the rebound of the conveyor, or bi–polar seesaw, could be the cause of glacial age terminations in the Northern Hemisphere.

The combined northern and southern impacts favors the idea of stronger AABW control of the conveyor, and confirms that as soon as AABW production is curtailed, the increase of NADW production converts a cold trend to warm trend and can reverse the cooling in the north, despite the ongoing northern meltwater event. Alternatively, an increase of AABW production would curtail NADW production and lead to cooling in the Northern Hemisphere, in a harmony with Broecker's idea (e.g., *Broecker* [2000]) of the ocean bi–polar seesaw.

4. DISCUSSION AND CONCLUSIONS

Based on their numerical experiments assessing Heinrich and Dansgaard–Oeschger type events, *Seidov and Maslin* [2001] presented an explanation of hemispheric asymmetry by variations in relative amount of deepwater formation in the two hemispheres and thus by the resulting heat piracy. For example, present–day oceanic cross–equatorial heat transport is northward, whereas during a Heinrich event the cross–equatorial transport was southward. Thus, South Atlantic Ocean post–glacial heat piracy, replacing the present–day North Atlantic Ocean piracy, causes the Southern Hemisphere oceans to warm, while the Northern Hemisphere oceans cool. When the iceberg armada ceases, the freshwater cap on NADW formation is removed and northward cross–equatorial heat transport kicks in to restore a North Atlantic heat piracy condition. Accordingly, the North Atlantic Ocean warms up the Northern Hemisphere while the Southern Hemisphere cools down. If the glacial boundary conditions re–assert themselves, the North and South Atlantic Oceans come back to an almost perfect balance (see the sketch in Figure 3 explaining the heat piracy idea).

An important result of the *Seidov and Maslin* [1999, 2001] computer simulations is that a high–latitudinal meltwater (or de–densification, in general) scenario, although it does not reject the upper–ocean impact, does NOT require a substantial change in the heat transported by the upper ocean currents. The imbalance between NADW and AABW productions, i.e. between the deep–ocean flows, is the primary control on cross–equatorial heat transport, and thus could be the sole agent responsible for the observed seesaw climate oscillations. This seems to be counter–intuitive, as it is usually assumed that only changes of the upper–ocean currents can affect the heat transport to high latitudes (see Appendix in *Seidov and Maslin* [2001] for an explanation for why the deep ocean heat transport can be of the same sign as the heat transport in the upper ocean). It is suggested that the deep–ocean currents could be the ultimate internal mechanism capable of reversing cross–equatorial heat transport within the required time scale of the oscillations (hundreds of years or longer).

Support for this deep–ocean driving force hypothesis comes from the relative timing of the north–south lead–lag observed in the ice cores [*Blunier et al.*, 1998]. For example, the substantial weakening of the conveyor would happen within a matter of years, as soon as a sufficient number of icebergs travel to the convection sites in the northern North Atlantic to provide the meltwater capping of the convection. However, the actual warming of high latitudinal waters, due to change of overturning and the related change of the heat transport may take hundreds of years. For instance, *Broecker* [2000] argues that 200 years may be needed to reduce the density of the sea surface waters in one hemisphere to the point when the hemispheres change the role and the conveyor flip–flops.

This is why, when the GRIP and Byrd ice cores are compared, the peak warming in Antarctica is delayed, occurring almost at the end of the Heinrich event [*Blunier et al.*, 1998]. Moreover, the dramatic warming at the end of the Heinrich event and the switching "on" of North Atlantic heat transport regime would take hundreds of years to steal enough heat to cool down the Southern Hemisphere. Hence it can explain the delay seen in the ice core data (*Broecker* [2000] estimated that the time needed to rebound of the conveyor may as long as several hundred years).

Our simulations demonstrate the stronger sensitivity of the global ocean circulation to variations in the southern than to northern deepwater source. A new vision of the role of the Southern Ocean role in driving the ocean conveyor becomes significant in view of possible consequences of the global warming process. In this scheme, the seesaw oscillating mechanism, prone to an AABW regime, follows a simple rule. The rule is that when the NADW subsides, the AABW picks up and affects the heat transport regime in the opposite way. Alternatively, when AABW subsides, NADW spins up and warms up the deep ocean.

