

A stratospheric influence on the winter NAO and North Atlantic surface climate

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[1] The North Atlantic Oscillation (NAO) has a profound effect on winter climate variability around the Atlantic basin. Strengthening of the NAO in recent decades has altered surface climate in these regions at a rate far in excess of global mean warming. However, only weak NAO trends are reproduced in climate simulations of the 20th Century, even with prescribed climate forcings and historical sea-surface conditions. Here we show that the unexplained strengthening of the NAO can be fully simulated in a climate model by imposing observed trends in the lower stratosphere. This implies that stratospheric variability needs to be reproduced in models to fully simulate surface climate variations in the North Atlantic sector. Despite having little effect on global mean warming, we show that downward coupling of observed stratospheric circulation changes to the surface can account for the majority of change in regional surface climate over Europe and North America between 1965 and 1995. **Citation:** Scaife, A. A., J. R. Knight, G. K. Vallis, and C. K. Folland (2005), A stratospheric influence on the winter NAO and North Atlantic surface climate, *Geophys. Res. Lett.*, 32, L18715, doi:10.1029/2005GL023226.

1. Low Frequency Variations in the NAO and the Stratosphere

[2] The NAO is an intrinsic mode of variability, accounting for half of the year-to-year variability in winter surface temperature over Northern Europe [Rodwell *et al.*, 1999] and a third of the variability in northern hemisphere surface temperature [Hurrell, 1996]. It corresponds to variations in the Atlantic storm track and is localised by regions of strong baroclinic eddy activity. Whereas the fundamental dynamics of the NAO are becoming clearer [Ambaum and Hoskins, 2002; Vallis *et al.*, 2004], it remains an outstanding problem to explain past variations in the NAO and to predict its future state. This is important because climate change signals can project onto the NAO [Gillett *et al.*, 2003; Thompson *et al.*, 2000] and internal variability from the NAO is still larger than the total anthropogenic change over some regions. The NAO has been measured in various ways, including the surface pressure difference between the Azores and Iceland [Jones *et al.*, 1997] (Figure 1) which shows large multidecadal variability. The 1960s exhibit lower values than any other decade, related to the cold North European winters such as that of 1962–1963 [Murray,

1966]. The trend between the 1960s and the 1990s exceeds that over any similar period and culminated in a series of warm stormy winters in Northern Europe [Alexandersson *et al.*, 1998]. The total change in pressure over the Atlantic in this period (Figure 1) projects strongly onto the NAO and has a dipole centred near 50°N and 30°W. Sea-level pressure changes between the 1960s and 1990s can therefore be thought of as an 11–12 hPa change in the NAO. The rapid shift towards more positive NAO since the 1960s is not well reproduced in climate models, even with prescribed sea-surface conditions from observations. Although a minimum in the NAO in the 1960s and a positive trend over the following decades is reproduced [Rodwell *et al.*, 1999], the magnitude of the trend is no more than half of the observed trend and only after normalisation do models and observations appear to agree [Rodwell *et al.*, 1999; Mehta *et al.*, 2000; Latif *et al.*, 2000; Cassou and Terray, 2001; Hoerling *et al.*, 2001]. It has been argued that this is due to the specification of sea surface temperature in models, when in reality the atmosphere strongly influences sea surface temperature but is itself only weakly influenced by the ocean [Bretherton and Battisti, 2000]. Nevertheless, model simulations with interactive oceans are also unable to simulate the recent NAO trend [Gillett *et al.*, 2003]. Some of the observed trend in the NAO and European climate has recently been attributed to anthropogenic change [Gillett *et al.*, 2003]. However, the combined effect of anthropogenic forcing and internal variability is still too small in current models to explain the observed trend [Osborn, 2004] and the NAO trend therefore remains unexplained.

