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# Factors influencing Northern Hemisphere winter mean atmospheric circulation anomalies during the period 1960/61 to 2001/02

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**Influences from the Tropics, the stratosphere and the specification of observed sea surface temperature and sea-ice (SSTSI) on Northern Hemisphere winter mean circulation anomalies during the period 1960/61 to 2001/02 are studied using a relaxation technique applied to the ECMWF model. On interannual time-scales, the Tropics strongly influence the Pacific sector but also the North Atlantic sector, although weakly. The stratosphere is found to be influential on the North Atlantic Oscillation (NAO) on interannual time-scales but is less important over the Pacific sector. Adding the observed SSTSI to the tropical relaxation runs generally improves the model performance on interannual time-scales but degrades/enhances the model's ability to capture the 42-year trend over the Pacific/Atlantic sector. While relaxing the stratosphere to the reanalysis fails to capture the trend over the whole 42-year period, the stratosphere is shown to be influential on the upward trend of the NAO index from 1965 to 1995, but with reduced amplitude compared to previous studies. Influence from the Tropics is found to be important for the trend over both time periods and over both sectors although, across all experiments, we can account for only 30% of the amplitude of the hemispheric trend.** 

*Key Words:* tropical impact; stratospheric impact; NAO; PNA; ERA-40

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## 1. Introduction

Over the Euro-Atlantic sector, the North Atlantic Oscillation (NAO) is the most important mode of variability in the atmospheric circulation (Greatbatch, 2000; Hurrell *et al.*, 2003), accounting for roughly 40% of the variance in winter mean sea level pressure (MSLP) and exerting a correspondingly strong influence on European winter mean climate. A positive (negative) NAO index implies stronger (weaker) westerly winds than usual leading, in turn, to winters that are generally milder (colder) than usual in Europe. In the North Pacific sector and over North America, it is the Pacific–North American (PNA) pattern that has the dominant influence (Wallace and Gutzler, 1981; Trenberth and Hurrell, 1994). It is well known that the PNA pattern is readily excited by diabatic heating anomalies in the tropical Pacific, for example in association with variability of El Niño–Southern Oscillation (ENSO) (e.g. Trenberth *et al.*, 1998). In the case of the NAO, the role of influences from the Tropics is less clear, although some studies suggest that such an influence could be significant (e.g. Greatbatch *et al.*, 2003; Lin and Derome, 2005; Lin *et al.*, 2009). Fraedrich and Müller (1992) provide evidence for a tendency towards the negative (positive) NAO during El Niño (La Niña) years whereas Greatbatch *et al.* (2004) indicate non-stationary behaviour (also Greatbatch and Jung, 2007) with a very different impact from the Tropics over the Euro-Atlantic sector before and after the 1976/77 Pacific climate shift (Trenberth *et al.*, 2002). Toniazzo and Scaife (2006) argue that the influence of El Niño over Europe depends on the strength of the event in the tropical Pacific and Ineson and Scaife (2009) have presented evidence that the stratosphere can act as a bridge between the tropical Pacific and Europe. The mechanism involves interference between the forced response and the seasonally evolving climatological stationary waves, an issue that has been explored for forcing from different parts of the Tropics by Fletcher and Kushner (2011). Indeed, there is mounting evidence that circulation anomalies in the stratosphere can influence the tropospheric circulation and the NAO in particular, and hence European winter climate (e.g. Baldwin and Dunkerton, 1999, 2001; Scaife *et al.*, 2005; Douville, 2009; Kunz *et al.*, 2009). Some authors (e.g. Cohen *et al.*, 2010) also argue that anomalies in Eurasian autumn snow cover can influence the following winter by means of an upward propagating Rossby wave train leading to anomalies in the stratosphere that, in turn, can influence the extratropical troposphere (Fletcher *et al.*, 2009; Smith *et al.*, 2011).

In a recent series of articles, Jung *et al.* (2010a, 2011) have used a relaxation technique to explore factors influencing the mean extratropical circulation in the troposphere in two different winters, 2005/06 and 2009/10, both of which were unusually cold in northern Europe. In the case of the 2005/06 winter, a significant influence from the tropical Atlantic sector, including South America, was found, whereas in the case of the 2009/10 winter it was difficult to find any determining factors, leading to the conclusion that the unusually negative value of the NAO index that winter may have been the result of internal atmospheric variability within the extratropical troposphere (assuming no model error). The relaxation technique is described in detail in Jung *et al.* (2010b) and is a powerful diagnostic tool that involves strongly constraining the atmospheric state in a model close to observations over part of the globe (e.g. the Tropics) while leaving the rest of the model domain unconstrained. The technique can be viewed as a way of accessing the impact of perfect predictability in the constrained parts of the model domain on the winter mean circulation in the unconstrained parts. A similar technique was used by Douville (2009) who found a strong influence from the stratophere on the winter mean NAO index, an issue we explore further here. The influence of the stratosphere on the upward trend of the NAO index from the mid-1960s to the mid-1990s had been investigated earlier by Scaife *et al.* (2005), who concluded that most of the upward trend can be explained as a response of the troposphere to an upward trend in the strength of the winter polar vortex in the stratosphere. Jung and Barkmeijer (2006) have also demonstrated an influence on the underlying troposphere from perturbing the stratospheric polar vortex using an adjoint technique applied to a version of the European Centre for Medium-Range Weather Forecasts (ECMWF) model. More recently, Simpson *et al.* (2011) have demonstrated the influence of the stratospheric variability for extending the time-scale of events in the troposphere by comparing two model runs, in one of which zonal mean quantities in the stratosphere are relaxed to climatology.

Here we present the results of a series of ensemble experiments in which the relaxation technique has been applied to diagnose the Northern Hemisphere winter mean circulation anomalies for 42 winters during the ECMWF reanalysis dataset (ERA-40) period from 1960/61 to 2001/02. The 42-year period begins at the time when the NAO index was predominently negative and there were a number of severe winters in Europe, notably 1962/63 for which January is the third coldest in the Central England Temperature (CET) record (Manley, 1974; Parker *et al.*, 1992), exceeded only by the Januaries of 1795 and 1740. There then followed the period during which the NAO exhibited a predominently upward trend from the 1970s to the mid-1990s, associated with the generally mild European winters of the 1980s and early 1990s, followed by the reappearance of colder winters starting with 1995/96. In the Pacific sector, the study period includes the 1976/77 climate transition to a warm regime in the tropical Pacific (e.g. Trenberth and Hurrell, 1994), followed by a return to more normal conditions around 1990.

The plan of this article is as follows. In section 2 the model set-up, the different experiments and the analysis techniques are described. Section 3 focuses on the results, first providing an overview using a pattern correlation analysis, then focusing on the interannual variability of the NAO and PNA and then discussing the circulation trend during the analysis period. Finally, section 4 provides a summary.

## 2. Methods

### 2.1. *Experimental set-up*

The numerical model used in this study is based on a version of the ECMWF atmosphere model which was used operationally between September 2009 and November 2010 and is very similar to the model described in Jung *et al.* (2010b,c). The horizontal and vertical resolution is the same as used for the ERA-40 reanalysis (thereby avoiding the need to do interpolation when carrying out the relaxation –see below). In particular, the model has spectral truncation T159 (compared to T1279 when used operationally) and there are 60 levels in the vertical, with about half the levels located above the tropopause (Untch *et al.*, 1998), and extending up to 0.1 hPa. All experiments were carried out using initial conditions and lower boundary conditions (daily sea surface temperature (SST) and sea ice) from the ERA-40 reanalysis. Each winter was integrated separately using 12 ensemble members for each experiment. Initial conditions for the ensemble members were taken from the reanalysis data at 6 h intervals from around 1 November of the respective year. All model runs were started at 1200 UTC on 1 November and the analysis was carried out on the following winter, where winter refers to the months December, January and February (DJF). All model experiments along with their abbreviations are discussed in detail below.

