Response of the Atlantic thermohaline circulation to increased atmospheric CO$_2$ in a coupled model

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Abstract

Changes in the thermohaline circulation (THC) due to increased CO$_2$ are important in future climate regimes. Using a coupled climate system model – Parallel Climate Model (PCM), regional responses of the THC in the North Atlantic to increased CO$_2$ and the underlying physical processes are studied here. The Atlantic THC shows a 20-year cycle in the control run, qualitatively consistent with observations. Compared with the control run, the simulated maximum of the Atlantic THC weakens by about 5 Sv or 14% in an ensemble of transient experiments with 1% CO$_2$ increase per year at the time of CO$_2$ doubling. The weakening of the THC is accompanied by reduced poleward heat transport in mid-latitude North Atlantic. Analyses show that oceanic deep convective activity strengthens significantly in the Greenland-Iceland-Norwegian (GIN) Seas, but weakens in the Labrador Sea and the south of Denmark Strait Region (SDSR). The strengthening of deep convective activity in the GIN Seas is mainly caused by an increased salty North Atlantic inflow, reduced sea ice volume fluxes from the Arctic into this region, all of which lead to a saltier (denser) upper ocean. The weakening of deep convective activity in the SDSR is induced by a reduced sea ice flux into this region and a reduced heat loss to the atmosphere which leads to a warmer (lighter) upper ocean. On the other hand, the weakening in the Labrador Sea is mainly attributed to increased precipitation that freshens the surface ocean. These regional changes produce the overall weakening of the THC in the Labrador Sea and SDSR, and more vigorous ocean overturning in the GIN Seas. The northward heat transport south of 60°N is reduced with increased CO$_2$, but increased north of 60°N due to the increased North Atlantic water across this latitude.
1. Introduction

The thermohaline circulation (THC) is primarily a density driven global-scale oceanic circulation. It plays an important role in global meridional heat and freshwater transport. Changes in the THC alter the global ocean heat transport, thus affecting the global climate. Human-induced global warming due to increased atmospheric CO$_2$ and other greenhouse gases can change the global evaporation minus precipitation pattern and terrestrial runoff. For example, there will likely be more fresh water input into the polar and sub-polar seas due to increased precipitation at high latitude (e.g., Dai et al. 2001a, b). The resulting surface buoyancy flux into these seas will be altered by the fresh water input anomaly and surface warming. Since the sinking branch of the THC is highly localized in the northern North Atlantic marginal seas and in the southern ocean, such variations in surface buoyancy will lead to more stably-stratified upper oceans and suppressed deep convection in these seas, resulting in a weakened THC and reduced poleward heat transport.

The impact of a weakened or possibly even collapsed THC due to human-induced global warming on global climate has raised considerable concern (e.g., Broecker 1987, 1997; Rahmstorf 1999). The IPCC Third Assessment Report states wide varying response of the THC to the projected forcing scenario – IS92a over the 21$^{st}$ century in different models (Cubasch et al. 2001). Most of the coupled climate models predict a weakened THC in response to increased atmospheric CO$_2$ levels, such as the GFDL model (Manabe and Stoufer 1994; Dixon et al. 1999), the Hadley Center model (HadCM3)(Wood et al. 1999), the NASA GISS model (Russell and Rind 1999), the Canadian model (Boer et al. 2000), a Hamburg model (ECHAM3/LSG)(Voss and Mikolajewicz 2001), and the Parallel Climate Model (PCM)(Washington et al. 2000; Dai et al. 2001b). Several intermediate-complexity models (e.g., Stocker and Schmittner 1997; Rahmstorf and Ganopolski 1999; Schmittner and Stocker 1999; Wiebe and Weaver 1999) also show a weakened THC. A few other coupled models, however, show little response of the THC to increased greenhouse gas forcing.
These include the ECHAM4/OPYC3 (Latif et al. 2000), the NCAR CSM1.3 (Gent 2001), and the GISS AGCM coupled with the HYCOM ocean model (Sun and Bleck 2001).

For those models with a weakened THC, the relative importance of the surface warming and freshening, in general, varies among different models. By specially designing a set of experiments, Dixon et al. (1999) concluded that the enhanced poleward moisture transport due to atmospheric process contributes most to the reduction of the Atlantic THC in the GFDL model, while the effect of heat flux changes on the THC is less important. In contrast, Mikolajewicz and Voss (2000), using the ECHAM3/LSG model, reported that the direct and indirect effects of the oceanic surface warming account for most of the THC weakening, while the effect of the enhanced poleward moisture transport is secondary. Thorpe et al. (2001) illustrated that the high-latitude temperature increases account for 60% of the THC weakening and the salinity decreases account for 40% in a model with ECHAM3 coupled to a Cox-type ocean model. Dai et al. (2003) show that the weakening of THC in the PCM results from larger surface ocean density decreases in the northern North Atlantic than in the Southern Ocean that are primarily caused by surface warming. They suggest that the THC’s response to increased greenhouse gas forcing depends on the initial THC state in the model. On the other hand, models with a stable THC show that the effects of surface warming are compensated by a salinity increase, resulting in little change in surface ocean density in northern North Atlantic region.

