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Potential causes of 15th century Arctic warming using coupled model simulations with data assimilation

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Abstract

An ensemble of simulations of the climate of the past millennium using a three-dimensional climate model of intermediate complexity are constrained to follow temperature histories obtained from a recent compilation of well-calibrated surface temperature proxies using a simple data assimilation technique. Those simulations provide a reconstruction of the climate of the Arctic that is compatible with model physics, the forcing applied and the proxy records. Available observational data, proxy-based reconstructions and our model results suggest that the Arctic climate is characterized by substantial variations in surface temperature over the past millennium. Though the most recent decades are likely to be the warmest of the past millennium, we find evidence for substantial past warming episodes in the Arctic. In particular, our model reconstructions show a particularly warm period at the end of the 15th century. This warm event is likely related to the internal variability of the climate system. We examine the roles of competing mechanisms that could potentially produce this anomaly. These examinations lead us to conclude that changes in atmospheric circulation, through enhanced southwesterly winds towards northern Europe, Siberia and Canada, are likely the main cause of the Arctic warming during the late 15th century.

1 Introduction

Studies of Arctic climate indicate considerable warming in this region in recent decades. For the past 100 years, the Arctic has warmed twice as much as the global average (Trenberth et al., 2007). This warming is associated to a substantial diminution of sea ice thickness (Serreze et al., 2000) and extent (Meier et al., 2005).

While recent Arctic warmth appears anomalous, observational data and proxy data both indicate substantial long-term temperature variability in the region. A multidecadal interval of relative warmth, for example, can be found during the early 20th century, between the 1920s and 1940s, when conditions were only modestly less warmth than today. Annual mean temperatures in the Arctic appear to have been substantially (roughly 2°C) warmer in the 1940s than in the 1910s (Bengtsson et al., 2004). While instrumental temperature data are relatively sparse during the first half of the last century, the early 20th century Arctic warm period appears to have been associated with a different large-scale spatial pattern of warmth from the current warm period. The early-century warming was largely confined to the Arctic alone (i.e. the region north of 60° N) while the latter warming is more widespread, with a pronounced warming in the Eurasian mid-latitudes (Kuzmina et al., 2008; Johannessen et al., 2004; Overland et al., 2004).

The dynamical processes underlying these two Arctic warm periods are also likely different. For the most recent decades, it is almost certain that forcing by anthropogenic greenhouse gas increases has dominated over the contribution from internal variability (Johannessen et al., 2004), though the relative role of natural multidecadal variability is not entirely resolved (Polyakov and Johnson, 2000). By contrast, during the early 20th century when anthropogenic forcing was considerably weaker than today, the observed Arctic warming event was likely due to the natural variability of the climate system. While natural external radiative forcing by variations in solar and volcanic forcing could have played some role in this early warming, the natural forcing is insufficient in producing more than a few tenths of a degree warming (e.g. Crowley, 2000), i.e. it is likely an order of magnitude too small to explain the observed early 20th century Arctic warming. It has been proposed that the early 20th century warming was caused by increased southwesterly winds and oceanic heat transport into the Barents Sea region (Bengtsson et al., 2004; Overland et al., 2004; Rogers, 1985). There is evidence that these changes were, in turn, associated with purely internal, multidecadal oscillatory variability of the climate system (Bengtsson et al., 2004; Johannessen et al., 2004; Overland et al., 2004; Delworth and Mann, 2000; Delworth and Knutson, 2000; Przybylak et al., 2000).

The absence of direct instrumental data before the mid-19th century requires the use of climate “proxies”, such as tree rings, corals, ice cores, lake sediments and historical

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documents, from which we can infer some key characteristics of climate changes in past centuries. Such compilations from high northern latitudes (e.g. Jiang et al., 2005; Jennings and Weiner, 1996; Massé et al., 2008; D'Arrigo and Jacoby, 1993; Jacoby and D'Arrigo, 1989; Overpeck et al., 1997; Ogilvie and Jónsson, 2001) suggest that similar Arctic warm events may have occurred in past centuries. In this study, we focus on the evidence and dynamical explanations for any such extended periods of Arctic warmth during the past millennium. Proxy reconstructions of global or hemispheric mean surface temperatures (e.g. Mann et al., 1999, 2005b, 2008; Briffa et al., 2001; Jones et al., 2001; Esper et al., 2002; Mann and Jones, 2003; Jones and Mann, 2004; Jansen et al., 2007) suggest the existence of a period of modest large-scale warmth from the 10th–14th centuries, though it does not rival current warmth. This so-called "Medieval Warm Period" is followed by a period of relative large-scale coolness over the 15th-19th centuries known as the "Little-Ice Age". At the hemispheric or global scale, these changes are consistent with the response of the climate system to external changes over the past millennium in natural (and after the 19th century, anthropogenic) radiative forcing (e.g. Crowley, 2000). At regional or local scales, however, the influence of the forced response of the climate may be overwhelmed by the contribution of internal climate dynamical processes (Goosse et al., 2005).

