Clim Dyn (2011) 36:1255–1276
DOI 10.1007/s00382-010-0958-0

Mechanisms of the atmospheric response to North Atlantic multidecadal variability: a model study

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Received: 7 August 2009 / Accepted: 24 November 2010 / Published online: 16 December 2010

Abstract The atmospheric circulation response to decadal fluctuations of the Atlantic meridional overturning circulation (MOC) in the IPSL climate model is investigated using the associated sea surface temperature signature. A SST anomaly is prescribed in sensitivity experiments with the atmospheric component of the IPSL model coupled to a slab ocean. The prescribed SST anomaly in the North Atlantic is the surface signature of the MOC influence on the atmosphere detected in the coupled simulation. It follows a maximum of the MOC by a few years and resembles the model Atlantic multidecadal oscillation. It is mainly characterized by a warming of the North Atlantic south of Iceland, and a cooling of the Nordic Seas. There are substantial seasonal variations in the geopotential height response to the prescribed SST anomaly, with an East Atlantic Pattern-like response in summer and a North Atlantic oscillation-like signal in winter. In summer, the response of the atmosphere is global in scale, resembling the climatic impact detected in the coupled simulation, albeit with a weaker amplitude. The zonally asymmetric or eddy part of the response is characterized by a trough over warm SST associated with changes in the stationary waves. A diagnostic analysis with daily data emphasizes the role of transient-eddy forcing in shaping and maintaining the equilibrium response. We show that in response to an intensified MOC, the North Atlantic storm tracks are enhanced and shifted northward during summer, consistent with a strengthening of the westerlies. However the anomalous response is weak, which suggests a statistically significant but rather modest influence of the extratropical SST on the atmosphere. The winter response to the MOC-induced North Atlantic warming is an intensification of the subtropical jet and a southward shift of the Atlantic storm track activity, resulting in an equatorward shift of the polar jet. Although the SST anomaly is only prescribed in the Atlantic ocean, significant impacts are found globally, indicating that the Atlantic ocean can drive a large scale atmospheric variability at decadal timescales. The atmospheric response is highly non-linear in both seasons and is consistent with the strong interaction between transient eddies and the mean flow. This study emphasizes that decadal fluctuations of the MOC can affect the storm tracks in both seasons and lead to weak but significant dynamical changes in the atmosphere.

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1 Introduction

The Atlantic meridional overturning circulation (MOC) is a major component of the climate system as it is the main oceanic contributor to the northward heat transport. The MOC exhibits strong decadal to multidecadal fluctuations and part of this variability appears to be potentially predictable one to two decades ahead (Collins et al. 2006; Msadek et al. 2010). However, a better understanding of the interaction between the atmosphere and the MOC is needed to develop a decadal prediction system (Latif et al. 2006). The mechanisms of variability of the MOC at decadal and longer timescales remain largely unknown.

There are large discrepancies in the simulated MOC in the different coupled models. Some show clear multidecadal fluctuations that are largely believed to be driven by atmospheric forcing without any back influence of the ocean (Delworth et al. 1993; Jungclaus et al. 2005; Dong and Sutton 2005), while others suggest a two-way coupling between the MOC and the atmosphere (Timmermann et al. 1998; Danabasoglu 2008). This coupling would imply an important feedback of the MOC on the atmosphere at decadal timescales, but the existence of an MOC influence on climate remains a subject of controversy.

Although there is no long-term measurements of the MOC, a few observable fingerprints of the MOC variability at decadal timescales have been investigated. Multidecadal variations of SST in the North Atlantic have been detected in observations (Bjerknes 1964; Kushnir 1994; Delworth and Mann 2000), becoming known as the Atlantic multidecadal oscillation[1] (AMO, Kerr 2000). The SST anomaly pattern associated with the AMO is interhemispheric in scale with opposite sign in the northern and southern hemispheres, and the strongest signal in the North Atlantic. The SST anomaly associated with the AMO is thought to be the surface expression of the MOC and the communicator of deep ocean variability to the atmosphere (Latif et al. 2004; Knight et al. 2005; Pardaens et al. 2008). Recent studies based on multimodel analysis suggest that the AMO is an internal mode of variability rather an externally forced signal (Ting et al. 2009; Knight 2009). The AMO has been associated with a wide array of significant global impacts of high societal relevance like India and Sahel summer rainfall and Atlantic hurricane activity (Goldenberg et al. 2001; Zhang and Delworth 2006; Knight et al. 2006). The AMO has been used in several studies as a forcing pattern to investigate the atmospheric response to decadal oceanic fluctuations forced by the MOC. Using an ensemble of equilibrium simulations of the HadAM3 atmospheric model forced by the observed AMO, Sutton and Hodson (2007) investigated the climatic impacts of the basin-scale warming and cooling of the North Atlantic Ocean associated with a positive and a negative phase of the observed AMO, respectively. They found a clear signal in all seasons but the largest changes occurred during summer. Over the North Atlantic, a warm SST induced low pressure anomalies and enhanced precipitation. The climatic impacts were largest in the lower troposphere and extended further to the whole northern hemisphere. The tropical signature was rather strong and featured a northward shift of the ITCZ associated with enhanced precipitation in the North and reduced precipitation in the South. The consistency of these results was revealed by comparing the atmospheric response to the AMO in five different atmospheric general circulation models (AGCM, Hodson et al. 2009). The models showed a consistent picture during summer with cyclonic anomalies in response to a warmer North Atlantic, mainly centered over the tropics and reflecting a weaker subtropical anticyclone. During winter the atmospheric changes were weak and not consistent between models.

The atmospheric circulation response to the AMO identified by Sutton and Hodson (2007) might reflect an influence of the natural variability of the MOC, but only partly, because the AMO can also be influenced by direct atmospheric forcings that are not related to the MOC (e.g. ENSO teleconnections). Therefore, the atmospheric response to the MOC variability should be best investigated using an SST anomaly directly forced by MOC changes. The lack of MOC observations hardly allows to extract such an SST anomaly from observations. However, it can be defined from an ocean–atmosphere coupled model. Msadek and Frankignoul (2009, hereafter MF09) have shown that the natural variability of the Atlantic MOC has a significant influence on the atmosphere in the IPSL–CM4 coupled climate model. They detected a significant and robust signal mainly during summer, and showed that it reached maximum about 5 years after an intensification of the MOC. As shown by MF09, in the IPSL model, the atmospheric response to the MOC is hemispheric in scale and projects onto the East Atlantic pattern (EAP) over the North Atlantic, which is the second mode of atmospheric variability in this region. As the EAP was also shown to drive the MOC variability at decadal timescales in that model, this result suggested a weak positive feedback of the MOC on the atmosphere. Using lagged statistical relationships, a robust response was identified, however the signal-to-noise ratio was too low to determine the physical processes of the MOC influence. The SST anomaly associated with the atmospheric response, which lags the MOC by about 5 years, features a warming in the North Atlantic subpolar gyre and a cooling in the Nordic Seas. This SST anomaly resembles the model AMO and indeed the climatic impacts induced by the intensified MOC were quite similar to those following a positive phase of the AMO in the IPSL model. This is consistent with the link between the MOC and the AMO suggested in previous studies.

The SST anomaly associated with the MOC variability will be used in the present paper as forcing to investigate the atmospheric circulation response to the MOC. A common approach to analyze the atmospheric response to extratropical SST anomalies has been to specify the SST in so-called forced experiments. Although the results strongly diverge among the different models, the use of higher resolution models over the past decade combined with linear model studies allowed to identify a common

[1] Because of the short observed timeseries of SST changes, it is difficult to identify an oscillation and thus the AMO is also referred to as the Atlantic multidecadal variability (AMV).

mechanism of the atmospheric response. As reviewed by Kushnir et al. (2002), the equilibrium response depends upon the nature and strength of transient eddy feedbacks, in particular the eddy vorticity fluxes resulting from interactions between the anomalous flow directly forced by diabatic heating and by the storm tracks (Kushnir and Lau 1992; Ting and Peng 1995; Peng and Whitaker 1999; Peng et al. 2003). The initial response to anomalous diabatic heating is baroclinic, as predicted by linear theory (Hoskins and Karoly 1981). Then, the transient eddy feedback comes into play and generally transforms the direct baroclinic response into a growing barotropic one (Peng et al. 2003). Depending on the location of the SST forcing, the response may project onto the dominant modes of internal atmospheric variability, which are known to be mainly eddy driven (Branstator 1992). When the magnitude and sign of the forcing are varied, several models showed that the response scaled linearly with SST amplitude (Palmer and Sun 1985; Deser et al. 2004; Magnusdottir et al. 2004) and non-linearly with polarity (Kushnir and Lau 1992; Magnusdottir et al. 2004; Conil and Li 2005). Perpetual conditions for a specific month were often used to underline the high dependence on the background climatological flow. Peng et al. (1997) showed that the same SST anomaly can lead to a very different response in January and February, as the result of transient eddies interaction with a different atmospheric mean state.

Most of these sensitivity studies focused on the winter season where the transient eddies are the strongest, while there has been little emphasis on the summer response. However, the atmospheric response to the MOC identified by MF09, which projected on the EAP was strongest in summer. The largest AMO impacts detected by Sutton and Hodson (2005) in the observations, then by Sutton and Hodson (2007) in HadAM3, and by Hodson et al. (2009) in a serie of AGCMs, also occurred during summer. A significant atmospheric response in summer, but for an SST anomaly in the North Pacific has also been detected in the observations (Frankignoul and Sennechael 2007). Yet it is not clear whether the better detection of the response during this season is only due to a higher atmospheric signal-to-noise ratio as suggested by Sutton and Hodson (2007), or if an additional mechanism proper to the summer season is at stake. The fraction of marine stratiform clouds is greatest during summer, and could enhance the atmospheric response to SST anomalies through a positive radiative feedback (Norris et al. 1998; Park et al. 2005). The analysis of Peng and Whitaker (1999) also suggests that the summer atmospheric response might be related to the seasonal modulation of the transient eddy feedback by the mean atmospheric flow as during winter. Yet the dynamics of the summer atmospheric response to boundary forcing need to be explored further.

Several studies have assessed the limitation of AGCM experiments with specified SST anomalies (Barsugli and Battisti 1998; Sutton and Mathieu 2002). The experiments in which SST anomalies are not allowed to respond to changes in atmospheric circulation can produce misleading results, as they may underestimate the atmospheric variance and overestimate the heat flux feedback and thus the thermal damping. However, the model experiments of Lee et al. (2008) showed that whereas the enhancement of variability due to coupling might be an accurate interpretation in the subtropics, it might not be in the midlatitudes where the SST is predominantly driven by wind forcing. An alternative way to provide a more realistic assessment of the role of the ocean in climate variability is to use a model configuration in which the atmosphere is coupled to an ocean mixed layer, adding a mean flux correction to compensate for the dynamical oceanic processes that are not explicitly represented, like heat advection by the oceanic circulation. In the North Atlantic response studies of Drévillon et al. (2001) and Peng et al. (2005), the mixed layer was only forced by surface heat exchanges. Peng et al. (2006) included Ekman transport in the slab mixed layer ocean and showed that the equilibrium state was reached earlier and the response was closer to observations. A more realistic mixed layer model was developed by Alexander et al. (2000) and used by Cassou et al. (2007) to investigate the atmospheric response to a North Atlantic SST anomaly. It included vertical entrainment and allowed to represent the winter to winter persistence of the SST anomalies that are isolated below the shallow mixed layer during summer and are reentrained into the mixed layer the following winter.

In this paper we investigate the atmospheric circulation response to the North Atlantic SST anomaly associated with MOC fluctuations and described in MF09. For comparison with the atmospheric response detected in the IPSL climate model, we use its atmospheric component, and couple it to a slab ocean mixed layer model. The SST anomaly is that diagnosed from the coupled model, and we prescribe a flux correction to ensure that the slab model has the same SST climatology as the coupled model. Section 2 describes the model set up and the sensitivity experiments. The atmospheric response to the MOC-induced SST anomaly is analyzed for the summer season in Sect. 3 and for the winter season in Sect. 4. Discussion and conclusions are given in Sect. 5.

