# On the Response of the Atlantic Ocean to Climatic Changes in High Latitudes: Sensitivity Studies With a Sigma Coordinate Ocean Model

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The oceanic adjustment processes during the transition from present climate conditions to atmospheric forcing conditions that are significantly different than present climate have been simulated with a sigma coordinate Atlantic Ocean model. An idealized "global warming" scenario has been simulated by imposing surface flux anomalies, representing warming and freshening in the high latitudes of the North Atlantic. The results show a relatively short oceanic adjustment process that takes place within a period of a few decades in which the thermohaline overturning circulation (THC) is adapted to a new state with a transport smaller by 4-5 Sv than that obtained with a control run without the surface anomalies. While the total change in the intensity of the THC in this ocean model is consistent with that obtained by coarse resolution coupled ocean-atmosphere climate models, this study shows a shorter adjustment time scale and more pronounced spatial changes than climate models do. Interesting results include a weakening and cooling of the Gulf Stream, reversal of the Labrador Sea circulation and considerable weakening in the deep western boundary current and in downslope near-bottom flows in high latitudes. Sensitivity experiments explore how the parameterization of horizontal diffusion in this sigma coordinate ocean model affects long-term climate simulations. These experiments show that horizontal diffusion in the model affects the transition process and local gyres, but the climate change in the THC and in the meridional heat flux are quite robust and insensitive to the way horizontal diffusion is parameterized in the model.

## 1. INTRODUCTION

The world oceans, and in particular the Atlantic Ocean where most deep-water formation occurs, play an important role in long-term climatic changes. For example, changes in freshwater fluxes associated with changes in ice coverage influence the thermohaline ocean circulation (THC), meridional heat transport and eventually global overturning circulation patterns through changes in the so-called "ocean conveyor belt" [*Broecker*, 1991]. Some numerical models suggest that the THC may even have multiple stable states [*Manabe and Stouffer*, 1995, 1997; *Marotzke and Willebrand*, 1991; *Delworth et al.*, 1993] associated with different climates. Paleoclimate records such as the Greenland ice cores indicate several abrupt changes in the North Atlantic climate, such as those occurred during the last cold Younger Dryas period, about 11,000 years BP. The last gla-

The Oceans and Rapid Climate Change: Past, Present, and Future
Geophysical Monograph 126

cial maximum (LGM), about 18,000 years BP, was also the subject of considerable research by observational and model studies [*Imbrie et al.*, 1992; *Prange et al.*, 1997; *Seidov and Haupt*, 1997, 1999]. Radiocarbon analyses suggest that during the LGM the THC was considerably weaker than today, with weaker and shallower North Atlantic Deep-Water (NADW) transport, but possibly larger Antarctic Bottom Water (AABW) transport.

While past climatic changes are documented in paleoclimate records, possible future climatic changes, such as those associated with increasing greenhouse gas concentrations, are based mostly on the prediction of coupled ocean-atmosphere climate models [e.g., *Manabe et al.*, 1991; *Manabe and Stouffer*, 1988, 1997; *Haywood et al.*, 1997]. Some of these climate models, when integrated for hundreds of years into the future, assuming different scenarios for greenhouse gas concentrations, indicate the possibility that the THC may collapse due to increasing surface temperatures and freshwater fluxes in high latitudes. In such a case, ocean circulation and consequently the world climate may be very different than current climate, and may resemble, in some aspects, past climates with weaker THC. In order to run global climate models for hundreds or even thousands of years, some simplifications must be done. For example, these models have low horizontal resolution (in the range of 3-4 degrees), they are highly diffusive, and they do not accurately represent many oceanic processes such as vertical mixing and the interaction between flow and bottom topography. Western boundary currents are only poorly resolved in most climate models. On the other hand, more realistic, high-resolution ocean models require too-much computations to allow long integrations of hundreds of years. However, if the ocean is adjusted to climatic changes within a period of time that is relatively short compared with the long time scale of the climate system, then a relatively short simulations with realistic models may be useful in order to study the adjustment process. This approach has been done before [e.g., *Seidov and Haupt*, 1997; 1999], but it has not been tried yet with a model of the typed used here. *Gerdes and Koberle* [1995] and *Döscher et al.* [1994] have shown that the anomalous temperature signal in high latitudes propagates by coastal waves along the western boundary and affects the Gulf Stream transport and the thermohaline circulation of the North Atlantic within a time scale of 10-20 years.

Since future climate prediction depends to large extent on model configurations and parameterizations, it is important to try variety of models with different levels of complexities. Most of the earliest climate ocean models were based on different versions of the z-level Bryan-Cox model [*Bryan* 1969] and its successor, the Modular Ocean Model, MOM, however, more recently isopycnal models become an attractive alternative for climate simulations [*Oberhuber*, 1993; *New and Bleck*, 1995; *Halliwell*, 1998]. A different class of models, such as the Princeton Ocean Model, POM [*Blumberg and Mellor*, 1987], originally developed for coastal and regional applications, are now being used for climate studies as well [*Ezer and Mellor*, 1994, 1997; *Häkkinen*, 1995, 1999; *Ezer et al.*, 1995; *Ezer*, 1999, 2001]. Sigma coordinate ocean models may be attractive for simulating processes involved in bottom boundary layers (BBL) and deep water formation along slopes; these processes are sometimes difficult to simulate with coarse resolution z-level models [*Gerdes*, 1993; *Beckmann and Döscher*, 1997; *Winton et al.*, 1998; *Pacanowski and Gnanadesikan*, 1998]. A recent comparison study between z-level and sigma coordinate ocean models using otherwise identical numerics, shows that sigma coordinate models are able to handle much lower horizontal diffusivities than z-level models do [*Mellor et al.*, 2001]. This model attribute may be important for simulations of interdecadal variabilities, which can be sensitive to the horizontal diffusivity of ocean models [*Huck et al.*, 1999]. On the other hand, a disadvantage of sigma coordinate models is that bottom topography smoothing and other remedies are needed in order to reduce pressure gradient and along-sigma diffusion errors near steep topographies [*Mellor et al.* 1994, 1998; *Ezer and Mellor*, 2000]. Of particular concern for climate studies using POM is the use of along-sigma diffusion. A common procedure to minimize diapycnal mixing along sloping bottoms and improve the representation of bottom boundary layers in the model is to subtract climatological temperature and salinity data from the diffusion terms [*Mellor and Blumberg*, 1985]. This technique may affect the model ability to simulate climatic changes, since it involves a weak relaxation of the model climatology to the observed (i.e., today's) climatology. Advanced advection schemes that do not require implicit diffusion [*Häkkinen*, 1999], or isopycnally oriented diffusion [*Barnier et al.*, 1998] are other alternative approaches. However, it is important also to evaluate the implications for climate simulations of the standard along-sigma diffusion as has been used by the majority of POM users and in previous climate simulations with this model [*Ezer and Mellor*, 1994, 1997; *Ezer et al.*, 1995; *Ezer*, 1999]. Here, a modification of the along-sigma diffusion scheme that eliminates the climatological relaxation in POM, is being tested.

The approach in this study is to look at various oceanic indicators, such as meridional heat fluxes, THC and temperature fields, and to evaluate how they change over time