Throughout this article, 'anomalies' for a particular experiment refer to departures of the ensemble mean or individual ensemble members from the mean winter state of the same experiment, the winter mean being the average over all ensemble members and all years comprising the experiment. By defining anomalies in this way, we ensure

 

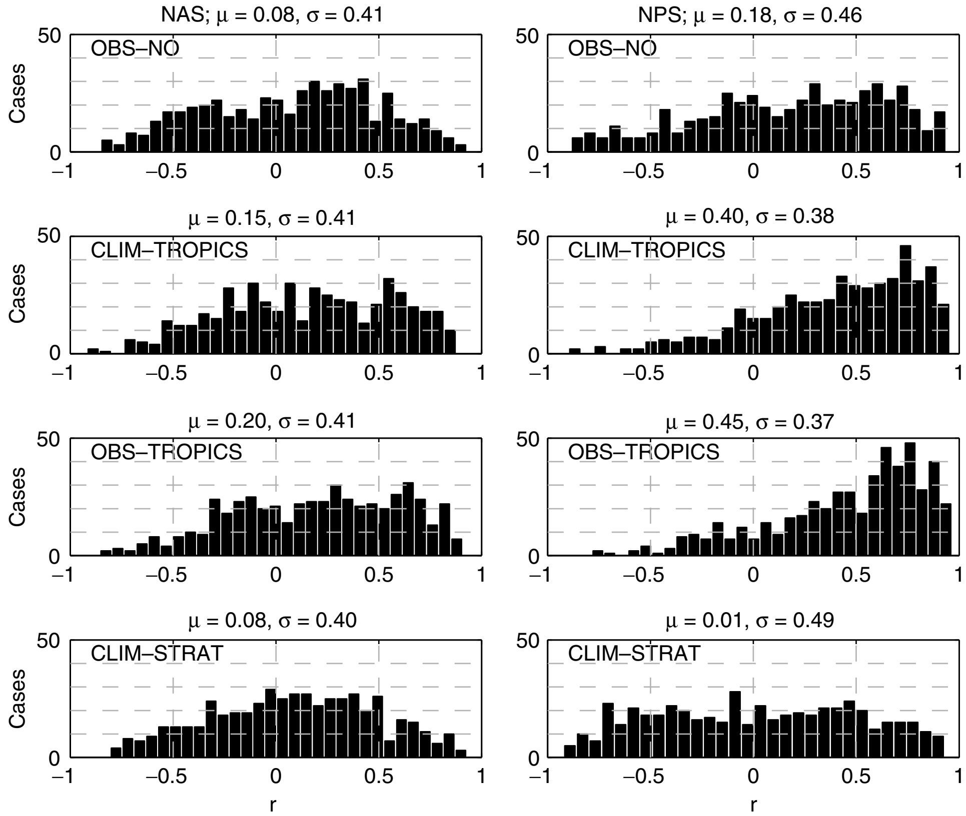


**Figure 1.** Histograms of pattern correlations ($r$) of 500 hPa geopotential height anomalies between the $12 \times 42 = 504$ model realisations in each experiment and the corresponding reanalysis fields. Left column shows the pattern correlation over the North Atlantic sector (20–80°N and 90°W–40°E), right column over the North Pacific sector (30–65°N and 160°E–140°W). Each histogram consists of 30 bins. $\mu$ indicates the mean correlation and $\sigma$ the standard deviation in each case.

that the anomalies reflect the anomalous conditions during the individual winters, with no contribution from the different model climates associated with each experiment.

### 2.2. Relaxation formulation

In the relaxation experiments, the model is drawn toward the ERA-40 reanalysis data in a specific region during the course of the integration; this is achieved by adding an extra term of the following form to the ECMWF model:

$$-\lambda(\mathbf{x} - \mathbf{x}_{\text{ref}}). \tag{1}$$

The model state vector is represented by $\mathbf{x}$ and the reference field toward which the model is drawn by $\mathbf{x}_{\text{ref}}$. The strength of the relaxation is determined by $\lambda = a\lambda_o$, where $a$ defines the geographic region and model levels where the relaxation is applied. Here $\lambda_o = 0.1\,\text{h}^{-1}$ defines the time-scale of the relaxation, here corresponding to a time-scale of 10 h. The parameters that are relaxed are zonal velocity, $u$, meridional velocity, $v$, temperature $T$, and the logarithm of the surface pressure, $\ln ps$ ($\ln ps$ is not relaxed in stratospheric relaxation experiments). $\mathbf{x}_{\text{ref}}$ is taken from the 6-hourly ERA-40 reanalysis data, linearly interpolated in time to each model time step.

To allow for an effective localization, the relaxation is carried out in gridpoint space. When applying masks to localize the relaxation, care was taken to reduce adverse effects close to the relaxation boundaries. Here the transition from relaxed to unrelaxed regions in the tropical relaxation cases is smoothed using a hyperbolic tangent function. The smoothing is such that the relaxation coefficient $\lambda$ goes from $\lambda_o$ to 0 within a 20° belt in latitude. Boundaries of 20°N and 20°S stated in the text below refer to the centre of the respective 20° belt (Figure 1 in Jung *et al.*, 2010b). Changes of $\lambda$ are also smoothed in the vertical in the stratosphere relaxation case. Here, the relaxation coefficient goes from $\lambda_o$ to 0 in a vertical layer encompassing about 13 model levels, as in Jung *et al.* (2010a,b). The values of $\lambda$ at 500, 200, 50, and 20 hPa are given by $1.1 \times 10^{-7}\lambda_o, 2.3 \times 10^{-6}\lambda_o, 0.018\lambda_o$ and $0.5\lambda_o\,\text{h}^{-1}$, respectively (Figure 2 in Jung *et al.*, 2010b). In the case of stratospheric relaxation, the design of the relaxation zone was chosen to test the influence of large-scale stratospheric circulation anomalies which have been observed to first appear in the upper stratosphere and subsequently 'propagate' downward into the lower stratosphere, where they are believed to affect tropospheric weather regimes (Baldwin and Dunkerton, 2001) and, hence, the winter mean circulation in the troposphere. (It has been shown by Jung and Leutbecher, 2007, that the ECMWF model reproduces such behaviour.) The use of a smooth, rather than an abrupt, transition zone is designed to reduce the spurious reflection of upward propagating planetary waves, although such effects are difficult to eliminate completely and should be borne in mind when interpreting the results. For discussion of the relaxation technique as applied to the Tropics, readers are referred to Hoskins *et al.* (2012).

