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Climate of the Past Discussions

Impact of precipitation intermittency on NAO-temperature signals in proxy records

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Abstract

In mid and high latitudes, the stable isotope ratio in precipitation is driven by changes in temperature, which controls atmospheric distillation. This relationship forms the basis for many continental paleoclimatic reconstructions using direct (e.g. ice cores) or indirect (e.g. tree ring cellulose, speleothem calcite) archives of past precipitation. However, the archiving process is inherently biased by precipitation intermittency. Here, we use two sets of atmospheric reanalyses (NCEP and ERA-interim) to quantify this precipitation intermittency bias, by comparing seasonal (winter and summer) temperatures estimated with and without precipitation weighting. We show that this bias can reach locally 6 to 10 °C and has large inter-annual variability. We then assess the impact of precipitation intermittency on the strength and stability of temporal correlations between seasonal temperatures and the North Atlantic Oscillation. Precipitation weighting reduces the correlation between winter NAO and temperature in some areas (e.g. Québec, South-East USA, East Greenland, East Siberia, Mediterranean sector) but does not alter the main patterns of correlation. The importance of the precipitation intermittency bias with respect to other processes affecting precipitation isotopic composition is further analysed using outputs of an atmospheric general circulation model enabled with stable isotopes and nudged to reanalyses (LMDZiso). In winter, LMDZiso shows similar correlation values between the NAO and both the precipitation weighted temperature and precipitation δ^{18}O, thus suggesting limited impacts of moisture origin. Correlations of comparable magnitude are obtained for the available observational evidence (GNIP and Greenland ice core data). Our findings support the use of archives of past precipitation δ^{18}O for past NAO reconstructions.

1 Introduction

The North Atlantic Oscillation (NAO) is known to have large impacts on mid to high latitudes climate, affecting daily to seasonal patterns of precipitation or temperature

(Hurrell et al., 2003). This is related with the associated shift in the storm track path leading to a quadrupole of surface temperature anomalies, consisting of a northern seesaw between Northern Europe and North America, and a southern seesaw with nodes centered in the Mediterranean area and the Southeastern United States (Pinto and Raible, 2012). While winter NAO has long been reported to affect Northern Hemisphere climate, climate impacts of summer NAO have also been depicted in specific regions such as Europe and Greenland (Folland et al., 2009). Instrumental records document large inter-annual and decadal variations in the NAO index (Jones et al., 1997), usually defined as the pressure difference between the Azores high and the Icelandic low (Hurrell, 1995). Obtaining longer records is crucial to assess the natural spectrum of NAO variability and its response to external forcings (such as changes in solar and volcanic activity, or anthropogenic perturbations) and to evaluate the ability of climate models to capture the NAO variability (IPCC, 2007).

Over the last 15 yr, several studies have combined different sets of documentary data and paleoclimate information in order to propose reconstructions of pre-instrumental NAO variability (Appenzeller et al., 1998; Luterbacher et al., 1999; Glueck and Stockton, 2001; Cook et al., 2002; Trouet et al., 2009, 2012; Pinto and Raible, 2012). These studies have used information related to precipitation and/or temperature in seasonal to annually resolved archives, such as dendrochronological series or Greenland ice core accumulation. Some stalagmites possess annual laminae (visible and/or fluorescent) that can be used to construct precise chronologies of the last decades/centuries (Baker et al., 1993; Genty, 1993). Thicknesses (from ~ 0.01 to $1\,\mathrm{mm\,yr^{-1}}$) that are controlled by rainfall and temperature have been used as proxy for high resolution climate reconstructions (Munoz et al., 2009; Tan et al., 2006). Among the rare examples that used stalagmite layer thickness, the one from the Uamh Cave system (North Scotland) showed significant correlation between fluorescent laminae thickness and winter instrumental NAO index (Baldini et al., 2006; Proctor et al., 2000; Baker et al., 2002) and, combined with tree ring data from Morocco, was used to derive fluctuations of NAO during the last millennium (Trouet et al., 2009).

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All NAO reconstructions show large multi-decadal variability. However, prior to the 19th century, they show a lack of consistency (Pinto and Raible, 2012). This may be due to the influence of other modes of variability which can affect North Atlantic climate such as the Atlantic Multidecadal Oscillation (Delworth and Mann, 2000), to variable teleconnections, and to the different source data sets and methodologies employed (Lehner et al., 2012). One way to reduce uncertainties is to expand the source data (type of proxies and spatial coverage) used for such reconstructions.

Recently, a significant imprint of NAO has been identified in precipitation isotopic composition ($\delta^{18}O$) in Greenland ice core records (Barlow et al., 1993; Vinther et al., 2003a, 2010), both in monthly precipitation $\delta^{18}O$ data from Central Europe (Baldini et al., 2008) and in daily $\delta^{18}O$ data from Norway (Theakstone, 2011). Simulations conducted with general or regional atmospheric models equipped with the explicit representation of water stable isotopes have also depicted a clear signature of NAO in precipitation $\delta^{18}O$ over Europe (Field, 2010; Langebroek et al., 2011), Greenland (Sjolte et al., 2011) and Svalbard (Divine et al., 2011). This imprint of the NAO is also observed in a growing number of tree ring cellulose isotope studies, from Europe (Treydte et al., 2007; Saurer et al., 2012) to Siberia (Sidorova et al., 2010). A recent study also describes a relationship between a high resolution record of Red Sea coral $\delta^{18}O$ and the NAO (Felis and Rimbu, 2010). All these findings motivate the combined use of direct (ice cores) and indirect (tree ring cellulose, speleothem calcite or fluid inclusions, coral aragonite ...) archives of precipitation isotopic composition to reconstruct past NAO variability.

