4.13 FEEDBACKS AFFECTING THE RESPONSE OF THE THERMOHALINE CIRCULATION TO INCREASING CO_2: A STUDY WITH A MODEL OF INTERMEDIATE COMPLEXITY.

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1 INTRODUCTION

Climate change simulations with coupled atmosphere-ocean models have shown that the ocean plays an important role in defining both transient and equilibrium responses of the climate system to changes in greenhouse gas (GHG) and aerosol concentrations in the atmosphere. There are, however, significant uncertainties in the oceanic response to an external forcing. The rate of heat uptake by the deep ocean differs from one ocean model to another (Cubash et al, 2001; Murphy and Mitchell, 1995; Sokolov and Stone, 1998, Sokolov et al., 2001). The causes and magnitude of weakening of the thermohaline circulation, which can lead to fundamental changes in the state of the climate system, is also rather different between different coupled models (Dixon et al., 1999; Cubash et al, 2001). Understanding of the dynamics of the changes in atmospheric and oceanic states then becomes crucial for estimating the likelihood of any particular prediction.

Several global-change simulations with coupled atmosphere-ocean models show the North Atlantic branch of the thermohaline circulation (THC) and associated heat transport weakening during the period of transient warming of the atmosphere (Cubash et al, 2001). The meridional overturning can even completely cease to exist, if external forcing is strong enough (Stouffer and Manabe, 1999). An increase in both the atmospheric moisture and heat transports leads to decrease in the meridional mass transport in the North Atlantic. It is however less obvious which of these transports is more important in causing the decay in the overturning. Changes in freshwater fluxes are claimed to be a main reason for the decay in THC in a number of studies (Manabe and Stouffer, 1994; Schmittner and Stocker, 1999; Dixon et al., 1999, Wiebe and Weaver, 1999). Mikolajewicz and Voss (2000) in contrast, report that heat flux changes were mainly responsible for THC weakening in their studies; see Dixon et al. (1999) for more details. In contrast to these studies, Latif et al. (2000) report no significant change in the THC during the period of the transient warming of the atmosphere due to an anomalous moisture flux from the tropical Atlantic, which acts to sustain the meridional circulation in the basin. The lack of agreement between different coupled models calls for a detailed analysis of the individual feedbacks in the system, as well as for an assessment of the methods used in the analysis.

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While many studies show that the THC starts to recover its strength after the increase in CO_2 concentration stops, the dynamics of the process has received little attention in the literature, with the exception of the study by Wiebe and Weaver (1999). In that study, the authors report the increase in the meridional gradient of the equilibrium steric height that is *consistent* with the intensified overturning in the Atlantic and attribute it to the low-latitude warming. The latitudinal distribution of the warming is, however, a product of several complicated feedbacks between surface forcing and the oceanic redistribution of heat, and a further investigation of these mechanisms is necessary. The dynamics of the recovery is likely to be dominated by the ocean, which also drives the recovery time scale (Manabe et al, 1991). Manabe and Stouffer (1994) suggest that the circulation recovery in their study is due to the removal of freshwater anomalies from the surface by circulating waters. Internal dynamics of the ocean, such as subgrid mixing of heat and salt by eddies, also plays a potentially equally important role in determining the temporal behavior of the circulation. Wiebe and Weaver (1999) also report a melting of the sea ice and an associated decrease in the surface ocean area insulated by ice; this process leads to both larger heat loss and freshwater gain in the area, which tend to counteract each other.

The present study analyzes the transient evolution of the system during the first 200 years after the increase in the CO_2 concentration stops, as well as during the increase. We estimate the contribution of the surface heat forcing to the recovery of the circulation, as well as investigate the contribution of oceanic feedbacks to the thermal surface forcing. We show that the ocean's role during the recovery is very different from that during the period of the increasing CO_2 concentration. Unlike the period of transient buildup of the atmospheric CO_2, during which the changes in the ocean are forced by the atmosphere, the circulation recovery at later stages is governed by the ocean itself. After stabilization the changes in atmospheric heating are to a large degree a response, not a cause of the changes in the oceanic circulation.

An extensive analysis of the feedbacks in a coupled system requires a large number of numerical experiments, which are not possible with detailed and computationally expensive models. Here, we use a model of intermediate complexity that combines computational efficiency with explicit representation of atmospheric and oceanic circulation and sophisticated parameterizations of the main physical and dynamical processes. The model is the climate component of the MIT Integrated Global System Model (Prinn et al., 1999) and includes a zonal-mean statistical dynamical atmosphere, 3D ocean and thermodynamic ice model. We give a brief description of the model in section 2. In section 3, we concentrate on the behavior of the thermohaline circulation and on the mechanisms causing its decay during the period of increasing CO_2 concentration and its subsequent recovery following stabilization with constant CO_2. The discussion and summary are presented in section 4.

2 MODEL DESCRIPTION

We use a coupled atmosphere/ocean model of intermediate complexity. The model matches sophisticated General Circulation Models (GCMs) in the number of dynamical processes, but achieves computational efficiency by coarse resolution. The model components, spin-up, and coupling procedure are described in detail in Kamenkovich, et al (2002). Here we give only a brief description.

The atmospheric component (Sokolov and Stone, 1998) was developed from the GISS GCM (Hansen et al, 1983). It solves the zonally averaged primitive equations as an initial value problem and includes parameterizations of all the main physical processes, radiation, convection, clouds, etc . It also includes parameterizations of heat, moisture, and momentum transports by large-scale eddies (Stone and Yao, 1987, 1990); and has complete moisture and momentum cycles. The model therefore can capture most of the atmospheric feedbacks rather realistically. The model allows four different types of surfaces in the same latitude belt, namely open ocean, sea-ice, land, and land-ice. The surface characteristics as well as turbulent and radiative fluxes are calculated separately for each kind of surface, while the atmosphere above is assumed to be well mixed zonally. The atmosphere model uses a realistic land/ocean ratio for each latitude. Land processes are represented by a two layer bucket model. Thermal conductivity, heat capacity and field capacity depend on soil type and vegetation. Runoff depends on ground wetness and precipitation rate. The detailed description of the model can be found in Sokolov and Stone (1998); and Prinn et al. (1999). A thermodynamic ice model is used for representing sea ice. The model has two layers and computes ice concentration (the percentage of area covered by ice) and ice thickness. It forms and melts ice on the bottom and on the side of the ice layer. The atmospheric model coupled to a mixed layer diffusive ocean model has been tested in a number of equilibrium and transient climate change simulations. It was shown that changes in both the global averages and spatial distributions of such climate variables, as temperature, precipitations, surface heat fluxes are similar to those obtained in the simulations with more sophisticated GCMs (Sokolov and Stone, 1998; Prinn et al., 1999: Sokolov et al., 2001).

