

Response of the Atlantic Thermohaline Circulation to Increased Atmospheric CO₂ in a Coupled Model

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(Manuscript received 17 July 2003, in final form 28 May 2004)

ABSTRACT

Changes in the thermohaline circulation (THC) due to increased CO₂ are important in future climate regimes. Using a coupled climate model, the Parallel Climate Model (PCM), regional responses of the THC in the North Atlantic to increased CO₂ and the underlying physical processes are studied here. The Atlantic THC shows a 20-yr cycle in the control run, qualitatively agreeing with other modeling results. Compared with the control run, the simulated maximum of the Atlantic THC weakens by about 5 Sv (1 Sv $\equiv 10^6$ m³ s⁻¹) or 14% in an ensemble of transient experiments with a 1% CO₂ increase per year at the time of CO₂ doubling. The weakening of the THC is accompanied by reduced poleward heat transport in the midlatitude North Atlantic. Analyses show that oceanic deep convective activity strengthens significantly in the Greenland–Iceland–Norway (GIN) Seas owing to a saltier (denser) upper ocean, but weakens in the Labrador Sea due to a fresher (lighter) upper ocean and in the south of the Denmark Strait region (SDSR) because of surface warming. The saltiness of the GIN Seas are mainly caused by an increased salty North Atlantic inflow, and reduced sea ice volume fluxes from the Arctic into this region. The warmer SDSR is induced by a reduced heat loss to the atmosphere, and a reduced sea ice flux into this region, resulting in less heat being used to melt ice. Thus, sea ice–related salinity effects appear to be more important in the GIN Seas, but sea ice–melt-related thermal effects seem to be more important in the SDSR region. On the other hand, the fresher Labrador Sea is mainly attributed to increased precipitation. 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 CO₂, but increased north of 60°N due to the increased flow of 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 CO₂ and other greenhouse gases can change the global evaporation minus precipitation pattern and affect terrestrial runoff. For example, there will likely be more freshwater input into the polar and subpolar seas due to increased precipitation at high latitudes (e.g., Dai et al. 2001a,b). The resulting surface buoyancy flux into these seas will be altered by the freshwater 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 potential impacts of a weakened or possibly even collapsed THC due to human-induced global warming on global climate have raised considerable concern (e.g., Broecker 1987, 1997; Rahmstorf 1999). The Intergovernmental Panel on Climate Change (IPCC) Third Assessment Report outlines widely varying responses of the THC to the projected forcing scenario—IS92a over the twenty-first century in different models (Cubasch et al. 2001). Most of the coupled climate models predict a weakened THC in response to increased atmospheric CO₂ levels, such as the Geophysical Fluid Dynamics Laboratory (GFDL) R15 model (Manabe and Stouffer 1994; Dixon et al. 1999), the Hadley Center model (HadCM3; Wood et al. 1999), the 1995 version of the National Aeronautics and Space Administration Goddard Institute for Space Studies (NASA GISS) coupled model (Russell and Rind 1999), the Canadian General Circulation Model (CGCM1; Boer et al. 2000), a Hamburg model [ECHAM3/Hamburg Large Scale Geo-

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strophic Ocean Circulation Model (LSG); Voss and Mikolajewicz 2001], and the Parallel Climate Model (PCM; Washington et al. 2000; Dai et al. 2001a). Several intermediate-complexity models [e.g., Stocker et al. (1992) coupled model, Stocker and Schmittner (1997) and Schmittner and Stocker (1999); CLIMBER-2, Rahmstorf and Ganopolski (1999); an energy moisture-balance model (EMBM) of the atmosphere coupled to the GFDL Modular Ocean Model 2 (MOM2), 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/isopycnic-coordinate oceanic general circulation model (OPYC3) (Latif et al. 2000), the National Center for Atmospheric Research (NCAR) Climate System (CSM1.3; Gent 2001), and the GISS AGCM coupled with the Hybrid Coordinate Ocean Model (HYCOM; Sun and Bleck 2001).

For those models with a weakened THC, the relative importance of the surface warming and freshening, in general, varies. By specially designing a set of experiments, Dixon et al. (1999) concluded that the enhanced poleward moisture transport due to atmospheric processes contributes the most to the reduction of the Atlantic THC in the GFDL model, while the effects of heat flux changes on the THC are 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 the HadCM3 coupled climate model. On the other hand, those models with a stable THC mentioned earlier show that the effects of surface warming are compensated for by a salinity increase, resulting in little change in the surface ocean density in the 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 examine oceanic changes along isopycnal surfaces, focusing on the regional processes underlying the THC's response to increased atmospheric CO₂ forcing in the PCM. The 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 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's) 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 used by Zhang and Hibler (1997) and 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. 2001a,c; Arblaster et al. 2002).

b. Experiments

The simulations examined here include a 300-yr control run with constant 1990 values for the greenhouse gases [e.g., CO₂ concentration is fixed at 355 ppm; Arblaster et al. (2002)] and an ensemble of four 80-yr transient climate runs with 1% CO₂ increases per year. The control run started from year 49 after the model was fully coupled. The four transient climate runs started from years 21, 101, 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 the CO₂ doubling in ET comparing with that in CON. However, the averaged precipitation north of 60°N increases by about 9% in ET. The equilibrium climate sensitivity of the PCM to the doubled CO₂ is 2.1°C (Meehl et al. 2000b).

c. Analysis method

In this paper, we chose to analyze the PCM ocean output in the density domain since the motion of seawater is along isopycnal surfaces underneath the ocean mixed layer. First, the potential density of the seawater 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 (2002, appendix D) 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 (panel a) and density (panel b) domain averaged over the whole length

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 five 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 four water classes are regrouped into one new class to represent the upper branch of the Atlantic THC based on Fig. 5. The new class is called the upper water (UW). Class 5 water is renamed as the deep water (DW).

