Mechanism of Interdecadal Thermohaline Circulation Variability in a Coupled Ocean–Atmosphere GCM

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ABSTRACT

Interdecadal variability of the Atlantic thermohaline circulation (THC) is studied in the third version of the Hadley Centre global coupled atmosphere–ocean–sea-ice general circulation model (HadCM3). A diagnostic approach is used to elucidate the mechanism that governs the variability and its impacts on climate. An irregular and heavily damped THC oscillation with a period around 25 yr is identified. The oscillation appears to be forced by the atmosphere but the ocean is responsible for setting the time scale. Following a minimum in the THC, the mechanism for phase reversal involves the accumulation of cold water in the subpolar gyre, leading to an acceleration of the gyre circulation and the North Atlantic Current. This acceleration increases the transport of saline waters into the regions of active deep convection, raising the upper-ocean density and leading, after adjustment, to acceleration of the THC. The atmosphere stimulates this THC variability in two ways: 1) by forcing the subpolar gyre through (North Atlantic Oscillation) NAO-related wind stress curl and heat flux anomalies; and 2) by direct forcing of the region of active deep convection, also through wind stress curl and heat flux anomalies. The latter is not closely related to the NAO. The mechanism for phase reversal has many similarities to that found in a previous study with a much lower resolution coupled model, suggesting that this mechanism may be quite robust. However the time scale, and details of the atmospheric forcing, differ.

The THC variability in HadCM3 has significant impacts on the atmosphere not just in the Atlantic region but also more widely, throughout the global Tropics. The mechanism involves modulation by the THC of the cross-equator SST gradient in the tropical Atlantic. The SST anomalies induce a displacement of the ITCZ in the Atlantic basin with knock-on effects over the other ocean basins. These findings highlight the potential importance of the Atlantic THC as a cause of interdecadal climate variability on a global scale.

1. Introduction

The thermohaline circulation (THC) is a planetary-scale pattern of ocean currents and an important component of the climate system. It plays an essential role in maintaining the mean climate by transporting a large amount of heat, $O(10^{15} \text{W})$, from low to high latitudes (e.g., Broecker 1991; Weaver et al. 1999; Ganachaud and Wunsch 2000). Substantial changes in the THC and the associated ocean heat transport (OHT) would provoke major climate changes (e.g., Manabe and Stouffer 1999; Vellinga and Wood 2002; Dong and Sutton 2002a). There is therefore an obvious need to better understand the behavior of the THC, and especially the physical mechanisms that govern its time dependence.

This paper is concerned with understanding the processes that govern natural variability of the THC on interdecadal time scales. Progress in this area may help to differentiate anthropogenic climate change from natural climate variability (CLIVAR 1998; Anderson et al. 1998).

Observations show that the North Atlantic climate system possesses pronounced interdecadal variability in its sea surface temperature (SST) and atmospheric component (e.g., Deser and Blackman 1993; Kushnir 1994; Hurrell 1995; Delworth and Mann 2000). The long time scale associated with this variability suggests that ocean dynamics, and in particular the THC, may play an important role. This evidence supports the pioneering work of Bjerknes (1964) who conjectured that ocean dynamics could lead to interdecadal SST variability along the Gulf Stream extension. Curry and McCartney (2001) have presented observational evidence of interdecadal variations in the ocean gyre circulation and in the baroclinic transport of the North Atlantic...
Current linked to changes in the North Atlantic Oscillation (NAO).

The role of ocean dynamics in North Atlantic climate variability has also been studied using coupled general circulation models (CGCMs; Delworth et al. 1993; Timmermann et al. 1998; Grotzner et al. 1998; Watanabe et al. 1999; Christoph et al. 2000; Holland et al. 2001; Wu and Gordon 2002). A well-defined 50-yr oscillation in the North Atlantic occurs in the coupled global model of Delworth et al. (1993). The mechanism responsible for the variability is the reduced heat transport of a weaker THC that leads to the formation of a cold dense pool in the central North Atlantic. This cold dense water results in a stronger subpolar gyre, and thereby enhances the salinity transport into the convection region, leading to an acceleration of the THC. Timmermann et al. (1998) studied interdecadal variability in the coupled ECHAM3–Large-Scale Geostrophic (LSG) model and found a pronounced oscillation with a period of 30–40 yr. They also identified an important role for the THC, but the mechanism was somewhat different to that discussed by Delworth et al.

One issue of debate has been the role of the atmosphere in variability of the THC. Studies with ocean GCMs have demonstrated that much of the variability in the North Atlantic Ocean, including the THC, can be explained as an ocean response to atmospheric variability (e.g., Halliwell 1997, 1998; Visbeck et al. 1998; Seager et al. 2000; Eden and Willebrand 2001; Eden and Jung 2001). The extent to which the ocean feeds back to affect the atmosphere is much less clear. Timmermann et al. (1998) found the atmosphere to be an active player in their interdecadal mode. By contrast, Delworth and Greatbatch (2000) showed that the interdecadal variability of the THC in the Geophysical Fluid Dynamics Laboratory (GFDL) model (as discussed by Delworth et al. 1993), may be explained by a passive response of the ocean to internally generated interdecadal atmospheric variability—that is, the THC variability is not coupled in the sense that, for example, ENSO is coupled. This said, studies with atmospheric models suggest that the ocean (but not necessarily the THC) plays a significant role in interdecadal climate variability in the North Atlantic (Rodwell et al. 1999; Latif et al. 2000; Mehta et al. 2000; Hoerling et al. 2001).

