MIT Joint Program on the Science and Policy of Global Change

Feedbacks Affecting the Response of the Thermohaline Circulation to Increasing CO_2.

A Study with a Model of Intermediate Complexity

Igor Kamenkovich, Andrei P. Sokolov and Peter H. Stone

Report No. 75
July 2001

The MIT Joint Program on the Science and Policy of Global Change is an organization for research, independent policy analysis, and public education in global environmental change. It seeks to provide leadership in understanding scientific, economic, and ecological aspects of this difficult issue, and combining them into policy assessments that serve the needs of ongoing national and international discussions. To this end, the Program brings together an interdisciplinary group from two established research centers at MIT: the Center for Global Change Science (CGCS) and the Center for Energy and Environmental Policy Research (CEEPR). These two centers bridge many key areas of the needed intellectual work, and additional essential areas are covered by other MIT departments, by collaboration with the Ecosystems Center of the Marine Biology Laboratory (MBL) at Woods Hole, and by short- and long-term visitors to the Program. The Program involves sponsorship and active participation by industry, government, and non-profit organizations.

To inform processes of policy development and implementation, climate change research needs to focus on improving the prediction of those variables that are most relevant to economic, social, and environmental effects. In turn, the greenhouse gas and atmospheric aerosol assumptions underlying climate analysis need to be related to the economic, technological, and political forces that drive emissions, and to the results of international agreements and mitigation. Further, assessments of possible societal and ecosystem impacts, and analysis of mitigation strategies, need to be based on realistic evaluation of the uncertainties of climate science.

This report is one of a series intended to communicate research results and improve public understanding of climate issues, thereby contributing to informed debate about the climate issue, the uncertainties, and the economic and social implications of policy alternatives. Titles in the Report Series to date are listed on the inside back cover.

Henry D. Jacoby and Ronald G. Prinn,
Program Co-Directors

For more information, please contact the Joint Program Office

Postal Address: Joint Program on the Science and Policy of Global Change
MIT E40-271
77 Massachusetts Avenue
Cambridge MA 02139-4307 (USA)

Location: One Amherst Street, Cambridge
Building E40, Room 271
Massachusetts Institute of Technology

Access: Phone: (617) 253-7492
Fax: (617) 253-9845
E-mail: globalchange@mit.edu
Web site: http://MIT.EDU/globalchange/

Printed on recycled paper

Feedbacks Affecting the Response of the Thermohaline Circulation to Increasing CO_2. A Study with a Model of Intermediate Complexity[#]

Igor V. Kamenkovich[†], Andrei P. Sokolov[*] and Peter H. Stone[*]

Abstract

A three-dimensional ocean model with an idealized geometry and coarse resolution coupled to a two-dimensional (zonally-averaged) statistical-dynamical atmospheric model is used to simulate the response of the thermohaline circulation to increasing CO_2 concentration in the atmosphere. The relative role of different factors in slowing down the thermohaline circulation was studied by performing simulations with ocean only and partially coupled models. The computational efficiency of the model allows an extensive and thorough study of the causes of changes in the strength of the thermohaline circulation, through a large number of extended runs. The increase in the atmosphere-to-ocean surface heat fluxes is shown to be the dominant factor in both causing the weakening of the circulation in response to an increasing external forcing as well as in controlling the subsequent recovery. Changes in the zonal distribution of heat fluxes serve as a positive feedback for both decrease and recovery of the meridional overturning, and turn out to be as important as changes in the zonal-mean values of heat fluxes. We also demonstrate that the recovery of the circulation in the ocean model cannot be sustained without feedbacks from the atmosphere. The dependency of global and regional responses on parameterization of eddy mixing, namely the Gent-McWilliams parameterization scheme versus horizontal diffusion, is also discussed.

Contents

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 gases (GHGs) and aerosols 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 (IPCC, 1996; Murphy and Mitchell, 1995; Sokolov and Stone, 1998). The causes and magnitude of weakening

[#] Joint Institute for the Study of Atmosphere and Ocean Contribution Number 841.
[†] *Corresponding author address*: Dr. Igor Kamenkovich, Department of Atmospheric Sciences, Joint Institute for the Study of the Atmosphere and the Oceans, Box 354235, University of Washington, Seattle, WA 98195-4235.
[*] Center for Global Change Science, Massachusetts Institute of Technology, Cambridge, Massachusetts

of the thermohaline circulation, which can lead to fundamental changes in the state of climate system, is also rather different between different coupled models (Dixon *et al.*, 1999). Understanding of the dynamics of the changes in atmospheric and oceanic states then becomes crucial for estimating the likelihood of any particular prediction.

The North Atlantic branch of the thermohaline circulation and associated heat transport typically weaken during the period of transient warming of the atmosphere. 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 lead to decreases 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 thermohaline circulation (THC) in a number of studies (Manabe and Stouffer, 1994; Shmittner 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 all previous studies, Latif *et al.* (2000) reports no change in the THC during the period of the transient warming of the atmosphere. The lack of agreement between different coupled models calls for a detailed analysis of the individual feedbacks in the system, as well as for assessment of the methods used in the analysis.

While many studies show that the THC starts to recover its strength after the increase in CO_2 concentration stops (see for example Manabe and Stouffer, 1994), the dynamics of the process has received surprisingly little attention in the literature. It remains unclear what mechanisms are responsible for the increase in the circulation in the still anomalously warm climate. Here again the relative contribution of heat and freshwater forcing can be very different. Manabe and Stouffer (1994) suggest that recovery occurs due to removing freshwater anomalies from the surface by circulating waters. An interesting question then arises whether circulation can recover on its own with no “help” from the atmospheric fluxes. Internal dynamics of the ocean also plays a potentially equally important role in determining the temporal behavior of the circulation.

A detailed 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 sufficient realism of the physics in the model. We give a brief description of the model in Section 2. We describe the temperature response of our model to increasing CO_2 in the atmosphere and its dependence on the parameterization of mesoscale eddies in Section 3. In Section 4, 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 summary and conclusions are presented in Section 5.

2. MODEL DESCRIPTION AND EXPERIMENTAL DESIGN

We use a coupled atmosphere/ocean model of intermediate complexity. The model combines computational efficiency with ability to reproduce most physical processes simulated by coupled atmosphere-ocean GCMs. The model components, spin-up, and coupling procedure is described in detail in Kamenkovich *et al.* (2001). Here we provide only a brief description.

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 model was spun up with a surface heat flux given by climatological heat fluxes and relaxation of surface temperature to its observed values, and with a climatological surface moisture and momentum fluxes, taken from Jiang *et al.* (1999). Observed fluxes and sea surface temperature (SST) used to spin up the model were zonally averaged over each of the two basins. The atmospheric component is a two-dimensional (zonally averaged) statistical-dynamical atmospheric model, which was developed from the GISS GCM (Hansen *et al.*, 1983). Details are given in Sokolov and Stone (1998). The model 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 reproduces most of the non-linear interactions and feedbacks simulated by atmospheric GCMs.

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

$$F_H = H_o(1-\gamma) + H_i\gamma + \left(\frac{dH_o}{dT}\right)(T-\bar{T})(1-\gamma) \tag{1}$$

$$F_w = F_o(1-\gamma) + F_i\gamma + \left(\frac{dF_o}{dT}\right)(T-\bar{T})(1-\gamma) \tag{2}$$

where H_o and F_o are the heat and fresh-water fluxes over the open ocean, dH_o/dT and dF_o/dT 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 $\bar{T}$ is the zonal mean of SST. The last terms on the right-hand sides of the equations account for the fact that the 2D atmospheric model computes heat and moisture fluxes using zonal mean SST. These terms allow for zonal variations in surface fluxes forcing the ocean, and mimic zonal transfers of heat and moisture from the warmer to colder areas. 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.