Many of the previous studies focused on basin-scale effects of surface warming and freshening on the THC. Because the sinking branch of the THC is highly localized, especially in the North Atlantic, it is important to analyze the oceanic processes on regional scales. Here we extend Dai et al. (2003)’s analysis by examining changes along isopycnal surfaces, focusing on the regional processes underlying the THC’s response to increased atmospheric CO₂ forcing in the PCM. Results of this paper should improve our understanding of the effects of the intensity changes in regional deep convective activity on the THC.
2. Model, experiments and analysis methods

a. Model

The PCM is a fully coupled climate system model (Washington et al. 2000) consisting of four component models – atmosphere, ocean, land and sea-ice. The atmospheric component is the National Center for Atmospheric Research’s (NCAR) Community Climate Model version 3 (CCM3) at T42 horizontal resolution (approximately 2.8°) with 18 hybrid levels vertically (Kiehl et al. 1998). The ocean component is the Los Alamos National Laboratory’s Parallel Ocean Program (POP)(Smith et al. 1995) with an average grid size of 2/3° (1/2° in latitude over the equatorial region) and 32 vertical levels. The Land surface model is NCAR’s Land Surface Model (LSM)(Bonan 1998). The sea ice model is a version of Zhang and Hibler (1997) optimized for the parallel computer environment required by the PCM. The control climate in the PCM is stable and comparable to observations (Washington et al. 2000). This coupled model has been used in a number of climate change studies (e.g., Meehl et al. 2001; Dai et al. 2001b, c; Arblaster et al. 2002; and Dai et al. 2003).

b. Experiments

The simulations examined here include a 300-year control run with constant 1990 values of the greenhouse gases (e.g., CO₂ concentration is fixed at 355 ppm)(Arblaster et al. 2002) and an ensemble of four 80-year transient climate runs with 1% CO₂ increase per year. The control run started from year 49 after the model was fully coupled. The four transient climate runs started from year 21, 91, 151, and 201 of the control run, respectively. Hereafter, the control run will be referred to as CON, and the ensemble of transient runs as ET.

The global mean surface air temperature at the time of CO₂ doubling around year 70 in ET increases by about 1.3°C and the globally averaged first level ocean temperature is increased
by about 1°C. The global averaged precipitation increases by 1.9% at the time of CO₂ doubling in ET comparing with that in CON. However, the averaged precipitation north of 60°N increases by about 9% in ET.

c. **Analysis method**

Since the motion of sea water is along isopycnal surfaces underneath the ocean mixed layer, this motivated us to analyze the PCM ocean output in the density domain. First, the potential density of the sea water was calculated using the model output with reference to the sea surface. Then, the water mass is interpolated to a set of isopycnal layers according to the water density using a method developed by Bleck (appendix D, 2002) that conserves mass, salt, heat and momentum. The density values for these layers are shown in the second column of Table 1.

Figure 1 shows the mean Atlantic meridional streamfunction (MSF) derived in both depth (upper panel) and density (lower panel) domain averaged over the last century of CON. The two panels represent a similar pattern of the meridional overturning circulation (MOC) with maximum values greater than 30 Sv (Sv ≡ 10^6 m³/s). The MOC derived in the density domain, however, shows more detailed structure of the upper ocean without losing details of the deep ocean. The lower panel also implies that as the upper ocean water flows northward, it becomes denser due to the atmospheric cooling effect. As the surface water becomes dense enough, it sinks into the deep ocean and flows southward.

It is worth pointing out that the locations of the maximum MOC centers seems different in the two panels (Fig. 1). The center is located between 30 and 40°N in the depth domain, but around 50°N in the density domain. A careful review found that there is a maximum MOC center between 30 and 40°N in the density domain although this center is weaker than that in the depth domain. This indicates that the vertical motion there has fallen into one isopycnal layer due to a set of limited number of isopycnal layers used here. By increasing the number of isopycnal layers, this
maximum MOC center will be able to be resolved in the density domain. On the other hand, the maximum MOC center around 50°N in the density domain indicates a vigorous diapycnal motion there. Observations indicate that most of the North Atlantic Deep Water (NADW) is generated in this region (e.g., McCartney and Talley 1984).