2 Model experiment

2.1 The model

The model consists of the atmospheric component of the IPSL coupled model, namely LMDZ (Hourdin et al. 2006), coupled to a slab mixed layer ocean. The coupling takes place in the North Atlantic, from 10°N to the climatological sea-ice boundary in winter, with 10°-latitude buffer zones. Elsewhere, the SST is prescribed from the coupled model climatology. The LMDZ AGCM has an horizontal resolution of about 3.75° by 2.5° in longitude and latitude directions, respectively, and 19 levels in the vertical. The slab mixed layer ocean is designed as an anomaly model in which the ocean temperature (T) is determined by the surface heat flux (Q) and the horizontal advection by the Ekman currents (U_E) as follows:

$$\frac{\partial T'}{\partial t} = \frac{Q'}{\rho C_p \bar{h}} - \mathbf{U}'_{\mathbf{E}} \cdot \nabla \bar{T} - \overline{\mathbf{U_E}} \cdot \nabla T' - \lambda T' \quad (1)$$

where overbars denote the daily climatology, and primes refer to the deviation from that climatology. $\mathbf{U_E}$ is the Ekman transport defined by $\mathbf{U_E} = \frac{\nabla \times \tau}{\rho f}$, τ is the wind stress and h the mixed layer depth, which is seasonally varying and prescribed from the control run of the IPSL coupled model. This ensures an SST variance close to that of the fully coupled model. Since entrainment and dissipation processes are not represented in the mixed layer model, we have limited the persistence of SST anomalies by the damping term $\lambda\ T'$, with λ taken equal to 6 months^{-1} consistent with the observational estimates of Frankignoul (1985). The daily climatology for Q and τ is based on the ensemble mean of 50 AGCM control runs forced by the SST climatology of the fully coupled model. The mixed layer temperature anomaly T' is calculated daily from Eq. 1, and then added to the SST climatology to drive the AGCM, which is broadly equivalent to the use of a flux correction in the traditional approach. The formulation is similar to the one described in Peng et al. (2006). The flux correction prevents temperature drift and partly compensates for the missing physics in the ocean such as heat transport by the geostrophic currents and diffusion, and to a lesser extent for model biases. Hereafter, the simplified coupled model is referred as the slab model and the coupled model as IPSL.

2.2 The control simulation

A 80-year control simulation was first performed with the slab model, producing a reasonable climatology in both ocean and atmosphere. In the region where the coupling is acting, the mean SST differs only slightly from the IPSL SST climatology (not shown). The model underestimates the SST by about 0.5°C following the North Atlantic current path, mainly due to the lack of geostrophic advection. Another bias reaching locally 1–1.5°C is found near the sea ice boundary region, probably in part because the climatological ice limit was derived from observations and is slightly different from that in IPSL (Marti et al. 2006). Nonetheless, the SST bias is small. In addition, it remains similar in the perturbed simulations and thus it will have negligible impacts on the differences between the control integration and the perturbed simulations.

The mean atmospheric circulation simulated by the slab model is illustrated for winter, when the variance is largest in the northern hemisphere. Figure 1 shows the mean 200 mb zonal wind, the mean 500 mb geopotential height (hereafter referred to as Z500) and its departure from the zonal mean for the slab model (left), IPSL (middle), and the NCEP reanalysis (Kalnay et al. 1996, right). These fields characterize the large-scale atmospheric circulation in the midlatitudes and their spatial structure must be correctly represented to get a realistic atmospheric response (Peng and Robinson 2001). The zonal winds at 200 mb are very similar in the slab model and in IPSL (Fig. 1a, b). Comparison with Fig. 1c indicates a fairly realistic representation of the North Atlantic jet, the East African jet and the Pacific jet, despite a slight overestimation of the Pacific jet in the slab model. The mean Z500 field shows a succession of troughs and ridges centered around 50°N, which are more visible when the zonal mean is removed. These stationary waves affect the strength and position of the jet and thus play a role in the atmospheric response to extratropical SST anomalies. They are well represented in the models, although the slab model tends to overestimate the ridge over the Rockies and to underestimate the jet diffluence over western Europe. The atmospheric patterns are also shifted southward in both models, which is linked to a systematic bias found in the AGCM (Hourdin et al. 2006; Marti et al. 2006). The spatial structures of the atmospheric internal variability are also well simulated in the slab model compared to IPSL and observations. The leading modes of atmospheric variability in the North Atlantic, namely the NAO and the EAP, are slightly shifted northeastward as in IPSL (MF09), but they are fairly close to observations (Fig. 2a, b). It can also be shown that the location and intensity of the storm tracks over the North Atlantic are realistic and compare well with those in IPSL described in Marti et al. (2010).

2.3 Sensitivity experiments

To identify the atmospheric response of the slab model to the MOC-associated SST anomaly, a 80-year simulation was performed with the SST anomaly added to the climatology in the North Atlantic. The model response is the mean difference between the perturbed and control integrations. Because of the weak atmospheric persistence, each year can be considered as independent so that each integration is considered as providing 80 independent realisations of the seasonal cycle. The statistical significance of the response is estimated by a Student's t test,

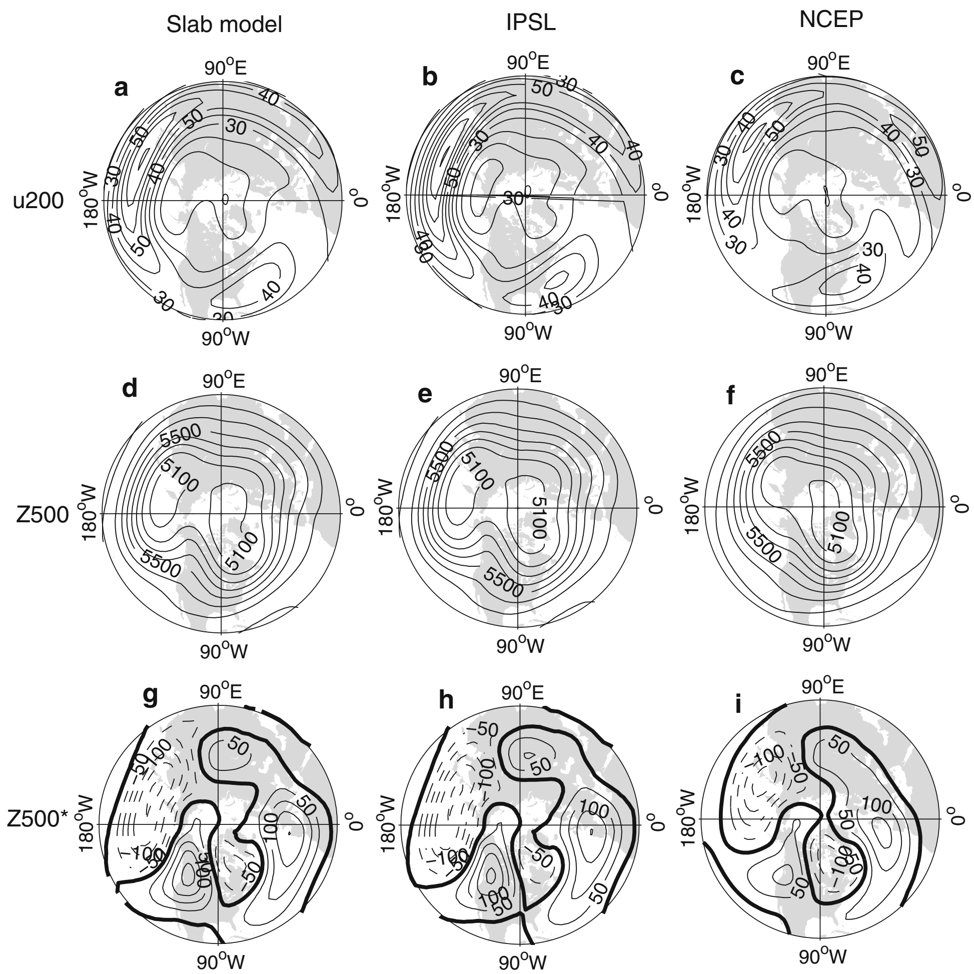


Fig. 1 Mean winter (DJF) atmospheric circulation in the slab model (*left*), in IPSL (*middle*), and in the NCEP–NCAR reanalyses (*right*). **a–c** Zonal mean wind at 200 mb (contour every 10 m/s). **d–f** Mean Z500 (contour every 100 m). **g–i** Eddy part of the Z500 field (contour every 50 m). The *heavy black line* is the zero contour

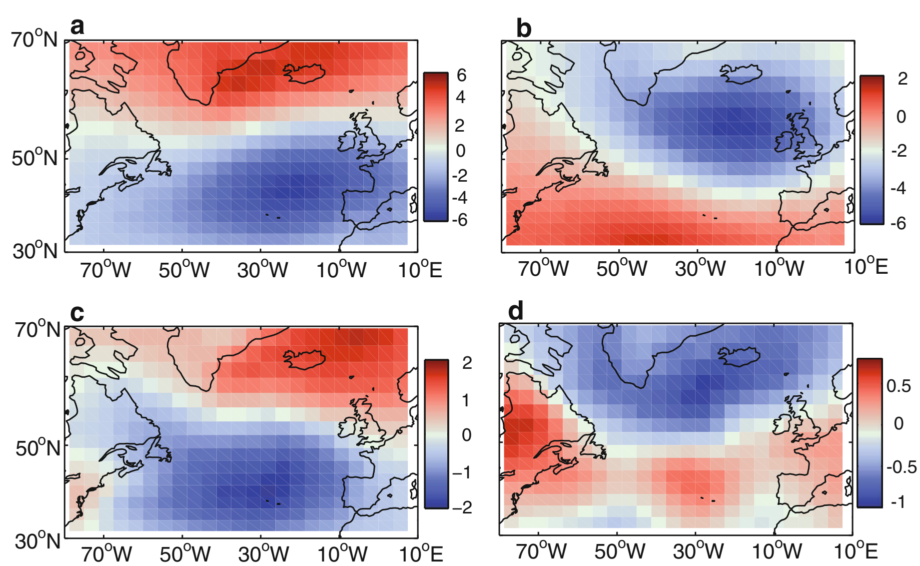


Fig. 2 Leading modes of atmospheric variability in the North Atlantic, namely the NAO (**a**) and the EAP (**b**), defined by the two leading EOFs of monthtly SLP in the control simulation. **c** North Atlantic SLP anomaly (in mb) in the slab model in winter and **d** in summer, in response to the SST anomaly in Fig. 3a, b, respectively. The statistical significance of the anomalies in **c** and **d** is not shown here but is given and discussed later in Figs. 4a and 13a

assuming similar variance in the anomalous and control runs. Six sensitivity experiments were performed. In the first two, we prescribed the two SST anomalies shown in Fig. 3 over the whole slab domain. Two additional experiments were performed with the same SST anomalies but reversing the sign in order to investigate the non-linearity of the atmospheric response. Finally two experiments were done with the SST anomalies in Fig. 3, but prescribing

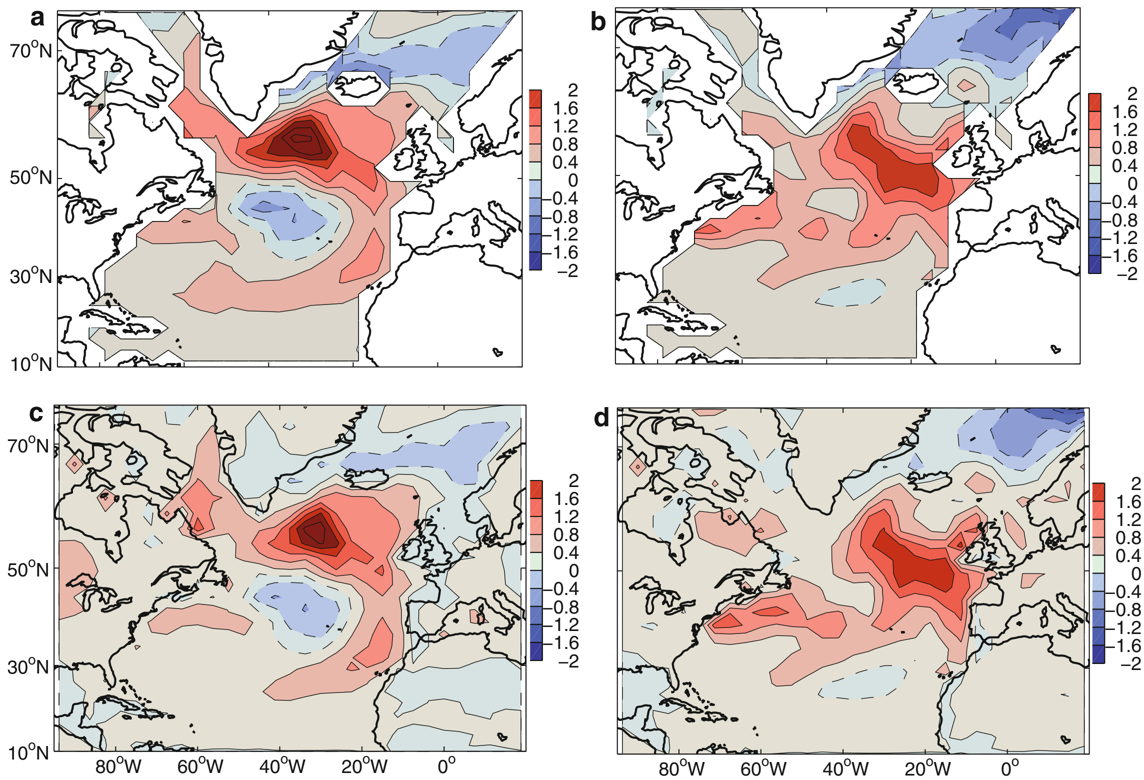


Fig. 3 SST anomalies (in °C) prescribed in the slab experiments and defined by the regression of the summer (JJA) SST anomaly in the IPSL model onto the MOC index, 5 years (**a**) and 10 years (**b**) after the MOC. The data are low-pass filtered in **b**. **c** Actual JJA SST anomalies computed by the slab model defined by the averaged difference between the 80-year control run and the run where the anomaly in **a** was added. **d** As in **c** in response to the anomaly shown in **b**

the climatology north of 60°N, thus restricting the slab domain in order to estimate the influence of the cooling in the Nordic Seas. The latter was found to have negligible impacts during summer and thus the last two experiments will be described with less details in this study.