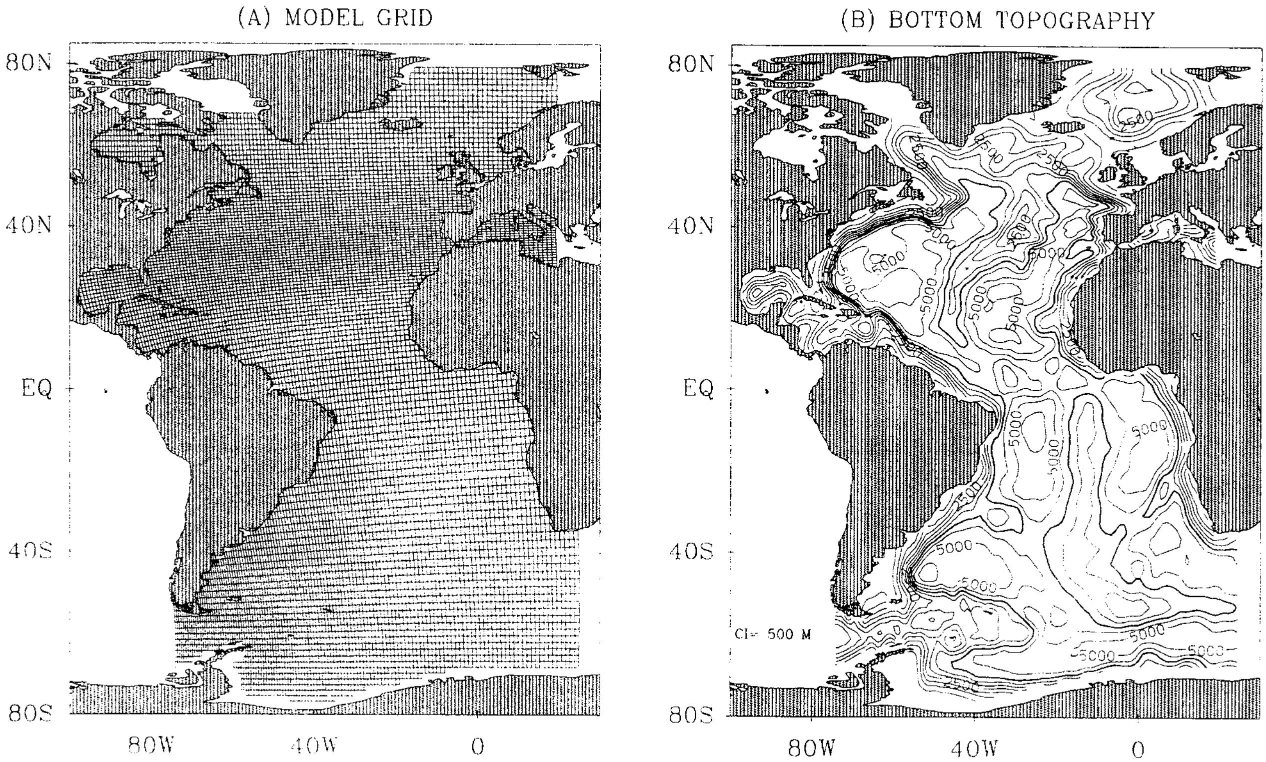


**Figure 1**. (a) Curvilinear orthogonal model grid and (b) bottom topography. Contour interval in (b) is 500m; a heavy line indicates the 4000m-isobath.

when atmospheric forcing conditions are significantly different than current conditions. These "climate change" simulations will be compared with control simulations representing today's climate. The experiments will be repeated with different model parameterizations, in order to evaluate the effect of model diffusion on current and future climate simulations. This study is a natural follow-up on the climatological simulations of *Ezer and Mellor* [1997] and the decadal variability simulations of *Ezer* [1999], using a similar model domain and configuration, but extending the calculations to resolve longer time scales.

Two main goals are included in this study. First, to study the processes involved in the adjustment of the ocean to extreme climatic changes in high latitudes, and second, to test the sensitivity of climate simulations with a sigma coordinate ocean model to diffusion parameterization.

The paper is organized as follows: First, in section 2, the ocean model and the experiments are described, then in sections 3, the results are discussed. Discussion and conclusions are offered in section 4.

## 2. THE OCEAN MODEL AND THE EXPERIMENTS

The Princeton Ocean Model, POM [*Blumberg and Mellor*, 1987] is a bottom-following sigma coordinate ocean model, which employs the Mellor-Yamada turbulence scheme [*Mellor and Yamada*, 1982] for vertical mixing parameterization. The model configuration, lateral boundary conditions and the Atlantic Ocean grid, between 80°S and 80°N (Figure 1), are similar to those used by *Ezer and Mellor* [1997] and *Ezer* [1999]. The horizontal curvilinear orthogonal grid has variable resolution with a grid size of 50-100 km; the vertical grid has 16 sigma layers with a higher resolution near the surface. The lateral boundary conditions include three open boundaries, two in the south for the inflow and outflow of the Antarctic Circumpolar Current, and one in the north for the connection with the Arctic Ocean. The distribution of vertically integrated transports along each open boundary is constant in time and specified according to estimates based on observations and models, but internal velocities at each level are controlled by radiation conditions [see *Ezer and Mellor*, 1997 for details]. The possible effect of the lateral boundary conditions on the calculations will be discussed later. There are three main differences between this study and the previous studies with this model. First, surface heat and salt fluxes are used as boundary conditions instead of surface temperature and salinity, second, the along-sigma diffusion has been modified, and third, longer simulations with forcing representing climate changes are performed.

The horizontal mixing coefficients for momentum (viscosity) and tracers (diffusivity) are calculated by a Smagorinsky-type formulation [*Smagorinsky et al.*, 1965] such that

$$(A_M, A_H)= (C_{vis}, C_{dif})\Delta x \Delta y \left[ \left(\frac{\partial u}{\partial x}\right)^2 + \frac{1}{2}\left(\frac{\partial v}{\partial x}+\frac{\partial u}{\partial y}\right)^2 + \left(\frac{\partial v}{\partial y}\right)^2 \right]^{1/2} \qquad (1)$$

where u and v are the horizontal velocity components in the x and y direction, respectively (x and y are along the curvilinear model grid lines). Therefore, horizontal viscosity and diffusion are reduced with decreasing grid size and velocity gradients. Sensitivity experiments using different values of the coefficients $C_{vis}$ and $C_{dif}$, have been the subject of the study by *Ezer and Mellor* [2000]; here these coefficients are set to a constant value of 0.2. The formulation of horizontal diffusion along sigma layers in POM, following *Mellor and Blumberg* [1985], has shown some advantages in maintaining bottom boundary layer structures; this approach, however, may cause undesired diapycnal mixing over steep bottom slopes, so steps must be taken to minimize this numerical problem as discuss below. The changes in temperature due to the along-sigma diffusion in the model can be expressed as

$$\frac{\partial T}{\partial t}=\frac{\partial}{\partial x}\left(A_H \frac{\partial (T-T_{clim})}{\partial x}\right) + \frac{\partial}{\partial y}\left(A_H \frac{\partial (T-T_{clim})}{\partial y}\right) \qquad (2)$$

where $T_{clim}(x,y,z)$ represents climatological data. A similar formulation applies also to salinity. The subtraction of climatological data removes most of the undesired diapycnal mixing, since only (small) anomalies are involved in the horizontal diffusion terms, and in fact, the results resemble calculations with zero external diffusion [*Ezer and Mellor*, 2000]. This formulation may affect climate simulations since the use of say, *Levitus* [1982] climatology for $T_{clim}$, adds a relaxation term to the model temperature, which may prevent the model from developing a climate state that is significantly different than the imposed $T_{clim}$. This hypothesis will be tested here by replacing the constant Levitus $T_{clim}$ with a time varying model climatology, thus eliminating any relaxation of the model fields to the observed climate.

The surface forcing includes the monthly climatological surface wind stress of the Comprehensive Ocean-Atmosphere Data Set (COADS) analyzed by *da Silva et al.* [1994]. Surface heat flux ($Q_T$) and surface salinity flux ($Q_S$) from the atmosphere to the ocean are calculated according to

$$Q_T = Q_c + \gamma_T \left(T_m^0 - T_c^0 - \Delta T\right) \qquad (3a)$$

$$Q_S = (E-P)_c S_c^0 + \gamma_S \left(S_m^0 - S_c^0 - \Delta S\right). \qquad (3b)$$

Subscripts "c" and "m" represent fields derived from the COADS monthly climatology and from the model, respectively. $T^0$ and $S^0$ are the surface temperature and salinity, respectively, and E-P is the observed evaporation minus precipitation. $\Delta T$ and $\Delta S$ are imposed departures of surface temperature and salinity from the observed climatologies; $\gamma_T$ and $\gamma_S$ are coupling coefficients. The coupling coefficient $\gamma_T$ has been chosen as -50 W m$^{-2}$ K$^{-1}$, similar to the value of $\partial Q/\partial T$ calculated from COADS observations. The chosen coefficient $\gamma_S = 5\times10^{-6}$ m s$^{-1}$ is based on sensitivity studies when simulating the current climate (i.e., with $\Delta T=\Delta S=0$). While surface flux boundary conditions as in (3) are common in ocean modeling, they do have considerable deficiencies as they do not represent realistic atmospheric feedback processes, which are better represented in coupled ocean-atmosphere models. In the previous studies of *Ezer and Mellor* [1997] and *Ezer* [1999] surface temperature and surface salinity were used as boundary conditions (the equivalence of setting the coupling coefficients in (3) to infinity). A boundary condition with an imposed surface property may constrain model variability more than a flux boundary condition does [*Greatbatch et al.*, 1995; *Seager et al.*, 1995; *Huck et al.*, 1999], so there is some advantage in the approach taken here.

To test the model response to surface forcing and diffusion parameterization, four simulations are performed, each one executed for 60 years:

(a) Experiment 1 is a control experiment with standard diffusion (labeled "CS"). Surface boundary conditions in (3) include monthly climatological values obtained from COADS and zero surface temperature and salinity anomalies ($\Delta T=\Delta S=0$); $T_{clim}$ (and $S_{clim}$) in (2) were obtained from the *Levitus* [1982] climatology (interpolated into the model grid).