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.

To investigate the dependence of the model's response to changes in external forcing on the sub-grid scale mixing parameterization, we performed several simulations with both Gent–McWilliams (GM) and horizontal diffusion (HD) schemes. In all of these simulations, the atmospheric CO_2 concentration increases at the rate of 1% per year (compounded) for 75 years; it is kept constant after that and the model is integrated for several hundreds years more. Two simulations were carried out: one with the standard version of the ocean model (simulations GM2CO2) and one in which the Gent-McWilliams parameterization of mesoscale eddies in the ocean is replaced by horizontal diffusion (HD2CO2). The model's sensitivity, defined as the equilibrium surface warming which would be caused by the CO_2 doubling in a simulation with the atmospheric model coupled to a mixed layer ocean model, was calculated to be 2.5^{o}C.

3. GLOBAL CHANGE EXPERIMENTS: TEMPERATURE RESPONSE

Changes in the globally averaged surface air temperature (SAT) during the initial stage are very similar in the GM2CO2 and HD2CO2 simulations (**Figure 1**). In particular, by the time of CO_2 doubling it increases by 1.58°C and 1.44°C (decadal means for years 66-75) in the GM and HD cases respectively; the difference becomes more visible at later stages. In contrast, the spatial distribution of surface warming is noticeably different between the two cases even during the period of increasing CO_2. Below we discuss differences in two regions in detail.

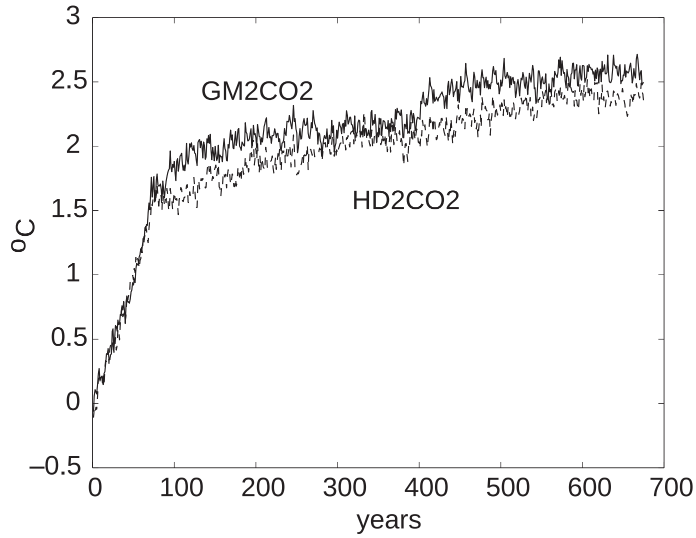

Figure 1. Increase of the global mean surface air temperature (SAT) for GM2CO2 (solid line) and HD2CO2 (dashed line). CO_2 concentration increases at the rate of 1% per year for the first 75 years, then is held constant for another 600 years.

3.1 North Atlantic

The increase in zonal mean SST is somewhat larger in the GM2CO2 simulation everywhere except around 50°N (**Figure 2**). As **Figure 3**a shows, there is a considerable cooling of the ocean surface in GM2CO2 centered at 52°N in the western North Atlantic (NA) with the SST decrease as large as –1.75 degrees. The rest of the region exhibits an overall warming with the maximum warming of 2.25 degrees immediately to the South from the described local minimum. The location of the cooling coincides with the northernmost extension of the western boundary current in the model. Since the meridional circulation substantially weakens due to the warming of the atmosphere (see Section 4), the local cooling can be explained by the decrease in the heat transport of the upper part of the western boundary current. The ability of our model to maintain a significant local minimum in

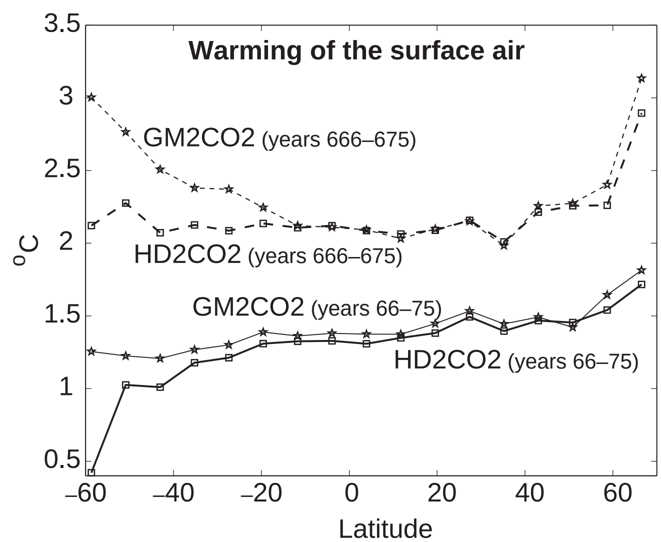

Figure 2. Increase in the zonally averaged SAT. The decadal means of the difference with the corresponding reference runs (see text) are shown for GM2CO2 (stars) and HD2CO2 (squares). Warming for years 66-75 is shown by the solid lines, warming for years 666-675 is shown by the dashed lines.

surface warming is interesting given the strong tendency of the zonal transfer term on the right-hand side of Eq. 1 to remove zonal gradients in SST.

It is interesting to note that HD2CO2 also shows a local minimum in warming centered at 52°N (Fig. 3b). This local minimum is analogous to that reported in other coupled GCM studies with HD (Manabe *et al.*, 1991; Cubash *et al.*, 1992; Russell and Rind 1999). However, the cooling is not observed, and the horizontal SST gradients across the "Gulf Stream" axis are much smaller than in GM2CO2. The most likely cause for the difference is the strong diapycnal mixing in the regions of steep isopycnal surfaces in the HD scheme that erodes horizontal temperature anomalies. In contrast, horizontal mixing in GM2CO2 is small throughout the ocean basin. Stronger warming of the western NA in HD2CO2 case then acts to weaken the convection in the region, and further reduce the overturning circulation (see Section 4.4).

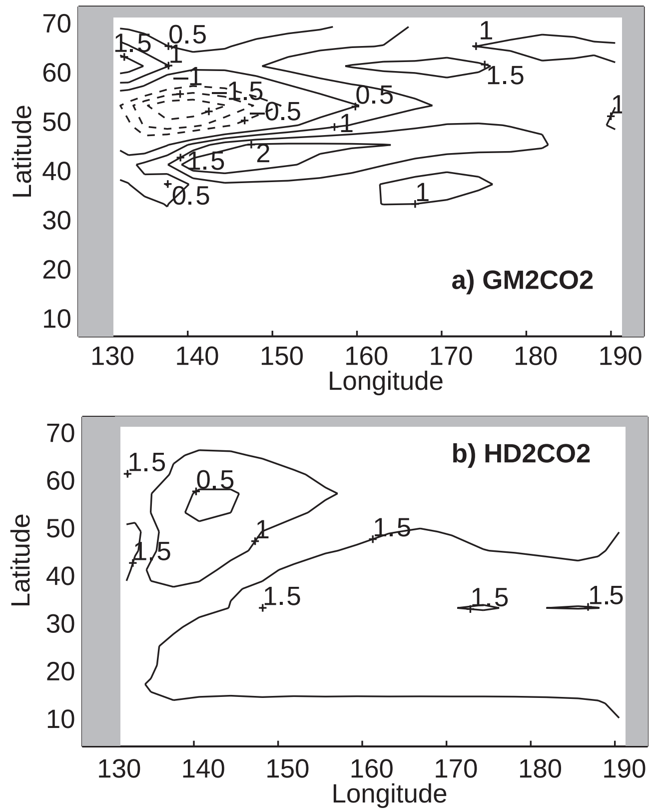

Figure 3. The annual mean of the SST warming at the year 76 with respect to corresponding reference runs: a) GM2CO2; b) HD2CO2. North Atlantic only is shown.