The lack of consensus regarding the specific mechanisms that govern interdecadal variability of the THC highlights the need for further studies to understand these mechanisms. In this paper we analyze the mechanism responsible for interdecadal variability of the THC in the third version of the Hadley Centre global coupled atmosphere–ocean sea-ice general circulation model (HadCM3). Our study is complementary to those of Vellinga and Wu (2004) and Knight et al. (2005, manuscript submitted to Nature, hereafter KAFVM), which focus on centennial time scale variability of the THC and its climatic impact. One attractive feature of the HadCM3 model is that, unlike the models analyzed by Delworth et al. (1993) and Timmermann et al. (1998), HadCM3 requires no flux adjustments. Such adjustments could potentially impact aspects of climate variability, and so are undesirable. A further advantage of HadCM3 is that the resolution in both the atmosphere and ocean components is higher than in these earlier studies.

The paper is organized as follows. In section 2, the coupled model is briefly described. The characteristics of, and mechanism responsible for, interdecadal variability of the Atlantic THC in HadCM3 are investigated in section 3. Section 4 discusses the relationship to previous studies and addresses the climate impacts of the THC variability. Conclusions are in section 5.

2. Coupled model

a. Model

The model we use is a version of the United Kingdom (UK) HadCM3, which is described in Gordon et al. (2000). The atmospheric model component in HadCM3 is a version of the Met Office unified forecast and climate model run with a horizontal grid spacing of 2.5° × 3.75° and 19 vertical levels using a hybrid vertical coordinate. The model uses a radiation scheme developed by Edwards and Slingo (1996), and the land surface scheme of Cox et al. (1999). The detailed description of the model formulation and their performance in a simulation forced by observed SSTs are described in Pope et al. (2000).

The oceanic component of the model is a 20-level version of the Cox (1984) model on a 1.25° × 1.25° latitude–longitude grid. The vertical levels are distributed to provide enhanced resolution near the ocean surface. The two components are coupled once a day. The model does not require flux corrections to maintain a stable climate. The mean climate and its stability in a 1000-yr control simulation are discussed in Gordon et al. (2000). In this paper, the interdecadal fluctuations in the THC, their impact on ocean heat content, sea surface temperature, and climate are investigated from this 1000-yr control simulation.

b. Climatology

Shown in Figs. 1a,b are the climatological annual mean sea surface temperature, surface salinity, and wind stress from observations and the model simulation. The observational SST and salinity are based on the Levitus climatology (Levitus 1982) and the wind stress is based on the Southampton Oceanography Centre (SOC) climatology (Josey et al. 2002). Although the model climatology is similar to that observed, there are a few disagreements. The most conspicuous disagreement is that the model-simulated SST front over the Gulf Stream is weaker than is observed. Both the zonal and meridional wind stresses are weaker than observed in midlatitudes. The associated underestimate of the
wind stress magnitude is also seen in the third version of the Hadley Centre uncoupled atmospheric model (HadAM3), and discussed in Pope et al. (2000).

The simulated mean meridional overturning circulation (MOC) in the North Atlantic is given in Fig. 1c, where positive values stand for an anticlockwise circulation. It reveals a maximum value of about 19 Sv ($10^6$ m$^3$ s$^{-1}$). The MOC shows an outflow of North Atlantic Deep Water (NADW) of about 14 Sv at the equator and an inflow of Antarctic Bottom Water (ABW) of 6 Sv into the North Atlantic. The sinking occurs in a broad region between 50° and 70°N, down
to a depth of \(\sim 2500\) m with strong sinking taking place primarily at about 65°N. The mean barotropic transport streamfunction is shown in Fig. 1d. The North Atlantic subtropical gyre has a strength of 30 Sv and the subpolar gyre of about 15 Sv. The coupled model reproduces the principal features of the North Atlantic THC and gyre circulations in a satisfying manner. The model climatology of mixed layer depth in March is shown in Fig. 1e. The dominant convective region occurs over the northeast North Atlantic, in the Greenland–Iceland–Norwegian (GIN) and Irminger Seas, where the mixed layer depth reaches 200–400 m. The model captures the observed separation of the deep water source by the Greenland–Iceland–Scotland ridge (Wood et al. 1999). The variability of the mixed layer depth is also large over the same region. The interannual standard deviation of the mixed layer depth is in a range of 150–250 m (Fig. 1f). In addition, in the Labrador Sea there is also a local maximum of the variability, even though the climatological mixed layer depth is relatively shallow (\(\sim 100\) m).