Although it is believed that there is less AABW variation than NADW variability during the glacial cycles of the Pleistocene, the reduction of NADW implies an increased impact of AABW even if the latter source does not change at all. Therefore, the perturbation of the conveyor only at the NADW sites could be a sufficient mechanism for sustaining the suggested climatic oscillations. If, however, sufficient evidence were found that glacial AABW production had varied substantially (e.g., *Stephens and Keeling* [2000]), theoretically it alone could drive the climate seesaw. *Broecker* [2000] suggests that AABW increase drove the Little Ice Age that ended after AABW production subsided. Hence we should anticipate a true bi–polar seesaw driven by two hemispheric sources. The key element is that the seesaw rebounds are driven by changes of deep–ocean currents rather than by sea–surface circulation alterations. However, although a southern meltwater might be a powerful near–future climate element, the role of the southern source of AABW in modulating variability has received much less attention then northern impacts, limiting the development of a complete understanding of decadal to millennial time–scale climate change. Therefore, our major task was to compare the roles of the hemispheric de–densification sources.

The source and character of the salinity perturbations in the Southern Hemisphere is substantially different from the Northern Hemisphere, and involves brine rejection in the formation of sea ice, freshwater fluxes from ablation and calving of the Antarctic Ice Sheet and the stability of the WAIS. Hence, the potential for changes in freshwater fluxes or salinity variations to influence the Southern Ocean is clearly evident. Our results demonstrate that changes in surface salinity of the Southern Ocean can significantly alter deepwater structure and temperatures. In addition, the experiments demonstrate that the high latitude sea surface need not be very warm, or very salty, or both to produce significant deep ocean warming.

In principle, an imbalance of meltwater impacts in high latitudes may lead to warming up of the deep ocean even if the ocean in high latitudes stays relatively cool in one of the hemispheres. In this case, the warming is in response to increased overturning in one of the hemispheres, whereas the other hemispheric source of deepwater might become stagnated. In other words, a strong meltwater impact in one of the two hemispheres may lead to a strengthening of the thermohaline conveyor driven by the source in another hemisphere. This, in turn, may lead to an increased surface poleward compensating flow in the active hemisphere. Further, this warming can have a substantial sea level impact even without ice sheet melting.

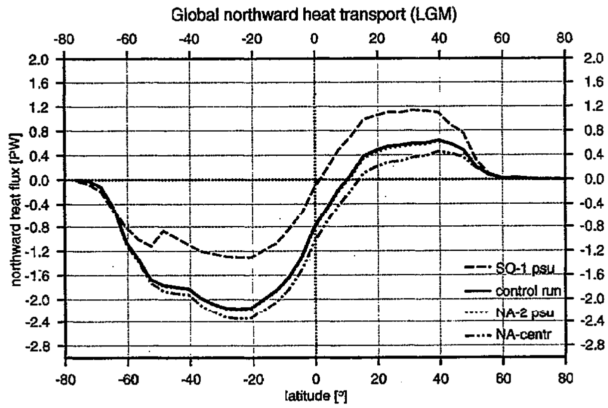

Figure 15. Northward heat transport in the Atlantic Ocean during LGM. As in Figure 9.

We have found that the warming of deep–ocean is caused by NADW intensification that accompanies the reduction of AABW, rather than by reduction of the southern source itself. The study presented here emphasizes the competitive nature of the northern and southern sources and indicates the role of the southern source as a strong modifier of NADW on centennial and millennial time scales. Further, this study calls attention to the important implications of changes in the THC, both in terms of heat transport and sea level rise.

Our results on the amplitude of deep–ocean warming during hypothesized near–future meltwater impacts can be compared with observations. As shown by *Levitus et al.* [2000], the global volume warming is 0.06°C per 40 years. Our global warming rates vary between 4°C to 7°C per 1000 years which is about two to three times higher than the observed present–day trend [*Levitus et al.*, 2000], would it continue for a thousand years. These rates could be dependent on the fact that there are no atmospheric feedbacks that could slow down the ocean warming induced by the southern meltdown (see below). Alternatively, the warming could speed up as the AABW weakens with the addition of meltdown (if is the major cause of the warming trend). Note that the ocean warming is not linear in time (see above), even if the low–salinity signal is kept constant.