Layer	Density (kg m^{-3})	Classes	New classes
1	21.31	Class 1 (UW)	UW
2	22.18	Class 1 (UW)	UW
3	22.96	Class 1 (UW)	UW
4	23.66	Class 1 (UW)	UW
5	24.29	Class 1 (UW)	UW
6	24.86	Class 1 (UW)	UW
7	25.38	Class 1 (UW)	UW
8	25.85	Class 1 (UW)	UW
9	26.27	Class 1 (UW)	UW
10	26.64	Class 2 (UIW)	UW
11	26.96	Class 2 (UIW)	UW
12	27.33	Class 3 (LIW)	UW
13	27.45	Class 3 (LIW)	UW
14	27.62	Class 4 (UDW)	UW
15	27.74	Class 4 (UDW)	UW
16	27.82	Class 5 (LDW)	DW
17	27.87	Class 5 (LDW)	DW
18	27.90	Class 5 (LDW)	DW

of the CON run. The two panels represent a similar pattern of the meridional overturning circulation (MOC) with maximum values greater than 30 Sv ($\text{Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$). The MOC derived in the density domain, however, shows a more detailed structure of the upper ocean without losing the details of the deep ocean. Figure 1b 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 seem 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 is concentrated in a particular isopycnal layer due to the limited number of isopycnal layers used here. By increasing the number of isopycnal layers, this maximum MOC center at 30° – 40°N could be stronger 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).

Note that the MOC in PCM is stronger than the observed estimations, about 17 Sv (e.g., Hall and Bryden 1982; McCartney and Talley 1984; Roemmich and Wunsch 1985). Seidov and Haupt (2003) suggest that the interbasin sea surface salinity (SSS) contrast between the North Atlantic and North Pacific controls the strength of the MOC. A higher interbasin SSS contrast indicates a stronger MOC. In comparison with the Levitus (1982) initial condition, the mean interbasin SSS contrast in this 300-yr control run is more than 30% higher. Thus, this stronger than observed MOC in PCM is likely to result in SSS biases in the North Atlantic and North Pacific Oceans.

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,

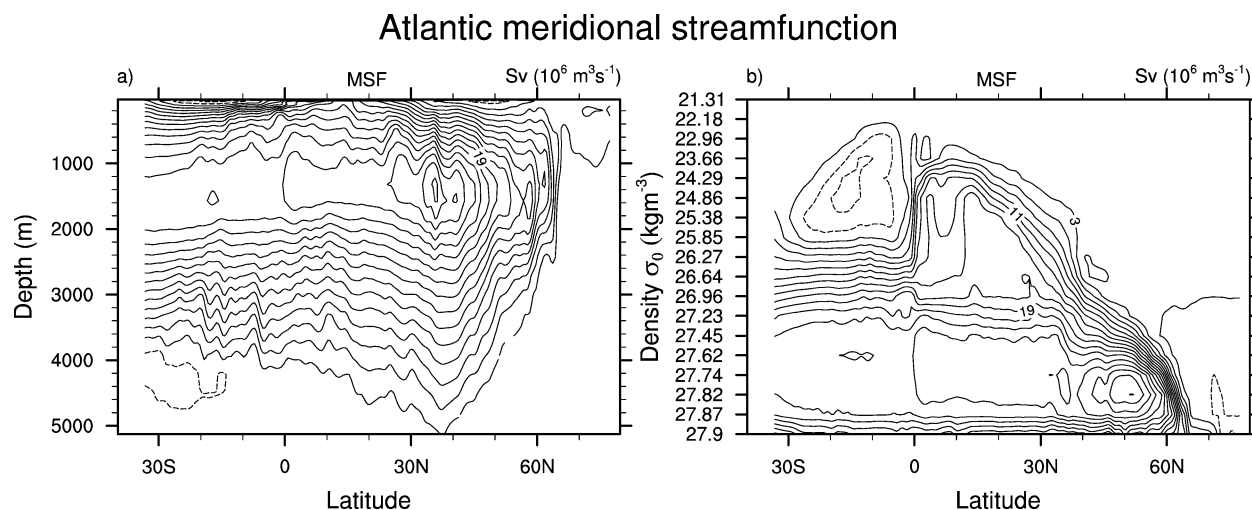


FIG. 1. Atlantic meridional streamfunction derived in the (left) depth domain and (right) density domain averaged over the whole length of the control run. Contour interval is 2 Sv. Note that the density coordinate is nonlinear.

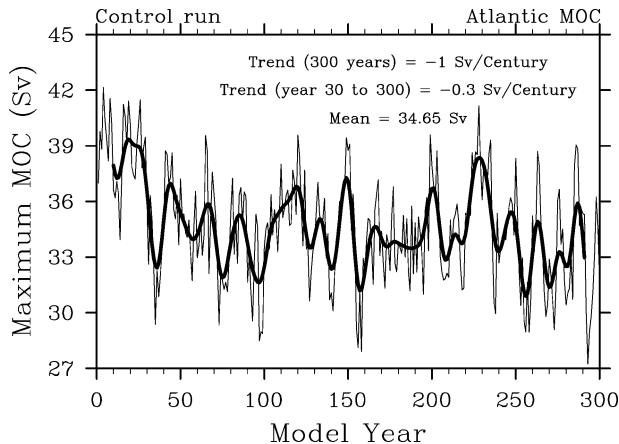


FIG. 2. Time series of the maximum MOC in the North Atlantic in the control run. The thick solid line is the 13-yr low-pass-filtered maximum MOC.

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-yr run is $-1 \text{ Sv century}^{-1}$. However, most of the changes occur during the first 30 yr. The linear trend from years 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-yr cycle significant at the 95% level. Similar decadal MOC oscillations are also suggested by other 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 beyond the scope of this paper.

The ensemble mean North Atlantic MOC shows a steady trend of weakening as CO_2 increases in ET in comparison with that in CON (Fig. 3a). Note that the solid line in Fig. 3a is an average of the four 80-yr segments in CON corresponding to those of ET. The big dip around year 55 in CON is a combined effect of the low MOC years in CON, resulting from the different starting points of each of the four 1% CO_2 ensemble members. For the four ensemble members, year 55 corresponds to years 75, 155, 205, and 255 respectively, in CON, which are all low MOC years.

