Detecting thermohaline circulation changes from ocean properties in a coupled model

Aixue Hu,[1] Gerald A. Meehl,[1] and Weiqing Han[2]

Received 9 April 2004; accepted 9 June 2004; published 3 July 2004.

[1] Significant changes of the thermohaline circulation (THC) are likely to cause abrupt climate change. Here we intend to find a simple measure to detect changes in THC through examining several factors proposed to control the THC variations using a coupled climate model. These factors are equatorial-South Atlantic upper ocean temperature, Southern Ocean freshening, inter-basin sea surface salinity contrast, and meridional steric height gradient. Three experiments are analyzed – a present-day control run, a freshwater hosing run and a 1% CO_2 run. Results show that if freshwater flux is the primary cause, all examined factors can predict the THC changes. If both thermal and haline forcings are involved, only the Atlantic meridional steric height gradient gives a consistent measure of the THC variations. A new result presented here is that the inter-basin sea surface temperature contrast between North Atlantic and North Pacific is found to be an indicator of THC changes. *INDEX TERMS:* 1620 Global Change: Climate dynamics (3309); 4532 Oceanography: Physical: General circulation; 4215 Oceanography: General: Climate and interannual variability (3309); 1635 Global Change: Oceans (4203). **Citation:** Hu, A., G. A. Meehl, and W. Han (2004), Detecting thermohaline circulation changes from ocean properties in a coupled model, *Geophys. Res. Lett.*, *31*, L13204, doi:10.1029/2004GL020218.

1. Introduction

[2] Reorganization of the global thermohaline circulation (THC) is the only well proposed mechanism responsible for the past abrupt climate change that has been reviewed in depth by an expert group [*Committee on Abrupt Climate Change et al.*, 2002]. THC involves a global scale three-dimensional water mass transport along with the atmospheric heating and cooling processes in which the warmer upper water is converted into dense deep water in the northern North Atlantic and around the periphery of the Antarctic. These highly localized deep convective activities could be fully suppressed by an outburst of a freshwater flood, especially in the North Atlantic. The shutdown of the THC could induce significant cooling in the North Atlantic region, thus triggering a cooling event.

[3] Although variations of the global THC are suggested to be important for climate change, detection of the THC change remains a challenging task. The challenge arises from our inability to obtain continuous three-dimensional global data. Consequently, strengths of the THC are often inferred from the variation of important factors that can cause changes of the THC steady state. These factors include: the meridional gradient of the zonally averaged steric height [*Hughes and Weaver*, 1994; *Thorpe et al.*, 2001], inter-basin salinity contrast [*Stocker et al.*, 1992; *Seidov and Haupt*, 2003], the southern ocean freshening [*Saenko et al.*, 2003], and the warming of tropical Atlantic upper ocean [*Rühlemann et al.*, 1999]. Other factors that can also affect the THC are: diapycnal mixing [*Munk*, 1966; *Bryan*, 1987], Southern Ocean winds [*Toggweiler and Samuels*, 1998], and geothermal heating [*Huang*, 1999].

[4] The goal of this paper is to explore some of the factors controlling the steady-state THC that can effectively reflect the THC response to century-long transient climate change. To accomplish the goal, we perform a set of experiments using a coupled climate model – the National Center for Atmospheric Research's Community Climate System Model version 2.0 (NCAR's CCSM2.0). Results from this research will increase our ability to detect the changes in THC and project its future status.

2. Model and Experiments

[5] The model used in this study is the NCAR's CCSM2.0 [*Kiehl and Gent*, 2004; *Holland*, 2003]. Its atmospheric component is the NCAR Community Atmospheric Model (CAM2) at T42 resolution and 26 hybrid levels vertically. The ocean model is a version of the Parallel Ocean Program (POP) developed at Los Alamos National Lab with 1° horizontal resolution and enhanced meridional resolution (1/2°) in the equatorial tropics, and with 40 vertical levels. The sea ice model is the Community Sea Ice Model (CSIM4) with the Elastic-viscous-plastic dynamics, a subgrid-scale thickness distribution, and energy conserving thermodynamics. The land model is the Community Land Model (CLM).

[6] Three experiments are analyzed: the CCSM millennial present day control run (CON) from year 500 to 699, a freshwater hosing run (HOS), and a 1% CO_2 transient climate run (TRC). The last two experiments are specifically designed to investigate the relation of THC with changes of various oceanic variables for the past and future climate, and therefore to determine the prominent factors that most efficiently reflect the THC change.