While precipitation $\delta^{18}O$ is related to temperature, it is an integrated tracer of the water cycle (Jouzel, 2003). In mid and high latitudes, its variability is driven by atmospheric distillation related to condensation temperature. Continental recycling (van der Ent et al., 2010) can also affect the precipitation isotopic composition. Likewise, changes in air mass trajectories can strongly impact precipitation $\delta^{18}O$ (Sodemann et al., 2008). Such effects can be evaluated using second order tracers such as deuterium excess (Kurita, 2011) and ^{17}O-excess (Landais et al., 2012), or quantified using

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isotopic models (Sturm et al., 2010). Here, we aim to quantify another source of uncertainty, related to precipitation intermittency. Indeed, precipitation $\delta^{18}O$ only records climate information when precipitation occurs. Because temperature and precipitation can co-vary at daily, seasonal or inter-annual time scales, this precipitation intermittency bias may distort the original NAO-temperature signal in climate archives. This source of uncertainty was recently investigated for Greenland, using atmospheric reanalyses (Persson et al., 2011). They showed that precipitation intermittency (fraction of summer/winter snowfall and its inter-annual variability) explained many aspects of NAO-isotope relationships in different Greenland ice core sites. A number of recent studies have highlighted the importance of precipitation and temperature covariance at synoptic scales (Persson et al., 2011; Sime et al., 2008).

In order to assess the impact of precipitation intermittency bias on temperature-NAO relationships, we need to have access to daily, coherent and global precipitation and surface air temperature datasets. For this purpose, we used different sources of information: (i) two atmospheric reanalyses datasets, and (ii) a simulation conducted with an atmospheric model equipped with water stable isotopes. The use of atmospheric reanalyses assures a consistency between temperature and precipitation, and large-scale weather patterns; however, the precipitation amounts are an output of the atmospheric models, albeit constrained by assimilation of several products. This motivates a comparison between two different reanalyses datasets (see Supplement Sect. A and Fig. S1). The use of an isotope-enabled model (LMDZiso), nudged to 3-dimensional wind patterns reanalyses from ERA40, then allows to quantify the relationships between precipitation-weighted temperature and precipitation isotopic composition, and to compare directly the simulated NAO-isotope relationships with observations.

After a description of the databases, simulations and methods (Sect. 2), the bias induced by precipitation intermittency are quantified at different time scales, for winter and summer (Sect. 3). This focus on DJFM (December–March) and JJAS (June–September) is motivated by the reported impacts of NAO on seasonal climate variables, and by the ability of some proxy records to resolve seasonal variations (e.g. ice cores).

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We then analyze the impact of precipitation intermittency on correlations between temperature and NAO and compare our results with observations. Finally, key conclusions and perspectives are presented in Sect. 4. Additional information is available in the Supplement.

2 Databases, simulations and methods

2.1 Atmospheric reanalyses datasets

Two reanalysis datasets are used in this study: (i) NCEP data (Kalnay et al., 1996), modified for consistency with CRU TS.3.1 temperature data (Maignan et al., 2011); and (ii) ERA-interim data (Dee et al., 2011).

The first dataset, CRU-NCEP, is based on a combination of the NCEP reanalysis (Kalnay et al., 1996), covering the period 1948 to 2010 with a 6-hourly $2.5^{\circ} \times 2.5^{\circ}$ spatial resolution, and the CRU TS 3.1 dataset from the University of East Anglia (Mitchell and Jones, 2005), which provides monthly climate variables (temperature, precipitation, cloudiness and relative humidity) from 1901 to 2010 with half a degree spatial resolution for both latitudes and longitudes on land areas. The daily NCEP fields are interpolated to the finer $0.5^{\circ} \times 0.5^{\circ}$ spatial grid, and their climatology is corrected for biases with the monthly fields from CRU TS 3.1 (for further details see http://dods.extra.cea.fr/data/p529viov/cruncep/). The objective of CRU-NCEP is double: firsts, to correct for biases in the NCEP dataset because fields such as precipitation or solar radiation are produced by the atmospheric model and can show significant biases; second, to improve the resolution of the dataset and better account for the effet of orography taking advantage of the higher CRU dataset resolution. Thus, this final mixed CRU-NCEP dataset provides a fine resolution both in time (6-hourly) and space (0.5°).

In order to assess the robustness of our results, the analysis will be also extended to the ERA-Interim reanalysis (Dee et al., 2011), for the period from 1990 to 2010.

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Comparisons between different reanalyses products (Supplement Sect. A) and observations have stressed the good performance of the ERA-interim spatial and temporal distribution of precipitation (Bosilovich et al., 2008; Ma et al., 2009; Chen et al., 2011). The performance of the reanalyses depends on the assimilation of humidity observations (Andersson et al., 2007; Janowiak et al., 2010) and the physical parameterization of the models especially regarding atmospheric convection (Jung et al., 2010). Caveats in the representation of soil moisture in the NCEP reanalyses have been related to an overestimation of summer land precipitation, due to excessive convective precipitation (Serreze and Hurst, 2000). At the monthly scale, these studies systematically report better performance for winter than for summer precipitation.

The robustness of the results obtained using products of atmospheric reanalyses has been assessed by comparing the difference between precipitation weighted temperature and seasonal mean temperature (see Sect. 2.4) from these datasets with the same calculations performed using the ECA&D (European Climate Assessment and Dataset, http://eca.knmi.nl) daily station data (see Supplement Sect. B, Figs. S1 and S2).