[3] In addition to studies of the surface boundary influence on the NAO, a few studies suggest a possible role for stratospheric conditions. Modelling experiments where the stratospheric circulation is weakened (strengthened) show a surface response that strongly resembles low (high) periods of the NAO [Boville, 1984; Polvani and Kushner, 2002; Norton, 2003]. Both observations and model experiments show that the surface effect appears within a few weeks [Baldwin and Dunkerton, 2001; Charlton *et al.*, 2004]. Stratospheric radiosonde observations exist for the period since the late 1950s [Pawson and Fiorino, 1998]. Satellite observations of the lower stratosphere give good global coverage from the late 1970s [Bailey *et al.*, 1993], and from the early 1990s data assimilation has been used to produce comprehensive stratospheric analyses [Swinbank and O'Neill, 1994]. These data show that over the 1965–1995 period when the NAO was increasing, lower stratospheric wind increased by approximately 7 ms⁻¹ (Figure 2). Marked multiannual covariability is also found in the NAO and stratospheric winds (Figure 2). The correlation coefficient between the NAO and lower stratospheric wind is close to 0.5 for the whole time series, and rises to 0.8 for the 1965–1995 period. In this paper we test how this

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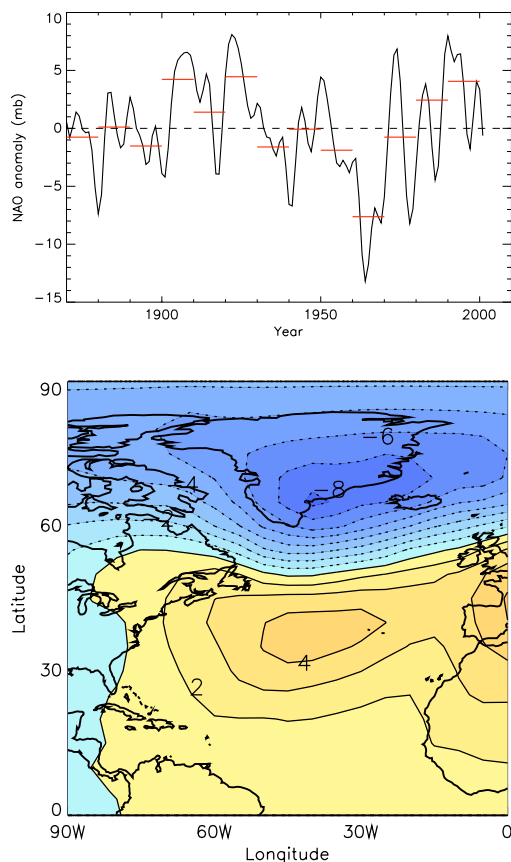


Figure 1. North Atlantic Oscillation index. Upper, Winter (DJF) NAO index 1867–2002 based on the difference in pressure at mean sea level between the Azores and Iceland [Jones *et al.*, 1997]. Three applications of a 1-2-1 filter have been made, corresponding to half amplitude at 6.7y. Decadal means are shown in red. Lower, change in winter mean sea-level pressure over the Atlantic region between the 1990s and the 1960s from the Hadley Centre HadSLP3 data set. Units are hPa.

stratospheric trend between 1965 and 1995 is likely to have affected surface climate.

2. Numerical Experiments

[4] A control experiment comprising of 6 simulations for the latter half of the 20th century was carried out with the 19 level Hadley Centre global atmospheric model, HadAM3 [Pope *et al.*, 2000]. The ensemble members differ only in their initial conditions and are forced with time-varying forcings from well mixed greenhouse gases including CO₂, CH₄, N₂O, CFC13, CF2Cl₂, tropospheric ozone and stratospheric ozone changes from 1975, changes in surface albedo and vegetation, anthropogenic sulphate and volcanic aerosols, and solar irradiance variations. Ocean surface temperatures and sea-ice were specified from historical observations [Rayner *et al.*, 2003]. This control experiment shows only a shallow minimum in the NAO index in the 1960s and a subsequent weak upward trend (Figure 3). The timing of the minimum agrees well with the observed NAO, but the control does not reproduce the deep negative NAO

values observed in the 1960s. The control also did not reproduce the observed trend in stratospheric wind; the modelled trend was only 1 ms^{-1} over the 1965–1995 period.