(b) Experiment 2 is another control experiment but with modified diffusion (labeled "CM"), i.e., it is similar to CS, but $T_{clim}$ and $S_{clim}$ were obtained from the model climatology of the previous 5 years, eliminating any relaxation to the observed climatology. The actual horizontal diffusion in this experiment (which varies temporally and spatially) is generally smaller than that used in CS and it is also reduces with time as the model approaches an equilibrium state.

(c) Experiment 3 represents a warm-climate experiment with standard diffusion (labeled "WS"). In (3) $\Delta T$ and $\Delta S$ are functions of latitude, with values of zero for latitudes south of 40°N and linearly varying values between 40°N and 80°N such that at 80°N $\Delta T$=3°C and $\Delta S$=-1psu. These warming and refreshing of the surface of the ocean in high latitudes resemble the climate changes under increase greenhouse gas concentrations as predicted by coupled ocean-atmosphere models [*Haywood et al.*, 1997; *Manabe et al.*, 1991; *Manabe and Stouffer*, 1995, 1997].

(d) Experiment 4 represents another warm-climate experiment, but with a modified diffusion (labeled "WM"), i.e., forcing is similar to that in WS, but diffusion is as in CM.

This set of experiments can isolate the effects of high latitude surface anomalies and model diffusion parameterization, since they show how horizontal diffusion affects both, today and future climate simulations. By comparing CS with WS and CM with WM any model drift other than that associated with the surface conditions is eliminated. In order to isolate buoyant-driven from wind-driven effects, the same COADS monthly climatological wind stresses are used in all the experiments. In reality, or in coupled ocean-atmosphere models, the wind is affected by the climatic changes of surface temperature [*Kushnir*, 1994; *Latif*, 1998]. The possible consequence of neglecting wind effects will be discussed later. The experiments start from the end of the 30-years integration of *Ezer and Mellor* [1997]. The control experiments will thus include the adjustment of the ocean model from surface boundary conditions of imposed temperature and salinity to surface flux boundary conditions, an interesting test by itself.

## 3. MODEL RESULTS

### *3.1 The Transient Adjustment Process*

We first look at the transient process and how the ocean model adjusts itself to a new climatic state under the imposed surface anomalies. Of particular interest is the question whether or not the relaxation of the model to climatology in (2) will prevent experiment WS from reaching a new oceanic state. Figure 2 shows the zonally averaged temperature difference between the two calculations with standard diffusion formulation (i.e., WS minus CS). The plume of relatively warmer water under the surface anomaly conditions propagated southward to about 40°N and downward, reaching almost a steady state after about 30 years. Around the tropics, and seemingly disconnected from the high latitude warming, subsurface warming grows in amplitude with time during the entire 60-years integration, we will explain this later. Despite the surface warming conditions, two regions show cooling trends, the upper layers around 40°N and the near-bottom deep layers; the former and the later relate to changes in the Gulf Stream and in the Antarctic Bottom Water cell, respectively, as discussed later. The spatial structure in the thermal field seen in Figure 2 is quite different than that obtained by course resolution climate models [e.g., *Manabe et al.* 1991], where warming of the upper layers in the entire North Atlantic is more homogeneous. The downslope penetration of the anomaly due to the near bottom advection and turbulent mixing is more pronounced in the sigma model than it is in the climate (z-level) models.

The temperature change at 500m in Figure 3 (which focuses only on the North Atlantic portion of the model domain) shows that the high latitude warm anomaly propagated along the western boundary and reached the tropics within a period of about 20 years. Our results are quite similar to the response of the Atlantic Ocean to high latitude anomalies, as described by previous studies. For example, *Gerdes and Koberle* [1995] describe adjustment process that involves propagation of anomalies by coastal topographic waves along the North America continent and advective response in the western boundary undercurrent when cooling of surface water near Iceland was imposed; the adjustment time scale was about 10 years. *Döscher et al.* [1994] describe an adjustment process of the Atlantic Ocean to high latitude anomalies that takes place within about 20 years. Despite the zonally averaged imposed surface anomalies, considerable zonal spatial variations in the subsurface temperature changes are apparent in Figure 3. Regions with relatively larger warming trend include the Greenland Sea, the Labrador Sea and the Gulf of Mexico/Caribbean Sea, where warm pools of water accumulate around local recirculation gyres. Cooling occurs downstream of the Gulf Stream and the North Atlantic currents; this cooling will be explained later in relation with weakening of the Gulf Stream and the reduction in the amount of warm subtropical water masses advected by the Gulf Stream into higher latitudes.

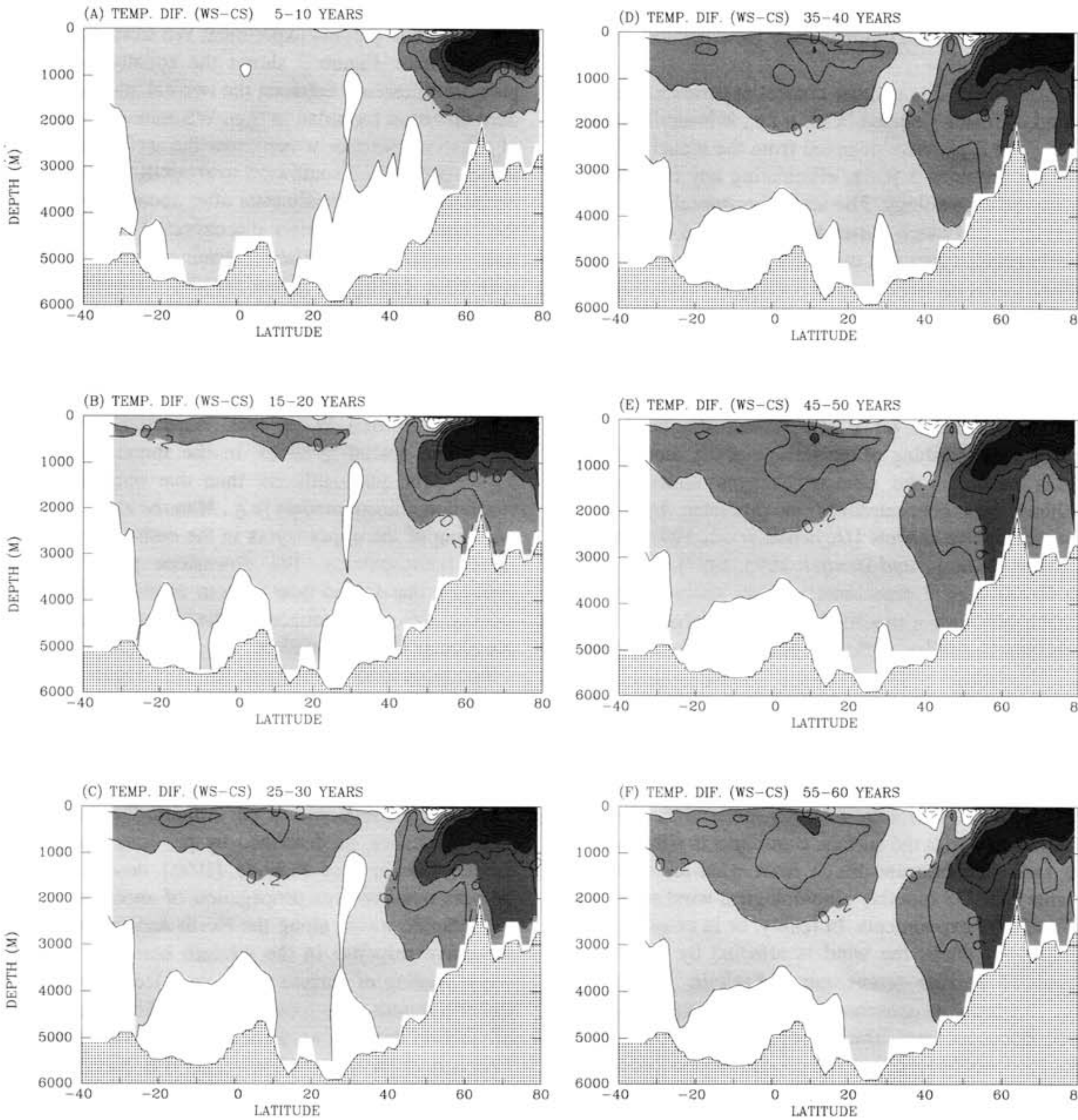


**Figure 2.** Zonally and 5-year averaged temperature differences (experiment WS minus CS) for years: (a) 5-10, (b) 15-20, (c) 25-30, (d) 35-40, (e) 45-50 and (f) 55-60. Contour interval is 0.2°C; white regions represent negative values and shaded regions represent positive values.

