European Climate Extremes and the North Atlantic Oscillation

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ABSTRACT

The authors estimate the change in extreme winter weather events over Europe that is due to a long-term change in the North Atlantic Oscillation (NAO) such as that observed between the 1960s and 1990s. Using ensembles of simulations from a general circulation model, large changes in the frequency of 10th percentile temperature and 90th percentile precipitation events over Europe are found from changes in the NAO. In some cases, these changes are comparable to the expected change in the frequency of events due to anthropogenic forcing over the twenty-first century. Although the results presented here do not affect anthropogenic interpretation of global and annual mean changes in observed extremes, they do show that great care is needed to assess changes due to modes of climate variability when interpreting extreme events on regional and seasonal scales. How changes in natural modes of variability, such as the NAO, could radically alter current climate model predictions of changes in extreme weather events on multidecadal time scales is also discussed.

1. Introduction

In the last decade there has been much interest in assessing whether climatic extremes have been changing because changes in temperature and rainfall extremes, in particular, are expected to be an important result of greenhouse warming on multicontinental to global scales (Karl et al. 1995; Frich et al. 2002; Kiktev et al. 2003; Groisman et al. 2005; Alexander et al. 2006) and for Europe (e.g., Klein Tank et al. 2002; Klein Tank and Können 2003; Moberg et al. 2006). These studies found observational evidence of increases in the frequency of various kinds of extreme precipitation and decreases in the frequency of low temperature events over the last few decades. However, in common with most studies, little attempt was made to separate out regional and seasonally varying atmospheric circulation influences, such as the North Atlantic Oscillation (NAO) (or the closely related Northern Annual Mode), from effects due to global warming.

The influence of the NAO on mean winter climate is now well established (Hurrell 1995). In its positive phase, the NAO corresponds to enhanced westerly flow over the North Atlantic and a northward shift of the midlatitude storm track. The signature of the positive phase of the NAO in surface temperature corresponds to warmer conditions in northern and central Europe and much of the eastern United States and cooler conditions over the Mediterranean, eastern Canada, and Greenland. This quadrupole pattern of temperature arises primarily from low-level temperature advection.
by the altered low-level winds over the Atlantic (Hurrell and van Loon 1997). NAO signals can also be found in the mean winter rainfall over Europe (Wibig 1999) but are generally less clear than temperature signals due in part to the restriction of long records of precipitation to land stations with incomplete spatial coverage. The pattern of rainfall changes over Europe due to a shift in the NAO has a similar structure to the change in surface temperature. During high NAO periods, higher winter precipitation accompanies higher temperatures over northern Europe, while lower precipitation accompanies lower temperatures over the Mediterranean and extratropical North Africa, although the largest changes in rainfall may occur over the Atlantic Ocean (Scaife et al. 2005).

Over the latter half of the twentieth century there have been notable increases in extreme winter precipitation over northern and central European regions (e.g., Groisman et al. 2005) and significant reductions in the frequency of frosty nights (e.g., Alexander et al. 2006), although the changes in strong precipitation extremes generally occur more to the north compared to those for frosty nights (Klein Tank and Können 2003). These results are broadly qualitatively consistent with climate model predictions of changes in climate extremes due to anthropogenic greenhouse gas emissions, although much more work has been done to quantify possible future changes in extremes (e.g., Durman et al. 2001; Meehl et al. 2005; Barnett et al. 2006; Tebaldi et al. 2006; Frei et al. 2006) than to verify the reproduction of past observed changes. As with other changes in extremes, less has been done to quantify how changes in modes of climate variability, such as the NAO, impact changes in extreme rainfall and temperature. Despite this, Wettstein and Mearns (2002) examined changes in temperature extremes over the North American continent between the 1920s and 1990s and found signals from increases in the Arctic Oscillation (AO)/NAO. These included increases in maximum temperature extremes in New England and decreasing minimum temperatures in Quebec. Easterling et al. (1997) also noted a reduction in the winter diurnal temperature range over Europe over 1950–93 associated with the increased frequency of the positive phase of the NAO. More recently, Thompson and Wallace (2001) have used observational reanalyses to show that the frequency of occurrence of cold events over the January to March period is strongly affected by the NAO throughout the extratropical Northern Hemisphere. They also pointed out that some of the change in observed frequency of cold events was due to a change in the shape of the distribution rather than just a mean climate shift. Osborn et al. (2000) found significant changes in U.K. precipitation extremes over the last few decades but they did not find any significant link with the NAO. However, a recent analysis (Haylock and Goodess 2004) has been carried out on extreme precipitation using rainfall data for the whole of Europe and the period 1958–2000. They showed a significant correlation between the NAO and the frequency of heavy daily winter rainfall events. This relationship was to a considerable extent interannual, but also included a significant trend due to the coincident trend in the NAO to more positive values.