The ocean component is based on the MOM2 GFDL model. The geometry is idealized and has two basins, "Atlantic" and "Pacific" connected by an "ACC" channel. There is no topography except a sill in the "Drake Passage"; the advantages and shortcomings of the idealized geometry are discussed in Kamenkovich, et al (2002). The Gent-McWilliams eddy parameterization scheme is used in all runs described in this paper. The model was spun up with a surface heat flux given by a sum of climatological heat fluxes and relaxation of surface temperature to its observed values. The surface boundary condition of heat is therefore a Haney-type boundary condition (Haney, 1971) linearized around SST. The condition allows for climatological estimates of heat flux to be recovered when the model reproduces observed SST (Jiang et al., 1999; Paiva and Chassignet, 2001), whereas the widely used relaxation of SST to its observed values leads to errors in the SST in regions of strong advection (Killworth et al, 2000) and to errors in the surface heat flux. Climatological surface moisture and momentum fluxes, taken from Jiang, et al. (1999), were used for the remaining surface boundary conditions. Moisture fluxes include river runoff and ice-calving data. We note that our choice of mixed boundary conditions eliminates unrealistic dependence of the surface freshwater fluxes on surface salinity and prevents surface moisture fluxes from being unrealistic (Jiang et al., 1999). Observed fluxes and SST used to spin up the model were zonally averaged over each of the two basins.

During the coupled regime, total surface heat and fresh-water fluxes into the ocean are:

$$F_H = H_o(1-\gamma) + H_i\gamma + [\,\partial H_o/\partial T\,](T-\overline{T})(1-\gamma) \quad (1)$$

$$F_w = F_o(1-\gamma) + F_i\gamma + [\,\partial F_o/\partial T\,](T-\overline{T})(1-\gamma) \quad (2)$$

where H_o and F_o are the zonally averaged heat and fresh-water fluxes over the open ocean, $\partial H_o/\partial T$ and $\partial F_o/\partial T$ their derivatives with respect to SST, H_i is the heat flux through the bottom of sea-ice, γ is the fractional ice area (ice concentration), F_i the fresh water flux due to ice melting/freezing, and $\overline{T}$ is the zonal mean of SST.

The last terms on the right-hand sides of the equations allow for zonal variations in surface fluxes. The near surface air temperature adjusts to changes in surface fluxes relatively fast, and inaccurate calculation of the derivatives $\partial H_o/\partial T$ and $\partial F_o/\partial T$ can lead to overestimation of the sensitivity of local heat fluxes to changes in SST. To account for this, we calculate the derivatives under assumption of exchange coefficients being fixed. That yields global averaged monthly mean values of $\partial SH/\partial T$ and $\partial LH/\partial T$ of 8-9 $Wm^{-2}K^{-1}$ and 17-19

$Wm^{-2}K^{-1}$, respectively. The values of $\partial SH/\partial T$ and $\partial LH/\partial T$, as well as their dependence on season and latitude are similar to those given by Rivin and Tziperman (1997). Possible changes in downward longwave radiation associated with changes in near surface air temperature and humidity are not taken into account. The last term in Eq. (2) represents variations in evaporation only, i.e., there are no longitudinal variations in precipitation in our model. To the extent that longitudinal variations in precipitation compensate longitudinal variations in evaporation, Eq. (2) can overestimate longitudinal variations in freshwater flux. Rivin and Tziperman (1997) however show that sensitivity of evaporation to changes in SST is about three times as large as that of local precipitation. Only globally averaged changes in river runoff are taken into account.

Similar to more sophisticated models, intermediate models are not as easily tunable as conceptual models and often requires flux adjustments (see a review in Claussen et al, 2001). Since the computational efficiency of our model is achieved not by simplification of physics, but by decreased resolution, the model climate drifts without flux adjustments; see a more detailed discussion of this subject in Kamenkovich et al (2002). Flux adjustments are used throughout this study and are calculated as differences between the values of heat, moisture and momentum fluxes used to spin up the ocean model and the corresponding values obtained in the simulation with the atmospheric model forced by observed SST and sea ice. Similarly, the atmosphere is forced by anomalies in the ocean generated SST added to the SSTs used to spin up the atmospheric model. Since the ocean model extends only from 64°S to 72°N, SST and sea ice outside of this area are simulated by a mixed layer ocean model.

Under present day conditions, the model surface reaches a near equilibrium state within first 200 years with zonal mean SST deviating from its initial values by less than 1°C if flux adjustments are used; see Kamenkovich et al (2002). The meridional overturning reaches a maximum of 26Sv, with waters sinking near the northern and northeastern boundaries of the "North Atlantic"; (stands for Northern Hemisphere Atlantic, NA hereafter). The annual-mean SST in this area varies between 5 and 6 degrees, being somewhat warmer than corresponding Levitus values in the Greenland-Norwegian seas. Due to simplified geometry and coarse resolution, the formation of the cold Labrador Sea Water (Talley and McCartney, 1982) is not captured in the model, which also makes forming deep water warmer than in reality. The poleward heat transport in the "Atlantic" reaches 1.06 PW at its maximum at 14°N, which is within the range of the available estimates (Trenberth et al 2001).

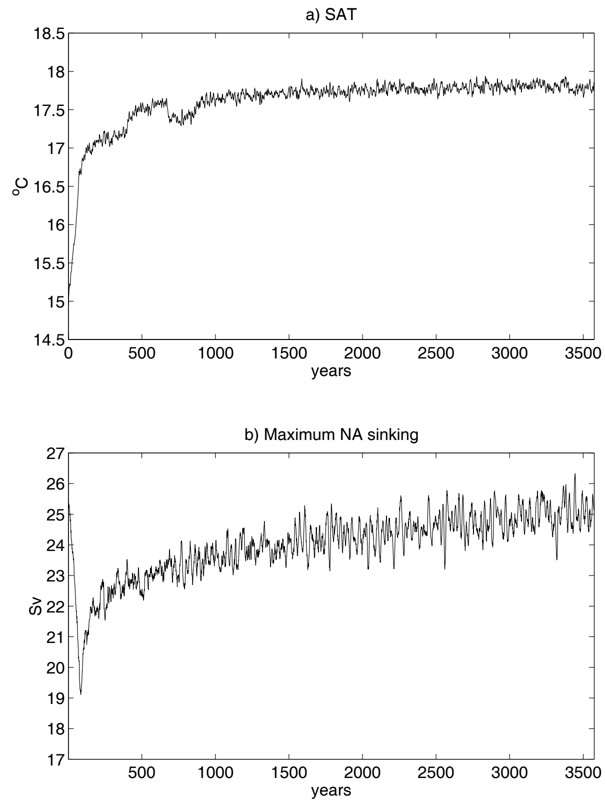

Fig.1 a) The global mean surface air temperature (SAT) for GM2CO2; the curve is smoothed with a 5-year sliding mean. CO_2 concentration increases at the rate of 1% per year for the first 75 years, and is held constant afterwards. b) The evolution in time of the maximum (subsurface) overturning in the "North Atlantic".

