THERMOHALINE CIRCULATION VARIABILITY IN THE NCAR CLIMATE SYSTEM MODEL (CSM)

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Abstract.

We describe here the decadal/interdecadal variability found in the thermohaline structure and circulation in the North Atlantic domain from the first multi-century numerical integration of the Climate System Model developed at the National Center for Atmospheric Research. Emphasis is placed upon the high latitude thermohaline behavior in the upper ocean and main thermocline, from about 30°N to 80°N, where the strongest variability occurs. A rich structure of oceanic variability is found and elucidated by relating it to some aspects of the air-sea and ice-ocean interactions. Important timescales of oceanic variability in the decadal/interdecadal range are found at 8-12 years and at 30-50 years, with distinct spatial structures. The 30-50 year band, in particular, is associated with patterns of surface anomalies similar to the ones found in the GFDL coupled model in the same frequency range. The evolution of these anomalies also appears to be quite similar in the two models. The decadal signals are associated with smaller spatial scales. Southward propagation of these anomalies along the eastern coast of North America suggests the possibility of an important role played by wave dynamics in the model thermohaline variability.
1. Introduction

Variability at interdecadal timescales has emerged in several observational studies of surface atmospheric and oceanic quantities (Deser and Blackmon 1993, Kushnir 1994, Hurrell 1995). Numerical simulations with ocean-only models have suggested that the ocean thermohaline circulation may play an important role in generating climatic signals with these timescales. In fact, interdecadal variability of the ocean thermohaline circulation has been found in a large number of ocean-only integrations and with different choices of the model parameters (Weaver and Sarachik 1991, Weaver et al. 1993, Greatbatch and Zhang 1995, Chen and Ghil 1995, Capotondi and Holland 1997). Some of these simulations (Weaver and Sarachik 1991, Weaver et al. 1993, Greatbatch and Zhang 1995, Chen and Ghil 1995) exhibited a spontaneous oscillatory model behavior in the presence of a steady surface forcing, while in other studies (Capotondi and Holland 1997) preferred timescales of variability resulted from the application of a surface stochastic forcing. The nature of the model variability can be very sensitive to the surface buoyancy forcing (Capotondi and Holland 1997). In fact, depending upon the surface thermohaline boundary conditions used, some feedbacks between sea surface temperature (SST) and sea surface salinity (SSS) and thermohaline circulation may be altered or even suppressed. Thus, it is clear that a realistic representation of the surface forcing and of all the feedbacks affecting the thermohaline circulation is fundamental for realistically modeling its variability.

Coupled ocean-atmosphere General Circulation Models (GCMs) have also shown the presence of climate variability at interdecadal timescales (Delworth et al. 1993, DMS hereafter, Zorita and Frankignoul 1997, Timmermann et al. 1997, Griffies and Bryan 1997, GB hereafter). However, the nature of such variability appears to be somewhat different in different models and in different integrations. Model limitations due to resolution and physical parameterizations, and the need to use "flux adjustments" to maintain the model mean state close to climatology may be responsible for these
The Geophysical Fluid Dynamics Laboratory (GFDL) coupled model (DMS) shows variability in the range 30-50 years. This variability is associated with fluctuations in the strength of the meridional overturning in the ocean model, and the patterns of SST variability show some similarity with the analysis of observational data by Kushnir (1994). The analysis performed by DMS does not seem to indicate any active atmospheric role in setting the timescale of these fluctuations. As suggested by Griffies and Tziperman (1995) from the analogy with the behavior of an idealized box model, the interdecadal variability described by DMS may be interpreted as a damped oscillatory ocean mode stochastically forced by the atmosphere. However, in a recent study, Weaver and Valcke (1998) are unable to reproduce the interdecadal signal when applying a stochastic forcing to the ocean-ice components of the GFDL model, thus indicating the possibility of a more active participation of the atmosphere in the coupled model interdecadal variability.

Variability at 20 years and 10 years has been found in the ECHAM-1/LSG coupled ocean-atmosphere model by Zorita and Frankignoul (1997). This variability affects only the upper ocean, and seems to be associated with fluctuations of the wind-driven circulation, directly forced by the atmosphere.

The ECHAM-3/LSG coupled model, on the other hand, shows fluctuations in the thermohaline circulation with a period of approximately 30 years (Timmermann et al. 1997). The authors suggest that changes in the thermohaline circulation lead to the development of SST anomalies in the high-latitude North Atlantic, which in turn determine changes in the North Atlantic Oscillation and consequent anomalous freshwater fluxes in the area of deep water formation in the North Atlantic. Thus, the atmospheric response plays an active role in this variability, which appears to be a coupled air-sea mode.

In this paper we analyze the thermohaline circulation variability in the Climate
System Model (CSM) recently developed at the National Center for Atmospheric Research (NCAR). This model includes ocean, atmosphere, ice and land surface model components (Boville and Gent 1998), and provides a fairly good simulation of the mean climate (Boville and Hurrell 1998) without the need for flux adjustments. Several aspects of the model climatology and variability have been analyzed in a series of papers which have been published as a special issue of Journal of Climate. However, an analysis of the model thermohaline circulation is missing. Our objective here is to fill that gap, and provide a comprehensive description of the model variability at interdecadal timescales and its connection with variability of the thermohaline circulation. Our emphasis will be on the analysis of the ocean component of the coupled system, and on the comparison with the findings of other coupled GCMs.

The paper is organized as following: in section 2 we summarize some aspects of CSM relevant to the present study. In section 3 we present a general overview of the thermohaline circulation variability found in CSM over the total duration of the integration (with the exclusion of the first 30 years). We will see that a long-term (about 150-170 years) variation in the strength of the thermohaline circulation leads to the establishment of different dynamical regimes in the North Atlantic. Thus, we will analyze these two regimes separately, focusing in particular on two periods, which are described in sections 4 and 5. Finally, in section 6 we will discuss our results and draw our conclusions.