3.2 Southern Ocean: A Brief Summary

The main differences between simulations with the GM and HD schemes are observed in the Southern Ocean, the region with steep isopycnal surfaces. The dependence of temperature response in the Southern Hemisphere on a particular sub-grid mixing scheme is in agreement with previous findings (Hirst *et al.*, 2000; Wiebe and Weaver, 1999). Here we provide only a brief description; the reader is referred to Kamenkovich *et al.* (1999, 2000) for a more complete discussion of our results.

Large horizontal diapycnal mixing of the HD scheme maintains the Deacon cell and excessive convection in the Southern Ocean. The GM scheme reduces the effects of the Deacon cell and convection resulting in smaller vertical heat flux into the Southern Ocean and a shallower mixed layer. This results in noticeable differences between the responses to the external forcing simulated by the GM and HD models. In the simulations with HD, the deep heat penetration in the Southern Ocean delays warming of the surface and results in the Southern Hemisphere being colder than the Northern one (Fig. 3). In contrast, the inter-hemispheric asymmetry is significantly reduced in the transient response in the model with the GM scheme. The heating is confined to the upper 1000*m* of the ocean.

The surface air temperature (SAT) in HD2CO2 and GM2CO2 simulations continues to increase for more than 500 years after the CO_2 concentration is fixed (Fig. 1). Although the SAT in the Northern Hemisphere (NH) becomes very similar between GM2CO2 and HD2CO2, the Southern Hemisphere (SH) remains warmer in GM2CO2 than in HD2CO2 (Fig. 3). The values of surface temperature achieved 600 years after the growth of the CO_2 concentration stops are therefore larger in the GM case; while the deep ocean warming and, as a result, sea level rise due to thermal expansion are smaller in the GM case. The difference between the GM and HD parameterizations is even larger in simulations in which the atmospheric model has a higher sensitivity (see Kamenkovich *et al.*, 2000). We also carried out two experiments in the directly coupled mode (no flux adjustment) to estimate the dependence of the results on the coupling method. In the case of our model, the directly coupled simulations produce qualitatively similar differences between the parameterization schemes, which suggests that the differences are robust.

4. GLOBAL CHANGE EXPERIMENTS: EVOLUTION OF THE CIRCULATION

The thermohaline circulation significantly slows in response to increasing atmospheric CO_2 concentration, but does not cease completely. The behavior of the sinking in time is noticeably different between experiments with the GM and HD schemes. The NA sinking in GM2CO2 (**Figure 4**) decreases by 5Sv (20%) by the end of the period with increasing CO_2 concentration, while the corresponding change in HD2CO2 is significantly larger and amounts to more than 15Sv (33%). In GM2CO2, the recovery of the NA sinking begins only about 10 years after the end of the CO_2 increase (Fig. 4). 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 675, the NA sinking in GM2CO2 recovers to 23Sv, which is 92% of its initial value. In contrast, the recovery is delayed in HD2CO2. After 75 years of integration with constant CO_2 level during which the NA sinking remains nearly unchanged, the circulation begins to recover at the rate similar to that in the GM2CO2 case; see Fig. 4. It is noteworthy that by the year 675 the NA sinking reaches only 32Sv, which is 72% of its initial value, which is another noticeable difference with the GM2CO2 case.

We now turn our attention to the causes of the changes in the meridional circulation in our ocean model. All conclusions from the analysis of the relative importance of surface fluxes of heat and moisture presented in Sections 4.1, 4.2, and 4.3 are very similar between GM and HD cases. We will therefore restrict ourselves to showing

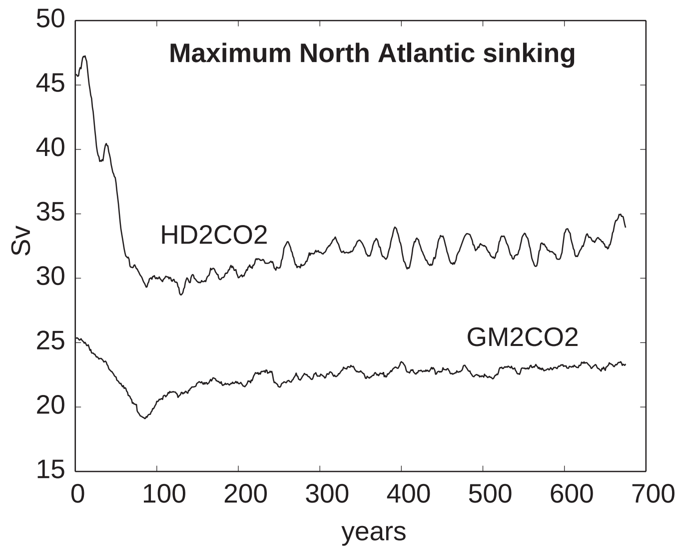

Figure 4. Evolution in time of the maximum (subsurface) overturning in the North Atlantic for the GM2CO2 and HD2CO2 runs (Sverdrups).

the results of the GM experiments only in these three Sections. As we pointed above however, the response of the Atlantic meridional overturning is different between GM and HD cases; possible causes for the difference will be discussed in Section 4.4.

4.1 Global Change Runs with Increasing CO_2

Warming of the atmosphere leads to increases in the heat and fresh water fluxes into the NA. Increases in each of these fluxes act to decrease surface density in the region, 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 $(gd_1/\rho)(d\rho/dt)$, where ρ is the density and d_1 is the thickness of the first layer in the model. The annual means of changes in these fluxes averaged over the North Atlantic area north of 44°N are plotted against time in **Figure 5**. 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 4.3.

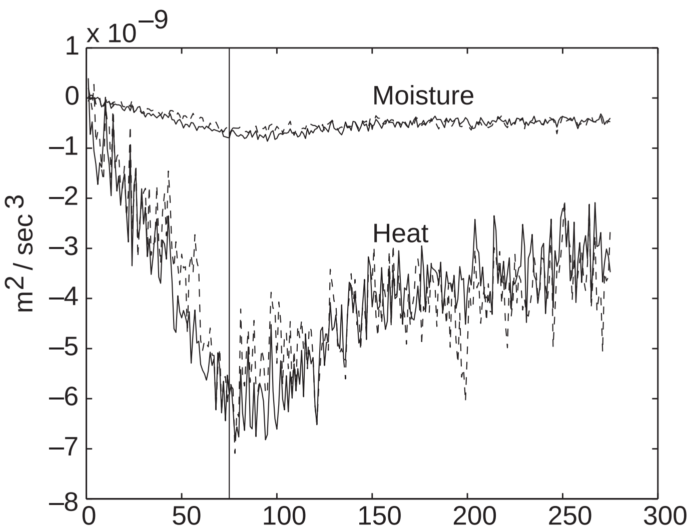

Figure 5. Changes in the surface buoyancy flux due to heat and moisture flux forcing averaged over the region north of 44°N in the Atlantic and expressed in the units of the buoyancy flux. The buoyancy flux is defined as $(gd_1/\rho)(d\rho/dt)$. Values for GM2CO2 are shown by the solid lines, values for HD2CO2 are shown by the dashed lines. Units are m^2sec^{-3}.