A useful diagnostic parameter in analysis of climate simulations is the thermohaline circulation (THC) index, defined here as the maximum value of the meridional stream function between 50°N and 70°N and expressed in Sv (1 Sverdrup = $10^6$ $m^3$ $s^{-1}$). A comparison of the THC index in high latitudes between all four experiments is shown in Figure 4. The experiments with modified (and very low) diffusivities (Figure 4b) show more transient changes during the first stages of the adjustment process and larger variations in the seasonal cycle. These experiments seem to be close to an equilibrium state during the last 30 years of the integration (longer simulations are probably needed to

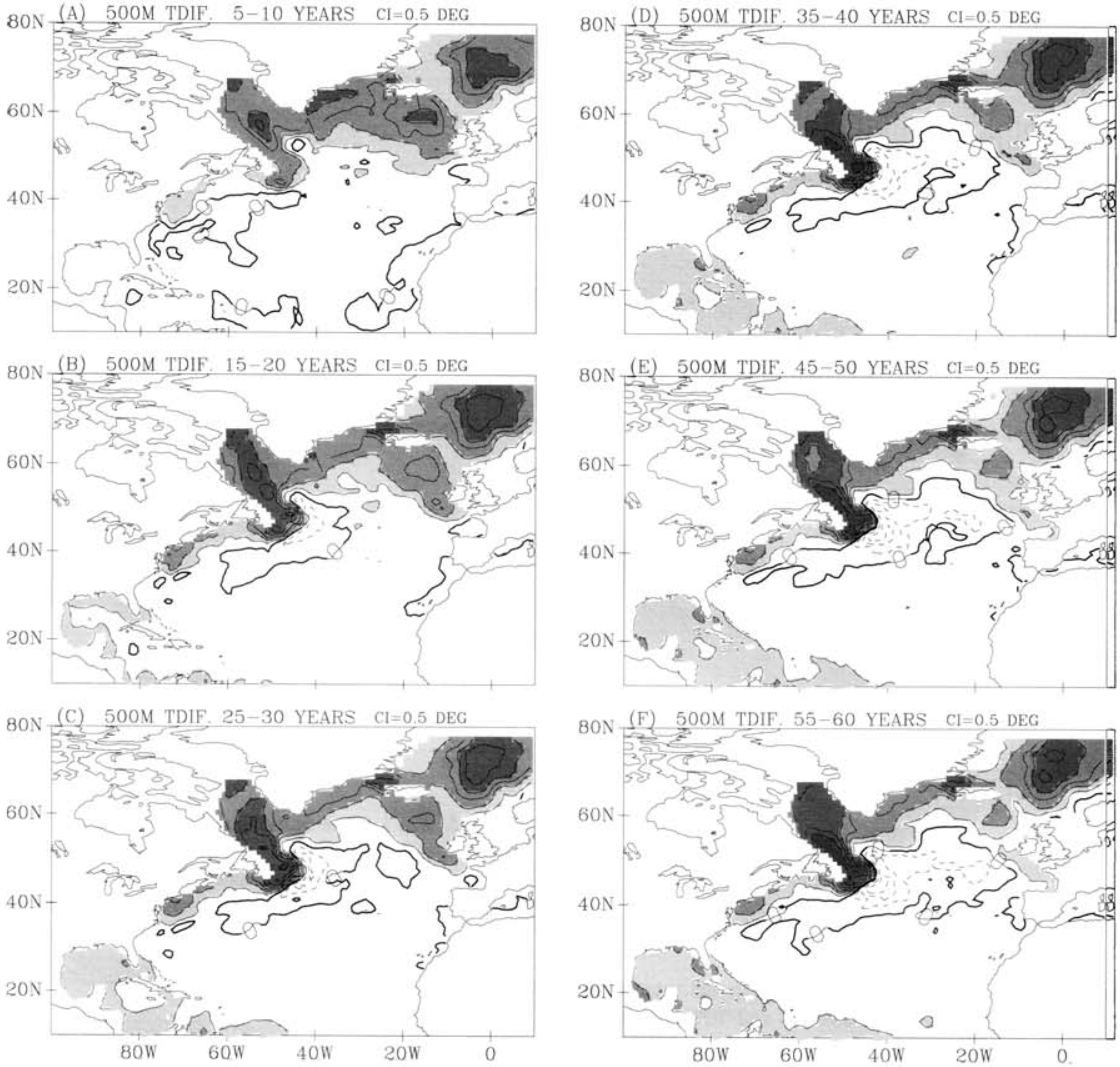


**Figure 3**. 5-year averaged temperature differences at 500m depth (experiment WS minus CS) for years: (a) 5-10, (b) 15-20, (c) 25-30, (d) 35-40, (e) 45-50 and (f) 55-60. Contour interval is 0.5°C; heavy lines indicate the zero contour, dashed lines negative values and shaded regions represent positive values above 0.5°C.

show if such a state exists at all). While one anticipates that the two control simulations (CS and CM) be very similar to each other and the two warm-climate experiments (WS and WM) be very different from each other due to the relaxation of CS and WS to today's climatology, surprisingly, the results are quite different than expected. In fact, both warm-climate experiments reached almost identical THC intensity after 60 years of integration, pointing to the conclusion that the relaxation effect in (2) is small enough so that it does not prevent the model from reaching a THC state that is different than current climate. Moreover, the results indicate that the changes in the intensity of the THC are primarily driven by surface fluxes rather than by diffusion. The increase in THC intensity in the two control runs during the first 10 years represents the adjustment from surface temperature and salinity boundary conditions, as used in *Ezer and Mellor* [1997], to flux surface boundary condition; similar results obtained by the box model experiments of *Huck et al.* [1999]. The THC index in the control run with modified diffusion reached

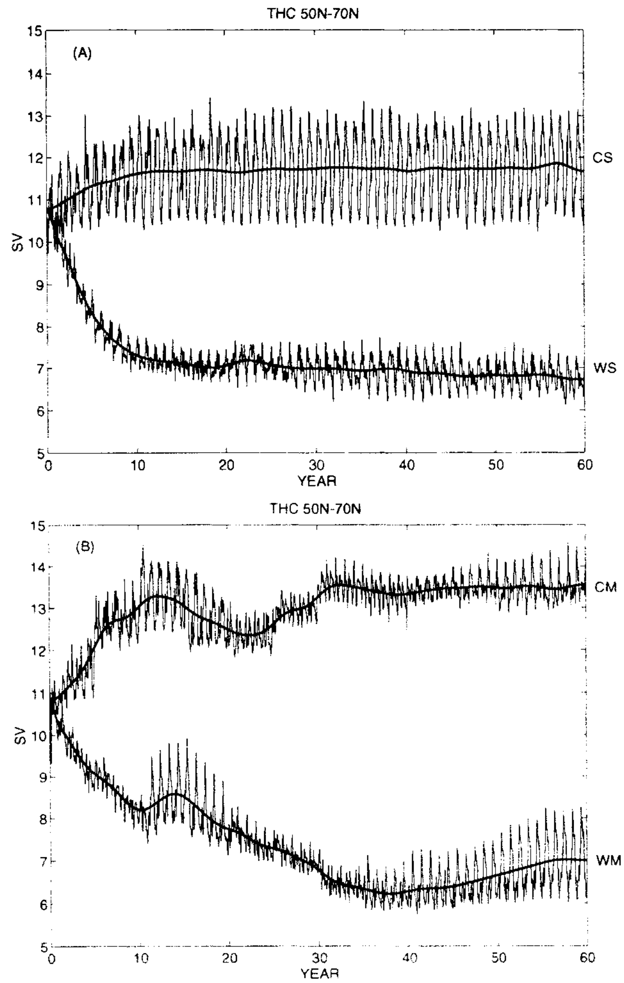


**Figure 4.** The THC index defined as the maximum value of the meridional stream function between 50°N and 70°N, for (a) experiments CS and WS and (b) experiments CM and WM. The heavy line has been low-pass filtered to remove variations shorter than 2 years.

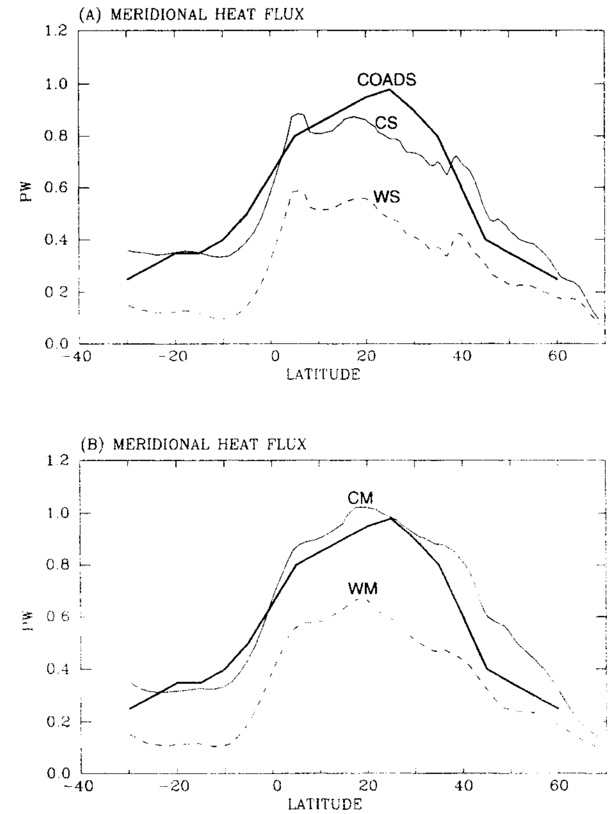


**Figure 5.** The average meridional heat flux calculated from the last 5 years of the integrations for experiments (a) CS and WS, thin solid and dashed lines, respectively, and (b) CM and WM, thin solid and dashed lines, respectively. Also shown, the heavy solid line in each panel, is the calculations based on the COADS climatology.

higher values at the end of the integration than the standard run did. When $T_{clim}$ in (2) is obtained from the model itself in run CM, the actual horizontal diffusion is smaller than that in CS, resulting in more intense circulation patterns, larger changes in convection and subduction, and a larger interannual variability.