2.2 LMDZiso simulation

In order to compare the correlations between NAO and precipitation weighted temperature with those related to changes in precipitation isotopic composition, a simulation for the period 1979–2008 with the isotopic model LDMZiso (Risi et al., 2010) was employed (Fig. S1). The simulation was nudged to the three dimensional horizontal winds of ERA-40 (Uppala et al., 2005) during the period 1979–2002, and from ECMWF operational analyses thereafter. No noticeable discontinuity due to the change in the nudging data was observed (Risi et al., 2010). The nudging with the reanalyses allows for a better representation of the synoptic and interannual variability of both δ^{18}O and temperature at mid and high latitudes. The LMDZiso simulation reproduces reasonably well the spatial and seasonal variations of δ^{18}O (Risi et al., 2010). However, its simulated precipitation δ^{18}O is overestimated over Central Greenland, associated to a warm temperature bias, also observed in other isotopic general circulation models (Steen-Larsen et al., 2011). Note also that the model resolution is relatively coarse (i.e. $2.5^\circ \times 3.75^\circ$ and 19 vertical levels) as compared to the two reanalysis datasets, which are thereby expected to better represent small scale climatic features, especially those related to complex orography. As for the two reanalyses, both the temperature biases due to precipitation intermittency and the spatial features of the associated NAO fingerprint will be also analysed in the LMDZiso outputs.

2.3 Precipitation isotopic composition datasets

Monthly precipitation $\delta^{18}O$ data are available from the Global Network of Isotopes in Precipitation (GNIP) stations (Rozanski et al., 1993; Baldini et al., 2008) and can be used to assess the robustness of the LMDZiso results. A key limitation arises from the brevity and discontinuities of most GNIP station records, with multi-decadal information being mostly provided by a few Central European stations. Our analysis will therefore focus on stations located north of 20° N, where the main centres of action of the NAO are located. A subset of GNIP stations where summer or winter measurements are available for at least 10 different years was extracted to evaluate the link with the NAO variability. Considering a 10 yr duration allows to expand the spatial coverage, while shorted records would preclude any statistical analysis. The data have been combined and re-gridded on a $1^\circ \times 1^\circ$ grid, resolving the high station density over Central Europe. When more than one station is available in a grid point, only the station with the longest record is employed. Under these conditions, 60 stations are finally selected (see Supplement Sect. C, Tables S1,S2).

Greenland ice core data are also used as an independent source of information, but most records are available only prior to e.g. 1970, limiting the comparison with the most recent and reliable part of the reanalyses and with LMDZiso datasets (Sodemann et al., 2008; Sjolte et al., 2011). In this analysis, we used a compilation of winter and summer ice core data (Vinther et al., 2010) complemented by the new data from the

NEEM ice core (Steen-Larsen et al., 2011). Specific details on the ice core data are given in Table S3.

2.4 Calculation of precipitation-weighted temperature

The biases associated with precipitation intermittency are expected to differ from one season to the next, for instance due to different radiative impacts of cloudiness. For each given year, we have therefore calculated seasonal (December–March, DJFM and June–September, JJAS) temperature and precipitation-weighted temperature averages.

The precipitation weighting was tested at three temporal resolutions: the 6 h step dataset, and both the daily and monthly averages of temperature and precipitation. It is calculated with the simplest approach:

$$T_{p,t}(\text{year,season,lat,lon}) = \frac{\sum_t T_{2\text{m}}(t,\text{lat,lon}) \times p(t,\text{lat,lon})}{\sum_t p(t,\text{lat,lon})}$$

where t represents the different temporal resolution (6 hourly, daily or monthly); lat and lon are respectively the latitude and longitude; $T_{2\text{m}}$ is the 2 m air temperature and p the precipitation amount (which is the sum of rainfall and snowfall amounts). Differences between the 6 hourly and daily precipitation weighted temperature may arise from differences in the diurnal cycle of precipitation and temperature; differences between the daily and monthly precipitation weighted temperature are expected to be caused by the covariance of temperature and precipitation at synoptic scales.

This calculation is performed for all CRU-NCEP, ERA-interim and LMDZiso datasets, and the respective results are compared and discussed in Sect. 3. Precipitation weighted temperature is not computed in grid points where annual precipitation is less than 5 mm. This threshold allows avoiding artefacts associated with very rare precipitation events in areas such as the Sahara; it does not affect NAO-influenced areas.

Precipitation weighted temperatures are found to be virtually identical calculated both at 6 hourly and daily timescales (not shown), thus suggesting that no important bias

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is introduced by the diurnal cycle. In contrast, large differences do emerge in the comparison of precipitation intermittency biases estimated from daily and monthly values (not shown). No remarkable differences are observed between mean temperature and monthly precipitation weighted temperature (not shown), probably because monthly precipitation amounts are rather regularly distributed, masking the co-variance between precipitation and temperature that takes place at the daily scale. In the next sections, we therefore focus our discussion on the precipitation intermittency bias calculated from daily values.

2.5 North Atlantic Oscillation index

Our reference NAO time series is the instrumental index defined by (Jones et al., 1997) and expanded by (Vinther et al., 2003b), calculated as the normalised sea level pressure difference between Gibraltar/Cadiz and Reykjavik. Following this same definition, NAO indices were computed for CRU-NCEP, ERA-interim and LMDZ-iso. All these time series correlate well with the instrumental data, as indicated by correlation coefficients above 0.8 and around 0.9 for CRU-NCEP (Fig. 1). Small differences may be caused by the resolution of atmospheric models and the choice of individual atmospheric model grid points. This coherency justifies the use of the reanalyses and LMDZiso for investigating NAO-climate relationships in a realistic framework. However, the small differences require that, for a given dataset, we use the NAO index most consistent with the dataset (e.g. LMDZiso surface climate parameters will be correlated with LMDZiso NAO index; GNIP data with the instrumental NAO index . . .).

