Climate Response to Basin-Scale Warming and Cooling of the North Atlantic Ocean

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ABSTRACT

Using experiments with an atmospheric general circulation model, the climate impacts of a basin-scale warming or cooling of the North Atlantic Ocean are investigated. Multidecadal fluctuations with this pattern were observed during the twentieth century, and similar variations—but with larger amplitude—are believed to have occurred in the more distant past. It is found that in all seasons the response to warming the North Atlantic is strongest, in the sense of highest signal-to-noise ratio, in the Tropics. However there is a large seasonal cycle in the climate impacts. The strongest response is found in boreal summer and is associated with suppressed precipitation and elevated temperatures over the lower-latitude parts of North and South America. In August–September–October there is a significant reduction in the vertical shear in the main development region for Atlantic hurricanes. In winter and spring, temperature anomalies over land in the extratropics are governed by dynamical changes in circulation rather than simply reflecting a thermodynamic response to the warming or cooling of the ocean.

The tropical climate response is primarily forced by the tropical SST anomalies, and the major features are in line with simple models of the tropical circulation response to diabatic heating anomalies. The extratropical climate response is influenced both by tropical and higher-latitude SST anomalies and exhibits nonlinear sensitivity to the sign of the SST forcing. Comparisons with multidecadal changes in sea level pressure observed in the twentieth century support the conclusion that the impact of North Atlantic SST change is most important in summer, but also suggest a significant influence in lower latitudes in autumn and winter.

Significant climate impacts are not restricted to the Atlantic basin, implying that the Atlantic Ocean could be an important driver of global decadal variability. The strongest remote impacts are found to occur in the tropical Pacific region in June–August and September–November. Surface anomalies in this region have the potential to excite coupled ocean–atmosphere feedbacks, which are likely to play an important role in shaping the ultimate climate response.

1. Introduction

Because of its large heat capacity and slow movement the ocean plays a central role in low-frequency climate variability. The role of the Atlantic Ocean is of particular interest because the North Atlantic is host to one of the few regions of deep-water formation on the planet, and therefore plays a vital role in the overturning circulation, which is responsible for a large fraction of the poleward heat transport accomplished by the oceans. There is evidence from palaeoclimate records that the Atlantic meridional overturning circulation (MOC) has undergone large, and sometimes rapid, changes (e.g., McManus et al. 2004). These changes are thought to have been triggered by releases of freshwater at high northern latitudes (Vidal et al. 1997), and to have caused major changes in climate such as a widespread cooling of the North Atlantic region (e.g., Vellinga and Wood 2002).

In the twentieth century, North Atlantic sea surface temperatures (SSTs) exhibited prominent multidecadal fluctuations with alternating warm and cool phases (Bjerknes 1964; Folland et al. 1986; Parker et al. 1991; Kushnir 1994; Schlesinger and Ramankutty 1994; Mann and Park 1994; Delworth and Mann 2000; Enfield et al. 2001; Sutton and Hodson 2005, hereafter SH05; Knight et al. 2005). Although small by comparison with palaeoclimate signals, the amplitude of these variations is large by comparison with interannual variability, and it

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has been suggested (e.g., Delworth and Mann 2000; Knight et al. 2005) that variations in the MOC were responsible. Kerr (2000) coined the phrase "Atlantic Multidecadal Oscillation" (AMO) to describe the idea of a persistent multidecadal signal in Atlantic sector climate. If the AMO is indeed a real phenomenon driven by variations in the MOC, then it is an aspect of the internal variability of the climate system. However, it is likely that Atlantic sea surface temperatures have also been influenced by changing external forcings. Rotstayn and Lohmann (2002) argued that, during the twentieth century, a combination of anthropogenic sulfate aerosol and greenhouse gas forcing could have forced an interhemispheric SST contrast that is similar, in the Atlantic basin, to the dipolar pattern associated with changes in the MOC. It is also likely that the increase in greenhouse gases has contributed to the most recent warming of the North Atlantic Ocean.

Whatever their cause, there is observational evidence that the recent multidecadal fluctuations in Atlantic SST were associated with important climate anomalies. Johannessen et al. (2004), following Delworth and Knutson (2000), showed that during the 1930s and 1940s, associated with warm conditions in the North Atlantic Ocean, surface air temperatures at northern latitudes (especially poleward of 60°N) were enhanced by up to 1°C in the zonal mean and more than 3°C locally in winter. Enfield et al. (2001) and McCabe et al. (2004) showed that the warm conditions in the North Atlantic are associated with reduced summer rainfall and increased drought frequency over much of the United States. AMO-like variations in Atlantic SSTs have also been implicated in the decadal variability of Sahel rainfall (Folland et al. 1986) and hurricane activity (Goldenberg et al. 2001). Last, Delworth and Mann (2000), following Kushnir (1994), showed evidence of AMO-related variations in sea level pressure (SLP)—and hence atmospheric circulation—in the North Atlantic region.

Other evidence that changes in the North Atlantic Ocean may affect climate comes from coupled models studies. For example, MOC shutdown experiments (e.g., Vellinga and Wood 2002; Dong and Sutton 2002) consistently lead to an interhemispheric contrast in surface air temperature anomalies and also have large impacts on precipitation, particularly in the Tropics. The climate impacts are not restricted to the Atlantic region but are experienced around the globe within a matter of years (Dong and Sutton 2002; Zhang and Delworth 2005). The impacts of internally generated fluctuations in the MOC have also been studied in control integrations with coupled models (e.g., Dong and Sutton 2005; Knight et al. 2005, 2006). For example, Knight et al. 2006 show evidence of significant impacts on rainfall in the Nordeste region of Brazil and in the Sahel.

Observational and coupled model studies sometimes suffer from the limitation that they show correlation rather than causality. Studies with atmospheric models are useful because they isolate the specific role of changes in SST. We (Sutton and Hodson 2003, hereafter SH03) analyzed an ensemble of atmospheric GCM integrations forced with observed global SST data for the period 1871–1999. Using an optimal detection methodology we found that in all four standard seasons [December–February (DJF, hereafter 3-month periods are denoted by the first letter of each respective month) MAM, JJA, and SON] the leading mode of ocean-forced multidecadal variability in the North Atlantic region was associated with an AMO-like pattern of SST. More recently we (SH05) performed experiments in which the same atmosphere model was forced with an idealized AMO SST pattern, describing a basinwide warming or cooling of the North Atlantic Ocean. Our analysis focused on the boreal summer season, and by careful comparison between the model results and observations we were able to demonstrate that changes in the Atlantic Ocean during the twentieth century were an important driver of multidecadal variations in the summertime climate of both North America and western Europe. The impacts on summer precipitation over North America were consistent with those identified in analyses of observations by Enfield et al. (2001) and McCabe et al. (2004).

This paper extends SH05 to consider all four seasons. We aim to address the following questions:

- What are the local and remote impacts on climate of a basinwide warming or cooling of the North Atlantic Ocean, and how do these impacts vary seasonally?
- What mechanisms govern the climate impacts and what are the roles of tropical versus midlatitude sea surface temperature anomalies?
- What are the implications of any findings for identifying and understanding the role of the Atlantic Ocean in climate variability and change?

The structure of the paper is as follows. In section 2 the model experiments and analysis procedure are described. The simulated responses to warming or cooling the North Atlantic Ocean are presented in section 3, and in section 4 these responses are compared with twentieth-century observations. These comparisons focus (as in SH05) on the contrast between the North Atlantic warm period, 1931–60, and the subsequent cool period, 1961–90. In section 5 the mechanisms re-

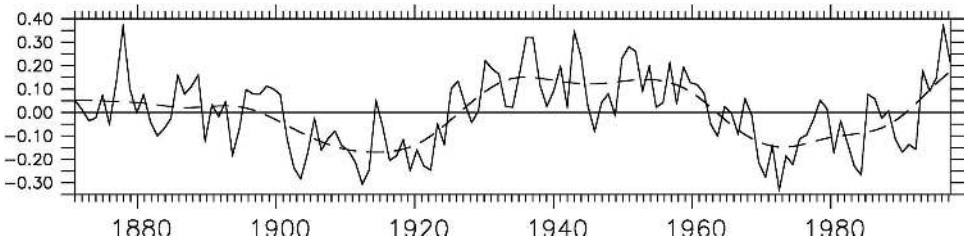

FIG. 1. Index of the North Atlantic SSTs from 1871 to 2003. The index was calculated by averaging annual mean HadISST Rayner et al. (2003) SST observations over the region 0°–60°N, 75°–7.5°W (solid line) and then removing the long-term mean. Low-pass filtering using a 37-point Henderson filter (Kenny and Durbin 1982) results in the dashed line. Both indices have been detrended. The units on the vertical axis are °C. This low-pass index explains 53% of the variance in the detrended unfiltered index and is very similar to that shown in Enfield et al. (2001) and SH05.

sponsible for generating the local and remote climate impacts are considered in more detail. Conclusions and implications are presented in section 6.