The year-to-year difference of the maximum MOC between ET and CON, the dashed line in Fig. 3a subtracting the solid line, shown in Fig. 3b (solid line), exhibits large interannual variations. The biggest interannual variation around year 55 is not due to a strengthening of the MOC in ET, but rather to a weakening of the MOC in CON as explained previously. The mean maximum MOC for the last 20 yr (years 61–80) is weaker by 4.7 Sv, approximately 14% of the control run. As a consequence, the northward heat transport is reduced by 0.04 PW ($1 \text{ PW} = 10^{15} \text{ W}$), about 4% at 30°N , and 0.08 PW, roughly 9% at 45°N . However, the heat transport north of 60°N increases, with a peak increase of 15%, about 0.03 PW at 65°N . Similar heat transport changes are reported by Meehl et al. (2000a). 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 Greenland–Iceland–Norway (GIN) Seas.

Figure 4 shows the mean MSF in ET and the MSF difference between ET and CON averaged over the last 20 yr of the 1% CO_2 runs. Comparing Figs. 4a and 1b, the MSF in ET is similar to that in CON, but with a decrease in density for the deep southward-flowing wa-

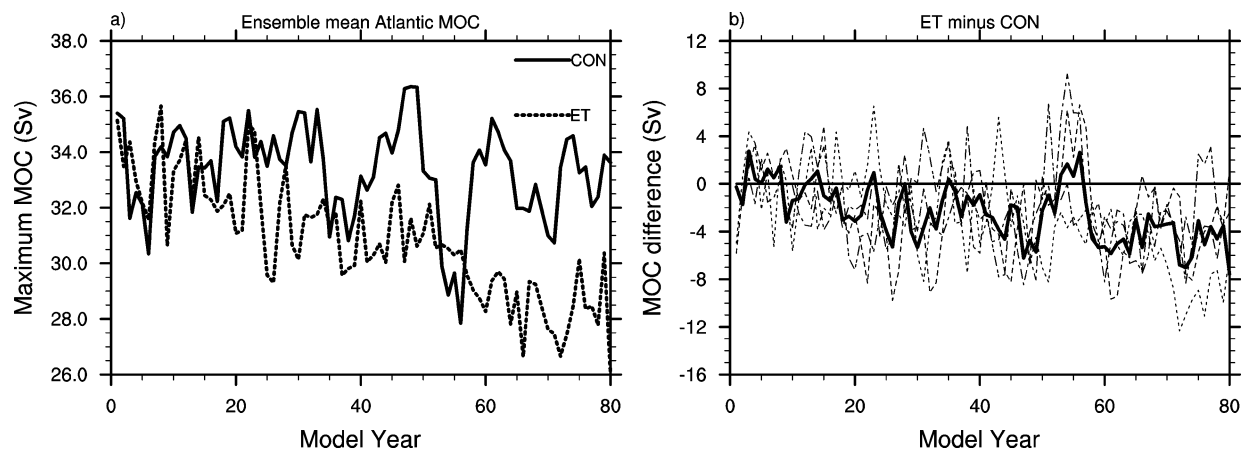


FIG. 3. Time series of (left) the maximum MOC (Sv) in CON and ET and (right) the maximum MOC difference $-1\% \text{ CO}_2$ runs minus that in the control run. In the left panel, the solid line is for CON and dashed line for ET. In the right panel, the thick solid line represents the changes of maximum MOC in ET from CON. The dashed lines are the differences between each ensemble member of the 1% CO_2 runs and the control run.

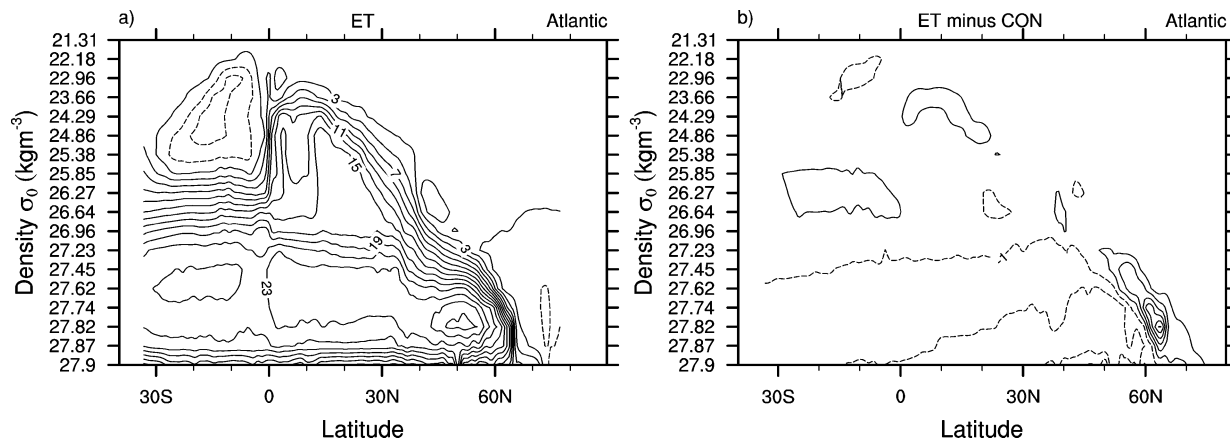


FIG. 4. (left) The mean MSF in ET averaged over years 61–80. (right) The MSF in ET minus that in CON, averaged over the last 20 yr of each of the 1% CO_2 runs for ET and the corresponding 20-yr period of the control run for CON. Contour interval is 2 Sv.

ter. But the MOC in ET is weaker by up to 5 Sv in the North Atlantic (south of 60°N). 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 the regional variability of deep convective activity in the northern North Atlantic marginal seas, the water mass is grouped into five 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 subdomains, namely the Labrador Sea ($45^\circ\text{--}65^\circ\text{N}$, west of 45°W), south of the Denmark Strait region (SDSR, $45^\circ\text{--}65^\circ\text{N}$, 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} generally flows northward and water with a density of 27.82 kg m^{-3} or heavier than 27.82 kg m^{-3} flows southward in the North Atlantic region in both CON and ET runs (Figs. 1b and 4a), we further simplify our three-dimensional analysis by grouping the first four water classes into a new class to represent the upper branch of the Atlantic THC, referred to as the upper water (UW). Class 5 water represents the lower branch of the THC, named the deep water (DW, Table 1). Hereafter, these two water classes are referred as the UW and the DW.