### 2.3. Model experiments

The model experiments are as follows:

1. **CLIM-NO**. In this case, the model sees climatological SST and sea-ice at the lower boundary and no relaxation is

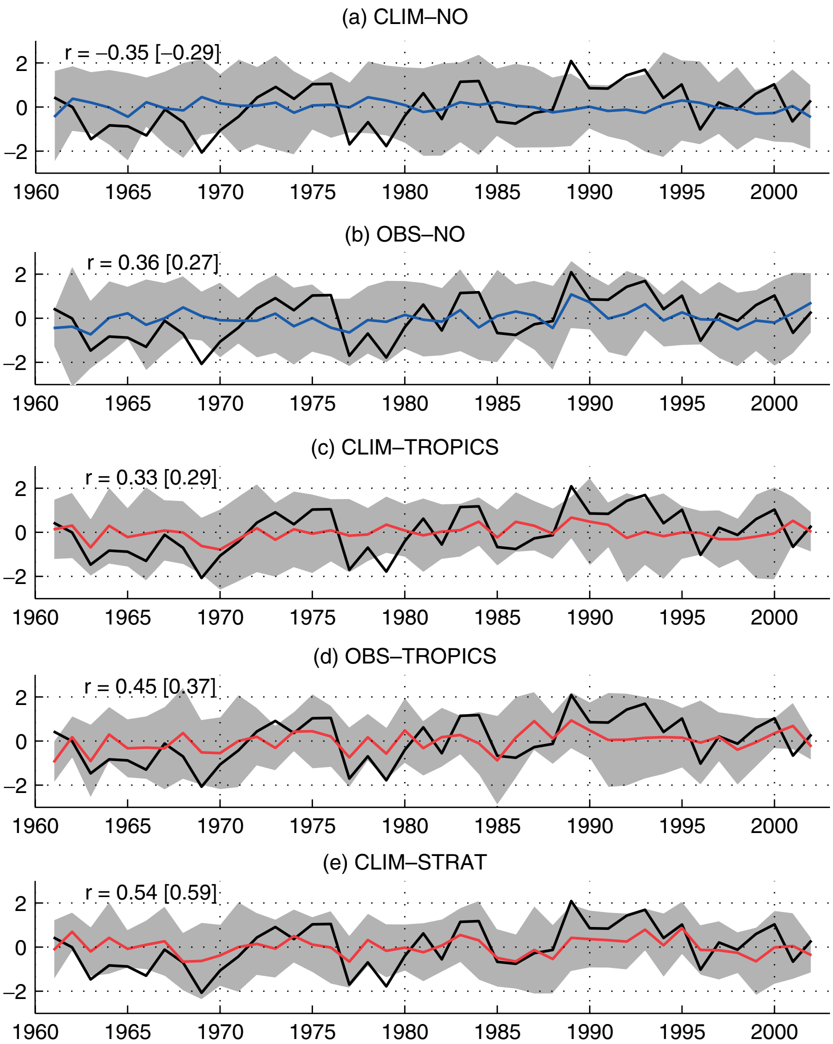


**Figure 2.** NAO indices for 500 hPa height anomalies. The observed NAO index (black) is the first principal component time series of an EOF analysis of 500 hPa geopotential height anomalies applied to the reanalysis data over the North Atlantic sector (30–80°N, 90°W–40°E). All other indices are obtained by projection of model anomalies onto the NAO pattern corresponding to the observed NAO index, normalised by the standard deviation of the observed index. Blue (red) indices are ensemble means without (with) relaxation. The grey shading indicates the range of ensemble members within two standard deviations of the ensemble mean. Correlation (*r*) of the (detrended) ensemble mean indices with the (detrended) observed index is given in the figure (in brackets).

used. The only information relevant to any particular year is contained in the initial conditions.

2. **OBS-NO**. In this case, the model sees observed SST and sea-ice at the lower boundary and no relaxation is used.

3. **CLIM-TROPICS**. In this case, climatological SST and sea-ice are specified at the lower boundary and relaxation is used between 20°N and 20°S throughout the whole depth of the model atmosphere.

4. **OBS-TROPICS**. In this case, observed SST and sea-ice are specified at the lower boundary and relaxation is used between 20°N and 20°S as in CLIM-TROPICS. Since the relaxation completely overwhelms the specification of observed SST in the Tropics, this experiment effectively looks at the additional information gained, compared to CLIM-TROPICS, by specifying the observed SST and sea-ice in the extratropics.

**5. CLIM-STRAT**. In this case, climatological SST and sea-ice are specified at the lower boundary and relaxation is used in the stratosphere (globally, including the Tropics).

A concern regarding CLIM-STRAT is the quality of the ERA-40 reanalysis prior to the satellite era, i.e. before the late 1970s. It is almost certainly the case that the realism of the stratosphere we use for the relaxation ($\mathbf{x}_{ref}$ in Eq. (1)) is not as good before the late 1970s as it is after this time. However, this is not to say that the reanalysis prior to the introduction of satellite data is completely unconstrained in the stratosphere since there is certainly an influence from the underlying troposphere, as can be shown using model experiments that employ relaxation only in the troposphere (not shown here). Nevertheless, the stratosphere as represented in the reanalysis is still a dynamically consistent realisation of the stratospheric state within the model used for the reanalysis. We believe, therefore, that the results we report in section 3 at least indicate the influence of the stratosphere on the underlying troposphere within the context of the model used for the reanalysis. Furthermore, since the model we use is a version of the ECMWF model with the same resolution as used for the reanalysis, the dynamics of the model we use is very similar to that used for the reanalysis. Similar considerations apply to other changes in the data stream used for the reanalysis (e.g. the availability of observations from the Tropics). This point is particularly important to remember when we consider the long-time-scale trends in section 3.4.

### 2.4. *Definitions used for the NAO and PNA*

We define the NAO as the first Empirical Orthogonal Function (EOF) of the winter mean 500 hPa height ($Z500$) anomalies taken from the ERA-40 data applied to the North Atlantic sector (the region 30–80°N and 90°W–40°E). This is similar to the definition given in Hurrell *et al.* (2003) but using only the region north of 30°N (instead of 20°N) to avoid intersecting the relaxation zone used in the tropical relaxation experiments. (Note that applying the EOF analysis to winter MSLP leads to an almost identical index.) The NAO index is then the principal component time series of this EOF and one unit of the index refers to one standard deviation departure from 0 (positive index corresponding to enhanced westerly winds over the northern North Atlantic). To obtain NAO indices from the model results, winter mean anomalies from the model are projected onto the spatial pattern associated with the NAO in the ERA-40 data and normalised by the standard deviation of the observed index (to enable comparison with the reanalysis data) where here, and in what follows, the 'observed' index refers to the index derived from the ERA-40 data (same for the PNA below). It should be noted that applying an EOF analysis to the winter mean model anomalies from CLIM-NO (the case with no added forcing) gives almost the same spatial pattern as obtained from the ERA-40 data. To define the PNA index, we use minus the area-weighted average of winter mean $Z500$ anomalies in the North Pacific sector (the region 30–65°N and 160°E–140°W, following Trenberth and Hurrell, 1994) normalised by the standard deviation of the time series (positive index corresponding to a deepened Aleutian low). The PNA index for the model experiments is defined in the same way, except that the time series is normalised by the standard deviation of the observed index.

### 2.5. *Monte Carlo methods*

To answer questions regarding the statistical significance of our results, we use Monte Carlo methods. In the following, we use the time series of the NAO index as an example. The same procedure is used in the case of the PNA index. We illustrate the method by describing how it is used to answer the following questions:

*1. Is the correlation between the ensemble mean NAO index and the observed NAO index for a particular model experiment significantly different from zero?*

We begin by choosing 12 time series for the NAO index by randomly selecting (without replacement) 42 values (one representing each year) 12 times from the $12 \times 42 = 504$ values available from all model runs comprising the model experiment. The ensemble mean NAO index is then computed from the 12 selected values for each year and the correlation between the time series of this ensemble mean index and the observed NAO is computed. The process is then repeated a large number (typically 10 000) of times and a histogram of the ensemble mean values is computed to produce the probability density function (PDF) of the correlation values. The resulting PDF is centred around zero correlation and significance levels can be derived by calculating the correlation ranges corresponding to percentiles. For example, the 95% range is defined so that 95% of all values sit within this range. The significance of the correlation between the actual ensemble mean NAO index from the model experiment and the observed index can then be assessed by noting into which percentile it falls. In the following, when we say that a correlation is significantly different from zero at the 95% level we mean that the correlation is found outside the 95% range.