3. MOC variability in future climate

In this section, variations of the annual mean maximum MOC and changes of the MSF in the North Atlantic ocean in the density domain will be reported first, followed by a three dimensional analysis of the regional water mass transport. Finally, the linkage of the MOC changes with the variations of water properties is discussed.

a. View of the MOC in one and two dimensions

Figure 2 shows the time evolution of the maximum values of the annual mean North Atlantic MOC in CON. The linear trend for this 300-year run is \(-1\) Sv per century. However, most of the changes occur during the first 30 years. The linear trend from year 30 to 300 is less than a third of that for the whole time series. Large decadal variations are evident in this figure. A spectral analysis reveals a 20-year cycle significant at 0.95 level. Similar decadal MOC oscillations are suggested by observations (e.g., Dickson et al. 1996) and modeling studies (e.g., Delworth et al. 1993, Cheng 2000). The cause of the decadal MOC variation in the PCM control run is under study, but out of the scope of this paper.

The North Atlantic MOC shows a steady trend of weakening as CO\(_2\) increases in ET. The difference of the maximum MOC in ET minus CON (of corresponding periods) shown in Fig. 3 exhibits large interannual variations, especially between year 50 and 60. The interannual variations are mainly related to the MOC variations in CON. The mean maximum MOC for the last 20 years
(year 61 – 80) is weaker by 4.7 Sv, approximately 14% of the control run. As a consequence, the northward heat transport is reduced by 4% at 30°N and 9% at 45°N. However, the heat transport north of 60°N increases, with a peak increase of 15% at 65°N. Similar heat transport changes are reported by Meehl et al. (2000). As shown below, the increase in northward heat transport north of 60°N is due to increased North Atlantic warm water across this latitude flowing into the GIN Seas.

Figure 4 shows the difference of the MOC between ET and CON averaged over the last 20 years of the 1% CO$_2$ runs. The MOC is weaker by up to 5 Sv in the North Atlantic (south of 60°N) in ET. However, in the region north of 60°N, the MOC is strengthened in ET, suggesting that the deep convective activity is intensified there.

b. Regional variability – a three dimensional view

To further analyze regional variability of deep convective activity in the northern North Atlantic marginal seas, the water mass is grouped into 5 density classes after Schmitz (1996). The definition of the water classes is given in the third column of Table 1. The regions of interest are divided into four sub-domains, namely the Labrador Sea (45°N to 65°N, west of 45°W), south of Denmark Strait region (SDSR, 45°N to 65°N, and east of 45°W), the GIN Seas and Baffin Bay (see Fig. 5 for details). The formation of the NADW mainly occurs in the first three regions.

Because water lighter than 27.82 kg/m$^3$ flows generally northward and water with a density of 27.82 kg/m$^3$ and heavier flows southward in the North Atlantic region (bottom panel of Fig. 1), we further simplify our three-dimensional analysis by grouping the first 4 water classes into a new class to represent the upper branch of the Atlantic THC, called as the upper water (UW). Class 5 water represents the lower branch of the THC, named as deep water (DW, Table 1). Hereafter, these two water classes are referred as the UW and the DW.

Figure 6 shows the ensemble-averaged isopycnal and diapycnal fluxes for CON (left two
panels) and ET (right two panels). The ensemble average is over the last 20 years of each of the transient climate runs (year 61–80) for ET and the corresponding 20-year periods of the control run for CON. Henceforth, all model data discussed are averaged over the same period as in Fig.6. The arrows indicate the direction of the isopycnal flow and the number next to the arrow is the amount of the volume flux in Sv. The numbers located at the center of each sub-domain represent the strength of the diapycnal fluxes. Negative values indicate a downward water mass conversion – water from a lighter class is converted to a denser class. The number located on top of Greenland represents the Bering Strait inflow from the Pacific into the Arctic.

The net northward flowing UW crossing 45°N in CON is about 27.2 Sv (37.5 Sv northward into SDSR and 10.3 Sv southward from the Labrador Sea), which is reduced to 23.8 Sv in ET (32.7 Sv northward and 8.9 Sv southward, see the upper two panels in Fig.6). On the other hand, the net North Atlantic water flowing into the GIN Seas increases from only 1.4 Sv in CON (2.0 Sv northward through Iceland-Norwegian channel and 0.6 Sv southward through Denmark Strait) to 9.6 Sv in ET. This increase of North Atlantic water into the GIN Seas occurs at both sides of Iceland. The flow between Iceland and Norway is doubled in the ET runs. At the Denmark Strait, the flow reverses direction to northward in ET in contrast with CON where it flows southward. About 5.3 Sv of North Atlantic water flows into the GIN Seas via Denmark Strait in the ET runs.