In this study, we seek, as in previous studies (e.g. Goosse et al., 2008a) to merge the observational information contained in actual proxy records with the physical and dynamical constraints present in forced climate model simulations, to interpret past climate changes. Here our focus is on using such analyses to interpret the impacts of large-scale dynamics, as well as radiative forcing changes, on the inferred pattern of past regional temperature changes.

Our model simulations employed LOVECLIM1.1 (Goosse et al., 2007). A set of five different experiments covering the last millennium were run with data assimilation. More specifically, the evolution of the model was constrained by selecting, among all available realizations, the realization of the internal variability that most closely matches the information from the proxies. These experiments allow us to advance hypotheses

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about the mechanisms associated with any intervals of regional Arctic warming. We performed a parallel ensemble of simulations without data assimilation. The ensemble mean in the latter case can be used to define the response of the system to external forcing alone, since the influence of natural internal variability, which differs from one realization to another, is removed by the averaging process. Comparisons between these two parallel sets of experiments allow us to isolate the relative contributions of both external forcing and internal variability.

We first describe the model and experimental design, the forcings applied and the data assimilation technique. The assimilated proxy records are taken from a recent compilation (Mann et al., 2008) of a large network of high-resolution (that is, decadally or annually-resolved) climate proxy data. Our focus is on a particularly warm period in the 15th century that is evident in the proxy data. Using the model data assimilation experiments, we analyze the role of various physical and dynamical processes that appear responsible for the pattern of observed Arctic warmth, and demonstrate that this pattern likely arises from dynamical variability.

2 Model description and experimental design

The different simulations analyzed in this study were performed with LOVECLIM1.1 (Driesschaert et al., 2007; Goosse et al., 2007), a three-dimensional climate model of intermediate complexity which includes representations of the atmosphere, the ocean and sea ice, the terrestrial biosphere, the oceanic carbon cycle and the polar ice sheets. As the last two components are not activated in this study, they will not be described here. The atmospheric component of LOVECLIM is ECBILT2 (Opsteegh et al., 1998), a quasi-geostrophic model of horizontal resolution T21 and three vertical levels, with simple parameterisations for the diabatic heating due to radiative fluxes, the release of latent heat, and the exchange of sensible heat with the surface. The oceanic component is CLIO3 (Goosse and Fichefet, 1999). This model is made up of a primitive-equation, free-surface ocean general circulation model coupled to a thermodynamic-dynamic sea-ice model. Its horizontal resolution is 3° in longitude and latitude, and there are 20 unevenly spaced vertical levels in the ocean. The terrestrial vegetation module VECODE (Brovkin et al., 2002) computes annually the evolution of trees, grass and deserts. It has the same resolution as ECBILT. More information about the model can be obtained at: http://www.astr.ucl.ac.be/index.php?page=LOVECLIM%40Description.

All the simulations were driven by the same forcings. It includes three natural forcings, namely the changes in orbital parameters, the volcanic activity and the variations in solar irradiance, as well as three anthropogenic forcings, i.e., the changes in greenhouse gas concentrations, including tropospheric ozone, the variations in sulphate aerosols loading, and the forcing due to changes in land-use. The temporal evolution of some of these forcings is shown in Fig. 1. The variations of orbital parameters follows Berger (1978). The effect of volcanism is derived from Crowley (2000) and is included through changes in solar irradiance. The evolution of solar irradiance follows the reconstruction of Muscheler et al. (2007). The evolution of greenhouse gas concentrations is based on a compilation of ice cores measurements (J. Flueckiger, personal communication, 2004). The influence of anthropogenic sulphate aerosols is taken into account through a modification of surface albedo (Charlson et al., 1991). The changes in land-use is based on Ramankutty and Foley (1999) and is applied in the model through a reduction of the area covered by trees and an increase in grassland as VECODE does not include a specific vegetation type corresponding to cropland.