In discussing the role of the Southern Ocean, we emphasize that neither the Weddell or Ross Sea deepwater sources, nor the meltwater around Antarctica are powerful enough to overwhelm the direct impact over the NADW source. Only the entire Southern Ocean impact has the power to control the world ocean deep circulation (albeit not directly but via modulating the NADW operation). Hence, the southern ice sheet meltdown and sea ice melting work synergetically to counterbalance the northern hemisphere influence.

Finally, we employed the last glacial maximum sea surface climatology to test our conclusions about the bi–polar seesaw sensitivity to the southern de–densification impact. All our above–formulated conclusions, as well as those presented in *Seidov and Maslin* [2001], stay intact in these paleoceanographic experiments. Regardless of the amplitude, origin or location of a meltwater impact in the North Atlantic, a counterbalancing de–densification of sea surface water in the Southern Ocean has a potential for again reversing the conveyor and restoring the present–day North Atlantic heat piracy pattern. As a restored northward cross–equatorial oceanic heat transport might cause a recurrence of northern cooling, the warming of the northern North Atlantic might revoke melting. Hence the ocean seesaw can rebound again, and the whole cycle will repeat multiple times until the northern meltwater resources are nearly exhausted (like at present or milder climates of the Holocene). Thus, in our simulations, we end up with a simple scheme of millennial scale climate oscillations driven by deep–ocean circulation, which is controlled by two hemispheric deepwater sources, in the North Atlantic and in the Southern Ocean. This scheme can also explain the apparent hemispheric asymmetry of the glacial record. And most importantly, it implies that the Southern ice melting impact can be a real threat in the climate change trends.

Although a stand–alone ocean model helps to outline the nature of the problem and enables a wealth of sensitivity experiments it has inherent limitations as a true climate change study. Basically, our approach is largely a first–order analysis. As such, the more extreme scenarios provide a higher level of sensitivity that clearly illustrate the potential response of the ocean without atmospheric and cryospheric feedbacks. The response of a coupled model may be different than a stand–alone ocean model because of the potential importance of feedbacks associated with the atmospheric response to an altered poleward ocean heat transport, or the impact of wind stress changes on global thermohaline overturning. Hence, our work is only a first step in assessing the climate response to changes in freshwater inputs at high latitudes. A use of a coupled ocean-atmosphere model may call for some corrections of the ocean seesaw dynamics as it is seen here. However, despite all the advances in getting supercomputers faster, scenario-type simulations of past climates are still hardly affordable. Palliative solution can be sought in using either "enhanced" ocean models, with atmospheric part downsized to energy balance models (e.g., *Bjornsson et al.* [1997]; *Weaver et al.* [1998]; *Weaver and Wiebe* [1999]; *Bjornsson and Toggweiler* [2001] (this volume), or so-called models of "intermediate complexity", with ocean-atmosphere and other climatic feedback included using simplified components of the climate systems (e.g., *Ganopolski et al.* [1998]; *Wang and Mysak* [2000, 2001]; *Ganopolski and Rahmstorf* [2001] (this volume)).

There are many other caveats that may be added to underline the limited nature of our "ocean-only" modeling results. Most of them are obvious and are due to absence of feedbacks between ice, ocean and atmosphere components of the climate system. However, we believe that the main result of importance of the southern freshwater impacts in seesaw behavior is valid despite of all the limitations of our approach. The Southern Ocean has the potential to overpower the Northern Hemisphere oceans and become a major climatic player in long–term climate change in some combination of climatic factors. Yet it is the NADW that is a universal driver of the conveyor. Amplified by southern meltwater episodes or reduced by the northern meltwater impacts, it remains the strongest player in bi–polar thermohaline conveyor variability. The two hemisphere sources, if reduced, lead to principally different consequences: If the NADW is reduced, the deep ocean cools down, whereas reduction of AABW leads to warmer abyssal waters. Therefore, a change of sea–ice or WAIS state tending toward less saline surface waters in the Southern Ocean can cause unfavorable sea–level changes, whereas the collapse of the northern source might cause cooling of the northern climate. Both scenarios may pose a serious threat to climate–sensitive environments.