2. Experiments and analysis

Figure 1 illustrates the observed recent multidecadal variations in North Atlantic SST in the Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) dataset (Rayner et al. 2003). It shows an index (1871–1999) created by averaging annual mean SST over the region 0°–60°N, 75°–7.5°W. A low-pass-filtered version of the time series is also shown. The multidecadal fluctuations have large amplitude by comparison with the interannual variations and are characterized by a cool phase in the early twentieth century, a warm phase between about 1930 and 1960, a further cool phase from about 1960 to 1990, and a warming at the end of the century. The low-pass-filtered time series is extremely similar if derived from seasonal mean (e.g., summer or winter) rather than annual mean data, suggesting that air–sea interactions in summer are insufficient to erase from the summer mixed layer the signature of longer-term oceanic memory.

The aim of our study is to investigate the impacts on climate of a basin-scale warming or cooling of the North Atlantic Ocean. To do so, we carried out experiments in which an atmospheric general circulation model was forced by idealized SST patterns representative of the multidecadal fluctuations shown in Fig. 1. The model we used was a version of the U.K. Hadley Centre Atmospheric Model version 3 (HadAM3; Pope et al. 2000). HadAM3 employs an Arakawa B grid with a horizontal resolution of 2.5° latitude × 3.75° longitude and 19 hybrid levels in the vertical. The model generally compares well with observations in terms of its mean climate, although there are of course some biases; for example, a high pressure bias at high latitudes. Other studies suggest that the variability of MSLP in the model is comparable to (perhaps somewhat weaker than) observations both when HadAM3 is forced by SST (Rodwell and Folland 2002) and when it is coupled to an ocean model (Collins et al. 2001).

The first SST forcing pattern, NA, is shown in Fig. 2a. It was formed by regressing annual mean HadISST data onto a standardized version of the low-pass-filtered index show in Fig. 1. If seasonal rather than annual mean SST data are used, then local regression coefficients typically vary by 10%–30%. We chose to use a seasonally invariant pattern because the large-scale multidecadal signal is common to all seasons. The North Atlantic part of the pattern was isolated by applying a mask of weights with weight zero outside of the Atlantic, poleward of 70°N, and in most of the South Atlantic. The zero weighting poleward of 70°N was employed because variations in sea ice extent are likely to be particularly important in this region and, except for recent decades, these variations are very uncertain. A cosine-squared smoothing was applied to the edges of the mask to prevent discontinuities in SST gradient. Finally the pattern was multiplied by a factor of 4, so the amplitude of the anomalies corresponds to 4 times the observed standard deviation. However the results were subsequently scaled by a factor ¼ to facilitate comparison with twentieth-century observations. This scaling procedure enabled us to estimate the linear response to forcing with shorter integrations than would otherwise have been required. It does, however, make more difficult the comparison with observations of nonlinear responses, a point we will discuss later.

We first carried out a control experiment (CTRL) by forcing the model with a monthly mean climatological SST field derived by averaging the HadISST data for the period 1961–90. We then carried out experiments in which the NA SST anomaly was added to (NA^+) or

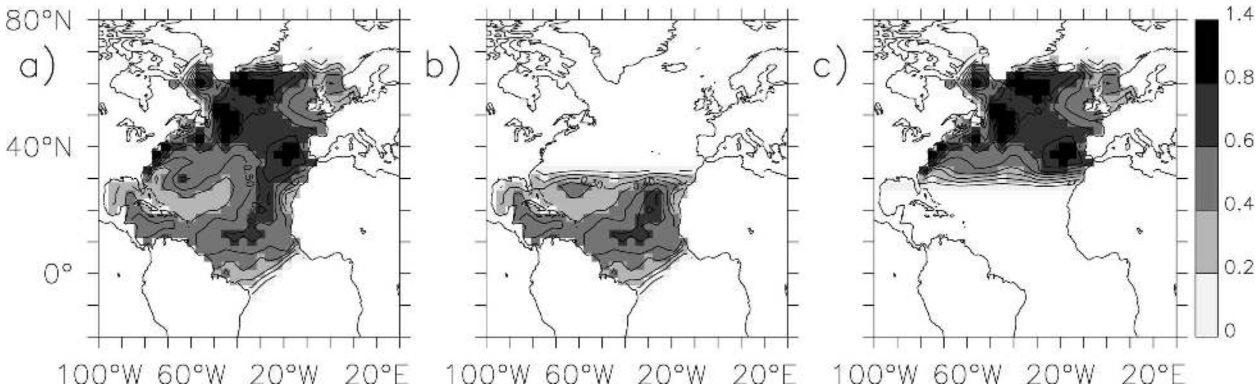

FIG. 2. SST Anomalies (°C): (a) NA^+SST pattern; (b) TNA^+SST pattern; and (c) XNA^+SST pattern. See section 2 for explanation of how the patterns were generated.

subtracted from (NA^-) the climatology. The control experiment was integrated for 40 yr of which the first 5 yr were discarded. The anomaly experiments were integrated for 20 yr of which the first year was discarded. Seasonal means were then computed by averaging three-month periods (DJF, MAM, JJA, and SON) in each year. Finally, seasonal mean time means were calculated by averaging each season over the period of each experiment. (Assuming that each year is statistically independent, this is equivalent for the anomaly experiments to an ensemble mean with a 19-member ensemble.) The temporal standard deviation was also computed for each season and experiment.

The linear response to the forcing is defined as the difference between the time means for the NA^+ and NA^- experiments, scaled by the factor of ¼. A *t* test was used to identify regions where the response was significant. For some fields, a signal-to-noise ratio, defined as the mean response divided by the interannual variability, was also computed. The interannual variability was estimated as

$$\sigma = \left(\frac{\sigma^2_{NA^+} + \sigma^2_{NA^-}}{2}\right)^{(1/2)}.$$

Unlike the *t* test the signal-to-noise ratio is independent of the duration of integrations (i.e., effective ensemble size) and provides a useful, physically relevant, measure of the response strength.

To determine which parts of the NA SST pattern have the greatest impact on climate we performed additional experiments using either the tropical part of the pattern (10°S:30°N; Fig. 2b; TNA^+ and TNA^-) or the extratropical part of the pattern (30°N:70°N; Fig. 2c; XNA^+ and XNA^-). A cosine-squared smoothing was applied at 30°N to form these SST anomalies. All the experiments were again integrated for 20 yr.

We compare the results from North Atlantic SST anomaly experiments with observed decadal changes in sea level pressure as represented in the Hadley Centre Sea Level Pressure dataset (HadSLP1). HadSLP1 provides mean sea level pressures between 1871 to 1998, on a 5° × 5° grid (Basnett and Parker 1997).

3. Results

a. *Response to a warming of the North Atlantic*

Figure 3 shows the response to NA^+–NA^- in SLP, precipitation, and 2-m air temperature (T2m). T2m is shown only over land. All seasons show low pressure over the warm North Atlantic, but the pattern of anomalies has a large seasonal cycle, particularly in the extratropics. In DJF and MAM the largest SLP anomalies are found in the extratropics, but in all seasons the highest signal-to-noise ratios are in the Tropics (with peak values >1.5 in JJA and SON). In the extratropics signal-to-noise ratios are generally much lower, indicating that the influence of the anomalous ocean conditions is weak by comparison with atmospheric internal variability. Poleward of 60°N the highest signal-to-noise ratios are seen in DJF and MAM, but the ratios here are still less than 0.5.

