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Changes in Thermohaline Circulation in Future Climate

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Abstract

Anthropogenically-induced changes of the thermohaline circulation could be potentially important in future climate regimes. By employing a coupled GCM (Parallel Climate Model, PCM), the maximum Atlantic thermohaline circulation weakens by about 3 Sv or 10% in an idealized forcing (1% CO2 increase) transient climate experiment at time of CO2 doubling compared with the control run. The weakening of the THC is accompanied by reduced poleward heat transport in mid-latitudes. Further analysis indicates that the weakening of the THC seems mainly caused by the freshening of the surface ocean in the northern North Atlantic region. However, variation of the deep convection intensity in this region is not uniform. In general, it shows a significant weakening of the deep convection in the Labrador Sea region and a mild strengthening of the deep convection at the south of the Denmark Strait region. Because increased CO2 induces warming and freshening effects, the density of the surface water is reduced when it reaches the Labrador Sea. Thus the intensity of deep convection there is weakened. Conversely, increased salinity at the south of Denmark Strait intensifies the deep convection there.

1. Introduction

Thermohaline circulation is a density driven global scale oceanic circulation, which plays an important role on global meridional heat transport. Variability of thermohaline circulation (THC) intensity would modify the global meridional heat transport, thus affecting the global climate. The anthropogenically induced global warming changes the global E-P pattern and also alters the runoff and fresh water storage on the landmass. It is likely that there will be more fresh water flux input in the polar and sub-polar seas. Therefore the surface buoyancy flux in these seas may change with the fresh water flux anomalies. Since the sinking branch of the THC is highly localized in the northern North Atlantic marginal seas and in the southern ocean, variations in surface buoyancy flux leads to a weakened deep convective process in these seas due to the additional fresh water input. The weakening of the deep convection results in a weakened THC, and a reduced poleward heat transport.

In this paper, we employ a global coupled GCM – the Parallel Climate Model (PCM) to investigate the variability of THC induced by anthopogenic forcing.

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2. Model and Experiments

The model used in this paper is the Parallel Climate Model, which includes four component models: atmospheric, ocean, land and sea ice. The atmospheric component of PCM is the National Center for Atmospheric Research's (NCAR) Community Climate Model (version 1.3); the oceanic component of PCM is the Los Alamos National Laboratory's Parallel Ocean Program (POP). The Land component is NCAR's Land Surface Model (LSM). And the sea ice model is a version of Zhang and Hibler (1997) optimized for the parallel computer environment required by PCM. For more details of the PCM, please see Washington et al (2000).

The simulations examined here are a 300-year control run with constant 1990 values of the greenhouse gases (e.g. CO2 concentration is fixed at 355 ppm) and four 1% CO2 increase transient climate ensemble runs. The globally averaged first level ocean temperature for all runs are shown in Figure 1. At the time of CO2 doubling, the global mean ocean surface temperature is increased about 1 °C.

3. Results

3.1 Analysis Method

Since the primary concern of this study is the THC, and the motion of sea water is along isopycnal surfaces underneath the surface mixed layer, this motivates us to analyze the PCM ocean output in the density domain. We first calculated the potential density of the model data, referenced to the sea surface. Then, the water mass is interpolated into a set of isopycnal layers in a manner of conserving mass, salt, heat and momentum. The method used here was developed by Rainer Bleck and Sumner Dean at Los Alamos National Laboratory.

