Changes of interannual NAO variability in response to greenhouse gases forcing

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Abstract Observations show that there was change in interannual North Atlantic Oscillation (NAO) variability in the mid-1970s. This change was characterized by an eastward shift of the NAO action centres, a poleward shift of zonal wind anomalies and a downstream extension of climate anomalies associated with the NAO. The NAO interannual variability for the period after the mid-1970s has an annular mode structure that penetrates deeply into the stratosphere, indicating a strengthened relationship between the NAO and the Arctic Oscillation (AO) and strengthened stratosphere-troposphere coupling. In this study we have investigated possible causes of these changes in the NAO by carrying out experiments with an atmospheric GCM. The model is forced either by doubling CO$_2$, or increasing sea surface temperatures (SST), or both. In the case of SST forcing the SST anomaly is derived from a coupled model simulation forced by increasing CO$_2$. Results indicate that SST and CO$_2$ change both force a poleward and eastward shift in the pattern of interannual NAO variability and the associated poleward shift of zonal wind anomalies, similar to the observations. The effect of SST change can be understood in terms of mean changes in the troposphere. The direct effect of CO$_2$ change, in contrast, can not be understood in terms of mean changes in the troposphere. However, there is a significant response in the stratosphere, characterized by a strengthened climatological polar vortex with strongly enhanced interannual variability. In this case, the NAO interannual variability has a strong link with the variability over the North Pacific, as in the annular AO pattern, and is also strongly related to the stratospheric vortex, indicating strengthened stratosphere-troposphere coupling. The similarity of changes in many characteristics of NAO interannual variability between the model response to doubling CO$_2$ and those in observations in the mid-1970s implies that the increase of greenhouse gas concentration in the atmosphere, and the resulting changes in the stratosphere, might have played an important role in the multidecadal change of interannual NAO variability and its associated climate anomalies during the late twentieth century. The weak change in mean westerlies in the troposphere in response to CO$_2$ change implies that enhanced and eastward extended mid-latitude westerlies in the troposphere might not be a necessary condition for the poleward and eastward shift of the NAO action centres in the mid-1970s.

Keywords The North Atlantic Oscillation (NAO) · NAO interannual variability · Greenhouse gases forcing · SST forcing

1 Introduction

The North Atlantic Oscillation (NAO) is the dominant mode of climate variability in the North Atlantic region, characterized by fluctuations in the atmospheric mass (or sea level pressure, SLP) between the Iceland and Azores regions (Hurrell 1995). It is a measure of the strength and preferred track of storms and depressions across the North Atlantic and into Europe, and of the strength of the prevailing westerly winds associated with the stormtrack (Osborn 2006; Vallis and Gerber 2008). NAO variability not only influences the mean winter climate over Europe but also influences climate extremes (e.g. Scaife et al. ...
A high NAO index is associated with a strong stormtrack with a northeast orientation, taking depressions into north-west Europe and giving an increase in heavy precipitation events. It is related to relatively warm, wet conditions in northern Europe and relatively cold, dry conditions in southern Europe, with the opposite situation for a low NAO index.

The transient eddies associated with the stormtrack play a key role in driving the midlatitude westerly jets. Variability of the resulting eddy-driven jet stream is dominated by shifting and pulsing of the jet, and variations in the NAO represent a combination of both these effects (Vallis and Gerber 2008; Woollings et al. 2010b). Observations and modelling studies suggest that a substantial amount of NAO variability is sustained by the eddy vorticity/momentum flux in the middle-high latitudes associated with the North Atlantic stormtrack (Lau and Nath 1991; Ting and Lau 1993; Hurrell 1995; Thompson et al. 2003). A positive feedback mechanism for the NAO between the eddy momentum flux and the mean wind structure was further demonstrated by Limpasuvan and Hartmann (2000) and Hartmann (2000). When the NAO is in its positive phase, the zonal flow exhibits considerably stronger westerlies and westerly vertical shear between 60°N–80°N than the negative phase. The stronger westerly zonal winds in the upper troposphere cause planetary waves to be less strongly refracted toward the pole, resulting in anticyclonic synoptic scale wave breaking, anomalous equatorward wave propagation and poleward momentum flux anomalies (Benedict et al. 2004; Franzke et al. 2004; Riviere and Orlanski 2007) that in turn support the high latitude westerlies and strengthen the polar vortex. In the sense of this maintaining mechanism, the centres of action of the NAO are substantially regulated by the intensification and displacement of the stormtrack in the North Atlantic basin.

In the late twentieth century there was a positive trend in the NAO that led to warmer and wetter winters in northern Europe, and cooler and drier winters in southern Europe (Hurrell 1995). It has been shown that this trend may be partly induced by anthropogenic increases in atmospheric greenhouse gases, although this trend is underestimated by the state-of-the-art coupled atmosphere ocean general circulation models (CGCMs) with large uncertainty (e.g., Gillett et al. 2005; Osborn 2004; Kuzmina et al. 2005; Stephenson et al. 2006). In addition, observations also show that the spatial pattern of interannual NAO variability shifted eastward since the late 1970s, resulting in a link between the NAO and ice export through Fram Strait that did not exist previously (Hilmer and Jung 2000; Jung et al. 2003). This eastward shift of the NAO centres of action appears to be unprecedented during the twentieth century and led to pronounced changes in NAO-related interannual climate variability, such as in near surface air temperatures, and net surface heat fluxes over the North Atlantic ocean (Jung et al. 2003). A natural question arises as to whether the NAO pattern shift could have occurred through internal atmospheric variability or in response to some external forcings such as anthropogenic climate change? The role of the mean state change on the eastward shift of NAO action centres has been emphasized in several studies (e.g., Lu and Greatbatch 2002; Peterson et al. 2003; Luo and Gong 2006). They showed that the change in the mean state acts to shift the axis of the North Atlantic stormtrack and the associated eddy and diabatic forcing feeds back to the mean flow in such a way as to drive an eastward-shifted NAO circulation pattern. Alternatively, a regime perspective shows that the anomalous pattern associated with positive phase of the NAO extends further east than the anomalous pattern associated with negative phase of the NAO, suggesting that the eastward shift of NAO action centres in both models and observations may be a consequence of increased occurrences of the positive NAO regimes (Cassou et al. 2004; Johnson et al. 2008; Woollings et al. 2010a).