3 EVOLUTION OF THE MERIDIONAL OVERTURNING CIRCULATION

In this section, we describe and analyze results of a global change simulation, in which the atmospheric CO_2 concentration increases at the rate of 1% per year (compounded) for 75 years reaching the concentration of 2.11 times the present-day value by the end of the year 75 (experiment GM2CO2). The concentration is kept constant after that and the model is integrated for additional 3425 years until it comes close to equilibrium with the globally averaged annual mean surface air temperature increasing by about 2.5°C (fig.1a). We now turn our attention to the mechanisms controlling evolution of the meridional overturning.

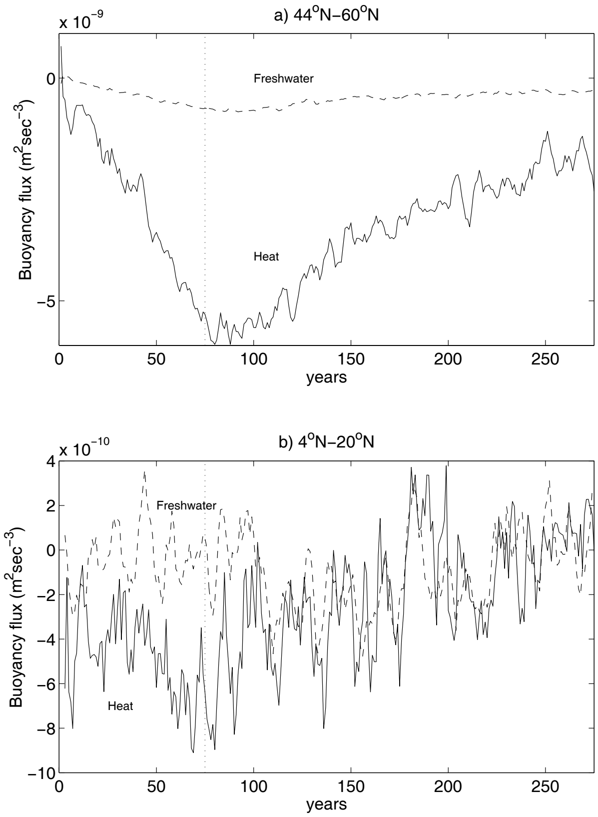

Fig.2 Changes in the area-averaged surface buoyancy flux due to heat (solid) and moisture (dashed) flux forcing in NA expressed in the units of the buoyancy flux. The buoyancy flux is defined as $(gd_1/\rho)(d\rho/dt)$ and is averaged over the region: a) between 44°N and 60°N; b) between 4°N and 20°N. Units are m^2sec^{-3}.

3.1 Global change simulations with increasing CO_2.

The thermohaline circulation (THC) significantly slows in response to increasing atmospheric CO_2 concentration, but does not cease completely. The NA sinking (fig.1b) decreases by 5Sv (20%) by the end of the period with increasing CO_2 concentration. Warming of the atmosphere leads to increases in the heat and fresh water fluxes into the northern NA. Amplification of each of these fluxes acts to decrease surface density in the region and meridional density contrast with low latitudes slowing the sinking. Our task is to determine which of the fluxes is the dominant reason for the THC weakening. We start by converting both fluxes into the units of a buoyancy flux, which is defined as $(gd_1/\rho)(\partial\rho/\partial t)$, where ρ is the density and d_1 is the thickness of the first layer in our model. The annual means of changes in these fluxes averaged over the area north of 44°N in the “Atlantic” are plotted against time in fig.2a. The averaging area includes a region to the south of the convection sites in the northern NA (Section 2) since these regions are crucial in setting meridional density gradients and controlling northward advection of heat and salt; analysis of the area northward of 55°N produces a qualitatively similar picture. It is clear, that the increase in buoyancy due to changes in the heat fluxes dominates over the effects of the freshwater fluxes and is likely to be the main reason for the slowdown of the THC in the “Atlantic”. These changes in the heat fluxes are caused by the direct radiative forcing associated with an increase in CO_2 concentration as well as an increase in the latitudinal heat transport, mainly transport of latent heat, by the atmosphere. Changes in the zonal heat transport from the “Pacific” to the “Atlantic” also play an important role in the modification of the heat fluxes in the “Atlantic”, as will be discussed in Section 3.3. We note that the dominance of the heat flux in the buoyancy forcing of our model would not change if the SST in the sinking region of the NA were more realistic. Also it is noteworthy that the magnitude of changes in the heat and moisture fluxes in low latitudes, shown for the 4°N-20°N band in fig.2b, are similar in buoyancy units and are both much smaller than changes in the heat flux north of 44°N (fig 2a). The changes in the moisture fluxes do not have a consistent pattern in time, exhibiting strong oscillations around the zero line. It should be kept in mind that local changes in runoff are not taken into account.

To study the relative roles of different surface fluxes in weakening the circulation, we conduct additional experiments with the ocean-only model forced by surface fluxes diagnosed from coupled runs. Our uncoupled approach is different from the partial coupling used by Mikolajewicz and Voss, (2000) and Dixon et al., (1999), in which some fluxes are prescribed, while others are simulated in a fully coupled mode. Although the partially coupled approach works equally well in the studies of the circulation evolution during changing CO_2 concentration, an analysis of the circulation recovery during the period of fixed concentration is more problematic, since, for example, it is not possible to restrict changes in atmospheric heating. In contrast, our uncoupled method is especially convenient for studying the mechanisms of the recovery described in the next section. To increase confidence in the conclusions of this section, we will also carry out partially coupled simulations and compare their results with those from the uncoupled runs.