The time evolution of the ensemble mean diapycnal fluxes, a conversion of UW to DW, is shown in Fig. 6 for the Labrador Sea, SDSR, and the GIN Seas. Relative

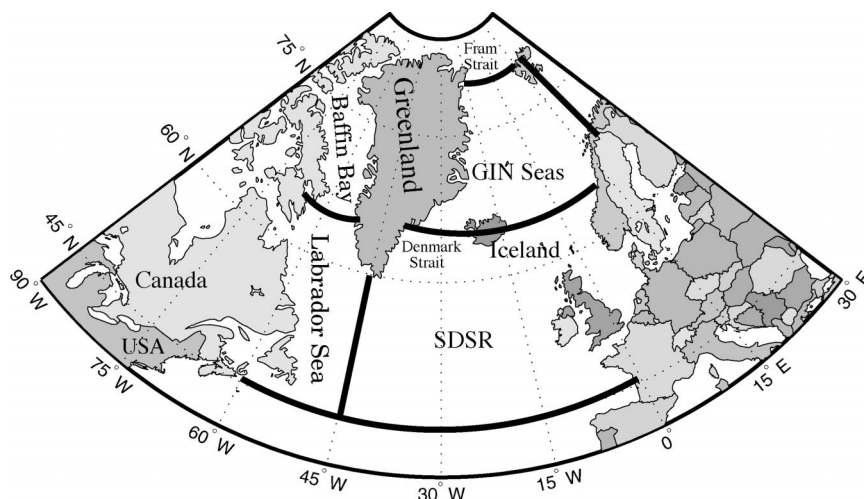


FIG. 5. Northern North Atlantic map. The four subdomains of interest are defined as the Labrador Sea ($45^\circ\text{--}65^\circ\text{N}$, west of 45°W), the south of the Denmark Strait region (SDSR; $45^\circ\text{--}65^\circ\text{N}$, east of 45°W), the GIN Seas (roughly $65^\circ\text{--}80^\circ\text{N}$, east of 45°W –west of 20°E), and the Baffin Bay (roughly $65^\circ\text{--}80^\circ\text{N}$, west of 45°E –east 80°W). The major deep ventilations occur in the GIN Seas, the Labrador Sea, and the SDSR.

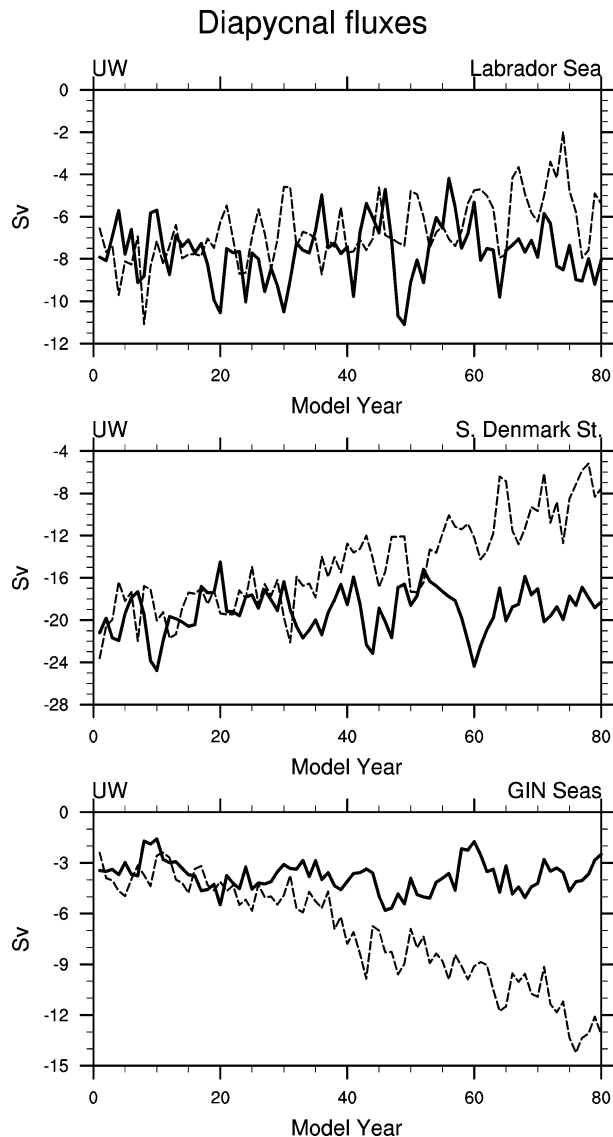


FIG. 6. Time series of the diapycnal fluxes in the Labrador Sea, the SDR, and the GIN Seas. Solid lines are for CON and dashed lines for ET.

to the control run, the diapycnal fluxes in ET exhibit a decreasing trend for the first two regions, and a trend toward strengthening for the last region. Those changes in diapycnal fluxes become significant after 30 yr of the CO_2 ramping in the SDR and GIN Seas. In the Labrador Sea, they show up primarily in the last 20 yr. In all three regions, the trend of diapycnal fluxes is never reversed. Therefore, by focusing on the last 20 yr, we should be able to detect the variations of the North Atlantic three-dimensional mass fluxes in ET compared with those in CON.

Figure 7 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 yr of each of the transient climate runs (years 61–80)

for ET and the corresponding 20-yr periods of the control run for CON. Henceforth, all model data discussed are averaged over the same period as in Fig. 7. The arrows indicate the direction of the isopycnal flow and the number next to the arrow is the amount of the volume flux (Sv). The numbers located at the center of each subdomain 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 near the top of Greenland represents the Bering Strait inflow from the Pacific into the Arctic.

In general, the net northward-flowing UW across 45°N is reduced, but the UW flowing into the GIN Seas is increased in ET. The upper two panels in Fig. 7 show that the reduction of this UW across 45°N is about 3.4 Sv, roughly a 12% decrease from that in CON. The net North Atlantic water flowing into the GIN Seas through the Iceland–Norwegian channel is doubled in ET. At Denmark Strait, the UW flow reverses its direction from a net southward flow to a net northward flow. The resulting net UW flowing into the GIN Seas from the SDR is close to 6 times higher in ET than it is in CON.