Ulbrich and Christoph (1999) and Hu and Wu (2004) suggested that a north-eastward shift of the NAO action centres can result from an increase of atmospheric greenhouse gases. However, there are still some questions that remain to be answered. Is the eastward shift of the NAO pattern in the mid-1970s associated with a change in stratosphere-troposphere coupling? In terms of the response to greenhouses gas concentration for the change of NAO pattern, what is the role of the change in sea surface temperature (SST) resulting from CO$_2$ change and what is the direct role of CO$_2$ change? What are the mechanisms responsible for the change in the NAO pattern and associated climate anomalies in response to different forcings? Addressing these questions is the main aim of this paper.

In this paper, we will provide further observational evidence that a change in the structure of extratropical climate variability and its climatic impacts occurred in the mid-1970s. Then, we will investigate the impact of increasing concentrations of atmospheric greenhouse gases on the pattern of interannual NAO variability by using an atmospheric general circulation model (AGCM), forced by either doubling CO$_2$, or increasing SST, or both. In particular, we will assess and elucidate the separate role of direct CO$_2$ change or indirect SST change induced by doubling CO$_2$ on the pattern of interannual NAO variability. The paper is organized as follows: Sect. 2 describes the observational data and the design of the experiments. The change in the pattern of interannual NAO variability and its effect on climatic impacts associated with the NAO based on observations is documented in Sect. 3. The
changes in the pattern of interannual NAO variability in response to different forcings in the model experiments are investigated in Sect. 4 and causes of the changes are elucidated in Sect. 5. Discussions and conclusions are given in Sects. 6 and 7, respectively.

2 Observational data and model experiments

2.1 Observational data

The observed data used in this study are monthly and daily mean NCEP-NCAR reanalysis data from 1949 to 2004 (Kalnay et al. 1996) and the Met Office Hadley Centre mean sea level pressure (MSLP) data set HadSLP2 from 1850 to 2004 (Allan and Ansell 2006). Various global gridded datasets of monthly precipitation are also used in the paper. These include the University of Delaware precipitation data from 1950 to 1999 (at http://www.cdc.noaa.gov), the DEKLIM precipitation data from 1951 to 2000 (Beck et al. 2005), and the PREC/L precipitation data from 1948 to 2004 (Chen et al. 2002).

2.2 Model experiments

The numerical model used is the United Kingdom Met Office Hadley Centre atmospheric general circulation model HadAM3 (Pope et al. 2000) at a horizontal grid-resolution of 2.5° by 3.75° with 19 levels in the vertical. The experiments performed are summarized in Table 1. For the control experiment, we use climatological monthly mean SST and sea-ice concentrations derived from HadISST observations (Rayner et al. 2003) for 1961–1990 and CO2 at the average concentration for 1961–1990. In the perturbed CO2 experiments, the CO2 concentration is doubled. For the perturbed SST experiment, monthly mean SST anomalies are added to the climatological fields used for the control experiment. The SST anomalies were derived from an experiment with the HadCM3 atmosphere-ocean GCM, which comprises HadAM3 coupled to an ocean GCM, in which CO2 increased at 1% per year from the pre-industrial CO2 concentration, reaching 4 times its preindustrial value at year 140 (Dong et al. 2009). To derive anomalies appropriate to a doubling of CO2, the difference between two 30-year means was used: (a) mean for a period (model run years 3–32) with CO2 values near present day and (b) mean for a period (model run years 73–102) with values near 2× present day CO2. The imposed SST anomalies in December, January and February (DJF) mean are illustrated in Fig. 1 which shows the familiar pattern of enhanced warming over the North Pacific and weak warming in the Southern Ocean and North Atlantic. These features are similar to the multimodel mean results based on IPCC AR4 models (Sutton et al. 2007). For the perturbed SST experiment, sea ice concentrations (not anomalies) are taken from the latter period in the coupled model coupled simulation. The perturbed SST and CO2 forcing experiment (2× CO2 climate) is similar to the SST only experiment but CO2 is doubled. For each of control and perturbed experiments, 2 × 25 year integrations with different initial conditions have been performed. The first 5 years of each integration will be ignored and this gives 40 years of data to analyse for each experiment. These experiments are designed to quantify the separate influences of SST change and CO2 change on the pattern change of interannual NAO variability.