The choice of an index based on pressure differences is questionable due to shifting positions of the NAO pressure centers of actions, both at the seasonal (Folland et al., 2009) and decadal scales (Pinto and Raible, 2012). An alternative definition of the NAO index is based on the first principal component of the sea level pressure field. This approach has been tested using the LMDZiso simulations (Supplement Sect. D, Figs. S3, S4 and S5) which provide a consistent framework for analysing the correlations of different NAO indices with climate and isotopic variables.

2.6 Statistical analyses

We focus on inter-annual relationships at the seasonal scale (December–March, DJFM and June–September, JJAS). For each grid point, the inter-annual relationships between DJFM or JJAS temperature, precipitation, precipitation-weighted temperature (using daily precipitation and temperature data) with the NAO index was quantified using the Pearson correlation coefficient, using the CDAT3.3 toolbox http://www2-pcmdi.llnl.gov/cdat). The significance of linear correlations was assessed with a Student's t-test, which takes into account the series autocorrelation to calculate the actual degrees of freedom and the linear correlation coefficient. Note that correlations were investigated on different time periods, depending on the data set (e.g. 1979–2008 with LMDZiso, 1990–2010 for ERA-interim). We have used the NCEP dataset to assess the stability of correlations over different time periods (1950–1970, 1970–1990 and 1990–2010; see Supplement Sect. E, Fig. S6). This allowed a systematic comparison of the results obtained from NCEP and ERA-interim over the time period 1990–2010. Correlations with the GNIP observation were calculated on different time periods when data were available for at least 10 yr.

3 Results and discussion

In this section, we describe the calculated mean precipitation intermittency biases (defined as $T_p - T$) and also its influence on the NAO-temperature relationship, for each season (JJAS and DJFM).

3.1 JJAS temperature biases

In summer, precipitation intermittency produces limited biases (in terms of spatial extent and in terms of magnitude) at mid and high latitudes (Fig. 2a). While NCEP suggests a small positive bias for high latitude lands, this result is not supported by ERA-interim (positive bias only for Greenland) or LMDZiso (no bias). By contrast, all datasets depict a cold bias on Western North America and around the Mediterranean Sea, albeit with different magnitudes (smaller for NCEP and larger for LMDZiso). Such a cold bias is simulated in areas characterized by low mean summer precipitation (Fig. S1) and may be linked with the albedo effect of clouds, or the impact of evaporation on land temperature (Hirschi et al., 2011).

The inter-annual standard deviation of the summer bias (Fig. 2b) remains smaller than for winter, with the exception of LMDZiso which produces both a large and variable bias around the Mediterranean area and Central Russia.

3.2 Impact of precipitation intermittency on JJAS NAO-temperature relationships

The correlation between respective NAO indices and reanalyses and LMDZiso variations of JJAS precipitation, temperature, precipitation weighted temperature and $\delta^{18}O$ is displayed in Fig. S8.

Correlations are generally patchy and weak for precipitation, with significant anti-correlations depicted by reanalyses for Western Europe and Northern Siberia; LMDZiso does not capture the strength of these relationships. Larger areas are detected for NAO-temperature relationships, mostly in N America, Greenland (especially South Greenland) and NE Siberia. LMDZiso displays a particularly strong correlation in Southern Greenland. For precipitation-weighted temperature, correlations are weaker and more patchy, with significant anti-correlation represented in some of the datasets for NW America, some areas of Greenland, and, in reanlayses, NE Siberia.

In LMDZiso, large anti-correlations are simulated between JJAS NAO and $\delta^{18}O$ for NW America, NE America, Greenland, and E Siberia. The fact that these correlations are stronger than for weighted temperature suggests that summer NAO also affects isotopic composition due to air mass origins; this has not yet been explored from specific trajectory calculations. The results obtained with LMDZiso must be taken with caution, due to the fact that the simulated correlations are stronger than observed (Table S3, Fig. S9) and sensitive to the position of NAO centers of action. Besides, Fig. S5 shows

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that weaker (stronger) correlations are simulated for Greenland (Scandinavia, Siberia) when considering an NAO index defined from the first principal component of LMDZiso pressure field.

While small, the precipitation intermittency bias complicates the signal detection. From these investigations, we cannot rule out a possibility to estimate past changes in JJAS NAO based on isotopic proxies, from selected areas (NW, NE America, E Siberia, Greenland) where T, T_p and δ^{18}O are related to NAO.

3.3 DJFM bias

Over most Northern Hemisphere land areas, wet days are warmer than the average winter days, and this bias reaches up to 6 °C in Arctic areas such as Greenland, Alaska or Siberia (Fig. 2a). This warm bias is consistent with the radiative impacts of water vapour and cloudiness on surface temperature. The spatial patterns of the DJFM bias are similar in CRU-NCEP, ERA-interim and LMDZ-iso, with a few regional differences in Central Greenland and Eastern Siberia, possibly due to differences in simulated precipitation amounts (Fig. S1). We note a specific band along the Rocky Mountains where precipitation weighting induces a cold bias, which could be related to an albedo effect of clouds and/or snow. The magnitude of the precipitation intermittency bias is much larger in ERA-interim at high latitudes (Fig. 2a). Surprisingly, LMDZiso produces a precipitation intermittency bias that is more similar to the magnitude derived from NCEP data, while its mean climate is closer to that of ERA-interim. Altogether, robust features of the spatial pattern of precipitation intermittency bias emerge from the three datasets investigated here, with a large bias found in several areas where climate is affected by the NAO (with the exception of Europe).