Table 1. Interannual correlation coefficients between the observed NAO and PNA index for Z500 and the corresponding ensemble mean index, with correlations for detrended indices in brackets.

| Experiment | NAO | PNA |
|---|---|---|
| CLIM-NO | −0.35* [−0.29 ] | 0.07 [0.08 ] |
| OBS-NO | 0.36* [ 0.27 ] | 0.53**[0.58**] |
| CLIM-TROPICS | 0.33* [ 0.29 ] | 0.79**[0.79**] |
| OBS-TROPICS | 0.45**[ 0.37* ] | 0.77**[0.80**] |
| CLIM-STRAT | 0.54**[ 0.59**] | 0.04 [0.07 ] |

One (two) asterisks denote correlations exceeding 0.31 (0.40) and are different from zero at the 95% (99%) level (section 2.5, Question 1).

For all model experiments, the method yields correlation thresholds of 0.31 and 0.4 for the 95% and 99% significance levels, respectively (the same as for a Student's $t$-test with 42 degrees of freedom). Serial correlation associated with memory from one year to the next, although not of serious concern for the NAO or PNA, will raise these thresholds slightly and should be born in mind when interpreting the correlations listed in Table 1.

*2. What is the strength of the added forcing in each model experiment?*

A large number (e.g. 10 000) of possible realisations are produced by picking one value for the NAO index from the 12 model runs for each year, keeping the order of the years as in the original experiment. Noting that the real world corresponds to a single realisation, it is clear that any of the selected realisations could correspond to the observed NAO time series within the context of the particular model experiment. The correlation is then computed between the time series of each realisation and the observed NAO index and a PDF of these correlation values is produced. The shift of the resulting PDF away from symmetry about zero can then be used to assess the strength of the forcing that has been added in each experiment (see discussion of the individual cases). This approach is much more powerful than looking at the ensemble mean alone. For example, even if the forcing is weak, then it is quite possible to have a highly significant correlation between the ensemble mean NAO index and the observed NAO index, as pointed out by Bretherton and Battisti (2000). However, in such a case, the PDF of correlations of single realisations and the observed NAO index will be centred close to zero, indicating the unimportance of the forcing in this case. (Bretherton and Battisti, 2000, give an informative discussion of this issue, particularly the discussion concerning their Figure 1.)

*3. How unusual is the observed NAO trend in the context of a particular model experiment?*

A large number (e.g. 10 000) of possible realisations are produced as in *2* above. The trend is then computed for each realization and a PDF produced of all possible trends. One can then see where the observed trend sits on the PDF and assess its likelihood of occurrence for a given model forcing. The PDF in this case will be centred around the trend of the ensemble mean NAO index from the experiment.

 

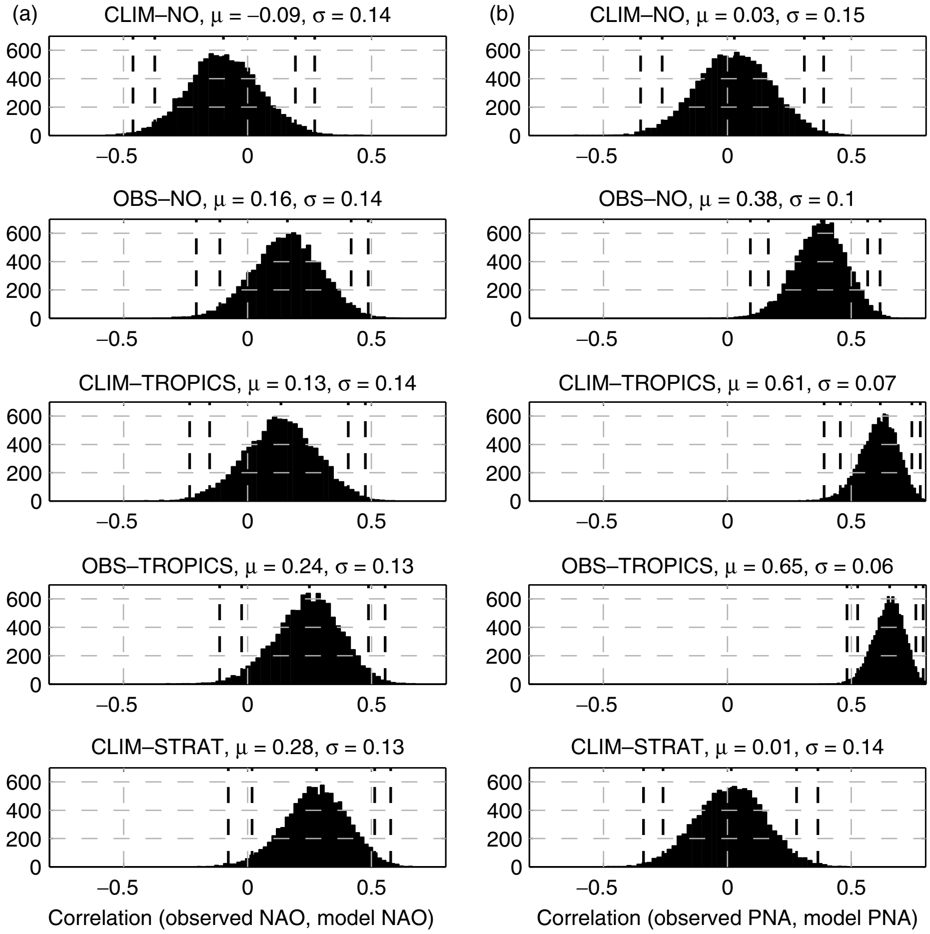


**Figure 3.** Histograms, using 50 bins, of correlations between the observed index and the NAO (left column) and PNA (right column) indices of 10 000 different model realisations for each experiment. The model realisations are created as described in section 2.5. Black dashed lines indicate the median as well as the 95% and the 99% ranges of the distribution. For the histograms shown in the figure, detrending of the time series was not carried out. However, detrending has little effect. $\mu$ indicates the mean correlation and $\sigma$ the standard deviation in each case.

## 3. Results

### 3.1. Pattern correlation

To give an overview of the results, we start with the pattern correlation between the $Z500$ anomalies from the individual ensemble members and the $Z500$ anomalies from the ERA-40 data. It should be noted that, to obtain meaningful pattern correlations, the area average is first removed from each field. Figure 1 shows histograms of the pattern correlations over the North Atlantic sector (the region 30–80°N and 90°W–40°E) and the North Pacific sector (30–65°N and 160°E–140°W) using 30 bins for each case. The impact of the forcing can be seen when there is a systematic shift in the histogram away from zero. This is particularly evident in the case of CLIM-TROPICS and OBS-TROPICS, especially in the North Pacific sector where the distribution is strongly skewed. But even in the North Atlantic sector, some influence from the Tropics is apparent. A similar but weaker behaviour can be seen in OBS-NO (this is the experiment that sees the observed SST and sea-ice but with no relaxation) and CLIM-STRAT. For CLIM-STRAT, there is a perceptable (although weak) shift towards positive correlation in the North Atlantic sector, as one would expect given the known coupling between the stratosphere and the NAO (Baldwin and Dunkerton, 1999; Douville, 2009). In the North Pacific sector, the distribution is quite flat for CLIM-STRAT compared to the North Atlantic sector but whether any meaning should be attached to the strong kurtosis is a moot point.