The diapycnal fluxes, representing the conversion of UW to DW, exhibit a significant weakening with increasing CO₂ in the Labrador Sea (from –7.6 Sv down to –4.8 Sv) and in the SDSR (from –20.8 Sv down to –12.2 Sv). This represents a 40% reduction of the UW to DW conversion between 45°N and 65°N from a total of 28.4 Sv in CON to 17 Sv in ET. In the GIN Seas, however, deep convective activity is dramatically strengthened. The diapycnal flux increases from –2 Sv in CON to –9.4 Sv in ET. The overall reduction of the diapycnal fluxes in these three sub-domains is moderate at about 4 Sv, from –30.4 in CON to –26.4 Sv in ET.
Because of the intensification of deep convective activity in the GIN Seas, the Denmark Strait Overflow Water (DSOW) in the lower branch of the THC is significantly increased from 8.6 Sv in CON to 13.9 Sv in ET (Fig. 6). Summing the flow values between Greenland and Norway, the net DW outflow from the GIN Seas is 2.6 Sv in CON, and 10.5 Sv in ET. The exchanges of DW between the SDSR and the Labrador Sea are reduced from 8.7 Sv in CON to 2.6 Sv in ET, resulting in a weakened western boundary current (16.6 Sv in CON versus 8 Sv in ET). The majority of the DW crossing 45°N (17.6 Sv out from a total of 25.6 Sv) flows southward via the east of the mid-Atlantic ridge, consistent with Wood et al. (1999).

It should be noticed that the volume fluxes in each of the sub-domains are not exactly balanced. This imbalance is related to the changes of the UW/DW volume in each sub-domain, also induced partly by the displaced North Pole ocean grid when we tried to calculate the isopycnal fluxes along a given latitude or longitude. In general, the imbalance is small.

c. Linkage to changes in water properties

The changes in diapycnal fluxes can be directly linked to variations in water mass properties in these sub-domains (Fig. 7). In the GIN Seas, UW is 1.59°C warmer and 0.18 ppt saltier in ET than in CON. The resulting density of UW is about 0.045 kg/m$^3$ heavier. This heavier UW intensifies the winter deep convective processes in the GIN Seas by weakening the oceanic vertical stratification. On the other hand, the Labrador Sea UW in ET is colder (−0.13°C) and fresher (−0.09 ppt), and the density is about 0.059 kg/m$^3$ lighter than in CON. In the SDSR, UW in ET is of 0.024 kg/m$^3$ lighter with a warmer (0.25°C) and saltier (0.02 ppt) upper ocean. Therefore, the lighter UW strengthens the oceanic vertical stratification and leads to weakened deep convective activity in the latter two regions.

The relative importance of the temperature and salinity changes in these three regions is different. In the SDSR, temperature variation dominates the UW density changes, and in the GIN
Seas and Labrador Sea, the salinity changes is more important on the UW density variations in the PCM. It should also be mentioned that the water temperature and salinity changes tend to be of the same sign, such as warmer and saltier or cooler and fresher. The net effect of these will minimize the changes in water density.

4. Physical processes

The water property changes can be caused by changes of isopycnal and diapycnal fluxes (oceanic transport processes), surface heating and cooling, net surface fresh water flux, and sea ice melting and freezing. Each of these processes will be addressed in this section.

a. Oceanic transport processes

The regional heat and salt balance can be changed due to changes in isopycnal and diapycnal flows. Figures 8 and 9 show heat and salt fluxes. The numbers in rectangles represent the net heat or salt gain (positive) or loss (negative) due to oceanic processes (including both isopycnal and diapycnal transports) in the region and the numbers in ovals represent the heat or salt fluxes due to diapycnal transport. The arrows indicate the direction of the along isopycnal flows and numbers next to the arrows represent the amount of the heat or salt fluxes.

The heat and salt transports are calculated as follows:

\[ H_t = \int_{y_1}^{y_2} \int_{x_1}^{x_2} \rho c_p T v dx dy, \]

\[ H_s = \int_{y_1}^{y_2} \int_{x_1}^{x_2} \rho S v dx dy, \]

where \( H_t \) represents heat flux, \( H_s \) represents salt flux, \( \rho = 1000 \text{ kgm}^{-3} \) is the water reference density, \( c_p = 3996 \text{ Jkg}^{-1}\text{deg}^{-1} \) specific heat of sea water, \( T \) water temperature in degree Kelvin, \( v \) velocity, \( S \) salinity. For the isopycnal heat or salt fluxes, the integration is from \( x_1 \) to \( x_2 \) along a given latitude or longitude, from layer \( y_1 \) to \( y_2 \), where \( dy \) represents the layer thickness. For the diapycnal heat
or salt fluxes, the integration is from $x_1$ to $x_2$ along a given latitude, and from $y_1$ to $y_2$ along a given longitude, where $v$ represents the diapycnal velocity.