The SST anomalies in Fig. 3a, b are diagnosed from the fully coupled run, namely IPSL. MF09 showed that during summer, an EAP-like response was forced by the main mode of low-frequency MOC variability, essentially describing a modulation of its strength. They found the maximum amplitude of the atmospheric response when the SST lags the MOC by 4–5 years, corresponding to a warming of the North Atlantic south of 60°N, a weak cooling at about 40°N, and a cooling of the Nordic Seas. The SST anomaly shown in Fig. 3a only slowly varies with increasing lags, although the warming is more uniform south of 60°N when the SST lags by 10 years (Fig. 3b). The latter SST anomaly resembles that associated with the model AMO, defined as in the observations by the low-passed filtered SST anomaly averaged over the North Atlantic. It also has quite similar impacts on the atmosphere (MF09). We prescribe the two SST anomalies (Fig. 3a, b) to determine the influence of the weak subtropical cooling seen in Fig. 3a. As the SST amplitude associated with a typical MOC fluctuation is small, the SST anomalies have been multiplied by a factor 9 to increase the signal-to-noise ratio and obtain a more robust atmospheric response. The SST patterns associated with the MOC showed negligible seasonal dependence. Hence, the anomalies in Fig. 3 were prescribed without seasonal variations. Figure 3c, d shows the actual SST anomalies computed by the slab model (Eq. 1). They are very similar to the SST anomalies diagnosed from the fully coupled run and their amplitude is not damped. This comparison ensures that the actual SST anomalies that are interacting with the atmosphere do not deviate much from the imposed SST anomalies. The atmospheric response analyzed hereafter thus corresponds to that forced by the North Atlantic SST anomalies associated with the MOC in IPSL.

The winter and summer North Atlantic SLP responses to the SST anomaly in Fig. 3a, b are shown in Fig. 2d, c. They are defined by the difference between the anomalous and control runs, averaged over December–January–February (DJF, Fig. 2c) and June–July–August (JJA, Fig. 2d). The winter atmospheric response projects onto the NAO, with a spatial correlation of 0.9, while the summer response projects onto the EAP, with a spatial correlation of 0.6. This is consistent with the coupled model results of MF09 and in

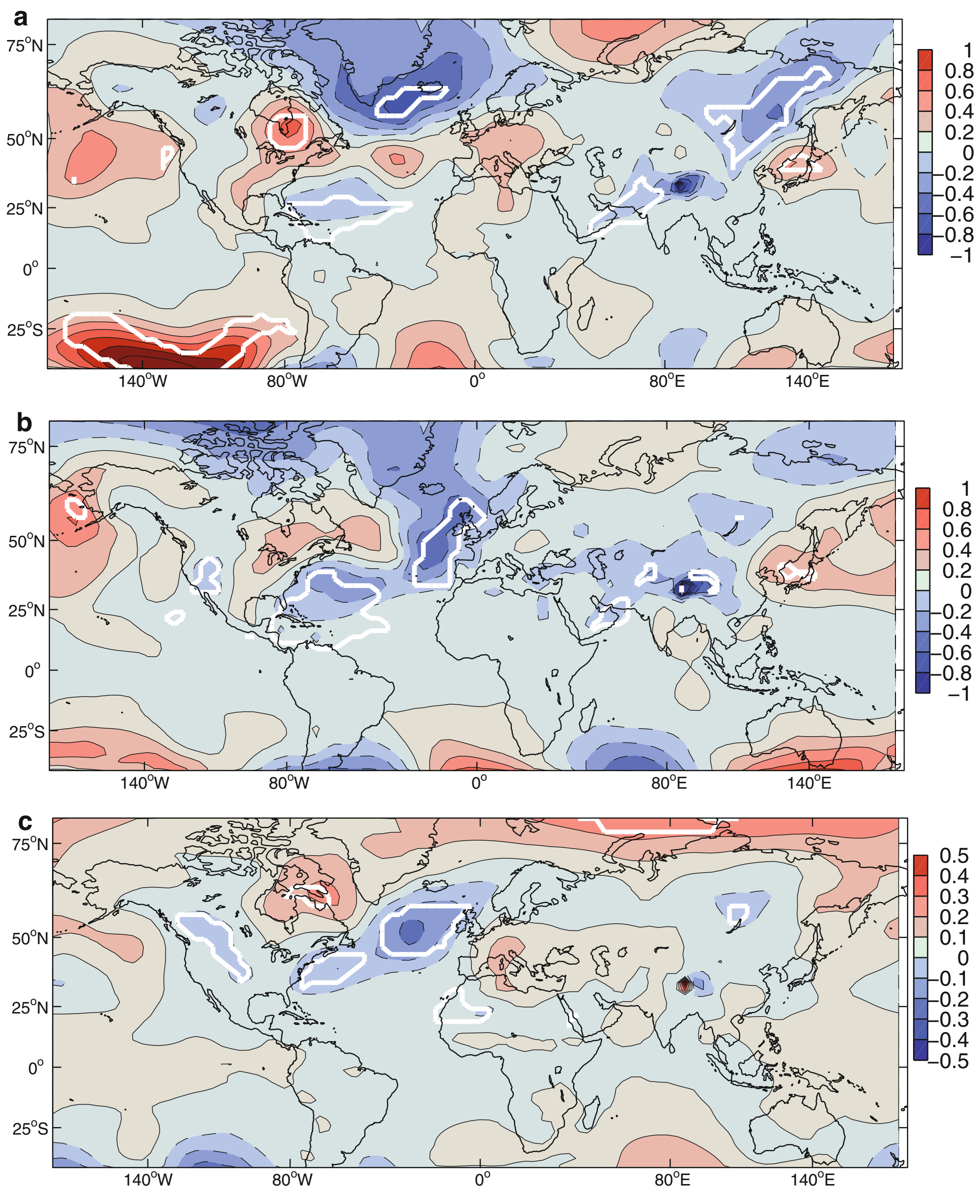


Fig. 4 **a** Summer (JJA) SLP anomalies (in mb) in the slab model in response to the SST anomaly in Fig. 3a, and **b** to the SST anomaly in Fig. 3b. **c** Regression of the JJA SLP anomalies (in mb) onto the MOC in the IPSL model, 5 years after the MOC. Note the differences in the scales between the slab model and IPSL. The *white contour* indicates statistically significant regions at the 5% level

agreement with previous model studies, which suggested that the atmospheric response to extratropical SST anomalies projects onto the internal variability (Deser et al. 2004; Peng et al. 2006). The mechanisms of the atmospheric response are investigated next, extending the description to the global scale, as for both seasons the prescribed North Atlantic SST anomaly leads to significant changes beyond the forcing region.

3 The summer signal

3.1 Response to an intensification of the MOC

As the summer response in IPSL was strongest about 5 years after the MOC, the summer atmospheric response was first investigated based on the SST anomaly in Fig. 3a. The global SLP response averaged over the months JJA is shown in Fig. 4a. The summer SLP response in the slab model shows some similarities with that in IPSL in the North Atlantic (Fig. 4c) with a significant EAP-like signal, as mentioned above. Although the comparison degrades when the whole northern hemisphere or the global domain is considered, it is good in a broad domain encompassing North America, the North Atlantic and Western Europe, with a pattern correlation of 0.40 in [95°W–30°E; 15°N–65°N]. In this domain, the null hypothesis of no response is rejected at the 5% level in 17% of the grid points, suggesting field significance as defined by Livezey and Chen (1983). In the slab model, the cyclonic anomalies are centered south of Greenland and are shifted north compared to IPSL, and the anticyclonic anomalies in the

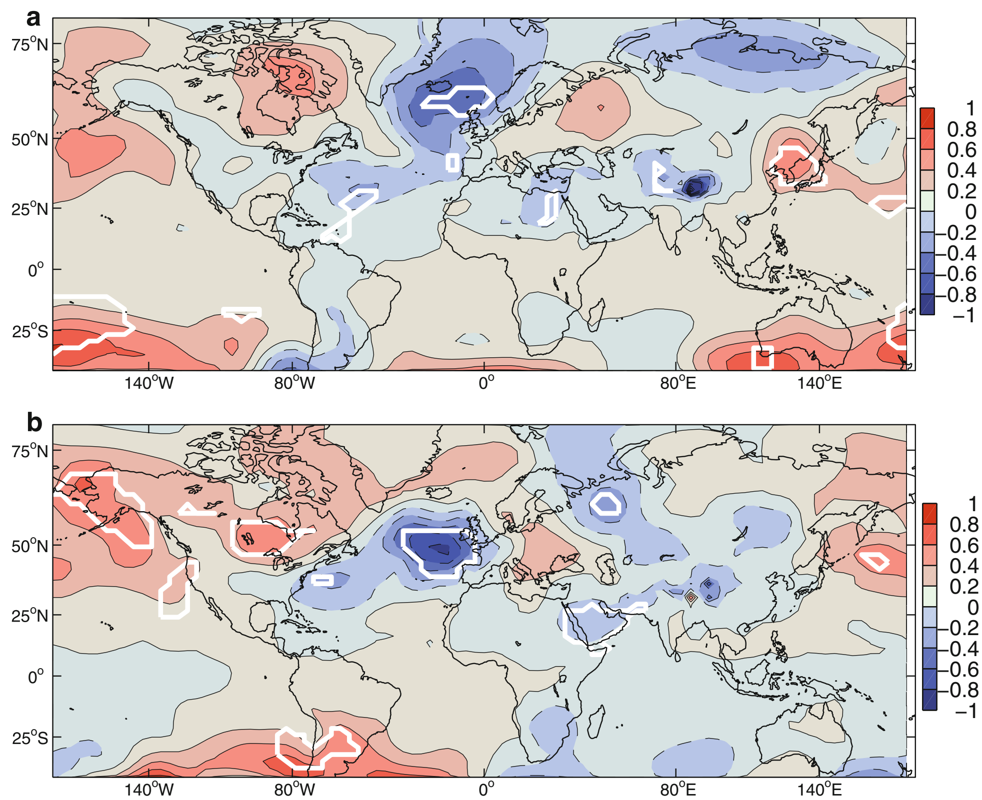


Fig. 5 **a** Summer (JJA) SLP anomalies (in mb) in the slab model in response to the SST anomaly in Fig. 3b removing the cold anomalies in the Nordic Seas. **b** Summer (JJA) SLP anomalies in a forced experiment prescribing the SST anomaly in Fig. 3b. The *white contour* indicates statistically significant regions at the 5% level

midlatitudes extend more zonally. The response over the tropical and subtropical Atlantic is negative both in IPSL and in the slab experiments, which indicates a weaker subtropical anticylone. Prescribing the SST anomaly in Fig. 3b leads to a SLP response closer to that in IPSL, with cyclonic anomalies over the whole North Atlantic (Fig. 4b). This suggests that the EAP-like response is sensitive to a uniform warming of the North Atlantic.