3 Eastward shift of the NAO action centres and downstream extension of climate anomalies associated with the NAO in observations

To describe the pattern change of interannual NAO variability that occurred around the mid-1970s (Hilmer and Jung 2000; Jung et al. 2003), sea level pressure (HadSLP2) was regressed onto an NAO index. This was defined as the difference in averaged pressure between two regions (35°N–40°N, 28°W–22°W) and (64°N–66°N, 26°W–21°W). The regression patterns for the periods 1945–1974 and 1975–2004 and their difference are shown in Fig. 2. We chose 1974 to separate the observations into two
periods since much of the increase of the NAO index in the last five decades occurred in the 1970s (Osborn 2006; Scaife et al. 2005). It has been found that the regression patterns for the two periods are not very sensitive to the year that was chosen to separate them so long as it is in the 1970s (not shown; note that Jung et al. (2003) chose 1977 to separate the observations in their study). The means of the NAO index are 18.5 and 22.0 hPa for the two periods and its standard deviations are 6.4 and 5.9 hPa, respectively. The change in the mean NAO index is significant at the 95% confidence level using the $t$ test while the standard deviations are not significantly different from each other using the $F$ test between two periods. Figure 2 shows that the locations of the action centres of interannual NAO variability were shifted eastward during the period 1975–2004 relative to the period 1945–1974. The high pressure maximum shifted eastward by about 10° in longitude. We have performed empirical orthogonal function (EOF) analyses of SLP anomalies over the North Atlantic region for the two periods. Results indicate that the pattern of EOF1 shows a similar eastward shift in the latter period relative to the former period and the principal component (PC) of EOF1 is highly correlated with the NAO index for both periods.

Fig. 1 The sea surface temperature anomalies in DJF (°C) from the coupled model (HadCM3) simulation

Fig. 2 Regression of SLP anomalies (hPa) that are associated with the NAOI in DJF based on HadSLP2 data (Allan and Ansell 2006). a For 1945–1974, b for 1975–2004, and c the difference between the patterns in b and a. d The difference of SLP interannual standard deviation between the two periods, and e the cumulative distribution function (CDF) of the spatial correlations between the regression patterns for two randomly selected independent 30-year periods from 1945–2004. The vertical line in e is the spatial correlation between the regression patterns in a and b. Shading in a and b indicates that regressions are statistically significant at the 95% confidence level using a $t$ test and shading in d indicates that differences are statistically significant at the 95% confidence level using an $F$ test.
(not shown), indicating that the eastward shift of the action centres of interannual NAO variability in the mid-1970s is a robust feature. In addition to the eastward shift, the global SLP regression pattern indicates that there are significant pressure anomalies over the Pacific sector associated with interannual NAO variability during the later period, in contrast to weak pressure anomalies during the earlier period (not shown). The relatively strong link between the North Atlantic and North Pacific in the later period is consistent with the findings of Pinto et al. (2010). This indicates that the interannual variability of SLP over the northern hemisphere in the later period is more zonally symmetric and AO-like.

The difference in the standard deviation of SLP interannual variability between the two periods is illustrated in Fig. 2d. It shows that the main differences are the enhanced variability over western Europe and the Barents Sea, where the difference in SLP regression patterns between the two periods is also large (Fig. 2c). This indicates that the enhanced interannual variability over these two regions in the later period relative to the early period is likely due to the eastward shift of the centres of interannual NAO variability as shown in Fig. 2c.

To quantify the likelihood that the observed difference in regression patterns between the two periods arises due to sampling error, the following Monte Carlo test is performed. We randomly select two independent 30 year long samples from the period 1945–2004 without any order, perform regression analysis onto the NAO index for each sample, and then calculate the pattern correlation between the two resulting regression fields in the domain shown in Fig. 2a. We repeat this process 1,000 times and calculate the probability density function (PDF) of the pattern correlations. The cumulative distribution function (CDF) against the pattern correlation coefficient is shown in Fig. 2e along with the pattern correlation of 0.89 between the patterns in Fig. 2a, b. As indicated by fig. 2e, the likelihood that the pattern correlation of 0.89 is due to sampling error is about 5%. This result suggests that the change of regression patterns between the two periods, i.e., the eastward shift of the action centres of interannual NAO variability in the mid-1970s is unlikely due to the sampling error, in agreement with Jung et al. (2003).

The associations between the NAO and the Atlantic jet during the two periods are illustrated in Fig. 3. The regression patterns show an equivalent barotropic tripole in the troposphere with a maximum in the upper troposphere. For the period of 1949–1974, a positive NAO is associated with enhanced westerlies around 40°N–60°N and reduced westerlies to the north and south. For the period of 1975–2004, the pattern of zonal wind anomalies associated with the NAO shifts poleward by about 5° of latitude and the associated peak upper tropospheric zonal wind anomalies increase by 0.5–1.0 m s⁻¹, as illustrated by the difference of regression patterns (Fig. 3c). Another distinct feature illustrated in Fig. 3b is the large zonal wind anomalies in the lower stratosphere associated with the NAO, indicating a connection between stratospheric variability and tropospheric variability associated with the NAO in the later period. The geopotential height patterns (Fig. 3d, e) show an equivalent barotropic dipole for both periods with anomalous increased height in the midlatitudes and anomalous low height in the polar region. The zonal flows associated with the NAO shown in Fig. 3a, b are close to being in geostrophic balance with the height anomalies shown in Fig. 3d, e (e.g., Woollings 2008). Figure 3d shows significant height anomalies in the troposphere with maximum centres in the upper troposphere while Fig. 3e shows significant anomalous low height in the polar region in the lower stratosphere, implying a different connection with the stratosphere. The connection of the NAO with stratospheric variability in the later period is much strong as indicated in Fig. 3e, f and associated with this is anomalous cooling of the polar vortex (not shown).

The stormtracks are diagnosed by using a fourth order binomial filter applied to the NCEP daily mean sea level pressure. This filter has maximum response in the 2–8 day periodicity range, typically associated with the passage of synoptic systems (Rogers 1997). Figure 4 shows the seasonal mean root mean square (rms) of band-pass filtered daily SLPs in DJF for the two periods and the difference. The maximum in both periods occurs over Newfoundland, with the stormtrack axis extending northeastward to Iceland and northwestern Europe, as described by Rogers (1997). Positive differences in most regions over the North Atlantic in Fig. 4c imply that storm activity was enhanced and positive anomalies over northwestern Europe indicate downstream extension of the stormtrack in the later period relative to the early period.