[5] We also ran a perturbation experiment of 2 simulations from December 1964, using 2 randomly selected start conditions from the 6 used in the control ensemble. These simulations are identical to the control except for a perturbation that mimics the observed trend towards stronger stratospheric westerlies. The method used to apply the perturbation is similar to that used in previous studies [Norton, 2003]. We damped the stratospheric zonal wind with a Rayleigh drag coefficient that increases linearly with altitude. The drag is set to zero just above the tropopause at 17 km, has a timescale of 2.5 days near the stratopause, and decays linearly in time from 1965 to zero in 1995. This perturbation results in an upward trend of 8.5 ms^{-1} over the 1965–1995 period in the lower stratospheric zonal wind at 50 hPa and 60°N, in contrast to the almost constant stratospheric winds in the control experiment. It agrees reasonably well with the observed trend of 7 ms^{-1} over the same period. Because the perturbation is only applied to winds above the tropopause it represents a clear test of the influence of transient stratospheric change on surface climate.

3. Modelled Stratospheric Influence on the Troposphere

[6] The imposed trend in stratospheric wind induces a dramatic increase in the surface NAO (Figure 3). The trend is very similar in both perturbed simulations. The 1965–1995 NAO increase is $14.2 \pm 5.0 \text{ hPa}$ compared with just $2.4 \pm 3.7 \text{ hPa}$ in the control (Figure 3). This agrees with the observed NAO change of $10.4 \pm 4.6 \text{ hPa}$. As far as we are aware, this is the first simulation with a comprehensive GCM of the full trend in the NAO between the 1960s and the 1990s. A similar result holds for the hemispheric wide

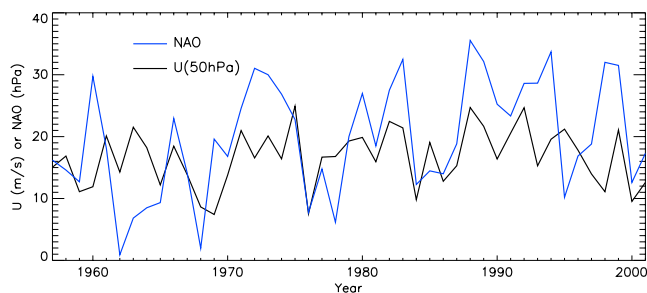


Figure 2. Winter NAO and stratospheric circulation. Winter stratospheric wind (black) and NAO index [Jones *et al.*, 1997] (blue). Winds are zonal averages at 60°N and 50 hPa to represent the lower stratosphere while being well above the tropopause. They are means of radiosonde analyses from the Free University of Berlin (1957–1996), Met Office SSU analyses (1979–1996) and Met Office assimilated analyses (1991 onwards). Geostrophic winds from FUB and SSU data have been scaled by a constant factor (approx. 0.9) to be consistent with the assimilated (non-geostrophic) winds. Year labelling corresponds to December.