### *3.2. Meridional Heat Flux and Overturning Circulation*

The poleward heat transport by the ocean plays an important role in maintaining the earth climate and may change as climate changes. For example, diagnostic calculations of past changes suggest changes in North Atlantic meridional heat flux of about 0.2 PW (1 PW= $10^{15}$W) over a period of 15 years, compared with a maximum value of about 1 PW. Considerable latitudinal variations due to changes in circulation were also obtained by these calculations [*Greatbatch and Xu*, 1993; *Ezer et al.*, 1995]. The meridional heat flux (MHF) as integrated zonally and vertically over the Atlantic basin is calculated for the four experiments and compared with the value obtained from COADS (Figure 5). Despite the fact that only latitudes north of 40°N were directly affected by the surface anomalies, considerable reduction in the MHF, up to 0.4 PW, are shown in both WS and WM experiments for all latitudes. The different diffusion parameterization in CS and WS compared with CM and WM experiments seems to affect the latitudinal variations in the MHF but had little effect on the climatic changes (i.e., the difference between the

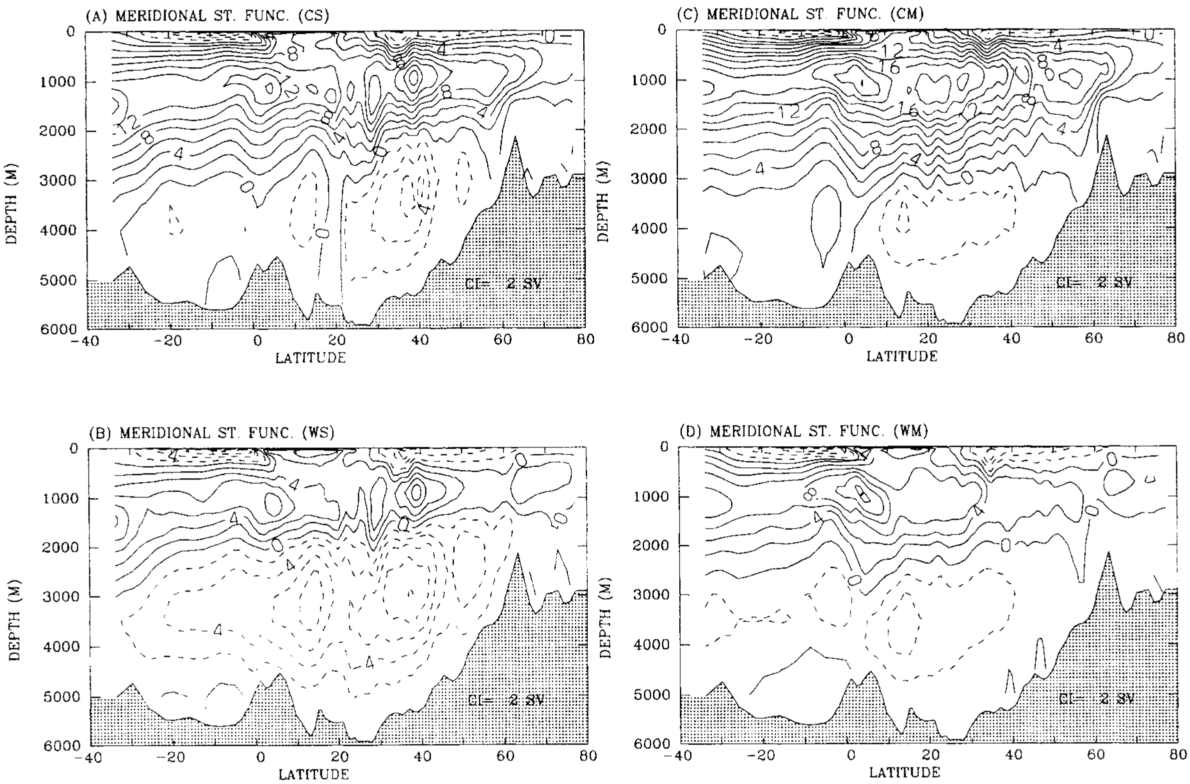


**Figure 6**. The zonally integrated meridional stream function calculated from the last five years of each experiment: (a) CS, (b) WS, (c) CM and (d) WM. Contour interval is 2 Sv and negative contours are indicated by dashed lines.

dashed and solid-thin lines in Figure 5a and 5b are similar). The experiments with modified diffusion (Figure 5b) have MHF that more closer resembles the observed values than the experiments with a standard diffusion (Figure 5a) showing that the modified diffusion formulation may be advantageous over the standard formulation.

The MHF is closely related to the THC, which is shown in terms of the meridional stream function in Figure 6. Both, the standard and the modified diffusion calculations, show considerable climatic weakening in the intensity of the THC (Figures 6b and 6d). These changes include a reduced North Atlantic Deep Waters transport between 1000 and 2500m depth and an almost complete shutoff in the transport of deep waters across the sills between the Greenland Sea and the North Atlantic Ocean. On the other hand, near bottom circulation, associated with northward transport of Antarctic Bottom Water (see also Figure 2), increased. The suggested climatic changes here due to climate warming resemble possible past changes such as the climate during the Last Glacial Maximum period as obtained from paleoclimate records [*Broecker*, 1991; *Imbrie et al.*, 1992] and models [*Seidov and Haupt*, 1999]. The results are generally consistent with previous modeling studies [*Döscher et al.*, 1994; *Gerdes and Koberle*, 1995], who show that cooling/warming in high latitudes causes intensifying/weakening of the THC and increasing/decreasing in MHF. It is also interesting to note some changes in the upper ocean transport and in particular the deepening of the shallow circulation cells (indicated by dashed lines in Figure 6) in both hemispheres. This change implies that while the majority of the poleward subsurface transport around 500m depth subducts at middle latitudes into deeper layers in the control experiments, in the warm-climate experiments larger portion of this flow joins the near surface equatorward transport instead. Decadal-scale changes in subduction and the ventilated thermocline [*Luyten et al.*, 1983] may thus affect the upper ocean connection between subtropical and tropical regions [*Liu et al.*, 1994]. Note also the development of a shallow circulation cell around 10°N, near the latitude of the local maximum warming seen in Figure 2. The development of this cell implies that under warm-climate conditions, larger