2. The model

The NCAR Climate System Model (CSM) has been described in detail in Boville and Gent (1998). Here we give a brief summary of its general philosophy and of those aspects of the model that are relevant for the present study, our emphasis being on the ocean component.

CSM includes the NCAR ocean model (NCOM; Gent et al. 1998) coupled to the
latest version of the NCAR Community Climate Model (CCM3; Khiel et al. 1998), a
dynamic sea-ice model (Weatherly et al. 1998), and a land surface model (LSM; Bonan
1998). The component models communicate through a flux coupler, where the turbulent
air-sea fluxes are computed given the surface air and sea temperatures. The present
version of CSM does not include a river runoff model, but the precipitation field over
the ocean is scaled by a constant factor to close the water budget. This factor is applied
daily, and averages to 1.03 (Boville and Gent 1998).

NCOM is derived from the GFDL Modular Ocean Model with the addition of a
mesoscale eddy flux parameterization along isopycnal surfaces (Gent and McWilliams
1990) and a nonlocal planetary boundary layer parameterization (Large et al. 1994).
The horizontal resolution is 2.4° in longitude and varies in latitude from a minimum
of 1.2° at the Equator and in the Arctic to a maximum of 2.3° at mid-latitudes.
The climatology produced by the uncoupled ocean model is in fairly good agreement
with observations (Large et al. 1997). In particular, the meridional heat transport
in the uncoupled ocean model is relatively close to the heat transport implied by the
atmospheric model. As a consequence, no artificial local adjustments to the surface heat
flux into the ocean have been necessary to keep the ocean model close to its uncoupled
equilibrium. Similarly, no “corrections” have been applied to the freshwater flux, except
for the constant scaling of the precipitation field to compensate for the lack of river
runoff.

The coupled integration was started with an oceanic steady state spun-up to be in
equilibrium with the CCM3’s climatology. The coupled model underwent a small initial
adjustment over the first few years of integration, but eventually it reached remarkably
stable SST and SSS distributions over most of the ocean during the remaining years of
the 300-year integration. However, the deep ocean shows significant drifts (Bryan 1998)
especially in the salinity field, probably due to the model overestimate of ice divergence,
leading to an export of freshwater from the ice formation regions and consequent
increase in surface salinity and rate of sinking in those regions. In order to account for
the initial model adjustment, we have excluded from the analysis the first 30 years of
integration. Thus, the period that we have considered consists of 270 years, from year 30
to year 299 of the CSM integration (with year 0 being the first year of the simulation).

One may wander how much of the variability that we describe in this study
is associated with the initial model adjustment after coupling. The ocean model
adjustment consists of a dynamical phase, in which the velocity field adjusts to the given
density structure (the timescale is determined by wave propagation, and may range
from a few years to a few decades), and a thermodynamic phase which is the adjustment
of the density field itself (the timescale is dictated by the very slow diffusion processes,
and it can be of the order of hundreds of years). In the present coupled calculation the
model density structure is not in equilibrium, but slowly drifting away from the initial
state in the deep ocean, as mentioned earlier. However, we believe that after 30 years
the dynamical adjustment has been achieved, and consider the model evolution and
variability around the (slowly varying) background density structure.

Since the North Atlantic is one of the two major sites of deep water formation
driving the ocean thermohaline circulation, we have concentrated our analysis on the
North Atlantic sector of the global ocean model, from 30°S to 80°N, as shown in figure
1. Also, as a first approach, we have considered annually averaged model data. A subset
of the analysis, repeated with winter data (December-March), gave very similar results.

Some studies (Bryan 1986, Weaver et al. 1993, Tziperman et al. 1994) have shown
that salinity perturbations in areas of deep-water formation can give rise to large, and
often catastrophic, responses in the ocean thermohaline circulation. Therefore, it is
very important to try to assess how "realistic" the freshwater flux in the high-latitude
North Atlantic is, given the presence of the precipitation scaling factor, in order to
understand the possible consequences of the river runoff scheme on the thermohaline
circulation variability. Doney et al. (1998) have compared the freshwater flux in the
coupled model (based on a 10-year climatology) with a surface flux climatology (denoted as NCEP-STR). At high latitudes, the major differences between the NCEP-STR climatological freshwater flux and the coupled model freshwater flux are associated with the too strong sea ice cycle in the coupled model. As we will show in a later section, the North Atlantic sea ice distribution extends too far south and east relative to observations, giving rise to an excessive meltwater input along the ice margin (a band extending in the southern Labrador sea, along the southern Greenland coast and around Iceland). Thus, although the freshwater flux field in the coupled model shows large discrepancies with observations (as large as 7 m/y) in the North Atlantic, the origin of these discrepancies is essentially determined by the model unrealistic ice distribution while the effect of the river runoff factor appears to be minor. The large meltwater input from the ice can be expected to affect the rate of deep water formation in an unrealistic way. However, we will show in the next section that deep water formation in the model seems to occur south of the ice edge, and is not affected by local ice processes. Also, as far as meridional overturning stability is concerned, we should notice that the studies mentioned above (Bryan 1986, Weaver et al. 1993, Tziperman et al. 1994), in which catastrophic collapses of the thermohaline circulation were found, were performed using a strong restoring for SST. It has been shown in other studies (Zhang et al. 1993, Rahmstorf and Willebrand 1995, Capotondi and Saravanan 1996, Capotondi and Holland 1997) that the possibility for the SST to respond to changes in the ocean circulation plays an important stabilizing role. The effective average restoring coefficient diagnosed by Doney et al. (1998) in the coupled CSM integration is indeed about a third of the one used in the standard bulk flux formula, thus suggesting a larger stability of the meridional overturning to salinity perturbations in the coupled model with respect to the results found in some ocean-only studies using strong SST restoring.
3. Thermohaline circulation variability: year 30 - year 299

In this paper we concentrate on the North Atlantic sector of the coupled model, where formation of North Atlantic Deep Water (NADW) occurs. The mean meridional streamfunction from year 30 to year 299 is shown in figure 2. The sinking occurs south of 60°N and reaches down to a depth of approximately 2500m. Its maximum intensity, around 30° – 40°N is 30Sv, which is somewhat larger than the observational estimates. In fact, the overturning strength increases from a value of about 20Sv in the uncoupled CSM ocean model (Gent et al. 1998) as a result of the changes in the surface heat and freshwater fluxes due to the coupling with the atmospheric and ice components.