1) **Heat transport**

The net northward heat transport across 45°N carried by UW in ET, which includes 39 PW northward and 10 PW southward for a net northward transport of 29 PW (bottom panel of Fig. 8), is reduced by 4 PW compared to 33 PW in CON (a 12% reduction, Fig. 8). The reduction is consistent with the weakening of the MOC. But the magnitude of the reduction is less than that of the MOC: the net northward volume transport of UW across 45°N is reduced by about 12.5%, from 27.2 Sv in CON to 23.8 Sv in ET (Fig. 6), which implies a warmer UW in ET.

In the SDSR, the net heat loss of the UW due to oceanic transport processes is 50% larger in ET (–3 PW) than in CON (–2 PW), indicating that these transport processes work against the UW warming in this region. The weakening of the diapycnal volume flux in the SDSR in ET as shown in section 3.b induces a reduced downward heat transport from –25 PW in CON to –15 PW (Fig. 8), which would have contributed to the warming of the UW in the SDSR. However, it is prevented by the reduced heat flux from subtropical Atlantic into this region and the increased heat transport from the SDSR into the GIN Seas.

In the GIN Seas, the net heat gain of the UW due to oceanic transport processes is higher in ET (3 PW) than in CON (2 PW). This higher heat flux convergence is mainly contributed by the increased North Atlantic water inflow through Denmark Strait and Iceland-Norway channels with a net heat flux of 11 PW in comparison with that of 1 PW in CON. Most of the increased heat flux into this basin is transported into denser layers due to the intensified diapycnal flux there (–11 PW in ET compared to –2 PW in CON). The rest is used to heat up the UW.

The net heat gain of the UW in the Labrador Sea is 1 PW in ET higher than that in CON (0 PW, Fig. 8). Thus the oceanic transport processes favor a warmer UW, opposite to the UW
temperature change shown in section 3.c.

2) Salt transport

Since the northward UW volume flux crossing 45°N is decreased in ET compared to CON as noted above, the northward salt transport is also reduced from a net transport in CON of $972 \times 10^6$ kgs$^{-1}$ ($1329 \times 10^6$ kgs$^{-1}$ northward minus $357 \times 10^6$ kgs$^{-1}$ southward) to $854 \times 10^6$ kgs$^{-1}$ ($1161 \times 10^6$ kgs$^{-1}$ northward minus $307 \times 10^6$ kgs$^{-1}$ southward) in ET (Fig. 9). The net reduction of about $118 \times 10^6$ kgs$^{-1}$ in ET is approximately a 12% decrease from CON. This percentage is slightly lower than the reduction in water volume transport (12.5%), thus the UW south of 45°N should be slightly saltier in ET than in CON.

In the SDSR, salt in UW is lost due to oceanic transport in both runs, however, the salt loss is higher in ET ($-90 \times 10^6$ kgs$^{-1}$) than in CON ($-73 \times 10^6$ kgs$^{-1}$), indicating that the UW should be fresher which is opposite to the salinity changes in this basin shown in Fig. 7. Therefore the oceanic transport processes work against the salinity increase in ET. Figure 9 also shows that the diapycnal salt flux is reduced from $-740 \times 10^6$ kgs$^{-1}$ in CON to $-437 \times 10^6$ kgs$^{-1}$ in ET due to the weakened diapycnal flow. This favors an increase of UW salinity in this region. However, increased salt exports from this basin to GIN Seas ($195 + 150 \times 10^6$ kgs$^{-1}$ in ET vs $70 - 15 \times 10^6$ kgs$^{-1}$ in CON), along with a reduced salt import from the subtropical region ($1161 \times 10^6$ kgs$^{-1}$ in ET vs $1329 \times 10^6$ kgs$^{-1}$ in CON), result in a higher salt loss in this region.

In the Labrador Sea, the salt gain in UW is increased from $13 \times 10^6$ kgs$^{-1}$ in CON to $26 \times 10^6$ kgs$^{-1}$ in ET (Fig. 9), opposite from the UW salinity changes in this basin. The increase in salt gain is related to the weakening of the diapycnal salt flux from $-269 \times 10^6$ kgs$^{-1}$ in CON to $-171 \times 10^6$ kgs$^{-1}$ in ET and the reduction of the southward salt export from $357 \times 10^6$ kgs$^{-1}$ in CON to $307 \times 10^6$ kgs$^{-1}$ in ET. Therefore the salinity changes in this sub-domain shown in Fig. 7 must be due to changes in fresh water flux from the atmosphere noted in the next section.
The UW salt flux entering the GIN Seas is dramatically increased from a net only $55 \times 10^6$ kgs$^{-1}$ in CON (70$\times$10$^6$ kgs$^{-1}$ northward and 15$\times$10$^6$ kgs$^{-1}$ southward) to a net $345 \times 10^6$ kgs$^{-1}$ northward in ET. On the other hand, the net salt flux from the Arctic ocean is reduced from 76$\times$10$^6$ kgs$^{-1}$ in CON (82$\times$10$^6$ kgs$^{-1}$ into GIN Sea via Fram Strait minus 6$\times$10$^6$ kgs$^{-1}$ exported to Barents Sea) to 66$\times$10$^6$ kgs$^{-1}$ in ET (83$\times$10$^6$ kgs$^{-1}$ into the GIN Seas via Fram Strait minus 17$\times$10$^6$ kgs$^{-1}$ exported to Barents Sea). The diapycnal salt flux in ET, however, is significantly increased from $-70 \times 10^6$ kgs$^{-1}$ in CON to $-330 \times 10^6$ kgs$^{-1}$. Overall, the net salt gain due to oceanic transport processes in this basin is actually increased from $62 \times 10^6$ kgs$^{-1}$ in CON to $82 \times 10^6$ kgs$^{-1}$ in ET, contributing to the salinity increase.