### 3.2. The NAO

Next we turn to the NAO. Figure 2 shows the time series of the ensemble mean NAO index, the observed NAO index and the spread of the model realisations about the ensemble mean (the grey shading indicates the range of ensemble members within two standard deviations of the ensemble mean). Even in CLIM-NO, the experiment run using only climatological forcing and no relaxation, the observed NAO index sits within the range of values of the NAO index found in the experiment, indicating that the time series of the observed index is a possible realisation in the context of this experiment and therefore need not have any cause beyond natural variability (and even though this experiment uses specified climatological SST and sea-ice). Such a conclusion is consistent with Semenov *et al.* (2008), who looked at time series of several thousand years for the NAO in two coupled models and concluded that the observed NAO variability during the instrumental record is not particularly unusual (also Wunsch, 1999). A curious feature of CLIM-NO is the marginally significant (at the 95% level) negative correlation of −0.35 between the ensemble mean NAO index and the observed index (Table 1; note that after detrending the

 

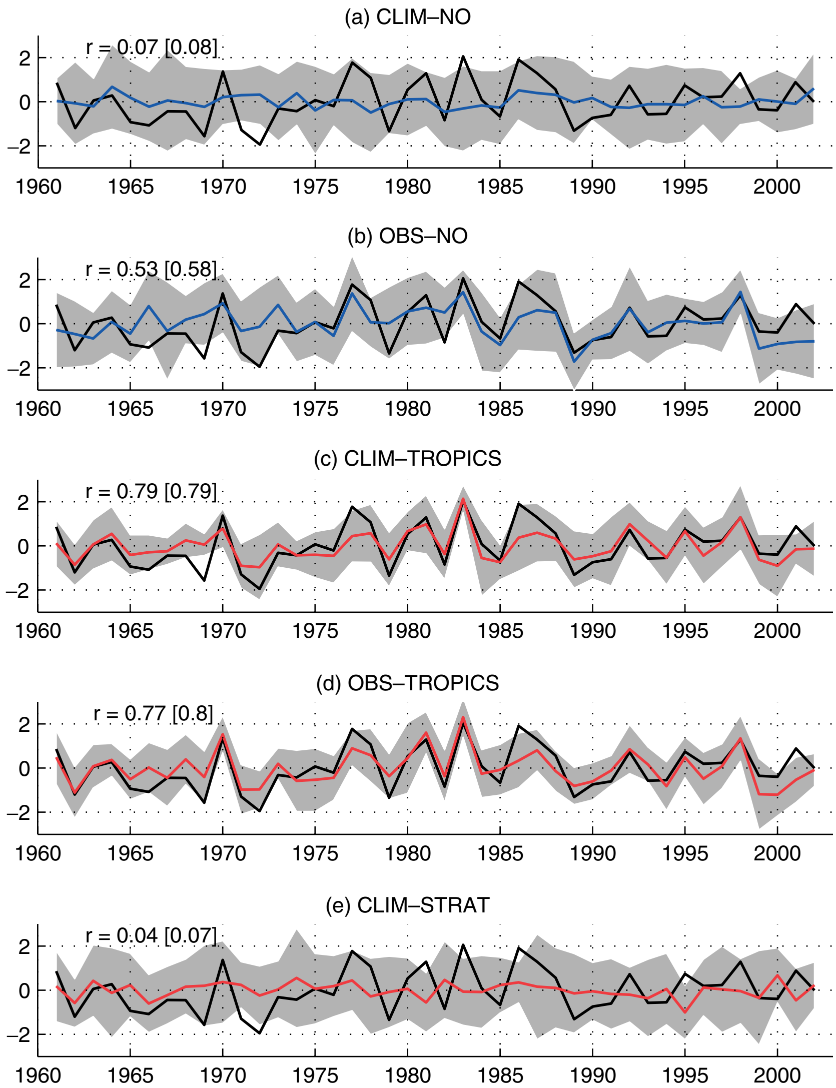


**Figure 4.** As Figure 2, but for the PNA index (black) at 500 hPa. The PNA index is defined as the weighted area mean of 500 hPa geopotential height anomalies over the region 30–65°N and 160°E–140°W, and multiplied by −1 so that positive index values correspond to a deeper than normal Aleutian low. Again, all indices are normalised by the standard deviation of the observed index.

correlation falls to −0.3). Since only the initial conditions distinguish one winter from another in this experiment, one might expect the correlation to be nearer zero. However, it is possible that there is memory of the initial conditions. For example, we know that the quasi-biennial oscillation in the stratosphere, present in the initial conditions, can influence the winter NAO (Boer *et al.*, 2008), although such an influence in CLIM-NO should lead to a positive, not a negative correlation, with the observed index. Examination of the spatial pattern of the point-by-point correlation between the ensemble mean fields from CLIM-NO and the reanalysis fields (Figure 4.8 in Gollan, 2012) shows (i) that regions where the correlations are significantly different from zero at the 95% level are spotty and (ii) that the pattern of the correlation fields changes between different 21 year windows. (i) and (ii) lead us to believe that the marginally significant correlation between the ensemble mean NAO index and the observed index in CLIM-NO is fortuitous. It should also be noted that in the case of the PNA, the correlation between the ensemble mean and the observed PNA index is effectively zero (Table 1).

In the other experiments (OBS-NO, CLIM-TROPICS, OBS-TROPICS and CLIM-STRAT), all of which include forcing from either relaxation and/or specified SST and sea-ice, the correlation between the ensemble mean NAO index and the observed NAO index is positive (Table 1). However it is only in OBS-TROPICS and CLIM-STRAT that the correlations for the detrended time series are significantly different from zero at the 95% level as determined using the Monte Carlo method (Question 1 in section 2.5; note that the threshold is strongly exceeded only in CLIM-STRAT). However, it is clear from Figure 2 that the amplitude of the ensemble mean signal in all experiments is quite small and not comparable to that of the observed time series. Our results contrast with those of Douville (2009), who not only found that the ensemble mean NAO index in his experiment with stratospheric relaxation has a similar amplitude to the observed NAO index, but that the ensemble mean NAO index is also correlated at 0.9 with the observed index –his Figure 1. We attribute these differences to the different model (and model set-up) being used in his study and view the difference between his results and ours as an indication of model sensitivity. To test the strength of the forcing in each of our experiments, Figure 3 (left column) shows histograms, using 50 bins, of the correlation between individual realisations of the time series of the

 