We have also tested the influence of the high latitudes cooling by performing an additional sensitivity experiment without the negative anomaly in the Nordic Seas. Figure 5a shows a SLP response similar to that induced by the SST in Fig. 3b, with the largest cyclonic anomalies over Greenland, and it also resembles the atmospheric response identified in IPSL. This indicates that the summer atmospheric circulation is mostly sensitive to the warming of the North Atlantic south of Iceland. It also provides an independent realization of the slab experiment showing consistent results, which reinforces the robustness of the summer atmospheric response described above. We further investigated the AGCM response by prescribing the SST anomaly in Fig. 3b, without a slab model (Fig. 5b). Significant negative SLP anomalies are again found in the North Atlantic, with a slightly stronger amplitude compared to the response in Fig. 4b. The patterns with and without a slab ocean show some local differences but they are both significant at the 5% level in the North Atlantic as in IPSL, suggesting some robustness. As discussed later, the forced AGCM response is not significant above the lower troposphere in summer and not for the SLP neither during winter, suggesting that using an interactive ocean does affect the detection of an atmospheric response in the extratropical North Atlantic.

The vertical structure of the North Atlantic summer response is presented in Fig. 6 for the total response (a) and its zonally asymmetric component (b). The North Atlantic response is equivalent barotropic in the mid and high latitudes and baroclinic in the tropics. The total response is dominated by extratropical anticyclonic anomalies that are weaker when the zonal mean is removed. The eddy response is mainly cyclonic and largest at high latitudes.

The Z500 response is shown in Fig. 7a. As it is largely equivalent barotropic over the North Atlantic, the height anomalies at other levels are qualitatively similar to those at 500 mb and are not shown. The Z500 response in summer is zonally elongated and global. In agreement with Fig. 6, it is dominated by anticyclonic anomalies that feature wave trains in the midlatitudes. Cyclonic anomalies are also found over Greenland and in the subtropical North Pacific and North Atlantic, extending northeastward to the Middle East and to China, but the latter anomalies are not statistically significant. The atmospheric sensitivity, defined as the maximum of the atmospheric response divided by the maximum of the SST anomaly, is about 5 m/K at 500 mb, with a

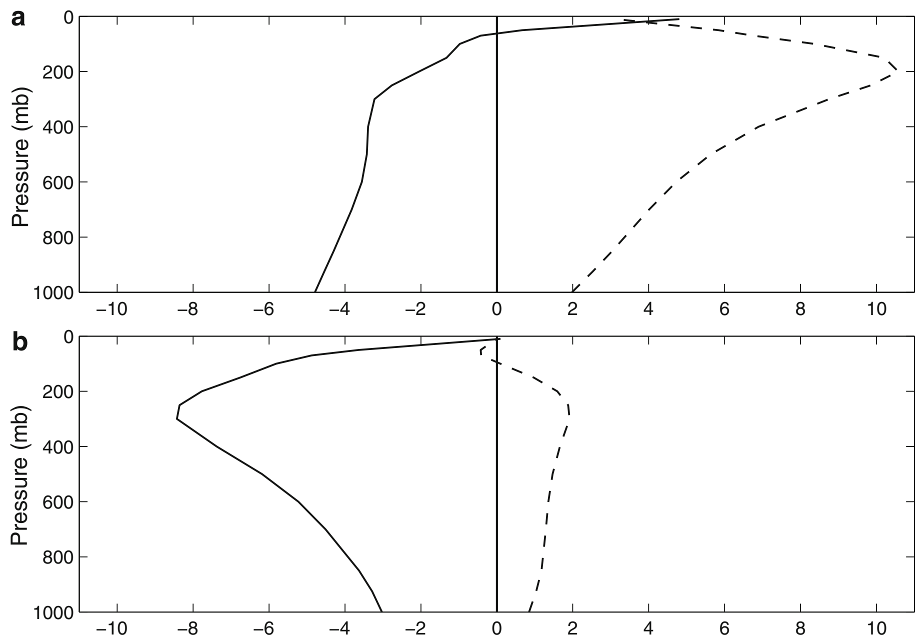


Fig. 6 Vertical profiles of the height anomalies (in m) in JJA averaged over the North Atlantic high latitudes ([60°W–20°W; 60°N–80°N], *solid line*), and the North Atlantic midlatitudes ([80°W–20°E; 40°N–50°N], *dashed line*). **a** Total response, **b** Eddy part (zonal mean subtracted)

stronger response when the zonal mean is not removed. It is weaker than that generally found during winter (e.g. Kushnir et al. 2002), presumably because of the different SST anomaly pattern prescribed in previous sensitivity studies and because of weaker dynamics during summer.

The zonal mean component of the response is large. To document it more clearly, the zonal mean part of the temperature and height responses are presented in Fig. 8, along with the zonal mean zonal wind anomalies. The anomalous warming extends to the whole troposphere and is associated with positive height anomalies centered at 45°N and 75°N, consistent with Fig. 7. The atmospheric response is broadly symmetrical with respect to the equator, with significant height and temperature anomalies in the southern hemisphere. By geostrophy, the increase (decrease) of pressure south (north) of 45°N is associated with a weaker (stronger) zonal wind. Consequently, the zonal mean zonal wind response in summer features a weakening of the subtropical jet and a strengthening of the subpolar jet. The comparison with the climatological mean state suggests that the anomalies represent a northward shift of the westerly jet.

The departure of the Z500 response from its zonal mean gives the eddy part of the response and indicates changes in the stationary waves (Fig. 7b). Compared to the total response, the ridges are weakened and the troughs are strengthened, especially in the Atlantic basin. Figure 7b also shows a significant response in the southern hemisphere subtropics, with wave-like anomalies, alternatively positive and negative, along 25°S.

The Z500 summer response in the slab model compares well with that in IPSL shown in Fig. 7c, particularly in the North Atlantic, which confirms that the SST anomaly induced by MOC changes is responsible for the atmospheric response identified in MF09. The comparison is better than near the surface, in particular when the zonal mean is removed. The zonal mean response in IPSL is weaker, probably because of different ocean–atmosphere adjustments that are not allowed in the slab model outside the North Atlantic. The sensitivity of the atmospheric response in the slab model is also weaker than in IPSL. Indeed the Z500 anomalies are about 5 m for a SST anomaly of 0.2°C in IPSL, against 10 m for an SST anomaly of about 2°C in the slab model. The amplitude of the response is thus about five times weaker in the slab model, assuming the response scales linearly with the amplitude of the forcing. This difference could be attributed to the non-linearity of the response as well as to the influence of other forcing than SST in IPSL, as will be discussed later. Despite its weak amplitude, the response is significant at the 95% confidence level. The significance was further investigated using a Monte Carlo approach, following Wilks (1997) and the results were identical to the *t* test (not shown). To address the robustness of the atmospheric response, the 80-year integration was divided into two parts and the Z500 anomalies examined in each half (Fig. 9). Despite somes local differences, the global scale anticyclonic anomalies are fairly well reproduced in each part and are statistically significant, with a pattern correlation of 0.43 in the domain [130°W–10°E; 0–70°N]. In this domain, the null hypothesis of no response can be rejected at the 5% level in 18% of the grid points, suggesting field significance (Livezey and Chen 1983).

As shown in MF09, the summer atmospheric response projects onto the EAP in the North Atlantic, both near the

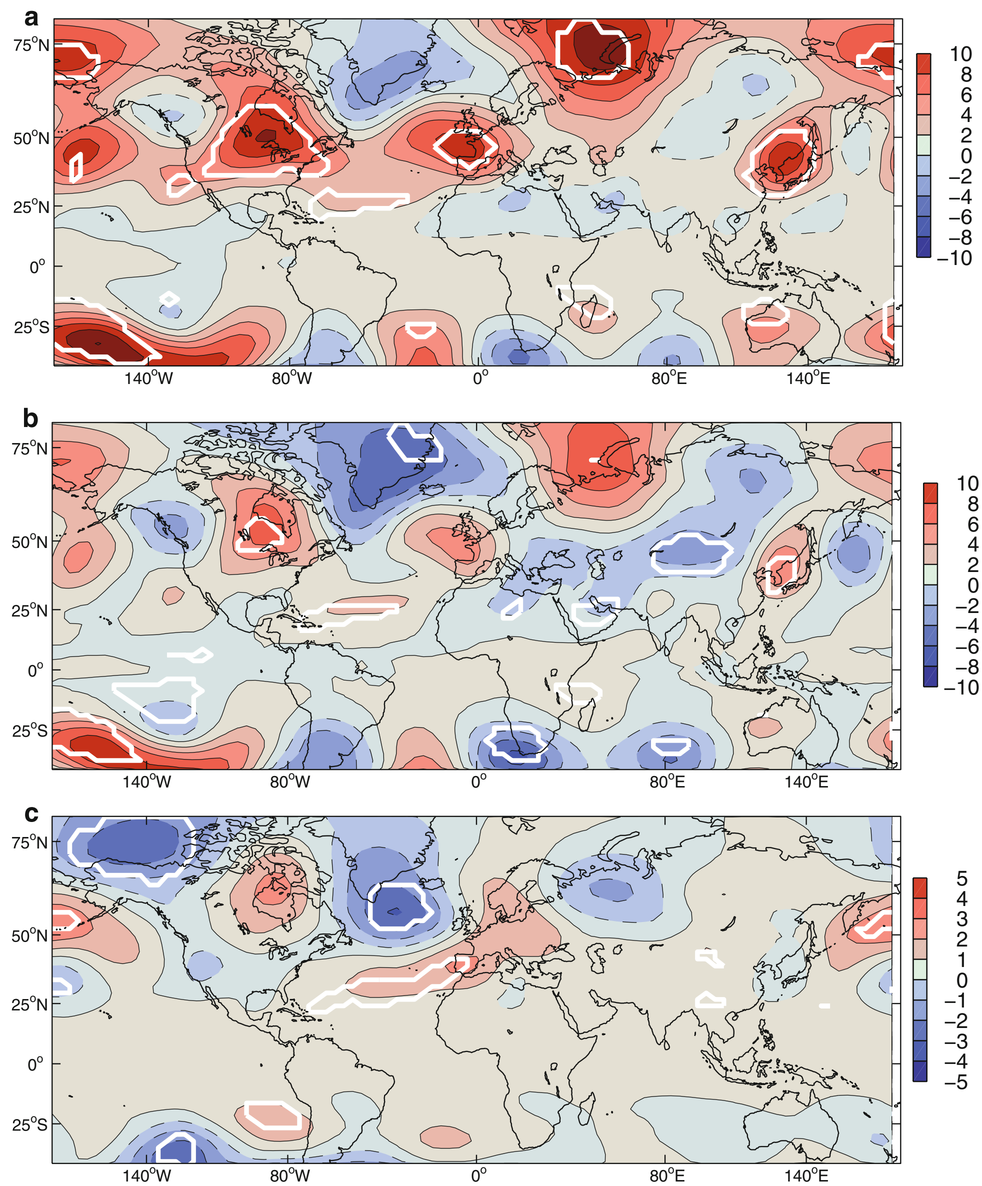


Fig. 7 **a** 500 mb geopotential height anomalies in the slab model averaged over the JJA months. **b** As in **a** with the zonal mean subtracted. **c** Regression of the JJA Z500 anomalies in the IPSL model onto the MOC index 4 years after a maximum of the MOC. Units are in m. The *white contour* indicates the 5% significance level

surface and in the troposphere. Indeed, the model EAP at 500 mb is very similar in the North Atlantic to the pattern in Fig. 7, with cyclonic anomalies south of Iceland surrounded by anticyclonic ones over North America and western Europe, extending southwestward towards the Gulf of Mexico. This result is consistent with previous studies (e.g. Deser et al. 2004; Peng et al. 2006) that showed a tendency of AGCM models to produce in forced experiments an atmospheric circulation that projects onto its natural low-frequency variability. Here, the atmospheric response resembles the mode that drives the low-frequency fluctuations of the MOC, suggesting a positive feedback on the atmosphere.

Quite similar results are obtained when the slab model is forced by the SST anomaly in Fig. 3b. The temperature response shows a warming of the whole troposphere associated with positive height anomalies (not shown). The eddy part of the response exhibits a trough over the warm SST but the signal is more noisy and is mostly significant only near the surface. Nevertheless it suggests that the atmosphere response is mostly sensitive to the warming of the North Atlantic and only weakly affected by the cooling in the subtropics.