In summary, the oceanic transport processes work against the salinity and temperature changes of the UW in the SDSR and in the Labrador Sea, and contribute positively to those changes in the GIN Seas.

b. Air-sea interaction

The ocean exchanges heat and fresh water with the atmosphere through air-sea fluxes, which induce temperature and salinity variations and thus affect oceanic buoyancy and stratification in the upper ocean.

Figures 10 and 11 show the net surface heat and fresh water fluxes in the four sub-domains. Positive values indicate an oceanic heat or fresh water loss. In general, the oceanic heat loss is lower in ET than in CON, and fresh water gain is higher in ET than in CON (except the SDSR) in all four sub-domains. The reduced heat loss is induced by a larger warming in surface air than in surface ocean. The increased fresh water gain, mainly due to precipitation increases, contributes to a fresher surface ocean.

The combined effects of surface heat and fresh water fluxes should lead to warmer and fresher UW in ET in the Labrador Sea. However, the model shows a colder and fresher UW (see Fig. 7
and the last line in Table 2). As shown in Table 2, the class 1 to 3 waters are warmer in ET than in CON and the class 1 and 2 waters are fresher in ET in the Labrador Sea. These changes result in a decrease in surface ocean density, which strengthens the upper ocean vertical stratification, and suppresses the vertical heat exchanges. As a result, the sub-surface water (i.e., the class 4 water) is cooler in ET (4.546°C) than in CON (4.630°C). Overall, the warming of the class 1 to 3 water does not overcome the cooling of the class 4 water in ET since the latter makes up about 60% of the UW. Therefore, the cooler UW in the Labrador Sea is a combined effect of air-sea interaction and upper ocean physics.

In the SDSR and GIN Seas, the reduced heat loss contributes to the warming of UW in ET. The fresh water flux does not contribute to the salinity anomaly in the SDSR, and works against the salinity increase in the GIN Seas in ET.

c. Sea ice

In general, the sea ice extent in the Arctic (not shown) and the ice export through Fram Strait are reduced in ET compared to CON. The sea ice extent, ice volume, thickness, and concentration in the GIN Seas are reduced by 30%, 45%, 25% and 7% in ET, respectively (Fig. 12). The ice exchange between the Arctic and GIN Seas through Fram Strait is decreased from 0.076 Sv in CON to 0.056 Sv in ET. The ice flux from the Barents Sea into the GIN Seas is reduced by 50% in ET from 0.022 Sv to 0.011 Sv. The ice flux exiting at the Denmark Strait is also reduced from 0.072 Sv in CON to 0.047 Sv in ET. The net ice flux convergence in the GIN Seas, a sum of ice flux at the Denmark Strait, Barents Sea, and Fram Strait, changes from 0.026 Sv to 0.020 Sv in ET. The reduced ice flux convergence helps the increase of the UW salinity. The reduction in ice covered area and concentration leads to an increased open ocean area in the GIN Seas. The combined effect of these two processes is an increase in wintertime heat loss, thus contributing to intensifying the deep convective activity in the GIN Seas.
In the SDSR, the ice flux convergence, a sum of ice flux at Denmark Strait and out of SDSR into the Labrador Sea, is reduced from 0.053 Sv in CON to 0.032 Sv in ET. The impact of this reduction in ice flux convergence in this region is two fold. First, it means a less fresh water input into this region, resulting in a positive contribution to the salinity changes. Secondly, this reduction also implies that the heat used to melt ice is reduced, leading to an increase in UW temperature. Therefore, the sea ice processes in this region contribute positively to the temperature and salinity changes in ET.