Acknowledgments. We are very grateful to Lawrence Mysak, Ron Stouffer, Andrew Weaver, Andreas Schmittner, and Katrin Meissner for their useful and extended comments, which helped to substantially improve the manuscript. This study was partly supported by NSF (NSF project #9975107 and ATM 00-00545).

REFERENCES

Anderson, J.B. and J.T. Andrews, 1999: Radiocarbon constraints on ice sheet advance and retreat in Weddell Sea, Antarctica, *Geology*(27): 179-182.

Andrews, J.T., 1998: Abrupt changes (Heinrich events) in late Quaternary North Atlantic marine environments, *Journal of Quaternary Science*, 13: 3-16.

Birchfield, G.E. and W.S. Broecker, 1990: A salt oscillator in the glacial Atlantic? 2. A "scale" analysis model, *Paleoceanography*, 5: 835-843.

Bjornsson, H., L.A. Mysak and G.A. Schmidt, 1997: Mixed boundary conditions versus coupling with an energy-moisture balance model for a zonally averaged ocean climate model, *Journal of Climate*, 10: 2412-2430.

Bjornsson, H. and J.R. Toggweiler, 2001: The climatic influence of Drake Passage, *Geophysical Monograph (This Volume)*, D. Seidov, B.J. Haupt and M. Maslin (Editors), American Geophysical Union, Washington, D.C.

Blunier, T., J. Chappellaz, J. Schwander, A. Daellenbach, B. Stauffer, T.F. Stocker, D. Raynaud, J. Jouzel, H.B. Clausen, C.U. Hammer and S.J. Johnsen, 1998: Asynchrony of Antarctic and Greenland climate change during the last glacial period, *Nature*, 394: 739-743.

Bond, G., H. Heinrich, W. Broecker, L. Labeyrie, J. McManus, J. Andrews, S. Huon, R. Jantschik, S. Clasen, C. Simet, K. Tedesco, M. Klas, G. Bonani and S. Ivy, 1992: Evidence for massive discharges of icebergs into the North Atlantic ocean during the last glacial period, *Nature*, 360: 245-249.

Bond, G., W. Showers, M. Cheseby, R. Lotti, P. Almasi, P. deMenocal, P. Priore, H. Cullen, I. Hajdas and G. Bonani, 1997: A pervasive millennial-scale cycle in North Atlantic and glacial climates, *Science*, 278: 1257-1366.

Bradley, R.S., 1999, Paleoclimatology: Reconstructing climates of the Quaternary, Academic Press, San Diego, 613 pp.

Broecker, W., 1991: The great ocean conveyor, *Oceanography*, 1: 79-89.

Broecker, W., 2001: The big climate amplifier ocean circulation-sea ice-storminess-dustiness-albedo, *Geophysical Monograph (This Volume)*, D. Seidov, B.J. Haupt and M. Maslin (Editors), American Geophysical Union, Washington, D.C.

Broecker, W.S., 1994a: Massive iceberg discharges as triggers for global climate change, *Nature*, 372: 421-424.

Broecker, W.S., 1994b: An unstable superconveyor, *Nature*, 367: 414-415.

Broecker, W.S., 1998: Paleocean circulation during the last deglaciation: A bipolar seesaw? *Paleoceanography*, 13: 119-121.

Broecker, W.S., 2000: Was a change in thermohaline circulation responsible for the Little Ice Age? *Proc. Nat. Acad. Sci.*, 97(4): 1339-1342.

Broecker, W.S. and G.H. Denton, 1989: The role of ocean atmosphere reorganizations in glacial cycles, *Geochimica Cosmochimica Acta*, 53: 2465-2501.

Broecker, W.S., S. Sutherland and T.-H. Peng, 1999: A possible 20th century slowdown of Southern Ocean deep water formation, *Science*, 286: 1132-1135.

Bryan, F., 1987: Parameter sensitivity of primitive equation ocean general circulation models, *Journal of Physical Oceanography*, 17: 970-985.

Church, J.A., J.S. Godfrey, D.R. Jackett and T.J. McDougall, 1991: A model of sealevel rise caused by ocean thermal expansion, *Journal of Climate*, 4: 438-456.