Runs in which the heat flux forcing of the ocean is taken from an unequilibrated segment of the coupled model can, however, present a problem. Since the specified heat flux is not equilibrated, i.e., there is a net heat flux into or out of the ocean, the ocean cannot reach an equilibrated state. To allow the ocean to reach equilibrium, in these runs the zonal-mean atmospheric heat flux into the ocean H_o (keeping the same notation as in equation 1) is represented by

$$H_o = H_o^c + \lambda\,(\overline{T}^{\,c} - \overline{T}\ ), \qquad (3)$$

where H_o^c and $\overline{T}^{\,c}$ are zonal mean surface heat flux and SST taken from a segment of an unequilibrated coupled run and T is the SST calculated by the ocean model. Because of the relaxation term, the global mean heat flux can now equilibrate to zero, whereas in the absence of this term, an external heating would be maintained in the system and equilibration is impossible. The value of λ depends on the interpretation of the feedback term. If we assume a purely long-wave radiative adjustment of the oceanic surface to changes in the SST, the climate feedback parameter becomes $\lambda = 4\sigma T^3$, where σ is the Stefan-Boltzman coefficient, and T is the SST. The resulting values are in the range between 4.5 and $6.2\ Wm^{-2}K^{-1}$. These values are close to the value of $[\ \partial H_o / \partial T\ ]$ (Eq. 1) calculated under an assumption that atmospheric temperature adjusts instantaneously to changes in SST. This approach, however, essentially neglects any role for the atmospheric (e.g. cloud, water vapor) feedbacks. Taking these feedbacks into account leads to the following alternative estimate of λ. In a fully coupled simulation with our model, doubling of the atmospheric CO_2 concentration corresponds to forcing of about $4Wm^{-2}$. Since the resulting SST increase is approximately 2.5^oC, the λ estimated this way equals $4/2.5 = 1.6\ Wm^{-2}K^{-1}$. While an equilibrium response will, obviously, depend on the chosen value of λ, our simulations showed almost no such dependency during either the slowing down or the initial stage of recovery of THC. In what follows, we show only the results with the first choice of λ.

We calculate H_o from equation (3) using zonally averaged heat fluxes and SST from a coupled simulation. The resulting value of H_o is used in equation (1); the derivatives $\partial H_o/dT$ and $\partial F_o/dT$ in the zonal transfer terms (see equations 1 and 2) are taken from the corresponding coupled simulations (see Table 1). The zonal deviations of SST from its zonal mean are calculated on every step by the ocean model; we therefore allow feedback of the zonal variations in SST on local heat flux anomalies.

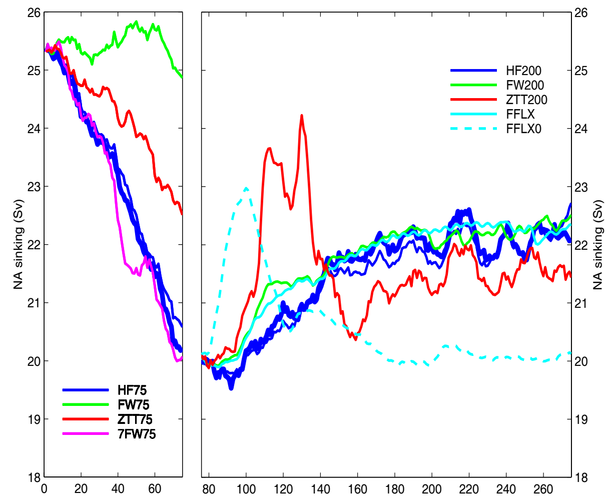

Fig.3 Evolution in time of the maximum (subsurface) overturning in the "North Atlantic" for the years 1-275 for experiments described in the Table 1. The thick line shows control case GM2CO2.

In the first uncoupled simulation (FW75) we use H_o^c, $\overline{T}^{\,c}$, $\partial H_o/\partial T$ and momentum fluxes taken from the reference run; the F_o and $\partial F_o/\partial T$ are taken from years 1-75 of the GM2CO2. The NA sinking in this case (fig.3) remains nearly constant, closely following the values in the reference run. We also performed a partially coupled run (see above discussion), in which the CO_2 concentration is kept constant at its reference level, whereas freshwater flux is diagnosed from the coupled global change run. As in our uncoupled experiment, the circulation stays very close to its initial value (not shown).

In the second uncoupled experiment HF75, F_o, $\partial F_o/\partial T$ and momentum fluxes are taken from the reference run, while the H_o^c, $\overline{T}^{\,c}$ in (3) and $\partial H_o/\partial T$ are taken from the corresponding global change experiments. The decrease of the NA sinking in this case matches that in the global change runs (fig.3). The behavior of the circulation does not noticeably change even if λ is set to zero, which is explained by the smallness of deviations of SST from the corresponding values in GM2CO2. The heating is therefore the surface forcing responsible for the weakening of the circulation. Similar results were obtained in a study by Mikolajewicz and Voss (2000), who used the Max-Plank Institute model, and reported that changes in surface heat fluxes are the main reason for the decrease in the THC. The results do not change significantly when we repeat this experiment with heating only north of the Equator taken from GM2CO2 and the rest of the forcing diagnosed from the reference

	Heat Flux			Freshwater Flux		ZASST	Wind stress
	H_o in eq. (1)	***$\partial H_o/dT$ in eq.(1)***	***H_o^c, T_c In eq.(3)***	***F_o In eq.(2)***	***$\partial F_o/dT$ in eq.(2)***		
GM2CO2 (1-275)	COUPLED CO_2 changes	COUPLED CO_2 changes	N/A	COUPLED CO_2 changes	COUPLED CO_2 changes	Ocean model	COUPLED CO_2 changes
REF (1-75)	COUPLED CO_2 fixed	COUPLED CO_2 fixed	N/A	COUPLED CO_2 fixed	COUPLED CO_2 fixed	Ocean model	COUPLED CO_2 fixed
HF75 (1-75)	Eq. (3)	GM2CO2 Yr. 1-75	GM2CO2 Yr. 1-75	REF Yr. 1-75	REF Yr. 1-75	Ocean model	REF Yr. 1-75
FW75 (1-75)	Eq. (3)	REF Yr. 1-75	REF Yr. 1-75	GM2CO2, Yr. 1-75	GM2CO2, Yr. 1-75	Ocean model	REF Yr. 1-75
7FW75 (1-75)	Eq. (3)	REF Yr. 1-75	REF Yr. 1-75	GM2CO2, Yr. 1-75, Changes Amplified	GM2CO2, Yr. 1-75	Ocean model	REF Yr. 1-75
ZTT75 (1-75)	COUPLED CO_2 changes	REF Yr. 1-75	N/A	COUPLED CO_2 changes	REF Yr. 1-75	REF, Yr.71-75	COUPLED CO_2 changes
HF200 (76-275)	Eq. (3)	GM2CO2 Yr.76-275	GM2CO2 Yr.76-275	GM2CO2, Yr. 71-75	GM2CO2, Yr. 71-75	Ocean model	GM2CO2, Yr.71-75
FW200 (76-275)	Eq. (3)	GM2CO2 Yr. 71-75	GM2CO2 Yr. 71-75	GM2CO2, Yr. 76-275	GM2CO2, Yr. 76-275	Ocean model	GM2CO2, Yr. 76-275
FFLX (76-275)	Eq. (3)	GM2CO2 Yr. 71-75	GM2CO2 Yr. 71-75	GM2CO2 Yr. 71-75	GM2CO2 Yr. 71-75	Ocean model	GM2CO2 Yr. 71-75
FFLX0 (76-275)	Eq. (3) with λ=0	GM2CO2 Yr. 71-75	GM2CO2 Yr. 71-75	GM2CO2 Yr. 71-75	GM2CO2 Yr. 71-75	Ocean model	GM2CO2 Yr. 71-75
ZTT200 (76-275)	COUPLED CO_2 fixed, doubled	GM2CO2, Yr.76-80	N/A	COUPLED CO_2 fixed, doubled	GM2CO2, Yr.76-80	GM2CO2 Yr.76-80	COUPLED CO_2 fixed, doubled