3. Mechanism of interdecadal variability in Atlantic THC/OHT

In this section, the characteristics of interdecadal variability of Atlantic THC and associated OHT are investigated. The physical mechanisms responsible for the variability are highlighted.

a. MOC index and its variability

In modeling studies the THC index is often defined as the maximum of the MOC in the North Atlantic (e.g., Delworth et al. 1993; Timmermann et al. 1998; Delworth and Greatbatch 2000). However, the latitude at which this maximum occurs may change with time, making the physical interpretation of the variability more difficult. Here we define an index as the mean value of MOC at the depth of 996 m in the latitude band 25.5°–32.5°N (where the OHT reaches a maximum). The power spectrum of this index is shown in Fig. 2. There is notable variability on interdecadal time scales with significant power at a time scale of about 100 yr and a marginally significant peak at time scale of about 25 yr. The physical mechanisms of the centennial variability and its climatic impact are investigated in Vellinga and Wu (2004) and KAFVM. In this paper, we focus on the variability on interdecadal time scales. To focus on these time scales the data are bandpass filtered using a Chebyshev recursive filter (Cappellini 1978) to retain the variability ranging from 8 to 65 yr to remove the variability associated with interannual and centennial time scales.

b. The interdecadal mode of Atlantic THC/OHT

Figure 3 shows the first empirical orthogonal function (EOF1) of the bandpass-filtered OHT and associated first principal component (PC1). This mode explains 58.4% of the variance in the bandpass-filtered OHT. The heat transport mode reaches peak amplitude around 35°N—where the meridional temperature gradients are greatest—decaying rapidly to the north of this latitude and more gradually to the south. This pattern is not unexpected. As shown in Dong and Sutton (2002b) in a 100-yr simulation with the same model (but with suppressed air–sea interaction in the tropical Pacific and Indian Oceans), variability in the northward transport of heat in the Atlantic is primarily governed by fluctuations of the advection of mean temperature by anomalous circulation rather than the fluctuations of the advection of anomalous temperature by mean circulation (i.e., variability in the OHT in the Atlantic is primarily governed by variability in the ocean circulation rather than variability in temperatures). The magnitude of the anomalous OHT peaks at around 0.025 PW (corresponding to a fluctuation of one standard deviation of PC1).

The EOF1 of bandpass-filtered MOC is shown in Fig. 3b, which explains 34.1% of the retained variance. It is characterized by a single cell extending south from 65°N and reaching well across the equator. It has a maximum value of about 0.7 Sv (for a fluctuation of one standard deviation of PC1) with the strongest sinking taking place primarily at about 65°N. The structure of this mode shows some similarities to the time-mean MOC shown in Fig. 1c, but note that the anomalous circulation is mainly confined to the North Atlantic and that anomalous cross-equator flow is very weak. This feature contrasts with the variability on centennial time scales, which is associated with significantly stronger cross-equator flow (Vellinga and Wu 2004 and KAFVM). Similar frequency dependence of the cross-equator flow is shown in the work of Johnson and Marshall (2002).

Figure 3c shows a comparison between the leading principal component of the MOC and the leading principal component of OHT. The time series are highly correlated with a correlation coefficient of 0.85, sug-
suggesting that low-frequency interdecadal variations of the dominant OHT mode are associated with the Atlantic MOC variability, in line with Dong and Sutton (2001). When the MOC index that was used to compute Fig. 2 was bandpass filtered, the resulting correlation with PC1 of the MOC was found to be 0.74. This high correlation is expected since the variability on interdecadal time scales is of the basin scale (Fig. 3b). Therefore, hereafter the PC1 of MOC is taken as our THC index.

Figure 3d shows the autocorrelations of PC1 of MOC and PC1 of OHT. The largest negative correlations are seen at the lag of about 12 yr, implying the dominant period of about 24 yr. The rapid decay with lag indicates that oscillations are heavily damped.

c. Interdecadal cycle of MOC

To examine the interdecadal variability of the THC, lead–lag regressions of the MOC on PC1 of the MOC have been calculated and are shown in Fig. 4. At the
lag of \(-12\) yr, the THC is in its weak phase with a clockwise anomalous circulation, associated with reduced northward warm surface flow and reduced southward deep flow with weaker downwelling at 65°N. Six years later, at the lag of \(-6\) yr, positive anomalies of meridional overturning circulation develop in the high-latitude North Atlantic, suggesting enhanced deep water formation. At lag 0, the THC reaches a maximum with enhanced northward warm surface flow, stronger southward deep returning flow, and enhanced downwelling at 65°N. At the lag of 6 yr, the positive anomalies of meridional circulation weaken and negative anomalies appear at high latitudes. At the lag of 12 yr, the pattern of MOC is similar to the pattern at the lag of \(-12\) yr, and one cycle is complete. Note that the amplitude of the anomalies at either lag \(-12\) or lag 12 is only about one-third that at lag 0, confirming the previous inference that the fluctuation is strongly damped.