As discussed in SH05 the SLP response for JJA shows two major low pressure centers, one situated over southern North America and the other situated over western Europe. The center over North America lies northwest of the highest signal-to-noise ratios. Close inspection shows that this center splits into two, with eastern and western lobes. The SLP response over North America in SON has a similar two-lobed struc-

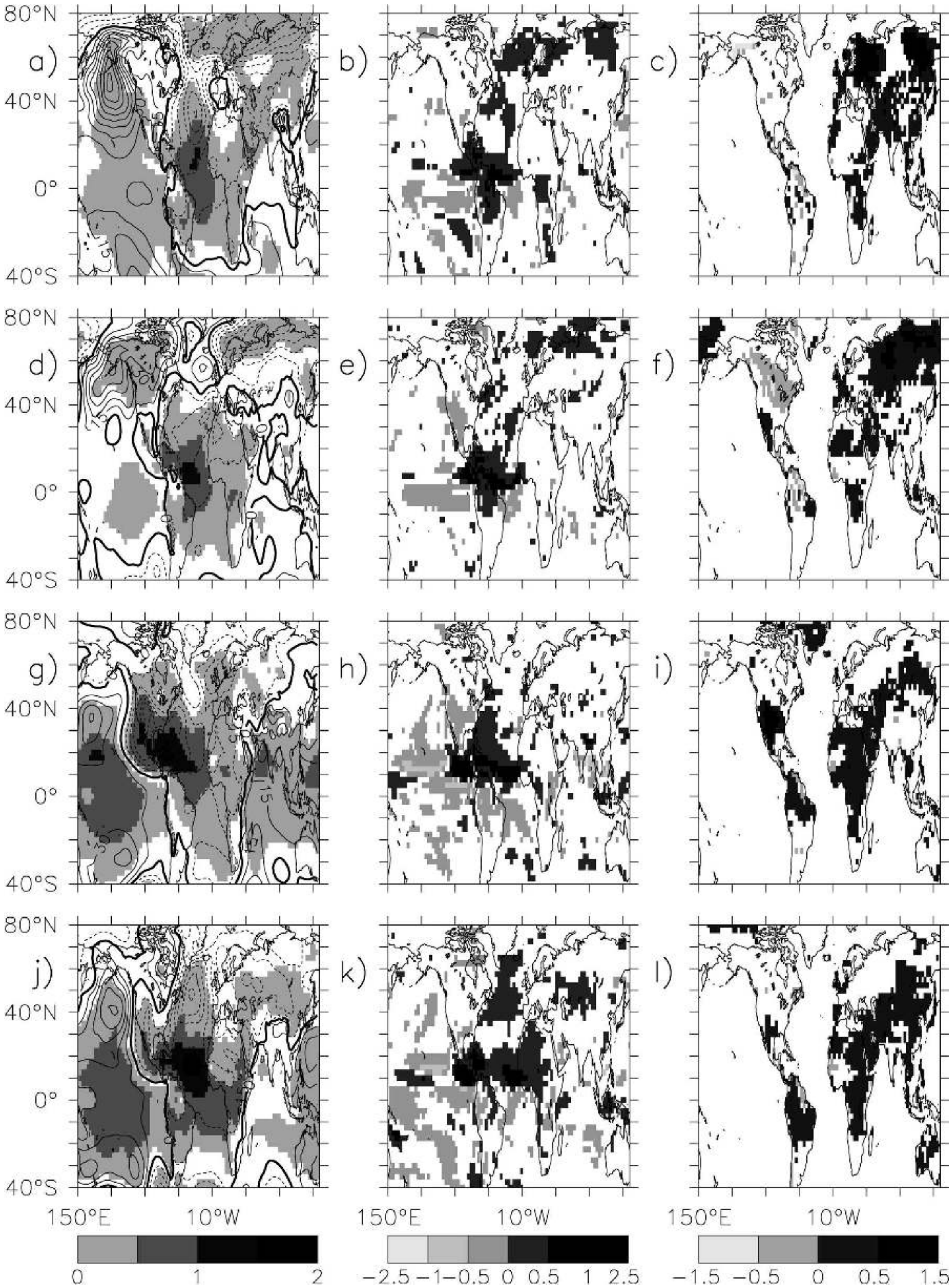

FIG. 3. Atmospheric response to warming of the North Atlantic (NA^+–NA^-). The panels show differences between seasonal mean time means of simulations with the HadAM3 model forced with the NA^+ and NA^- patterns. (As explained in section 2, results have been multiplied by 0.25 to aid comparison with Fig. 7.) Mean sea level pressure anomalies for (a) DJF, (d) MAM, (g) JJA, and (j) SON. (Contours in Pa with an interval of 15 Pa, shading indicates signal-to-noise ratio.) (b), (e), (h), (k) Same as in (a), (d), (g), (j), but for precipitation anomalies (mm day^{-1}). (c), (f), (i), (l) Same as in (a), (d), (g), (j), but for land surface air temperature anomalies (°C). Note that the scales for both precipitation and temperature are nonlinear. In all panels, regions where anomalies are not significant at the 95% level are shaded white. The fraction of significant coverage is greater than 17% in all panels.

ture, but the eastern lobe is more pronounced and reaches much farther north over the Atlantic Ocean. In DJF the SLP response features a low pressure center over the midlatitude North Atlantic Ocean and a second center that reaches eastward from the Greenland Sea over the Barents Sea and northern Europe. Two similarly located low pressure centers are also seen in MAM, but the midlatitude center is weaker than in DJF. Comparison with Fig. 2 suggests that the anomalous meridional temperature gradient at the northern boundary of the SST anomaly (60°–70°N) might play a role in forcing the higher-latitude SLP anomalies.

The SLP fields show a significant remote response outside of the Atlantic basin. Low pressure anomalies are found over Africa and parts of Asia (particularly in SON and DJF). Regions of high pressure are prominent over the Pacific and, in MAM, over northwestern North America. In JJA, as noted in SH05, the remote response extends throughout the Tropics, with high pressure anomalies over the whole Indo-Pacific basin.

There is enhanced precipitation over the warm North Atlantic (Fig. 3). The largest anomalies (peak values exceeding 2 mm day^{-1}) are found in the Tropics, particularly in JJA and SON when the underlying mean SSTs are warmest. The close association between the largest precipitation anomalies and the highest signal-to-noise ratios for SLP suggests that anomalous latent heat release is likely to be forcing the circulation anomalies, consistent with the theory of Gill (1980). In DJF and MAM a dipolar pattern of precipitation anomalies in the tropical Atlantic indicates a northward shift of the Atlantic ITCZ. Precipitation in the Pacific is generally suppressed, but there is enhancement near the coast of central America.

Figure 3 also shows impacts on precipitation over land. Over most of northern South America precipitation is enhanced in DJF and MAM, but is suppressed in JJA. North America shows in JJA the prominent negative precipitation anomalies over the United States and Mexico that were discussed in SH05. There are also significant negative anomalies in the western part of this region in MAM. In JJA there is an interesting hint of an impact on the Asian monsoon, with negative precipitation anomalies over India and the Bay of Bengal, and positive anomalies to the North. These anomalies will be discussed further in section 5.

The T2m fields in Fig. 3 show warm anomalies (up to 1.5°C) over parts of northern Eurasia in DJF and MAM. Comparison with the SLP fields suggests that the cause is likely to be enhanced advection of warm maritime air by the anomalously strong geostrophic westerly winds. In westernmost Europe the local influence of the warmer Atlantic SST may also be a factor. Over North America there are no significant anomalies in DJF, but in MAM there is a notable region of cold anomalies.[1] Comparison with the SLP fields indicates northeasterly wind anomalies in this region, which may be bringing cooler air from higher latitudes. It appears, therefore, that the extratropical surface air temperature anomalies in DJF and MAM are substantially controlled by the dynamical circulation response, rather than simply reflecting a thermodynamic response to the warming of the ocean.

In JJA there are warm anomalies over the United States and Mexico, as discussed in SH05. There are also warm anomalies over much of Africa, northern South America, and parts of central Asia. The pattern of anomalies in SON is similar but over North America the anomalies are weaker and less extensive. Along the northeast coast of South America is a strip of cold anomalies that can also be seen in the other seasons (especially DJF and MAM). There is also enhanced precipitation in this region, which suggests that increased cloud cover could be reducing surface solar radiation.

In summary, the atmospheric response to NA^+–NA^- is strongest (in the sense of highest signal-to-noise) in the Tropics and notably weaker in middle and high latitudes. There is substantial seasonal variation with the strongest response in JJA and SON. The tropical response features large, positive, precipitation anomalies over the tropical Atlantic and low SLP. Significant precipitation anomalies are also found over northern South America and southern North America, and there is a tendency for warming over land. The northern extratropical circulation response is relatively weak but is associated with significant temperature anomalies over the northern continents in boreal winter and spring. There is a remote response outside the Atlantic basin, which is most prominent in JJA and SON. High pressure and generally negative precipitation anomalies over the tropical Pacific are the most robust features of the remote response.

b. Relative roles of low-latitude and midlatitude warming

We have seen that the strongest atmospheric response to the NA^+–NA^- forcing is found in the tropical Atlantic region. In this section we investigate whether this response is locally forced by the SST anomalies in the same region and, if so, what the role of the midlatitude SST anomalies is.

[1] A similar region of cold anomalies is seen in the coupled model simulation of the AMO discussed by Knight et al. (2006).

Figure 4 shows the response to forcing by tropical Atlantic SST anomalies alone (i.e., $TNA^+–TNA^-$). Comparison with Fig. 3 shows that the response in the Tropics is extremely similar to that found in the response to $NA^+–NA^-$. This similarity suggests a common mechanism, the nature of which will be discussed in section 5.