[7] In the HOS experiment, a 0.1 Sv (1 Sv $\equiv 10^6$ $m^3 s^{-1}$) additional freshwater flux is uniformly distributed in the northern North Atlantic between 50° and 70°N, which mimics a freshwater pulse during the last deglaciation. This additional freshwater flux represents a 50% increase in net freshwater input into this region relative to the control experiment. In TRC experiment, the atmospheric CO_2

[1]Climate and Global Dynamics Division, National Center for Atmospheric Research, Boulder, Colorado, USA.
[2]Program in Atmospheric and Oceanic Sciences, University of Colorado, Boulder, Colorado, USA.

0094-8276/04/2004GL020218$05.00

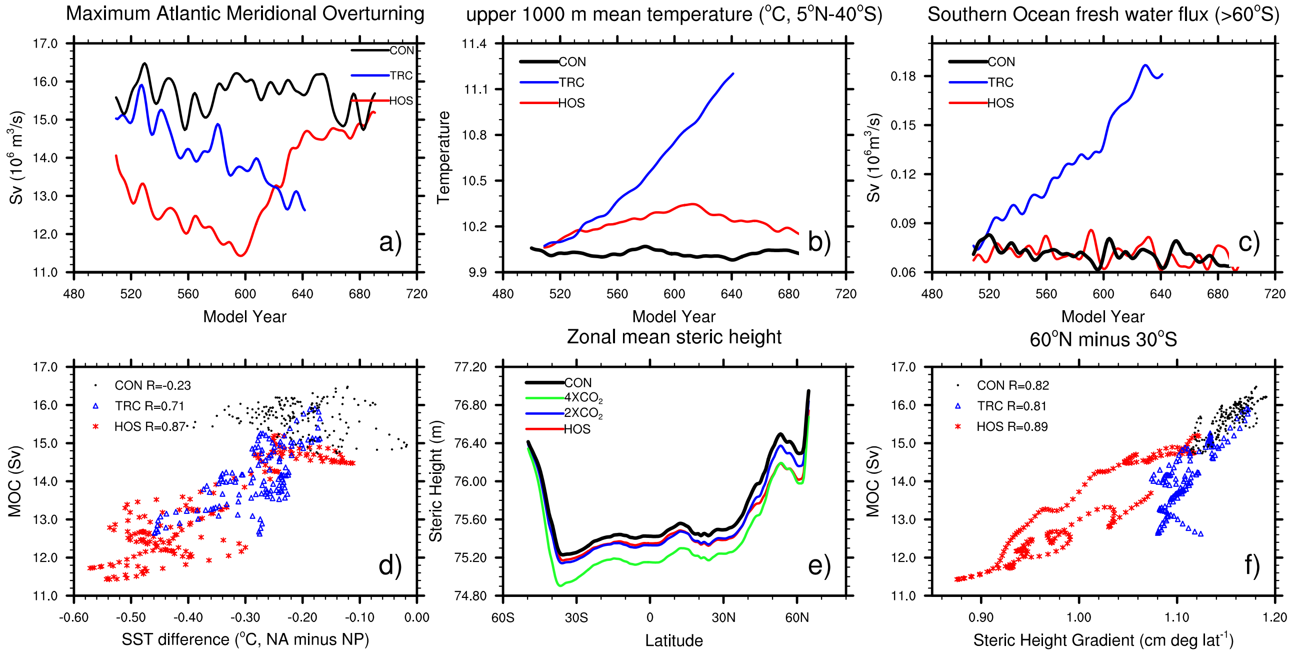

Figure 1. Time series: a) the Atlantic meridional overturning circulation (MOC), b) the mean upper 1000 meter ocean temperature of equatorial to South Atlantic (5°N–40°S), c) the freshwater flux into Southern Oceans (>60°S). Regression plots: d) the SST differences and MOC, and their correlation coefficients, f) the zonally averaged meridional steric height gradient and MOC, and their correlation coefficients. The SST differences are defined as the mean SST of North Atlantic minus the mean SST of North Pacific. North Atlantic (Pacific) is the region between 10 and 60°N. The steric height gradient is defined as the steric height difference between 60°N and 30°S. Panel e is the zonal mean steric height for CON (a 200-year average), HOS (averaged over years 580–599), 2 × CO_2 (averaged over years 560–579), and 4 × CO_2 (averaged over years 620–639). A 13-year low-pass filter is applied to all data used here, except those in panel e. The definition of the line patterns and markers for each panel are shown in the legends.

increases 1% per year compound. These two experiments are branched from the CON run at year 500. The TRC run is integrated for 150 years where CO_2 doubles at year 70 and quadruples at year 140. The HOS run is integrated with the additional freshwater flux for 100 years, and that flux is then switched off for another 100 years.