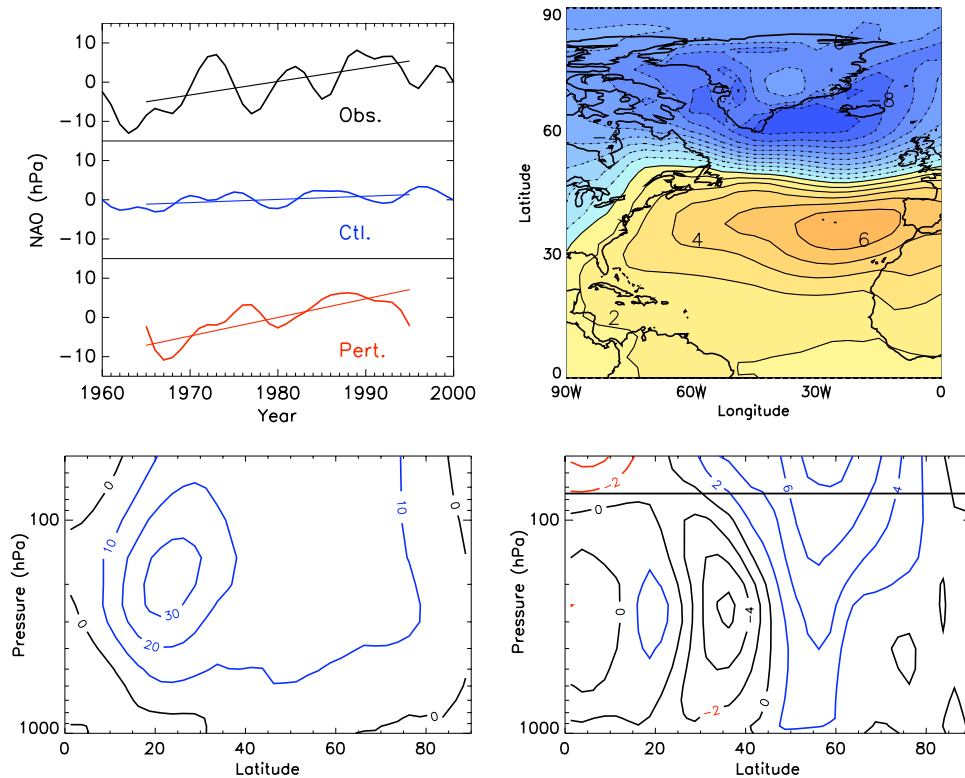


Figure 3. Modelled tropospheric response to the stratospheric circulation trend. Upper left, Winter NAO index in observations (black), the ensemble means of 6 control simulations (blue) and 2 perturbed stratosphere simulations (red) with 30 year linear trends superimposed and smoothing as in Figure 1. Upper right, 30 year change in sea-level pressure in the model (hPa) due to the imposed stratospheric circulation trend for comparison with Figure 1. Lower left, simulated zonal winds (ms^{-1}) in the control experiment. Lower right, difference between zonal mean wind in perturbed stratosphere and control simulations. Direct perturbations to the model were only applied above the black horizontal bar.

Arctic Oscillation (AO), defined as the first EOF of the pressure field. Our perturbation experiment shows an average increase of 1.12 standard deviations in the winter AO while the control ensemble increases by just 0.18 standard deviations (the EOF patterns have similar strength in the two ensembles). As suggested by observational analyses [Baldwin, 2003], stratospheric variations are therefore linked to the hemisphere wide AO and produce a similar but slightly weaker dipole anomaly over the Pacific, as well as influencing the more regional NAO in our model. Our model produces a near-barotropic tropospheric response to the imposed stratospheric perturbation (Figure 3) that is qualitatively similar to tropospheric signals in more idealised models that omit water vapour or stationary planetary waves [Polyani and Kushner, 2002]. When the stratospheric winter jet strengthens, the response is strengthening of the tropospheric westerlies at mid- to high-latitudes, weaker westerlies at lower latitudes and an increase in the NAO index. This is consistent with a mechanism of ‘downward control’ amplified by a feedback involving baroclinic eddies [Song and Robinson, 2004]. Use could also be made of this response as a “dial” to control the NAO in other perturbation experiments where different NAO states are needed.