In contrast to the Tropics, the response to the TNA forcing over the northern extratropics shows notable differences from the response to NA forcing. The positive T2m anomalies that were seen over northern Eurasia in DJF and MAM are not reproduced with TNA forcing alone, and the SLP patterns differ. Note, however, that with TNA forcing alone there is still a significant circulation response over the extratropical North Atlantic. In DJF there is a large low pressure center over the midlatitude Atlantic. Analyses of 200-hPa streamfunction (not shown) suggest that this anomaly is associated with the northward propagation of stationary Rossby waves from the tropical Atlantic (Hoskins and Karoly 1981).

We also examined the response to forcing by extratropical (midlatitude) Atlantic SST anomalies alone (i.e., $XNA^+–XNA^-$). The responses (not shown) are generally very weak. T2m naturally shows a warming over the region of forcing but almost no significant response over land. The SLP response is significant over the North Atlantic in JJA (as discussed in SH05) but not in the other seasons. Vertical sections of geopotential height anomalies for JJA (not shown) indicate that the response is shallow (significant anomalies are confined below about 700 mb) and may therefore simply reflect a local warming of the boundary layer by the underlying SST anomalies.

The weak response to midlatitude SST anomalies is consistent with previous studies (see Kushnir et al. 2002). However, comparison of Figs. 3 and 4 suggests that in the presence of tropical SST anomalies the midlatitude SST anomalies can exert some influence. For example, the response of northern Eurasian temperatures in DJF and MAM is notably stronger in the NA experiments than in the TNA experiments. The fact that these temperature anomalies are not reproduced by forcing with XNA anomalies alone indicates a nonlinear interaction between the responses to the tropical and midlatitude SST anomalies. The mechanism for this nonlinearity is unclear, but it could be that the tropical forcing perturbs the stationary wave structure over midlatitudes in such a way as to alter (in this instance, increase) its sensitivity to the midlatitude SST anomalies. There is considerable evidence that the response to midlatitude SST anomalies can be very sensitive to changes in the mean state (Kushnir et al. 2002). Last, the fact that XNA forcing does not excite the low SLP anomalies that were seen at high latitudes in DJF and MAM in response to NA forcing (Fig. 3) counters the hypothesis that the anomalous meridional temperature gradient at the northern boundary of the SST anomaly might have played an important role in exciting these anomalies.

c. *Nonlinearity of the response*

We also examined the possibility of nonlinearities associated with changing the sign of the SST forcing. To do this we considered the differences between individual experiments, NA^+ and NA^-, and the control experiment, CTRL. In the case of a linear response we would expect the result of NA^+–CTRL to be equal and opposite to the result of NA^-–CTRL. In the Tropics we found a high degree of linearity in all seasons in the sense that the pattern of anomalies is very similar. However the magnitude of anomalies is generally larger for the positive SST forcing (NA^+–CTRL). The major precipitation anomalies, for example, are 10%–40% larger. Larger anomalies are expected as a consequence of the Clausius–Clapeyron relationship. Over southern North America (15°–45°N, 230°–290°E) the JJA SLP anomaly for NA^+–CTRL is -125 ± 62 Pa, while for NA^-–CTRL the anomaly is 87 ± 34 Pa. Thus the response is stronger in the case of positive SST forcing, but the difference is of marginal statistical significance.

In contrast to the Tropics the response in the northern extratropics is highly nonlinear, especially in SON, DJF, and MAM. Figure 5 shows the results for DJF. In the Tropics there is an SLP dipole between the tropical Atlantic and eastern tropical Pacific that reverses sign between the two panels (Figs. 5a and 5b). By contrast, at around 40°N over the midlatitude Atlantic *both* panels show significant high pressure anomalies. Farther north, NA^+–CTRL shows significant low pressure anomalies whereas NA^-–CTRL shows no significant anomalies.

Figure 5 also shows cross sections of geopotential height in the latitude band 35°–45°N. Over the Atlantic basin NA^+–CTRL has an equivalent barotropic structure whereas NA^-–CTRL has a baroclinic structure. This finding suggests different mechanisms at work. One hypothesis is that NA^+–CTRL is more strongly influenced by stationary Rossby waves propagating from the tropical Atlantic region. Such Rossby waves are expected to have an equivalent barotropic structure in the extratropics and, furthermore, we might expect stronger excitation of these waves in NA^+–CTRL because the positive SST anomalies in NA^+ give rise to

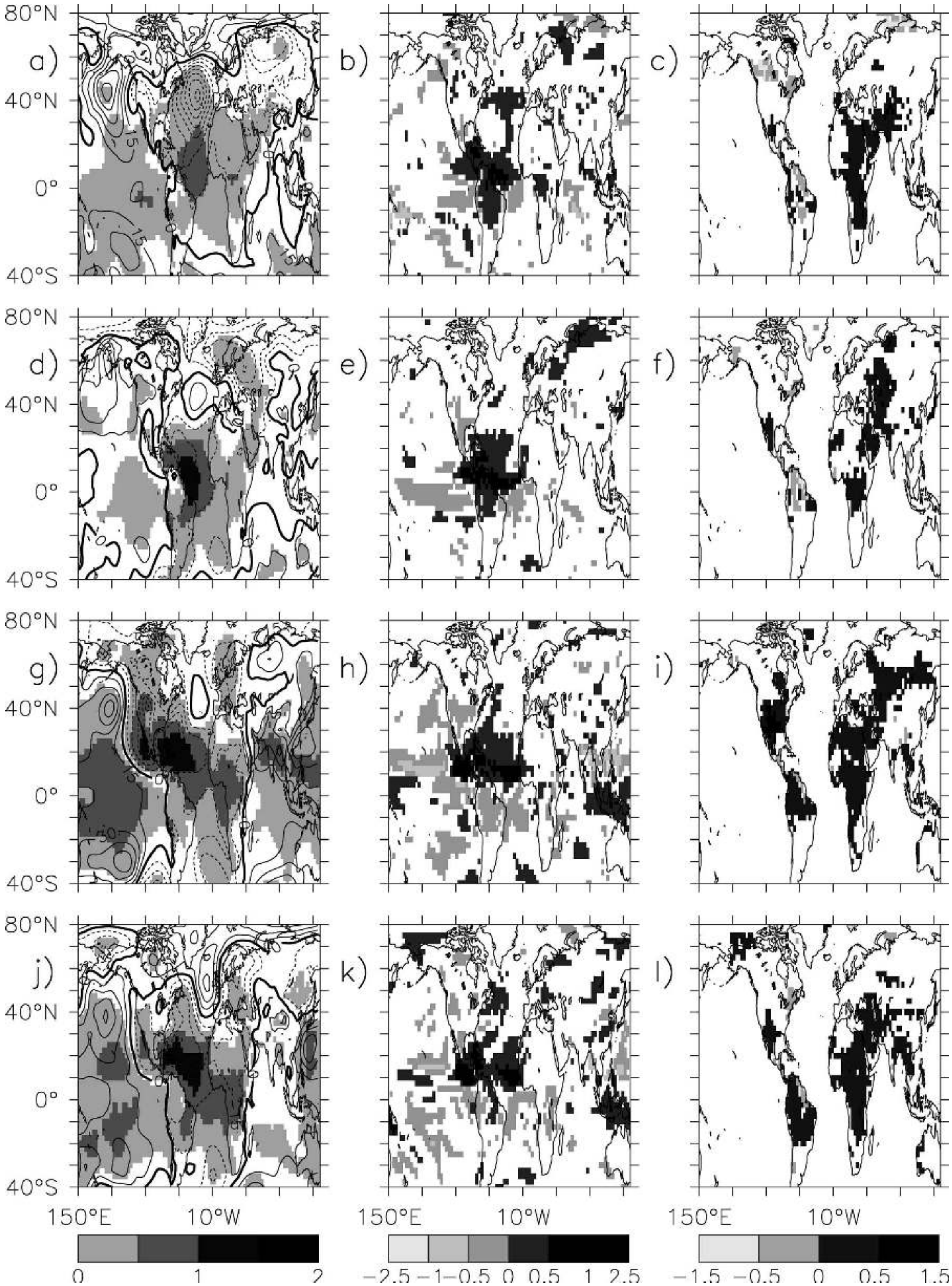

FIG. 4. Atmospheric response to warming of the tropical Atlantic. Same as in Fig. 3, but for TNA^{+}–TNA^{-}. Again, the fraction of significant coverage is greater than 17% in all panels.