CLIMAP, 1981: Climate: Long-Range Investigation, Mapping, and Prediction (CLIMAP) Project Members, Seasonal reconstructions of the Earth's surface at the Last Glacial Maximum. *Map and Chart Ser. MC-36*. Geological Society of America, Boulder, Colorado, pp. 1-18.

Cronin, T.M., 1999, Principles of Paleoclimatology, Perspectives in Paleobiology and Earth History. Columbia University Press, New York, 560 pp.

Crowley, T.J., 1992: North Atlantic deep water cools the southern hemisphere, *Paleoceanography*, 7: 489-497.

Dowdeswell, J.A., M.M. A., A.J. T. and M.I. N., 1995a: Iceberg production, debris rafting, and the extent and thickness of Heinrich layers (H1, H2) in North Atlantic sediments., *Geology*, 23: 301-304.

Dowdeswell, J.A., M.A. Maslin, J.T. Andrews and N.I. McCave, 1995b: Iceberg production, debries rafting, and the extent and thickness of Heinrich layers (H1,H2) in North Atlantic sediments, *Geology*, 23: 301-304.

Duplessy, J.-C., L. Labeyrie, A. Julliet-Lerclerc, J. Duprat and M. Sarnthein, 1991: Surface salinity reconstruction of the North Atlantic Ocean during the Last Glacial Maximum, *Oceanologica Acta*, 14: 311-324.

Duplessy, J.-C., L. Labeyrie, M. Paterne, S. Hovine, T. Fichefet, J. Duprat and M. Labracherie, 1996: High latitude deep water sources during the Last Glacial Maximum and the intensity of the global oceanic circulation, In: G. Wefer, W.H. Berger, G. Siedler and D. Webb (Editors), *The South Atlantic*. Springer, NY, pp. 445-460.

Duplessy, J.C., N.J. Shackleton, R.G. Fairbanks, L. Labeyrie, D. Oppo and N. Kallel, 1988: Deepwater source variations during the last climatic cycle and their impact on the global deepwater circulation, *Paleoceanography*, 3: 343-360.

England, M.H., 1992: On the formation of Atlantic intermediate and bottom water in ocean general circulation models, *Journal of Physical Oceanography*, 22: 918-926.

Ezer, T., 2001: On the response of the Atlantic Ocean to climatic changes in high latitudes: Sensitivity studies with a sigma coordinate ocean model, *Geophysical Monograph (This Volume)*, D. Seidov, B.J. Haupt and M. Maslin (Editors), American Geophysical Union, Washington, D.C.

Ganopolski, A. and S. Rahmstorf, 2001: Stability and variability of the thermohaline circulation in the past and future: a study with a coupled model of intermediate complexity, *Geophysical Monograph (This Volume)*, D. Seidov, B.J. Haupt and M. Maslin (Editors), American Geophysical Union, Washington, D.C.

Ganopolski, A., S. Rahmstorf, V. Petoukhov and M. Claussen, 1998: Simulation of modern and glacial climates with a coupled global model of intermediate complexity, *Nature*, 391: 351-356.

Gent, P.R. and J.C. McWilliams, 1990: Isopycnal mixing in ocean circulation models, *Journal of Physical Oceanography*, 20: 150-155.

Gill, E.G., 1982, Atmosphere-ocean dynamics, International Geophysical Series, 30. Academic Press, San Diego, 666 pp.

Godfrey, J.S. and G. Love, 1992: Assessment of sealevel rise, specific to the South Asian and Australian situations In: Seal Level Changes: Determination and Effects, *Seal Level Changes: Determination and Effects. Geophys. Monograph 69*. AGU, Washington, D.C., pp. 87-94.

Goodman, P.J., 1998: The role of North Atlantic Deep Water formation in an OGCM's ventilation and thermohaline circulation, *J. Phys. Oceanogr.*, 28: 1759-1785.

Goosse, H. and T. Fichefet, 1999: Importance of ice-ocean interactions for the global ocean circulation: A model study, *Journal of Geophysical Research*, 104(C10): 23337-23355.

Gordon, A., 1986: Interocean exchange of thermocline water, *Journal of Geophysical Research*, 91: 5037-5046.