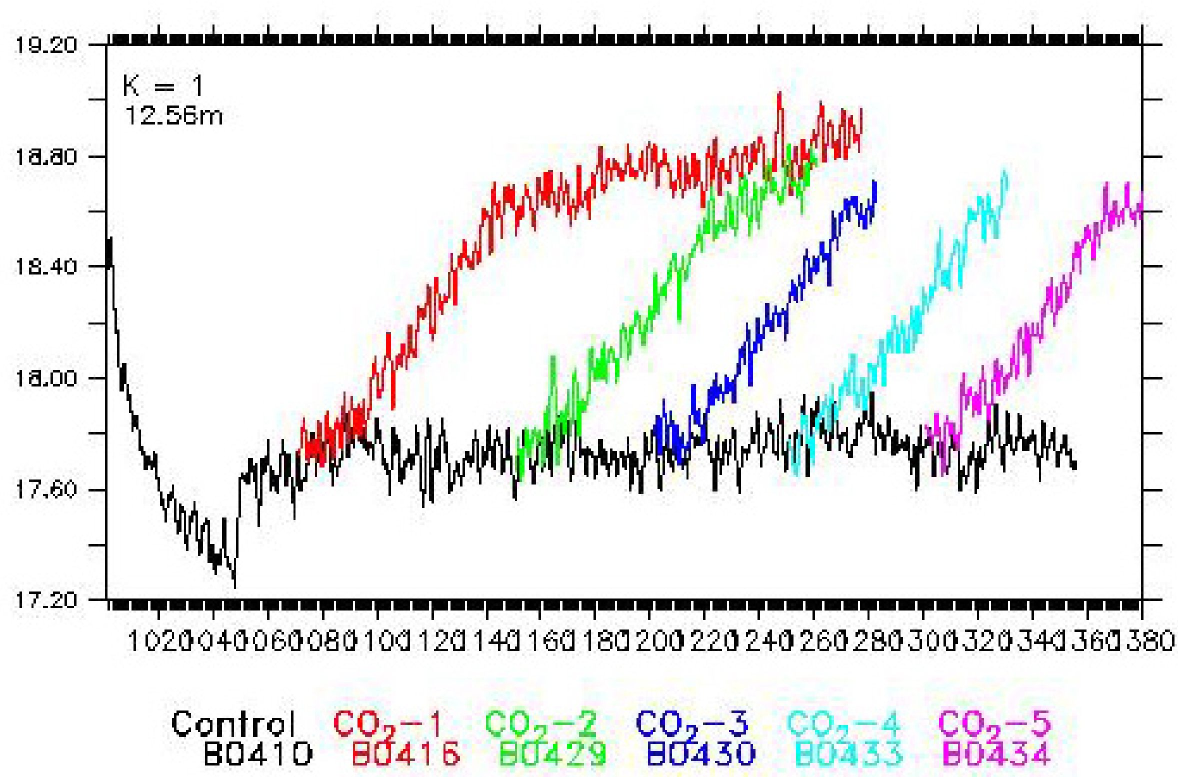

Fig. 1: Globally averaged first level ocean temperature. The dark solid line is for the control run and the colored lines are for each of the ensemble members of the 1% CO2 transient experiments. Only the first four of the ensemble experiments are used in this paper.

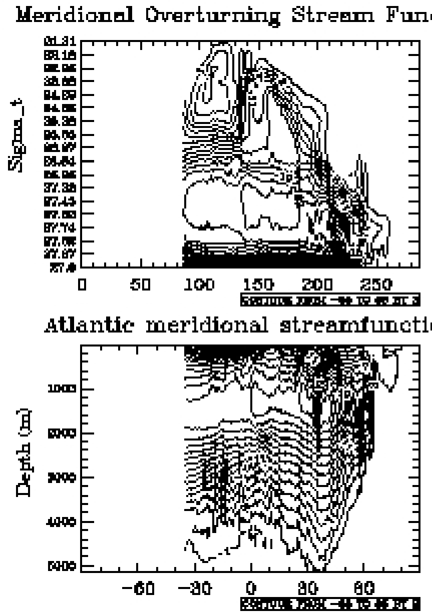

Fig. 2 The Atlantic Meridional Stream Function (MSF). The upper panel is the MSF derived from density domain and the lower panel is the MSF derived from depth domain. The x axis of the upper panel is the model grid from Southern Hemisphere to Northern Hemisphere. That of the lower panel is the latitude.