3.2 Non-linearity of the response

The atmospheric response to a weakening of the MOC is investigated by forcing the slab model with an SST anomaly that has the same structure than that in Fig. 3a, but the opposite sign. As in the previous section, the response,

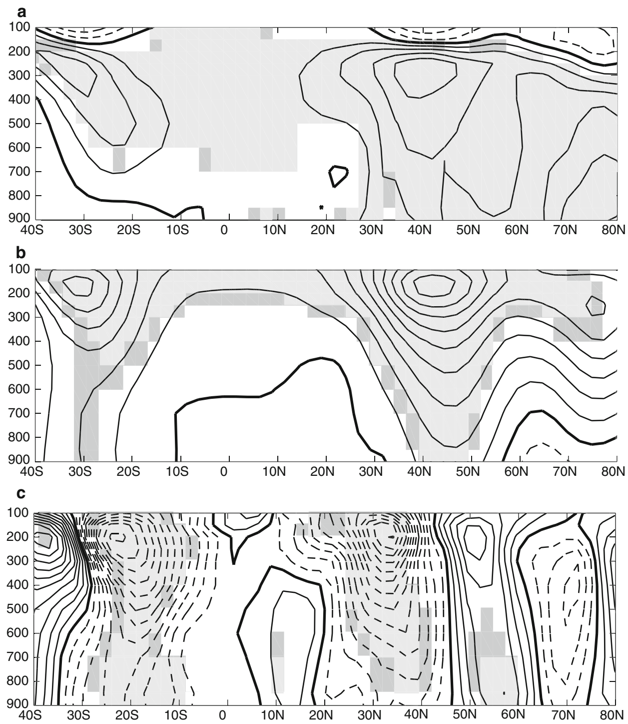


Fig. 8 Zonal mean summer response (JJA). **a** Temperature anomalies (contour every 0.05 K). **b** Height anomalies (contour every 1 m). **c** Zonal wind anomalies (contour every 0.05 m/s). Negative contours are *dashed*. The *heavy black line* is the zero contour. The areas *shaded* in *light* (*dark*) *grey* are significant at the 5% (10%) level

which is less significant, is decomposed onto a zonal mean and an eddy part. The zonal mean temperature (Fig. 10a) reveals a cooling of the northern hemisphere that is shallower than the warming in Fig. 8a, reaching its maximum near the surface. The cooling does not extend to the southern hemisphere while the warming in Fig. 8a was global. Negative height anomalies are associated with this cooling. Reversing the sign of the SST anomaly thus broadly changes the sign of the zonal mean temperature and pressure responses in the northern hemisphere. However, because of small meridional shifts, the zonal mean of the zonal wind anomalies keeps the same sign north of about 30°N, reflecting a weakened subtropical jet and a stronger subpolar jet (Fig. 10b). The Z500 response in Fig. 11 is consistent with this strong non-linearity and shows a response very similar to that in Fig. 7, i.e. no change of sign in the northern North Atlantic. The zonal mean is weaker than in Fig. 7a, thus the eddy part of the Z500 response is very similar to that in Fig. 11 (not shown). Peng and Whitaker (1999) showed that the equilibrium atmospheric response to midlatitude SST anomalies was modulated by the interaction between the zonal wind and the transient eddies. Therefore the similarity between the Z500 responses for both signs of the SST forcing is in agreement with that of the zonal wind response. Previous studies mainly emphasized the non-linearity of the winter atmospheric response. Our results show strong non-linearities in summer, suggesting a large influence of transient eddies during summer as well. Reversing the sign of the SST forcing leads to positive Z500 anomalies in the tropical Atlantic, suggesting a linear response in this region.

3.3 Physical mechanisms of the summer atmospheric response

In order to describe more precisely the atmospheric response to the MOC-induced SST anomaly in terms of stationary wave activity, the Plumb vector **F** (Plumb 1985) is estimated. It is derived from a locally applicable conservation relation for quasi-geostrophic waves on a zonal flow and provides an estimation of the three-dimensional stationary wave activity. The plumb vector **F** is computed from the monthly zonally asymmetric part (denoted by *)

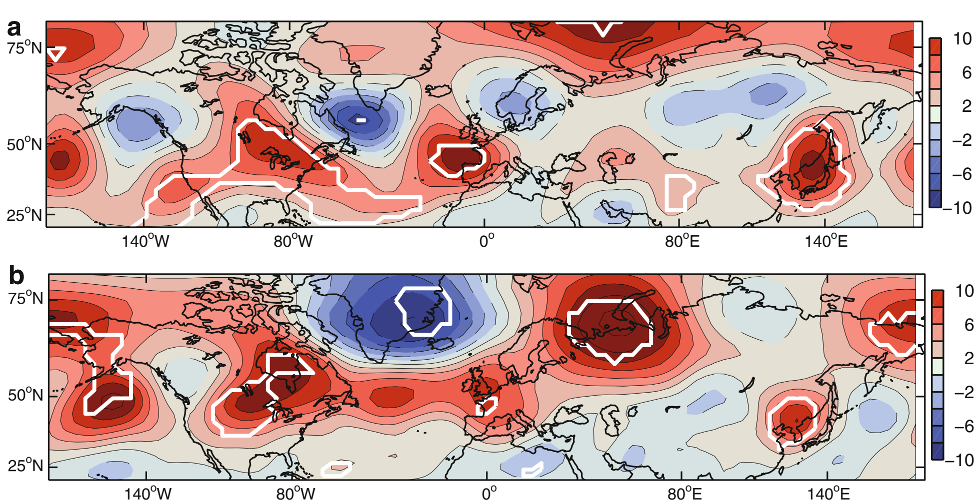


Fig. 9 500 mb geopotential height anomalies in JJA in each half of the 80-year-long integration. **a** Part 1 (averaged over year 1–40). **b** Part 2 (averaged over year 41–80). Units are in m. The *white contour* indicates the 10% significance level deduced from Monte Carlo simulations using 500 realizations

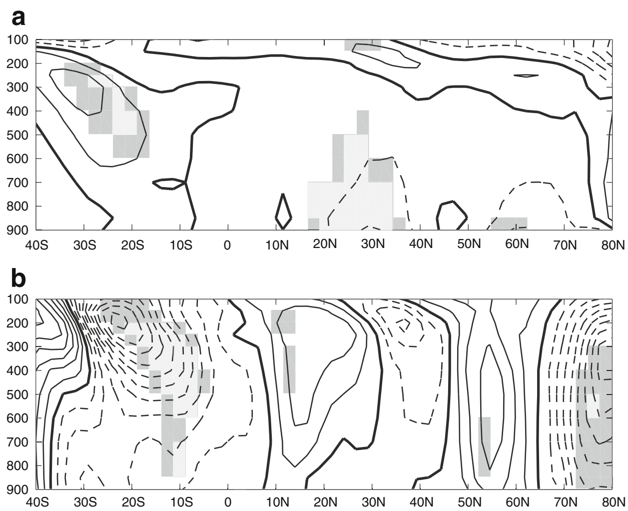


Fig. 10 Zonal mean summer response to a cooling of the North Atlantic associated with a weakening of the MOC (SST anomaly in Fig. 3a, with the sign reversed). **a** Temperature anomalies (contour every 0.05 K). **b** Zonal wind anomalies (contour every 0.05 m/s). Negative contours are *dashed*. The *heavy black line* is the zero contour. The areas *shaded* in *light* (*dark*) *grey* are significant at the 5% (10%) level

of the time mean geopotential, written in cartesian coordinates as in Fraedrich et al. (1993):

$$\mathbf{F} = \{F_x, F_y\} = \frac{\sigma}{2f^2}\left\{\Phi_x^{*2} - \Phi^*\Phi_{xx}^*; \Phi_x^*\Phi_y^* - \Phi^*\Phi_{xy}^*\right\} \quad (2)$$

where F_x and F_y are the zonal and meridional component of **F**, $\sigma = p/1{,}000$ hPa, and $\Phi^* = gz^*$ with z the geopotential height at 500 mb. Φ_x denotes the partial derivative of Φ with respect to x. An anomalous divergence (convergence) of **F** is associated with the creation (dissipation) of an anomalous stationary wave. **F** is also parallel to the direction of propagation of stationary waves. It was verified that the major sources of stationary waves activity in the northern hemisphere were well reproduced in the mean state of the model, with three main areas in Eastern Asia, Western North Atlantic, and North Pacific-Western North America.

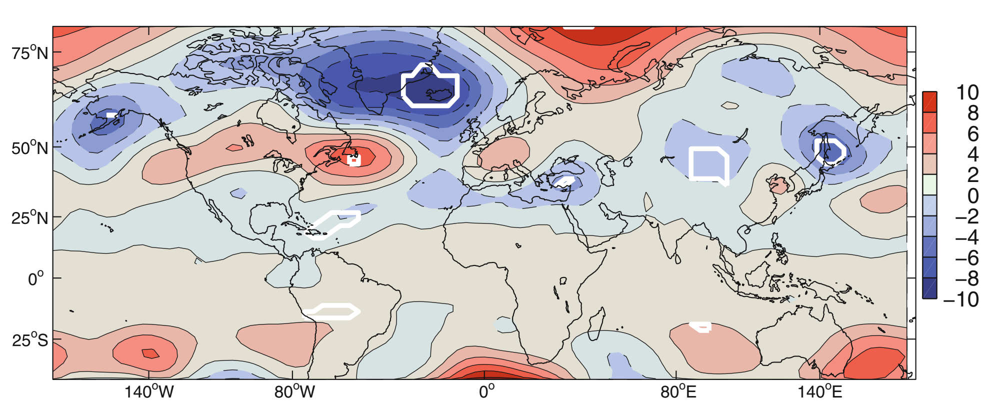
The stationary wave activity response to the MOC-induced SST anomaly in Fig. 3a is displayed in Fig. 12a. The anomalous Plumb vectors show a significant stationary wave activity over the northern part of the North Atlantic basin, with the strongest anomalies over Greenland and the Barent Seas. These changes are consistent with the eddy part of the Z500 response (Fig. 7b). The direction of the Plumb vector suggests a wave propagation from the Barents Seas to the north-eastern US through Greenland.

As discussed above, the strong non-linearity of the response suggests a forcing by the transient eddies. However, only monthly means were saved in the control and sensitivity experiments, preventing us to consider the transient eddy activity. In order to investigate the latter, another set of experiments was performed with the SST anomaly in Fig. 3a, keeping the daily means for 20 years. The storm track activity is defined following Hoskins and Valdes (1990) by the root mean square value of the 2.2–6 days fluctuations of the Z500 field ($\sqrt{z'^2}$, where the prime refers hereafter to the band-pass filtered daily data). The JJA storm track activity response to the SST anomaly associated with an intensified MOC is displayed in Fig. 12b. It shows an enhanced storm track activity over eastern US centered at about 45°N and extending northeastward to the whole Atlantic subpolar gyre. The comparison with the climatological mean state (not shown) suggests an enhancement and a northward shift of the mean storm track activity, consistent with the jet displacement in Fig. 8c.

The regions of large baroclinic activity are characterized by strong anomalous transient eddy heat fluxes. It can be shown that in response to an intensified MOC, more heat is transported northward by the eddies, with the strongest heat transport located over eastern US, consistently with the storm track anomaly showed in Fig. 12b.