[7] Surface temperature in our experiment shows a large, quadrupolar response to the stratospheric circulation change (Figure 4), compared to weak uniform changes in the control simulation. There are areas of warming over Europe and southern North America and 2 areas of cooling, one

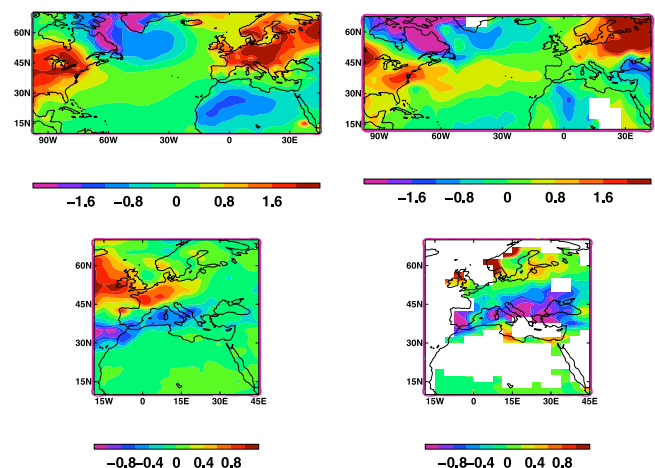


Figure 4. Surface climate response to the stratospheric circulation trend and comparison with observed changes. Upper left, modelled winter surface temperature change. Upper right, observed winter surface temperature change (K). Lower left, modelled winter precipitation change. Lower right, observed winter precipitation change over land (mm day^{-1}). Differences between 1990–95 and 1965–70 are shown. Model results are the difference between perturbed and control experiments.

over northeastern North America and Greenland, the other over southern Europe and northern Africa. This signal is remarkably similar in both pattern and magnitude to the surface temperature change that occurred between the 1960s and the 1990s (Figure 4). It appears to have dominated regional surface temperature trends over that period. For example, over Northern Europe (10W–50E and 50N–70N) the observed winter surface temperature trend was 0.53 K decade⁻¹. The perturbed stratosphere experiment reproduces 0.59 K decade⁻¹; in good agreement with the observations. Our control experiment reproduced a trend of only 0.15 K decade⁻¹ despite the inclusion of a comprehensive set of historical climate forcings. This effect is also responsible for much of the observed increase in rainfall over northern Europe and the decrease in rainfall over southern Europe during the same period, although regional differences between the model response and the observed precipitation change occur in some places such as Scandinavia (Figure 4). In summary, surface temperature and precipitation trends, as well as the NAO, were strongly influenced by the stratospheric circulation between 1965 and 1995.

4. Discussion

[8] Although our experiment reproduces observed change in regional surface climate, this does not necessarily imply stratospheric control of surface climate because the stratospheric trend could have been driven from the troposphere [Plumb and Semeniuk, 2003]. Positive feedback between tropospheric and stratospheric climate also seems likely because a positive surface NAO pattern corresponds to a more zonal tropospheric flow with weaker planetary waves. This results in strong stratospheric westerlies which will subsequently strengthen the positive NAO anomaly at the surface as shown here. A similar feedback may also strengthen negative NAO anomalies. Note that the model responds to stratospheric change in the very first winter of the perturbation experiment (Figure 3). Given the short timescale between stratospheric and subsequent NAO changes, we expect that interannual variations in the NAO might also be reproduced if we relaxed the stratosphere to its observed state on interannual timescales.

[9] Our results may also reconcile seemingly opposed results on the importance of stratospheric resolution for modelling the NAO [Shindell et al., 1999; Gillett et al., 2002] because we have shown that to capture effects on the NAO, it is important to simulate the low frequency variability in the stratospheric circulation rather than to simply enhance stratospheric model resolution. Reproducing this variability in models may be difficult though because large inter-decadal variations often arise through internal variability [Butchart et al., 2001]. Changes in the NAO and regional temperature (Figure 4) over the last few decades may therefore also be largely unpredictable but this does not affect the anthropogenic interpretation of increasing global mean temperature because the surface temperature signal associated with the NAO is a quadrupole pattern with little impact on the global mean.

[10] Regardless of the predictability of the NAO and whether it is controlled mainly by the troposphere or the stratosphere, we have demonstrated that stratospheric trends over the last few decades and the downward links to surface

climate are strong enough to explain much of the prominent trend in the NAO and regional climate over Europe and North America between the 1960s and the 1990s.

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