To verify this preliminary conclusion, 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. Our method allows us to focus on the response of the THC to changes in the surface fluxes that *exactly* repeat those in the analyzed global change experiments, whereas in a partially coupled run simulated fluxes can be different from the corresponding fluxes in the fully coupled experiments. The disadvantage of our technique is that it lacks the feedback between SST and surface fluxes; we will estimate its role by carrying out partially coupled counterparts to uncoupled experiments.

In the first experiment (FW75), the heat and momentum fluxes are taken from the corresponding reference runs with constant present-day CO_2 concentration, while the moisture flux is taken from the global change experiments. The NA sinking in this case (**Figure 6**) remains nearly constant, closely following that in the reference run. To verify this finding, we perform an additional partially coupled run (see above discussion), in which the CO_2 concentration is kept constant at its normal level, whereas freshwater flux is diagnosed from the global change run. As in our uncoupled experiment, the circulation stays very close to its initial value (not shown).

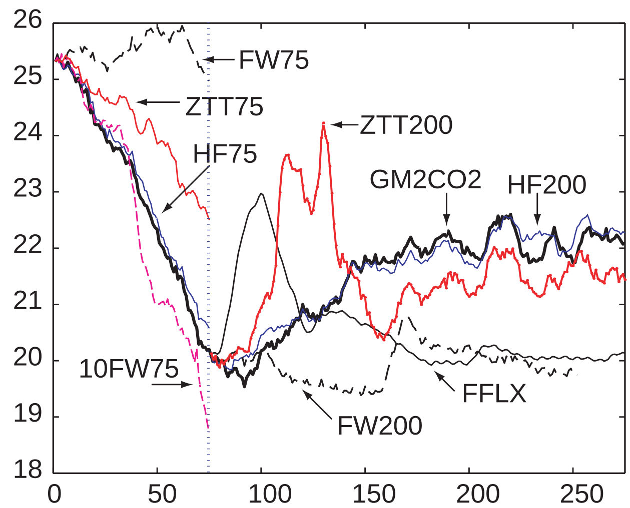

Figure 6. Evolution in time of the maximum (subsurface) overturning in the North Atlantic for the years 1-275 for experiments described in the **Table 1**. Corresponding experiment names are given next to the curves. The thick line shows control case GM2CO2.

In the second setting (HF75), moisture and momentum fluxes are taken from the reference run, while the heat flux is taken from the corresponding global change experiments. The decrease of the NA sinking in this case matches that in the global change runs, proving that the anomalous heating of the oceanic surface is, indeed, the main reason for the slowdown of the thermohaline circulation in our model (Fig. 6). Our results therefore agree with the conclusions of a study by Mikolajewicz and Voss (2000), who used the Max-Plank Institute (MPI) model and also reported that changes in surface heat fluxes are the main reason for the decrease in the THC. Their model was run until the CO_2 concentration in the atmosphere quadrupled at the year 120. The rate of the decrease in the circulation in MPI study is similar to that in our additional 150 year run (see Kamenkovich *et al*., 2001), by the end of which CO_2 also quadruples; there is however a significant delay in the start of the decrease in the THC strength in the MPI model.

Table 1. Experiments reported in the Figure 6 with the corresponding methods of surface forcing. Surface fluxes are either taken from a different fully coupled run (the name of the run is given in the table) or computed in a coupled mode ("Atmos. model"). GM mixing scheme.

Experiment	*Heat fluxes*	*Moisture fluxes*	*Wind stress*	*Zonal transfer terms*
GM2CO2	Atmos. model	Atmos. model	Atmos. model	Allowed to change
HF75	GM2CO2, yrs 1-75	Reference run	Reference run	Allowed to change
FW75	Reference run	GM2CO2, yrs 1-75	Reference run	Allowed to change
10FW75	Atmos. model CO_2 fixed	GM2CO2, yrs 1-75, *changes amplified*	Atmos. model	Allowed to change
ZTT75	Atmos. model	Atmos. model	Atmos. model	Fixed, yrs 71-75
HF200	GM2CO2, yrs 76-275	Fixed, yrs 71-75	Fixed, yrs 71-75	Allowed to change
FW200	Fixed, yrs 71-75	GM2CO2, yrs 76-275	Fixed, yrs 71-75	Allowed to change
FFLX	Fixed, yrs 71-75	Fixed, yrs 71-75	Fixed, yrs 71-75	Allowed to change
ZTT200	Atmos. model	Atmos. model	Atmos. model	Fixed, yrs 76-80

The results do not change significantly when we repeat this experiment with only heat fluxes north of the Equator taken from a global change run and the rest of forcing diagnosed from the reference 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 investigate a little further the role of the moisture fluxes and the insensitivity of the THC to their changes. The relatively small magnitude of the changes in the buoyancy flux due to changes in the moisture fluxes shown in Fig. 5 may not be the only reason for their small 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. It is therefore interesting to investigate what would happen if changes in the moisture fluxes in density units were as large as those in the heat fluxes, and heat fluxes in the NA did not increase. Will the decrease in the THC be the same as in the global change GM2CO2 run?

To address this question, we run an uncoupled run with heat fluxes diagnosed from a coupled run with fixed CO_2 and the moisture fluxes taken from GM2CO2 and multiplied by 10; the effects of moisture fluxes on density in this run are therefore similar in magnitude to that of the heat fluxes in GM2CO2. The circulation (not shown) started to decline at a rate similar to that in GM2CO during the first 15 years. However, the lack of SST feedbacks on the surface forcing made the results on later stages unreliable (circulation abruptly recovers with unrealistic SST structure), which dissuaded us from continuing the experiment. The result however suggests that the small amplitude of the changes in the moisture fluxes is the main reason for their secondary role in causing weakening of the meridional circulation.

We then carry a partially coupled experiment with fixed CO_2 concentrations, in which we force the ocean model with changes in moisture fluxes taken from GM2CO2 run and multiplied by a factor of 10. The heat fluxes are calculated interactively. The changes in the moisture flux in this experiment (10FW75 in Fig. 6) then have similar magnitudes (in density units) to the changes in heat flux in the run GM2CO2. The circulation weakens noticeably faster than in the GM2CO2 case; however the results are difficult to interpret for the following reason. Our analysis shows, that although CO_2 concentration was fixed and there was no increase in the zonal-mean heat flux into ocean, the heat flux into NA was increasing at a rate similar to that in the fully coupled GM2CO2 global change run. The main cause of this increase is the inter-basin zonal heat flux into the North Atlantic from the North Pacific (NP), associated with a gradual change in SST difference between NA and NP due to decreasing meridional oceanic circulation (see Eq. 1). We discuss the role of the inter-basin heat transport in slowing done the thermohaline circulation in detail in Section 4.3. The observed large drop in the circulation is then the result of both the increasing moisture and heat fluxes in the NA. The role of the moisture fluxes can therefore be easily overestimated in a partially coupled run, if heat flux changes are not carefully analyzed. The problem can be especially serious, if the contribution of heat fluxes to the weakening of the THC is not negligible, as in 10FW75 or in Dixon *et al*. (1999).

4.2 Runs with Constant Doubled CO_2

We now attempt to understand what mechanisms cause the initial recovery of the circulation in our coupled model. During the period of fixed doubled CO_2 concentration, the surface air temperature continues to increase. However, the surface buoyancy fluxes due to anomalous heat and moisture fluxes each decrease in magnitude in the NA, as Fig. 5 demonstrates. This decrease in the strength of the anomalous surface forcing is consistent with the recovery of the circulation. As during the years 1-75, the changes in heat fluxes clearly dominate over changes in the moisture fluxes.