Table 1.
Experiments reported in fig.3 with their names given in the table (with the simulated years in brackets). Surface fluxes are either taken from a fully coupled run (its name and corresponding years are in the table) or computed in a coupled mode ("COUPLED"). The equation 3 is not used when heat fluxes are computed in a coupled mode, as in GM2CO2, REF, ZTT75 and ZTT200. Zonal anomalies in SST, or ZASST are $\overline{T}$ - T in the equations (1) and (2) and are either calculated by the ocean model ("ocean model" in the table) or taken from another coupled run as in ZTT75 and ZTT200.

experiment. The resulting values of THC (not shown) again closely follow those in the global change experiment. Northern Hemisphere heat forcing is therefore the main factor causing the decrease of the THC in our simulations.

It is interesting to explore a little further the role of the moisture fluxes and the sensitivity of the THC to their changes. The relatively small magnitude of the changes in the buoyancy flux due to changes in the moisture fluxes in the mid- to high latitudes shown in fig.2a is not necessarily the only reason for their secondary role. Even if magnitudes were the same, the sensitivity of the THC to changes in heat and moisture fluxes may be different because of the different spatial structure of salinity and temperature. What would happen in FW75 if changes in the moisture fluxes in density units were as large as changes in the heat fluxes in 2CO2GM? Will the decrease in the THC be the same as in the global change GM2CO2 run? To address this question, we repeat FW75 with the changes in moisture fluxes (taken from GM2CO2) multiplied by a factor of 7.5. The effects of the increase of moisture fluxes on density in this run, 7FW75, are then similar in magnitude to that of the heat fluxes in GM2CO2. The circulation (fig.3) declines similarly to GM2CO2 except a period during years 40-55 when it first falls faster then recovers and continues following the evolution in GM2CO2. Despite the obvious differences between HF75 and 7FW75, one can conclude the THC is mostly responsive to the magnitude of surface buoyancy flux and not very sensitive to whether changes in the surface buoyancy fluxes are caused by the heat or moisture fluxes. We also carried out a partially coupled experiment with fixed CO_2 concentrations, in which we force the ocean model with changes in moisture fluxes taken from the GM2CO2 run and multiplied by a factor of 7.5. The circulation (not shown) weakens somewhat faster than in 7FW75, which is most likely explained by positive feedbacks from an active atmosphere in the partially coupled case (Nakamura et al., 1994), but the overall behavior is very similar.

From the results of 7FW75, one may conclude that the moisture fluxes would have the same effect on the circulation as the heat fluxes, were the buoyancy amplitude of the changes in moisture fluxes as large as that of the heat fluxes. However, although the global heat flux H_o^c used in 7FW75 is equal to zero, the NA heat flux is increasing in 7FW75 (fig.4a), which is caused by cooling of the NA (fig.4c) due to decrease in the meridional circulation. The NA cooling leads to anomalous heating of the NA due to both the feedback term (Equation 3) and an inter-basin zonal heat flux into NA from the “North Pacific” (Equation 1). In other words, changes in the circulation caused by freshwater fluxes bring in changes in surface heating despite our attempt

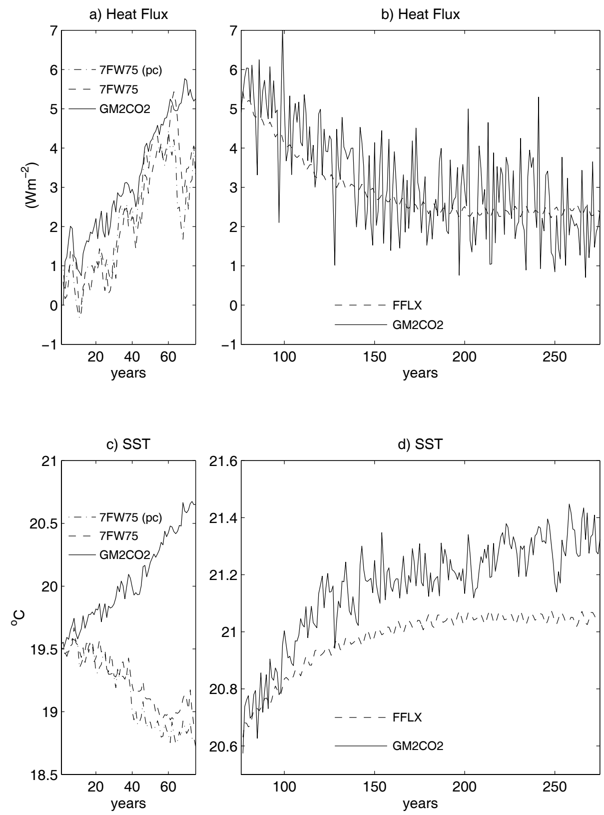

Fig.4 Annual mean heat flux change averaged over the NA (north of the equator): a) GM2CO2 (solid), 7FW75 (dashed), partially coupled equivalent of 7FW75 (dashed-dotted), smoothed for better presentation; b) GM2CO2 (solid), FFLX (dashed); c) as in (a) but for annual mean SST averaged over the NA, not smoothed; d) as in (b) but for the NA SST. Note different scales between (c) and (d).

to keep heating fixed. Simultaneous analysis of the heat fluxes is always needed in similar experiments, uncoupled or partially coupled alike.

3.2 Simulations with constant doubled CO_2.

The recovery of the NA sinking begins only about 10 years after the end of the CO_2 increase (fig.1b). The quick start of the recovery is very similar to that in Wood, et al. (1999) in their model with no flux adjustments, which also uses the GM scheme. By the year 1000, the NA sinking recovers to 23Sv, which is 92% of its initial value; the circulation fully recovers its strength during the next 2500 years of integration.