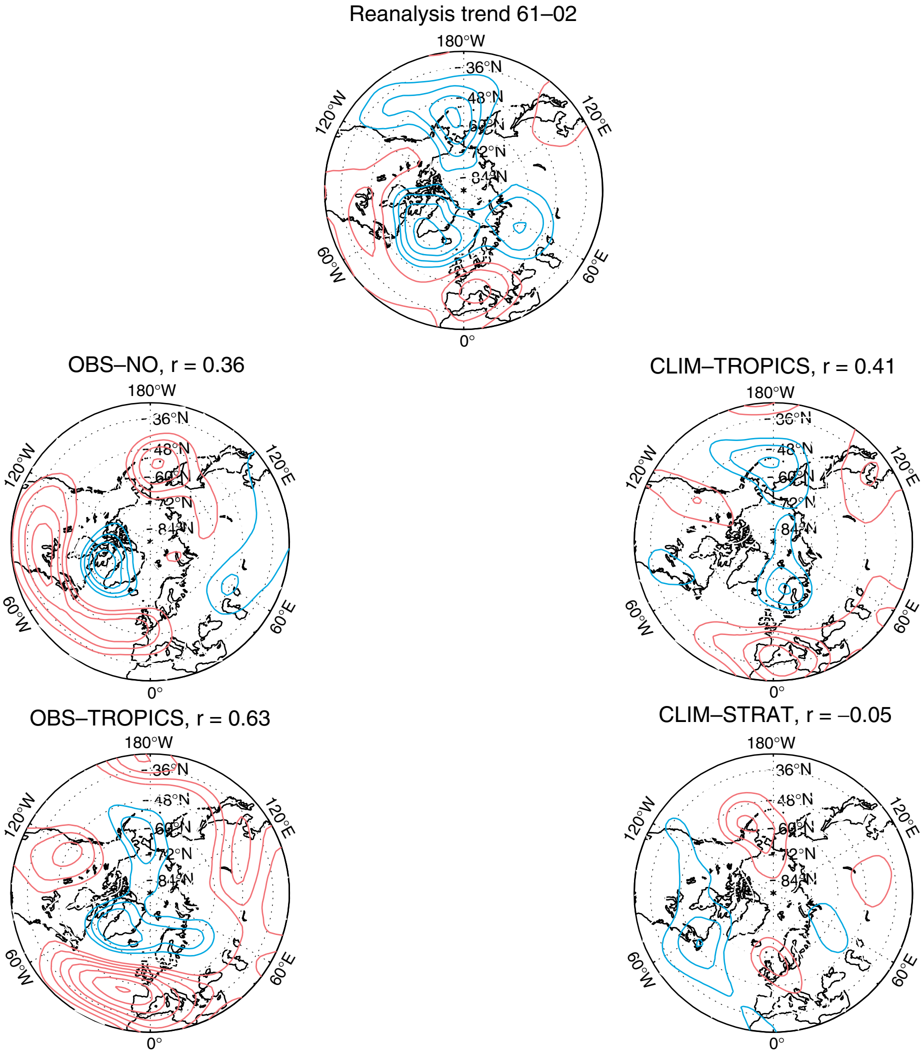


**Figure 5.** Linear trends of 500 hPa geopotential heights in the reanalysis (top) and ensemble means (middle and bottom) for winters from 1961 to 2002. Red denotes rising geopotential and blue lowering geopotential heights. The contour interval is 5 m $(10\,y)^{-1}$ for the reanalysis and 1.25 m $(10\,y)^{-1}$ for the ensemble means; the zero contour line is omitted. The pattern correlations ($r$) with the reanalysis are given above the plots.

NAO index from the experiments and the observed NAO index (Question 2 in section 2.5). In the case of CLIM-NO, the histogram is shifted to the left from zero and has the same cause as the negative correlation between the ensemble mean and the observed time series in this experiment noted earlier. In the other cases, the histograms are shifted noticably towards positive correlation. Both OBS-NO (observed SST and sea-ice specified globally) and CLIM-TROPICS (relaxation in the Tropics) show a similar shift. However, adding the observed extratropical SST and sea-ice to CLIM-TROPICS, as in OBS-TROPICS, clearly shifts the histogram further to the right. In both this experiment and CLIM-STRAT, the occurrence of negative correlations is extremely rare. The strong showing by CLIM-STRAT, in which the stratosphere is constrained to the reanalysis, is perhaps surprising in view of our finding when discussing Figure 1 that constraining the stratosphere does not greatly influence the pattern correlation between the circulation anomalies in the model and the observed circulation anomalies, even in the North Atlantic sector. These results seem to say that constraining the stratosphere has an impact on the NAO in the model (as implied by Baldwin and Dunkerton, 1999, for example) but is less effective at constraining other modes of variability in the North Atlantic sector (e.g. the East Atlantic and Scandinavian patterns; Rogers, 1990). Overall, we can say on the basis of these results that, on interannual time-scales, the Tropics do indeed influence the NAO in the model, (although weakly), that perfect knowledge of extratropical SST and sea-ice adds to that influence, and that perfect knowledge of the stratosphere has about the same level of influence. Of course, it is possible that some of the influence from the stratosphere is in reality initiated in the Tropics and that, similarly, influence from the Tropics is carried to the North Atlantic sector via the stratosphere (Ineson and Scaife, 2009, give an example), an issue for further study. Nevertheless, since the average correlation between individual model realisations and the observed NAO index is at best 0.3 in these experiments, none of the influences considered in our model experiments can account for more than about 10% of the observed interannual variance in the NAO index during the 42 winters being studied here. In the ensemble mean, this percentage increases upto as much as 25% in CLIM-STRAT, but still indicates that much of the interannual variance in the observed NAO index cannot be explained by the forcing that has been added to the model.

 

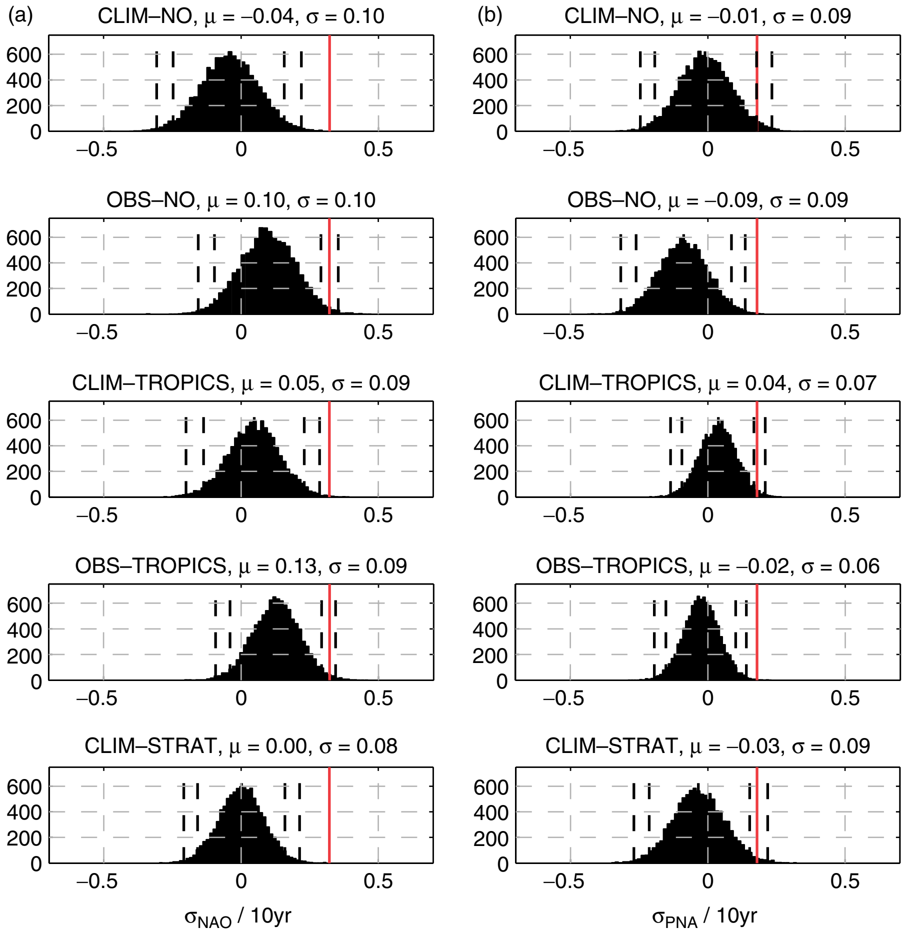


**Figure 6.** Histograms of 500 hPa NAO trends (left column) and PNA trends (right column) (in units of standard deviation, $\sigma_{NAO/PNA}$, per 10 years) between 1961 and 2002 in individual model realisations generated as described in section 2.5. Red lines show the NAO trend in the reanalysis. Black dashed lines indicate the 95% and the 99% ranges of the distribution. $\mu$ indicates the mean trend and $\sigma$ the standard deviation in each case.