To check if heat fluxes are a primary cause for the circulation changes, we conduct uncoupled experiments similar to the uncoupled experiments of Section 4.1. It should be noted, that equilibrium is not achieved in these experiments, and our only aim is to reproduce the transient behavior during the initial 200 years after CO_2 stops increasing. In the first experiment (HF200 in Fig. 6), we force the ocean model by the heat fluxes from the fully coupled experiment after year 75. For the moisture flux, we take values from the 5-year mean computed for years 71-75; 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.

Our next step is to run an uncoupled experiment with the ocean model forced by the moisture fluxes taken from a years 76-275 and heat flux and wind stress taken from the time mean for years 71-75 (run FW200 in Fig. 6). The THC fails to recover. It is clear, that the evolution of the moisture fluxes cannot explain the recovery of the circulation. The THC also fails to recover in another run, in which all surface fluxes are fixed to a 5-year mean for years 71-75 of GM2CO2, preserving a seasonal cycle (FFLX in Fig. 6). The overturning abruptly increases up to the year 100 but then quickly drops back. The circulation seems to have a tendency to remove positive buoyancy anomalies at the surface and regain its strength. It is noteworthy, that both HF200 and FW200 do not show this initial peak in the circulation strength. Therefore either the zonal transfer terms or the zonal means of the heat and moisture fluxes act to suppress the tendency of THC to recover. We will come back to this in the next section. This recovery however has to be supported by a feedback from the surface heat fluxes and is clearly not self-sustaining.

4.3 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 (Eqs. 1 and 2). The latter effect mimics zonal transfers of heat and moisture that cannot be explicitly computed by our zonally averaged atmospheric model. Our simple parameterization guarantees that there is a zonal flux of heat and moisture against zonal SST gradient.

Figure 7 shows the difference between the heat fluxes averaged between 44^{o}N and 58^{o}N in the Pacific and Atlantic basins (buoyancy flux units). This difference depends on the inter-basin SST contrast (Eq. 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. 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°N-60°N latitude band in the Atlantic is 3 times as large as those changes over the same latitude band in the Pacific; the NA moisture flux changes are almost 3.5 times as large as in the NP. 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 Atlantic. Therefore, the second term on the right hand side of Eq. 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 zonal fluxes of heat (Fig. 7) 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.

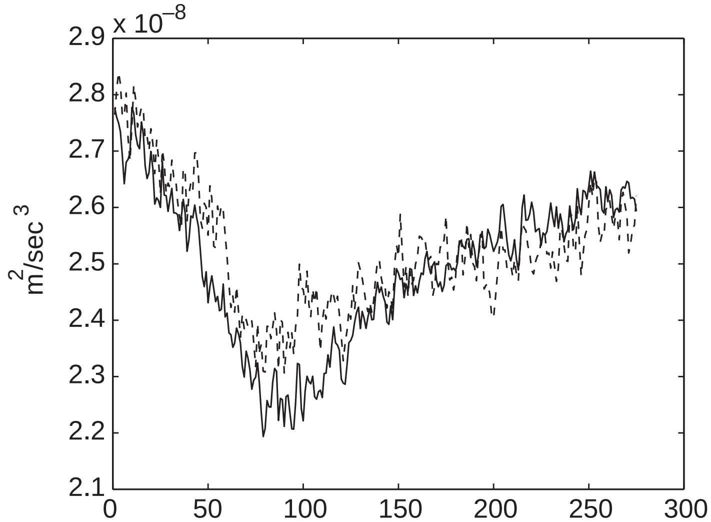

Figure 7. Difference between Pacific and Atlantic surface buoyancy fluxes due to heat flux forcing. Values are averaged between 44°N and 58°N and the units are m^2sec^{-3}. The solid line shows GM2CO2, and the dashed line shows HD2CO2.

To evaluate the role of zonal transfer terms, we ran two additional coupled GM experiments, in which zonal transfer terms were not allowed to change. The first run is the repeat of the global change experiment GM2CO2 for years 1-75 with zonal transfer terms taken from a 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. 6). 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 NP. Thus, the zonal heat fluxes supply a positive feedback to the THC during the initial decline. In a second experiment (ZTT200 in Fig. 6), we fix zonal transfer terms to their values for the years 71-75 (day-to-day variations are preserved) and run a coupled model for 200 years starting from the end of GM2CO2 run. The circulation initially exhibits a sharp increase analogous to that in the run with fixed fluxes (FFLX in Fig. 6). The circulation then drops back and continues its 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. Our model does not simulate latitudinal anomalies in precipitation. In fact, we observe an anomalous moisture flux from the Pacific into the Atlantic at the low latitudes, in the direction opposite to that in Latif *et al*. (2000).

4.4 Differences between GM and HD

As was described in the beginning of this section, the response of the THC to external forcing differs in the experiments with the two mixing schemes. The decrease in NA overturning during years 1-75 is much larger in the HD case both in absolute and relative terms and subsequent recovery is weaker than in the GM case. Nevertheless, the analysis of Sections 4.1-4.3 repeated for the HD cases did not provide much information for understanding the difference. The changes in buoyancy fluxes over the Northern Hemisphere are very similar between the GM and HD cases and the NA heat fluxes remain the main cause of the weakening and recovery of the circulation in the HD case.

Wiebe and Weaver (1999) also looked at the different effects of HD and GM on the evolution of the THC in global warming experiments, very similar to our HD2CO2 and GM2CO2, with a model of intermediate complexity. In our simulations, the initial decrease in the THC is considerably larger in both absolute and relative terms than in the Wiebe and Weaver experiments, especially in the HD case. The stronger decrease in our experiments compared to theirs is because the buoyancy flux in the North Atlantic in our experiments is an order of magnitude greater; it is also dominated by the heat flux, while in their experiment this flux is initially dominated by the moisture flux.

In addition, the THC recovery after CO_2 is fixed in our experiments is much weaker than in the Wiebe and Weaver experiments. In the recovery phase of their experiments, the changes in the buoyancy flux are dominated by the heat flux, which becomes larger and positive. They attribute this to the retreat of sea-ice in the area of the deep water formation, which enhances the cooling of the surface waters. In our model, there is no sea-ice in the vicinity of the NA deep water formation. The surface response of the ocean is, however, noticeably different between the two cases. As discussed in Section 3.1, reduced oceanic heat transport in the GM case results in the cooling of the western NA, which acts to sustain the sinking and slow the weakening of the meridional overturning. In contrast, the western NA is warmer in the HD case, since the cooling effect of the weakening meridional circulation is much reduced by strong diapycnal mixing, which acts to further weaken the sinking in the region.

The other noticeable difference is in the surface response of the Southern Hemisphere (SH). The meridional mass transport in the upper layers has been shown to depend on the magnitude of the northward pressure gradient (Rahmstorf, 1996; Gnanadesikan, 1999), which is in turn controlled by density structure in the upper layers, and by the SH surface buoyancy in particular (Wang *et al.*, 1999). In the current climate, the South Atlantic (between 44^{o}S and Equator) is warmer than the North Atlantic (NH). Because the heating of the SH is delayed in the HD case due to the deep heat penetration (see Section 3, Fig. 3), the temperature difference between South Atlantic and NH decreases during years 1-75 in HD2CO2, as **Figure 8** shows. This decrease acts to reduce the magnitude of the North-South surface density contrast and therefore weaken the northward pressure gradient, which further slows the circulation. In contrast, the inter-hemispheric SST contrast increases with time in GM2CO2 (note that SAT, not SST is reported in Fig. 3), which acts to sustain the THC. During years 76-275, the South-North SST contrast remains below its initial value in HD2CO2 consistent with the incomplete recovery of the meridional circulation in that case.