The inter-annual standard deviation of the precipitation intermittency bias is represented on Fig. 2b, reaching 2 °C in several areas. NCEP and ERA-interim produce large inter-annual variability of the winter bias in Northern North America, Greenland, and Siberia, while LMDZiso produces the largest variability in NE North America and has a significantly lower standard deviation in Siberia. The magnitude of high latitude

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precipitation intermittency bias and its variability has a potential to distort the fingerprint of NAO in precipitation archives, which is investigated in Sect. 3.4.

3.4 Impact of precipitation intermittency on DJFM NAO-temperature relationships

Figure 3 shows the comparison of the correlation coefficients between DJFM NAO and precipitation, temperature, and precipitation-weighted temperature (T_p) as calculated for CRU-NCEP, and ERA-interim from 1990–2010 and for LMDZiso from 1979 to 2008.

Winter precipitation is positively correlated with NAO in Northern Europe, and areas of Russia in all datasets; LMDZiso seems to produce larger correlations for Northern Europe than reanalyses. Negative correlations are also systematically produced for areas of NW North America (from Quebec to Ellesmere Island), Western or South Greenland, Eastern Siberia and the Mediterranean area, where the strongest negative correlations are depicted in both reanalysis datasets.

The spatial correlation patterns of NAO with temperature are different from the pattern previously described for precipitation. The results appear more coherent at a continental scale, with positive correlations for SE USA, Western to Northern Europe, Russia and Central Siberia, and negative correlations for NW America, Greenland, the Southern Mediterranean land areas and Eastern Siberia. While larger anti-correlations are observed in ERA-interim, especially over Greenland where the strongest anti-correlation is depicted ($R < -0.75$ for ERA), the strongest positive correlations are consistently obtained in Northwestern Europe ($R > 0.75$) for the three datasets. Overall, LMDZiso shows correlations of smaller amplitude (both positive and negative) than the reanalyses, only capturing the centers of action over Greenland and Eurasia.

The main patterns of NAO-temperature correlation are preserved when considering precipitation-weighed temperature, with the exception of East Greenland and East Mediterranean/Middle East areas. For LMDZiso, similar patterns of correlation are obtained for T and T_p, with even a stronger signal in a large Central Europe area ($R > 0.75$) and a shift in the strongest negative correlation coefficient from Southern to

Eastern Greenland. In the reanalyses, correlations seem to be sensitive to local conditions, such as orography. As a result, in CRU-NCEP correlations appear generally reduced in Siberia/Russia, North America and Greenland. Likewise, for ERA-interim, the most remarkable differences are a general reduction of correlation in Greenland and South-Western Russia, but enhanced correlation coefficients to the Southeast of Finland and over the Central Siberian Plateau. The difference between the different datasets seems closely related to differences in the winter precipitation climatologies (see Supplement Sect. A, Fig. S1).

In this section, analyses were focused on the most recent period (1990–2010), where there is an overlap between our different datasets, and which is best described thanks to the observations that are assimilated in the analyses systems. We are aware that this time interval is characterised by high DJFM NAO levels in the early 1990s. The stability of the NAO-climate relationships has been assessed using different time periods in the NCEP dataset (1950–1970, 1970–1990; Supplement Sect. E, Figs. S6 and S7). The spatial pattern of NAO-T_p correlation appears remarkably stable through time. However, the strength of the correlation is rather variable in a few high latitude areas (North and Eastern Greenland, Russia and Siberia) where winters are particularly dry in average (Supplement Sect. F, Fig. S8).

3.5 Comparison between DJFM NAO-T_p correlation and NAO-δ^{18} O correlation

As stated in the introduction, water stable isotopes are integrated tracers of the water cycle, and the isotopic composition of precipitation itself is related not only to local condensation temperature, but also to all phase changes occurring from evaporation to condensation. Second order parameters (deuterium excess or ^{17}O excess) may provide evidence for changes in evaporation conditions, but the observational records (precipitation monitoring) are too short to be used to analyse correlations between these parameters and NAO, and therefore the relationships between NAO and changes in moisture origin. For Greenland, where ice core records of past precipitation are available, numerous seasonally resolved δ^{18}O records are available (Vinther et al., 2010)

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(Supplement Sect. C), while only a few datasets of deuterium excess (Sodemann et al., 2008; Steen-Larsen et al., 2011) or ^{17}O excess (Landais et al., 2012), which require higher accuracy, are available. The most robust results have been produced for combined δ^{18}O from several shallow ice cores. So far, based on observations, it has not been possible to quantify the impact of NAO on moisture sources and therefore the possible biases introduced in local climate-δ^{18}O relationships. One moisture back-trajectory study has been conducted for Greenland precipitation, suggesting a 5 °C sea-surface-temperature difference between moisture sources for positive versus negative NAO winter conditions (Sodemann et al., 2008). We are therefore fully aware that relationships between NAO and δ^{18}O can involve processes other than condensation temperature. To quantify these other processes, we compare the pattern and strength of the correlation between NAO and T_p, with the pattern and strength of the correlation between NAO and precipitation δ^{18}O from LMDZiso (Figs. 3b and 4).