In this analysis we will be concerned with the variability of the meridional overturning around this mean state. To characterize the strength of the meridional overturning and its variability we have chosen to consider the maximum value of the meridional streamfunction as a function of time (the alternative approach of using the value of the meridional streamfunction at a fixed point, such as, for example, the location of the maximum mean overturning, gave very similar results). The time series of this index, from year 30 to year 299, is shown in figure 3.

The meridional overturning strength decreases over the first 100 years, remains low until approximately year 170, and then increases again, remaining relatively large over the last 100 years. The variation of the signal around its mean value is as large as 6Sv, which is about 20% of the mean value.

Most of this variability is captured by the leading EOF of the meridional streamfunction (figure 4), which accounts for 60% of the total variance. The pattern of this mode (figure 4, top), has a basin-wide character extending as deep as 3000 m, and extending into the South Atlantic. There is some resemblance between this pattern of variability and the pattern of meridional overturning EOF-1 found in the GFDL coupled model (GB). However, in the GFDL model that pattern represents a basin-wide trend of the overturning (or a fluctuation not resolved by the 200 year record considered) and
not a long-term fluctuation of the thermohaline circulation as in the present case.

These large excursions in the strength of the meridional overturning can be expected to have a very large impact on the North Atlantic model climate. In fact, changes in the strength of the meridional overturning are associated with changes in the northward advection of heat and salt, so that the resulting temperature and salinity conditions at high northern latitudes can be expected to be very different. In order to illustrate these differences we have focused on two periods, as shown in figure 3. The first period (period 1 hereafter) ranges from year 110 to year 189, while the second period (period 2) ranges from year 200 to year 299. Period 1, which is 80 years long, corresponds to the phase of weaker than average overturning, while period 2 spans the last 100 years of larger than average overturning strength. In figure 5 we compare the mean northward heat transport in period 1 and period 2 as a function of latitude. In both periods the largest heat transport is around 15°N. Differences in heat transport between the two periods are of the order of .1PW. The differences in temperature and salinity between the two periods (Period 2 minus Period 1) are shown in figure 6 for different depths (surface, 580m, and 2000m). At the surface, the largest differences in both SST and SSS are found in the area directly affected by the North Atlantic current system. Differences in temperature (left panels) are as large as 6°C, and differences in salinity (right panels) are as large as 1.5ppt. These differences decrease with depth, but even at 2000m they are still appreciable.

A EOF analysis of the SST and SSS fields (figure 7) over the 270 year period shows dominant patterns of variability (top panels) very similar to the surface difference fields in figure 6 (top panels). The EOF calculation has been performed considering the North Atlantic domain from 30°S to 80°N. However, only the model domain north of 30°N is displayed. EOF-1 of SST accounts for 60% of the variance, and EOF-1 of SSS accounts for 44% of the variance. Both EOFs have been normalized by the corresponding standard deviation, so that they can be considered as mean surface anomalies over the
total period of interest.

The associated time series (figure 7, bottom panels, thick solid line) follow very closely the time series of the meridional overturning (dashed line). All time series are normalized to unit standard deviation. The time series of the leading SST and SSS EOFs appear to be in phase, and to slightly lag the overturning. Therefore, one can interpret the SST and SSS anomalies in figure 7 as the result of larger than average (or lower than average) northward advection of heat and salt from the surface tropical waters to high latitudes by the thermohaline circulation.

The changes in SST and SSS associated with the large excursion of the overturning strength lead to different ice distributions at high latitudes in the North Atlantic. Figure 8 compares the mean ice concentrations and ice concentration standard deviations in period 1 and period 2, respectively. The largest differences in ice concentration are found in the Labrador Sea where, during period 1, the ice coverage is as large as 20% almost all the way to Newfoundland. We will see below that the Labrador Sea is an active area of deep water formation and the larger ice coverage (as well as the extensive freshwater capping) is responsible for the reduced rate (and variability) of deep water formation in this area during period 1 with respect to period 2. Ice concentration standard deviations are slightly larger in the Greenland-Iceland-Norwegian (GIN) sea during period 2, probably reflecting thinner ice conditions.

In both periods the ice distribution extends too far south and east compared with observations, and the ice coverage is particularly unrealistic in period 1 because of the extensive ice coverage in the Labrador sea. The unrealistic ice distribution may be partially a consequence of the model inability to represent the overflow of dense water over the sills between Greenland, Iceland, and Scotland (Ganopolski et al. 1997, Rahmstorf 1998). NADW forms in the model south of 60°N (figure 2), and there may be no contribution to this water mass from deep convection in the Greenland Sea. As a consequence, the northward heat transport by the thermohaline circulation would not
reach as far north as in the real world. In coupled models using flux adjustments this heat deficit can be corrected, but if no flux adjustments are applied colder temperatures and larger ice coverage in the Greenland Sea can be expected (Rahmstorf, 1998, personal communication).

Figure 2 suggests that deep water formation occurs south of 60°N. But what is the geographical distribution of the model deep convection? Convection is handled in the model by the surface mixed layer scheme (Large et al. 1994). An analysis of the vertical mixing in an ocean-only integration forced with monthly atmospheric fields (Large et al. 1997) indicates that convection mainly occurs in the Irminger Sea in late winter. However, in the coupled model, where the surface fluxes are from an interactive atmospheric component, the pattern of convection may be different. Meridional sections of mean temperature and salinity in period 1 and period 2 (not shown) indicates areas of weak vertical gradients between 45°N and 55°N. A surface layer of colder temperatures and lower salinities is present in both periods, but more enhanced in period 1. The corresponding density sections show that the stratification is weaker in those areas. As an example, figure 9 shows density sections at 42°W for period 1 and period 2, respectively. The doming of the isopycnals, considered a preconditioning factor for the occurrence of deep convection is especially pronounced in period 2, where higher surface salinities contribute to a higher surface density.