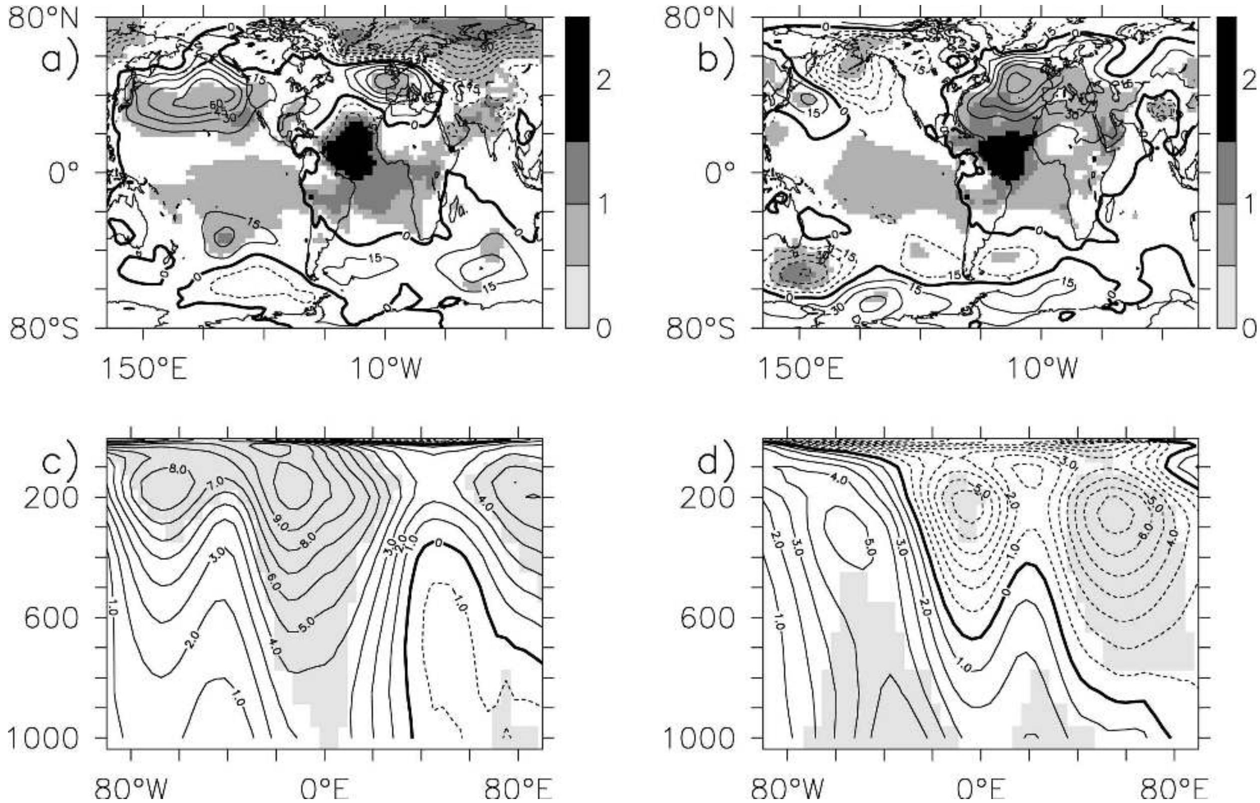

FIG. 5. Nonlinearity of the response to warming or cooling the North Atlantic. (a) Difference in DJF mean SLP between NA^+ and the control experiment (i.e., NA^+–CTRL). Contour interval is 15 Pa; shading indicates signal-to-noise ratio. (b) Same as in (a), but for NA^-–CTRL. (c) Difference in geopotential height averaged between 35° and 45°N for NA^+–CTRL. Contour interval is 1 m. (d) Same as in (c), but for NA^-–CTRL. In all panels, regions where anomalies are not significant at the 95% level are shaded white.

larger latent heating anomalies than the negative SST anomalies in NA^-.

The nonlinearity of the midlatitude atmosphere's response to modest SST anomalies is well known and arises particularly from the important role for changes in the transient eddies in shaping any response (e.g., Kushnir et al. 2002). However, it illustrates the need for great caution in attributing midlatitude climate variations to changing forcing by the North Atlantic Ocean. The results of this study suggest that attempts at such attribution are much more likely to be successful in the Tropics where the signal-to-noise is higher and the responses more linear.

4. Comparison with observations

Because our experiments used idealized patterns of SST forcing, comparison with observations has to be approached with care. In SH05 we made comparisons with simple composite differences of observations between the North Atlantic warm period, 1931–60, and the subsequent cool period, 1961–90. Here we extend this analysis to all four seasons. Differences in the average conditions between these two periods may arise in response to the changes in the North Atlantic Ocean, in response to changes in other ocean basins, as a consequence of variability generated internally in the atmosphere, or as a direct consequence of changing external (natural or anthropogenic) forcings. By a "direct consequence" we mean an impact that is not mediated by changes in the oceans.

Figure 6 shows the SST difference between 1931–60 and 1961–90. As expected it shows positive anomalies in the North Atlantic and the pattern is similar to the NA pattern (Fig. 2a). In particular, the largest anomalies are found in the region ~40°–60°N, and there is a secondary local maximum in the tropical North Atlantic at ~10°N. However, there are also some differences. First, the NA SST anomalies have larger amplitude (especially taking into account the factor 2 associated with the difference between NA^+ and NA^-). As explained in section 2, the results from the NA experiments were scaled to provide a fair (linear) comparison with the observational composite differences. However, nonlin-

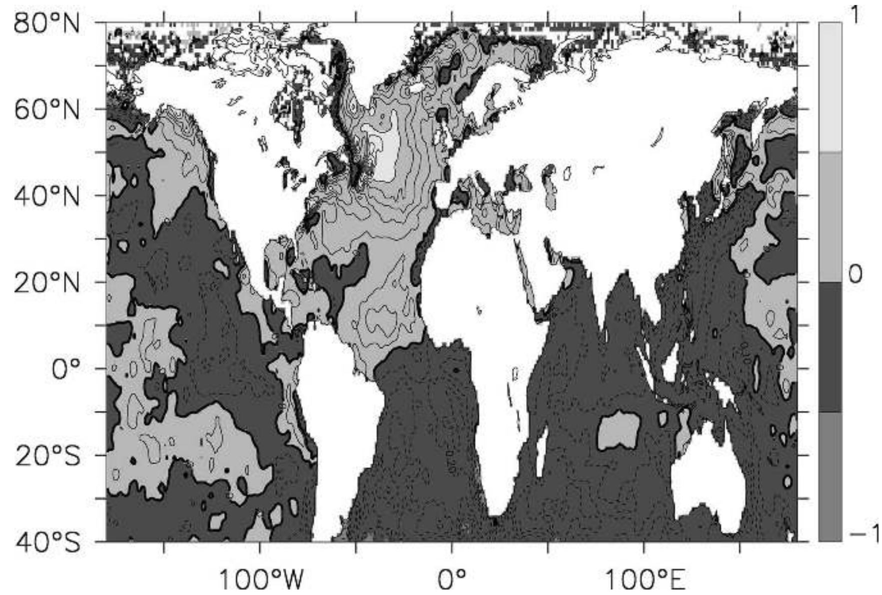

FIG. 6. Composite of annual mean SST (from HadISST). Mean of 1931–1960 minus mean of 1961–1990. Contour interval is 0.1°C.

earities associated with the different amplitude of SST forcing could be a source of differences in the comparisons discussed here. Second, Fig. 6 shows SST anomalies north of 70°N (where the anomalies in the NA pattern are zero), and also a fringe of negative anomalies along the east coasts of Canada and Greenland. These latter anomalies are associated with decreasing sea ice extents in recent decades. Note, however, that because of a lack of data HadISST assumes a climatological sea ice distribution between 1940 and 1952 (Rayner et al. 2003). Hence SST anomalies in these regions must be considered subject to large uncertainty. Third, Fig. 6 shows SST anomalies outside the North Atlantic: in the South Atlantic, Indian Ocean, and western Pacific. Anomalies in these regions could potentially impact SLP over the North Atlantic.

Figure 7 shows the composite difference of observed SLP for (1931–60)−(1961–90) for all four seasons. Recall that we expect the observational results to be noisy because they represent a single realization rather than an ensemble mean. In all seasons there are low pressure anomalies over most of the tropical and midlatitude North Atlantic overlying the warm SST anomalies. This association is consistent with that described using annual mean observational data by Kushnir et al. (1997) and Delworth and Mann (2000).

For DJF the observations show low pressure anomalies at low and midlatitudes and high pressure anomalies at higher latitudes. The anomalies are statistically significant in the tropical North Atlantic and over North Africa. The good agreement with the model results (Fig. 3a) in these regions suggests that these anomalies are a response to the changes in North Atlantic SST. The lack of agreement at higher latitudes is most likely due to atmospheric internal variability.

For MAM the observational composites show a pattern over the North Atlantic that resembles the negative phase of the NAO, and the anomalies in both centers of action are significant. No such pattern is seen in the model results (Fig. 3b). Because the observed anomalies appear to be statistically significant, the difference is unlikely to be explained by internal variability of the atmosphere; it could indicate roles for any or all of the following: differences in the SST patterns shown in Figs. 2a and 6; a direct impact of external forcings; model error; or errors in the observational data.