3. Results

3.1. Variation of MOC

[8] Since the meridional overturning circulation (MOC) of Atlantic is dominated by THC, hereafter, we use MOC to represent THC. The mean MOC strength, the maximum value of the Atlantic meridional stream function, in CON is 15.7 Sv, agreeing well with observed estimates [e.g., *McCartney and Talley*, 1984]. In comparison with that in CON, MOC weakens by 4.4 Sv in HOS averaged over years 580–599, and by 1.5 (2.9) Sv in TRC when CO_2 doubles (quadruples). Empirical orthogonal function analysis of the Atlantic meridional stream function shows that the production of the North Atlantic Deep Water and the Antarctic Bottom Water varies in an opposite sign in both CON and HOS runs (figure not shown)– when the production of the former is stronger, the latter is weaker, and vice versa, consistent with the bipolar seesaw theory of *Broecker et al.* [1999]. In TRC, the production of both waters varies in the same direction – as CO_2 increases, the formations of both are reduced.

[9] The meridional heat transport (MHT) north of the equator is proportional to the MOC strength – a weaker MOC transports less heat into North Atlantic (figure not shown). The MHT reduction at 24°N, a location with an abundance of repeated observations, is 0.17, 0.05 and 0.09 PW (1 PW $\equiv 10^{15}$ W) for HOS averaged over years 580–599, and for CO_2 doubling and quadrupling, respectively. In the South Atlantic, the northward MHT reduces by 0.15 PW in HOS. In TRC, the MHT reduction induced by the weaker MOC is compensated by the CO_2 induced warming effect. The resulting MHT is almost the same as in CON.

3.2. Upper 1 km Mean T in South Atlantic

[10] Comparing with Figure 1a, as MOC spins down, the upper 1000 meter equatorial-South Atlantic ocean temperature (T_{up}, Figure 1b) increases in HOS. An opposite situation occurs for a stronger MOC. These are consistent with *Rühlemann et al.* [1999] and *Stocker* [2002] in that a weaker MOC transports less heat from South Atlantic to North Atlantic. In TRC, the weaker MOC is also related to a higher T_{up}. However the increase in T_{up} is much larger than that in HOS. This higher T_{up} increase is primarily attributed to the CO_2 induced warming. In CON, the MOC induced T_{up} variation is small. Therefore, T_{up} could give us a good signal to detect the MOC changes for future climate only if the direct CO_2 contribution on T_{up} could be removed.

3.3. Southern Ocean Freshwater Forcing

[11] *Saenko et al.* [2003] proposed that the MOC strength in steady state is controlled by freshening in the Southern Oceans. As for the MOC, MHT and associated air-sea heat

Table 1. Mean SST, SHF, SSS and SFWF in NA and NP[a]

	Case	SST °C	SHF w/m^2	SSS ppt	SFWF Sv
NA	CON	20.0 ± .06	−6.5 ± .4	35.14 ± .03	−.325 ± .007
	HOS	19.4 ± .05	−4.3 ± .2	34.71 ± .02	−.292 ± .005
	2·CO_2	20.8 ± .03	−4.8 ± .2	35.13 ± .01	−.351 ± .003
	4·CO_2	21.7 ± .07	−3.7 ± .3	35.07 ± .01	−.376 ± .002
NP	CON	20.2 ± .05	−18.4 ± .3	34.35 ± .02	−.059 ± .004
	HOS	20.0 ± .03	−19.0 ± .1	34.37 ± .01	−.064 ± .002
	2·CO_2	21.1 ± .04	−17.4 ± .2	34.34 ± .01	−.071 ± .004
	4·CO_2	22.1 ± .03	−16.4 ± .1	34.26 ± .01	−.088 ± .003