The goal of this study is to obtain a simulation of the Arctic climate for the last millennium that is not only consistent with our model and the forcings applied, but also with the data available for that period. For that purpose, we constrain the model results using the recent compilation of well-calibrated surface temperature proxy records of Mann et al. (2008). This is achieved thanks to a new version (see Goosse et al., 2008b) of the data assimilation technique described in Goosse et al. (2006). We proceed in the following way: we start the simulation at the year 1000 AD, from a condition obtained from a long simulation covering the whole Holocene (Goosse et al., 2007). By intro-

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ducing small perturbations on the streamfunction in the atmosphere we generate an ensemble of 96 simulations for a short period of time (1, 5, 10 or 20 years). Then, we choose among those 96 representations of the model internal variability the one that is the closest to the proxy records available for the period of time investigated. This is achieved by using the following cost function:

$$CF_k(t) = \sqrt{\sum_{i=1}^{n} w_i \left(F_{\text{obs}}(t) - F_{\text{mod}}^k(t)\right)^2}.$$

$CF_k(t)$ is the value of the cost function for each member k of the ensemble for a particular period t. n is the number of proxies used in the model/data comparison. $F_{\text{obs}}(t)$ is the value of the variable F (the surface temperature in this case) in the proxy records at the location where they are available, and $F_{\text{mod}}^k(t)$ is the value of the same variable simulated by the model in the simulation k at the same location as the proxy record. w_i is a weight factor. The experiment k which minimizes the cost function CF_k is selected for that particular period of time, and the end of this simulation is used as the basis for the initial condition of the new ensemble of simulations performed over the next period. The procedure follows in the same way for the whole millennium. As this method requires a large number of simulations, LOVECLIM coarse resolution and low computer-time requirements is appropriate.

A set of 56 annually-resolved multiple proxy series (Mann et al., 2008) are used to constrain the model. They essentially consist of tree-rings, ice cores, some lake sediments and historical documents. Through the screening analysis, only the proxies having a statistically significant correlation with local instrumental surface temperature data during the calibration interval (1850–1995) were selected. When proxy records reflected temperature variations at sub-annual resolution, they were averaged to obtain annual mean values. The proxy records were decadally-smoothed using a low pass filter. Then, the series were converted to standardized anomalies by centering and scaling them to have zero mean over the calibration interval. The proxies were then averaged and scaled to the same mean and decadal standard deviation as the nearest available instrumental surface temperature gridbox (5° in latitude and longitude) series over the calibration period. We have kept only the records available back to the year 1400 AD, and which extend through AD 1995. The proxy data are primarily terrestrial, and cover tropical, extratropical, and polar regions, though the largest numbers of proxy records lie northward of 30° N. The location of proxy series available in the Arctic region is shown in Fig. 2. The available sampling favors Scandinavia, Siberia and western North America, while there is a dearth of data in certain regions such as eastern North America.

We present in this paper the results obtained from 5 different numerical experiments using data assimilation. They start from the same initial conditions, but use different ways to constrain the model and different periods of time in the computation of the cost function. In the first experiment, the weight factors w_i are the same for all the proxy records and the cost function is evaluated for 1-year averages. In the other four simulations, in order to give a larger weight to proxies which are more reliable, the value of the weight factors w_i is proportional to the correlation between the proxy records and the observations of temperature obtained during the instrumental period. In these 4 experiments, the averaging period in the computation of the cost function is set to 1, 5, 10 and 20 years, in order to test if this has an impact on our results. For instance, for 20-year mean, processes responsible for interannual variability may be filtered while they can play an important role in the selection of the best experiment when 1-year mean are analyzed. These different experiments help us testing the robustness of our results, making sure that for slightly different ways to constrain the model evolution with the proxy data, we obtain similar results, and that the technique yields internally consistent results. The mean of the 5 experiments can provide a better estimate of the variability by filtering the noise, while the standard deviation of those 5 members give us an estimate of the uncertainty of our results.

In addition, an ensemble of 10 simulations is performed without data assimilation. This ensemble is run with the same model and the same forcings than the ones used for the simulations performed with data assimilation, but with slightly different initial conditions for each of the members of the ensemble. The ensemble mean will help us assessing the response of the system only to the external forcings, and by comparing it with the experiments with data assimilation, we will be able to separate the changes caused by the natural variability to those caused by the external forcings in the model.