As a proxy for identifying locations more prone to the occurrence of convection we have used a vertically average measure of the stratification strength. Figure 10 shows a horizontal map of the buoyancy frequency averaged over the top 1500m for both periods. The mean surface velocities in both periods are also shown for comparison. In period 2 the band of weaker stratification (buoyancy frequency of the order of $5 \times 10^{-5}s^{-1}$) appears to be related to the cyclonic flow of the subpolar gyre and Labrador current. In period 1 the stratification appears generally stronger, except in the area east of 25°W and north of 50°N. Figure 10 gives us an indication of the areas where the
meridional overturning may be more sensitive to the presence of SST and SSS anomalies. This information will be used in the next two sections to understand the overturning variability and its link with surface anomalies.

The different climate scenarios in the high-latitude North Atlantic associated with different overturning strengths appear to be characterized by different dynamical regimes. It is evident from figure 11(bottom) that the dominant time scales of variability change with changing overturning strength. This visual impression is expressed more quantitatively by a wavelet spectrum of the meridional overturning time series, shown in figure 11(top). The wavelet spectrum shows the frequency content of the time series as a function of time, and it is displayed in the time/period domain. A Morlet function was used for the wavelet transform. Variability at 10 year period, as well as in the band 40-50 years, appears much more pronounced during the phases of large overturning (years 30-100 and years 200-299) than during the weak meridional overturning phase (years 110-190).

Another indication of a difference of regimes is shown in figure 12. Here, using a Hovmöller diagram of the variability in the northward velocity component along 40.4° N and at a depth of 262m, we see evidence of westward propagation during the strong overturning phase (years 200-270) but little during the weak overturning phase (years 105-200). This variability has a near decadal period when present and represents shifts in the Gulf Stream axis at this latitude. We will return to these westward propagating anomalies in section 5, where the variability in period 2 is analyzed in detail.

Because of these indications of differences in dynamics between Period 1 and Period 2, a separate analysis of the variability during the two different overturning phases appears necessary. It will be presented in the next two sections. We will see that the different ice conditions seem to have a large effect on the variability found in the two different periods.
4. Period 1 (year 110-189)

Period 1 spans the 80 years characterized by weaker than average thermohaline circulation, and larger than average ice coverage (figure 8). The dominant patterns of variability in the surface ocean properties are captured by the leading EOFs of the SST and SSS fields. Figure 13 shows the first two SST EOFs, while figure 14 displays the first two SSS EOFs. EOF-1 and EOF-2 of SST account for a comparable amount of variance (EOF-1 accounts for 21%, while EOF-2 accounts for 20%). The first SST EOF has a dipole-like structure, with a band of positive anomalies along the southeastern tip of Greenland, across the Denmark Strait and Norwegian Sea, and a tongue of negative anomalies off of Newfoundland. EOF-2, on the other hand, is characterized by a circular, one-signed anomaly, centered at approximately 55°N, 35°W. The SSS EOFs (figure 14) show similar patterns, but in reverse order: EOF-1 has a major center of action at approximately the same location as EOF-2 of SST, while EOF-2 has a dipole-like pattern which resembles the one of SST EOF-1. A comparison of the corresponding time series (not shown) shows, in fact, that EOF-2 of SST is correlated with EOF-1 of SSS (maximum correlation coefficient is .8 at lag 0), and EOF-1 of SST is correlated with SSS EOF-2 (maximum correlation coefficients are .6 at lag 0 and -.6 at lag 4). Thus, hereafter we will refer to these patterns as the “dipole” pattern (SST EOF-1 and SSS EOF-2), and the “monopole” pattern (SST EOF-2 and SSS EOF-1). Notice that the dominant time scales for the monopole pattern are in the range 10-20 years. The dipole pattern, on the other hand, seems to be associated with somewhat shorter time scales, in the range 8-10 years.

The first issue we would like to explore is the relationship between these dominant patterns of SST and SSS variability and the ocean thermohaline circulation. The lag-correlation between the time series of meridional overturning and the time series of SST and SSS for both the monopole and dipole patterns are shown in figure 15. In both cases correlations are smaller than .5, indicating a very weak link between SST and SSS.
anomalies and meridional overturning during this period. In fact, the major centers of action for both patterns appear to be displaced (especially the dipole pattern) with respect to the area of weaker stratification during this period, as shown in figure 10. This means that the largest surface anomalies are not found in areas where the rate of deep convection can most effectively be altered. Conversely, changes in the overturning strength do not seem to be responsible for the dominant patterns of SST and SSS variability.

This suggests that the sources of these anomalies may be found in the ocean interaction with the ice and/or atmospheric components of the CSM coupled model. In the remainder of this section we will concentrate upon the dipole-pattern. The location of the positive anomaly band in the dipole pattern is roughly coincident with the area of largest gradients in ice concentration (see figure 8) and largest ice concentration standard deviation during period 1. Also, anomalies in the form of bands of opposite polarities, as in the dipole pattern, are often associated with ice variability due to changes in ice extent and to ice advection processes (Fang and Wallace 1994). In order to test the possible link between the dipole pattern and ice processes, we have computed the correlation between the time series of SST for the dipole pattern and the ice concentration at each ocean model grid point. The largest correlations are found at lag -1 year (not shown, maximum value is -.63, with the ice leading) and at lag 0. The correlation map at lag 0 is shown in figure 16(top). The largest correlations are negative, implying that positive SST anomalies are associated with reduced ice concentration, and vice versa. Since the dipole pattern is characterized by salinity anomalies of the same sign as the SST anomalies, reduced ice extent would also be associated with higher than normal salinities. The largest correlation between SST and ice concentration is .83 east of Iceland. To further characterize the correlation between the dipole pattern and ice variability we compare in figure 16(bottom) the SST time series for the dipole pattern with the time series of ice concentration in the area of largest ice concentration.
variability (ice concentration standard deviation larger than 10%). The agreement between the two time series is remarkable.