The diapycnal fluxes vary differently in ET among the Labrador Sea, SDR, and the GIN Seas. In the former two regions, the diapycnal fluxes exhibit a significant weakening with increasing CO_2 . The total reduction of the UW to DW conversion is roughly 45% from a total of 26.6 Sv in CON to 14.7 Sv in ET. In the GIN Seas, however, deep convective activity is dramatically increased by 2 times from 3.8 Sv in CON to 11.3 Sv in ET. Because of this strengthening of the deep convective activity in the GIN Seas, the overall change of the diapycnal fluxes in these three subdomains is only moderately reduced by about 4.4 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 by approximately 60% in ET (Fig. 7). The DW inflow through the Iceland–Norwegian channel is reduced by 40%. Summing the flow values between Greenland and Norway, the net DW outflow from the GIN Seas increases from only 2.6 Sv in CON to 10.5 Sv in ET. The exchanges of DW between the SDR and the Labrador Sea are also reduced in ET, resulting in a western boundary current that is weakened by 50%. A significant amount of the DW crossing 45°N flows southward to regions east of the mid-Atlantic ridge, consistent with Wood et al. (1999).

Overall, variations of the Atlantic MOC in ET are determined by two competing processes in PCM: a weakening of UW to DW conversion in the Labrador Sea and SDR, and an intensification in the GIN Seas.

Although we do not focus on the Arctic, it also plays a role on the MOC. As shown in Figs. 7a,b, about 1 Sv of DW from the GIN Seas is converted into UW in the Arctic.

It should be noticed that the volume fluxes in each

Isopycnal and diapycnal fluxes (Sv)

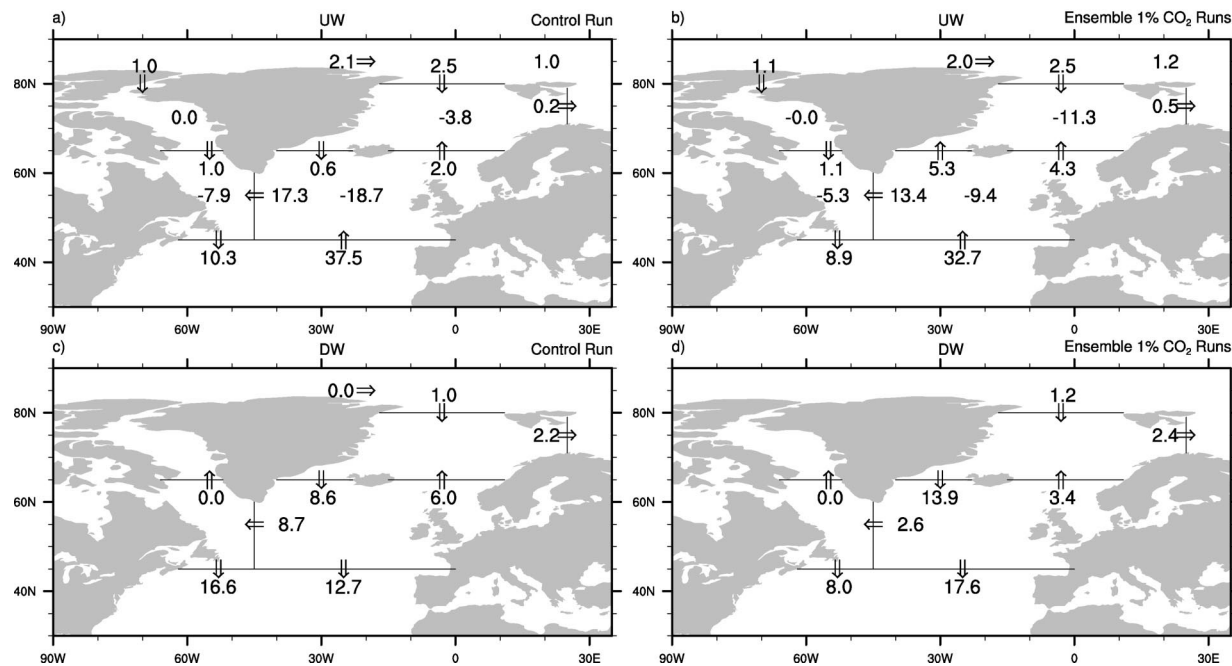


FIG. 7. Isopycnal and diapycnal fluxes (Sv) for (left) CON and (right) ET. (top) The UW representing the upper branch of the THC. (bottom) DW, which represents the lower branch of the THC. Arrows indicate the direction of the isopycnal flows, while the numbers next to the arrows give the amount of the isopycnal flows. The numbers at the center of each box represent the strength of the diapycnal fluxes. The numbers in upper-right corner in the upper two panels are the diapycnal fluxes in the Arctic. A negative number means lighter water converted into denser water.

of the subdomains are not exactly balanced. In general, the imbalance is small. This slight imbalance is related to the changes of the UW/DW volume in each subdomain, which is also induced partly by the displaced North Pole ocean grid when we tried to calculate the isopycnal fluxes along a given latitude or longitude.

c. Linkage to changes in water properties

The changes in diapycnal fluxes can be directly linked to variations in water mass density in these subdomains. Water density is primarily controlled by temperature and salinity. However the contribution of the temperature and salinity to the UW density variations in ET is different in these subdomains of the North Atlantic. Table 2 shows the UW temperature, salinity, and density changes in ET relative to those in CON averaged for

the whole time series and the last 20 yr. Basically, the direction of those changes is the same no matter whether it is for the 80-yr mean or for the last 20-yr mean in all concerned subdomains, for example, a cooling and freshening in the Labrador Sea for these two periods. From Table 2, it is also clear that the changes in UW properties become more obvious in the last 20 yr, which indicates a cumulative effect induced by changes of the atmospheric CO_2 level. In the GIN Seas, UW is heavier in ET. This heavier UW intensifies the winter deep convective processes thereby weakening the oceanic vertical stratification. This UW density increase is primarily induced by a salinity increase. The UW temperature change in the GIN Seas works against this density change. In the Labrador Sea, the salinity change also dominates the UW density change. The freshening there, overcoming the cooling effect, results in a decrease in UW density. On the other hand, the decrease of the UW density in the SDSR is caused by a warming, which offsets the effect of the salinity increase. In the latter two regions, the lighter UW strengthens the oceanic vertical stratification and leads to weakened deep convective activity.