There is some agreement between the observed and simulated SLP anomalies for MAM. Both show a similar SLP gradient over North America, indicative of anomalous northeasterlies. As was discussed in section 3, in the NA experiments there are cold surface temperature (T2m) anomalies in this region. A composite difference [(1931–60) − (1961–90)] of observational T2m data (Jones and Moberg 2003) also shows negative temperature anomalies in this region (not shown).

In contrast to MAM, the observational composite for JJA shows excellent agreement in the North Atlantic region with the model results (Fig. 3c). The signal-to-noise ratio for the observations is higher over the

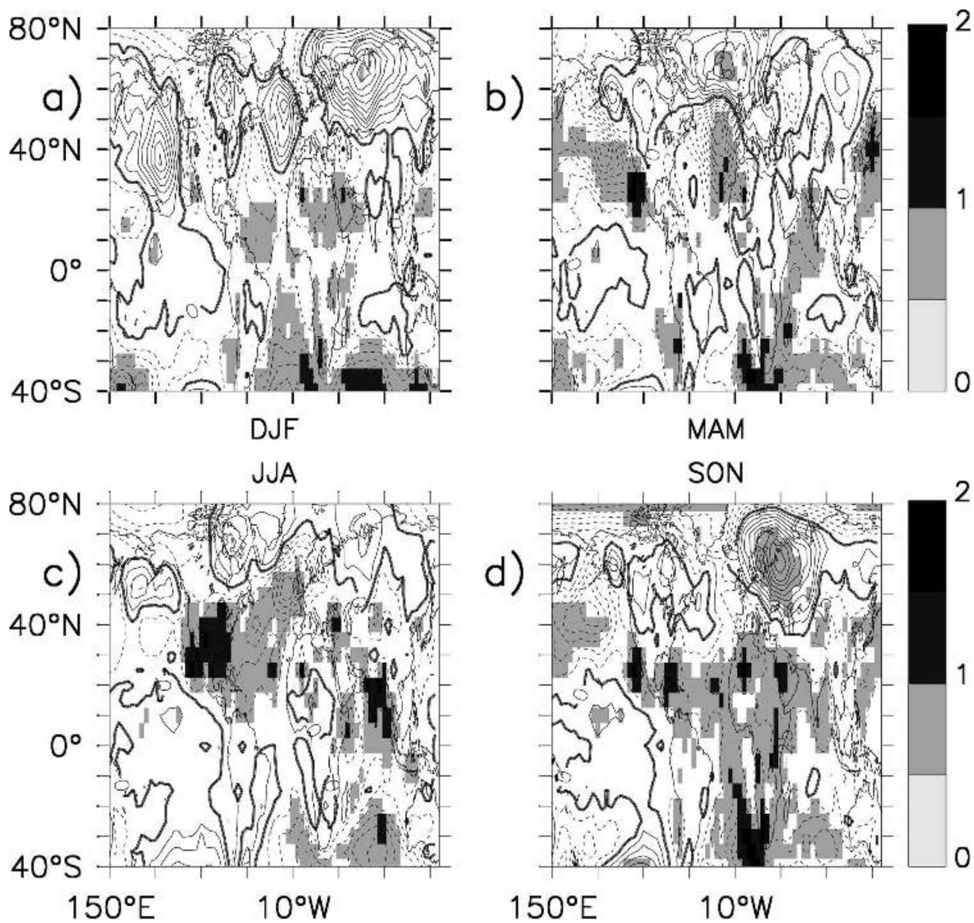

FIG. 7. Observed SLP anomalies for the difference between 1931–60 (warm North Atlantic) and 1961–90 (cool North Atlantic): (a) DJF, (b) MAM, (c) JJA, and (d) SON. Anomalies are computed from the HadSLP observational dataset. Contours are in Pa with an interval of 25 Pa and shading indicates signal-to-noise ratio. In all panels, regions where anomalies are not significant at the 95% level are shaded white. The fraction of significant coverage is greater than 17% in all panels.

United States in JJA than in any other season. As discussed in detail in SH05, the agreement between the observed and simulated anomalies strongly suggests a central role for changes in North Atlantic SST in forcing the observed multidecadal change in summertime circulation. This finding does not negate the evidence from other studies that changes in the tropical Pacific are also an important influence on North American summer climate (e.g., Schubert et al. 2004; Seager 2006, manuscript submitted to *J. Climate*).

The observational composite for SON is quite similar to that for DJF, but, consistent with the model results, the signal-to-nose is higher. In both the model results and observations negative SLP anomalies are found over the tropical North Atlantic and North Africa, suggesting an important role in these regions for forcing by North Atlantic SST. However, the high pressure anomalies seen in the observed composite over central Europe are not simulated.

In summary, the comparison of results from the model experiments with the observational composite differences suggests that changes in North Atlantic SST were an important factor forcing decadal changes in atmospheric circulation in the tropical and midlatitude North Atlantic region. The evidence suggests that changes in North Atlantic SST were most important in JJA (as discussed in SH05) but also exerted a significant influence (at lower latitudes) in SON and DJF. Changes in the North Atlantic Ocean may also have played a role in forcing changes in circulation at higher latitudes, but it is not possible to be conclusive about this suggestion because in our NA experiments the SST forcing was set to zero north of 70°N. Last, it is important to note that—in view of the evidence discussed in section 3c of nonlinearity in the mid- and high-latitude responses—the factor 4 by which we scaled the SST anomalies is likely to diminish their relevance to understanding twentieth-century climate records. However these results may still be relevant to interpretation of proxy climate records for earlier times, when larger amplitude changes in the Atlantic Ocean are believed to have occurred.

5. Discussion

a. Understanding the local climate impacts of SST anomalies in the tropical North Atlantic

The results discussed in section 3 showed that a large part of the climate response to the North Atlantic SST pattern was forced by the SST anomalies in the tropical Atlantic. In this region large precipitation anomalies were found, and it was suggested that the associated latent heating anomalies might be responsible for driving the anomalous atmospheric circulation and associated climate impacts. Here we consider this possibility in more detail. We focus on the JJA season when the response is strongest.

In his classic paper Gill (1980) discussed how the stationary response of the tropical atmosphere to diabatic heating anomalies can be described in terms of stationary Rossby and Kelvin waves. For an off-equatorial heating anomaly the theory predicts a response involving low pressure (at low levels) to the northwest of the heating, associated with a stationary Rossby wave. The pattern of heating implied by the precipitation anomalies shown in Fig. 3h cannot be described as a simple point source, but the SLP response (Fig. 3g)—with low SLP anomalies to the northwest of the main heating region—is very much in line with Gill's simple theory. Furthermore, the fact that in all seasons high signal-to-noise (shown for SLP) is closely associated with high tropical precipitation is entirely consistent with the circulation anomalies being forced by anomalous latent heating.

Figure 8 shows some additional fields for JJA diagnosed from the NA experiments (i.e., NA^{+}–NA^{-}). The 1000-hPa winds (Fig. 8a) show the expected cyclonic anomaly, consistent with low SLP, over southern North America, and also strong southwesterly anomalies over the tropical North Atlantic and eastern Pacific. In Gill's model the vertical structure of the tropical atmosphere is described by a single baroclinic mode, such that circulation anomalies in the upper troposphere are opposite to those in the lower troposphere. Figure 8b shows the 200-hPa streamfunction anomalies from the NA experiments. The pattern of twin upper-level anticyclones indicates that, as predicted by Gill's theory, the upper-tropospheric flow anomalies are indeed opposite to those in the lower troposphere. A similar pattern is found in all seasons (not shown). In view of the many simplifications included in Gill's model the agreement is impressive.

It was noted in section 3 that close inspection of the JJA SLP response (Fig. 3g) shows two lobes rather than a single center of low pressure over North America. These two lobes may reflect the two local maxima in the diabatic heating (as inferred from the precipitation field of Fig. 3h) on the eastern side of the Atlantic basin and in the far eastern Pacific. In addition, interaction between the stationary Rossby wave train and the Rocky Mountains may perturb the response.

Figure 8c shows omega at 500 hPa. The pattern of upward vertical motion corresponds closely with the pattern of enhanced precipitation (Fig. 3h). In addition there is enhanced subsidence in regions of suppressed precipitation, including over North and South America. Because there is no SST forcing in these subsiding regions it is likely that the suppressed precipitation is partly caused by the enhanced subsidence, which is itself a remote response to the extra heating in the regions of enhanced precipitation. Rodwell and Hoskins (2001) showed that the mean diabatic heating associated with the North American monsoon induces descent over the eastern subtropical Pacific. Our results suggest that warming of the subtropical North Atlantic can move some of the heating eastward, and hence move some of the descent onto the American continent. In addition, the remote precipitation anomalies will be influenced by the effect of the circulation anomalies on moisture flux convergence. For example, Ruiz-barradas and Nigam (2005) highlighted the importance of the stationary moisture flux convergence for warm season rainfall over the U.S. Great Plains region. The 1000-hPa wind anomalies (Fig. 8a) suggest that a reduction in the advection of warm moist air from the Gulf of Mexico contributes to the negative precipitation anomalies simulated over the United States.