Correlations between ice concentration and SST anomalies at decadal time scales has emerged in some observational studies. In particular, Deser and Blackmon (1993) find a strong correlation between winter SST anomalies east of Newfoundland and ice concentration in the Davis Strait/Labrador Sea (from Agnew 1991) at decadal time scales. In their case the largest correlation is found when the ice concentration leads the SST anomalies by 1-2 years. Larger than normal ice concentrations in the Davis Strait/Labrador Sea precede colder than normal winter SST east of Newfoundland. The (speculative) interpretation offered by Deser and Blackmon (1993) for this correlation is in terms of advection of ice anomalies out of the Labrador Sea. Although some correlation is found off the Newfoundland coast between the dipole pattern and the ice concentration in the CSM model, such correlation is much smaller than the one found by Deser and Blackmon (1993) in the same area. Also, the largest correlation is found at lag zero, and not with the ice anomalies leading the SST anomalies as in Deser and Blackmon (1993). In the CSM model the area of largest correlation is east of Iceland. These differences may be due to an unrealistic ice distribution in the CSM model, especially in period 1, so that a local ice redistribution may be responsible for the development of SST and SSS anomaly.

Having identified a link between ice extent and SST (SSS) anomalies we would now like to understand how this correlation is established. Since the connection between the meridional overturning and the dipole pattern appears to be weak, we consider the possible effects of the surface atmospheric circulation. Figure 17 shows the pattern of surface wind stress regressed upon the time series of EOF-1 SST. The contours of the correlation between EOF-1 SST and ice concentration are also shown in the same figure for comparison. The pattern of surface wind stress is characterized by anomalous southeasterly winds converging toward the area east of Iceland where the largest
correlation is found. The effect of southeasterly winds may be twofold: they can advect
the ice away from the area east of Iceland and can produce a positive heat flux into the
ocean. This would explain the negative correlation between SST and ice concentration
in this area. Ice export from the area is equivalent to an effective freshwater export
(ice leaving the salt produced by brine rejection behind), thus explaining the negative
correlation between SSS and ice concentration. Conversely, the surface wind stress is
toward the southeast in the Labrador Sea. This is consistent with negative SST (and
SSS) anomalies east of Newfoundland being associated with larger than normal ice
anomalies due to ice advection from the Labrador Sea.

The main conclusion of this section is that a large fraction of the variability found
in the SST and SSS fields can be related to variability in ice concentration. Changes in
the surface wind stress (and resulting ice advection) appear to be consistent with the
correlation found between SST and SSS variations and local ice concentration. However,
an explanation of the dynamical links will require further analyses.

5. Period 2 (year 200-299)

Period 2 spans the last 100 years of coupled integration, and it is associated with a
larger than average overturning strength. The ice coverage during this period (figure 8)
is not as extensive as during period 1, and closer to present day ice conditions. We have
also seen in section 3 that there is evidence of westward propagation during period 2,
which is absent during period 1. In this section we would like to clarify the nature of
these westward propagating features, whose period seems to be close to 10 years (figure
12).

As can be seen from figure 3, the amplitude of the variability in the meridional
overturning during period 2 is larger than in period 1, and appears associated with
timescales in the decadal range (8-12 years) as well as with a longer term modulation of
30-50 years. EOF analysis of the meridional streamfunction during period 2 results in a
partial separation of these two ranges of timescales (figure 18). EOF-1, which accounts for 50% of the variance, is associated with the longer term variations (although decadal oscillations are also present), while EOF-2, which accounts for 15% of the variance, is essentially associated with the decadal range. The patterns of the two EOFs are very different. EOF-1 has a basin-wide character, similar to the pattern of the overturning over the whole period (figure 4), while EOF-2 has a major center of action localized around 50°N and a smaller anomaly of opposite polarity south of 40°N. The 30-50 year timescale, although marginally resolved by the 100 year record we are considering here, is of particular interest because of its similarity with a dominant timescale of variability found in the GFDL coupled model (DMS, GB). The pattern of overturning EOF-1, however, has a larger scale structure with respect to the EOF associated with interdecadal fluctuations of the meridional streamfunction in the GFDL model (GB), which has dominant variability in the northern half of the North Atlantic.

Similar timescales of variability are also found in the temperature and salinity fields. Figure 19 shows the leading EOFs of SST (top-left) and SSS (top-right) for this period, and the associated time series (lower panels). The patterns are very similar for both fields, with major centers of action around the southern tip of Greenland and in the Labrador Sea, and weaker anomalies of opposite polarity in the southern Greenland Sea and off the eastern coast of the United States. The principal components of both fields are also very similar, with fluctuations at decadal scales as well as a lower frequency modulation. This longer term modulation becomes relatively more dominant at depth, as shown in figure 20, where temperature and salinity EOFs at model level 25 (approximately 950 m) are displayed. Notice that at this depth it is the second EOF (and third, not shown) of both fields that captures fluctuating signals. The first EOFs are, in fact, associated with the long term model trend (Bryan 1998).

In order to more clearly isolate the interdecadal signal from the decadal one, we have repeated the EOF analysis on low-pass filtered model data. For that purpose we
have used a Fourier filter with half power at 18 years. Figure 21 shows the two leading SST EOFs for the filtered data. EOF-1 has a pattern very similar to the corresponding pattern obtained with unfiltered data (figure 19), but the fraction of total variance accounted for by this mode increases to 61% for the filtered data. The pattern of EOF-1 is very similar to the pattern of SST found in the GFDL coupled model (DMS) by taking difference maps between model SST for a period of anomalously weak circulation and anomalously strong circulation. Also, as noted by DMS, this pattern bears a resemblance with the pattern of variability resulting from the analysis of observational data by Kushner (1994).