The Eliassen–Palm vector (**EP**) is a prominent diagnostic tool to analyze the local interaction between transient eddies and the time mean flow (Holopainen 1984). As shown by Hoskins et al. (1983), the divergence (convergence) of the **EP**-vector depicts a tendency of the eddies to

Fig. 11 Z500 anomalies (in m) in JJA in response to a cooling of the North Atlantic (Fig. 3a with the sign reversed). The *white contour* indicates the 5% significance level



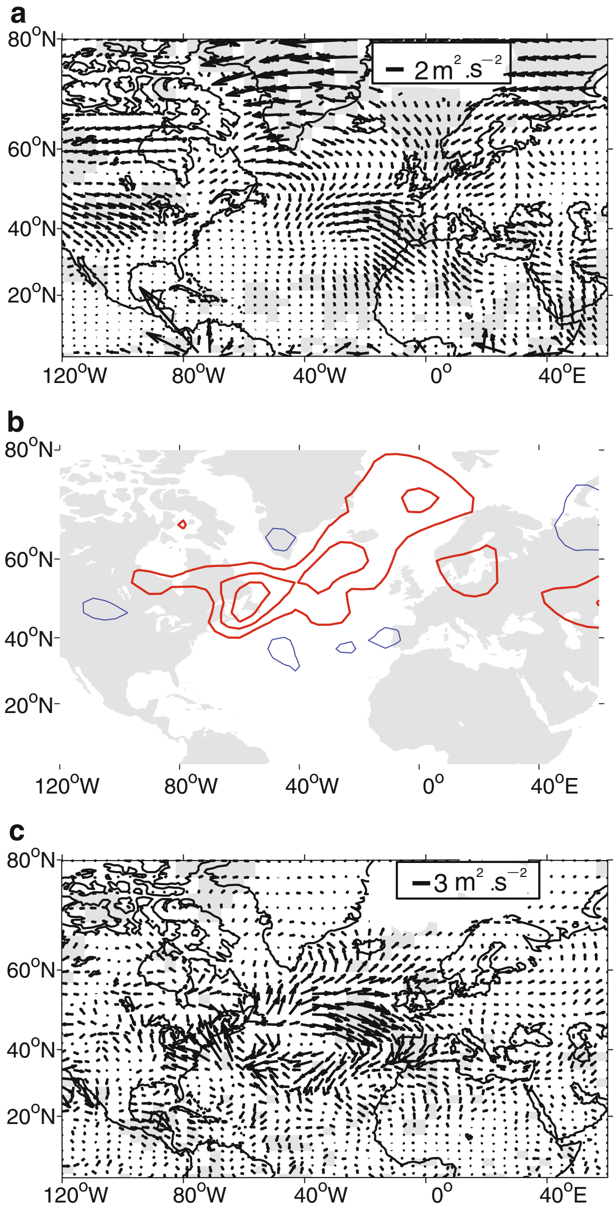


Fig. 12 **a** Plumb vector anomalies **F** at 500 mb. **b** 500 mb storm track activity response. Contours are every 1 m with the zero contour omitted. *Red* (*blue*) contours are for positive (negative) values. **c** Horizontal Eliassen–Palm vector (**EP**) anomalies at 200 mb. For all figures the anomalies correspond to the JJA response to the SST anomaly in Fig. 3a. Statistically significant regions at the 5% level are *shaded* in *grey* in **a** and **c** and omitted in **b** as the displayed anomalies are all significant

increase (decrease) the westerly mean flow. Furthermore, the shape of the eddies and the direction of the group velocity can also be determined. In the barotropic case, **EP** is in the direction of the group velocity of the transient eddies relative to the local time-mean flow (Trenberth 1986; Lau 1988). Following Trenberth (1986), the zonal and meridional **EP** components are defined, respectively, by the momentum flux of transient eddies $1/2(\overline{v'^2} - \overline{u'^2})$, and $-\overline{u'v'}$ where u and v are the zonal and meridional wind components at 200 mb. The anomalous **EP** vector in response to the SST anomaly in Fig. 3a is presented in Fig. 12c for JJA. It shows westerly divergent **EP** vectors over the North Atlantic in JJA. It is associated with a local acceleration of the jet stream by the eddies that is likely to induce the zonally orientated anticyclonic and cyclonic anomalies found in the Z500 field in Fig. 7a, north and south of the storm track, respectively. The larger **EP** flux anomaly south of the storm tracks and its southeastward direction are consistent with the anticyclonic anomalies that dominate the Z500 North Atlantic response (Fig. 7a). The **EP** flux response depicts the tendency of the eddies to strengthen the baroclinicity which first made them grow and acts thus as a positive feedback. The eddy vorticity flux divergence (convergence) thus plays a role in driving and maintaining the anomalous Z500 anticyclonic (cyclonic) circulation.

These results suggest that the atmospheric response in summer is primarily induced by changes in the Atlantic region. The EAP-like response is first created by an anomalous stationary wave and then reinforced and maintained by the transient eddies. The zonal mean circulation changes are thus mainly the result of interactions with the short-time scale fluctuations. The same type of influence was found by Drévillon et al. (2001) to explain the link between a summer North Atlantic SST anomaly and the following winter atmospheric circulation over Europe. Our study suggests that this mechanism is also important during summer.

Although the summer atmospheric response appears to be strongly controlled by local changes in the transient eddy activity in the North Atlantic, the equilibrium response might also be affected by non-local changes via atmospheric teleconnections. Prescribing an SST anomaly in the Atlantic ocean leads to a warming of the whole northern hemisphere that is associated with global-scale anticyclonic anomalies. Our results suggest that these global anomalies could be forced by changes in the Indo-Asian region, which is an active monsoon region during summer. An increased upper-level divergence is found over the Indo-Asian region suggesting a strengthening of the summer monsoon (not shown). As the SST anomaly was only prescribed in the North Atlantic, this suggests that the North Atlantic can influence the Indian summer monsoon region. A teleconnection between the Indian monsoon region and the North Atlantic was found in observational and model studies (Zhang and Delworth 2006; Goswami et al. 2006; Kucharsky et al. 2007; 2008). The warming of the Indo-Asian continent induces a direct circulation with rising air above. This alters the local Hadley cell, and the upper level divergence can act as a source of anomalous Rossby waves that propagate away and can lead to a global response (Hoskins and Karoly 1981; Branstator 2002). The Rossby wave sources were computed based on daily data for the summer season. Significant vorticity anomalies were found over south India (not shown), consistently with the increased upper level divergence. However the response was very noisy and the design of our experiments does not allow us to explore further the mechanism of this teleconnection.

The atmospheric response in the tropical and subtropical Atlantic is linear and baroclinic as expected from linear theory. It can be shown that the upper level streamfunction reveals an anomalous anticyclone above the negative SLP anomaly characteristic of a Gill-type response to diabatic forcing, and similar to the response in other AGCMs described by Hodson et al. (2009) and Kushnir et al. (2010).

4 The winter signal

4.1 Response to an intensification of the MOC

The winter SLP response was shown in Fig. 2c for the North Atlantic and is described here on larger scales (Fig. 13a). In the IPSL coupled model, the MOC has also a significant influence on the atmosphere during winter, with anomalies that project onto the NAO (Fig. 13b). This response was found to be significant 8–11 years after the MOC and it was maximum at lag 9. It was not described in details by MF09 because it was less robust than the summer signal. The consistency of this signal is assessed in this study by analyzing the winter atmospheric response to the prescribed SST anomalies in Fig. 3. As the winter response in IPSL was strongest about 10 years after the MOC, to allow a better comparison with the fully coupled model only the response to the SST anomaly in Fig. 3b is described. However, very similar results were obtained when using the pattern in Fig. 3a. The winter atmospheric response in the slab model is very similar to that in IPSL, with the strongest signal over the North Atlantic forcing region, projecting on a negative phase of the NAO in the case of an intensified MOC. Note that the southern lobe of the NAO is not significant at the 95% confidence level in IPSL but it is at the 90% level. In the North Atlantic domain [80°W–5°E; 25°N–80°N, the null hypothesis of no response is rejected at the 5% level in 20% of the grid points, suggesting field significance (Livezey and Chen 1983). Significant anomalies are found beyond the forcing region, in particular over the North Pacific with positive pressure anomalies that indicate a weakening of the Aleutian low. This extension of the North Atlantic anticyclonic anomalies to the North Pacific, leads to a pattern that projects onto the Arctic Oscillation. The Z500 anomalies are mostly similar to those near the surface in the

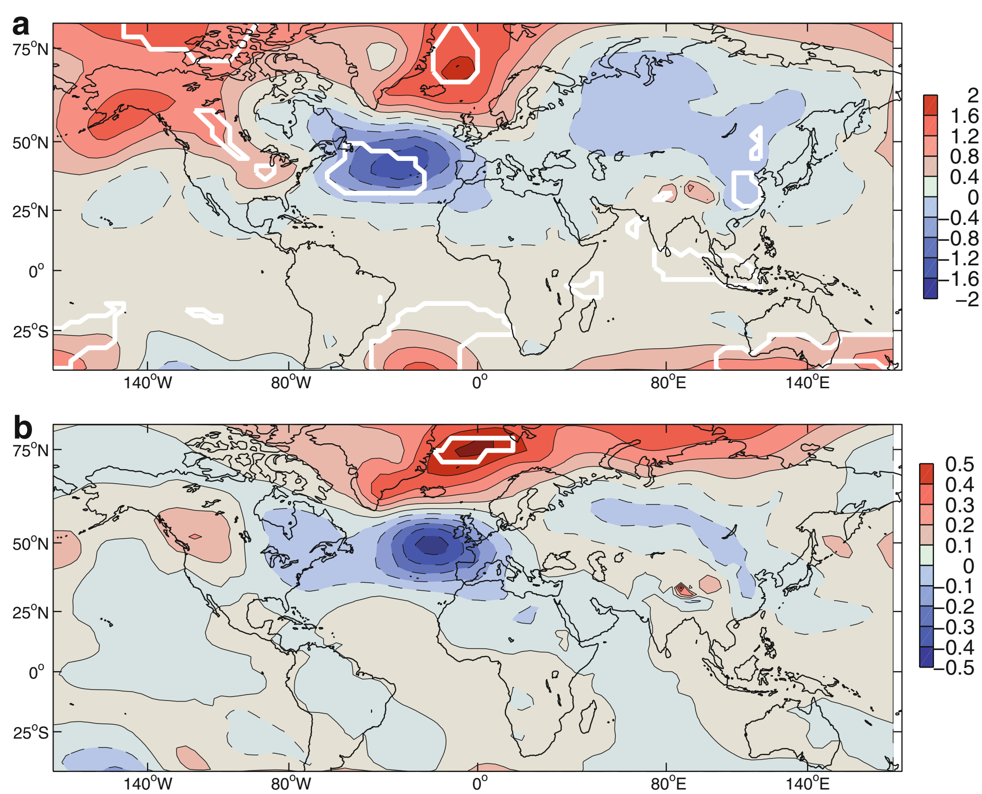


Fig. 13 **a** Winter (DJF) SLP anomalies (in mb) in the slab model in response to the SST anomaly in Fig. 3b. **b** Regression of the DJF SLP anomalies (in mb) onto the MOC in the IPSL model, 9 years after the MOC. Note the differences in the scales between the slab model and IPSL. The *white contour* indicates statistically significant regions at the 5% level

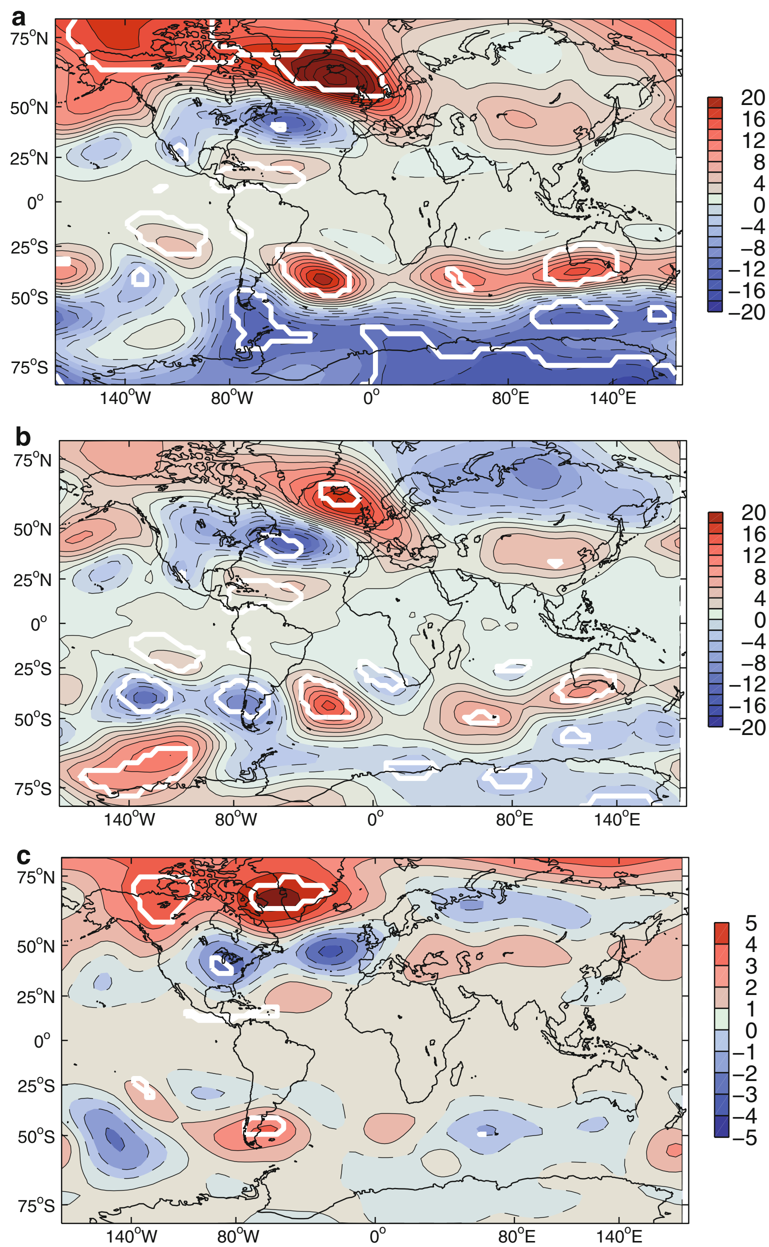


Fig. 14 **a** DJF Z500 response in the slab model. **b** Zonally asymmetric part of the DJF Z500 response. **c** Regression of DJF Z500 anomalies in the IPSL model 9 years after a maximum of the MOC. Units are in m. The *white contour* indicates the 5% significance level. Note the differences in the scale between the slab model and IPSL

extratropical North Atlantic and North Pacific, a consistent feature of the equivalent barotropic response (Fig. 14a).