The interannual variability of the Atlantic stormtrack associated with the NAO in the two periods is illustrated in Fig. 5. Some striking differences occur between the two periods. In the early period, a positive NAO is associated with more storm activity at high latitudes in the North Atlantic and less storm activity at midlatitudes of the North Atlantic (Fig. 5a). However, the increased storm activity occurs more northward and downstream of the North Atlantic into Scandinavia with a positive NAO for the later period (Fig. 5b, c). This change in stormtrack activity associated with interannual NAO variability between the two periods is consistent with the northward shift and enhancement of the Atlantic jet activity and interannual variability associated with the NAO shown in Fig. 3. In addition, the regression pattern of Fig. 5b is very similar to the dominant mode of Atlantic stormtrack variability in the later period (not shown). The NAO index and the principal component of the first EOF mode of North Atlantic stormtrack variability are highly correlated for the later period with a correlation.
A coefficient of 0.72, in contrast to 0.32 for the early period. This result indicates that the relationship between the NAO and interannual stormtrack variability in the North Atlantic during the past five decades is not stable. There is decadal-multidecadal modulation of this relationship, with a strong relationship only after the mid-1970s. These results are very similar to Jung et al. (2003), who showed that the eastward shift of the action centres of interannual NAO variability was also associated with an eastward shift of the stormtrack and changes in the number of deep extratropical cyclones over the North Atlantic. The non-stationarity of the NAO-stormtrack relationship may explain the contrasting findings of Rogers (1997) and Wettstein and Wallace (2010) on the strength of this relationship.

The surface air temperature anomalies over Eurasia associated with the NAO in the two periods are illustrated in Fig. 6. Positive NAO winters in the early period, for example, are associated with anomalously high surface air temperature over northern Europe and western Russia, and anomalously low surface air temperature over North Africa and Middle East, as discussed in many previous studies (e.g., Hurrell 1995; Stephenson et al. 2006). However, as Fig. 6b indicates, large positive surface air temperature anomalies associated with the positive NAO in the later period extend much further eastward than those in Fig. 6a. This indicates that a downstream extension of climate anomalies accompanied the eastward shift of the NAO action centres in the mid-1970s (Fig. 6c).

The precipitation anomalies also show a similar change in the pattern between the two periods (Fig. 6d–f). During positive NAO winters in the early period, enhanced precipitation occurs mainly over northern Europe, while in the
later period the positive precipitation anomalies also extend into northern Russia. Using other precipitation datasets generate similar patterns as shown in Fig. 6d–f.

4 Eastward shift of the NAO action centres in an atmospheric model forced by SST and CO$_2$ changes

4.1 Eastward shift of the NAO action centres in model experiments

The SLP pattern associated with the NAO is well simulated by the model in the control simulation (Fig. 7a). Both locations of the action centres and the standard deviation (6.4 hPa) of NAOI are close to those for the period 1945–1974 based on observations. The standard deviations of NAO index in the forced simulations are 6.1, 5.8, and 7.8 hPa, respectively and they are not significantly different from the control experiment at the 95% confidence level using the $F$ test. The mean of NAOI in the control experiment is 19.5 hPa while it is 21.4, 21.1, and 18.7 hPa in the forced simulations, respectively, indicating an increase in mean NAOI in response to both SST and CO$_2$ changes or in response to SST change with a magnitude about the half of that between two periods in observations.
In response to joint SST and CO₂ change, as Fig. 7b illustrates, the action centres of the NAO shift eastward by about 10° in longitude. Both mid and high latitude anomalies extend downstream and the NAO pattern becomes more zonally oriented. These results are consistent with previous studies of NAO change in response to increasing atmospheric greenhouse gas concentration using coupled GCMs (Ulbrich and Christoph 1999; Hu and Wu 2004). Interestingly, separate SST (Fig. 7c, g) or CO₂ (Fig. 7d, h) changes also lead to an eastward shift of the NAO action centres. The regressed SLP anomalies in the CO₂ change experiment show a distinct AO-like mode structure with negative pressure anomalies in the Arctic and mid-latitude high pressure anomalies in both the North Atlantic and North Pacific (not shown). This indicates an enhancement of North Atlantic-North Pacific connection of interannual atmospheric variability in the model in response to CO₂ change.

A similar Monte Carlo test as that used for quantification of the observed differences has been performed for the model experiments. In this case, we combine the control simulation with each sensitivity experiment, and then randomly select two 40 year samples from the combined data. The resulting CDF is shown in Fig. 7e, and it indicates that the likelihood that difference of regression patterns between the sensitivity experiments and the control experiment is due to sampling is about 8, 35, and 2.5% for the joint SST and CO₂ forcing, SST forcing, and CO₂ forcing experiments respectively. This indicates that the changes in the pattern of the NAO in response to CO₂ and to joint SST and CO₂ forcing are unlikely to be a result of sampling.
The standard deviation of interannual SLP variability in the control simulation and its change in the perturbation experiments are shown in Fig. 8. The control simulation shows a maximum in interannual variability around Iceland. This feature and the pattern of interannual variability are similar to those for the period of 1945–1974 based on observations (not shown). The changes of interannual variability show clearly that responses to SST and CO₂ change...
do not add linearly. Similarly, Butler et al. (2010) showed that even in a simple atmospheric GCM the jet stream response to heating perturbations does not behave linearly when different heating forcings are combined. Understanding this nonlinearity is an important issue which deserves further investigation. In response to the joint SST and CO₂ change, as Fig. 8b indicates, enhanced interannual variability occurs over western Europe and in the Arctic, where the SLP regression patterns in the two experiments show large differences (Fig. 7f). This indicates that the enhanced interannual variability over these two regions in response to SST and CO₂ change is likely due to the eastward shift of the action centres of interannual NAO variability. In response to the SST change, as Fig. 8c indicates, the increases in interannual SLP variability over western Europe and Barents are relatively small, but consistent with small changes shown in the SLP variability over western Europe and in the Arctic (Fig. 8d), where the difference between the SLP regression patterns in the two experiments is also large (Fig. 7f).