SST EOF-2 (figure 21, top-right) has a major center of action east of Newfoundland, and a weaker anomaly of opposite polarity east of Iceland and along the southeastern tip of Greenland. Notice that the position of the negative anomaly seems to follow very closely the average position of the ice edge during this period. This appears to be an indication of the possible role played by ice melting/freezing processes (maybe associated with changes in the overturning and northward heat transport) in creating SST anomalies.

The time series of EOF-1 and EOF-2 (figure 21, lower panels) appear correlated at some lag. Although the timescale of the fluctuations varies over the 100 year period, there are events (such as the large oscillation around years 260-270) in which an approximate quadrature relationship exists between the two time series. A quadrature relationship between two EOFs is an indication of propagation. To explore the nature of this propagation we show, in figure 22, the evolution of EOF-1 plus EOF-2 from year 264 to year 280, which is approximately one half of a full oscillation. Year 264 corresponds to a maximum of EOF-1. The large positive anomaly around the southern tip of Greenland moves southeastward and becomes the smaller scale positive anomaly east of Newfoundland characteristic of EOF-2. In the mean time, a negative anomaly develops on either sides of Iceland along the ice edge, and slowly propagates
southwestward along the coast of Greenland to develop in a large negative anomaly south of Greenland and in the Labrador Sea half a period later. The positive anomaly east of Newfoundland moves southwestward along the eastern coast of North America while decaying in amplitude. Propagation of anomalies along the east coast of North America was also noted by GB. By looking at the evolution of dynamic topography EOF-1 and EOF-2, they found a westward propagation from the center of the basin and subsequent southward propagation along the eastern coast of North America.

Propagation characteristics are less clearly detected in the SSS fields. However, the leading SSS EOF (not shown), which accounts for 54% of the variance, has a very similar structure to SST EOF-1, and its time series is very highly correlated with the time series of SST EOF-1. Warmer SSTs are associated with higher salinities, and vice versa. Temperature and salinity anomalies seem to develop at the ice edge, suggesting a possible important role played by ice-ocean interactions in the thermohaline circulation variability. The low-passed filtered time series of meridional overturning is highly correlated with the dominant patterns of SST and SSS variability (not shown). In particular, during the large oscillation around year 270 the anomalous overturning strength is perfectly in phase with SST EOF-1 and SSS EOF-1. Thus, when surface waters are anomalously warm and salty south of Greenland and in the Labrador Sea, the overturning is at its maximum. The associated increase in northward heat transport may lead to ice melting around the ice edge and to the formation of colder and fresher surface waters. The latter would then be advected southward by the East Greenland current into the subpolar gyre to produce, half a period later, a cold and fresh anomaly south of Greenland and in the Labrador Sea. The center of action of EOF-1 for both SST and SSS is, in fact, partially overlapping the area of weaker stratification shown in figure 10, where convection is more likely to occur and could more easily be affected by surface density anomalies. This mechanism would thus explain the longer term variations of the thermohaline circulation purely in terms of ice-ocean interactions,
in the spirit of what was suggested by Yang and Neelin (1997) using a hierarchy of ice-ocean coupled models.

Delworth et al. (1997) find that the interdecadal variations in the thermohaline circulation of the GFDL coupled model are coherent with the variability in SST and SSS in the Greenland Sea. In the GFDL model, SST and SSS anomalies appear to be generated in the Arctic, be advected southward through the East Greenland Current, and then propagate around the subpolar gyre into the Labrador Sea and central North Atlantic. The thermohaline circulation is weakest when SSS over the sinking region is at a minimum, in agreement with what we find in our analysis. Reduced salinities over the sinking area would produce a "fresh water capping" responsible for isolating the cold near-surface layers from the warmer deeper layers, and thus reducing the heat loss at the sea surface, as documented by Lazier (1980) from observational data. Thus, there is a strong similarity between the evolution of SST and SSS anomalies and interdecadal thermohaline circulation variability in the GFDL coupled model and in CSM. The main differences appear to be associated with the geographical origin of the surface anomalies. While in the GFDL model they originate from ice variations in the Arctic, in CSM they seem to develop along the model ice margin, in the southern Greenland Sea. These differences are probably due to the very different ice distributions in the two models.

Having analyzed the interdecadal fluctuations of the thermohaline circulation, we now consider the variability in the decadal range. In particular we would like to understand how the propagating features shown by the Hovmöller diagram in figure 12 relate to the dominant patterns of variability and their evolution. To isolate the decadal range, we consider band-pass filtered model data between 6 and 14 years. Figure 23 shows the first two EOFs of the band-passed temperature field at model layer 15 (depth of approximately 260 m), the same layer for which the Hovmöller diagram in figure 12 was computed. Both patterns (top panels) have a dipole-like structure, with largest intensities off Newfoundland, and along the Labrador coast and east coast
of North America. They are approximately in quadrature, as it can be seen from the associated time series (figure 23, bottom panels). Using the same approach applied for the interdecadal time scales, we show, in figure 24, yearly maps of the combination of the two patterns from year 211 to year 219, covering almost a full period of the oscillations. The anomalies seem to originate in the Labrador Sea, reach their largest amplitude off Newfoundland, and then propagate southwestward along the eastern coast of North America. The evolution shown in figure 24 clarifies the nature of the westward propagation noticed at 40°N along a zonal line (figure 12). The propagation is actually southwestward along the coast. Although the amplitude of the anomalies decays southward, it is still appreciable at 30°N. While the southward movement of the anomalies along the Labrador coast may be interpreted in terms of advection by the mean currents, the southward propagation along the coast of North America can only be explained in terms of wave dynamics, since the mean currents are northward and large (up to 30 cm/s at this depth).

Numerical simulations in higher resolution ocean models (Doescher et al. 1994) have shown that the initial phase of the dynamical model adjustment to changes in the rate of deep water formation in the North Atlantic is accomplished through southward propagation of coastal Kelvin waves, in agreement with Kawase’s (1987) theory of the spin-up of the deep circulation. It may be that the southward propagation seen here is also a manifestation of the model adjustment to changes in the overturning strength. A clarification of the nature of the southward propagating anomalies and in particular of the role they play in the decadal model variability is offered in Capotondi (1999).