First, the LMDZiso model results are described (Fig. 3). The model results clearly show the same pattern of winter NAO correlation with T_p and δ^{18}O, with negative correlations over parts of Greenland and NW America (Québec area), and positive correlations in Northern Eurasia. The width of the areas with significant correlation (> 0.35) and the strength of the correlation appears significantly smaller in δ^{18}O than in T_p, suggesting that other processes (e.g. changes in moisture sources, changes between condensation and surface temperature ...) act to reduce the imprint of NAO in δ^{18}O. This supports the idea that the investigation of the NAO-T_p relationships provide a "best case" background for searching for NAO-δ^{18}O relationships.

The LMDZiso NAO-δ^{18}O and the NCEP and ERA-interim NAO-T_p correlations are now compared to observations of NAO-δ^{18}O correlations, using instrumental records of precipitation isotopic composition and Greenland ice cores. This comparison is strongly limited by the lack of consistency of the time frame where these observations are available (Supplement Sect. C, Tables S1 and S2).

The GNIP data (Fig. 4) confirm the simulated structure of winter NAO-δ^{18}O relationships (Baldini et al., 2008), with positive correlations encountered in Western

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Europe, North America and Russia, and small, negative correlations over Iceland and Svalbard. No data is available to test the simulated negative NAO-T_p correlation depicted by NCEP and ERA-interim for the extreme east of Siberia and for the Québec/Baffin Bay/Ellesmere Island area. The strongest correlation coefficients are obtained for Western and Central Europe, with R values up to 0.75, and therefore comparable with the strength of the correlations produced in this area for T_p and δ^{18}O by NCEP, ERA-interim and LMDZiso. These large correlations were also reported in several observational (Baldini et al., 2008) and modelling (Field, 2010; Langebroek et al., 2011; Sjolte et al., 2011) studies, which, however, often explored the anomalies between specific positive and negative NAO situations, rather than linear correlations, preventing us from making quantitative comparisons.

Winter signals extracted from Southern and Central Greenland ice cores (Vinther et al., 2010) systematically depict a negative correlation coefficient with NAO, varying between −0.13 and −0.42 for the time intervals where raw data are available (Fig. 4, Table S3). The comparison between the ice core signals and our results is also delicate because of the different time periods available, as most of the ice core records are only available prior to the 1970s. In NW Greenland, the new NEEM ice core shows no correlation between winter δ^{18}O (one shallow ice core) and NAO for the last decades (Steen-Larsen et al., 2011). This ice core lies on an area of sharp gradients in atmospheric datasets (Figs. 3, 4) where all datasets (CRU-NCEP, ERA-interim and LMDZiso) show large variations from significant anti-correlation with local T_p and δ^{18}O to zero correlation. The larger winter NAO correlation depicted by reanalyses and LMDZiso compared to the NEEM data could be explained by two possibilities: either all atmospheric models poorly capture the NW Greenland weather patterns and precipitation intermittency (in the Greenland area where annual precipitation is dominated by summer snowfall), or the NAO signal present in local temperature is masked in δ^{18}O due to changes in moisture source, and moreover LMDZiso fails to capture these effects. The latter option cannot be excluded, as in situ, summer surface water vapour isotopic composition shows large subseasonal deuterium excess variations associated with different

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air mass origins which further cannot be captured by LMDZiso (Steen-Larsen et al., 2012).

4 Conclusions

The precipitation intermittency biases and their implications for correlations with the North Atlantic Oscillation have been explored, using two atmospheric analyses and one atmospheric model equipped with water stable isotopes.

Summer NAO appears to leave a discernable imprint on temperature and isotopes in very specific areas (NE and NW America, Greenland, E Siberia), but precipitation intermittency reduces the spatial extent and the magnitude of the correlation with temperature. From existing isotopic datasets, LMDZiso seems to overestimate the strength of the summer NAO imprint on Greenland water stable isotopes, possibly due to the isotopic composition changes associated with different air mass origins. We therefore remain extremely cautious about reconstructions of summer NAO using precipitation-sensitive archives. Further investigations should be performed using daily meteorological and water vapour isotopic data combined with back trajectories, as well as higher resolution simulations, because of the uncertainties on daily precipitation amounts in atmospheric models.

More robust conclusions can be achieved for winter NAO. For that season, precipitation intermittency only weakly affects the strong imprint of NAO on surface air temperature, and NAO-climate relationships further appear stable throughout the investigated datasets and time periods. The correlations between winter NAO and T_p likely capture correctly the actual signs and patterns of NAO-δ^{18}O correlation. However, other factors do affect δ^{18}O and may alter the relationship observed for T_p. The NAO signal in Central European δ^{18}O appears as a robust feature, observed from the long GNIP data and captured in all atmospheric simulations with the right location and strength. The same conclusion applies to the anti-correlation between NAO and Greenland winter

$\delta^{18}O$. Other areas depict a strong simulated NAO imprint which cannot be compared to direct observations (e.g. North America, Russia). The consistency of several of our findings between NCEP, ERA-interim and LMDZiso, and with daily station data, suggest that they are robust. We have also shown that they are not affected by a different definition of NAO accounting for changes in action centers.