Figure 2 gives the Atlantic Meridional Stream Function derived from both density (upper panel) and depth (lower panel) data. The two panels represent a similar pattern of the Meridional Overturning Circulation (MOC). The MOC derived from thedensity domain shows more detailed structure of the upper ocean without losing any details of the deep ocean. The upper panel also shows that as the upper ocean water flows northward, due to the atmospheric cooling effect, the water becomes denser. When the surface water is dense enough, it sinks to the deep ocean and flows southward.

3.2 Results

Figure 3 is the time series of the North Atlantic maximum MOC difference between each of the ensemble members of the 1% CO_2 transient runs and the control run. The thick solid line is the mean difference between ensemble 1% CO_2 transient runs and the control run. In general, the ensemble 1% CO_2 runs show a weakening of the Atlantic MOC after year 20. Except for the decadal time scale oscillation, the Atlantic MOC is weakened about 3 Sv (10^6 m^3/s), which is about 10% of MOC strength in the control run.

Fig. 4 shows the difference of the MOC between the ensemble mean and the control run averaged over the last 5-year of the 1% CO_2 runs. The strength of the MOC is weakened more than 3 Sv in the North Atlantic. This weakening of the MOC implies a decreased poleward heat transport by ocean, which would contribute to variations in the northern North Atlantic marginal seas and the Arctic.

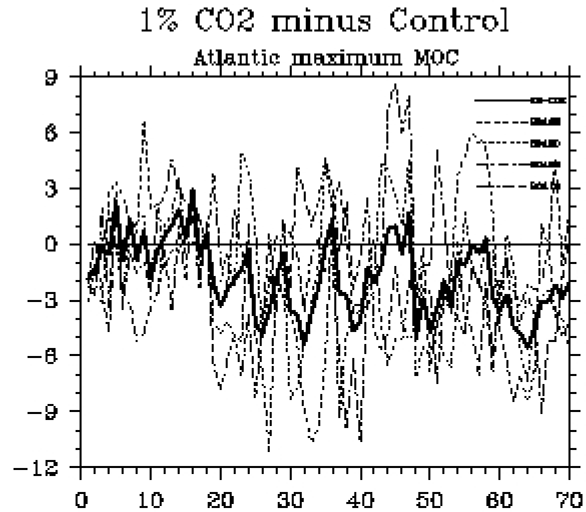

Fig. 3 Time series of the maximum North Atlantic MOC rate differences between each of the ensemble 1% CO_2 runs and the control run. Dashed lines represent each of the ensemble members and solid line represent the mean of the ensembles.

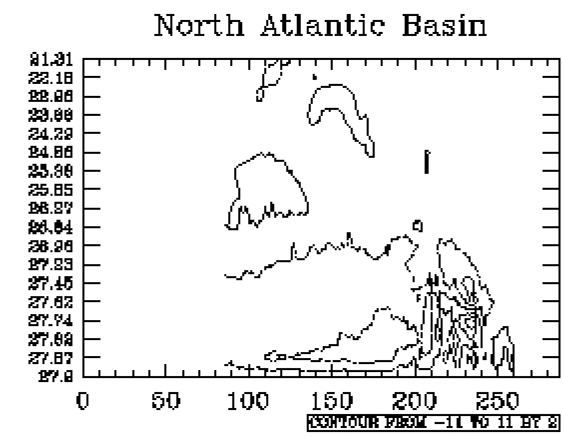

Fig. 4 Difference of the MOC between ensemble mean and the control run averaged over the last five years of the 1% CO_2 transient runs.

It should be also noticed in Figure 4 that in the region of Northern North Atlantic marginal seas, the variation of the MOC is not uniform vertically. It seems the convection process is weakened in the upper ocean, and strengthened at mid-depth. Since the MOC is zonally and vertically integrated, the zonal differences of the deep convective activity are neglected in this figure.