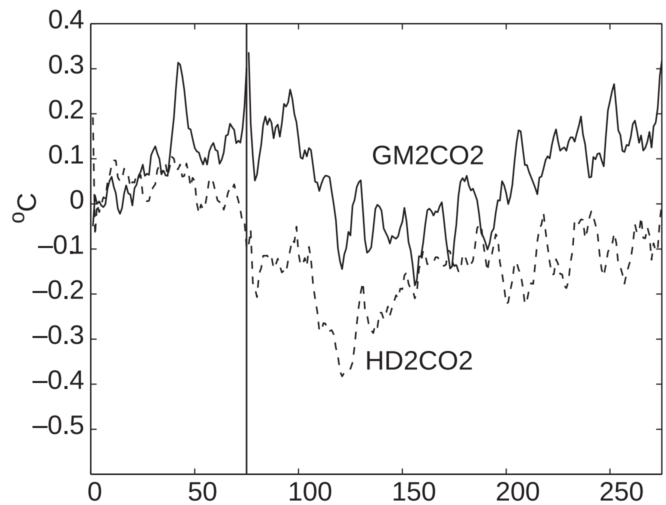

Figure 8. Change (with respect to the years 1-5) in the difference between the South Atlantic (between 44S and the Equator) and the Northern Hemisphere SSTs as a function of time (smoothed by a 5-year sliding mean). GM2CO2 (solid) and HD2CO2 (dashed).

5. SUMMARY AND DISCUSSION

We use a coupled atmosphere/ocean model of intermediate complexity for the analysis of the behavior of the thermohaline circulation in response to increasing CO_2 concentration in the atmosphere. The model captures the main features of the response of the temperature and circulation documented by more sophisticated coupled models. In addition, 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 advantages. In particular, explicit parameterization of the zonal transfers of heat and moisture at the surface allows us to easily change them in a numerical run separately from other terms.

As in a number of other studies, the NA meridional overturning in our model initially weakens during the period of increasing CO_2 concentration in the atmosphere and then partially recovers after the 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 behavior of the circulation in time is very similar between our standard fully coupled case and our uncoupled runs, in which only heat fluxes are taken from a coupled run with increasing CO_2 concentration and fluxes of moisture and momentum are diagnosed from a reference run. This demonstrates 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), who used the Max-Plank Institute (MPI) model and demonstrated a major role of the heat fluxes in slowing the circulation. In contrast, other studies (Manabe and Stouffer, 1994; see also a summary in Dixon *et al.*, 1999) point to the formation of the surface fresh-water anomalies in NA associated with an increase in the atmospheric moisture transport as the dominant factor in weakening the sinking in the region. Wiebe and Weaver (1999) suggest that the warming of the surface acts to intensify the NA circulation by increasing the density and pressure contrast between the high latitudes and tropics. This effect, however, depends on the latitudinal distribution of the surface warming. With uniform warming, the sea level rise would indeed be larger in the tropics due to the nonlinearity of the equation of state. The anomalous surface warming however increases northward (Fig. 3), which acts to decrease the meridional pressure gradient in the upper pycnocline counteracting the effect reported by Wiebe and Weaver (1999).

This study also presents a first attempt, to our knowledge, to investigate the causes of the circulation recovery, which starts shortly after the increase in the atmospheric CO_2 concentration stops. Results from our uncoupled experiments demonstrate that the recovery is caused by decreasing surface heating of the NA. This process increases surface density in the region, which enhances the formation of the deep water. Our analysis also suggests, that although the circulation tends to remove surface buoyancy anomalies and initially starts recovering, the recovery cannot be sustained, if surface buoyancy fluxes do not change. For example, the meridional circulation in the ocean model forced with unchanging surface fluxes fails to recover, despite initial fast intensification followed by the equally rapid weakening. It is also unlikely that the changes in the atmospheric fluxes to the ocean are not sensitive to the evolving oceanic circulation. The recovery of the circulation is therefore essentially a coupled mechanism, in which feedbacks between oceanic THC and atmospheric fluxes into the ocean play the primary role.

The estimates of the relative importance of anomalous fluxes of heat and moisture depend not only on the specifics of a particular coupled model but on the method of analysis as well. Our uncoupled method of analysis is different from a partially coupled approach, in which some fluxes are computed by an active atmospheric model, while others are taken from a separate run (Mikolajewicz and Voss, 2000; Dixon *et al.*, 1999). 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 moisture and heat fluxes for decrease in the THC in the coupled run are comparable, as in the study by Dixon *et al.* (1999), taking just one of them to force the ocean model produces results that can be easily misinterpreted. The latter statement is illustrated by our experiment, in which we artificially increase the rate of change of moisture flux anomalies in an uncoupled run and in a partially coupled experiment. In the partially coupled case, we observe an anomalous zonal heat flux into the NA, which increases with time due to the gradual cooling of the NA surface without increase in the zonal-mean value of heat flux. Since it is impossible to control changes in the heat fluxes in the partially coupled setting, results are hard to interpret.

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 can be potentially overestimated by this crude parameterization. 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 response of the circulation differs between the two mixing schemes. NA sinking in our model with the GM scheme decreases at a slower rate during the stage of increase in CO_2 than in the experiment with the HD. THC recovers at a similar rate in both cases, which results in the HD final values being well below the initial levels. Nevertheless, surface fluxes in the NA evolve in a very similar manner between GM and HD cases and cannot be a reason for the difference. A complete explanation of the reasons for the difference would require a full understanding of the mechanisms controlling the meridional overturning in the Atlantic. We can nevertheless suggest several reasons to explain the difference between the GM and HD cases. The western NA appears to be cooling in GM2CO2 due to the weakening of the meridional circulation; the cooling acts to sustain the sinking in the region and slow the weakening of the overturning. In contrast, the SST gradients are reduced by the strong diapycnal mixing in HD2CO2 resulting in the more uniform warming of the NA, which acts to reduce the sinking. In addition, the meridional circulation has been shown to depend on inter-hemispheric density contrast in the Atlantic (Rahmstorf, 1996; Wang *et al.*, 1999). The delayed Southern Hemisphere surface warming in the HD2CO2 case then will act to weaken the meridional circulation, enhancing the decrease in the circulation during years 1-75 and slowing recovery during the rest of the experiment. We note that a similar difference in the behavior of the circulation between the GM and HD schemes is observed in the CO_2 quadrupling experiment by Wiebe and Weaver (1999). Their CO_2 doubling GM and HD experiments, in contrast, exhibit very similar decrease and recovery of the circulation.

This study outlines advantages of using models of intermediate complexity for detailed analysis of the mechanisms controlling the thermohaline circulation. This analysis typically requires a large number of numerical experiments, which are not possible with a detailed model. An intermediate model that combines computational efficiency with realistic representation of physics makes this analysis both plausible and meaningful. After an extensive study with the intermediate model singles out mechanisms that control evolution of the THC in response to external forcing, the quantitative conclusions of the study can be refined with the use of more detailed models.

Acknowledgements

Igor Kamenkovich was supported by the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) under NOAA Cooperative Agreement #NA67RJ0155, Contribution # 841. This research was supported in part by MIT's Joint Program on the Science and Policy of Global Change, and in part by the U.S. Department of Energy's Office of Biological and Environmental Research under Grant No. DE-FG02-93ER61677 to MIT.