The use of numerical models to study this problem has been hampered by the inability of general circulation models (GCMs) to reproduce the large observed increase in the NAO over the latter part of the twentieth century. This is true not only of coupled ocean–atmosphere models (e.g., Gillett et al. 2003; Osborn 2004; Stenchikov et al. 2006) but also of simulations of the twentieth-century climate with observed climate forcings and observed ocean conditions (Cohen et al. 2005). Even the internal atmospheric circulation variability in GCMs is unlikely to explain the observed increase in the NAO (Kuzmina et al. 2005). This may be due to inadequacies in model representation of NAO variability, including the poor representation of positive feedback between the troposphere and stratosphere in these models. The NAO is characterized by a deep, nearly barotropic structure with zonal wind signatures extending from the troposphere to the stratosphere (Thompson et al. 2003). It may be essential that models are able to simulate these dynamical links so as to produce realistic variability. Furthermore, it has recently been shown in GCM experiments that the magnitude of decadal variations in the lower-stratospheric winds and the strength of their downward influence are large enough to reproduce the observed rapid increase in the surface NAO over the 1960s to 1990s (Scaife et al. 2005). By imposing stratospheric anomalies, these experiments also allowed the surface NAO change and the observed mean surface temperature and mean precipitation changes to be reproduced. In this paper, we use these model experiments to determine the impact of changes in the NAO on the frequency of extreme winter temperature and rainfall events. We compare the modeled changes with some of the observational results of changing temperature and precipitation extremes over Europe and estimate the extent to which changes in extreme winter precipitation and temperature can be explained by the trend in the NAO between the 1960s and 1990s. A brief analysis is also made of observed relationships between the NAO and European temperature and precipitation extremes over the
entire twentieth century, to obtain a longer time perspective.

2. Model experiments and mean climate responses

We make use of two ensembles of model simulations carried out with the 19-level Third Hadley Centre Coupled Atmospheric Model (HadAM3), which has a spatial resolution of 2.5° latitude by 3.75° longitude (Pope et al. 2000). The first ensemble is a control (CTL) comprising six simulations for the latter half of the twentieth century. The ensemble members differ only in their initial atmospheric conditions and are forced with time-varying boundary conditions from well-mixed greenhouse gases including CO$_2$, CH$_4$, N$_2$O, CFCl$_3$, CF$_2$Cl$_2$, tropospheric and stratospheric ozone, changes in surface albedo and vegetation, anthropogenic sulfate and volcanic aerosols, and solar irradiance variations. Apart from ozone (which is held at constant climatological values until 1975 and varies thereafter), all other forcings are annually varying from the start of the simulations according to observed estimates (Johns et al. 2003). Variations in ocean surface temperature and sea ice extent were specified from an analysis of historical observations (Rayner et al. 2003).

The CTL ensemble shows only a shallow minimum in the NAO index in the 1960s and a subsequent weak upward trend (Fig. 1). The timing of minimum values of the simulated NAO in the 1960s agrees well with the observed NAO, but the rate of increase of the NAO index in the control experiment is several times smaller than the observed increase from the 1960s to the 1990s.

The second ensemble (NAO↑) consists of two simulations for the period 1964 to 1995, using two randomly selected start conditions from the six used in the control ensemble. These simulations are identical to CTL except for a perturbation that mimics the observed trend toward stronger stratospheric westerlies. The method used to apply the perturbation is similar to that used in previous studies (e.g., Norton 2003) and has been used to simulate the increase in the NAO in a climate model (Scaife et al. 2005). The method damps the stratospheric zonal wind at all latitudes with a body force whose drag coefficient increases linearly with altitude and is zero below 17 km. The e-folding drag time scale decays linearly with height to the top of the model and has a time scale of 2.5 days near the stratosopause. The drag also decreases linearly in time from its maximum in 1965 to zero in 1995. Over this period the perturbation results in an upward trend of 8.5 m s$^{-1}$ at 60°N near the core of the polar night jet at 50 hPa in the lower stratosphere. This is in contrast to the almost constant stratospheric winds in the control experiment. The trend agrees reasonably well with the observed trend of $\sim 6$ m s$^{-1}$ over the same period.