[a]SST stands for sea surface temperature, SSS for sea surface salinity, SHF for net surface heat fluxes, SFWF for net surface freshwater fluxes. The sign convention for SHF and SFWF is that positive indicates ocean gaining heat or freshwater. NA (NP) stands for North Atlantic (Pacific) between 10 and 60°N. 2·CO_2 (4·CO_2) represents CO_2 doubling (quadrupling). Numbers in this table are averaged over NA or NP, and the same time average as in Figure 1e is applied here. The numbers after the ± signs are the standard deviations.

flux anomalies are all proportional to the southward moisture transport from subtropical to subpolar regions in the Southern Hemisphere. In our two transient climate runs (Figure 1c), the freshwater flux into the Southern Oceans (a proxy of the poleward water vapor transport) increases dramatically in TRC, and there is only a small change in HOS compared to CON. In comparison with Figure 1a, fresher subpolar Southern Oceans are related to a weaker MOC in TRC, in contrast with Saenko et al.'s conclusion. However, the Antarctic deep water production is reduced and the Southern Ocean upwelling also increases (figure not shown), agreeing with Saenko et al. To further investigate the discrepancy between our model and Saenko et al., two new sensitivity experiments are conducted: A. the freshwater flux in the CON run is replaced by that in the TRC run; B. the freshwater flux in the TRC run is replaced by that in the CON run. In the other word, the CO_2 induced thermal effect was removed in experiment A, and the CO_2 induced hydrological change was removed in experiment B. Results from these two experiments show that MOC strengthens in A relative to CON, and weakens further in B relative to TRC. Thus, the CO_2 induced warming effect on deep convective activity overcomes the Southern Ocean freshening in TRC, producing an overall weaker MOC in our model.

[12] Although the variations of the freshwater flux into the Southern Oceans is small in CON, the relation between MOC and this freshwater flux is consistent with Saenko et al. In HOS, the warming of the equatorial-South Atlantic induces a small increase in the freshwater flux into the Southern Oceans, but the much larger freshwater forcing in northern North Atlantic leads to a weakened MOC. Thus the effect of the freshwater flux changes in the Southern Oceans under transient climate on MOC could be offset by other factors.

3.4. MOC and Inter-basin SST, SSS Contrast

[13] *Seidov and Haupt* [2003] pointed out that the inter-basin sea surface salinity (SSS) contrast between North Atlantic and North Pacific (10 to 60°N) controls the strength of the steady-state MOC. A stronger salinity contrast causes a stronger MOC. In our transient climate runs, the North Atlantic freshening in HOS decreases this salinity contrast to 0.34 ppt from 0.79 ppt in CON (Table 1), resulting in a weaker MOC, agreeing well with Seidov and Haupt. In TRC, this salinity contrast has no change when CO_2 doubles, and increases to 0.81 ppt when CO_2 quadruples. This increase is related to a weaker surface freshening in the North Atlantic, and a stronger one in the North Pacific (Table 1). Since the net freshwater loss actually increases in both oceans, the overall freshening in these two oceans is due to CO_2 induced ice melting. The exported melt-ice water from the Arctic is the major contributor to the freshening in North Atlantic. The local melt-ice water dominates the Pacific freshening with minor contribution from Arctic ice export via Bering Strait driven by biased model wind. The MOC weakening as CO_2 increases in TRC (Figure 1a) implies that the overall warming and freshening in the North Atlantic, suppressing the deep convective activity there, dominates the increased inter-basin salinity contrast, producing a weaker MOC.

[14] Here we further studied the thermal effect on MOC from analyzing the sea surface temperature (SST) contrast between North Atlantic and North Pacific. In TRC, the surface ocean in both regions becomes warmer with a stronger warming in the North Pacific (Table 1). In HOS, both oceans become colder with a stronger cooling in the North Atlantic. The resulting surface temperature contrast increases in both runs. This increased temperature contrast is related to a weaker MOC (Figure 1d). In the second half of the HOS run, when MOC spins up due to switching off the additional freshwater flux, the temperature contrast decreases. Therefore, the sea surface temperature contrast can be used as a new indicator of the MOC changes.