Gordon, A.H., S.E. Zebiak and K. Bryan, 1992: Climate variability and the Atlantic Ocean, *Eos, Transactions, American Geophysical Union*, 79: 161,164-165.

Grousset, F.E., L. Labeyrie, J.A. Sinko, M. Cremer, G. Bond, J. Duprat, E. Cortijo and S. Huon, 1993: Patterns of ice-rafted detritus in the glacial North-Atlantic (40-55°N), *Paleoceanography*, 8: 175-192.

Gwiazda, R.H., S.H. Hemming and W.S. Broecker, 1996: Tracking the sources of icebergs with lead isotopes: The provenance of ice rafted debris in Heinrich event 2., *Paleoceanography*, 11: 77-93.

Heinrich, H., 1988: Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years., *Quaternary Research*, 29: 142-152.

Hellerman, S. and M. Rosenstein, 1983: Normal monthly wind stress over the world ocean with error estimates, *Journal of Physical Oceanography*, 13: 1093-1104.

Hirschi, J., J. Sander and T.F. Stocker, 1999: Intermittent convection, mixed boundary conditions and the stability of the thermohaline circulation, *Climate Dynamics*, 15(4): 277-291.

Hirst, A.C., S.P. O'Farrell and H.B. Gordon, 2000: Comparison of a coupled ocean-atmosphere model with and without oceanic eddy-induced advection. Part I: Ocean spinup and control integrations, *Journal of Climate*, 13: 139-163.

Houghton, J., 1997, Global warming. Cambridge University Press, N.Y., 251 pp.

Hulbe, C., 1997: An ice shelf mechanism for Heinrich layer production, *Paleoceanography*, 12: 711-717.

Jackett, D.R., T.J. McDougall, M.H. England and A.C. Hirst, 2000: Thermal expansion in ocean and coupled general circulation models, *Journal of Climate*, 13: 1384-1405.

Kamenkovich, I.V. and P.G. Goodman, 2001: The effects of vertical mixing on the circulation of the AABW in the Atlantic. *Geophysical Monograph (This Volume)*, D. Seidov, B.J. Haupt and M. Maslin (Editors), American Geophysical Union, Washington, D.C.

Karl, T.R., 1993: A new perspective on global warming, *Eos, Transactions, American Geophysical Union*, 74: 25.

Knutti, R. and T.F. Stocker, 2000: Influence of the thermohaline circulation on projected sea level rise, *Journal of Climate*, 13(12): 1997-2001.

Labeyrie, L.D., J.J. Pichon, M. Labracherie, P. Ippolito, J. Dupart and J.-C. Duplessy, 1986: Melting history of Antarctica during the past 60,000 years, *Nature*, 322: 701-706.

Levitus, S., J.I. Antonov, T.P. Boyer and C. Stephens, 2000: Warming of the World Ocean, *Science*, 287: 2225-2229.

Levitus, S. and T.P. Boyer, 1994: World Ocean Atlas, vol. 3, Salinity, 99 pp., Natl. Ocean and Atmos. Admin., Washington, D. C.

Levitus, S., R. Burgett and T.P. Boyer, 1994: World Ocean Atlas, vol. 4, Temperature, 117 pp., Natl. Ocean and Atmos. Admin., Washington, D. C.

Lorenz, S., B. Grieger, H. P. and K. Herterich, 1996: Investigating the sensivity of the atmospheric general circulation Model ECHAM 3 to paleoclimate boundary conditions., *Geol. Rundsch.*, 85: 513-524.

MacAyeal, D.R., 1992: Irregular oscillations of the West Antarctic ice sheet, *Nature*, 359: 29-32.

Maier-Reimer, E., U. Mikolajewicz and K. Hasselmann, 1993: Mean circulation of the Hamburg LSG OGCM and its sensitivity to the thermohaline surface forcing, *Journal of Physical Oceanography*, 23: 731-757.

Manabe, S. and R. Stouffer, 1997: Coupled ocean-atmosphere model response to freshwater input: Comparison to Younger Dryas event, *Paleoceanography*, 12: 321-336.