The anomalies are weaker and baroclinic in the tropical North Atlantic and over central Europe and Asia. There is also evidence, in particular in the slab experiments, of a wave-like structure in the North Atlantic, with zonally orientated anomalies that are tilted northeastward. This tripolar structure is found in each half of the integration, with the eastward shift appearing in the second part only (not shown). As in summer, the Z500 response in the slab model is closer to that in IPSL when the zonal mean is removed (Fig. 14b). The sensitivity of the winter response is also about five times weaker in the slab model, as in the summer, reaching about 10 m/K. However, this is comparable to the amplitude found in previous model studies that focused on the winter (Kushnir et al. 2002).

Significant anomalies are also found in the southern hemisphere, with zonally elongated low-pressure anomalies south of 50°S, and high-pressure anomalies northward (Fig. 14a, b). This response projects onto a positive phase of the Southern Annular Mode and suggests teleconnections with the North Atlantic. Note that these anomalies were not found in IPSL (Fig. 14c).

The zonal mean response in DJF for temperature and zonal wind (Fig. 15) reveals a tropospheric warming in the northern hemisphere, which is mainly confined to the high latitudes, close to the maximum of the prescribed SST anomaly. It can be shown that this warming is associated

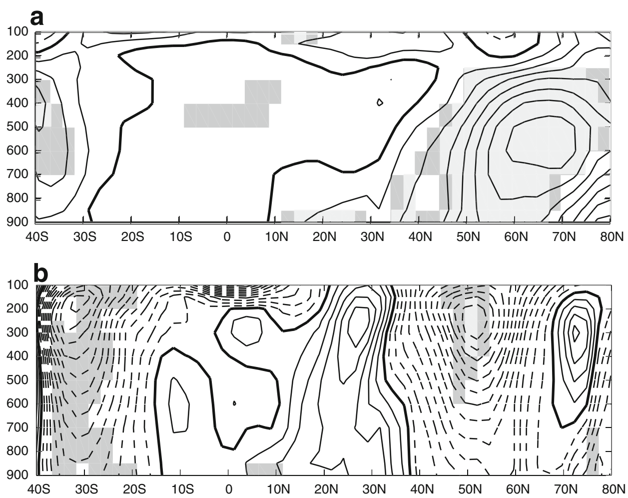


Fig. 15 Zonal mean response in winter (DJF). **a** Temperature anomalies (contour every 0.05 K). **b** Zonal wind anomalies (contour every 0.05 m/s). Negative contours are *dashed*. The *heavy black line* is the zero contour. The areas *shaded* in *light* (*dark*) *grey* are significant at the 5% (10%) level

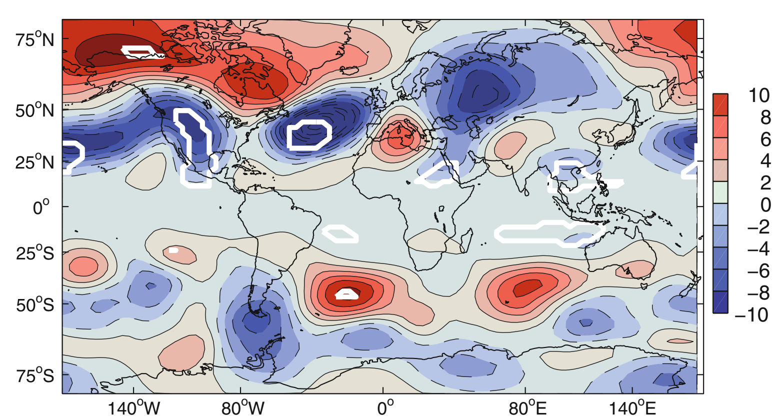
with positive geopotential height anomalies north of 50°N. The zonal mean zonal wind anomalies are thus positive in the subtropics and negative north of 40°N, which depicts a southward shift of the jet. Consistently with the limited hemispheric extension of the zonal mean warming, very similar results were obtained when subtracting the zonal mean (Fig. 14b). Removing the negative anomaly in the Nordic Seas leads to larger differences in winter than in summer (not shown). The anticyclonic anomalies at high latitudes are shifted south-eastward and the response does not project on the NAO anymore. The teleconnection in the North Pacific is also different. This suggests that the winter atmosphere is more sensitive to the high latitude cooling associated with MOC changes in the model. The winter atmospheric response in the AGCM case forced by the SST anomaly in Fig. 3b without a slab ocean model is not significant and does not project on the NAO (not shown). It suggests that the use of an interactive ocean model is needed to detect a response in winter.

4.2 Non-linearity

The non-linearity of the winter atmospheric response to changes in the MOC was investigated as in the summer case by prescribing an SST anomaly of the opposite sign as in Fig. 3b, which corresponds to a weakening of the MOC. Over the North Atlantic, the Z500 response projects onto a negative phase of the NAO, like the one forced by a strengthening of the MOC (Fig. 16). The North Pacific and the southern hemisphere anomalies also have the same sign as those in Fig. 14a, although less significant.

In the tropics, there is a change of sign with that of the SST, showing negative Z500 anomalies between 20°S and 20°N while Fig. 14 showed positive ones. This suggests that the winter atmospheric response is broadly non-linear in the midlatitudes and linear in the tropics. Analyzing the zonal mean temperature response indicates that the cooling prescribed in the North Atlantic extends to the northern hemisphere, suggesting some linearity in the temperature response (Fig. 17a). Despite the change of sign with the SST polarity, the temperature response is asymmetric and it can be shown that the geopotential height anomalies are shifted northward as well, compared to the response forced by a positive phase of the SST anomaly. Therefore, the zonal mean zonal wind anomalies in Fig. 17b indicate a weakening of the subtropical jet and a strengthening of the subpolar jet, similarly to what was found in Fig. 15b. Hence, as in summer, the zonal wind anomaly keeps the same sign in the midlatitudes whatever the sign of the SST. This does not explain the origin of the non-linearity but it is consistent with the non-linearity of the geopotential response. The non-linearity is stronger in winter as the result of stronger anomalous baroclinic activity as shown below. Note that the strong non-linearity of the winter response makes it difficult to be detected in the coupled simulation using linear methods, as in MF09.

Fig. 16 DJF Z500 response (in m) to a weakening of the MOC (SST anomaly in Fig. 3b, with the sign reversed). The *white contour* indicates the 5% significance level



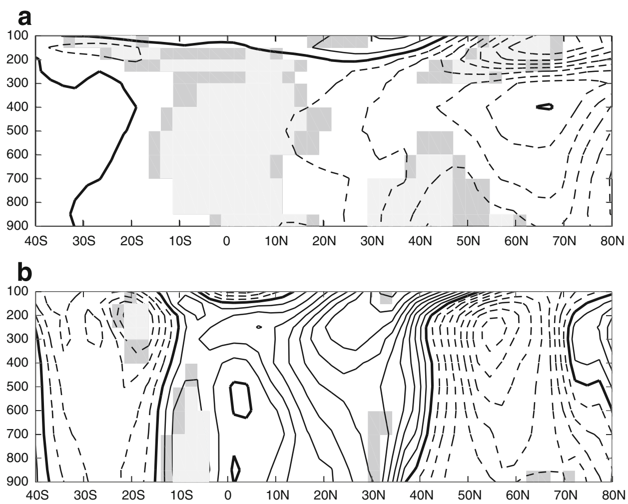


Fig. 17 Zonal mean DJF response to a weakening of the MOC (SST anomaly in Fig. 3b, with the sign reversed). **a** Temperature anomalies (contour every 0.05 K). **b** Zonal wind anomalies (contour every 0.05 m/s). Negative contours are *dashed*. The *heavy black line* is the zero contour. The areas *shaded* in *light* (*dark*) *grey* are significant at the 5% (10%) level

4.3 Physical mechanisms of the winter atmospheric response

The zonal wind anomalies in the upper troposphere are mainly confined to the Atlantic sector, with a banded structure consisting of negative anomalies over the tropical Atlantic, positive anomalies over the subtropics, and negative anomalies at mid and high latitudes, suggesting a southward shift of the storm tracks for a positive SST forcing (Fig. 18a). The changes in the upper troposphere zonal wind are consistent with those of the **EP** fluxes, suggesting that the equatorward shift of the jet is driven by a southward shift in the baroclinic activity. The **EP** vector anomalies show an increased transient eddy activity south of the mean storm track location and a smaller decrease north of it (Fig. 18b). The increase in the subtropical jet tends to move the region of maximum transient eddy activity toward the south, partly through an increase in the vertical wind shear. The southward shift in the baroclinic activity then induces a southward shift of the eddy-driven polar jet. The transient eddy activity is also significantly altered outside the North Atlantic. Anomalous **EP** fluxes over the North Pacific indicate that the transient eddies tend to decelerate the mean flow (no shown). This is consistent with the weakening of the winter Aleutian low described in Fig. 13b. A similar response was found in other AGCMs, as discussed in Sect. 5. Figure 18a also indicates significant changes in the southern hemisphere with zonally elongated wind anomalies. This atmospheric response projects onto a positive phase of the Southern Annular Mode with intensified westerlies south of 50°S and weaker winds northward, suggesting a link between the North and South Atlantic, as suggested by the Z500 response. However, further investigations are needed to identify the mechanism of this teleconnection.

The mechanisms of the winter atmospheric response are also associated with changes in the subtropics. The velocity potential changes associated with the MOC-induced SST anomaly show a significant anomalous upper-level divergence over the Gulf of Mexico, extending eastward to Southern Europe and West Africa (Fig. 18c). This anomalous divergence weakens the mean convergent state and depicts a weakening of the subtropical anticyclone, consistently with the Gill-type response described in previous studies (Hodson et al. 2009; Kushnir et al. 2010). Unlike the latter studies, the tropical SST anomalies imposed in the tropics are very weak, suggesting that the weakening of the subtropical anticyclone is forced by the extratropics.

5 Summary and discussion

The aim of this study is to investigate the mechanisms of the atmospheric response to an MOC-induced SST anomaly in the North Atlantic. It is based on the results of MF09 who showed that in the IPSL coupled model, changes in the MOC have a significant influence on the atmosphere, with a robust response in summer and hints in winter. To confirm these findings, we ran and analyzed sensitivity experiments with a slab version of the coupled model. We investigated the atmospheric response to the prescribed SST anomalies that were associated with low-frequency changes in the Atlantic MOC. As in the IPSL model, we found a significant response in the North Atlantic that projects onto the EAP during summer, and onto the NAO during winter, with an equivalent barotropic structure. This confirms the statistical analysis of MF09. Although the sensitivity of the atmospheric response was found to be substantially weaker in the slab model than in IPSL, we assessed its robustness by analyzing each half of the 80-year long integration and using complementary non-parametric tests of significance. We found our results to be statistically significant at the 95% confidence level, although the signal-to-noise ratio is small. Moreover the sensitivity of the winter atmospheric response is comparable to previous AGCM studies with an amplitude of 10 m/K at 500 mb. The JJA response is about half the strength of the winter response but this is expected as the summer is a less dynamically active season. The different amplitude between the fully coupled and the slab models is probably due to the non-linearity of the atmospheric response with respect to the amplitude of the SST forcing. It also suggests that other forcings than SST drive the atmospheric response in IPSL. For instance, the sea-ice coverage was kept to its climatological value. However, the