There are important mean surface climate differences between the CTL and NAO↑ ensembles. The lower troposphere in the NAO↑ ensemble responds by increasing the winter NAO (Fig. 1). Pressure and temperature fields in both NAO↑ members show a response that closely matches the observed NAO pattern with a surface NAO index that increases approximately linearly with time from 1965 to 1995. The resulting NAO increase between 1965 and 1995 in CTL, NAO↑ ensemble means, and the observations are 2 ± 3, 14 ± 2, and 10 hPa, respectively (the errors here are the uncertainty in the ensemble means due to the different trends in individual members: note that they are too small to explain the difference between CTL and NAO↑). We can therefore use the difference between the CTL and NAO↑ ensembles as a good representation of the observed NAO change over this multidecadal period. Note that the magnitude and pattern of the surface pressure response throughout the North Atlantic and Europe agrees well with the observed changes, as well as those of mean surface temperature and rainfall. For example, winter changes in north European temperature over the main warming region of the NAO dipole (50°–70°N, 10°W–50°E) are 0.53 K decade$^{-1}$ in observations, 0.15 K decade$^{-1}$ in the CTL ensemble, and 0.59 K decade$^{-1}$ in the NAO↑ ensemble over 1965–95 (Scaife et al. 2005). Also, the difference between the CTL and NAO↑ ensembles effectively eliminates the influence of background climate warming due to in-
creasing greenhouse gases, which is present in both ensembles.

3. Extreme temperature events

a. Changes in daily minimum temperature

Modeled changes in the frequency of very low daily minimum temperatures in winter are shown in Fig. 2. Changes in the CTL ensemble represent the direct effects of global warming, that is, of increasing greenhouse gases and the other well-known climate forcings listed above. Changes in the NAO↑ ensemble represent not only the effects of these climate forcings but also the increase in the NAO. Details of the method used to calculate the frequency of extremes are given in the figure captions. Over Europe, area mean changes are approximately five times as large in the NAO↑ ensemble as in CTL (Fig. 2). This indicates that, in our model, the effects of circulation changes on temperature minima far outweigh those due to climate forcings on this 30-yr time scale. These changes are qualitatively consistent with the changes in mean temperature, which are also several times larger in NAO↑ than CTL (Scaife et al. 2005).

Observed changes in daily minimum temperature can be estimated from land station data and such studies indicate an increase in daily minimum temperatures over the last few decades and a strong tendency for days with extreme minimum temperatures to occur less often toward the end of the twentieth century, particularly in northern Europe. Given the dominance of the NAO-induced changes in our model simulations it seems likely that the NAO could explain some of these observed changes in extremes. Here we use data from the study of Alexander et al. (2006) to calculate observed temperature extremes. Observed signals are consistent with that from the NAO↑ ensemble. In particular, the reduction in occurrence of very low minimum temperatures in northern Europe agrees well with that reproduced in NAO↑ (Fig. 2). The observed changes do not agree with the

![Fig. 2](image-url)
weak changes in the CTL ensemble. There are also signs of the NAO dipole in the observational data in Fig. 2 that correspond to a tendency toward a greater frequency of very low minimum temperatures over southern Europe and vice versa over northern Europe. A similar pattern to that seen in our modeled and pure observational data in Fig. 2 has also been reproduced for daily minimum temperatures from reanalysis data (Thompson and Wallace 2001). Circulation changes therefore appear to be crucial to changing the frequency of winter temperature extremes over Europe.

By pooling gridpoint data across large European areas, we can calculate changes in daily maximum and minimum temperatures across the whole probability distribution. Figure 3 shows this change for absolute daily minimum winter air temperatures between the 1990s and 1960s in the NAO$^{\uparrow}$ simulations for northern and southern Europe. The modeled change in temperature of the percentiles for northern Europe (Fig. 3) indicates changes of over 3$^\circ$ in the 10th percentile for NAO$^{\uparrow}$. This change is much greater than the change in CTL and occurs mainly in the northeast of Europe. The observed change in minimum temperature is also especially strong at the cold end of the distribution, where changes of 2$^\circ$–3$^\circ$C have occurred (Caesar et al. 2006).