d. Dependence of THC fluctuation on the mixing over the Nordic Seas

It is well established that fluctuations of the deep water formation rate drive the fluctuations in THC (e.g., Rahmstorf 1995; Mauritzen and Håkkinen 1999). There are also related fluctuations in ocean density. The association between ocean density and the overturning circulation is illustrated in Figs. 5a,b, which shows the regression coefficients of the mixed layer depth in March and annual mean upper-ocean density onto PC1 of the MOC when the THC is lagging by 4 yr. This is the time when the regression coefficients are the largest (Fig. 5c). The large positive mixed layer depth anomalies (Fig. 5a) of about 40 m over the GIN Sea and about 20 m over the Irminger Sea suggest that convection at these sites is actively involved in the interdecadal variability of the THC. The density regression (Fig. 5b) indicates that a maximum in the MOC is associated with positive density anomalies over the GIN Sea and the Irminger Sea 4 yr previously. At this time, mixed layer depth and density anomalies over the Labrador Sea are small, suggesting that Labrador Sea convection has a lesser role in THC variability on the time scale considered here. Both the mixed layer depth and density regression patterns show similar characteristics when they lead the MOC by 6 to 3 yr.

To further illustrate the relationships between convective mixing and THC fluctuation, two mixing indices are defined. One is the GIN Sea mixing index, defined as the mixed layer depth over the region (60°–80°N, 20°W–15°E). The second is the Irminger Sea index, defined as the mixed layer depth over the region

![Fig. 4. Regression coefficients (Sv per unit of PC1) of MOC at various lags to PC1 of MOC. Positive lags mean PC1 is leading.](http://journals.ametsoc.org/jcli/article-pdf/18/8/1117/3796547/jcli3328_1.pdf)
The lead–lag correlations between these two mixing indices and PC1 of the MOC give maximum values of 0.45 for the Irminger Sea index at a lead of 5 yr, and 0.40 for the GIN Sea index at a lead of 3 yr. The correlations of upper-ocean density indices over the same two regions with the PC1 of MOC give a maximum value of 0.45 and 0.46, respectively, when density leads by 5 and 3 yr, respectively. Consequently we define new mixed layer depth and density indices by averaging over the larger region (50°–80°N, 45°W–15°E), which includes both the GIN and the Irminger Seas. The lead–lag correlation coefficients between these new indices and PC1 of the MOC are shown in Fig. 5c. The largest positive correlation (0.44 and 0.49) occurs when both indices lead the THC by 4 yr.

Figure 5d shows the vertical profile of density regression coefficients over the deep convection region (i.e., over the GIN and Irminger Seas). Positive density anomalies extend from the surface to about 1000 m with weak negative density anomalies from about 1000 to 2500 m, thereby reducing the vertical stability of the water column, enhancing the deep convection rate, and driving a stronger THC with a time lag of 3–6 yr. This lag is the time for the basin-scale MOC to adjust to a new density distribution in deep convection region, and is consistent with the adjustment time scale found in other studies (e.g., Eden and Willebrand 2001; Dong and Sutton 2002a; Bentsen et al. 2004; Cheng et al. 2004). The adjustment process involves the excitation of a coastally trapped Kelvin wave that propagates southward along the western boundary, eastward along the equator, and then poleward along the eastern boundary in both hemispheres. The poleward propagation leads to the excitation of Rossby waves that carry the perturbation signal into the ocean interior (e.g., Kawase 1987; Doscher et al. 1994). Because of the important role for boundary waves the adjustment time scale may be sensitive to the ocean model resolution (Doscher et al. 1994).
e. Relative role of temperature and salinity variability in driving THC fluctuations

In the previous section, it has been shown that the THC fluctuation is driven by density variations in the deep convection region. To understand the relative importance of thermal versus haline processes for the THC fluctuations, density components due to variable temperature and salinity are computed. Figure 6 shows lead–lag correlations and regressions of the total density, the density due to temperature, and the density due to salinity in the upper 200 m in the deep convection region with PC1 of the MOC. The results indicate that density fluctuations attributed to salinity anomalies lead the THC variability by about 4 yr. In contrast, temperature fluctuations make their most important contribution to density variations at positive lags, with the negative correlation peaking at a lag of +5 yr. The sign of the correlations is such that temperature variations are acting to damp the density anomalies that are initially established by salinity variations. Hence the results suggest that salinity-forced density changes are largely responsible for the changes in the stability of the water column in the deep convection region that lead to enhanced deep water formation and interdecadal variability of the THC. A similar conclusion was reached by Delworth et al. (1993) and Timmermann et al. (1998) based on analysis of very different models.

f. Relative importance of salinity advection versus surface fluxes for THC variability