6. REFERENCES

Dixon, K.W., T.L. Delworth, M.J. Spelman and R.J. Stouffer, 1999: The influence of transient surface fluxes on North Atlantic overturning in a coupled GCM climate exchange experiment, *Geophys. Res. Let.*, **26**: 2749-2752.

Hansen, J., G. Russell, D. Rind, P. Stone, A. Lacis, S. Lebedeff, R. Ruedy and L. Travis, 1983: Efficient three dimensional global models for climate studies: Models I and II. *Mon. Weather Rev.*, **111**: 609-662.

Hirst, A.C., H.B. Gordon and S.P. O'Farrell, 1996: Global warming in a coupled model including oceanic eddy-induced advection. *Geoph. Res. Let.*, **23**: 3361-3364.

IPCC, *Climate Change 1995. The Science of Climate Change*. Cambridge University Press, Cambridge, UK, 1996.

Kamenkovich, I., A. Sokolov and P. Stone, 1999: An intermediate coupled model for studying the sensitivity of the thermohaline circulation to climate change, Preprints, 12th Conference on Atmospheric and Oceanic Fluid Dynamics, pp. 193-197.

Kamenkovich, I., A. Sokolov, and P. Stone, 2000: A coupled atmosphere-ocean model of intermediate complexity for climate change study. MIT Joint Program on the Science and Policy of Global Change, Report No. 60 (http://mit.edu/globalchange/MITJPSPGC_Rpt60.pdf).

Kamenkovich, I., A. Sokolov, and P. Stone, 2001: An efficient climate model with a 3D ocean and statistical-dynamical atmosphere. Submitted to *Clim. Dyn*.

Knutii, R., and T. Stocker, 1999: Influence of the thermohaline circulation on projected sea level rise. Submitted to *J. Climate*.

Latif, M., E. Roeckner, U. Micolajewicz and R. Voss, 2000: Tropical stabilization of the Thermohaline Circulation in a greenhouse warming simulation. *J. Climate*, **13**, 1809-1813.

Manabe, S., and R.J. Stouffer, 1994: Multiple-century response of a coupled ocean-atmosphere model to an increase of atmospheric carbon dioxide. *J. Climate*, **7**: 5-23.

Manabe, S., R.J. Stouffer, M.J. Spelman and K. Bryan, 1991: Transient responses of a coupled ocean-atmosphere model to gradual changes of atmospheric CO_2. Part I: Annual mean response. *J. Climate*, **4**: 785-818.

Mikolaewicz, U., and R. Voss, 2000: The role the individual air-sea flux components in CO_2-induced changes of the ocean's circulation and climate. *Climate Dyn.*, **16**: 627-642.

Mitchell, J.F.B, T.C. Johns and C.A. Senior, 1998: Transient response to increasing greenhouse gases using model with and without flux adjustment. *Hadley Centre Technical Note #2.*

Murphy, J.M., and J.F.B Mitchell, 1995: Transient response of the Hadley Centre coupled ocean-atmosphere model to increasing carbon dioxide. Part. II: Spatial and temporal structure of response. *J. Climate*, **8**: 57-80.

Nakamura, M., P.H. Stone and J. Marozke, 1994: Destabilization of the Thermohaline Circulation by atmospheric eddy transports. *J. Climate*, **7**: 1870-1881.

Rahmstorf, S., 1996: One the freshwater forcing and transport of the Atlantic thermohaline circulation. *Climate Dynamics*, **12**: 691-637.

Russell, G.L., and D. Rind, 1999: Response to CO_2 transient increase in the GISS coupled model: Regional cooling in a warming climate. *J. Climate*, **12**: 531-539.

Schmittner, A., and T.F. Stocker, 1999: The stability of the thermohaline circulation in global warming experiments, *J. Climate*, **12**: 1117-1133.

Sokolov, A., and P. Stone, 1998: A flexible climate model for use in integrated assessments. *Climate Dyn.*, **14**: 291-303.

Stocker, T.F., and A. Schmittner, 1997: Influence of CO_2 emission rates on the stability of the thermohaline circulation, *Nature*, **388**: 862-865.

Stouffer R.J., and S. Manabe, 1999: Response of a coupled ocean-atmosphere model to increasing atmospheric carbon dioxide: sensitivity to the rate of increase, *J. Climate*, **12**: 2224-2237.

Wiebe, E.C., and A.J. Weaver, 1999: On the sensitivity of global warming experiments to the parameterization of sub-grid scale ocean mixing. *Climate Dyn*., **15**: 875-893.

Wood, R.A., A.B. Keen, J.F.B. Mitchell, and J.M. Gregory, 1999: Changing spatial structure of the thermohaline circulation in response to atmospheric CO_2 forcing in a climate model, *Nature*, **399**: 572-575.

REPORT SERIES of the **MIT** *Joint Program on the Science and Policy of Global Change*