Consistent with the Gill-type response suggested by Fig. 8, the North Atlantic SST forcing also affects the vertical shear in the main development region (MDR; 10°–20°N) for Atlantic Hurricanes (see Fig. 9). Hurricane formation is favored by warmer SST but is inhibited by high vertical wind shear, which can prevent the axisymmetric organization of deep convection (e.g., Goldenberg et al. 2001). Goldenberg et al. suggested that recent increases in Atlantic hurricane activity result from "simultaneous increases in North Atlantic sea-surface temperatures and decreases in vertical wind shear." Our results (Fig. 9) support the hypothesis that the increases in the Atlantic SST are in fact an impor-

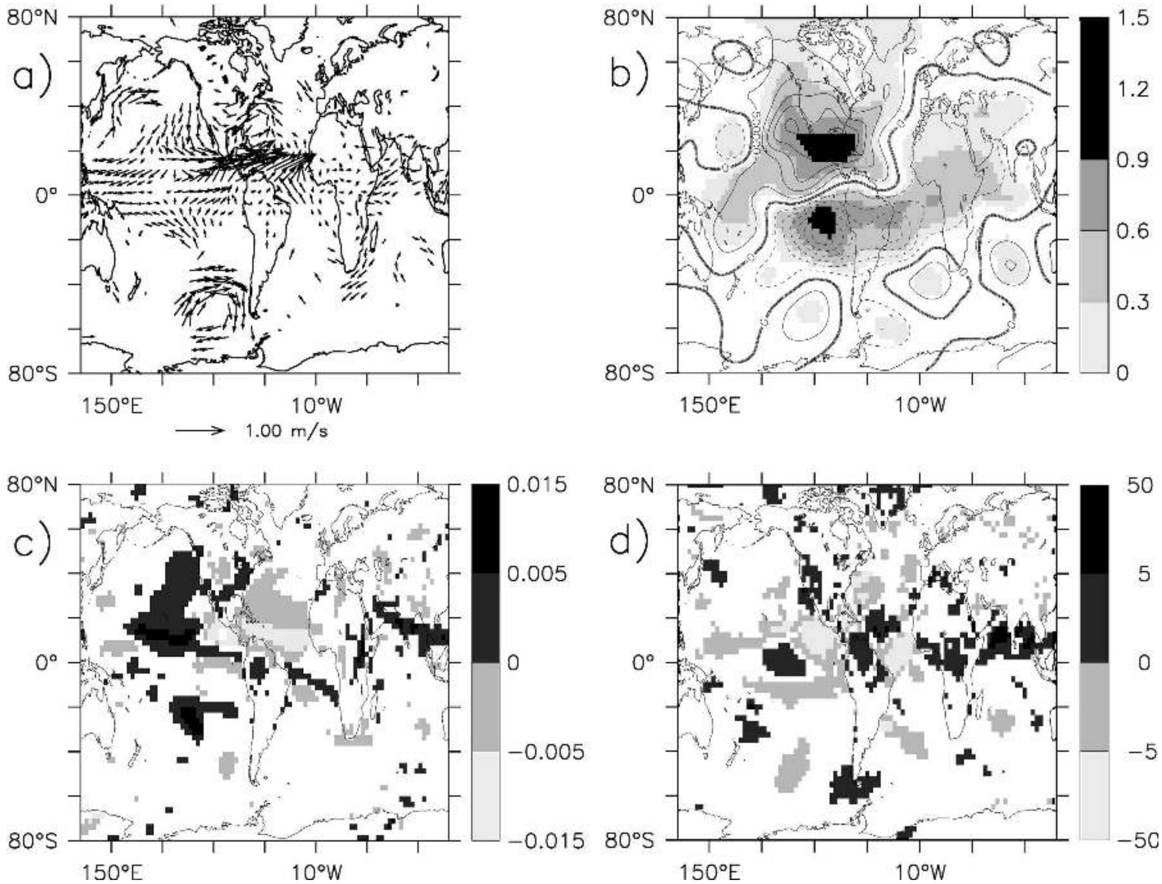

FIG. 8. Atmospheric response to warming of the North Atlantic (NA^+–NA^-) in JJA. Same as in Fig. 3, but for (a) 1000-hPa winds. Only regions where at least one component of the wind is significant at the 95% level are shown. (b) 200-hPa streamfunction. Shaded regions are significant at the 95% level and show signal to noise ratio. (c) Omega at 500 hPa. Units are Pa s^{-1}. Negative values denote ascent. (d) Total heat flux *into* the ocean. Units are W m^{-2}.

tant cause of the reductions in vertical shear (see also Vitart and Anderson 2001; Shapiro and Goldenberg 1998).

In quantitative terms Fig. 9 suggests that the change in North Atlantic SST that was observed between 1931–60 and 1961–90 forced a change in the vertical shear in the MDR of $\sim$1.9 m s^{-1} (based on an average over the whole region; larger anomalies are found on the western side of the MDR). This compares with a mean MDR shear of $\sim$8 m s^{-1} in the control integration. Goldenberg et al. note that shear $>\sim$8 m s^{-1} is unfavorable for hurricane development, so our results clearly imply that the change in shear induced by changing Atlantic SST could have a significant impact on the likelihood of hurricane formation.

b. Understanding the remote climate impacts of SST anomalies in the tropical North Atlantic

As noted in section 3, the response to the North Atlantic SST anomalies extends far beyond the North Atlantic region. This finding suggests that the changes in the Atlantic Ocean could be an important driver of multidecadal variability on a global scale, as hypothesized by Dong and Sutton (2002), who argued that Rossby wave propagation provided a potential route for information about Atlantic changes to propagate efficiently into the Pacific basin.

Figure 3 suggests that the impacts on the Pacific are strongest in JJA and SON. Because in our experiments there is no seasonal cycle in the SST anomalies, the reason must be seasonal evolution of the mean state. Seasonal mean SST in the tropical North Atlantic is highest in JJA and SON, and the peak values are located farther north and west than in the other seasons. In JJA seasonal mean SSTs in the Gulf of Mexico and Caribbean Sea exceed 28°C, and as a consequence small SST anomalies can have a large impact on convection. The more northward location of the precipitation anomalies in JJA and SON may enable Rossby waves, excited by the associated latent heating, to

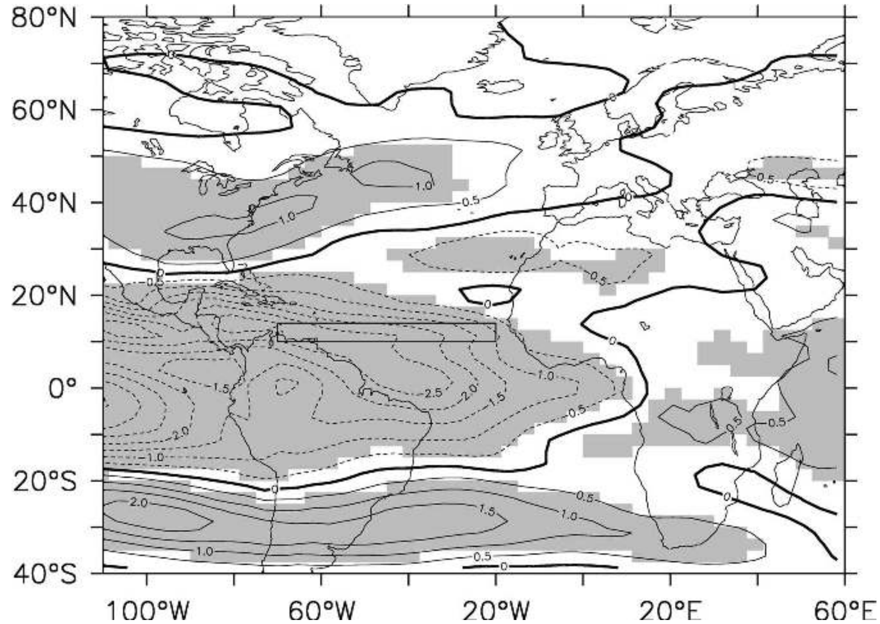

FIG. 9. Response of vertical shear of the zonal wind to warming of the North Atlantic (NA^+–NA^-) in ASO. Shown are anomalies in the vertical shear of the zonal wind between 200 and 850 mb. Units are m s^{-1}. The box shows the MDR for Atlantic hurricanes. ASO is the primary season for hurricanes.

propagate westward more freely than in the other seasons when the blocking effect of the Andes is more important. The excitation and propagation of Rossby waves will also be affected by seasonal evolution of the mean winds.