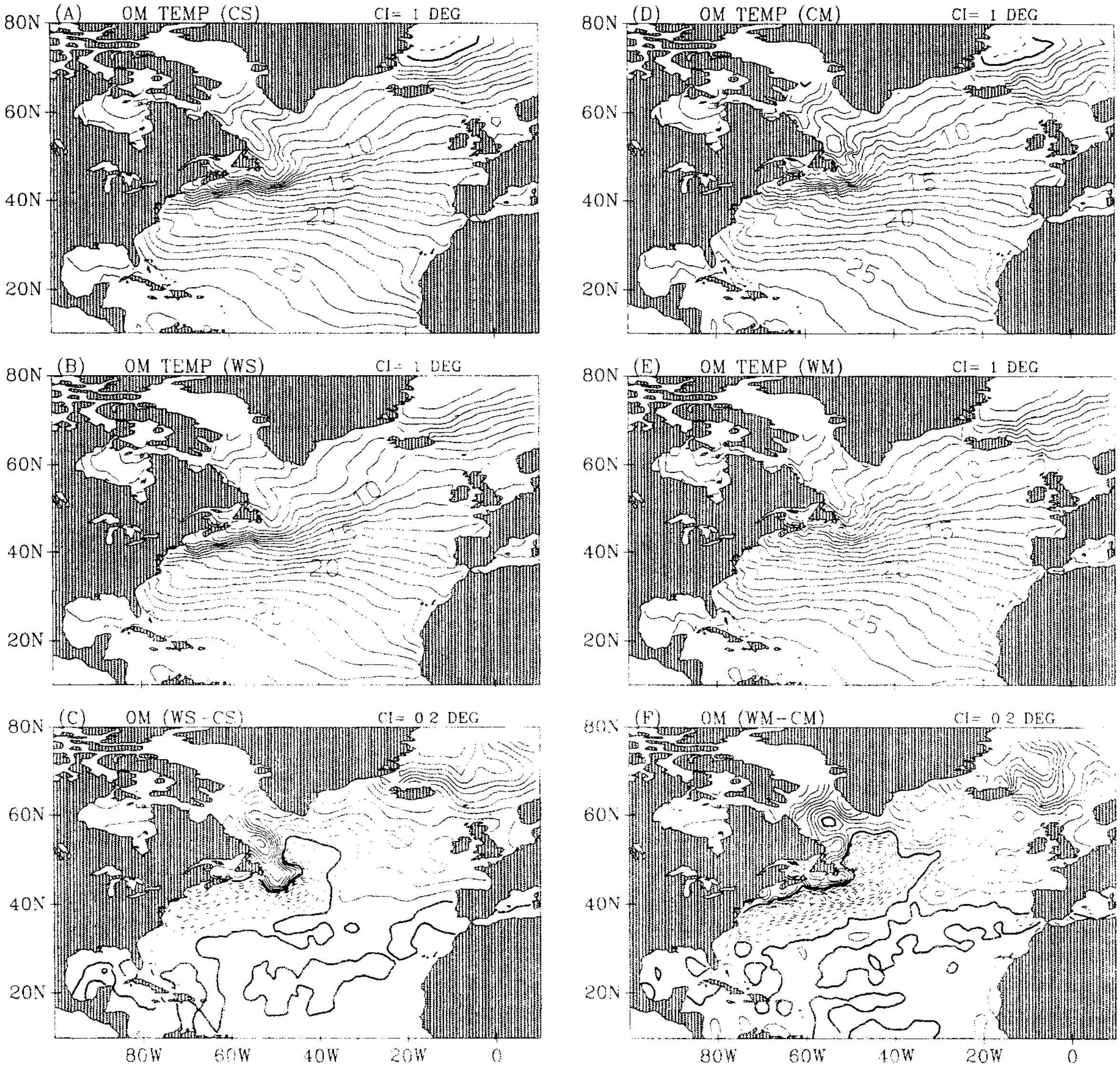


**Figure 7**. Sea surface temperature and the difference between the warm climate and the control experiments, (a) CS, (b) WS, (c) WS-CS, (d) CM, (e) WM and (f) WM-CM, calculated from the last 5 years of the integrations. Contour interval is 1°C, except (c) and (f) were it is 0.2°C; dashed lines represent negative values.

portion of the northward surface transport in the tropical North Atlantic recirculates back equatorward below the surface instead of transporting tropical waters into higher latitudes.

### *3.3 Spatial Changes*

The imposed surface anomalies in (3), $\Delta T$ and $\Delta S$, are zonally uniform, thus any spatial departure from a zonally averaged change (e.g., Figures 2 and 3) must be the result of changes in ocean circulation. In fact, the use of surface fluxes instead of surface temperature and salinity [as was the case in *Ezer and Mellor*, 1997 and *Ezer*, 1999] allows the model to develop its own surface temperature change (Figure 7) which is very different than the imposed zonal anomaly. Note that surface temperature gradients across the Gulf Stream are reduced in WM (and to lesser extent in WS), indicating a weakening of the geostrophic component of the Gulf Stream. The weaker Gulf Stream reduces the northward transport of warm subtropical waters, resulting

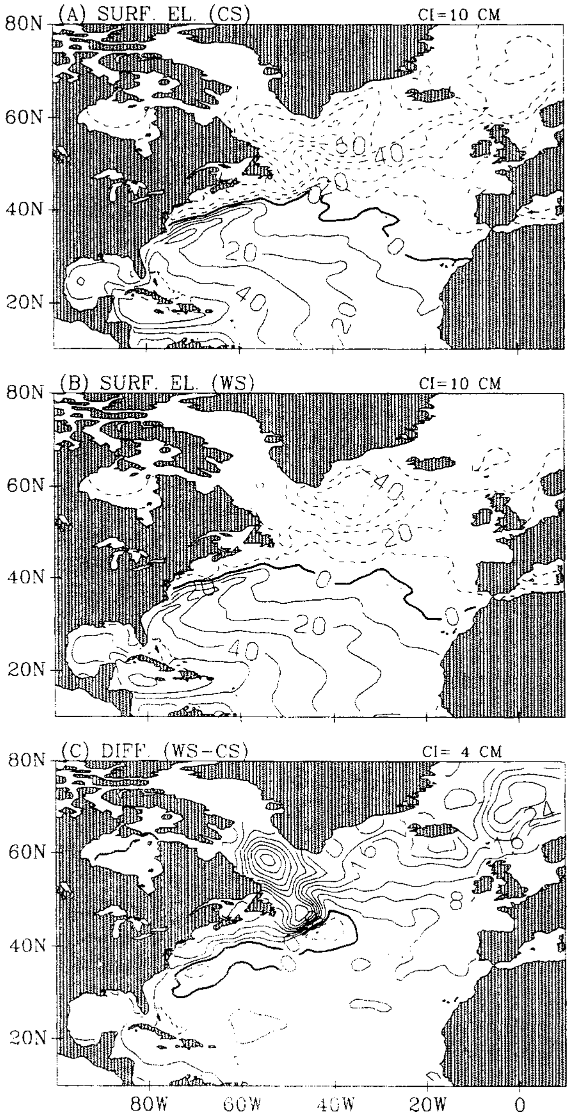


**Figure 8.** Same as Figure 7, but for the sea surface elevation. Contour interval is 10 cm, except (c) and (f) were it is 4 cm.

in relative cooling along its path. This upper ocean cooling occurs despite the fact that the imposed surface heat flux anomaly implies an additional surface warming north of 40°N in the warm-climate calculations. Figure 8 shows the corresponding surface elevation (and thus surface geostrophic velocity) at the end of the integrations. In the warm-climate experiments, WS and WM, the circulation of the subpolar gyre and the Labrador Sea weakened considerably, and the gradients of surface elevation across the Gulf Stream smaller. Although quantitative spatial differences exists between the experiments with different diffusions (comparing the right and left panels), qualitatively the climatic changes show similar trends. The process of spatial climatic changes can be explained as follows. The warmer than normal surface water masses in high latitudes are entrained with subsurface layers and mixed downward (Figure 2) while the signal also propagates along the western American coast (Figure 3 and Figure 7) around Newfoundland and into the northern recirculation gyre north of the Gulf Stream. The warming of the cold side of the Gulf Stream results in smaller thermal gradients across the Stream and in its intensity. As

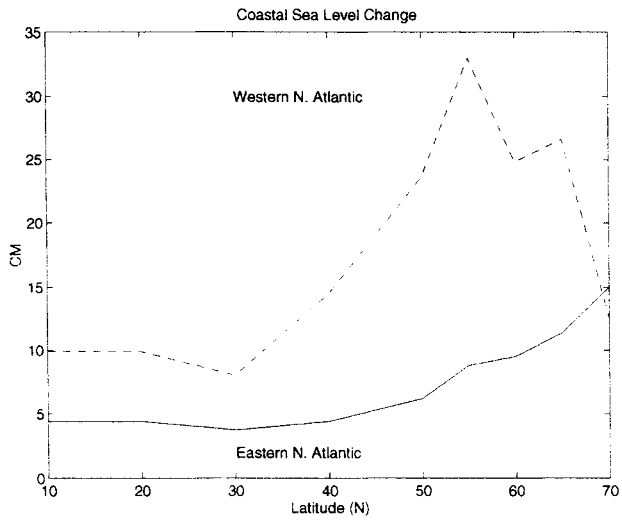


**Figure 9**. The coastal sea level change (WS-CS) along the eastern coast (solid line) and the western coast (dashed line) of the North Atlantic.

a result, the northeastward transport of warm waters by the Gulf Stream decreases, which further decreases the thermal gradients, creating a positive feedback mechanism. The process does not continue forever. A semi-equilibrium is reached due to the negative feedback of the surface heat flux (i.e., if surface model properties depart from the imposed values, the last term in 3a and 3b produces a flux correction). It should be noted that the freshening effect due to the imposed salinity anomaly works in the same manner (not shown) as the temperature does. However, the spatial change in the salinity field is not as pronounced as in the temperature field, because the imposed heat flux anomaly is at the coldest regions while the maximum imposed salinity anomaly is at regions that are already relatively fresh.