### 3.3. *The PNA*

As for the NAO, the time series of the observed PNA index falls within the range of values for the PNA index found in CLIM-NO and so is a possible realisation under climatological SST and sea-ice. However, it is known that the PNA is strongly influenced by forcing from the tropical Pacific Ocean, e.g. in association with ENSO variability (Trenberth *et al.*, 1998), so it is not surprising to find a strong influence from the applied forcing in some of the other model experiments. Figure 4 shows the time series of the ensemble mean PNA index and the observed PNA index, together with the spread (as in Figure 2). The correlation between the ensemble mean index and the observed index is significantly different from zero at the 99% level in each of OBS-NO, CLIM-TROPICS and OBS-TROPICS (whether or not detrending is applied) and reaches near 0.8 for the time series (detrended or not) in CLIM-TROPICS and OBS-TROPICS. It is also clear from the time series that the amplitude of the ensemble mean signal in CLIM-TROPICS, for example, is much closer to that of the observed signal than is the case for the NAO. By contrast, the influence of the stratosphere, as seen in the CLIM-STRAT experiment, appears to be weak. This is confirmed in Figure 3 (right column) which shows the histogram of correlations of the individual realisations and the observed PNA index for the different experiments. Whereas the histogram for CLIM-STRAT is centred close to zero, indicative of, at best, very weak influence from the stratosphere, the influence of the added forcing in CLIM-TROPICS and OBS-TROPICS is particularly strong. Not only does the average correlation between the individual model realisations and the observed index lie between 0.6 and 0.7 (and so accounting for almost 50% of the variance in the observed time series, rising to more than 60% in the ensemble mean) but the histograms are also much narrower in the CLIM-TROPICS and OBS-TROPICS cases than in the other experiments. Both the strong shift to the right and the narrowness of the distributions indicate the strong constraint imposed by tropical forcing, with the suggestion of some additional narrowing of the distribution from the observed extratropical SST and sea-ice.

### 3.4. *Trends*

We begin by looking at the trend over the whole time period 1960/61 to 2001/02 and then take a closer look at the subperiod 1965–1995 which was studied by Scaife *et al.* (2005) and during which the NAO showed a particularly strong upward trend. Figure 5 shows the pattern of the trend in the ensemble mean of selected experiments compared with the same trend from the ERA-40 reanalysis. For the ensemble mean trends, the pattern correlation with the reanalysis trend is shown above the figure. It should be noted that in the plots using model output, the contour interval is only one quarter of that used for the reanalysis

 

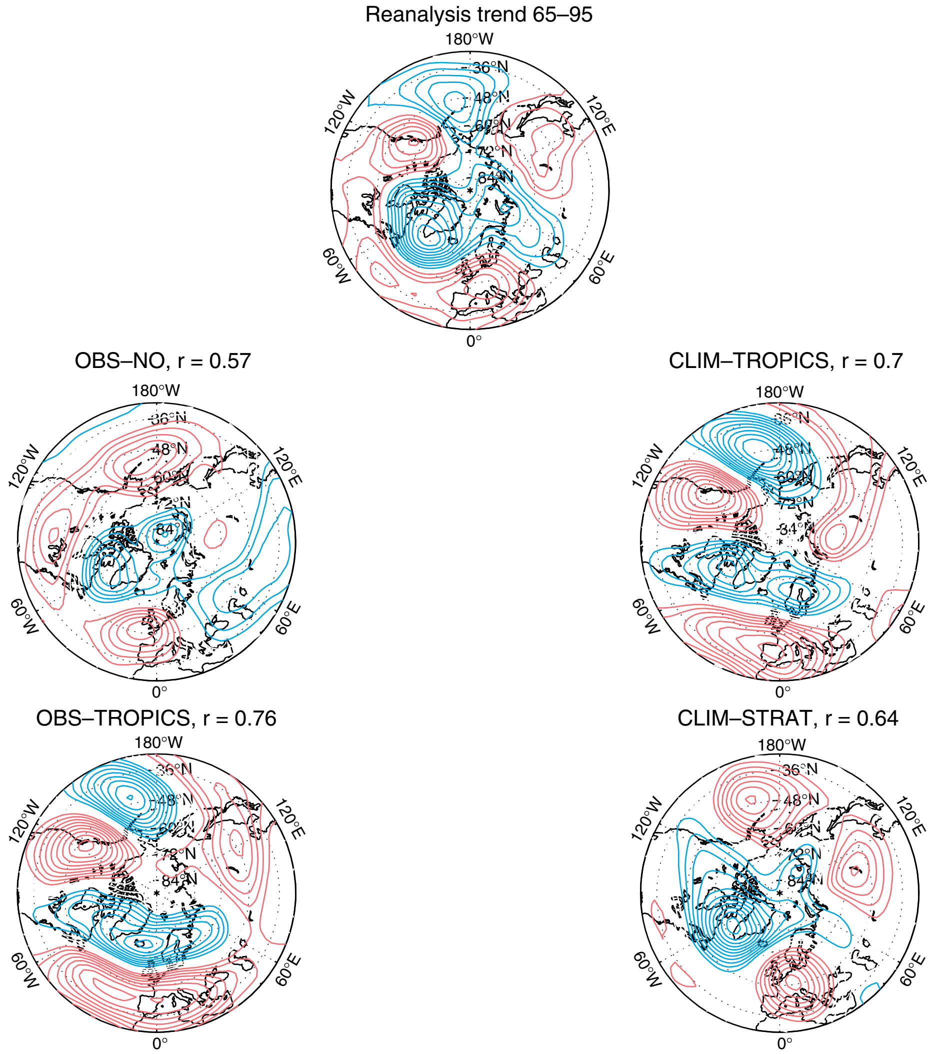


**Figure 7.** As Figure 5, but for the subperiod 1965 to 1995.

trend, an indication that the amplitude of the trends in the ensemble mean have reduced amplitude compared to the reanalysis trend. Similar to the period 1949–1999 studied by Lu *et al.* (2004) (also Ostermeier and Wallace, 2003), the reanalysis trend over the 42-year period resembles the Cold Ocean Warm Land (COWL; Wallace *et al.*, 1996) pattern and is associated with a deepening of both the Aleutian and Icelandic lows. The best performance in terms of pattern correlation is obtained in OBS-TROPICS, although it is clear that the trend in the Pacific sector is less well reproduced in both the experiments (OBS-NO and OBS-TROPICS) that use the observed SST and sea-ice (an issue discussed further later). Comparing CLIM-TROPICS and OBS-TROPICS, we see that, while the addition of the observed extratropical SST and sea-ice in OBS-TROPICS degrades the model performance in the Pacific sector, it leads to a stronger trend towards a more positive NAO state over the North Atlantic. However, the ensemble mean trend in CLIM-STRAT compares poorly with the reanalysis trend when considered over the full 42 years, an issue we return to below.

Figure 6 shows histograms of the trends for the NAO (left column) and PNA (right column), from selected model experiments; the histograms are derived using the Monte Carlo method, as under Question 3 in section 2.5. In each case, the trend in the reanalysis is shown by the red vertical line. It can be seen that, for the NAO, the reanalysis trend over the 42-year period would be an unusual event in the context of all the model experiments, but mostly likely to occur in the case of OBS-NO and OBS-TROPICS, the two cases that use the observed SST and sea-ice. In these experiments, the reanalysis trend falls between the 95% and 99%iles, pointing, as noted earlier, to an influence (although small) for extratropical SST and sea-ice on the trend in the NAO in the reanalysis. For the PNA (Figure 6, right column), the reanalysis trend is actually most likely to occur in the CLIM-NO ensemble for which there is no added forcing. For CLIM-TROPICS, the case with tropical relaxation, it falls between the 95% and the 99%iles and is therefore a possible realisation in the model. The degradation over the Pacific sector obtained by introducing the extratropical SST and sea-ice is apparent by comparing CLIM-TROPICS and OBS-TROPICS. Interestingly these experiments suggest that, while using observed SST and sea-ice improves the trend over the North Atlantic sector, it generally degrades the trend in the model over the Pacific sector.