Once Rossby wave signals reach the eastern Pacific there is potential for interaction with the Pacific ITCZ. In our simulations this interaction generates further, mostly negative, heating anomalies and there is an associated anomalous Walker Circulation between the tropical Atlantic and eastern Pacific (Fig. 8c). Furthermore the Pacific heating anomalies generate further circulation anomalies, evidence of which can be seen in the 1000-hPa winds (Fig. 8a, e.g., the anomalous northeasterly winds around 10°N, 150°–200°E).

In our atmosphere model experiments the Pacific Ocean cannot respond to the anomalies in the surface atmosphere, but in reality there will be a response. There are significant anomalies over the Pacific in wind stress (implied by Fig. 8a), surface heat flux (Fig. 8d), and precipitation (Fig. 3), all of which have the potential to force changes in the Pacific Ocean. Coupled ocean–atmosphere feedbacks (potentially including the Bjerknes positive feedback that plays a central role in ENSO) could be excited, leading to larger changes in the Pacific Ocean and associated aspects of climate (Dong and Sutton 2002). The importance of coupled feedbacks in shaping the response over the Pacific is supported by the recent work of Zhang and Delworth (2005).

Equally interesting are the impacts on the Indian Ocean and its surrounding continents. Figure 3 suggested an impact on the Asian monsoon with negative rainfall anomalies over southern India and the Bay of Bengal. Figure 8a shows that these rainfall anomalies are associated with a weakening of the Somali Jet. The simplest explanation for the weakening of the monsoon is that additional diabatic heating over the tropical Atlantic region produces a widespread warming of tropical mid- and upper-tropospheric temperatures, thereby tending to inhibit convection in regions outside the tropical Atlantic (this widespread warming is seen in the simulations but is not shown here.)

Interestingly the association found here between a warm North Atlantic and a weakened Asian monsoon is the opposite to that found by Zhang and Delworth (2005), who found that the monsoon weakened in response to a *cooling* of the North Atlantic (induced by a reduction in the MOC). A possible explanation is that coupled feedbacks in the Indian Ocean (or in the Pacific but with knock-on consequences for the Indian Basin) modify the direct response to such an extent as to reverse the sign of the monsoon anomalies. This hypothesis is supported by recent experiments we have carried out with a coupled model (Dong et al. 2006; Dong and Sutton 2007).

6. Conclusions

In this paper we have used an atmospheric general circulation model to investigate the response of seasonal mean climate to a seasonally invariant basinwide warming or cooling of the North Atlantic Ocean. Temperature variations with this spatial structure, and multidecadal time scales, are present in the instrumental record (approximately the last 150 yr), and may be caused by variations in the Atlantic MOC, changes in external forcing, or—most likely—a combination of these factors. Larger-amplitude variations in North Atlantic SST are believed to have occurred in the more distant past.

Our major findings are as follows:

- In all seasons the climate response is strongest, in the sense of highest signal-to-noise ratio, in the Tropics (subject to the caveat that we included no SST forcing north of 70°N). The largest anomalies are not necessarily coincident with the regions of highest signal-to-noise ratio. The largest precipitation anomalies are found in the Tropics in all seasons, but in some seasons the largest temperature anomalies are found at higher latitudes.
- There is a large seasonal cycle in the climate impacts. Some features, such as enhanced precipitation in the tropical North Atlantic region, are consistent across seasons but many features are not. For example, there is a large impact on precipitation and temperature over the United States and Mexico in summer but little impact on the same region in winter. By contrast, impacts on Eurasian temperatures are largest in winter. In winter and spring temperature anomalies in the extratropics are largely governed by dynamical changes in circulation rather than simply reflecting a thermodynamic response to the warming or cooling of the ocean. For example, warming of the North Atlantic Ocean induces *negative* temperature anomalies over North America in spring.
- The tropical climate response is primarily forced by the tropical SST anomalies. The main features are in line with a Gill-type Rossby wave response to diabatic heating anomalies. The extratropical climate response is influenced both by tropical and higher-latitude SST anomalies. The direct response to extratropical SST anomalies is strongest (highest in signal-to-noise ratio) in JJA but very weak in other seasons. However the impact of the extratropical Atlantic SST anomalies appears to be enhanced in the presence of the tropical Atlantic SST anomalies.
- Consistent with other studies (Vitart and Anderson 2001) North Atlantic SST anomalies modulate the vertical shear in the main development region for Atlantic hurricanes. This impact is closely associated with the above mentioned Gill reponse, and may aid the interpretation of recent observed trends in hurricane activity (Goldenberg et al. 2001; Emanuel 2005; Webster et al. 2005).
- Comparisons of warming versus cooling the North Atlantic showed the tropical response to be rather linear but the extratropical response to be highly nonlinear in SON, DJF, and MAM. We hypothesized that the dominant mechanisms that govern the extratropical response change with the sign of the SST forcing.
- Comparison between the model results and observed changes in sea level pressure between the North Atlantic warm period, 1931–60, and the susbequent cool period, 1961–90, suggested that changes in North Atlantic SST exerted their most important influence in JJA (as discussed by SH05), but that their influence was also significant (primarily in lower latitudes) in SON and DJF. Changes in the North Atlantic Ocean may also have played a role in forcing changes in circulation at higher latitudes, but it is not possible to be conclusive about this suggestion because in our experiments the SST forcing was set to zero north of 70°N.
- Significant climate impacts are not restricted to the Atlantic basin. Impacts on other regions, especially the tropical Pacific, are largest in JJA and SON, probably because the excitation and westward propagation of Rossby waves is most favored in these seasons. Our results suggest that warming or cooling of the North Atlantic Ocean can induce anomalies in wind stress and surface heat and freshwater fluxes, which could influence the Pacific Ocean. Coupled feedbacks are likely to be important in shaping the ultimate climate impacts (Dong and Sutton 2002; Zhang and Delworth 2005). The findings from this study support previous suggestions (Dong and Sutton 2002; Zhang and Delworth 2005; SH05) that the Atlantic Ocean could be an important driver of decadal variability in other regions, possibly including decadal variability of ENSO.

If our model results provide a reliable indication of the behavior of the real world, then they have implications for the interpretation of climate records. In particular it is important that the large seasonal variation of the climate response, both in spatial pattern and signal-to-noise ratio, be taken into account in attempts to infer the role of the Atlantic Ocean in modulating proxy climate records. Our findings suggest that attempts at such attribution are much more likely to be successful in the Tropics where the signal-to-noise is higher and the responses more linear.

The evidence from our results of the key role for SST anomalies in the tropical Atlantic Ocean also points to the importance of further work to identify and better understand the causes of these SST anomalies. Specifically, there is a need to distinguish more clearly the roles of 1) variations in the Atlantic MOC and 2) changing external forcings in modulating SST in this critical region. Conclusions regarding this issue will have implications for projections of climate change, in which competition between greenhouse gas–induced warming and a greenhouse gas–induced weakening of the MOC is an important uncertainty affecting projections for the Atlantic region especially.

It is important to acknowledge the limitations of our study. In particular, we have carried out experiments with a single atmospheric model, and other models may yield climate responses that differ in some respects from those we have found. Qualitative intermodel comparisons do suggest consistency with respect to important features such as the climate impacts over North America in boreal summer (M. Hoerling 2005, personal communication), but there is a need for more detailed, controlled, and quantitative comparisons. [A set of such comparisons are being undertaken as part of a European Union (EU) Framework 6 project, understanding the dynamics of the coupled climate system (DYNAMITE).] Related to this, there is a need to explore the extent to which specific climate impacts may be sensitive to modest changes in the SST forcing pattern or amplitude.

Another potential limitation is the use of a prescribed SST boundary condition. Several studies have pointed out that this choice of boundary condition may sometimes give misleading results (Bretherton and Battisti 2000; Sutton and Mathieu 2002). The notably good agreement between the model results and observed changes in JJA (SH05) suggest that this issue is not always a problem, but there is a need for further work to better understand the circumstances under which a prescribed SST boundary condition is valid, and when it is not.

Our study also highlights several other avenues where further research is needed. There is a need to better understand the nature of multidecadal variability in the high-latitude North Atlantic and its impacts on climate, including in particular the role of sea ice changes. There is also a need to better understand the role of the Atlantic Ocean as a cause of *global* multidecadal variability. In this context clarifying the relative importance of oceanic teleconnections associated with changes in the MOC circulation (e.g., Timmermann et al. 2005), and atmospheric teleconnections, which can excite remote coupled responses (Dong and Sutton 2002; Zhang and Delworth 2005), is a key challenge.

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