Layer	Density	Class
1	21.31	
2	22.18	
3	22.96	
4	23.66	Class 1 UW
5	24.29	
6	24.86	
7	25.38	
8	25.85	
9	26.27	
10	26.64	Class 2 UIW
11	26.96	
12	27.23	Class 3 LIW
13	27.45	
14	27.62	Class 4 UDW
15	27.74	
16	27.82	
17	27.87	Class 5 LDW
18	27.90	

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), upper intermediate water (UIW), lower intermediate water (LIW), upper deep water (UDW), and lower deep water (LDW).

To further analyze the variability of deep convective activity in the Northern North Atlantic marginal seas, the water mass is grouped into 5 density classes after Schmitz (1996). As shown in figure 5, the southward water transport is 28 Sv in the control run and about 25 Sv in the ensemble 1% CO2 run. Locally, the weakening of the diapycnal fluxes is not uniform. In the Labrador Sea region, the diapycnal flux changes from 4.1 Sv in the control run to 1.6 Sv in 1% CO2 runs, a decrease of 61%. In the GIN Seas and south of the Denmark Strait region, the diapycnal flux varies from 23.2 Sv in the control run to 21.3 Sv in the 1% CO2 runs, only a decrease of 8%. This indicates a non-uniform change of deep convective activity induced by the global warming effect due to anthropogenic forcing.

The diapycnal fluxes from one class to another in the Labrador Sea and GIN Seas are summarized in Table 2. It shows the diapycnal fluxes in the upper ocean have weakened by more than 3 Sv in the Labrador Sea in the 1% CO2 run relative to the control run. The decrease of the diapycnal fluxes from class 4 to class 5 is not as large as that in the upper ocean. This indicates an upward movement of the lower branch of the MOC since the density of this southward flow is lighter.

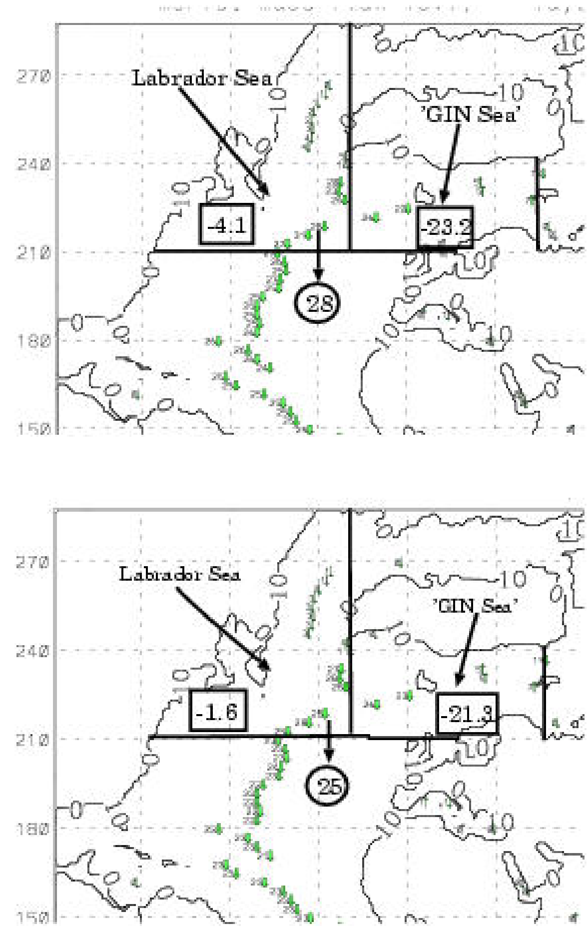

Fig. 5 Meridional transport of the class 5 water (LDW) in the northern North Atlantic region. The upper panel is for the control run and the lower panel for the ensemble 1% CO2 runs. All data are averaged over the last 5-years of the 1% CO2 transient runs. The numbers in circles represent the meridional transport. The numbers in squares represent diapycnal fluxes. Negative means downward water mass conversion (such as from class 4 (UDW) to class 5 (LDW)).