Variations in salinity in the deep convection region may arise from variations in the transport of salinity in the ocean or from variations in the surface freshwater flux. To identify the dominant processes responsible for the fluctuations of salinity in the deep convection region, we have performed salinity budget analysis in a similar way to Delworth et al. (1993). The volume mean regression coefficients of salinity budget on the THC index are shown in Fig. 7. It shows that positive advection of salinity leads the THC by 5–6 yr with surface fluxes and convection playing a weak damping role. Figure 7 indicates that anomalous transport of salinity into the deep convection region results in anomalous salinity. Anomalous salinity, in turn, drives the THC variability (as shown in Fig. 6). The fact that the transport of salinity into the deep convection region leads the THC by 5–6 yr indicates that the transport fluctuations are not driven by the THC fluctuation. The processes responsible for the fluctuation of salinity transport will therefore be discussed next.

g. Oceanic conditions associated with THC variability

Shown in Fig. 8 are oceanic conditions at the lag of −6 yr (THC is lagging by 6 yr), 0 and +6 yr (THC is leading by 6 yr). At the lag of −6 yr, there are negative OHC anomalies over most of the North Atlantic and positive salinity anomalies in the northeast North Atlantic. Associated with negative OHC and positive salinity anomalies are positive density anomalies (Fig. 8c). Note that whereas in the deep convection region at this lag density variations are dominated by variations in salinity (Fig. 6), over the subpolar gyre as a whole both temperature and salinity variations contribute to the variations in density. The density anomalies are as-
The variations in the strength of the subpolar gyre, which peak prior to an extreme in the MOC, have the most effect on the salinity in the deep convection region (as shown in Fig. 6), whereas the variations in the MOC itself have the most effect on temperature in this region (as measured in terms of the impact on density). A second factor is that temperature and salinity anomalies are affected differently by air–sea interactions. In particular, surface temperature anomalies are actively damped whereas salinity anomalies are not. Figure 9, to be discussed shortly, shows the damping of temperature anomalies in the deep convection region.

h. The role of the atmosphere in subpolar gyre variability

In the previous section, it has been shown that density variations in the regions of active deep convection are related to variability in the subpolar gyre. In this section the role of the atmosphere in driving the subpolar gyre and related aspects of the THC variability is investigated. Shown in Fig. 9 are the regressions of annual mean sea level pressure (MSLP) and surface heat flux at lags $-6, -3, 0$ (the atmosphere is leading) to the THC index. Similar regressions using seasonal mean data suggest that the pattern seen in Fig. 9 is predominantly due to the atmospheric fluctuation in the northern winter and spring (not shown). At the lag of $-6$ yr, the atmospheric circulation anomalies are characterized by an anomalous cyclonic circulation at high latitudes and an anticyclonic circulation in midlatitudes. This circulation pattern projects onto the positive phase of the NAO. The wind stress curl anomalies associated with the anomalous cyclonic circulation will induce Ekman upwelling, tending to raise the thermocline, cool the upper ocean, and drive a stronger subpolar gyre (as seen in Fig. 8). At the same time negative surface heat flux anomalies are also acting to cool the ocean surface in the subpolar gyre.

At lags of $-3$ and $0$ the center of anomalous cyclonic circulation is located farther northeast over the GIN Seas region of active deep convection. This pattern suggests that, at this stage, the atmospheric circulation acts directly to favor (or precondition) the development of deep convection through the impact of enhanced Ekman upwelling on the stability of the water column.
FIG. 8. Linear regression coefficients of various oceanic variables, averaged in the upper 500 m, onto PC1 of MOC. (Left) PC1 is lagging by 6 yr, (middle) simultaneously, and (right) PC1 is leading by 6 yr. (top) OHC (°C), (upper middle) salinity (psu), (lower middle) density (kg m⁻³), and bottom (d) upper-ocean current (cm s⁻¹). Shading scale indicates regressions are 95% significant using the Student’s t test.
At the same time anomalous heat fluxes are acting to cool the ocean in the GIN Seas region, also favoring the development of deep convection. Thus, Fig. 9 suggests that the atmosphere acts to drive the THC variability both indirectly, by modulating the strength of the subpolar gyre, and hence the advection of saline waters into the region of active deep convection and more directly, by forcing the convective regions themselves (both through wind stress curl and heat fluxes).

To further investigate the relationship between the THC variability and variability in the subpolar gyre, a subpolar density index and an NAC index have been defined as area-averaged density over the subpolar region \((40°-60°N, 70°W-0°W)\), and area-averaged zonal current over the Gulf Stream region \((35°-45°N, 70°-20°W)\), respectively. Figure 10a shows the lead–lag correlation coefficients between the THC index, the subpolar gyre density index, and the NAC index. Positive correlations when the density index and NAC index are leading indicates that density variations over the subpolar gyre region lead THC variations by about 7 yr (consistent with Fig. 8), while variations in the NAC index lead THC variations by about 4 yr. A notable feature of the NAC correlations is the strong asymmetry with respect to lag 0. This strong asymmetry suggests that the impact of NAC variability on THC variability is stronger than the impact of the THC on the NAC.