In the Labrador Sea, the ice extent and volume are decreased by 15% and 20% in ET, respectively. The ice flux convergence, a sum of ice fluxes across northern, eastern and southern boundaries of the Labrador Sea (see Fig. 12) is also decreased from 0.025 Sv in CON to 0.022 Sv in ET which would have contributed to a saltier UW. However, the net impact of the sea ice processes does not overcome impact of the air-sea interaction processes, especially the increased precipitation, which stabilize the upper ocean stratification. In fact, the decrease in sea ice extent and volume indicates a weakened wintertime sea ice production, and less salt ejection into the upper ocean due to ice formation. The latter also contributes to a stable upper ocean stratification in ET.

d. Mechanism of the GIN Seas’ MOC strengthening

It is now more clear that the weakening of deep convective activity in the Labrador Sea and the SDSR is mainly due to the stabilization of the upper ocean induced by increased CO$_2$. However, the intensification of the deep convective activity in the GIN Seas seems at odds with the expected oceanic response. In fact, the warming induced by increased CO$_2$ causes an increase in upper ocean temperature almost everywhere, resulting in a warmer and saltier northward flowing Atlantic water. When this water reaches the Labrador Sea and the SDSR, it is not dense enough to sink into the abyssal sea due to warming and/or freshening effects. Therefore this warmer and saltier North Atlantic water overshoots into the GIN Seas (an increase in water volume transport
into the GIN Seas as shown in Fig. 6 and discussed in section 3.b) since there is nothing to prevent it.

A further analysis indicates that the warming and salinity increase of the UW in the GIN Seas occur in concert with the increase in the warmer and saltier North Atlantic water inflow through the channels on both sides of Iceland. Deep convective activity in the GIN Seas is also intensified simultaneously. The correlation between the annual UW salinity (temperature) increase in the GIN Seas and the North Atlantic inflow on both side of the Iceland is 0.95 (0.95) for the 80-year integration in ET. And the correlation between the increase in annual diapycnal flux in the GIN Seas and the North Atlantic inflow is about 0.94. At the latitude of the GIN Seas, the salinity perturbation becomes more important to the variation in water density. In winter, salinity induced increases of density dominates the decreases of density induced by temperature, which destroys the weak stratification in the GIN Seas, and enhances the deep convective activity there.

On the other hand, the decrease of ice volume flux into the GIN Seas in ET also induces an increase in UW salinity, which acts to weaken the upper ocean stratification, thus contributes to strengthen the deep convective activity in the GIN Seas.

Caveats: in this paper, the effects of wind variations induced by increasing atmospheric CO$_2$ concentration are not explicitly discussed. However, since the motion of UW is mostly wind driven, those effects are implicitly included in the variations of the UW transport studied in section 3.b. Also, the effects of surface winds on air-sea heat and water fluxes are implicitly included in the surface fluxes discussed in section 4.b.
5. Summary and conclusions

The North Atlantic THC in a 300-year control run and a four-member ensemble of 1% CO$_2$ transient runs using PCM has been analyzed for decadal and longterm changes. The Atlantic THC show a 20-year cycle in the control run, qualitatively agreeing with observations (Dickson et al. 1996) and other coupled climate system models (e.g., Delworth et al. 1993; Cheng 2000). Compared with the control run, the North Atlantic THC weakens by about 5 Sv (14%) at the time of CO$_2$ doubling. Spatially, the changes of the diapycnal fluxes are not uniformly distributed isopycnally. Our analyses reveal that the diapycnal fluxes weakens by roughly 40% in the Labrador Sea and the South of Denmark Strait Region, and strengthens by more than 3 times in the GIN Seas.

Analyses of the various processes in the model indicate that the weakening (strengthening) of the diapycnal fluxes is related to the strengthening (weakening) of the upper ocean stratification induced by increased CO$_2$. Processes controlling the THC responses identified here include the oceanic transport processes, air-sea interaction, sea ice processes, and upper ocean physics. The relative importance of these processes is different in the various sub-domains.

In the Labrador Sea, the increased net fresh water input (precipitation minus evaporation) in ET plays a crucial role in stabilizing the surface ocean, thereby suppressing the deep convective activity there. The warming of the surface water is also very important in the Labrador Sea. In the SDSR, warming in the upper ocean is the key factor for reduced oceanic deep convection. This warming is induced by a reduced heat loss to the atmosphere and a reduced sea ice volume flux from the GIN Seas into this region. The latter reduces the heat used to melt ice, indirectly contributes to the warming in SDSR. Directly, the reduced ice flux into this region induces an increase of the upper ocean salinity. However, its effect on density is smaller than the density decrease induced by the overall temperature increase. In the GIN Seas, salinity changes induced by sea ice and increased oceanic transport of warm and salty water into the region are the main factors for increases in the
upper ocean density and deep convection. The effect of UW warming on density is smaller than that of the salinity increase in the GIN Seas.