The eastward shift of the NAO action centres in response to either SST change or CO₂ change, or the joint SST and CO₂ changes bears similarity to the observed change which occurred in the mid-1970s. This is especially true for the response to either CO₂ change or joint SST and CO₂ change. The changes of the NAO action centres in response to CO₂ doubling in the coupled model (HadCM3) have also been analyzed (not shown) and the results indicate a similar eastward shift as that in the atmospheric model forced by joint SST and CO₂ change illustrated here.

4.2 Changes in zonal wind anomalies associated with the NAO in model experiments

The zonal wind anomalies over the Atlantic sector associated with the NAO in various simulations are illustrated in Fig. 9 and they show an equivalent barotropic tripole in the troposphere with a maximum in the upper troposphere. In the control simulation (Fig. 9a), the positive phase of the NAO is associated with enhanced westerlies at around 40°N–60°N, with reduced westerlies to the north and south. These features are similar to those shown in Fig. 3a for the period 1949–1974 based on observations. In response to SST and CO₂ changes, the pattern of zonal wind anomalies associated with the NAO shifts poleward by about 5° latitude (Fig. 9b, e). This change is similar to the change that occurred in the mid-1970s in observations (Fig. 3b, c). Another important feature in Fig. 9b is the associated zonal wind anomalies in the lower stratosphere, a feature also observed in Fig. 3b for the observations. The poleward shift of zonal wind anomalies associated with the NAO is also seen in the response to SST change (Fig. 9c, f) and to CO₂ change (Fig. 9d, g). Interestingly, the zonal wind anomalies associated with the NAO in response to CO₂ change features a clear signal in the lower stratosphere, indicating a strong stratosphere-troposphere connection in response to the direct CO₂ change.

Fig. 8 Standard deviation of interannual SLP variability (hPa) in DJF and changes in perturbation experiments relative to the control. a For the control experiment. b The change for the SST and CO₂ experiment, c the change for the SST experiment, and d the change for the CO₂ experiment. Shading indicates that changes are statistically significant at the 95% confidence level using an F test.
4.3 Changes in the NAO-related stormtrack variability in model experiments

The association between the interannual NAO variability and the North Atlantic stormtrack variability in the control simulation is illustrated in Fig. 10a. As in observations shown in Fig. 5a for the period 1949–1974, positive NAO winters in the control simulation are associated with more storm activity in high latitudes and less storm activity in midlatitudes of the North Atlantic. In response to the joint

Fig. 9 Regression of zonal wind (m s\(^{-1}\)) over the Atlantic sector (60°W–0°E) onto the NAOI in DJF for model simulations. a For the control experiment, b for the SST and CO\(_2\) experiment, c for the SST experiment, and d for the CO\(_2\) experiment. e, f, and g are the anomalies relative to the control simulation. Shading indicates that regressions are statistically significant at the 95% confidence level using a \(t\) test.
SST and CO$_2$ change, the NAO-related stormtrack variability shows a downstream extension (Fig. 10b). The difference relative to the control simulation shows positive anomalies in the north and negative anomalies in the south and an eastward extension of these anomalies (Fig. 10e), implying a strengthening, and an extension of NAO-related stormtrack variability. In comparison with the control simulation, SST forcing alone leads to weak changes of

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**Fig. 10** Regression of the seasonally mean rms of band-pass filtered daily sea level pressures (hPa) onto the NAOI in DJF for model simulations. a For the control experiment, b for the SST and CO$_2$ experiment, c for the SST experiment, and d for the CO$_2$ experiment.  

**e, f, and g** are the anomalies relative to the control simulation. *Shading* indicates that regressions are statistically significant at the 95% confidence level using a $t$ test.
stormtrack activity associated with interannual NAO variability (Fig. 10c, f). However, CO$_2$ forcing leads to large changes in stormtrack variability associated with the NAO, characterized by a northward shift, amplification in strength and a downstream extension (Fig. 10d, g), similar to the changes that occurred in the mid-1970s in observations (Fig. 5b, c).

4.4 Changes in NOA-related surface air temperature variability in model experiments

The surface air temperature anomalies over Eurasia associated with interannual NAO variability in various experiments are illustrated in Fig. 11. For the control experiment (Fig. 11a), positive NAO winters, for example, are
associated with anomalously high surface air temperature over northern and western Europe and anomalously low surface air temperature over North Africa and the Middle East. However, the high latitude positive surface air temperature anomalies associated with the positive NAO do not extend eastward enough in comparison with those based on observations (Fig. 6a), indicating that the model has difficulties in reproducing the NAO-related climate anomalies. Despite this deficiency, as Fig. 11 indicates, large NAO-related surface air temperature anomalies extend further eastward in the forced experiments than those in the control simulation. This is especially true for the doubled CO\textsubscript{2} experiment (Fig. 11d, g), which shows strong surface temperature anomalies associated with the NAO extending far east, similar to the period 1975–2004 in observations (Fig. 6b, c).

5 Understanding the eastward shift of the NAO action centres in the model experiments

What processes are responsible for the eastward shift of the pattern of NAO action centres in response to SST change, CO\textsubscript{2} change, or joint SST and CO\textsubscript{2} change?