Our study shows the potential of using direct (ice cores) or indirect archives of winter precipitation $\delta^{18}O$ for reconstructing past changes in the winter NAO index. Many published reconstructions referenced in Sect. 1 rely on precipitation or temperature sensitive proxies. The winter NAO-precipitation relationship does not have large-scale characteristics that are robust over different atmospheric models and time periods, apart from its negative imprint in SW Greenland and Mediterranean area, and positive imprint in Northern Europe. Our analysis has suggested a limited impact of precipitation intermittency, showing similar structure of winter NAO-T and NAO-T_p correlations. The NCEP data even suggest that the NAO-T_p correlation is stable in time for different 20 yr time periods (Supplement Sect. E, Fig. S6).

We suggest using the T_p variable (from in situ meteorological data, or from analyses) as a target for calibrating $\delta^{18}O$ proxy records, rather than the mean winter temperature, because of the magnitude and variability of the precipitation intermittency bias (Fig. 1). Of course, obtaining pure winter $\delta^{18}O$ records from natural archives other than ice cores is extremely difficult. However, deposition and archiving processes may propagate and preserve the winter NAO imprint in precipitation $\delta^{18}O$ into summer or annual mean signals. This could be the case for tree ring cellulose, for instance if soil moisture is refilled by snow melt in mountainous or high elevation areas. Such processes probably explain why a winter NAO signal is found in tree ring cellulose $\delta^{18}O$ from Siberia, where data from Yakoutia appear in anti-phase with those from Taimyr, and in phase with the variability depicted in Greenland ice cores (Sidorova et al., 2010). Annually resolved speleothems may also preserve a winter precipitation $\delta^{18}O$ signal, despite the complexity of infiltration and calcification processes.

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Compiling records of past precipitation δ^{18}O inter-annual variability derived from different natural archives is the next step required for using them towards NAO reconstructions. The present study has allowed us to identify the regions where precipitation weighted temperature and δ^{18}O exhibit the maximum correlation or anti-correlation with the NAO over the recent instrumental era. The combination of records from the Mediterranean area (where winter precipitation is important), with records from Central and Northern Europe, Québec, Greenland and Russia/Siberia/Yakoutia has the potential to allow new, more robust reconstructions of the NAO. This could be achieved from precipitation δ^{18}O estimated using lake sediments, speleothems, ice cores and tree ring cellulose, albeit with the difficulties related to the seasonality and complexity of each archiving process. Multi-centennial simulations conducted with coupled ocean-atmosphere models equipped with water stable isotopes could also be used to assess the stability of the NAO-δ^{18}O relationships through time, in response to different forcings and under different states of e.g. the Atlantic Multi-decadal Oscillation (Lehner et al., 2012). Although restricted to the last decades, our study suggests to systematically combine at least data from Europe and from Greenland in order to capture the structure of NAO imprints.

Supplementary material related to this article is available online at: http://www.clim-past-discuss.net/8/4957/2012/cpd-8-4957-2012-supplement.pdf.

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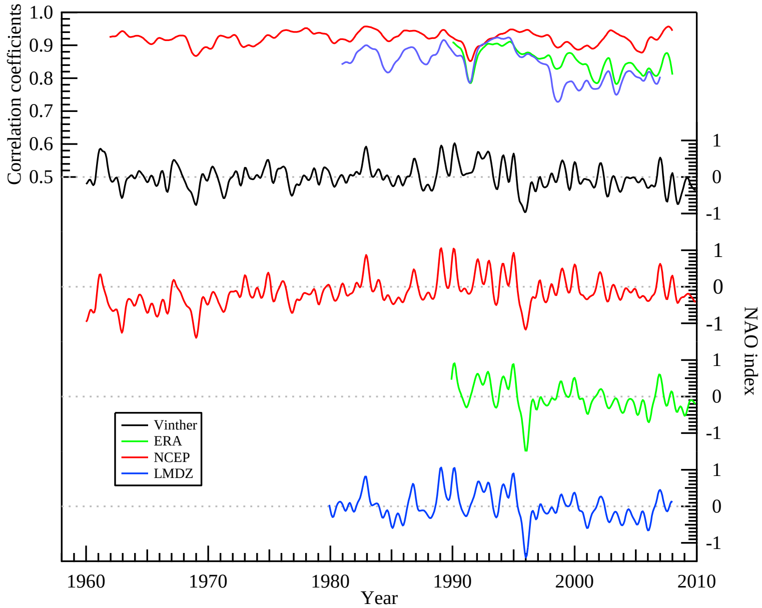

Fig. 1. Monthly NAO indices from 1960 to 2010 defined as normalized pressures differences between Gibraltar and Reykjavik. The lower panels depict individual time series: instrumental data (black), NCEP (red) and ERA (green) analyses, and LMDZiso simulation (blue). An 11 month binomial filter was applied on monthly data to stress annual minima and maxima. Upper panel: 25 month running correlation coefficient between the reference instrumental time series and ERA (green), NCEP (red) and LMDZiso (blue) outputs.