3 Comparison of model results with proxy data

Before analyzing the climate evolution obtained in our simulations, we have to assess the robustness of the technique of data assimilation and the quality of model results by comparing them with the proxy records used to constrain the model. The comparison between the anomaly of the spatial pattern of annual mean surface temperature proxies and the model results (we have retained from the model only the points where the proxies are available) is shown in Fig. 2. We represent a typical warm and cold period, averaged over 50 years, taking place during years 1470–1520 AD and 1600–1650 AD, respectively. In general, the spatial pattern of surface temperature simulated in the model is reasonably close to the proxy data, although some local large differences can be observed, for example in the North American region. To understand those local discrepancies we have to keep in mind that some proxies contained sizeable non-climatic sources of noise which will not be correlated over local scales, and that the model may be deficient in representing the variability at such scales (i.e. one model grid scale). Both factors could lead to substantial local differences between model results and the proxy reconstructions. On the other hand, as show in Fig. 3, the model results show a better agreement with proxy records at regional scales. The temporal evolution of surface temperature averaged over three small regions where proxies are available are showed (boxes in Fig. 2 define these different regions). We obtain a good agreement between the surface temperature computed in each one of the 5 model simulations and the proxy-based reconstruction. For the average over each region, we measure the misfit between model results (mean of the 5 experiments) and proxy series by calculating the root mean-square error (RMSE) for the period 1400–1996. In the first

(RMSE=0.08) and second (RMSE=0.1) regions, all simulations are in good agreement with the proxy records. The third region (RMSE=0.21) presents as well good results, although some discrepancies with proxy data and a larger variance between model simulations is observed. For instance, the amplitude of the early 17th century cooling in that region is larger in the proxies than in the different model simulations, and this minimum is shifted.

Finally, in Fig. 3d, we compare the annual mean surface temperature averaged over the whole Arctic (region northward of 64° N) obtained in the different simulations with the high-latitude summer-weighted annual temperature reconstruction of Overpeck et al. (1997). The "Little Ice Age" and subsequent warming recorded by this compilation are reproduced in the model simulations. The agreement between model and proxy data is quite good overall, though the mid-19th century is colder in the Overpeck et al. reconstruction than in the model. The model also tends to simulate slightly too high temperatures at the end of the 20th century.

4 The late 15th century warm period

The annual mean surface temperature in the Arctic in the 5 simulations including data assimilation (Fig. 4, blue curve) shows the relatively warmth during the first five centuries that is evident in hemispheric climate reconstructions (e.g. Jansen et al., 2006; Mann et al., 2008). The mean surface temperature northward of 64° N during the 12th century is about 0.2°C warmer than over the reference period 1600–1950. The cooling that follows, starting at the beginning of the 13th century, is interrupted by some warming periods. Two important peaks of temperature are observed at the beginning and the end of the 15th century. They correspond to the warmest periods of the last millennium before the industrial period for the mean over the five experiments, i.e. that, in our simulations, they are warmer than the so-called "Medieval Warm Period" in the Arctic. The "Little Ice Age" then follows, with relatively cool temperatures during the 16th, 17th and 19th centuries. From the beginning of the 20th century to the present, there was an abrupt increasing trend in surface temperature, associated with anthropogenic forcing.

The scatter between the 5 experiments with data assimilation is measured by the standard deviation of the 5 members. During the first 4 centuries of the last millennium, a fewer number of proxies is available. The variance between the different model simulations is thus larger than for the next centuries. The short standard deviation observed for the 15th century period (standard deviation=0.06) indicates that the uncertainty of our results is relatively small.

To interpret the simulated temperature changes, we compare our experiments with data assimilation with those without data assimilation (forced response). The peak medieval Arctic warmth is greater in the simulations without data assimilation (Fig. 4, green curve). Averaged over the years 1100 to 1150, the temperature is almost 0.5°C higher than the mean over the reference period in the forced response. The millennial-scale cooling trend (approximately half a degree over the millennium) is thus more pronounced in the forced response than in the simulations with data assimilation. Several causes could be responsible for this discrepancy. The forcing used in the model (and thus the forced response) is uncertain and prone to potential systematic error (e.g. Jones and Mann, 2004). Internal variability of the system at any low-frequency may induce a cooling in the Arctic, counterbalancing the effect of the forcing. There are also uncertainties in the proxy temperature reconstructions themselves, which become increasingly substantial in the earlier centuries of the past millennium (Mann et al., 2008).