Fractional changes across the distribution indicate a reduction in the frequency of lowest percentile events and an increase in the frequency of highest percentile events for northern Europe. We conclude that much of the observed change in winter minimum temperatures over northern Europe between the 1960s and 1990s is related to the NAO increase. Figures 2 and 3 also show how the sign of these changes is reversed in the Mediterranean and extratropical North African regions to give an unequal dipole structure with opposite changes in the two regions. This structure is similar to the change in winter mean temperature, which also shows opposite changes in these two regions and larger changes over northern Europe (Scaife et al. 2005).

Although these results do not radically affect our interpretation of global mean changes in minimum temperature, they do indicate that regional changes in observed winter temperature minima across Europe and
northern Africa over the 1960s to 1990s can be largely explained by the increase in the winter NAO.

b. Changes in the number of frosts

The modeled change in the number of frosts in Europe due to the NAO change is shown in Fig. 4. Here frosts are said to occur if the daily minimum surface air temperature is below 0°C. Changes in frost numbers in the CTL ensemble are much smaller than those in NAO↑. The model indicates a large decrease in the number of frosty nights over northern Europe between the 1960s and 1990s due to the increase in the NAO and a simultaneous but smaller increase over extratropical North Africa for the same reason. Note that many mountainous and high-latitude regions, such as Scandinavia, show only small changes in the number of frosts because mean minimum temperatures in these regions are often well below zero in the December–February season and the change from the NAO is too small to warm many individual nights above the freezing point. Particularly large changes are seen over the Baltic Sea.

This is presumably because temperature advection due to changes in the NAO explains much more of the variability in frost frequency in this region than over land where orography and variations in nighttime radiative cooling are important. The Baltic Sea feature is also weakly present in the control simulation, presumably due to the change in sea ice which is specified in both ensembles as part of the sea surface boundary condition. The observed signal is very similar to that from the NAO increase in the NAO↑ ensemble. In particular, the reduction in frosty nights in Europe agrees well with that reproduced in NAO↑ (Fig. 4). The observed changes do not agree with the weak changes in the CTL ensemble. There are also signs of the NAO dipole in the observational data in Fig. 3 that correspond to the largest reductions in frosts over northern Europe, particularly in the Baltic region. This corroborates observational studies that suggest a strong link between the NAO and the frequency and severity of Baltic sea ice (Koslowiski and Loewe 1994). Our simulations suggest that both the decrease in winter frosts over northern

![Fig. 4. Change in the number of winter frosts per year in (top left) NAO↑, (right) CTL ensembles, and (bottom left) observations. Units are frost days per year calculated over the same winters as in previous figures and using the DJF daily minimum temperature values as above. In order that the existing observational dataset could be used (Alexander et al. 2006), frost days are calculated over the whole year rather than just winter, but by definition, for the most part, only sample the winter season.](image-url)
Europe and the increase over North Africa over the 30 years considered here can be attributed to the change in temperature from the North Atlantic Oscillation. Note again, though, that the influence of the winter NAO on Europe does not greatly affect existing interpretations of globally averaged reductions in frost nights.

4. Extreme precipitation events

Differences between the NAO↑ and CTL simulations used here reproduce both the pattern and amplitude of mean precipitation changes to a reasonable extent, and maximum changes of around 1 mm day⁻¹ are found in both observations and the model (Scaife et al. 2005). We might therefore expect a similar dipolar response in extreme rainfall events over the European region. This would lead to increased occurrence of heavy precipitation over northern Europe and decreased occurrence over southern Europe and extratropical North Africa during high NAO periods. Note that in this section we use observed changes in the frequency of 90th percentile rainfall events. A gridded precipitation dataset was created using 130 climate stations assessed as being homogeneous (Wijngaard et al. 2003) and 2187 former Soviet Union stations adjusted to be as homogeneous as possible (Groisman and Rankova 2001). The result was a gridded dataset that could be compared directly with the model output.

a. Changes in heavy rainfall events over Europe

Despite doubts about the ability of such relatively coarse grained models to simulate the intensity of heavy precipitation events, using the frequency of exceeding percentile-based thresholds appears to give robust results when compared to higher-resolution models (Durman et al. 2001) and we use a similar measure here. Our modeled change in the frequency of 90th percentile events (Fig. 5) shows a dipolar response that is qualitatively similar to the mean rainfall change with increased heavy rainfall over northern and western Europe and decreased heavy rainfall over the Mediterranean and North Africa. The observed signal is most