The spatial changes described above may have important implications for the detection of climatic changes from coastal sea level observations. Sea level records reflect the thermal expansion of the global ocean, but local changes in circulation due to the oceanic adjustment to spatial thermal changes can be significant [*Mellor and Ezer*, 1995]. Climatic changes detected at coastal sea level stations vary significantly from one station to another even when only short distance separates between them, however, ocean models can simulate these variations quite well, from climatic changes in the open ocean [*Ezer et al.*, 1995]. The coastal sea level change (WS-CS in Figure 8) shows for example, different responses for the western and for the eastern coasts (Figure 9). The sea level rise along the eastern coast resembles that expected from the imposed zonal surface temperature anomaly, but the sea level rise along the western boundary is much larger and spatially varying due to the ocean circulation changes described before. Therefore, these calculations indicate the difficulty in predicting local sea level change without taking into account changes in ocean dynamics.

### *3.4 Changes in Near-Bottom Flows and Deep Water Formation*

Figure 10 shows the vertically integrated stream function and the climatic changes. Note first the differences between the two control runs (Figure 10a and Figure 10d), in particular, the differences in the recirculation gyres in the Gulf of Mexico and south of the Gulf Stream. The relaxation of the diffusion terms to the observed climatology in CS and WS may have caused inconsistencies between the model and the observed climatologies that affected those gyres. On the other hand, the control case with modified diffusion, CM, shows more realistic, though somewhat weak, northern recirculation gyre north of the Gulf Stream; it has been demonstrated by *Ezer and Mellor* [1992] that the intensity of this gyre influences the position of the Gulf Stream. The climatic changes under warm-climate conditions include weakening (and even a possible reversal in WS) of the cyclonic circulation in the Labrador Sea, owing to the warm pool created there (Figure 3). The weakening of the Gulf Stream is more pronounced in the case with no relaxation to climatology (Figure 10e) than in the standard case (Figure 10b), as the relaxation helps to maintain the Gulf Stream signature despite the climatic changes.

Since the model uses a bottom-following sigma coordinate system and has a prognostic turbulence closure scheme, dynamics of bottom boundary layers are represented in the model quite well; one can also easily analyze the near-bottom flow, looking at the lowest active layer in the model. *Ezer and Mellor* [1997] shows that even coarse resolution sigma models generate strong bottom flows, such as the deep western boundary current (DWBC), while z-level models may need much higher horizontal and vertical resolutions to generate intense bottom currents [*Winton et al.*, 1998]. The mean kinetic energy at the bottom layer (Figure 11) shows that the DWBC, at about 1500m depth, dominates the near bottom flows. The DWBC plays a major role in the mechanism of transporting cold water masses to the deep North Atlantic from the formation zone near the straits between the Greenland Sea and the Atlantic Ocean. In the warm-climate cases (WS and WM) the DWBC intensity is reduced significantly, and thus contributing to the weakening of the THC (Figure 6) and to the reduction in northward MHF (Figure 5). This result is

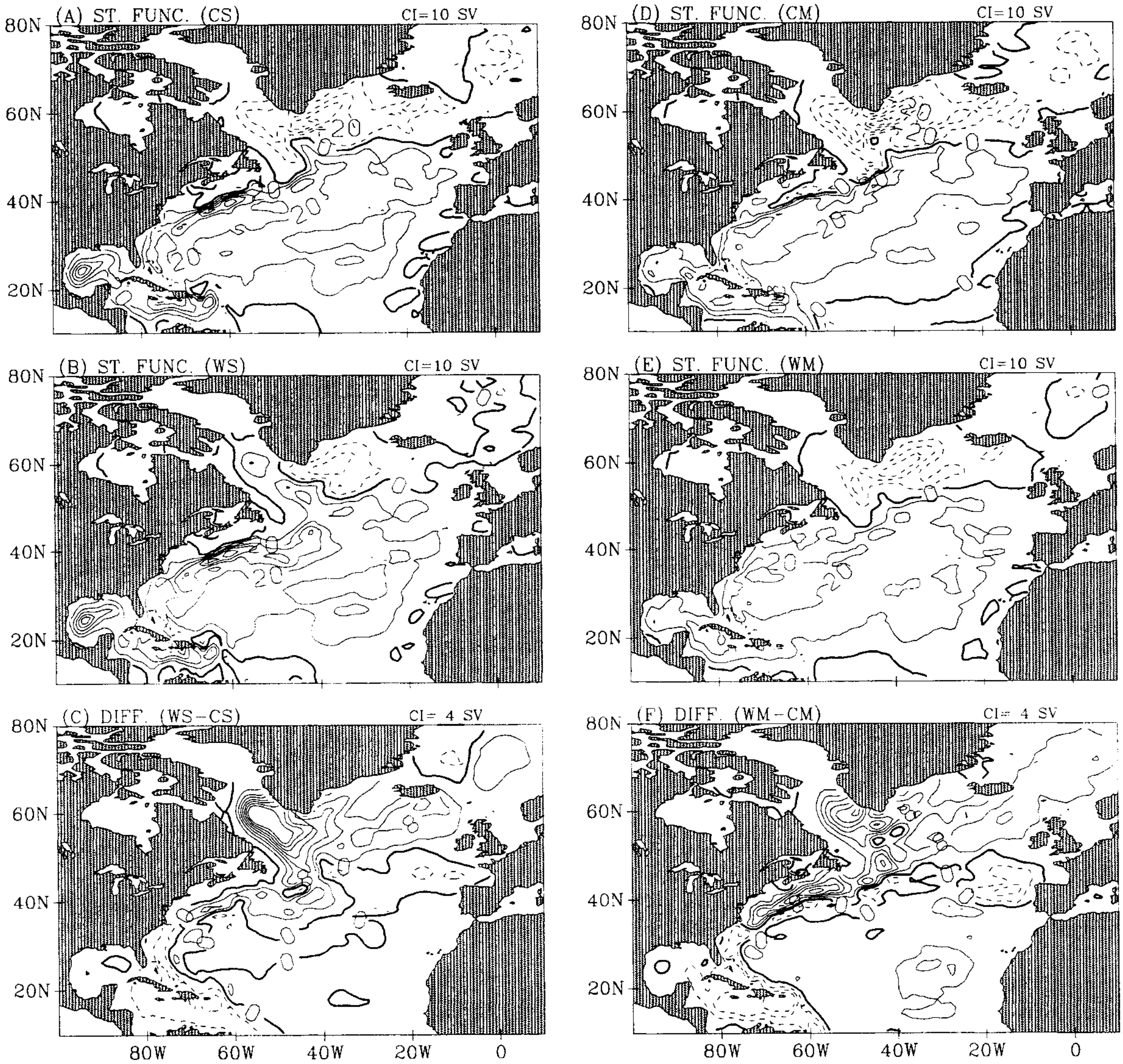


**Figure 10.** Same as Figure 7, but for the vertically integrated stream function. Contour interval is 10 Sv, except (c) and (f) were it is 4 Sv.

consistent with previous studies using other models [*Döscher et al.*, 1994; *Gerdes and Koberle*, 1995]. However, it is also demonstrated here that horizontal diffusion in a sigma coordinate ocean model may have a significant effect on the bottom flow, which could explain the changes seen in the deep gyres of the THC (Figure 6).

Changes occurred at the region of deep-water formation in high latitudes are of particular importance, thus the vertical component of the bottom layer flow is shown in Figure 12 (only the case with modified diffusion is shown here since the standard diffusion case is almost identical in those calculations). In the control run (Figure 12a), downslope flows (shaded areas) are evident in the Denmark Straits, the Faroe Island-Iceland Straits and along the Greenland coast, with particular strong downslope component south of Iceland and south of Greenland. This downslope flow is consistent with bottom boundary layer dynamics, where Ekman veering of the near-bottom flow to the left of the mean current is expected. Under warm-climate conditions, the stratification of the upper layers in high latitudes is more stable (Figure 2) so downward mixing of cold surface waters is reduced, and the boundary current is warmer and