Manabe, S. and R.J. Stouffer, 1988: Two stable equilibria of a coupled ocean-atmosphere model, *Journal of Climate*, 1: 841-866.

Manabe, S. and R.J. Stouffer, 1995: Simulation of abrupt change induced by freshwater input to the North Atlantic Ocean, *Nature*, 378: 165-167.

Maslin, M., D. Seidov and J. Lowe, 2001: Synthesis of the nature and causes of rapid climate transitions during the Quaternary, *Geophysical Monograph (This Volume)*, D. Seidov, B.J. Haupt and M. Maslin (Editors), American Geophysical Union, Washington, D.C.

Maslin, M.A., N.J. Shackleton and U. Pflaumann, 1995: Surface water temperature, salinity, and density changes in the northeast Atlantic during the last 45,000 years: Heinrich events, deep water formation, and climatic rebounds, *Paleoceanography*, 10: 527-544.

McWilliams, J.C., 1998: Oceanic general circulation models, In: E.P. Chassignet and J. Verron (Editors), *Ocean Modeling and Parameterization*. Kluwer Academic Publishers, Boston, pp. 1-44.

MOM-2, 1996: Documentation, User's Guide and Reference Manual (edited by R. C. Pacanowski), GFDL Ocean Technical Report No. 3.2. Geophysical Fluid Dynamics Laboratory/NOAA, Princeton, N.J.

Oeschger, H., J. Beer, U. Siegenthaler, B. Stauffer, W. Dansgaard and C.C. Langway, 1984: Late glacial climate history from ice cores, In: J.E. Hansen and T. Takahashi (Editors), *Climate Processes and Climate Sensitivity. Geophys. Monogr. Ser.* American Geophysical Union, Washington D.C., pp. 299-306.

Oppenheimer, M., 1998: Global warming and the stability of the west Antarctic ice sheet, *Nature*, 393: 325-332.

Seidov, D. and B.J. Haupt, 1999b: Numerical study of glacial and meltwater global ocean, In: J. Harff, J. Lemke and K. Stattegger (Editors), *Computerized Modeling of Sedimentatry Systems*. Springer, New York, pp. 79-113.

Seidov, D. and M. Maslin, 1996: Seasonally ice freee glacial Nordic Seas without deep water ventilation, *Terra Nova*, 8: 245-254.

Seidov, D. and M. Maslin, 1999: North Atlantic Deep Water circulation collapse during the Heinrich events, *Geology*, 27: 23-26.

Seidov, D. and M. Maslin, 2001: Atlantic Ocean heat piracy and the bi-polar climate sea-saw during Heinrich and Dansgaard-Oeschger events, *Journal of Quaternary Science*, 16: in press.

Seidov, D., M. Sarnthein, K. Stattegger, R. Prien and M. Weinelt, 1996: North Atlantic ocean circulation during the Last Glacial

Maximum and subsequent meltwater event: A numerical model, *Journal of Geophysical Research*, 101: 16305-16332.

Stephens, B. and R. Keeling, 2000: The influence of Antarctic sea ice on glacial-interglacial CO_2 variations, *Nature*, 404,: 171-174.

Stocker, T.F., 1998: The seesaw effect, *Science*, 282: 61-62.

Stocker, T.F., R. Knutti and G.-K. Plattner, 2001: The future of the thermohaline circulation - a perspective, *Geophysical Monograph (This Volume)*, D. Seidov, B.J. Haupt and M. Maslin (Editors), American Geophysical Union, Washington, D.C.

Stocker, T.F., D.G. Wright and W.S. Broecker, 1992: The influence of high-latitude surface forcing on the global thermohaline circulation, *Paleoceanography*, 7: 529-541.

Stoessel, A., S.-J. Kim and S.S. Drijfhout, 1998: The impact of Southern Ocean sea ice in global ocean model, *J. Phys. Oceanogr.*, 28: 1999-2018.

Stommel, H. and A.B. Arons, 1960: On the abyssal circulation of the world ocean, I, Stationary planetary flow patterns on a sphere, *Deep Sea Research*, 6: 140-154.

Stommel, H., A.B. Arons and A.J. Faller, 1958: Some examples of stationary planetary flow patterns in bounded basins, *Tellus*, 10: 179-187.