Figure 12 shows the zonal mean temperature and zonal wind changes over the Atlantic sector in response to various forcings. In response to joint SST and CO\textsubscript{2} change, the relatively strong polar warming in the lower troposphere is clear in the winter hemisphere, with a maximum warming of 5.0°C at the surface near the North Pole (Fig. 12a). The zero contour follows the tropopause, with cooling in the stratosphere and a maximum tropical warming of 3.5°C at about 300 hPa on the equator. The cooling in the stratosphere is due to the well-known stratospheric adjustment via increased infrared emission (e.g., Shine et al. 2003). The strong upper tropospheric warming in the tropics enhances the poleward temperature gradient in the upper troposphere. As Fig. 12b, c show, the sector mean temperature anomalies in the troposphere are predominantly the result of atmospheric response to SST change. As expected, the cooling in the stratosphere seen in Fig. 12a is mainly the result of the direct response to CO\textsubscript{2} change (e.g., Shine et al. 2003).

Associated with the changes in air temperature are changes in zonal wind (Fig. 12d–f). In response to the joint SST and CO\textsubscript{2} change, the Atlantic jet is strengthened and extends poleward, as illustrated by enhanced westerlies poleward of 45\degree in both hemispheres, consistent with the increased poleward temperature gradient in midlatitudes in the upper troposphere (Fig. 12a). This is consistent with Yin (2005). The tropospheric jet responses, seen in Fig. 12d, predominantly result from the response to SST change (Fig. 12e). The response is also associated with a significant strengthening of the zonal-mean westerlies at the surface at about 45\degree N and 50\degree S. This poleward shift of westerlies in the Atlantic sector in both hemispheres also bears a similarity to the changes in westerlies between the two periods in the observations (not shown).

Changes in zonal wind at 500 hPa in response to various forcings are illustrated in Fig. 13a–c. They clearly show that the Atlantic jet is strengthened in response to SST and CO\textsubscript{2} change and the jet also extends downstream. Once again, the changes in tropospheric circulation seen in Fig. 13a are predominantly the result of SST change (Fig. 13b) since the CO\textsubscript{2} change induced zonal wind anomalies are very weak and insignificant (Fig. 13c).

A useful local measure of the susceptibility of the basic state to baroclinic instability is the growth rate of the fastest growing Eady wave (e.g., Hoskins and Valdes 1990). This is defined as 0.3f(N/f)(dv/dz) where f is the Coriolis parameter, N is the Brunt-Vaisalla frequency, z is the upward vertical coordinate and v is the horizontal wind vector. The changes of this quantity at 500 hPa in response to different forcings are illustrated in Fig. 13d–f. The characteristic stormtrack regions in the control simulation are well delineated with high Eady growth rate (not shown), supporting the hypothesis that this diagnostic is of some use in identifying the linkage between the time mean flow and transient eddy activity, despite being unable to identify the geographical extent of regions where storms are decaying. The enhanced growth rate in response to SST and CO\textsubscript{2} changes is found over the North Atlantic around 45\degree N and over western Europe. The changes of rms band-pass filtered daily SLP show similar changes (not shown), indicating enhancement and downstream extension of high frequency transient eddy activity in response to SST and CO\textsubscript{2} change. These changes predominantly reflect the response to SST change as indicated by Fig. 13e. This suggests that the changes in the mean flow in the Atlantic sector, and resulting changes in the transient eddy activity in response to SST change are responsible for the change in interannual NAO variability (e.g., Hartmann 2000; Limpasuvan and Hartmann 2000; Lu and Greathatch 2002; Peterson et al. 2003; Thompson et al. 2003; Luo and Gong 2006).

Figures 12, 13 show that the change in the westerly wind and Eady growth rate in the troposphere in response to direct CO\textsubscript{2} change is very small and insignificant. However, as indicated in the previous section, the pattern of interannual NAO variability in this simulation also shows an eastward shift relative to the control simulation. The mean response in the stratosphere is characterized by a strengthened polar vortex since there is significant enhancement of zonal wind over the Pacific sector (not shown) in response to CO\textsubscript{2} change although the change in the Atlantic sector is weak (Fig. 12f). Similar to
Castanheira and Graf (2003), we define an index of stratospheric polar vortex strength as the zonal mean zonal wind at 60\(^\circ\)N and 50 hPa. In response to joint SST and CO\(_2\) change or to SST change, the mean polar vortex strength increases by 16.7 and 7.1% relative to the control experiment while the interannual variability increases by 20.1 and 10.3% respectively. In response to CO\(_2\) change, the mean polar vortex strength increases by 7.0%. However, the interannual variability of the polar vortex index increases by 44.7%. The enhancement of stratospheric interannual variability is further illustrated in Fig. 14 that shows that the variability in the stratosphere increases.

![Fig. 12](image-url)
about 10–20% for the joint SST and CO₂ change and SST change. However, the interannual variability in the CO₂ only experiment increases by about 40–50% in the polar region and 60–70% in the North Atlantic and western Europe. These results indicate that the changes of interannual NAO variability in response to CO₂ change in the model experiment are associated with a significantly enhanced interannual variability in the lower stratosphere, highlighting the role of the stratosphere-troposphere coupling. The increased westerlies at about 45°N in the lower stratosphere over the Pacific sector in response to the CO₂ change imply that the critical line is further equatorward, so wave activity can extend toward the equator and is less strongly guided into the polar stratosphere, resulting in a less disturbed polar vortex (Holton and Tan 1980). This might be a factor in the enhanced troposphere-stratosphere coupling.