The net effects of these processes are to weaken the THC in the Labrador Sea and the SDSR, but to strengthen it in the GIN Seas. The northward heat transport south of 60°N is reduced with increased CO₂, but increased north of 60°N due to the increased North Atlantic water across this latitude.

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Reference


to an increase of atmospheric carbon dioxide. *J. Climate*, 7, 5-23.


Table 1. Density and water mass classes. The ocean model output has been converted into 18 isopycnal layers. After Schmitz (1996), the ocean water is grouped into 5 density classes, named the upper water (UW, Class 1), upper intermediate water (UIW, Class 2), lower intermediate water (LIW, Class 3), upper deep water (UDW, Class 4), and lower deep water (LDW, Class 5). In this paper, the first 4 water classes are regrouped into one new class to represent the upper branch of the Atlantic THC based on Fig. 1. This new class is called the upper water (UW). Class 5 water is renamed as the deep water (DW).

Table 2. Temperature, salinity, and water volume in the Labrador Sea
Fig. 1. Atlantic meridional stream function (Sv, 1 Sv $\equiv 10^6$ m$^3$/s) derived in depth domain (upper) and density domain (lower) averaged over the last century of the control run. Contour interval is 2 Sv. Note that the density coordinate is non-linear.

Fig. 2. Time series of the maximum MOC at North Atlantic in the control run. The thick solid line is the 13-year lowpass filtered maximum MOC.

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Fig. 4. Ensemble mean MOC in ET minus that in CON, averaged over the last 20-year of each of the 1% CO$_2$ runs for ET and the corresponding 20-year period of the control run for CON. Contour interval is 1 Sv.

Fig. 5. Northern North Atlantic map. The four sub-domains of interest are defined as the Labrador Sea (45$^\circ$N to 65$^\circ$N, west of 45$^\circ$W), South of Denmark Strait region (SDSR, 45$^\circ$N to 65$^\circ$N, east of 45$^\circ$W), the GIN Seas (roughly 65$^\circ$N to 80$^\circ$N, east of 45$^\circ$W to west of 20$^\circ$E), and the Baffin Bay (roughly 65$^\circ$N to 80$^\circ$N, west of 45$^\circ$W to east 80$^\circ$W). The major deep ventilations occur in the GIN Seas, the Labrador Sea and SDSR.

Fig. 6. Isopycnal and diapycnal fluxes (Sv) for CON (left panels) and ET (right panels). The upper panels are for the UW representing the upper branch of the THC. The lower panels for DW which represents the lower branch of the THC. Arrows indicate the direction of the isopycnal flows, the numbers next to the arrows give the amount of the isopycnal flows. The numbers at the center of each box represent the strength of the diapycnal fluxes. Negative means lighter water converted into denser water.

Fig. 7. Regional mean temperature, salinity, and density difference of the UW between ET and CON. “Lab” represents the Labrador Sea, “GIN” the GIN Seas, “SDSR” the south of Denmark.
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**Fig. 8.** Total isopycnal and diapycnal heat fluxes in PW ($10^{15}$ W) of the UW. Upper panel represents the control run, and the lower panel for the ensemble 1% CO$_2$ runs. Arrows indicate the direction of the heat fluxes. Numbers next to the arrows are the amount of heat transport. Numbers in ovals represent the diapycnal heat flux through the bottom of the UW, in which negative means heat from the UW to the DW. Numbers in rectangles are the net heat gain (positive) or loss (negative).

**Fig. 9.** Total isopycnal and diapycnal salt fluxes ($10^6$ kgs$^{-1}$) of the UW. Upper panel represents the control run, and the lower panel for the ensemble 1% CO$_2$ runs. Arrows indicate the direction of the salt fluxes. Numbers next to the arrows are the amount of salt transport. Numbers in ovals represent the diapycnal salt flux through the bottom of the UW where negative means salt from the UW to the DW. Numbers in rectangles are the net salt gain (positive) or loss (negative).

**Fig. 10.** Net surface heat fluxes (W/m$^2$) in CON (number in oval) and in ET (number in rectangle). Positive indicates an oceanic heat loss.

**Fig. 11.** Evaporation minus precipitation (myr$^{-1}$) in CON (number in oval) and in ET (number in rectangle). Positive indicates an oceanic fresh water loss.

**Fig. 12.** Sea ice variations in CON and ET. Numbers in circles represent the ice volume flux in CON, those in rectangles are the ice volume flux in ET. Arrows point out the direction of the ice volume fluxes. “A” represents the ice covered area, “V” the ice volume, “T” the ice thickness, “C” the ice concentration. The percentage number is the percentage changes of the sea ice properties in ET relative to CON. Unit for the ice volume flux is $10^{-3}$ Sv.
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Table 2: Temperature, salinity, and water volume in the Labrador Sea

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Isopycnal and diapycnal salt fluxes ($10^6$ kgs$^{-1}$)

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