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.

TABLE 2. Temperature, salinity, and density changes.

	Mean (yr)	Labrador Sea	GIN Seas	SDSR	Baffin Bay
Temperature ($^{\circ}\text{C}$)	80	-0.06	0.88	0.10	-0.065
	20	-0.14	2.02	0.25	-0.176
Salinity (ppt)	80	-0.042	0.087	0.004	-0.022
	20	-0.097	0.231	0.014	-0.058
Density (kg m^{-3})	80	-0.027	0.011	-0.013	-0.013
	20	-0.065	0.043	-0.028	-0.033

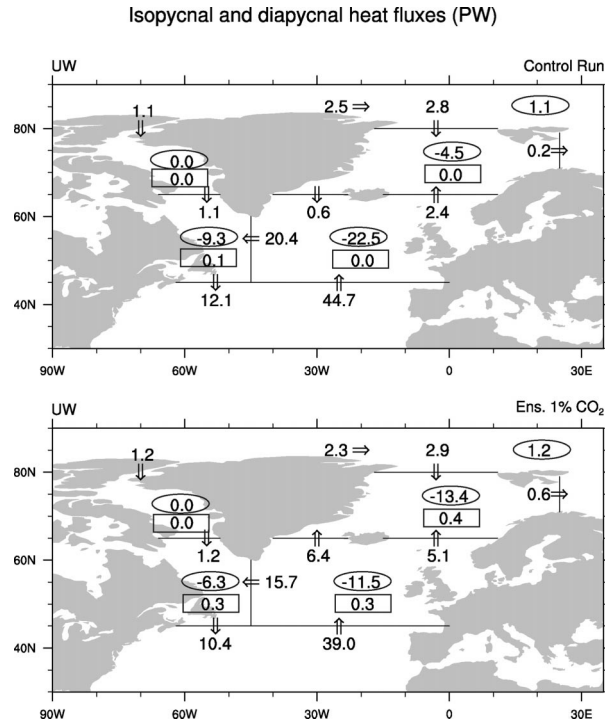


FIG. 8. Total isopycnal and diapycnal heat fluxes in PW (10^{15} W) of the UW. (top) The control run, and (bottom) the ensemble 1% CO_2 runs. Arrows indicate the direction of the heat fluxes. Numbers next to the arrows are the amount of heat transport. Numbers inside the ovals represent the diapycnal heat flux through the bottom of the UW, in which a negative value means heat travels from the UW to the DW. Numbers inside the rectangles are the net heat gain (positive) or loss (negative).

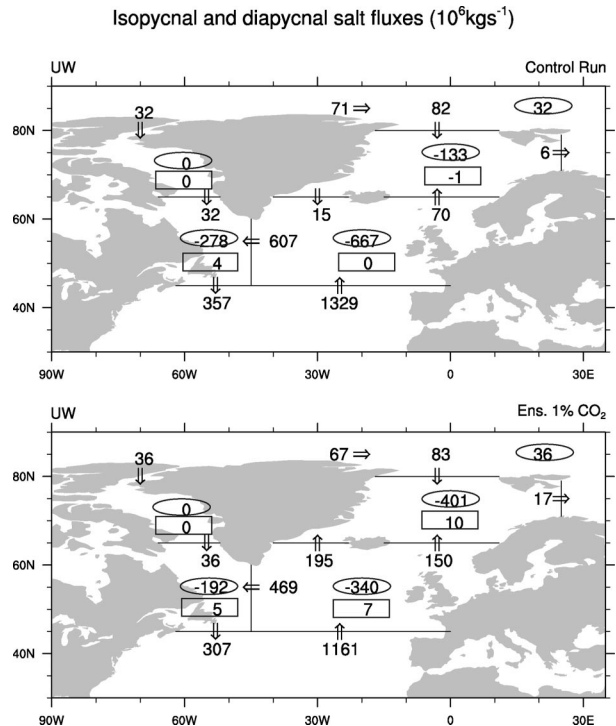


FIG. 9. Total isopycnal and diapycnal salt fluxes (10^6 kg s^{-1}) of the UW. (top) The control run, and (bottom) the ensemble 1% CO_2 runs. Arrows indicate the directions of the salt fluxes. Numbers next to the arrows are the amount of salt transport. Numbers inside the ovals represent the diapycnal salt flux through the bottom of the UW, where a negative value means salt travels from the UW to the DW. Numbers inside the rectangles are the net salt gain (positive) or loss (negative).

4. Physical processes

The water property changes in the North Atlantic have been linked to the variations of the diapycnal flux intensity there in the previous section. In turn, these water property changes can be caused by changes in oceanic transport processes, surface heating and cooling, net surface freshwater 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 variations in isopycnal and diapycnal flows, resulting in a change in water temperature, salinity, and density. Figures 8 and 9 show heat and salt fluxes. The numbers inside the rectangles represent the net heat or salt gain (positive) or loss (negative) due to both isopycnal and diapycnal transport processes in the region and the numbers inside the 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:

$$H_t = \int_{x_1}^{x_2} \int_{y_1}^{y_2} \rho c_p T v \, dx \, dy \quad \text{and} \quad (1)$$

$$H_s = \int_{x_1}^{x_2} \int_{y_1}^{y_2} \rho S v \, dx \, dy, \quad (2)$$

where H_t represents heat flux, H_s represents salt flux, $\rho = 1000 \text{ kg m}^{-3}$ is the water reference density, $c_p = 3996 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat of seawater, T is the water temperature in kelvins, v is velocity, and S is 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 is reduced by 4 PW (bottom panel in Fig. 8). This number is significantly larger than the net reduction of meridional heat transport (MHT) as shown

in section 3a, which implies that the southward MHT carried by DW also decrease by a similar amount.

In the SDSR, there is no net heat gain or loss due to oceanic transport processes in CON. In ET, this region shows a 0.3 PW net heat gain, indicating that the UW warming there is at least partly due to these oceanic transport processes. The weakening of the diapycnal volume flux in the SDSR in ET as shown in section 3b induces a 50% reduction in downward heat transport (Fig. 8), and a weaker heat transport from the SDSR to the Labrador Sea also contributes to this net heat gain. But the increased MHT from the SDSR to the GIN Seas keeps this net gain small.