Figure 10b shows lead–lag correlations between the NAC index and indices of density, wind stress curl, and surface heat flux averaged over the subpolar gyre region. It shows that positive density anomalies and anomalous cyclonic circulation in the atmosphere over the subpolar region lead to a stronger NAC with a time lag of 2–3 yr. As already seen in Fig. 9 surface heat fluxes act to cool the subpolar gyre at the same time as wind stress curl anomalies act to induce Ekman upwelling and gyre acceleration. The relationship identified here in the coupled model between the wind stress curl and NAC is supported by the observational study of Curry and McCartney (2001). They showed that the transport associated with the Gulf Stream and NAC gradually weakened during the low NAO period of the 1960s and then intensified in the subsequent period of high NAO with the oceanic index lagging the atmospheric index by 1–2 yr. The results are also consistent with the OGCM study in Eden and Willebrand (2001), which showed enhanced subpolar gyre circulation 2–3 yr after the positive phase of NAO. They further

![Fig. 9. Linear regression coefficients of (top) annual mean MSLP (hPa) and (bottom) surface heat flux (W m\(^{-2}\)) onto PC1 of MOC. (left) PC1 is lagging by 6 yr, (middle) PC1 is lagging by 3 yr, and (right) simultaneously. Sign convention is that downward heat flux is positive. Shading scale indicates regressions are 95% significant using the Student’s t test.](image-url)
showed that the surface heat flux and wind stress curl associated with the NAO play a roughly equal role.

Further confirmation of the importance of atmospheric variability in driving variations of the THC is provided by a cross-spectral analysis between a NAO index (unfiltered) derived from the model and the (also unfiltered) MOC index in Fig. 2. The NAO index is taken as the leading principal component of December–January–February (DJF) MSLP. The spatial pattern, shown in Fig. 11a, has the expected dipole structure with centers of action near Iceland and the Azores.

![Figure 10: Lead-lag correlations of PC1 of MOC with density over subpolar region and NAC strength. (b) Lead-lag correlation of NAC strength with density, wind stress curl, and surface heat flux over the subpolar region (40°–60°N, 70°W–0°).](image)

![Figure 11: (a) EOF1 of DJF MSLP, (b) squared coherence, and (c) the phase (unit in °) spectrum between the THC index and PC1 of MSLP. Positive phase angles indicate that PC1 is leading THC. This mode explains 41.4% of the total variance. Figure 11b shows that there is notable coherence on inter-decadal time scales between the NAO index and the MOC index, with a peak at about 26 yr. Figure 11c...](image)
shows that, at this period, the atmosphere leads the ocean by about 4–5 yr, corresponding to a phase angle of \(\sim 60^\circ\). This result is consistent with the role of the atmosphere inferred from Fig. 9.

4. Discussion

a. Summary of mechanism for interdecadal variability in the THC

Based on the results shown in the previous sections, a schematic diagram of the mechanism responsible for the phase reversal of the THC on decadal time scales, and the role of the atmosphere in driving this interdecadal THC variability, is summarized in Fig. 12. We describe first the ocean processes before discussing the atmospheric driving.

Starting from a minimum in the THC, negative OHC anomalies begin to build up north of the Gulf Stream region due to anomalous OHT divergence. After about 5–6 yr, the negative OHC anomalies reach a maximum. Associated with the buildup of negative OHC anomalies are positive anomalies in the upper-ocean density and an acceleration of the subpolar gyre and NAC. The stronger subpolar gyre increases the transport of salinity into the region of active deep convection and leads to a maximum in the upper-ocean density in this region. Enhanced convection is triggered, leading to an increase in the rate of deep water formation, and acceleration of the THC. The THC reaches a maximum approximately 4 yr after the maximum of upper-ocean density in the region of deep convection (Fig. 6a) and approximately 6 yr after the maximum in the salinity transport into this region (Fig. 7). The total time for the phase reversal is 12–14 yr, consistent with a period of about 24–28 yr. The above summary makes it clear that the mechanism for phase reversal relies on interactions between the subpolar gyre circulation and meridional overturning circulation. Figure 13 is a further schematic that summarizes these interactions.

The above sequence of events makes no reference to the atmosphere and might plausibly be considered an
The mechanism we have identified in the HadCM3 model has many similarities to that described by Delworth et al. (1993), and further discussed by Delworth and Greatbatch (2000), in the GFDL model. In particular, the mechanism for phase reversal summarized in Fig. 12 is essentially the same. This consistency suggests the mechanism may be robust since the two models differ in many respects, in particular the HadCM3 model has significantly higher resolution in both its ocean and atmosphere components. There are, however, some differences. First, the time scale we find is shorter than that in the GFDL model. This difference is mainly due to the different time scales for anomalies in the upper OHC to accumulate in the subpolar gyre, following a maximum or minimum in the THC. This time scale is 5–6 yr in our model while it is 12–14 yr in the GFDL model, and corresponds to one-quarter of the oscillation period. The shorter accumulation time scale in our model may be due to the higher resolution of the ocean model, which simulates stronger currents and stronger meridional temperature gradients.