Scaife *et al.* (2005) have argued that the stratosphere played an important role in dynamics of the strong upward trend in the NAO index over the North Atlantic during

 

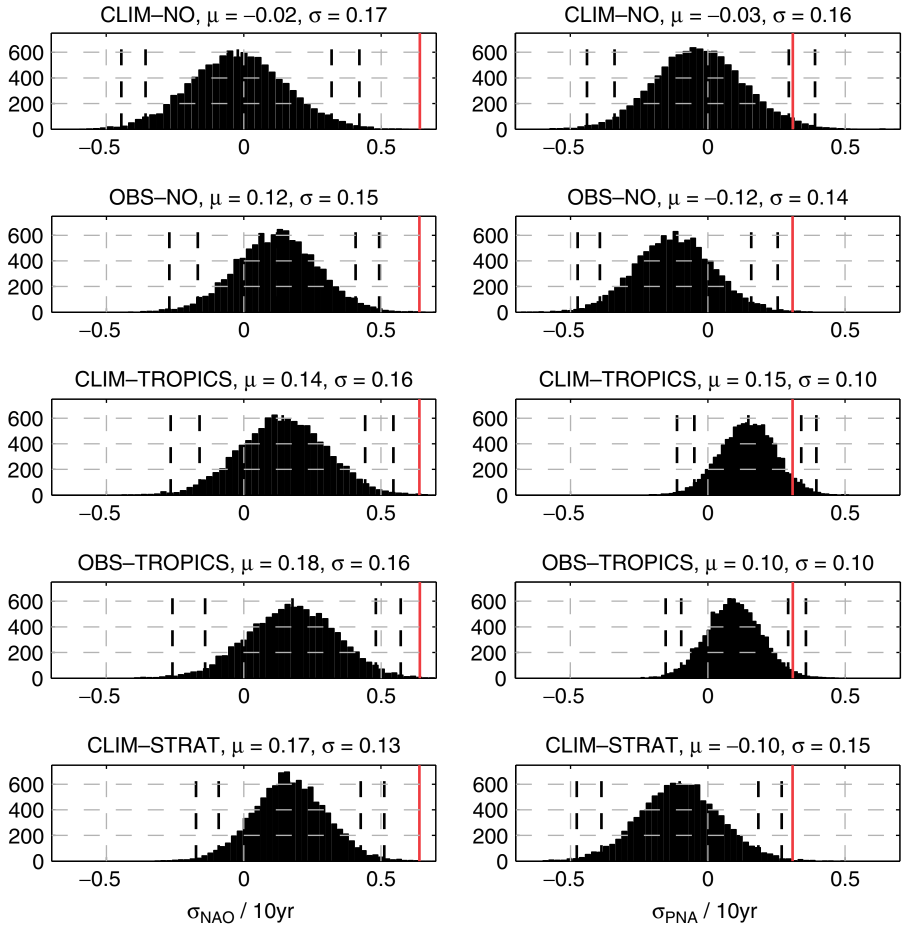


**Figure 8.** As Figure 6, but for the subperiod 1965 to 1995.

the period 1964/65–1994/95. Figure 7 shows the ensemble mean trends in our model experiments, as for Figure 5 but for the subperiod 1964/65–1994/95, together with the corresponding reanalysis trend. For the subperiod 1964/65–1994/95, CLIM-STRAT shows a much better performance than before, the pattern correlation rising from −0.05 for the period 1960/61–2001/02 to 0.6 for the subperiod 1964/65–1994/95. It is clear that the improved performance comes from the Atlantic/Eurasian sector, while the stratospheric influence on the Aleutian low has the wrong sign, as for the full period 1960/61–2001/02. These results add support to Scaife *et al.* (2005), but also argue that the influence of the stratosphere on the Atlantic sector can depend on the time period being considered (a topic for further research). Also, whereas Scaife *et al.* (2005) claim to be able to account for the full amplitude of the upward trend of the NAO during this period, here we can account for at most 30%. This difference is similar to what we found in section 3.2, when comparing our results using stratospheric relaxation with those from Douville (2009) for the interannual variability of the NAO, and probably are a reflection of model sensitivity associated with using different models and different model set-ups. Turning to experiments CLIM-TROPICS and OBS-TROPICS, there is a clear influence of tropical forcing over both the Pacific and Atlantic sectors. Of course, as for the interannual variability, influences from the stratosphere may be of tropical origin and influences from the Tropics may be communicated to the Atlantic sector via the stratosphere (Ineson and Scaife, 2009). Histograms of the trend from individual model realisations are shown for the subperiod 1964/65–1994/95 in Figure 8. For the NAO, the reanalysis trend is an unlikely event in all experiments, including CLIM-STRAT, since it lies outside the 99% range in every case, despite the clear shift away from zero in all the histograms that use added forcing (that is, all except CLIM-NO). For the PNA, the reanalysis trend is most likely to occur in CLIM-TROPICS, where it falls inside the 95% range but is now more likely to occur in OBS-TROPICS than for the period 1961–2002.

A feature of both time periods is the much improved performance of CLIM-TROPICS compared to OBS-NO in the Pacific sector, suggesting that the error in OBS-NO is in the Tropics and, in particular, in the model response to the specified SST (Copsey *et al.*, 2006, provide a discussion of this issue). Indeed, the much improved performance of CLIM-TROPICS nicely illustrates the advantage of using the relaxation technique in an atmosphere-only model compared to specifying the tropical ocean SST. In the latter case, coupled ocean–atmosphere interactions are missing whereas these effects are fully included by using tropical relaxation. Likewise, differences between CLIM-TROPICS and OBS-TROPICS (especially in the 1960/61–2001/02 period) reflect the degrading influence of extratropical SST and sea-ice on the model's ability to reproduce the trend in the Pacific sector. Finally we note that the importance of tropical forcing over the Pacific sector, and the influence from the stratosphere over the Atlantic sector, which we have found here is consistent with Blessing *et al.* (2008) who used an adjoint approach to determine the optimal forcing for exciting the observed trend over the period 1948/49 to 1998/99.

## 4. Summary

We have investigated the impact of influences from the Tropics, the stratosphere and the specification of observed SSTSI on the simulation, by a recent version of the ECMWF model, of Northern Hemisphere winter mean circulation anomalies during the period 1960/61 to 2001/02. Results from a number of different experiments have been reported, some of which use a relaxation technique (Jung *et al.*, 2010a,b) in which the model is relaxed towards the reanalysis in either the Tropics or the stratosphere. The Tropics are found to act as a major constraint on the interannual variability of the PNA pattern but also influence the interannual variability in the North Atlantic sector, although weakly. The stratosphere is found to be particularly influential on the interannual variability of the NAO but otherwise seems to be unimportant for other modes of interannual variability in the North Atlantic sector and appears to have no significant influence over the Pacific sector on interannual time-scales. Adding the observed extratropical SSTSI to the tropical relaxation runs generally improves the model performance on the interannual time-scales in both sectors, but degrades/enhances the model's ability to capture the 42-year trend over the Pacific/Atlantic sector. While relaxing the stratosphere to the reanalysis gives a poor performance at reproducing the trend in the ERA-40 reanalysis over the full 42-year period, the stratosphere is shown to be influential in the upward trend of the NAO index from 1964/65 to 1994/95, consistent with Scaife *et al.* (2005) and Blessing *et al.* (2008), although the influence of the Tropics (perhaps communicated to the Atlantic sector via the stratosphere; Ineson and Scaife, 2009) is also important. The results suggest that the influence from the stratosphere can vary in importance between different time periods. Overall, however, we find a weaker influence from the stratosphere than in the studies of Scaife *et al.* (2005) and Douville (2009). The experiment that uses observed SSTSI (and no relaxation) is notable in its failure to capture the trend over both time periods in the Pacific sector, a failure we attribute to inaccuracy in the model's response to the specified SST in the Tropics (e.g. Copsey *et al.*, 2006), indicating the need to consider coupled ocean/atmosphere processes. Across all experiments, we are unable to account for more than a small fraction (at best 25% using the ensemble mean) of the interannual variance of the winter mean NAO, although a significantly higher fraction (60%) in the case of the PNA. The amplitude of the ensemble mean trends produced by the model are also typically one quarter to one third of the corresponding trends found in the reanalysis (Figures 5 and 7; note that the contour interval used to plot the model results is one quarter of that used to plot the trends in the reanalysis).

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