Toggweiler, J.R., K. Dixon and K. Bryan, 1989: Simulations of radiocarbon in a coarse-resolution world ocean circulation model, 1, Steady state prebomb distribution, Journal *of Geophysical Research*, 94: 8217-8242.

Toggweiler, J.R. and B. Samuels, 1980: Effect of sea ice on the salinity of Antarctic Bottom Water, *J. Phys. Oceanogr.*, 25: 1980-1997.

Vaughan, D.G., J.L. Bamber, M. Giovinetto, J. Russel and A.P.R. Cooper, 1999: Reassessment of net surface mass balance in Antarctica, *Journal of Climate*, 29(4): 933-946.

Wang, X., P. Stone and J. Marotzke, 1999a: Global thermohaline circulation. Part I. Sensitivity to atmospheric moisture transport, *J. Climate*, 12: 71-82.

Wang, X., P. Stone and J. Marotzke, 1999b: Global thermohaline circulation. Part II. Sensitivity with interactive atmospheric transports, *J. Climate*, 12: 83-91.

Wang, Z. and L.A. Mysak, 2000: A simple coupled atmosphere-ocean-sea-ice-land surface model for climate and paleoclimate studies, *Journal of Climate*, 13: 1150-1172.

Wang, Z. and L.A. Mysak, 2001: Ice sheet-thermohaline circulation interactions in a climate model of intermediate complexity, Journal of Oceanography: in press.

Warrick, R.A., E.M. Barrow and T.M.L. Wigley (Editors), 1993. Climate and Sea Level Change: Observations, Projections and Implications. Cambridge University Press, N.Y.

Weaver, A., 1999: Millennial timescale variability in ocean/climate models, In: P.U. Clark, S.R. Webb and L.D. Keigwin (Editors), *Mechanisms of global climate change*. AGU, Washington, DC, pp. 285-300.

Weaver, A.J., 1995: Driving the ocean conveyor, *Nature*, 378: 135-136.

Weaver, A.J., S.M. Aura and P.G. Myers, 1994: Interdecadal Variability in an Idealized Model of the North Atlantic, *Journal of Geophysical Research - Oceans*, C99: 12423-12441.

Weaver, A.J., C.M. Bitz, A.F. Fanning and M.M. Holland, 1999: Thermohaline circulation: high latitude phenomena and the difference bewteen the Pacific and Atlantic, *Annual Review of Earth and Planetary Sciences*, 27: 231-285.

Weaver, A.J., M. Eby, A.F. Fanning and E.C. Wiebe, 1998: Simulated influence of carbon dioxide, orbital forcing and ice-sheets on the climate of the last glacial maximum., *Nature*, 394: 847-853.

Weaver, A.J. and T.M.C. Hughes, 1994: Rapid interglacial climate fluctuations driven by North Atlantic ocean circulation, *Nature*, 367: 447-450.

Weaver, A.J. and E.C. Wiebe, 1999: On the sensitivity of projected oceanic thermal expansion to the parameterisation of sub-grid scale ocean mixing, *Geophysical Research Letters*, 26: 3461-3464.

Weyl, P.K., 1968: The role of the oceans in climatic change: a theory of ice ages, *Meteorological Monographs*, 8: 37-62.

Zahn, R., J. Schonfeld, H.-R. Kudrass, M.-H. Park, H. Erlenkeuser and P. Grootes, 1997: Thermohaline instability in the North Atlantic during meltwater events: Stable isotopes and ice-rafted detritus from core SO75-26KL, Portuguese margin, *Paleoceanography*, 12: 696-710.

E.J. Barron, EMS Environment Institute, Pennsylvania State University, University Park, PA 16802. (eric@essc.psu.edu)

B.J. Haupt, EMS Environment Institute, Pennsylvania State University, University Park, PA 16802. (bjhaupt@essc.psu.edu)

M. Maslin, Environmental Change Research Centre, Department of Geography, University College London, 26 Bedford Way, London. WC1H 0AP, UK. (mmaslin@ucl.ac.uk)

D.Seidov, EMS Environment Institute, Pennsylvania State University, University Park, PA 16802. (dseidov@essc.psu.edu)