Further differences between the mechanism in the HadCM3 model and that in the GFDL model are associated with the role of the atmosphere. We agree with Delworth and Greatbatch (2000) that the THC variability is forced by the atmosphere rather than a coupled phenomenon, but with regard to the spatial structure of the forcing we have drawn a distinction (e.g., in Fig. 12) between 1) forcing of the subpolar gyre circulation, which has subsequent impacts on the convectively active region, and 2) direct forcing of the convectively active region. No such distinction was made in Delworth et al. (1993) or Delworth and Greatbatch (2000), probably because in the lower-resolution GFDL model the location of the convectively active region is broader and difficult to clearly distinguish from the location of the subpolar gyre. Figure 9 in this paper suggests, however, that such a distinction is appropriate.

In this study we have not attempted to achieve a quantitative separation between the role of anomalous momentum fluxes (i.e., wind stress curl), heat fluxes, and freshwater fluxes in driving the THC variability.
Because the fluctuations in these quantities are correlated such a separation would require additional uncoupled experiments such as were performed by Delworth and Greatbatch (2000). The conclusions of Delworth and Greatbatch (2000) and also Eden and Jung (2001) was that variability in the heat fluxes is the most important driver of interdecadal variability in the THC. This may be the case, but we think it appropriate to highlight the fact that a significant role for wind stress curl anomalies in preconditioning (via Ekman upwelling) the convectively active region is strongly suggested by our model results (Figs. 9c,e) and may have been overlooked in previous studies either because of the low ocean resolution (in the case of the GFDL model) or because attention was focused solely on the response to NAO changes (in the case of Eden and Jung 2001). We note that whereas the atmospheric pattern that forces the subpolar gyre (Fig. 9a) does project strongly onto the NAO, the atmospheric pattern that forces the convective regions does not.

c. Impact of THC fluctuations on climate

It has long been suspected that changes in the THC may be an important factor in climate fluctuations (e.g., Bjerknes 1964; Delworth et al. 1993; Kushnir 1994; Timmermann et al. 1998; Delworth and Mann 2000). Thus far we have studied the mechanism of interdecadal THC variability in HadCM3 but have yet to consider whether this variability has significant climate impacts. Note that such impacts may exist irrespective of whether ocean–atmosphere coupling plays any significant role in the mechanism of the variability.

Climate impacts of fluctuations in the THC are likely to be mediated through changes in sea surface temperature (SST). In Fig. 14 we therefore examine, again using regression analysis, the evolution of SST following a maximum in the THC, and compare it to the evolution of upper OHC, which we have already related to changes in the THC. Note that it is not obvious that variations in SST will be consistently related to fluctuations in the THC, since SST changes may be governed more strongly by shallow processes confined to the mixed layer rather than larger-scale, deep-reaching, ocean dynamics. It is well known, for example, that most of the variability in SSTs on interannual time scales is a response to variations in surface heat fluxes and near-surface Ekman currents. However, Fig. 14 shows that, on interdecadal time scales, there are in fact significant changes in SST related to THC variability.

At lag 0 there are significant SST anomalies over most of the North Atlantic, and the pattern of anomalies at mid- and high latitudes is very similar to that seen in OHC. In the Tropics, however, there is an additional feature that is not prominent in the OHC field: a dipolar pattern of SST anomalies straddles the equator with positive anomalies in the Northern Hemisphere and weaker negative anomalies in the Southern Hemisphere. Closer inspection of the OHC field shows that the negative anomalies are seen south of the equator in the eastern tropical Atlantic but there are no significant positive anomalies to the north of the equator. The appearance of this SST dipole is notable because 1) other studies of the response to THC changes (e.g., Yang 1999; Dong and Sutton, 2002a; Vellinga and Wood 2002) have found a similar feature in the SST, and 2) the atmosphere is known to be very sensitive to fluctuations in the cross-equatorial SST gradient in the Atlantic (e.g., Moura and Shukla 1981; Sutton et al. 2000; Folland et al. 2001).

Figure 14 shows that, as time advances, the positive OHC anomalies in the North Atlantic amplify and expand to fill the whole subpolar gyre, as was already seen in Fig. 8. The SST anomalies in the North Atlantic largely follow the evolution of OHC suggesting that the SST changes are primarily controlled by changes in the large-scale ocean circulation related to the THC. There are, however, one or two differences, most notably at lag 6 in the region of the GIN Seas. Here we find that positive OHC anomalies are associated with (marginally significant) negative SST anomalies. This suggests that anomalous surface fluxes (which can be seen in Fig. 9f) have been effective at damping the temperature anomalies at the surface while having relatively little impact on the heat content integrated over the upper ocean. Note finally that the SST dipole in the tropical Atlantic is present but weaker at lag +3 and absent at lags +6 and +9.