[15] The different rate of surface temperature change in these two oceans is primarily related to changes in MOC. Ideally, a weakened MOC transports less heat into the North Atlantic and less cold deep water into the North Pacific, inducing a cooling effect in the former and a warming effect in the latter for a steady-state MOC. In our transient experiments, the atmospheric effect is also important. In TRC, the CO_2 induced warming suppresses the upwelling induced cooling effect in the North Pacific, resulting in an enhanced warming there. In the North Atlantic, the weaker warming is due to less meridional heat transported to there and the downward heat transport by convective activity. In HOS, we see a weaker cooling, instead of a warming in the North Pacific as MOC spins down. The net surface heat budget shows an increased oceanic heat loss (Table 1), indicating that colder air draws more heat from the North Pacific. This colder air is possibly initiated at North Atlantic since the cooling in both oceans shows no significant time lag. In CON, since the MOC is at a steady-state, the surface temperature variations in both regions are small and do not have a systematic trend. Therefore, this temperature contrast is very likely acting as an indicator, not a trigger of the MOC changes since it is a result of a systematic MOC change due to transient forcing.

3.5. MOC and Meridional Steric Height Gradient

[16] Steric height is defined as vertically integrated density. A linear relation between MOC strength and meridional steric height gradient has been found by many authors [e.g., *Hughes and Weaver*, 1994; *Thorpe et al.*, 2001]. Comparing with that in CON, the zonally averaged steric height, a density integration from surface to 2750 m, is reduced more significantly in mid-high latitude North

Atlantic due to freshwater forcing than that in mid-latitude South Atlantic due to less northward heat transport, resulting in a decreased meridional steric height gradient and a weaker MOC in HOS (Figure 1e). In TRC, the change in steric height is almost uniform meridionally, resulting in a smaller change in meridional steric height gradient, and a smaller MOC change than that in HOS. A regression plot (Figure 1f) further demonstrates that as meridional steric height gradient decreases, MOC weakens for all three cases. But the coefficient of the linear regression between MOC and meridional steric height gradient is 17 Sv per cm/deg-lat in HOS and CON, and that increases to 29 Sv in TRC. This higher regression coefficient in TRC is caused by an increased net evaporation in subtropical South Atlantic, resulting in a weaker steric height decrease there.

4. Discussion and Conclusions

[17] In contrast with *Hirschi et al.* [2003] who suggested the use of moorings across a longitude-depth section and zonal wind stress to obtain continuous MOC measurement, here we have tried to find an indirect measure to detect the THC changes using NCAR's CCSM2.0 since a direct and sustained measurement of the THC is not available from observations. Three experiments are analyzed - a CON run, a TRC run with 1% CO_2 increase per year, and a HOS run with an additional 0.1 Sv freshwater flux into the northern North Atlantic.

[18] Our results show that MOC response to transient climate changes is a result of several competing processes. In the model, MOC weakens when CO_2 increases, agreeing with many climate models [*Cubasch et al.*, 2001]. Surface warming and freshening in North Atlantic, suppressing deep convective activity there, are the primary cause for a weaker MOC. The increased poleward moisture transport in the Southern Hemisphere and the increased sea surface salinity contrast between North Atlantic and North Pacific tend to strengthen the MOC, but are overcome by changes in the North Atlantic, resulting in a moderate MOC reduction. This result is in line with several recent studies of the MOC response to climate change scenarios [*Dixon et al.*, 1999; *Mikolajewicz and Voss*, 2000; *Thorpe et al.*, 2001]. In the HOS run, the additional freshwater flux causes a freshening in the North Atlantic, a decreased surface salinity contrast, and a slight increase in poleward moisture transport in the Southern Hemisphere. The first two lead to a greater weakening of the MOC than in the TRC run.

[19] Factors examined here to detect the MOC changes are the equatorial-South Atlantic upper ocean temperature, Southern Ocean freshening, sea surface salinity contrast between North Atlantic and North Pacific, and meridional steric height gradient. Variations in these factors will change the steady state of MOC. Our study indicate that the effect of some of these factors on MOC strength is dominated by others under transient climate conditions. If the freshwater input into the ocean is the primary cause, all examined factors can be used as an indicator to detect MOC changes. If the thermal forcing is the primary cause, only the meridional steric height can be used. A new factor found in this research is the sea surface temperature contrast between North Atlantic and North Pacific which can be used to detect MOC changes for the two forced runs.

[20] Next, we will sort out more detailed underlying physics influencing the effectiveness of these factors on transient MOC changes. And the results in this study are also going to be compared with other coupled climate models.

[21] **Acknowledgments.** 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. The National Center for Atmospheric Research is sponsored by the National Science Foundation. Weiqing Han was supported by NSF grant OCE-0136836.