6 Discussion

During the second half of the twentieth century, the NAO underwent two kinds of interdecadal change. Firstly, the NAO showed the well known pronounced positive trend from low values during the 1960s to high values in the 1990s (Hurrell 1995; Gillett et al. 2005; Kuzmina et al. 2005; Scaife et al. 2005). Secondly, the NAO action centres of interannual variability shifted eastward in the late 1970s (Hilmer and Jung 2000; Jung et al. 2003). In this study, we have focused on understanding the eastward shift of the interannual NAO variability and the associated climate anomalies. Based on observations the eastward shift of the NAO action centres was associated with a poleward shift of anomalous westerlies and a poleward and downstream extension of the stormtrack after the mid-1970s. In addition, analysis has demonstrated that the interannual NAO
variability for the period after the mid-1970s has an annular mode structure with significant anomalies over the North Pacific and an NAO-related signal penetrating deep into the stratosphere, implying a strengthened relationship between the NAO and the AO and an enhanced role of the stratosphere in interannual NAO variability.

Castanheira and Graf (2003) showed that stratospheric circulation plays a role in controlling the connection between North Atlantic and North Pacific pressure patterns. A teleconnection between SLP over the North Pacific is found during the strong polar vortex regime but not when the polar vortex is weak. The increased NAO connection to the North Pacific in the later period in observations found in this study is therefore consistent with the observed polar cooling trends in the upper troposphere and lower stratosphere in recent decades in both hemispheres (Folland et al. 1998) which are associated with increased zonal wind in mid-high latitudes in the stratosphere (Scaife et al. 2005). This is also consistent with Christiansen (2003) who showed an abrupt stratospheric regime shift in the late 1970s with more frequent strong polar vortex regimes in the period after the late 1970s, and Honda et al. (2007) and Pinto et al. (2010) who showed that the inter-basin dynamical link between the North Pacific and the North Atlantic was strengthened in recent decades.

The AGCM modelling results forced by both SST and CO₂ changes in this study are in agreement with Ulbrich and Christoph (1999) and Hu and Wu (2004) in suggesting an eastward shift of the NAO. However, additional sensitivity experiments indicate that both SST and CO₂ change independently force an eastward shift of the pattern, similar to that seen in observations. The effect of SST change can be understood in terms of mean changes in the troposphere, especially in the Eady growth rate. The large warming in the upper troposphere in the tropics in response to SST change increases the poleward temperature gradient in midlatitudes in the upper troposphere. Corresponding to this is an increase and poleward shift in Eady growth rate which, in turn, leads to a poleward shift and a downstream extension of the Atlantic stormtrack and eddy-driven jet stream, and the associated climate anomalies. This result is consistent with Peterson et al. (2003) who suggested that the eastward shift of the NAO action centres is a consequence of the upward NAO trend in recent decades. Using an AGCM driven by diabatic forcing, Peterson et al. (2003) implicated nonlinear dynamics, such as eddy fluxes either by transient eddies or the mean flow, in the circulation and NAO changes.

The changes in interannual NAO variability in response to CO₂ change are associated with a strengthened mean polar vortex and strongly enhanced interannual variability, highlighting the dynamical role of the stratosphere. The changes in variability in this simulation can not be understood in terms of the mean changes in the troposphere, which are very weak (Fig. 13c, f). The influence of the wintertime stratospheric polar vortex and heating

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**Fig. 14** Standard deviation of interannual variability of 50 hPa height (m) in DJF and changes in perturbed experiments relative to the control. **a** For the control experiment, **b** changes in response to SST and CO₂ change, **c** changes in response to SST change, and **d** changes in response to CO₂ change. *Shading* indicates that changes relative to the control experiment are statistically significant at the 95% confidence level using an *F* test.
perturbations on the tropospheric circulation has been a topic of increasing interest in recent years (e.g., Baldwin et al. 2003; Castanheira and Graf 2003; Norton 2003; Charlton et al. 2004; Haigh et al. 2005; Limpasuvan et al. 2005; Kuroda 2008; Castanheira et al. 2009, Simpson et al. 2009). Different processes have been proposed to explain the dynamical stratosphere-troposphere coupling, such as the mechanisms of downward control by the meridional circulation (Haynes et al. 1991; Thompson et al. 2006), geostrophic/hydrostatic adjustment of the troposphere to perturbations of the stratospheric polar vortex (Ambaum and Hoskins 2002), or planetary wave propagation and breaking (e.g., Limpasuvan and Hartmann 2000; Perlwitz and Harnik 2004; Limpasuvan et al. 2005; Kunz et al. 2009).

The change in NAO-related zonal wind anomalies we find in response to CO₂ change (Fig. 9g) bears a similarity to the zonal wind anomalies of Fig. 4d in Haigh et al. (2005), who performed a sensitivity study using a simple model to an increased temperature gradient in the lower stratosphere. Simpson et al. (2009) further demonstrate that the temperature perturbation in the lower stratosphere leads to a change in refractive index which influences wave propagation, and results in changes in eddy momentum flux and mean meridional circulation in the troposphere that in turn drive zonal wind anomalies on a fast time scale of about a month. The pattern of zonal wind anomalies is also similar to that in Fig. 3 of Scaife et al. (2005) resulting from a perturbation mimicking the observed trend towards stronger stratospheric westerlies. The similarity of the zonal wind changes associated with the NAO in our modeling experiments and those resulting from a perturbation in the stratosphere (e.g., Haigh et al. 2005; Scaife et al. 2005) confirms that the changes in the pattern of the NAO in response to CO₂ forcing might be a consequence of changes in the stratosphere with the change in polar vortex strength and its interannual variability playing an important role.