To investigate whether there is a significant atmospheric response to the variability in the THC we regressed sea level pressure and precipitation on our THC index (PC1 of the MOC) and varied the lag between −9 and +9 yr. At negative lags we see the role of the atmosphere in driving the THC fluctuations as discussed in the previous section with significant anomalies confined in the Atlantic and Eurasian sector (not shown). At lags of +6 and +9, atmospheric anomalies are weak and marginally significant, although over the North Atlantic the same patterns as were seen in Fig. 9, but with the opposite phase, are present as we would expect. The strongest indications of an atmospheric response are found at lag 0 and, more weakly, at lag 3. Figure 15 shows the patterns of SST, MSLP, and precipitation at lag 0. The interpretation of this figure must be approached with caution because significant anomalies may be associated either with the ocean forcing the atmosphere, or with the atmosphere forcing the ocean, or both. In the extratropics the likelihood is that the atmosphere is forcing the ocean, which has already been discussed. In the Tropics, however, we argue that the reverse is most likely to be the case. First, other studies (Yang 1999; Dong and Sutton 2002a; Vellinga and Wood 2002) have demonstrated that fluctuations in the tropical Atlantic SST dipole do arise as a response to changes in the THC. Second, Fig. 15 shows that the SST dipole over the tropical Atlantic is mirrored in a dipolar pattern of precipitation, indicating a northward
shift of the ITCZ, which many studies (e.g., Moura and Shukla 1981; Sutton et al. 2000; Folland et al. 2001) have shown to be a robust response of the atmosphere to variations in the cross-equatorial SST gradient. Third, if the significant anomalies in MSLP and precipitation that are seen throughout the Tropics in Fig. 15 are not forced by the ocean then there is no reason for them to be consistently associated with the variations in the MOC. The Pacific pattern seen in Fig. 15 resembles the interdecadal Pacific oscillation (IPO) discussed, for example, by Power et al. (1999). It therefore raises the interesting possibility that fluctuations in the THC could force the IPO. Alternatively, Fig. 15 could reflect an impact of the IPO on the Atlantic; however, this is unlikely since significant atmosphere and SST anomalies at negative lags are confined in the Atlantic and Eurasian sector (not shown).

The Tropicwide nature of the atmospheric anomalies seen in Fig. 15 is striking. It suggests that interdecadal variations in the Atlantic THC have the potential to modulate climate globally. Dong and Sutton (2002a) argued for the same possibility (based on different evidence). The most likely mechanism via which the Atlantic changes can impact atmospheric circulation over the other ocean basins is via propagation of equatorially trapped Rossby and Kelvin waves. Such waves will be excited in response to the anomalous diabatic heating associated with the displaced ITCZ in the tropical Atlantic.

5. Conclusions

It has been shown that an irregular interdecadal oscillation of the thermohaline circulation (THC) in the Atlantic Ocean exists in HadCM3, a coupled GCM without flux adjustment. The variability is primarily forced by the atmosphere, rather than being a coupled phenomenon, but the ocean is responsible for setting the time scale. The mode is heavily damped. Following a minimum in the THC the mechanism for phase reversal involves: 1) the accumulation of cold water in the subpolar gyre, leading to an acceleration of the gyre circulation including the North Atlantic Current, 2) enhanced advection of salinity into the GIN and Irminger Seas, leading to an increase in the upper-ocean density in this region, 3) enhanced convection in the GIN and Irminger Seas, leading to an increase in the rate of deep water formation, and acceleration of the THC. The total time for the phase reversal is 12–14 yr, consistent with a period of about 24–28 yr.

Excitation of the THC variability by the atmosphere has two elements: first, forcing of the subpolar gyre by wind stress curl and heat flux anomalies, and second, direct forcing of the region of active deep convection (especially the GIN Seas region) also by wind stress curl (which modifies convective stability through modulation of Ekman upwelling) and heat flux anomalies. The first forcing is associated with an NAO-like pattern of atmospheric circulation while the second is associated with a different pattern that features MSLP anomalies over the GIN Seas.

The fluctuations in the THC cause SST anomalies in both the tropical (0.1°–0.2°C) and North (0.8°–1.0°C)
Atlantic and modulate the cross-equatorial SST gradient in the tropical Atlantic. The climate impacts are certainly modest but the tropical atmosphere in particular is very sensitive to small changes in SST. The atmospheric response involves a displacement of the ITCZ in the Atlantic basin but also climate anomalies throughout the Tropics. The results from this study therefore suggest that interdecadal variations in the Atlantic THC have the potential to modulate the climate globally. The most likely mechanism via which the Atlantic changes can impact atmosphere circulation over the other ocean basins is via the propagation of equatorially trapped Rossby and Kelvin waves, in response to the anomalous diabatic heating associated with the displaced ITCZ. Last, we note there is some evidence (e.g., Johannsen et al. 2004) that in the real world THC variability may have larger impacts on both SST and climate than is suggested by the HadCM3 model.

Perhaps surprisingly, the mechanism found in the HadCM3 model has many similarities to that found in the GFDL model by Delworth et al. (1993). The HadCM3 model has significantly higher resolution in both its atmosphere and, especially, the ocean (1.25° versus approximately 4°) component. The fact that a similar mechanism arises in both models suggests that it may be quite robust. It also highlights the need for more research to examine this robustness further. One question that needs to be answered is the following: is a similar mechanism seen in other models, especially models with still higher resolution?

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