In the GIN Seas, the net heat gain of the UW due to oceanic transport processes is 0.4 PW in ET, but 0 PW in CON. This higher heat flux convergence is mainly contributed by the increased North Atlantic water inflow through Denmark Strait and the Iceland–Norway channels with a net heat flux of 11.5 PW in comparison with that of 1.8 PW in CON. Most of the increased heat flux into this basin is transported into denser layers due to the intensified diapycnal flux there. Only a small portion of that heat flux is used to heat up the UW.

The net heat gain of the UW in the Labrador Sea also increases in ET relative to that in CON (Fig. 8), owing to the reduced diapycnal heat transport and a weaker southward heat transport. Thus the oceanic transport processes favor a warmer UW in this region, opposite to the UW temperature change shown in section 3c.

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 (Fig. 9). The net reduction of about $118 \times 10^6 \text{ kg s}^{-1}$ in ET is approximately a 12% decrease from CON. This percentage is slightly lower than the reduction in water volume transport (12.5%), implying a slightly saltier UW to the south of 45°N in ET than in CON. In fact, the increase in the atmospheric CO_2 level causes an increase in net evaporation in the subtropical Atlantic, which induces a saltier UW in those regions.

In the SDSR, the oceanic transport processes do not cause a change in the total salt in UW in CON, but they induce a $7 \times 10^6 \text{ kg s}^{-1}$ net salt gain in ET. This salt gain is consistent with the salinity change in ET. This salt gain is related to a weaker diapycnal salt flux, and a reduced salt transport from the SDSR to the Labrador Sea; however, the increase in salt transport from the SDSR to the GIN Seas means the salt gain is small.

In the Labrador Sea, the salt gain in UW is increased by 25% in ET compared to that in CON (Fig. 9), which is opposite from the UW salinity changes in this basin. The increase in salt gain is related to the weakening of the diapycnal salt flux and the reduction of the southward salt export in ET. Therefore the salinity changes in this subdomain shown in Table 2 must be due to

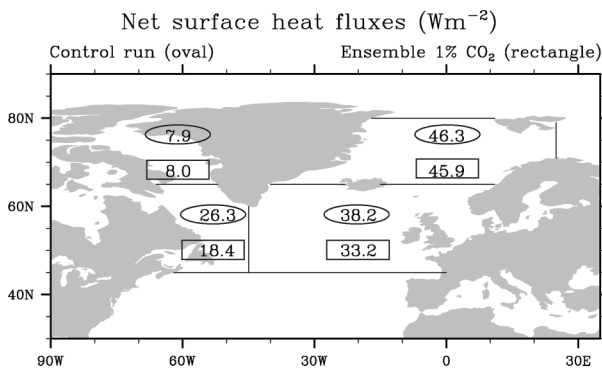


FIG. 10. Net surface heat fluxes (W m^{-2}) in CON (number inside the oval) and in ET (number inside the rectangle). Positive indicates an oceanic heat loss.

changes in precipitation and evaporation from the atmosphere noted in the next section.

The UW salt flux entering the GIN Seas is dramatically increased. The net salt flux from the SDSR to the GIN Seas jumps 6 times in ET relative to CON. The net salt flux from the Arctic Ocean is slightly reduced. The diapycnal salt flux in ET, however, is significantly increased due to the strengthening of the deep convection there. Overall, the net salt gain due to oceanic transport processes in this basin is about $10 \times 10^6 \text{ kg s}^{-1}$ in ET in comparison to a net salt loss in CON.

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

b. Air–sea interaction

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

The net surface heat flux, a sum of the latent and sensible heat fluxes, and the net shortwave and longwave radiation fluxes, is shown in Fig. 10. The net surface freshwater flux, a sum of evaporation, precipitation, and river runoff, is shown in Fig. 11. Positive values indicate an oceanic heat or freshwater loss. In general, the oceanic heat loss is lower in ET than in CON, and the freshwater gain is higher in ET than in CON (except the SDSR) in all four subdomains. The reduced heat loss is induced by a larger warming in the surface air than in the surface ocean. The increased freshwater gain, mainly due to precipitation increases, contributes to a fresher surface ocean.

The combined effects of surface heat and freshwater fluxes should lead to warmer and fresher UW in ET in the Labrador Sea. However, the model shows a colder and fresher UW (see Table 2 and the last line in Table 3). As shown in Table 3, the class 1–3 waters are warmer in ET than in CON and the class 1 and 2 waters are

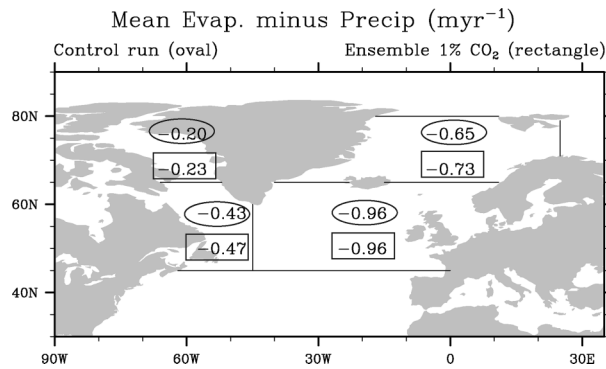


FIG. 11. Evaporation minus precipitation (m yr⁻¹) in CON (number inside the oval) and in ET (number inside the rectangle). Positive indicates an oceanic freshwater loss.