![Fig. 5. Changes in the frequency of heavy precipitation events in (top left) NAO↑, (right) CTL simulations, and (bottom left) observations. The fractional change in occurrence of above 90th percentile 5-day events is plotted. The calculation of 90th percentiles and winters used are consistent with previous figures. Gridded observations of precipitation data were calculated for this study: 5-day mean winter (DJF) precipitation station data were gridded onto the model grid using an angular distance weighting method (Alexander et al. 2006). Provided there were at least two stations within the "correlation decay distance" then values were calculated for each grid box.](image-url)
consistent with that from the NAO increase in the NAO↑ ensemble. In particular, the increase in heavy rainfall in northern Europe shows reasonable agreement with that reproduced in NAO↑ (Fig. 5). The observed changes do not agree as well with the weak changes in the CTL ensemble. There are also some signs of the NAO dipole in the observational data in Fig. 5 that correspond to a tendency toward a reduced frequency of heavy rainfall over southern Europe and an increase over northern Europe. Despite these similarities, careful comparison of the change in observed and model rainfall extremes shows that the modeled signal does not extend as far into eastern Europe as the observed signal and it is shifted slightly southwest. This shift can be explained by a slight southwest bias in the modeled change in mean rainfall and temperature compared to the observed NAO signal. This results from a climatological bias in the climatological mean model rainband (Scaife et al. 2005).

Across the British Isles, the model indicates a range of increases in the frequency of heavy winter rainfall, but the area mean indicates an average increase in the frequency of events as found in observations for winter using U.K. rainfall over a similar period to that considered here (Osborn et al. 2000). Osborn et al. (their Fig. 11) found a roughly 50% increase in the frequency of similar heavy precipitation events for this period and tens of percent increases are seen in NAO↑ over the United Kingdom. This result suggests again that the increase in the NAO has very likely contributed to the increase in heavy U.K. winter rainfall in recent decades and, by implication, winter flooding events in the United Kingdom in the late 1980s and early 1990s.

b. Changes in the distribution of heavy precipitation values

As many European locations have a significant fraction of winter days with little or no rainfall, the 5-day precipitation distribution has a fundamentally different shape to the temperature distribution and peaks at zero. The shape of the distribution also affects the change in percentile values that arises from a shift in the NAO (Fig. 6) and leads to the largest changes occurring at the heavy rainfall end of the distribution. For
northern Europe, the fractional change shows a very clear reduction in low rainfall events and an increase in the heavy events with the opposite response for the Mediterranean and North African regions. Note that, as with temperature, changes are typically several tens of percent and are largest in the wings of the distribution as might be expected.

5. Longer-term links between extremes and the NAO

We can also use observational data to estimate the relationship between the NAO and extreme precipitation and temperature over the entire twentieth century to obtain a longer time perspective on the results presented above. In this case we use indices for temperature and precipitation extremes for the period 1901–2000 (Moberg et al. 2006). We use indices that count the number of days below the 10th and above the 90th percentile for minimum temperature and rainfall, respectively. The correlations between the observed station-based NAO index between Iceland and the Azores and these European station-based extremes are shown in Fig. 7. Unfortunately, the spatial coverage of century-long daily records is rather poor; hence the information is missing in large parts of Europe. However, there is a clear anticorrelation between the frequency of low minimum temperatures and the NAO over large areas in central and northern Europe. An anticorrelation is also clearly seen between the NAO and the number of frost days, although not in the northernmost parts where daily minimum temperatures in winter are mostly below zero, regardless of the NAO phase. These observational results for century-long records agree qualitatively well with the model results (Figs. 2 and 4), suggesting that the simulated changes in winter temperature extremes behave realistically in response to the NAO and that the changes associated with the NAO are of the same sign (though not necessarily linear) over the whole twentieth century.

Note also that the Iberian Peninsula and northern Europe, especially southern Scandinavia, show opposite changes in extreme precipitation associated with the NAO. The changes are of the same sign as the
modeled changes due to increasing NAO (Fig. 5) and demonstrate that the relationship between the multidecadal changes in NAO and heavy precipitation events simulated over the 1960s to 1990s is similar to the observed relationship on a year-by-year basis between the NAO and extreme precipitation over the whole twentieth century. To study if the observed NAO trend after 1965 has a strong influence on the correlations over 1901–2000, we repeated all calculations using observational data only for 1901–64. The results (not shown) were nearly the same, indicating that correlations between the NAO and the temperature and precipitation extremes are little affected by the choice of period.