We now consider the mechanisms of the recovery of the circulation in our model after the CO_2 concentration in the atmosphere stabilizes. During the period of fixed doubled CO_2 concentration (after year 75), the surface air temperature continues to rise. However, the surface buoyancy fluxes due to anomalous heat and moisture fluxes each decrease in magnitude in the NA, as fig.2 demonstrates. As during the years 1-75, the changes in heat fluxes clearly dominate over changes in the moisture fluxes. This decrease in the strength of the anomalous surface forcing is coupled with the recovery of the circulation. In this section, we evaluate the role of atmospheric moisture and heat fluxes in forcing the recovery. Since changes in the heat forcing are in large part induced by changes in SST due to evolving oceanic circulation, we also want to estimate the contribution of the ocean to modulating surface heat fluxes. In particular, does the ocean assume a leading role in the process by the circulation recovering on its own with no significant help from the atmosphere? Does, alternatively, the atmosphere force the ocean to recover its circulation?

To check if heat fluxes are a primary cause for the circulation changes, we conduct uncoupled experiments similar to the uncoupled experiments of section 3.1. It should be noted, that equilibrium is not achieved in these experiments, and our only goal is to reproduce the transient behavior during the initial 200 years after CO_2 stops increasing. In the first experiment (HF200 in fig.3), we force the ocean model by the H_o^c, $\overline{T}^c$ in (3) and $\partial H_o/\partial T$ taken from years 76-275 of the 2CO2GM. For F_o, $\partial F_o/\partial T$ and wind stress, we take values from the 5-year mean computed for years 71-75 of GM2CO2; the seasonal cycle is retained. The initial conditions are taken from the end of the year 75. The circulation very closely repeats its behavior in the fully coupled experiment. The changes in the heat flux are therefore essential for the circulation's further evolution. It is noteworthy that the evolution of the circulation does not noticeably depend on the value of λ, with nearly identical behavior even if λ is zero, which as in the previous section is explained by the smallness of deviations of SST from the corresponding values in GM2CO2.

Our next step is to run an uncoupled experiment with the ocean model forced by the F_o and $\partial F_o/\partial T$ taken from a years 76-275 and H_o^c, $\overline{T}^c$, $\partial H_o/\partial T$ and wind stress taken from the time mean for years 71-75 (run FW200 in fig.3). In this run, the last term on the right-hand side of (3) has the form of a weak relaxation of surface temperature toward its values at the end of coupled run with increasing CO_2, when the THC is still weakening. The SST is therefore not constrained to follow its values in the GM2CO2 run and this term just allows a feedback of the ocean surface on heat fluxes. The THC recovery is as strong as that in GM2CO2 in this run. Note that H_o^c and $\overline{T}^c$ are fixed in FW200, and all changes in the surface heating can only come from the evolving ocean. We next set the λ in (3) to zero, which, as discussed in section 3.1, corresponds to maintaining external heating. The global SST at the end of the run increases by more than 3 degrees from its value at year 75, whereas the SST increases by only 0.8 degrees during years 76-275 of GM2CO2. This continuing strong warming suppresses the recovery of the circulation in the case of λ equal to zero (not shown here).

Is the recovery then a process, in which the ocean plays the leading role by forcing the changes in the surface heating through oceanic feedbacks on SST? To address this question, we run an additional experiment FFLX, in which *all* surface forcing variables, H_o^c, $\overline{T}^c$, $\partial H_o/\partial T$, F_o, $\partial F_o/\partial T$ and wind stress, are taken from time means for years 71-75 of GM2CO2. The atmospheric forcing is not changing, but the ocean can alter heat fluxes due to the evolution of the circulation. The THC recovery is remarkably similar to that in FW200 and its rate is fairly close to that in GM2CO2. We can conclude that freshwater forcing is not an important term in the recovery of the circulation, which takes place mainly as an oceanic process, with an additional contribution from the equilibrating atmosphere (note the difference between GM2CO2 and FFLX). It is noteworthy that the circulation fails to recover in the repeat of FFLX with zero oceanic feedbacks, $\lambda=0$ (FFLX0 in fig.3). The overturning abruptly increases up to the year 100 but then quickly drops back. The circulation seems to have an initial tendency to regain its strength. This recovery is then suppressed by continuing strong warming of the oceanic surface, since fixing heat fluxes implies maintaining atmospheric heating forcing (section 3.1). The SST feedbacks are crucial for the ability of the model to equilibrate and recover its circulation.

The evolving ocean circulation alters SST and, through the radiative feedback introduced in the experiment, the surface heat fluxes. The globally averaged surface heat fluxes in FFLX in fact evolve similarly to that in GM2CO2 (fig.4b). One notable difference, however, is that the interannual variability in the global heat uptake by the ocean in FFLX is considerably reduced compared to GM2CO2 (fig.4b). The variability is therefore essentially a coupled process with the atmosphere providing a positive feedback for the surface heat flux anomalies (see also Nakamura et al., 1994. Saravanan, 1998), while our simple radiative ocean feedback acts to damp the heat flux anomalies. The interannual variability was retained in the heat fluxes when they were used in the uncoupled

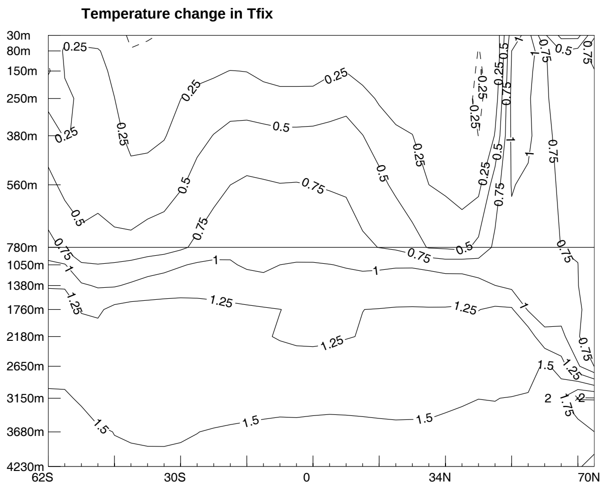

Fig.5 Changes in zonal mean Atlantic temperature during equilibration in TFIX: difference between decadal mean for equilibrium in TFIX and the annual mean for year 75 of GM2CO2.

experiments of the previous section; smoothed values are shown in the fig.4a.