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 subsurface 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 1–3 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 freshwater 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

Both the UW temperature and salinity in a region can be affected by variations in sea ice conditions, which include variations in net sea ice flux convergence in a region and the changes in sea ice cover. An increase in ice flux convergence implies 1) an increased freshwater input, contributing to a decrease of UW density, and 2) more heat is needed to melt this ice, resulting in a decrease in UW temperature, contributing to an increase in UW density. An increase in ice cover indicates 1) a decrease in oceanic heat loss due to enhanced insulation effect, and 2) an increase in UW salinity due to increased salt ejection during the ice formation period. The net

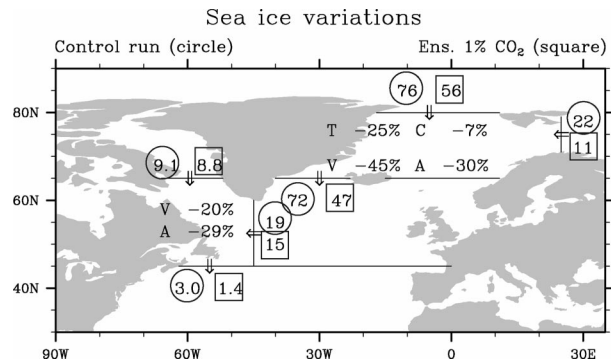


FIG. 12. Sea ice variations in CON and ET. Numbers inside the circles represent the ice volume flux in CON, those inside the squares are the ice volume flux in ET. Arrows point to the direction of the ice volume fluxes. Here, “A” represents the ice covered area, “V” the ice volume, “T” the ice thickness, and “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.

contribution of sea ice condition changes is not expected to be the same in different regions as discussed later in this section.

In general, the sea ice extent in the Arctic (not shown) and the ice export through the 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 fluxes from the Arctic Ocean to the GIN Seas through the Fram Strait and through the Barents Sea are reduced by 26% and 50% in ET compared to those in CON, respectively. And the ice flux exiting at the Denmark Strait is also reduced by 35% 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 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 UW salinity and 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 the ice flux at the Denmark Strait and out of the SDSR into the Labrador Sea, is reduced from 0.053 Sv in CON to

TABLE 3. Temperature, salinity, and water volume in the Labrador Sea.

Class	CON			ET		
	<i>T</i> (°C)	<i>S</i> (ppt)	<i>V</i> (10 ¹⁴ m ³)	<i>T</i> (°C)	<i>S</i> (ppt)	<i>V</i> (10 ¹⁴ m ³)
1	2.378	32.596	0.287	2.637	32.548	0.330
2	-0.023	33.491	0.765	0.025	33.472	0.891
3	1.443	34.223	1.290	1.934	34.264	1.232
4	4.630	34.995	4.513	4.546	34.976	3.768
1–4	3.420	34.582	6.856	3.289	34.492	6.221

0.032 Sv in ET. The impact of this reduction in ice flux convergence in this region is twofold. First, it means a less freshwater input into this region, resulting in a positive contribution to the salinity changes. Second, 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 both 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 the 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 the 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. At the same time, the evaporation is significantly increased in the subtropical Atlantic region. As a result, the northward-flowing Atlantic water becomes warmer and saltier in ET than in CON. When this water reaches the Labrador Sea and the SDSR, it is not dense enough to sink into the abyssal sea. Therefore part of this warmer and saltier North Atlantic water flows into the GIN Seas, leading to an increase in water volume transport into the GIN Seas as shown in Fig. 7 and discussed in section 3b, since no physical processes or topography can totally constrain this from happening.

A further analysis indicates that the warming and the 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, and the intensification of the deep convective activity in the GIN Seas. 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 contributing to strengthening the deep convective activity in the GIN Seas.

Now the question is whether the intensified deep convective activity induces an increase in North Atlantic

water inflow, or the other way around. This cannot be answered by direct model data diagnosis. Our speculation is that the intensified deep convective activity in the GIN Seas draws in more North Atlantic water. As atmospheric CO_2 increases, the ice cover in the GIN Seas reduces and the ice flux from the Arctic into the GIN Seas also decreases. These lead to a saltier UW in summer, and thus a less stratified upper ocean, since there is less available ice to be melted in ET than in CON. At the latitudes of the GIN Seas, the salinity perturbation becomes more important to the variation in water density. In winter, a salinity-induced increase of density, plus the winter cooling effect, destroy the weak stratification in the GIN Seas and enhance the deep convective activity there. This enhanced deep convective activity drives a stronger Denmark Strait overflow—more DW flowing out of the GIN Seas as shown in Fig. 7d. The latter requires more North Atlantic water to flow into the GIN Seas. The salt brought in by the North Atlantic water further increases the UW salinity in the GIN Seas, causing a positive feedback to the strength of the deep convective activity there and the strength of the Denmark Strait overflow. Model data support this theory since the increase in UW density and diapycnal fluxes mainly happens in the Greenland Sea and along the east coast region of Greenland, where they are most strongly influenced by sea ice condition changes.

A couple of caveats must be included here. 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 3b. Also, the effects of surface winds on air–sea heat and water fluxes are implicitly included in the surface fluxes discussed in section 4b.

5. Summary and conclusions

The North Atlantic THC in a 300-yr control run and a four-member ensemble of 1% CO_2 transient runs using PCM have been analyzed for decadal and long-term changes. The Atlantic THC shows a 20-yr cycle in the control run, qualitatively agreeing with other coupled climate 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 mass fluxes weaken by roughly 45% in the Labrador Sea and the south of the Denmark Strait region, and strengthen by approximately 2 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

CO₂. 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 subdomains.

In the Labrador Sea, the increased net freshwater 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 in the reduced oceanic deep convection. This warming is induced by a reduced heat loss to the atmosphere, an increased heat convergence due to oceanic transport processes, and a reduced sea ice volume flux from the GIN Seas into this region. The latter reduces the heat used to melt ice, indirectly contributing to the warming in the 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 the increased oceanic transport of salty water into the region are the main factors for the increases in the upper-ocean density and deep convection. The effect of UW warming on density, induced by an increased heat convergence by the oceanic transport process and reduced heat loss to the atmosphere, 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 flow of North Atlantic water across this latitude.

Acknowledgments. The authors thank the two anonymous reviewers for their constructive comments and suggestions. The authors thank Dr. Rainer Bleck for his kindness in providing the Fortran code that was used to convert POP ocean data to the isopycnic domain. The authors also thank Dr. Dan Seidov for his comments on the mechanisms affecting THC strength in the PCM. Thanks also go to Julie Arblaster and Gary Strand for their help in processing the PCM data. A portion of this study was supported by the Office of Biological and Environmental Research, U.S. Department of Energy, as part of its Climate Change Prediction Program.

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