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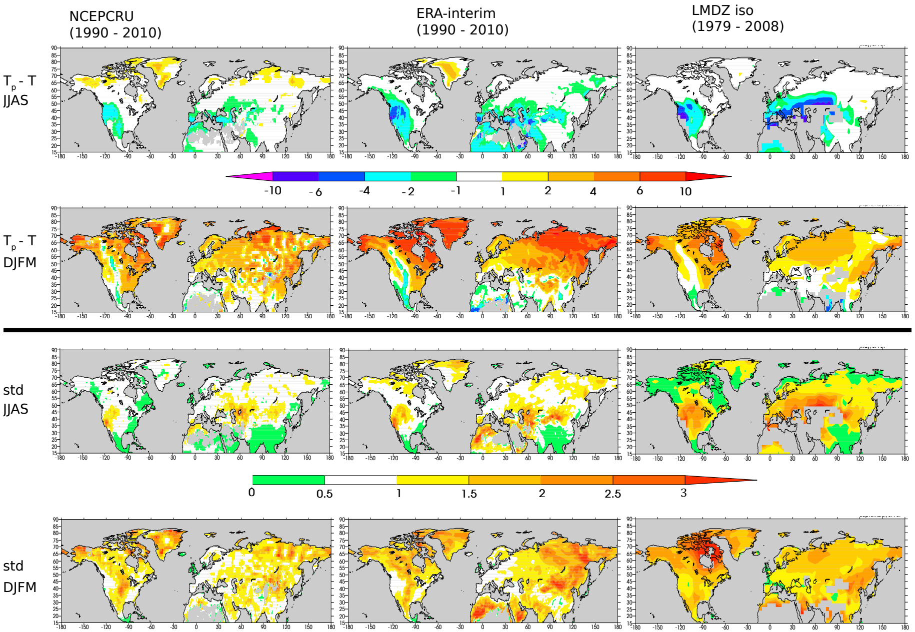

Fig. 2. (a) Mean precipitation intermittency bias, defined as the difference between the daily precipitation weighted temperature (T_p) and the mean seasonal temperature (T) in CRU-NCEP, ERA-interim and LMDZiso, for both JJAS and DJFM; **(b)** Same as the upper panels but showing the inter-annual standard deviation (std) of the biases.

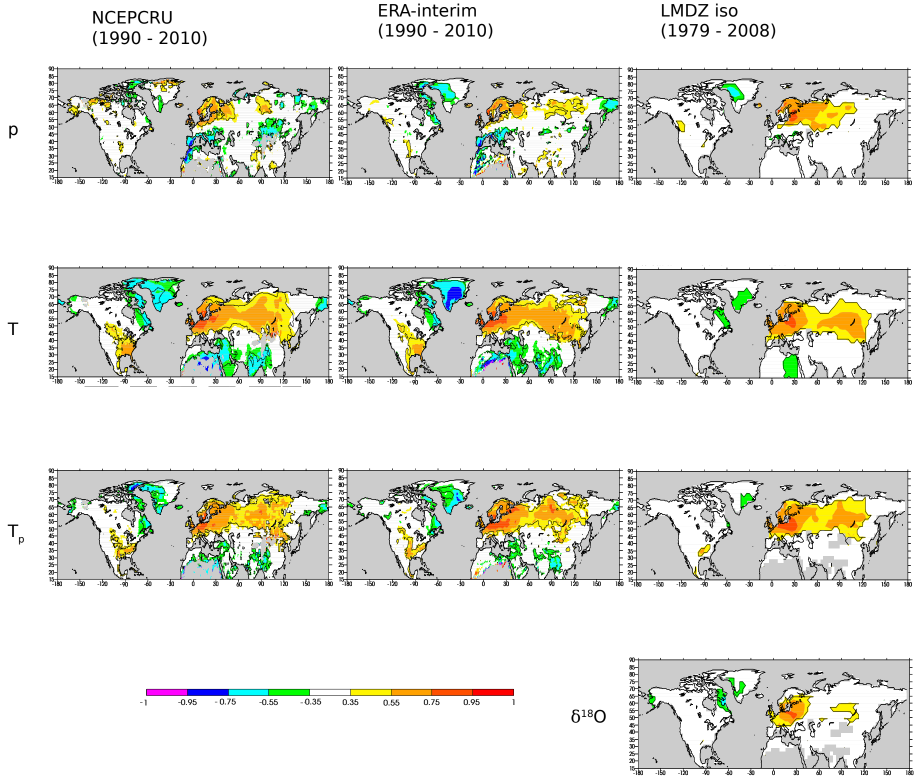

Fig. 3. Linear correlation coefficient between the respective DJFM NAO indices and precipitation (p), temperature (T) and precipitation-weighted temperature (T_p) fields in CRU-NCEP (left), ERA-interim (middle) and LMDZiso (right). Significant correlations are highlighted using a Student t-test ($p < 0.05$) (thin black line). Lowest panel: correlation coefficients of the DJFM NAO index with the winter $\delta^{18}O$ time series in LMDZiso.

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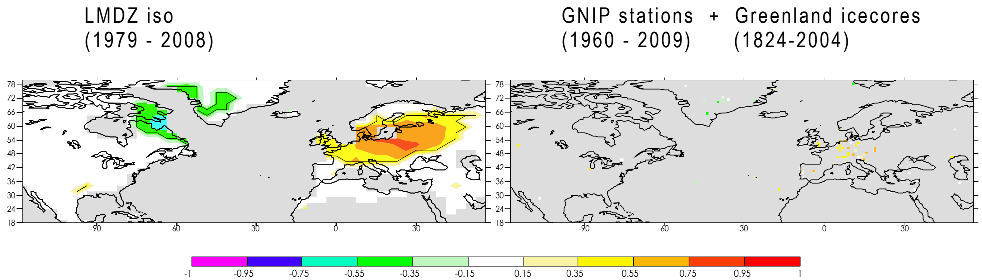

Fig. 4. Linear correlation coefficient between the respective DJFM NAO indices and δ^{18}O from GNIP and Greenland ice core data (left; see text and Supplement Sect. C for details) and for LMDZiso (right); zoomed on the areas where observations are available. Note that a different color scale is used compared to Fig. 3, which allows to show the sign of weak but significant correlations (light green and light yellow, with absolute correlation coefficients between 0.15 and 0.35). White areas indicate small (non significant) correlation coefficients.