6. Discussion and Conclusions

In this paper we have described the variability in the North Atlantic thermohaline circulation of the CSM coupled model. The evolution of the meridional overturning over the 270 years of integration that we have considered is dominated by a long-term
variation with a timescale of approximately 150-170 years. The strength of the meridional overturning decreases by several Sverdrups over the first 100 years, remains lower than normal between year 100 and year 180, and then increases again to larger than normal values over the last 100 years of the integration. These changes in the strength of the meridional overturning are associated with large variations in the northward transport of heat and freshwater, and have a considerable impact on the North Atlantic model climate. In particular, the colder sea surface temperatures during the phase of weaker than average overturning result in a larger ice concentration in the Labrador Sea when compared to the period of stronger overturning.

The different ice conditions in the model North Atlantic are associated with different characteristics of the model variability. During the phase of weak overturning, when the ice concentration is larger, the amplitude of the thermohaline circulation variability is somewhat smaller, and it is only weakly correlated with the changes in the sea surface temperature and sea surface salinity fields. In fact, the dominant patterns of sea surface temperature and sea surface salinity appear to be displaced with respect to the area of weaker stratification where convection is more likely to occur. An important pattern of variability that emerges in the surface ocean during this period has a dipole-like structure and appears to be correlated with changes in ice concentration and anomalous wind stress.

During the phase of larger than average overturning (and reduced ice concentration), on the other hand, the thermohaline circulation variability appears more vigorous, and dominant timescales at 8-12 years as well as 30-50 years emerge. The 50 year timescale is of particular interest because a similar timescale is found in the GFDL coupled model (DMS, GB, Delworth et al. 1997). More importantly, the patterns of variability in the sea surface temperature and sea surface salinity fields are very similar in the two coupled models, and a southward propagation of anomalies along the eastern coast of North America is also found in both cases. Another interesting similarity is associated with the
origin of the sea surface temperature and sea surface salinity anomalies in the sinking region. In the GFDL model these anomalies appear to originate in the Greenland Sea and be advected southward by the East Greenland current and then propagate into the subpolar gyre and central North Atlantic (Delworth et al. 1997). They seem to be correlated with ice variability. Similarly, in CSM surface anomalies appear to originate at the ice edge, which in this model is further south, and propagate along the eastern coast of Greenland and into the subpolar gyre. The correlation of these anomalies with the overturning variability is similar to the one found by Delworth et al. (1997), namely the overturning is weakest when sea surface salinity is at a minimum over the sinking region.

Apart from this interdecadal signal, variability in the decadal range (8-12 years) is also found. The corresponding patterns of variability are characterized by smaller spatial scales, with a major center of action east of Newfoundland. Propagation of these anomalies along the eastern coast of North America is also found, reminiscent of the propagation of coastal Kelvin waves accompanying model adjustments to changes in the rate of deep water formation (Doescher et al. 1994). This boundary propagation (which cannot be attributed to advection because of the opposite direction of the mean currents) is also reminiscent of the boundary disturbances described by Greatbatch and Peterson (1996) in an idealized ocean model. The connection of these “boundary waves” with the model thermohaline circulation variability is analyzed in Capotondi (1999).

In this study we have mainly concentrated on the variability in the ocean component of the climate model. The role played by the atmosphere in the model interdecadal variability is largely unexplored. This will be analyzed in later studies.

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References


Delworth, T., S. Manabe, and R. J. Stouffer, 1997: Multidecadal climate variations in the


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Tziperman, E., J. R. Toggweiler, Y. Feliks, and K. Bryan, 1994: Instability of the


Figure Captions

**Figure 1.** Subdomain of the global ocean model considered for the analysis. The domain extends from 30°S to 80°N.

**Figure 2.** Mean meridional streamfunction (in Sv) for the CSM coupled integration over the period 30-299 years. Contour interval is 1Sv for negative values, and 2Sv for positive values.

**Figure 3.** Time evolution of the meridional overturning strength (in Sv) in the CSM coupled integration over the period 30-299 years. A period of weaker than average overturning strength (Period 1) and a period of stronger than average overturning strength (Period 2) are also indicated.

**Figure 4.** (Top) Leading EOF of meridional overturning variability (in Sv) over the period years 30-299. Contour interval is .5Sv. (Bottom) Time evolution of the meridional overturning EOF-1 (thick solid line), and (normalized) meridional overturning time series (dashed line). The two time series are almost coincident.

**Figure 5.** Comparison between the mean northward heat transport (in PW) in period 1 (dashed) and period 2 (solid) as a function of latitude.

**Figure 6.** Differences between the mean temperature (left panels) and salinity (right panels) distribution in period 2 and period 1. Temperatures are in °C and salinity in ppt. Difference fields are shown at the surface (top panels), at model level 20 (480m depth), and at level 30 (2000m depth).
Figure 7. (Top-left) EOF-1 of North Atlantic SST based on yearly averages of the CSM coupled results over the period 30-299 years. The EOF calculation was performed for the whole North Atlantic domain, but only the area north of 30°N is displayed. The EOF is normalized by its standard deviation. Contour interval is 0.5°C. Maximum value is 3.6°C. This mode accounts for 60% of the variance. (Top-right) EOF-1 of North Atlantic SSS. Contour interval is .1ppt. This mode accounts for 44% of the variance. (Bottom-left) Time series of EOF-1 of SST (solid line) normalized to have unit standard deviation, and normalized time series of meridional overturning index (dashed line). (Bottom-right) Time series of EOF-1 of SSS (solid line) normalized to have unit standard deviation, and normalized time series of meridional overturning index (dashed line).

Figure 8. Mean ice concentrations in period 1 (top-left) and period 2 (top-right), and ice concentration standard deviations for period 1 (bottom-left) and period 2 (bottom-right). Maximum value of standard deviation in period 1 is .14.