The possibility of a downward influence from the lower stratosphere by direct modulation of tropospheric baroclinic waves has been suggested by several studies (Charlton et al. 2004; Wittman et al. 2004, 2007; Chen et al. 2007; Castanheira et al. 2009; Kunz et al. 2009), in the sense of altered tropospheric synoptic scale wave breaking characteristics. Using a hierarchy of models, Wittman et al. (2007) found that increasing stratospheric shear increases the phase speed of growing baroclinic waves, increasing the growth rate of modes with low synoptic wave numbers, and decreases the growth rate of modes with higher wave numbers. These changes in baroclinic instability result in changes in mid-latitude stormtracks which in turn lead to a change in the NAO (Vallis and Gerber 2008).

The model responses in this study have been based on experiments with a single model HadAM3 that has relatively poor representation of the stratosphere (Gill et al. 2002; Cordero and de F. Forster 2006). However, it is still not clear whether a well-resolved stratosphere is a necessary condition for the production of reliable tropospheric climate-change predictions (e.g., Shindell et al. 1999; Sigmond et al. 2008). Shindell et al. (1999) have argued that a well-resolved stratosphere is required to reproduce observed tropospheric circulation trends, in particular that part of the trend that projects positively onto the Northern Hemisphere Annual Mode (NAM). Other studies have argued differently (Fyfe et al. 1999; Gill et al. 2002) and shown that such positive NAM responses can arise in global warming simulations in the absence of a well-resolved stratosphere. The effects of increasing stratospheric resolution are not clear. Sigmond et al. (2008) gave an example of one climate model in which the altered response to CO₂ doubling in a stratosphere-resolving version of the model could be attributed to the changes in the gravity wave drag scheme, rather than an active role in changing stratospheric dynamics. Instead, a recent study by Sigmond and Scinocca (2010) suggests that the inter-model response sensitivity might be better understood in terms of the lower-stratospheric biases of their control climates. It will be important in future work to investigate the extent to which the changes identified in our perturbed experiments are sensitive to the model vertical resolution in the stratosphere.

Finally, it is interesting to compare the mean model responses presented here with the response of the coupled model HadCM3 to increased greenhouse gas forcing. Figure 12d shows that, when forced by SST and CO₂ changes, the AGCM exhibits a clear equivalent barotropic mean jet stream response which projects onto the NAO. In contrast, the coupled model HadCM3 predicts a very weak mean response of the NAO to greenhouse gas forcing (Stephenson et al. 2006; Woollings et al. 2010a). This suggests that model biases associated with the coupling to ocean models may result in underestimates of the NAO responses to forcing in the climate model projections.

7 Conclusions

In this paper, the change in the nature of interannual NAO variability and its associated climate anomalies in the mid-1970s in observations have been revisited. The possible causes of these changes in the NAO have been investigated by carrying out experiments with an atmospheric GCM. The model is forced either by doubling CO₂, or increasing sea surface temperatures (SST), or both. In the case of SST forcing the SST anomaly is derived from a coupled model
simulation forced by increasing CO$_2$. The key results are as follows:

- Observations show there was a change in interannual NAO variability in the mid-1970s. This change was characterized by an eastward shift of the NAO action centres, poleward shift of zonal wind anomalies, and downstream extension of climate anomalies associated with the NAO, confirming previous research (e.g., Jung et al. 2003). In addition, it has been demonstrated that the interannual NAO variability for the period after the mid-1970s has an annular mode structure that penetrates deeply into the stratosphere with significant anomalies over the North Pacific. This implies both a strengthened relationship between the NAO and the AO, and an enhanced role of the stratosphere in interannual NAO variability during the later period.

- SST and CO$_2$ change both force an eastward shift in interannual NAO variability and a poleward shift of zonal wind anomalies, similar to those seen in observations.

- The effect of SST change on the interannual NAO variability can be understood in terms of mean changes in the troposphere, especially the change in the baroclinicity as shown by the Eady growth rate. Tropical heating associated with warm SSTs generates strong upper tropospheric warming in the tropics, enhances the poleward temperature gradient in the upper troposphere, and affects the subtropical jet. This leads to the change in baroclinicity and therefore the change in stormtrack which affects the eddy-driven jet and NAO interannual variability.

- The direct effect of CO$_2$ change on the interannual NAO variability cannot be understood in terms of mean changes in the troposphere. Instead, the stratospheric response, characterized by a strengthened mean polar vortex with greatly enhanced interannual variability is important. Stronger interannual variability in the stratosphere, either internally generated or resulting from altered upward planetary wave propagation due to the changed polar vortex, might have played an important role through stratosphere-troposphere coupling (Norton 2003; Haigh et al. 2005) and resulted in changes to the baroclinic waves (Wittman et al. 2007). These results also imply that a strengthened and extended tropospheric jet stream is not a necessary condition for the eastward shift of NAO interannual variability (Lu and Greatbatch 2002; Peterson et al. 2003; Luo and Gong 2006).

- There are many similarities between the changes in several characteristics of NAO interannual variability in the modelling results in response to doubling CO$_2$ and those in observations after the mid-1970s. Although the model forcing is considerably stronger, this implies that the increase of greenhouse gas concentration in the atmosphere, and the resulting change in the stratosphere, might have played an important role in the middecadal change of interannual NAO variability and its associated climate anomalies during the twentieth century.

- The increased regional impacts of interannual NAO variability, both in the recent decades and in the model experiment with doubled CO$_2$, imply that the NAO might be an important predictor for winter climate over Asia. The improvement of short-term NAO forecasts could lead to the improvement of winter climate forecasts for Asia, in both the present day and the future. The predictability of Asian winter climate and its association with NAO variability deserves further investigation.

Both the SST change and CO$_2$ change imposed in our perturbation experiments are larger than the changes in observations between the two periods in observations. It is, therefore, important to perform similar experiments with observed SST and CO$_2$ change to verify the robustness of our findings.

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