On the other hand, the diapycnal flux shows a significant strengthening in the upper ocean in the GIN Seas in the 1% CO2 runs, especially the diapycnal flux increases from class 2 to class 3 from 0.2 Sv in the control run to 3.2 Sv in the 1% CO2 run. This implies that the effects from increase CO2 induced oceanic warming and decreased the density of the northward flowing water in North Atlantic. This water needs to move into the GIN Sea to convert into a denser class water. In the other words, the natual effect of atmospheric cooling on the North Atlantic water is reduced due to the greenhouse gas induced warming.

Diapycnal Fluxes for each class (Unit: Sv = 10^3 m^3/s)

	Ens-Control Run		Ens-CO_2 doubling	
	Lab. Sea	GIN Sea	Lab. Sea	GIN Sea
Class 1	0.0	0.1	-0.2	-0.8
Class 2	-7.7	-0.2	-4.5	-3.2
Class 3	-7.5	-14.1	-4.3	-16.6
Class 4	-4.1	-23.2	-1.6	-21.4
Class 5	0.0	0.0	0.0	0.0

Table 2. Comparison of the diapycnal fluxes in the Labrador Sea and GIN Sea (including the regions south of the Denmark strait, see Figure 4.) between control run and the 1% CO2 runs

The volume temperature and salinity in the upper 1000 m show a warmer and more saline feature GIN Sea region and a cooling and freshening in the Labrador Sea region. As shown in Table 2, the changes in temperature do not stop the weakening of the deep convection in the Labrador Sea and the strengthening of the diapycnal mixing in the GIN Seas. Thus, the variation of the diapycnal flux in these seas is mainly controlled by the salinity variation in our model.

Recall in Figure 4, variations of the diapycnal fluxes in the northern North Atlantic are not uniform vertically. In Table 3, this region has been divided into 3 sub-regions. The sum of the downward diapycnal fluxes in class 3 is lower in the 1% CO2 run. The weakening mainly happens in sub-regions 1 and 2. The same flux in class 4 shows a stronger diapycnal mixing process in sub-region 2. This stronger diapycnal mixing process is related to the higher salinity there in the 1% CO2 run.

4. Summary

The North Atlantic MOC in a 300-year control run and four-member 1% CO2 transient ensemble runs are analyzed from a global coupled model, the PCM. The model results show a weakening of about 3 Sv of the North Atlantic MOC due to greenhouse gas induced warming relative to the control run. Locally, the changes of the diapycnal fluxes are not uniformly distributed horizontally. The weakening of deep convection is strongest in the Labrador Sea, mainly related to the freshening effect. On the other hand, the diapycnal fluxes are strengthened in the upper ocean in the GIN Seas. This strengthening is due to the increased salinity which is greater than the increased warming thus leading to increased density and more diapycnal flux there.

	Region 1	Region 2	Region 3
Class 1	0.07	-0.16	0.27
Class 2	0.24	-0.05	0.09
Class 3	2.03	0.82	0.14
Class 4	0.30	-4.55	0.09
Class 5	0.00	0.00	0.00

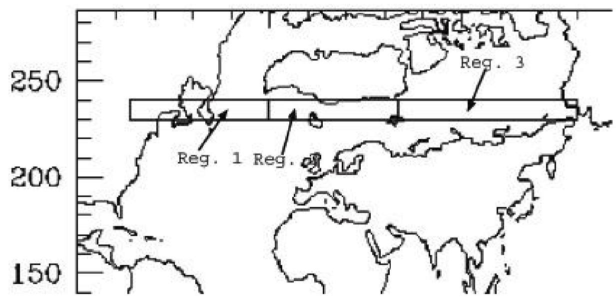

Table 3 Diapycnal flux differences in narrow regions of the Northern North Atlantic between the control run and the 1% CO2 run. Positive values indicate a strengthened downward water mass conversion.

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