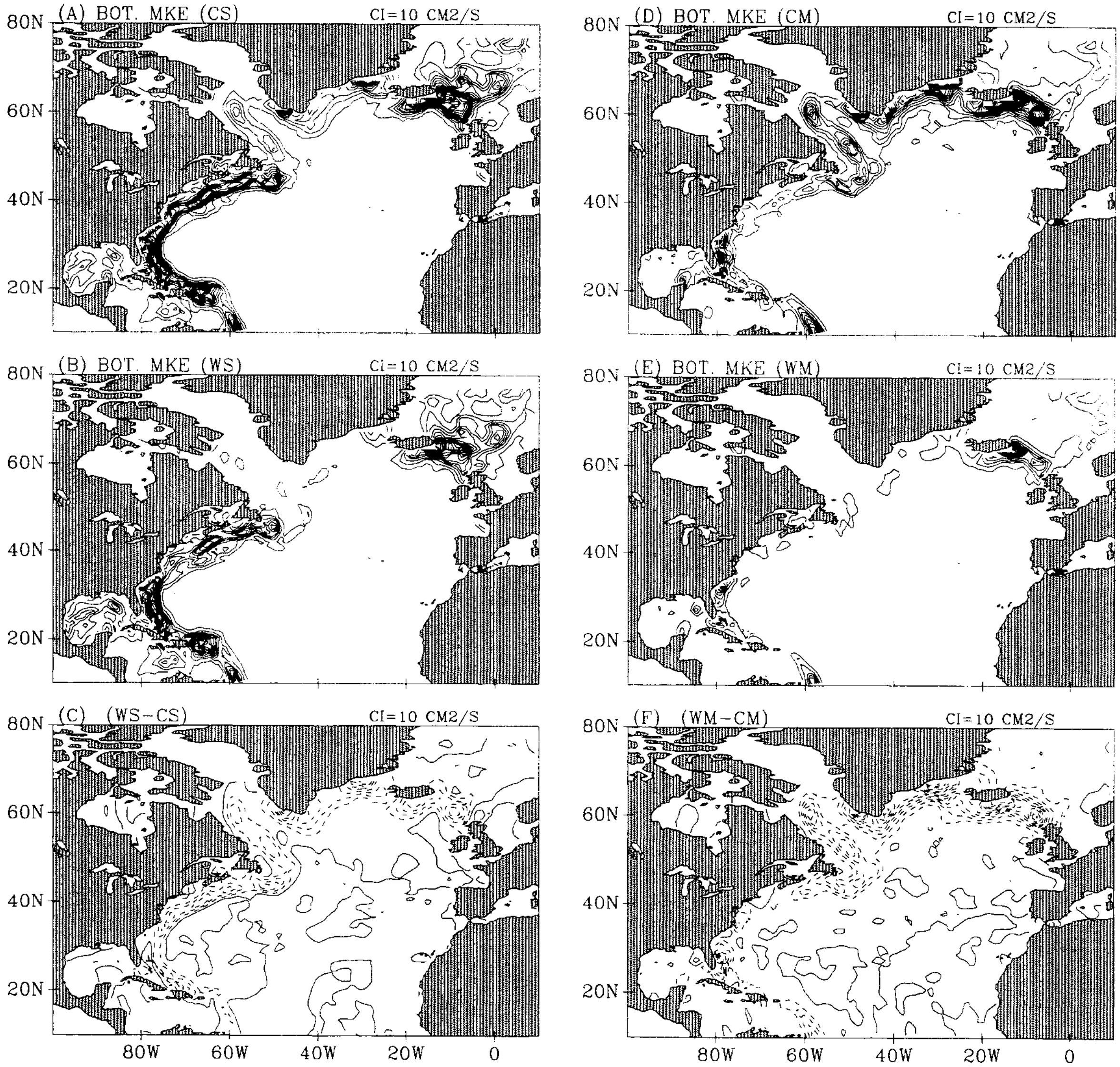


**Figure 11**. Same as Figure 7, but for the near-bottom (lower sigma level) mean kinetic energy (MKE). Contour interval is 10 $cm^2 s^{-1}$.

weaker (Figure 11). The result is a significant reduction of downslope flows in regions most important for deep-water formation. The offshore upwelling southeast of Greenland is also reduced. The reduction of surface mixing and the change in circulation also affect another important area of water formation, the Labrador Sea.

## 4. DISCUSSION AND CONCLUSIONS

This study follows closely on the footsteps of previous studies [*Ezer and Mellor*, 1994, 1997; *Ezer et al.*, 1995; *Ezer*, 1999, 2001] which used the same sigma coordinate coastal ocean model (POM) as a tool to study basin-scale climate processes previously studied mostly with other types of ocean models. The model, with its low horizontal diffusivities, turbulence mixing scheme and sensitivity to coastline features and bottom topography, provides a different perspective on some processes that may not be well-resolved in standard coarse resolution climate models.

The ocean model response to an idealized "warm-climate" forcing is consistent with those obtained by coupled ocean-atmosphere climate models [*Manabe et al.*, 1991;

*Manabe and Stouffer*, 1995, 1997; *Haywood et al.*, 1997] to some extent, for example, in the net change of the intensity of the THC. However, significant differences between the sigma coordinate ocean model calculations and previous studies require further explanations. The adjustment time-scale to abrupt climate changes in the ocean-only model, 10-20 years, is generally shorter than the adjustment time-scale in coupled ocean-atmosphere climate models, 50-100 years. This difference can be partly attributed to the lack of realistic atmospheric feedback in the surface forcing. However, it is also quite likely that the more intense deep flows and more realistic mixing processes in the higher resolution ocean model contribute to a faster transfer of the climate change signal to the ocean interior. The horizontal and overturning circulation in this model have more spatial gyre structure than coarse resolution climate models do and thus climate changes results in spatial differences and local circulation changes that are absent from coarse resolution models. Flows over sills and in particular the exchange of water masses between the Greenland and the Atlantic Ocean are represented in the sigma ocean model, while in many climate models most of the deep water formation occurs unrealistically south of the sills. An interesting result was the relative cooling of middle latitudes (despite the imposed surface heat flux) owing to the weakening of the Gulf Stream current, which transports warm subtropical waters to higher latitudes. Reduction in deep water formation near the Greenland and Iceland coasts have been clearly identified, a result of weakening of the western boundary current and consequently a reduction in its downslope Ekman component.

The study also examine the sensitivity of climate simulations to the along-sigma diffusion in the Princeton Ocean Model, and in particular, the effect of the relaxation of model temperature and salinity fields to the observed climatology. An alternative approach, using the model own climatology (with some time-delay), has been tested successfully. An unexpected result was the fact that the diffusion relaxation term affects the control runs more than the warm-climate runs (Figure 4), implying that the climatic changes are primarily controlled by surface fluxes rather than by horizontal diffusion; near bottom flows were affected by the diffusion parameterization to some extent. Changes in thermohaline circulation and meridional heat flux from current climate to a warm climate conditions are quite robust independent of the particular diffusion used, and the relaxation term does not seem to prevent the model from simulating climatic oceanic changes, as originally feared.

While idealized studies of the type presented her are useful in studying the adjustment process that takes place in

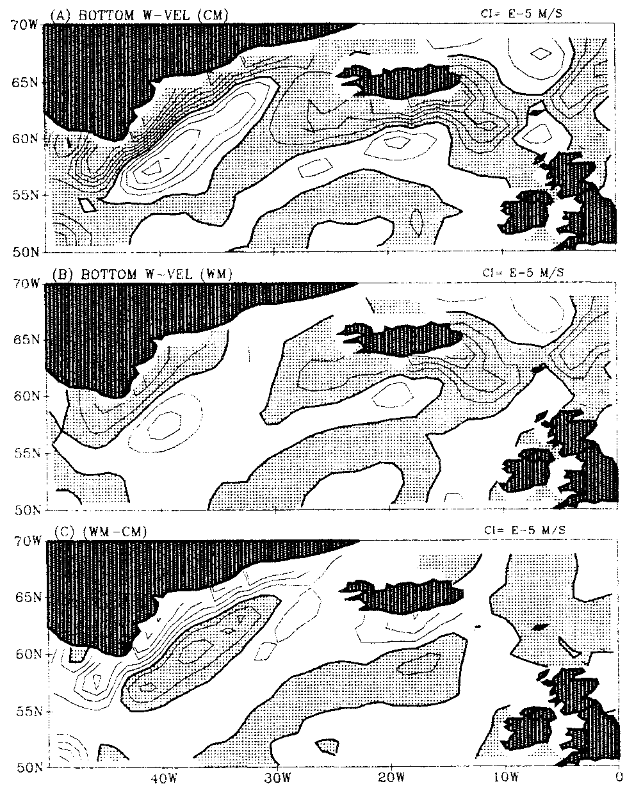


**Figure 12**. The near-bottom vertical velocity component calculated from the lower sigma level in the model averaged over the last 5 years of the model calculations: (a) CM, (b) WM and (c) WM-CM. Shaded regions indicate negative values, i.e., downslope flows. Contour interval is $10^{-5}$ m $s^{-1}$.

the ocean in response to abrupt climate changes, one should also mention the possible limitations of the study compared with studies with global coupled climate models. For example, the atmospheric feedback would have modified the heat, salt and momentum (wind stress) fluxes, but was ignored here (climate models indicate, though, that the THC is more affected by surface heat and salt fluxes than by wind stress, so neglecting wind changes may not significantly alter the results). The use of a basin model requires specification of open boundary conditions; climatic changes in the exchange of heat and momentum between the Greenland Sea and the Arctic Ocean were ignored, as well as effects associated with sea-ice. The lateral boundary conditions in the south should not have a significant effect on the results for the time scales involved here, but the boundary conditions in the north may have, since anomalies have predominantly propagated from the north southward. In any case, the implications of unresolved oceanic spatial changes and mixing processes in coupled climate models, as indicated here, should be further investigated.

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