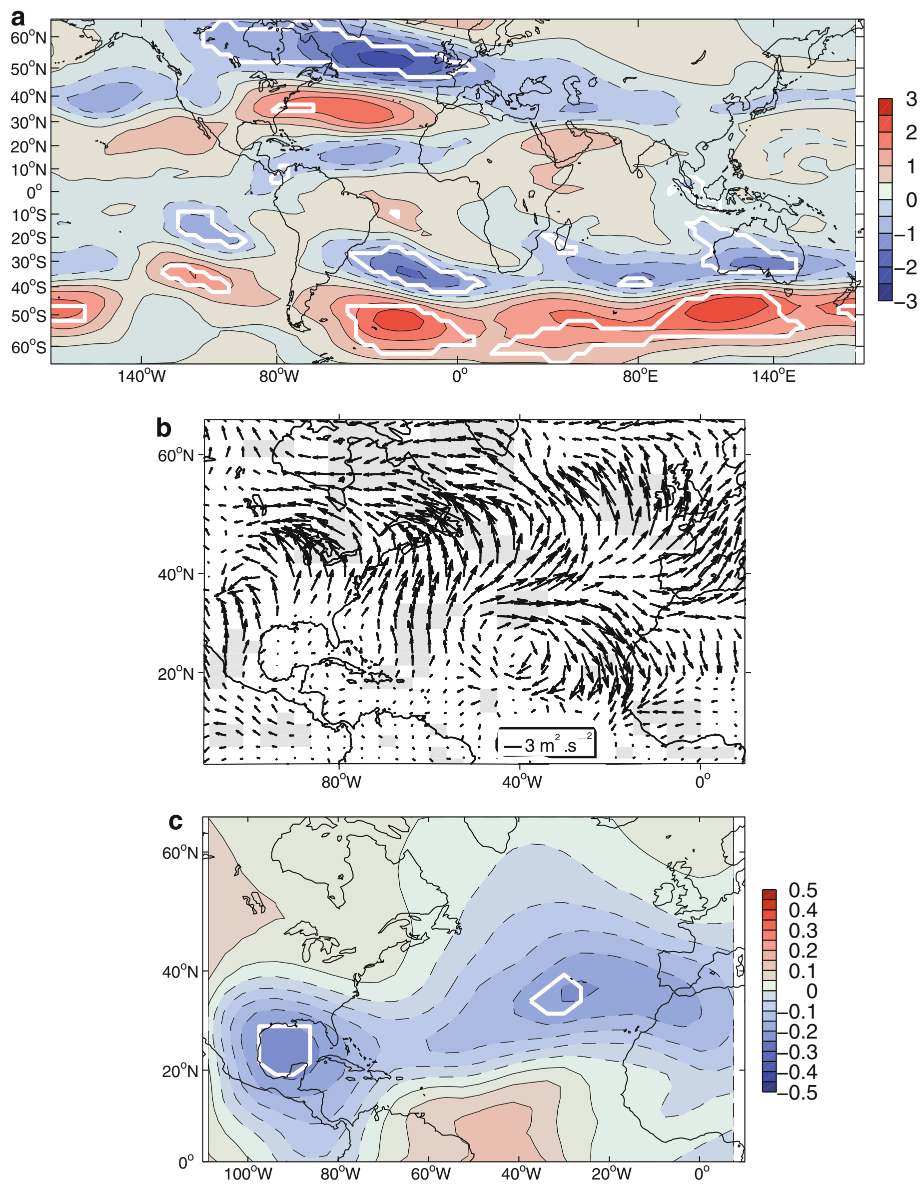


Fig. 18 **a** DJF zonal wind anomalies at 200 mb (in m s^{-1}). **b** DJF **EP** vector anomalies at 200 mb. The scale is given. **c** DJF velocity potential anomalies at 200 mb (in 10^6 kg s^{-1}). All the anomalies correspond to the response to the SST anomaly in Fig. 3b. The *white contour* in **a** and **c** and the *grey shading* in **b** indicate the 5% significance level

MOC might also have an influence on the atmospheric circulation via changes in the sea-ice extent. Conil and Li (2005) showed that although the SST forcing dominates, significant non-linear changes appear in response to sea ice changes in LMDZ. The importance of sea-ice forcing was also underlined by Deser et al. (2004) in the NCAR model. The atmospheric response to the MOC should thus be better investigated by using both SST and sea-ice forcings.

During summer, the response to the North Atlantic SST anomaly that corresponds to an increase in the MOC is dominated by a large scale tropospheric warming associated with global-scale zonally elongated anticyclonic anomalies. The storm track activity is enhanced in the North Atlantic, and is associated with a northward shift and an intensification of the mean westerlies. Our results suggest that the divergence and convergence of eddy vorticity flux leads to the growth and maintenance of the zonally elongated anticyclonic anomalies south of the storm tracks, and the cyclonic anomalies north of the storm tracks, leading to a pattern that projects on the EAP and is best seen when the zonal mean of the geopotential response is removed. The largest signal was detected in the North Atlantic with negative anomalies over Greenland, projecting on the EAP. This response is linked to the occurrence of a stationary wave anomaly over Greenland, and is consistent with the anomalous cyclonic circulation in the transient eddy response and the associated wave propagation. The atmospheric response to the MOC-induced SST changes is thus likely first created by anomalous stationary wave activity and then reinforced and maintained by transient eddies. Changes in the stationary wave activity were found in other atmospheric models. Drévillon et al. (2001) found an increased stationary wave activity over the North Atlantic, forced by a summer SST anomaly, that subsequently led to an anticyclonic anomaly over Europe during winter, which was amplified and maintained by the Atlantic storm track. The dominant influence of transient-eddy activity has been found in other atmospheric models (e.g. Peng et al. 2003; Deser et al. 2004) but the focus was on the winter season. Here we have shown that similar mechanisms are likely to occur during summer, albeit with a weaker intensity.

The response to the North Atlantic SST anomaly extends beyond the North Atlantic-European region and is global in summer, which suggests that changes in the Atlantic ocean could be an important driver of multidecadal variability on a global scale, as argued by Latif et al. (2006). An increased atmospheric convection detected in our experiments over the Indo-Asian continent during summer suggests a potential propagation of the anomalies from that region to the North Atlantic, through Rossby waves propagation. Several observational and model studies showed that even a small perturbation in the region of strong mean convection in the Indo Pacific region could generate an atmospheric response with a substantial zonal mean component (Lau et al. 2005; Fu et al. 2006; Cassou et al. 2009). The link between the Atlantic decadal variability and the summer Indian monsoon has been further suggested in previous model and observational studies (e.g. Zhang and Delworth 2006; Goswami et al. 2006). However, further experiments are needed to determine the physical mechanisms that allow the SST to influence the Asian atmospheric circulation, and to more firmly establish the role of the latter in affecting in return the Atlantic and global responses.

The sensitivity experiments emphasized the non-linearity of the summer atmospheric response with respect to the sign of the SST anomaly. We showed that both warm and cold phases of the SST anomaly, corresponding to an enhanced and reduced MOC, respectively, lead to an EAP-like signal of the same polarity. The temperature response scaled linearly with the sign of the forcing but the geopotential one was non-linear. This result is consistent with previous studies of the extratropical atmospheric response to SST anomalies, in particular that of Conil and Li (2005) who used a different version of the same AGCM. We emphasized that the asymmetry in the zonal mean temperature and geopotential anomalies leads to the same zonal mean zonal wind response in the midlatitudes, where the storm tracks develop. The zonally asymmetric atmospheric response which results from the interaction between the transient eddies and the mean flow was found therefore to be very similar whatever the sign of the SST anomaly.

We have also investigated the atmospheric response to MOC changes during winter. We found a significant NAO-like response in the North Atlantic that extend zonally to the whole northern hemisphere, projecting on the Arctic Oscillation. Our experiments revealed statistically significant alterations of the North Atlantic storm tracks during winter, with an intensified subtropical jet and a weaker subpolar jet. We showed that the Eliassen–Palm vector had a dipolar structure suggesting that transient eddies tend to accelerate the mean flow between 30°N and 40°N, and decelerate it to the north. The prescribed SST anomaly induces a NAO-like response mainly through an eddy feedback mechanism by perturbing the North Atlantic storm tracks. The response is consistent with that seen in the IPSL coupled model, even though the latter was not robust. It is also in agreement with a large number of GCM experiments which showed a prominent influence of transient-eddy feedbacks during winter, and a key role of the transients in the maintenance of the atmospheric response (Peng and Robinson 2001; Kushnir et al. 2002; Peng et al. 2003). Consistently with the large influence of baroclinic instabilities, we showed that the winter response is highly non-linear and projects on a negative phase of the NAO for both polarities of the SST anomaly. The strong non-linearities explain the difficulty to detect the winter signal in the coupled simulation using linear methods like regression. As in summer, the temperature response is relatively symmetrical with respect to the sign of the forcing and the zonal mean zonal wind anomalies keep the same sign in the storm track development region. The non-linearity of the Z500 response is then consistent with the eddy-mean flow interaction, although the mechanisms of such a strong non-linearity has not been identified. A significant response was also found over the North Pacific with a weakening of the Aleutian low forced by an MOC-induced warming of the North Atlantic. A similar response was found in other AGCMs forced by an AMO-like SST anomaly (Schubert et al. 2009), and the recent study of Kushnir et al. (2010) suggests that it is forced by the tropical Atlantic. This Atlantic-Pacific teleconnection was also found in other coupled models in response to a weakening of the MOC, hence with the sign reversed, i.e. a cooling of the North Pacific associated with a deepening of the Aleutian low (Zhang and Delworth 2005; Timmermann et al. 2007). Although Okumura et al. (2009) showed that both oceanic and atmospheric pathways could allow MOC-induced North Atlantic changes to impact remotely on the North Pacific, only an atmospheric pathway can explain the teleconnection in our slab model experiments. The winter atmospheric response in the forced AGCM case without a slab ocean model is not significant and does not project on the NAO in the North Atlantic. It suggests that the use of an interactive ocean model is needed to detect a response in winter. Further work would be needed to investigate the difference between the forced and slab configurations and identify the physical processes that shape the atmospheric response and are missing in the forced case in this model.

The atmospheric response in the slab model shows some similarities with that in other AGCMs in response to the observed AMO (Sutton and Hodson 2007; Hodson et al. 2009). The largest signal-to-noise ratio is found during summer with cyclonic anomalies over the North Atlantic European region, in agreement with our study. The precipitation response shows less agreement in particular over the Indian Monsoon region, that is poorly represented in most AGCMs. The winter response in HadAM3 also projects onto the NAO, but with the opposite phase compared to the slab model. The winter atmospheric response shows strong non-linearities in both models.

The role of the tropics differs between models. Hodson et al. (2009) and Kushnir et al. (2010) argue that the atmospheric response to AMV is largely forced by the tropics. The SST anomaly in our experiments is very weak in the tropics, so the response is largely driven by the extratropics even if a weak influence of the tropics cannot be excluded. Our results in the subtropics are however consistent with the Gill-type mechanism proposed by Hodson et al. (2009) and Kushnir et al. (2010). The discrepancies in the extratropics might be the result of different climatologies leading to different eddy-mean flow interactions (Peng et al. 1995), and of differences in the prescribed SST anomalies. Indeed, although the SST anomaly used in our experiments shows some similarities with the model AMO, we have not investigated the atmospheric response to the AMO. The SST pattern associated with the observed AMO and used by Sutton and Hodson (2007) and Hodson et al. (2009) shows larger anomalies in the tropical Atlantic than the SST following the MOC.[2] Correspondingly, the summer atmospheric response identified by Sutton and Hodson (2007) is baroclinic and does not project onto the NAO or the EAP, whereas in our slab model, it is equivalent barotropic and resembles the EAP.

The atmospheric response described in this study was detected prescribing a 2°C SST anomaly and using a 80-year-long simulation. Given the weak signal-to-noise ratio, it is useful to estimate the ensemble size that would be required to detect a significant response to specified SST anomalies that have a realistic magnitude. Assuming a normal distribution, the mean increases as the square root of the sample size (von Storch and Zwiers 1999). To detect a similar signal-to-noise ratio using a 0.4° SST anomaly, which has a 5% chance to be observed for a Gaussian distribution, we would need an ensemble size 25 times larger, thus a 2000-year long simulation. As a model study, the extent to which the proposed mechanisms accurately reflect the actual behavior of the observed coupled system remains uncertain. Additional multi-model comparisons using the SST forcing associated with the MOC are needed to identify the robust results.

[2] This is not a particularity of the IPSL model as in several climate models the surface signature of the MOC is strongest at high latitudes and weak in the tropics.

Acknowledgments The authors thank Christophe Cassou, Jérome Sirven, Isaac Held, Ying Li, and two anonymous reviewers for useful comments on earlier versions of this manuscript. Computer resources were allocated by IDRIS, the computer center of the CNRS in Paris. The research leading to these results has received funding from the European Community's 7th framework programme (FP7/2007-1013) under grant agreement No. GA212643 (THOR: "Thermohaline Overturning-at Risk", 2008–2012. Support to CF from the Institut Universitaire de France, is gratefully acknowledged.

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