References

Broecker, W. S., S. Sutherland, and T.-H. Peng (1999), A possible 20th-century slowdown of Southern Ocean deep water formation, *Science*, *286*, 1132–1135.

Bryan, F. (1987), Parameter sensitivity of primitive equation ocean general circulation models, *J. Phys. Oceanogr.*, *17*, 970–985.

Committee on Abrupt Climate Change, Ocean Studies Board, Polar Research Board, Board on Atmospheric Sciences and Climate, Division on Earth and Life Studies, and National Research Council (2002), *Abrupt Climate Change: Inevitable Surprises*, Natl. Acad., Washington, D. C.

Cubasch, U., et al. (2001), Projections of future climate change, in *Climate Change 2001: The Scientific Basis-Contribution of Woring Group I to the Third Assessment Report of the Intergovernment Panel on CLimate Change*, edited by J. T. Houghton et al., pp. 525–582, Cambridge Univ. Press, New York.

Dixon, K. W., T. L. Delworth, M. J. Spelman, and R. J. Stouffer (1999), The influence of transient surface fluxes on North Atlantic overturing in a coupled GCM climate change experiment, *Geophys. Res. Lett.*, *26*, 2749–2752.

Hirschi, J., J. Baehr, J. Marotzke, J. Stark, S. Cunningham, and J.-O. Beismann (2003), A monitoring design for the Atlantic meridional overturning circulation, *Geophys. Res. Lett.*, *30*(7), 1413, doi:10.1029/2002GL016776.

Holland, M. M. (2003), The North Atlantic Oscillation–Arctic Oscillation in the CCSM2 and its influence on Arctic climate variability, *J. Clim.*, *16*, 2767–2781.

Huang, R. X. (1999), Mixing and energetics of the oceanic thermohaline circulation, *J. Phys. Oceangr.*, *29*, 727–746.

Hughes, T. M. C., and A. J. Weaver (1994), Multiple equilibria of an asymmetric two-basin ocean model, *J. Phys. Oceanogr.*, *24*, 619–637.

Kiehl, J. T., and P. R. Gent (2004), The community climate system model, version two, *J. Clim.*, in press.

McCartney, M. S., and L. D. Talley (1984), Warm-to-cold water conversion in the northern North Atlantic Ocean, *J. Phys. Oceangr.*, *14*, 922–935.

Mikolajewicz, U., and R. Voss (2000), The role of the individual air-sea flux components in CO_2-induced changes of the ocean's circulation and climate, *Clim. Dyn.*, *16*, 627–642.

Munk, W. H. (1966), Abyssal recipes, *Deep Sea Res.*, *13*, 707–730.

Rühlemann, C., S. Mulitza, P. J. Müller, G. Wefer, and R. Zahn (1999), Warming of the tropical Atlantic Ocean and slowdown of the thermohaline circulation during the last deglaciation, *Nature*, *402*, 511–514.

Saenko, O. A., A. J. Weaver, and A. Schmittner (2003), Atlantic deep circulation controlled by freshening in the Southern Ocean, *Geophys. Res. Lett.*, *30*(14), 1754, doi:10.1029/2003GL017681.

Seidov, D., and B. J. Haupt (2003), Freshwater teleconnection and ocean thermohaline circulation, *Geophys. Res. Lett.*, *30*(6), 1329, doi:10.1029/2002GL016564.

Stocker, T. F. (2002), North-south connections, *Science*, *297*, 1814–1815.

Stocker, T. F., D. G. Wright, and W. S. Broecker (1992), The influence of high-latitude surface forcing on the global thermohaline circulation, *Paleoceanography*, *7*, 529–541.

Thorpe, R. B., J. M. Gregory, T. C. Johns, R. A. Wood, and J. F. B. Mitchell (2001), Mechanisms determining the Atlantic thermohaline circulation response to greenhouse gas forcing in a non-flux-adjusted coupled climate model, *J. Clim.*, *14*, 3102–3116.

Toggweiler, J. R., and B. Samuels (1998), On the ocean's large-scale circulation near the limit of no vertical mixing, *J. Phys. Oceanogr.*, *28*, 1832–1852.

W. Han, Program in Atmospheric and Oceanic Sciences, University of Colorado, Duane-Physics Bldng., Rm D318, Campus Box 311, Boulder, CO 80309, USA.

A. Hu and G. A. Meehl, Climate and Global Dynamics Division, National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307, USA. (ahu@ucar.edu)