In all experiments described in this section, the SST was allowed to substantially deviate from its values at year 75, since the feedback parameter λ corresponds to a long relaxation time scale. The SST changes substantially during the course of the equilibration even in the runs with fixed $\overline{T}^c$, although changes are smaller than in GM2CO2 (fig. 4d). One is tempted to conclude that the SST structure at year 75 is the *cause* of the weakened circulation and the SST evolution is necessary for the recovery. The bulk of the ocean, however, is not in equilibrium at year 75, and the steady-state THC corresponding to the same distribution of SST as at year 75, can be very different. To determine the strength of this steady-state THC, we repeat run FFLX with λ set to a large number of $42\ Wm^{-2}K^{-1}$ and T_c taken from year 75 (run TFIX), and therefore force the SST to stay close to its value at year 75. We then run the ocean model to equilibrium for 3,500 years and observe that the final value of the circulation intensity is 25.5 Sv, which is very close to the THC intensity in the present-day model climate. The increase in the THC strength is accompanied by the evolution of the temperature structure. Although the "Atlantic" SST changes very little south of 50^{o}N, subsurface temperature undergoes a significant adjustment, as seen in Figure 5 that shows the difference in zonal mean Atlantic temperature between the final state of TFIX and year 75 of GM2CO2. The warming increases toward the equator, which enhances the meridional density gradient in the NA and is consistent with intensifying meridional circulation (see also Wiebe and Weaver, 1999). It is noteworthy, that the high latitudes in NA warm in TFIX, which is a consequence of increasing meridional circulation but acts as a negative feedback for intensifying convection.

3.3 Role of the zonal surface exchanges of heat and moisture.

In our model, changes in the local heat and moisture fluxes come from two sources: from changes in the zonal-mean fluxes predicted by the 2D atmospheric model related to changes in zonal-mean SST; and as a result of the evolving zonal distribution of SST (equations 1 and 2). The latter effect mimics zonal transfers of heat and moisture that cannot be explicitly computed by our zonally averaged atmospheric model and guarantees that there is a zonal flux of heat and moisture down the zonal SST gradient. The magnitude of these transfers should however be compared against results from more sophisticated models.

Fig.6a shows the difference between the heat fluxes averaged between 44^{o}N and 60^{o}N in the "Pacific" and "Atlantic" basins (buoyancy flux units). This difference depends on the inter-basin SST contrast (Equation 1). During years 1-75, the latter gradually decreases as a result of the slowdown of the "conveyor belt" circulation, which maintains inter-basin temperature contrast in the Northern Hemisphere in our model. The result is anomalous zonal heat (and moisture) fluxes from the "Pacific" into the "Atlantic", which act to further weaken the overturning. In fact, the changes in the heat flux integrated over the 44^{o}N-60^{o}N latitude band in the "Atlantic" is 3 times as large as those changes over the same latitude band in the "Pacific" (not shown); the moisture flux changes are almost 3.5 times as large as in the "Pacific". This asymmetry in the evolution of the moisture fluxes between basins is very similar to that observed by Mikolajewicz and Voss (2000), who observed 70% of the total change in the moisture flux going into the "Atlantic". Therefore, the second term on the right hand side of equation 2 mimics the asymmetry in the distribution of freshwater flux anomalies between two ocean basins observed in at least one of the more sophisticated models. Analogously, the inter-basin SST contrast increases when "conveyor belt" circulation starts to intensify during the years 76-275, resulting in anomalous zonal fluxes of heat (fig.6a) and moisture from the "Atlantic" into the "Pacific", further enhancing the sinking in the former basin. The zonal transfer terms therefore act as positive feedbacks for changes in the NA sinking.

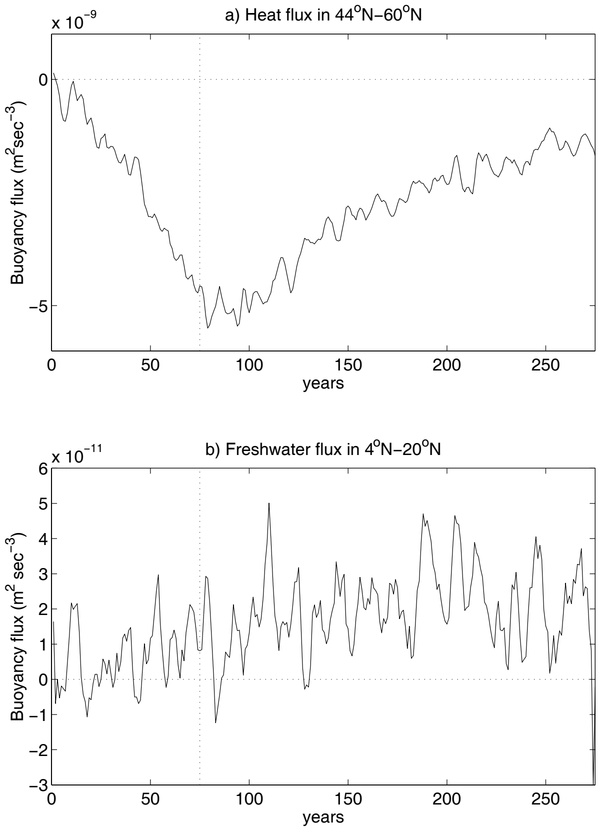

Fig.6 Difference between the "Atlantic" and "Pacific" surface buoyancy fluxes. Negative values imply positive buoyancy inter-basin transport into the "Atlantic" from the "Pacific" (heating or freshening of the "Atlantic"). a) Heat fluxes averaged between 44°N and 60°N; b) freshwater fluxes averaged between and 4°N and 20°N. The units are m^2sec^{-3}.

To evaluate the role of zonal transfer terms, we ran two additional coupled experiments, in which zonal transfer terms were not allowed to change, and $\partial H_o/\partial T$, $\partial F_o/\partial T$ and SST deviations from zonal mean (eqs.1 and 2) are taken from a fully coupled reference run. The first simulation is the repeat of the global change experiment GM2CO2 for years 1-75 with zonal transfer terms taken from the reference run with unchanging CO_2 concentrations. The decrease in the circulation is 2.5Sv smaller than that in the fully coupled run (ZTT75 in fig.3). Therefore, approximately half of the total decrease can be attributed to the additional flow of heat into the NA due to its cooling and the decrease in its SST contrast with the "North Pacific". Thus, the zonal heat fluxes supply a positive feedback to the THC during the initial decline. In a second experiment (ZTT200 in fig.3), we fix zonal transfer terms to their values for the years 76-80 (day-to-day variations are preserved) and run a coupled model for 200 years starting from the year 75 of GM2CO2 run. The circulation initially exhibits a sharp increase, but later drops back, and continues a slow recovery for the rest of the run. The overturning strength at year 275 is 0.5Sv smaller than that in the control GM2CO2 case. We can conclude that zonal transfer terms in the heat and moisture fluxes first tend to suppress the initial tendency of THC to recover, but then act to enhance the recovery of the THC in NA. Thus, the atmosphere zonal heat fluxes generally supply a positive feedback to changes in the THC, analogous to that associated with meridional heat fluxes (Nakamura et al. 1994).