The relatively mild conditions during the early 15th century appears largely consistent with the forcing (low volcanic activity, relatively high total solar irradiance, see Fig. 1). By contrast, the late 15th century warm period is less clear in the forced response of the model. It is possible that the response of the model to the external forcings is actually not correct and that the data assimilation technique takes charge to head the system in the good direction. For instance, the response of the atmospheric circulation to external forcing, such as solar and volcanic forcing, is weak in LOVECLIM (Goosse and Renssen, 2004), while it has been suggested that the Arctic

Oscillation/North Atlantic Oscillation (e.g. Shindell et al., 2001) and El Niño-Southern Oscillation (Mann et al., 2005a) response to external radiative forcing has a strong impact on past regional climatic changes. However, during the 15th century, no particular event in external forcing, no special solar neither volcanic activity change, can be seen (Fig. 1). It is thus reasonable to attribute this event to a particular realization of the internal variability of the system.

In order to find the causes of the changes in temperature during the late 15th century simulated in our model including data assimilation, we analyze the anomalies in atmospheric and oceanic heat transports, an information not available from proxy records. The mean of the 5 model simulations performed with data assimilation is used in the following patterns.

The simulated spatial distribution of annual surface temperature anomaly for the warm period averaged over the years 1470 to 1520 AD (Fig. 5), shows an overall warming over the Arctic region. The few proxy records available in this region for that period are in good agreement with the model results (Fig 2a). This pattern is robust in our model as each individual simulation gives similar ones (not shown). The largest warming is located in the Canadian Archipelago and Eurasian Arctic, with the maximum in the Barents Sea, whose temperature is almost 0.6°C higher than in the reference period.

The pattern of the annual mean anomaly of the geopotential at 800 hPa, averaged over the period 1470–1520 (Fig. 6), is consistent with the particularly warm conditions of that period. The negative anomaly west of Iceland produces an increased inflow of warm air coming from the south, leading to the warming over northern Europe, the Barents Sea and the western Siberian region. Similarly, the negative anomaly centered over the Bering Strait induces a warming over Canada. By contrast, in regions characterized by winds anomaly coming from the north, such as the Baffin Bay and the eastern Siberia, the temperature anomaly is weak and even negative in some regions. The negative geopotential anomaly at high latitude and the positive one at mid latitudes present some similarities with the circulation pattern associated with the Arctic

Oscillation (AO). Changes in the position and intensity of the Aleutian Low, as shown in Fig. 6, are characteristics of the PDO (Gedalof et al., 2002). However, the circulation anomaly responsible for the warm conditions in the model could not be clearly linked to those modes. Rather, it corresponds to the right combination of anomalies in both the Atlantic and Pacific sectors that leads to a warming of nearly all regions in the Arctic and thus a clear signal on the regional mean shown in Fig. 4.

Changes in oceanic circulation could also have an impact on regional temperature changes during the last millennium. However, the model does not simulate any clear oceanic signal during the period 1470–1520. For instance, Fig. 7 shows that the meridional transport of heat in the North Atlantic Ocean towards the Arctic does not experience any large variations over the last millennium in our simulations. Consequently our results do not support attribution of the warming observed in the Arctic seas during the late 15th century to changes in oceanic circulation. A slight continuous general increase of the poleward heat transport is observed during the millennium, as well as some similarities with Fig. 4. Nevertheless, changes are not significantly different to zero. This weak oceanic response in the model may be due to the experimental design: we are not constraining directly the oceanic changes since the proxies selected for the data assimilation are located only on continents and continental shelves. Though some oceanic proxies at high latitudes are available, including, for instance, records derived from benthic and planktonic foraminifera, stable isotopes and diatom assemblages (Eiriksson et al., 2006; Lund et al., 2006; Klitgaard Kristensen et al., 2004; Jiang et al., 2002; Mikalsen et al., 2001; Black et al., 1999), the quantity of continuous high-resolution marine sedimentary proxy records in the Arctic Ocean over the last millennium is rather low. Furthermore, the uncertainty associated with the calibration and dating of the marine records is generally larger than the other types of proxy records (Jones and Mann, 2004). As a consequence, incorporating those proxies in our data assimilation procedure is not possible at this stage. Most studies suggest that some regional temperature variability coincides with changes of oceanic circulation in the North Atlantic region, in particular, some indicate a role of the ocean in the Atlantic

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decadal variability. However, none of these studies highlight particular conditions during the 15th century that would suggest a clear underestimation of the role of the ocean in our simulations.

5 Conclusions

In our simulations over the past millennium using the model LOVECLIM with a data assimilation technique, the highest mean temperature of the last millennium before the 20th century over the Arctic occurred at the end of the 15th century. During this period, the simulated temperatures are even higher than during the so-called "Medieval Warm Period". As the forced response of the model does not produce such an event, this warm period can be interpreted as induced by internal adjustments of the climate system.