Figure 9. Meridional sections of potential density at 42°W associated with the mean temperature and salinity distributions in period 1 (top) and period 2 (bottom). Potential density is referenced to the surface. Contour interval is .2 sigma units.

Figure 10. Spatial distribution of vertically averaged (0-1500m) buoyancy frequency. Units are $10^{-4}s^{-1}$. Contour interval is $10^{-5}s^{-1}$. Mean surface velocities are also shown for comparison. Light shading defines the bathymetry at 1500m, while darker shading defines the model coastline at the surface.

Figure 11. (Top) Wavelet spectrum of the overturning index time series displayed in the bottom panel. The amplitude of the wavelet transform is shown as a function of time (horizontal axis) and timescale (vertical axis). The 10 and 50 year timescales are indicated by the horizontal dashed lines. (Bottom) The time series of the strength of the anomalous meridional overturning. This variability is around a mean value of 30Sv.
Figure 12. A Hovmöller diagram showing the northward component of velocity at a latitude of 40.4°N and a depth of 262m. The abcissa is longitude from North America to Europe and the ordinate is time (each block showing 90 years of the simulation). The left, middle and right panels show years 30–119, 120–209 and 210–300 respectively, time advancing from bottom to top.

Figure 13. EOF-1 (top-left) and EOF-2 (top-right) of CSM North Atlantic SST based on yearly averages of the CSM coupled results over the period 110-189 years. The EOF calculation was performed for the whole North Atlantic domain, but only the area north of 30°N is displayed. The EOFs are normalized by their standard deviations. Contour interval is 0.2°C for negative contours (dashed line), and 0.3°C for positive contours (solid line). EOF-1 accounts for 21% of the variance, and EOF-2 for 20%. (Bottom-left) Time series of EOF-1 (dashed) and smoothed with a low-pass (50% power at 6 year period) Fourier filter (solid). The time series has been normalized to unit standard deviation. (Bottom-right) As for (bottom-left), but for EOF-2.

Figure 14. EOF-1 (top-left) and EOF-2 (top-right) of CSM North Atlantic SSS based on yearly averages of the CSM coupled results over the period 110-189 years. The EOF calculation was performed for the whole North Atlantic domain, but only the area north of 30°N is displayed. The EOFs are normalized by their standard deviations. Contour interval is 0.1ppt. Negative contours are dashed. EOF-1 accounts for 17% of the variance, and EOF-2 for 12%. (Bottom-left) Time series of EOF-1 (dashed) and smoothed with a low-pass (50% power at 6 year period) Fourier filter (solid). The time series has been normalized to unit standard deviation. (Bottom-right) As for (bottom-left), but for EOF-2.

Figure 15. (Top) Autocorrelation of meridional overturning at different lags (solid), and lag-correlation of meridional overturning with SST (dotted) and SSS (dashed) for the “monopole” pattern (EOF-1 of SSS, and EOF-2 of SST). (Bottom) Lag-correlation of meridional overturning with SST (dotted) and SSS (dashed) for the “dipole” pattern (EOF-1 of SST and EOF-2 of SSS). Positive lags indicate that the overturning lags the SST and SSS anomalies, and vice versa for negative lags.
Figure 16. (Top) Correlation between the time series of EOF-1 of SST and ice concentration in period 1. [Positive contours solid, negative contours dashed]. (Bottom) Comparison between the time series of EOF-1 of SST (solid, sign inverted) and the time series of ice concentration in the area where the ice concentration standard deviation is larger than 10% during period 1 (dashed). Refer to figure 8 for the ice concentration standard deviation map.

Figure 17. Surface wind stress (in dyne/cm²) regressed upon the time series of EOF-1 of SST. Contours indicate the correlation between the EOF-1 SST time series and the ice concentration.

Figure 18. EOF-1 (top-left) and EOF-2 (top-right) of meridional streamfunction from year 200 to year 299. Countour interval is .5 Sv. EOF-1 accounts for 50% of the variance, while EOF-2 accounts for 15% of the variance. (Bottom-left) Time series of meridional streamfunction EOF-1. (Bottom-right) Time series of meridional streamfunction EOF-2.

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Figure 20. EOF-2 of temperature (top-left) and salinity (top-right) at model layer 25 (depth 959m) over the period 200-299 years. The EOF calculation was performed for the whole North Atlantic domain, but only the area north of 30°N is displayed. The EOFs are normalized by their standard deviations. Contour interval for temperature is .05°C for negative values and .1°C for positive values. Contour interval for salinity is .01ppt. Negative contours are dashed. Notice that at this depth the model multidecadal variability is captured by EOF-2, EOF-1 describing the model long-term trend at depth. (Bottom-left) Time series of temperature EOF-2 (dashed) and smoothed with a low-pass (50% power at 18 year period) Fourier filter (solid). The time series has been normalized to unit standard deviation. (Bottom-right) Time series of salinity EOF-2.

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Figure 22. Combination of EOF-1 and EOF-2 for the low-pass filtered SST over years 264 to 280, showing the evolution of the anomalies over approximately a half period. Every second year is shown. Contour interval is .5°C. Notice that only the domain north of 20°N is shown.

Figure 23. EOF-1 (top-left) and EOF-2 (top-right) of band-pass filtered temperature data at model layer 15 (depth of approximately 260 m). Data were filtered using a Fourier filter with 50% power at 6 years and 14 years. The EOF calculation was performed for the whole North Atlantic domain, but only the area north of 20°N is displayed. The EOFs are normalized by their standard deviations. Contour interval is .1°C. (Bottom-left) Time series of SST EOF-1 (solid) and SST EOF-2 (dashed). (Bottom-right) Time series of SST EOF-2.
Figure 24. Combination of EOF-1 and EOF-2 for the band-pass filtered temperature at model level 15 (262m depth) over years 211 to 219, showing the evolution of the anomalies over approximately one period. Every year is shown. Contour interval is .5°C for values larger than 1°C, and .25°C for values lower than 1°C. Notice that only the domain north of 20°N is shown.
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