Latif et al (2000) in their global warming study describe an anomalous moisture flux from the tropical "Atlantic", which acts to sustain the meridional circulation in the basin. This effect was attributed to the ENSO-like warming of the tropical "Pacific", leading to enhanced precipitation over the "Pacific", which in effect causes an anomalous moisture flux from the tropical "Atlantic" to the "Pacific". Although our model simulates an anomalous moisture flux out of the tropical "Atlantic" after year 50 (fig6b), the magnitudes of these anomalies are small (note different scaling in figs.6a and 6b), and unlikely to have a significant effect on the circulation. Our model does not simulate zonal variations in the precipitation, and it is therefore not surprising that the effect described in Latif et al (2000) is weak in our model.

4 DISCUSSION AND SUMMARY OF THE RESULTS

As in a number of other studies, the meridional overturning in the "North Atlantic" in our model initially weakens during the period of increasing CO_2 concentration in the atmosphere and then recovers when CO_2 concentration is held constant. For the analysis of the causes of these changes in the circulation, we carry out a number of uncoupled runs, in which the ocean model is forced by the surface fluxes taken from coupled runs. The ocean feedback on the surface fluxes is represented by a relaxation term (with the coefficient λ), which permits the ocean to alter surface forcing and equilibrate its heat uptake.

We demonstrate that the increasing heat fluxes in the global change runs are the main reason for the slowdown of the NA sinking in our model. Our conclusion agrees with the findings in Mikolajewicz and Voss (2000), but contradicts some other studies (Manabe and Stouffer, 1994; see also a summary in Dixon et al., 1999). The main reason for a secondary role of the moisture fluxes in causing changes in the meridional circulation in our model is their relatively small contribution to the surface buoyancy flux changes. We note that this is generally true for the models in Coupled Model Intercomparison Project (Huang et al., 2003). It is noteworthy, however, that local changes in the river runoff are not taken into account in our model, which could have affected the magnitude of the moisture fluxes in some regions. For example, Dixon et al (1999) report an important role of the increased river flow into the NA (north of 50^{o}N) in their model. When we artificially amplified changes in the moisture fluxes to make them comparable (in buoyancy units) to that of the heat fluxes, the drop in the circulation strength was similar to that in the coupled run, even though the heat flux in this experiment was taken from the present-day climate simulation. Our ocean model, therefore, appears nearly equally sensitive to the changes in heat and moisture fluxes. It is noteworthy, however, that changes in the ocean circulation in this run lead to an anomalous heat flux into the "Atlantic", which further weakens the overturning circulation. This positive feedback can complicate the analysis. Generally speaking, combining surface fluxes from two different runs, ocean only or partially coupled alike, works well only when one of these fluxes is the cause for circulation changes, as is the case in our study and in that of Mikolajewicz and Voss (2000). When contributions of both the moisture and heat fluxes to the weakening of the thermohaline circulation (THC) in the coupled run are significant, as in our experiment with amplified changes in the freshwater fluxes or in the study by Dixon et al. (1999), evaluation of their relative roles requires careful interpretation of the results.

Our uncoupled experiments led to several conclusions about the dynamics of the recovery of the THC after the increase in the atmospheric concentration stops:

1. The equilibration of the global surface heating is a necessary condition for the recovery in our model. To demonstrate this, we run an uncoupled experiment with the heat fluxes fixed to their values at the time of the doubling of CO_2 (years 71-75) without a feedback between the SST and the heat flux (*λ=0*), which is equivalent to maintaining a global heat forcing. In this run the circulation fails to recover.

2. An active atmosphere is not required for the recovery of THC. In the uncoupled experiments with the ocean feedbacks included and the heat fluxes fixed, the recovery is very similar to that in the fully coupled run. The SST-induced changes in the surface heat fluxes are in fact similar in magnitude to the changes in heating in the coupled experiment. One noteworthy difference is the substantially reduced interannual variability. The initial recovery is also insensitive to the magnitude of the feedback coefficient, although the final SST values are expected to depend on it.

3. The freshwater fluxes have a secondary role in the recovery, as the uncoupled runs with fixed and varying freshwater forcing lead to qualitatively similar results. The ability of the thermohaline circulation to completely recover its strength on "its own" through feedbacks on surface heating can, therefore, be a consequence of the dominant role of the heat fluxes. If freshwater forcing were to take the dominant role in causing changes in the circulation, as is the case in several model studies cited above, the role of the atmosphere may have been different, since the feedback of the ocean circulation on moisture fluxes is limited.

4. Despite importance of the surface heating, the evolution of the SST structure itself is not required for the recovery. The uncoupled experiment, in which the ocean model is driven to equilibrium with the SST distribution kept very close to that at the year 75, produces an overturning that is much stronger than at the year 75 and is very similar to that in the unperturbed model climate. The intensification of THC is explained by an increase in the meridional density gradients in the subsurface layers. This result emphasizes the transient nature of the ocean state at the year 75. Subsequent changes in the subsurface density structure during the following slow equilibration can lead to intensified THC without modifications in the SST. We conclude that although surface density changes serve as a trigger to the reduction of the "North Atlantic" sinking, they alone cannot control the evolution of the circulation.

To mimic the zonal distribution of heat and moisture fluxes, we represented "zonal transfer terms" in our model, by making surface fluxes proportional to SST deviations from its own zonal mean. The simplicity of our approach allows separating effects of the zonal mean fluxes' contribution of heat and moisture and corresponding zonal transfers of these quantities. However, the strength of the dependence of the surface heat and moisture fluxes on the zonal structure of surface temperature may be represented inaccurately by this crude parametrization. The approach is more problematic for zonal moisture transports, since we account for zonal variations in the evaporation only and do not allow any compensation of local changes in evaporation by local changes in precipitation. This simple technique however results in an inter-basin asymmetry in changes of the moisture fluxes that is very similar to the one in the MPI model (Mikolajewicz and Voss, 2000). Experiments, in which zonal transfer terms are not allowed to change, show that the contribution of the changes in zonal distribution of heat flux to the weakening of the NA overturning is roughly equal to that of changes in its zonal-mean value. The recovery of THC is somewhat weaker when the former are not taken into account. The zonal transfer terms therefore represent a positive feedback for the "Atlantic" meridional overturning circulation during both the decline and advanced recovery stages of the THC evolution. Fixing zonal transfer terms however resulted in a strong initial recovery of the circulation, analogous to that in the experiment with fixed surface forcing described above. The terms therefore act to suppress the initial tendency of the THC to recover its strength.

The model's computational efficiency allows a large number of numerical experiments needed for a detailed analysis of the dynamical mechanisms leading to changes in the circulation. Some simplifications in the model's physics also have the advantage of a greater flexibility. However, idealized geometry and low resolution, in particular in the zonal direction in the atmosphere, have several shortcomings. For example, the model cannot explicitly represent the Labrador Sea Water formation, and therefore its NA deep water is warmer than the observed. The regional changes in the surface fluxes of heat and freshwater cannot be represented realistically, which can also affect some of the conclusions quantitatively.

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