1. **Uncertainty in Climate Change Policy Analysis** *H.D. Jacoby and R.G. Prinn*, December 1994
2. **Description and Validation of the MIT Version of the GISS 2D Model** *A. Sokolov and P. Stone*, June 1995
3. **Responses of Primary Production and Carbon Storage to Changes in Climate and Atmospheric CO_2 Concentration** *X. Xiao* et al., October 1995
4. **Application of the Probabilistic Collocation Method for an Uncertainty Analysis** *M. Webster* et al., Jan. 1996
5. **World Energy Consumption and CO_2 Emissions: 1950-2050** *R. Schmalensee* et al., April 1996
6. **The MIT Emission Prediction and Policy Analysis (EPPA) Model** *Z. Yang* et al., May 1996
7. **Integrated Global System Model for Climate Policy Analysis** *R. Prinn* et al., June 1996 (*superseded by No. 36*)
8. **Relative Roles of Changes in CO_2 and Climate to Equilibrium Responses of Net Primary Production and Carbon Storage** *X. Xiao* et al., June 1996
9. **CO_2 Emissions Limits: *Economic Adjustments and the Distribution of Burdens*** *H.D. Jacoby* et al., July 1997
10. **Modeling the Emissions of N_2O and CH_4 from the Terrestrial Biosphere to the Atmosphere** *Y. Liu*, Aug. 1996
11. **Global Warming Projections: *Sensitivity to Deep Ocean Mixing*,** *A. Sokolov and P. Stone*, September 1996
12. **Net Primary Production of Ecosystems in China and its Equilibrium Responses to Climate Changes** *X. Xiao* et al., Nov. 1996
13. **Greenhouse Policy Architectures and Institutions** *R. Schmalensee*, November 1996
14. **What Does Stabilizing Greenhouse Gas Concentrations Mean?** *H. Jacoby* et al., November 1996
15. **Economic Assessment of CO_2 Capture and Disposal** *R. Eckaus* et al., December 1996
16. **What Drives Deforestation in the Brazilian Amazon?** *A.S.P. Pfaff*, December 1996
17. **A Flexible Climate Model For Use In Integrated Assessments** *A. Sokolov and P. Stone*, March 1997
18. **Transient Climate Change & Potential Croplands of the World in the 21st Century** *X. Xiao* et al., May 1997
19. **Joint Implementation: *Lessons from Title IV's Voluntary Compliance Programs*** *E. Atkeson*, June 1997
20. **Parameterization of Urban Sub-grid Scale Processes in Global Atmospheric Chemistry Models** *J. Calbo* et al., July 1997
21. **Needed: A Realistic Strategy for Global Warming** *H.D. Jacoby, R.G. Prinn and R. Schmalensee*, Aug. 1997
22. **Same Science, Differing Policies; *The Saga of Global Climate Change*** *E.B. Skolnikoff*, August 1997
23. **Uncertainty in the Oceanic Heat and Carbon Uptake and their Impact on Climate Projections** *A. Sokolov* et al., September 1997
24. **A Global Interactive Chemistry and Climate Model**, *C. Wang, R.G. Prinn and A. Sokolov*, September 1997
25. **Interactions Among Emissions, Atmospheric Chemistry and Climate Change: *Implications for Future Trends*** *C. Wang and R.G. Prinn*, September 1997
26. **Necessary Conditions for Stabilization Agreements** *Z. Yang and H.D. Jacoby*, October 1997
27. **Annex I Differentiation Proposals: *Implications for Welfare, Equity and Policy*** *D. Reiner & H. Jacoby*, Oct. 1997
28. **Transient Climate Change and Net Ecosystem Production of the Terrestrial Biosphere** *X. Xiao* et al., Nov. 1997
29. **Analysis of CO_2 Emissions from Fossil Fuel in Korea: 1961–1994** *K.-H. Choi*, November 1997
30. **Uncertainty in Future Carbon Emissions: *A Preliminary Exploration*** *M. Webster*, November 1997
31. **Beyond Emissions Paths: *Rethinking the Climate Impacts of Emissions Protocols in an Uncertain World*** *M. Webster and D. Reiner*, November 1997
32. **Kyoto's Unfinished Business** *H.D. Jacoby, R.G. Prinn and R. Schmalensee*, June 1998
33. **Economic Development and the Structure of the Demand for Commercial Energy** *R. Judson* et al., April 1998
34. **Combined Effects of Anthropogenic Emissions and Resultant Climatic Changes on Atmospheric OH** C. *Wang and R.G. Prinn*, April 1998
35. **Impact of Emissions, Chemistry, and Climate on Atmospheric Carbon Monoxide** *C. Wang & R. Prinn*, Apr. 1998
36. **Integrated Global System Model for Climate Policy Assessment: *Feedbacks and Sensitivity Studies*** *R.G. Prinn* et al., June 1998
37. **Quantifying the Uncertainty in Climate Predictions** *M. Webster and A. Sokolov*, July 1998
38. **Sequential Climate Decisions Under Uncertainty: *An Integrated Framework*** *L.J. Valverde* et al., Sep. 1998
39. **Uncertainty in Atmospheric CO_2 Concentrations from an Uncertainty Analysis of a Ocean Carbon Cycle Model** *G. Holian*, October 1998

Contact the Joint Program Office to request a copy. The Report Series is distributed at no charge.

40. **Analysis of Post-Kyoto CO_2 Emissions Trading Using Marginal Abatement Curves** *A.D. Ellerman and A. Decaux*, October 1998
41. **The Effects on Developing Countries of the Kyoto Protocol and CO_2 Emissions Trading** *A.D. Ellerman* et al., November 1998
42. **Obstacles to Global CO_2 Trading: *A Familiar Problem*** *A.D. Ellerman*, November 1998
43. **The Uses and Misuses of Technology Development as a Component of Climate Policy** *H. Jacoby*, Nov. 1998
44. **Primary Aluminum Production: *Climate Policy, Emissions and Costs*** *J. Harnisch* et al., December 1998
45. **Multi-Gas Assessment of the Kyoto Protocol** *J. Reilly* et al., January 1999
46. **From Science to Policy: *The Science-Related Politics of Climate Change Policy in the U.S.*** *E. Skolnikoff*, Jan. 1999
47. **Constraining Uncertainties in Climate Models Using Climate Change Detection Techniques** *C. Forest* et al., April 1999
48. **Adjusting to Policy Expectations in Climate Change Modeling** *S. Shackley* et al., May 1999
49. **Toward a Useful Architecture for Climate Change Negotiations** *H.D. Jacoby* et al., May 1999
50. **A Study of the Effects of Natural Fertility, Weather and Productive Inputs in Chinese Agriculture** *R.S. Eckaus and K. Tso*, July 1999
51. **Japanese Nuclear Power and the Kyoto Agreement** *M. Babiker, J. Reilly and A.D. Ellerman*, August 1999
52. **Interactive Chemistry and Climate Models in Global Change Studies** *C. Wang and R.G. Prinn*, Sep. 1999
53. **Developing Country Effects of Kyoto-Type Emissions Restrictions** *M. Babiker and H.D. Jacoby*, Oct. 1999
54. **Model Estimates of the Mass Balance of the Greenland and Antarctic Ice Sheets** *V. Bugnion*, Oct. 1999
55. **Changes in Sea-Level Associated with Modifications of the Ice Sheets over the 21st Century** *V. Bugnion*, October 1999
56. **The Kyoto Protocol and Developing Countries** *M. Babiker, J. Reilly and H.D. Jacoby*, October 1999
57. **A Game of Climate Chicken: *Can EPA regulate GHGs before the Senate ratifies the Kyoto Protocol?*** *V. Bugnion and D. Reiner*, Nov. 1999
58. **Multiple Gas Control Under the Kyoto Agreement** *J. Reilly, M. Mayer and J. Harnisch*, March 2000
59. **Supplementarity: *An Invitation for Monopsony?*** *A.D. Ellerman and I. Sue Wing*, April 2000
60. **A Coupled Atmosphere-Ocean Model of Intermediate Complexity for Climate Change Study** *I. Kamenkovich* et al., May 2000
61. **Effects of Differentiating Climate Policy by Sector: *A U.S. Example*** *M. Babiker* et al., May 2000
62. **Constraining Climate Model Properties using Optimal Fingerprint Detection Methods** *C. Forest* et al., May 2000
63. **Linking Local Air Pollution to Global Chemistry and Climate** *M. Mayer* et al., June 2000
64. **The Effects of Changing Consumption Patterns on the Costs of Emission Restrictions** *S. Lahiri* et al., August 2000
65. **Rethinking the Kyoto Emissions Targets** *M.J. Babiker and R.S. Eckaus*, August 2000
66. **Fair Trade and Harmonization of Climate Change Policies in Europe** *L. Viguier*, September 2000
67. **The Curious Role of "Learning" in Climate Policy: *Should We Wait for More Data?*** *M. Webster*, Oct. 2000
68. **How to Think About Human Influence on Climate** *C.E. Forest, P.H. Stone and H.D. Jacoby*, October 2000
69. **Tradable Permits for Greenhouse Gas Emissions: *A primer with particular reference to Europe*** *A.D. Ellerman*, November 2000
70. **Carbon Emissions and The Kyoto Commitment in the European Union** *L.L. Viguier* et al., Feb. 2001
71. **The MIT Emissions Prediction and Policy Analysis (EPPA) Model: *Revisions, Sensitivities, and Comparisons of Results*** *M.H. Babiker* et al., February 2001
72. **Cap and Trade Policies in the Presence of Monopoly and Distortionary Taxation** *D. Fullerton and G. Metcalf*, March 2001
73. **Uncertainty Analysis of Global Climate Change Projections** *M. Webster* et al., March 2001
74. **The Welfare Costs of Hybrid Carbon Policies in the European Union** *M. Babiker* et al., June 2001
75. **Feedbacks Affecting the Response of the Thermohaline Circulation to Increasing CO_2** *I. Kamenkovich* et al., July 2001

Contact the Joint Program Office to request a copy. The Report Series is distributed at no charge.