6. Conclusions

Using a novel set of experiments with a general circulation model, we have examined the effect of a realistic multidecadal increase in the NAO on surface climate extremes over Europe in winter. Using ensembles whose difference removes the modeled effects of anthropogenic forcing, we were able to assess the role of multidecadal changes in the NAO in changing extreme weather events. The large changes in the modeled extreme temperatures and precipitation found here show that changes in the NAO are likely to be responsible for much of the observed change in the frequency of above 90th percentile winter precipitation, numbers of frost days, and below 10th percentile minimum temperature extremes between the 1960s and 1990s over Europe and extratropical North Africa. Accordingly, accounting for regional atmospheric circulation change is a crucial step when interpreting observed changes in regional extremes. This is especially important in individual seasons.

The response of extreme surface temperature and precipitation to changes in the NAO has a qualitatively similar pattern to changes in mean temperature and precipitation from the NAO. Thus the observed increase in the NAO gave rise to an increase in heavy precipitation events and a decrease in frosts in northern Europe with smaller and opposite changes in southern Europe and North Africa. This is to be expected if the whole probability density function of daily surface temperature and rainfall values is displaced in the same direction as the mean climate by the NAO. Changes in frequency of these events exceeded 50% over parts of northern Europe. Consequently, for northern Europe, heavy winter rainfall events are much more likely and frosts and extreme low temperature events are much less likely during high NAO conditions. More extreme events than have been considered here lie in the outer wings of the temperature and precipitation distributions and may well show a different response to the NAO. However, we are currently limited by having only two perturbed model simulations, and more simulations would be needed to estimate the NAO related changes in the more extreme weather events without resorting to extreme value theory and associated assumptions to fit to the tail distribution. It may of course still be the case that more extreme events are affected differently or only weakly by the NAO.

Although comparison with observations is restricted by data missing from the observational record, the use of quality-controlled datasets of European land station data do show patterns of observed changes in extreme rainfall and temperature that are consistent with the modeled signals. Recent developments in extending the length of these records back in time mean that we are also able to demonstrate that similar changes in extremes associated with the NAO are likely to have occurred over the whole of the twentieth century.

Our results also have implications for the importance of stratosphere–troposphere coupling for surface climate. As the perturbation in the model is only applied to winds above the tropopause, the results presented here represent a clear test of the influence of stratospheric circulation on surface climate. They show that a realistic change in stratospheric circulation has a downward link from the stratosphere to surface climate that is strong enough to have had large impacts on the frequency of extreme precipitation and very low temperatures over Europe.

We also note that projected climate change signals in winter extremes for the twenty-first century (e.g., Frei et al. 2006 for precipitation) are of the same order of magnitude as the changes found here from the NAO between the 1960s and 1990s. This emphasizes the need to accurately model low-frequency NAO variability in the numerical models used for regional climate prediction. An important part of the evaluation of regional climate models is to establish whether they are able to reproduce past changes in mean climate and extreme events. This may not be the case given that the global models from which the boundary conditions for regional models are derived do not reproduce the observed NAO trend.

It is important that we determine the source of observed NAO changes. The results presented here do not in themselves have any bearing on the extent to which changes in the NAO are natural or anthropogenic, although the recent downturn in the winter NAO from strong positive values in the early to mid-1990s to near-average values a decade later suggests that a significant part of the increase from 1965 to 1995 may be
due to natural climate variability. If so, attribution of changes in regional, seasonal climate extremes needs to be carried out much more thoroughly than hitherto, taking into account the details of atmospheric circulation changes in given seasons. Furthermore, if the NAO increase between the 1960s and 1990s is natural in origin, future decades could easily see a reversal of regional trends in European winter climate because NAO effects can dominate the effects of global warming on Europe in winter, even on multimodel timescales. Indeed, this may already be underway given the recent decrease of the winter NAO. Alternatively, if the overall NAO increase since the 1960s is anthropogenic in origin, then the rapid changes in Eurasian winter climate and its extremes over the last few decades could continue in the future. In this case, current climate prediction models severely underestimate future changes in European winter precipitation and temperature change because they have great difficulty in producing the observed rate of increase in the NAO.

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