Our model results, consistent with both the available proxy data and the model physics, clearly show that the simulated 15th century Arctic warming is almost entirely explainable in terms of changes in atmospheric circulation. The negative geopotential anomalies west of Iceland and in the North Pacific provide a coherent explanation for the increase in surface temperature during that period. The late 15th century patterns of surface temperature and sea level pressure is somewhat similar to the early 20th century Arctic warm event. The available data indicates that the winter times in the 1920s were characterized by increased warm air inflow into Europe, while the Baffin Bay experienced a cooling (Overland et al., 2004; Bengtsson et al., 2004). The pattern of sea level pressure (SLP) anomalies during this period is comparable with the pattern of the 15th century warming obtained in our model reconstructions (the geopotential height being the closest variable to the SLP in the model). The early 20th century warm event might thus not have been unique in the recent past. Furthermore, changes in the intensity and position of the Aleutian Low are responsible of the warming over the Canadian Archipelago. The relatively large event in the 15th century appears thus as a consequence of coincident changes in the European and Pacific sectors that also

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play a role in variations of Arctic climate during the 20th and early 21st centuries (e.g. Overland and Wang, 2005).

No robust change in the patterns of oceanic circulation could be found in our model results to explain the changes observed in the Arctic seas during the 15th century warm event. The absence of strong response of the ocean in our simulations covering the last millennium may be due to the data assimilation and in particular to the lack of well calibrated oceanic proxies for the last millennium. Furthermore, the existence of a mode of multidecadal variability in the North Atlantic, related to fluctuations in the intensity of the thermohaline circulation, has been put in evidence in several studies (Delworth and Mann, 2000; Knight et al., 2005), and could explain part of the temperature variations at high latitudes (Zhang et al., 2007). The important intensification of the Atlantic water inflow to the Arctic, which explains some of the recent warming of the Arctic Ocean (Zhang et al., 1998; Gerdes et al., 2003), could have been reproduced in past times. As a consequence, additional work is required to investigate the role of oceanic circulation in past changes in the Arctic. In this framework, it would be very interesting to include, when available, marine proxies in the data selected to constrain our model results.

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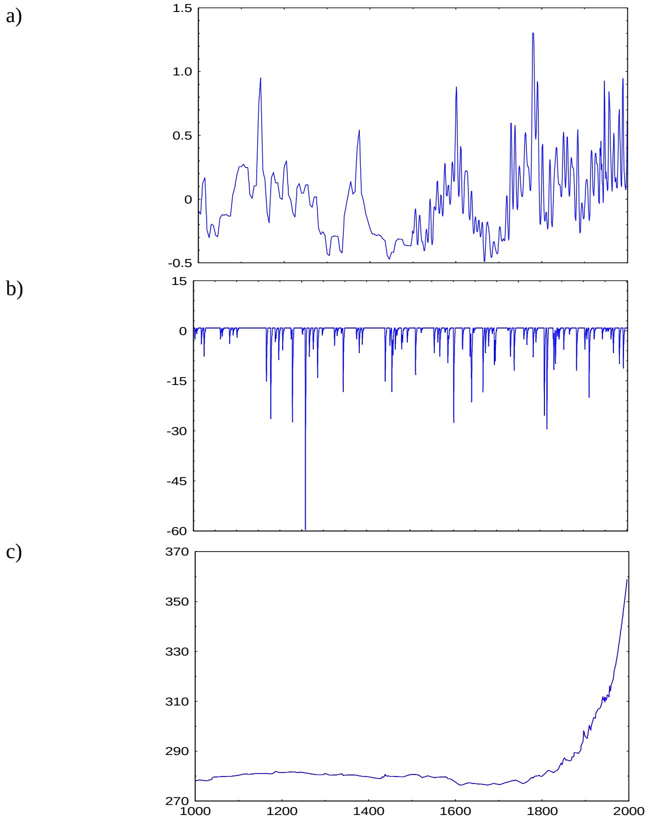

Fig. 1. (a) Global mean radiative forcing (W/m^2) used to drive LOVECLIM simulations for the last 1000 years associated to variations in the total solar irradiance based on Muscheler et al. (2007). **(b)** Radiative forcing (W/m^2) associated to volcanic activity according to Crowley (2000) for the region including latitudes from 35° N to 90° N, included in LOVECLIM by a modification in the solar irradiance. **(c)** Time series of CO_2 concentrations (ppmv).

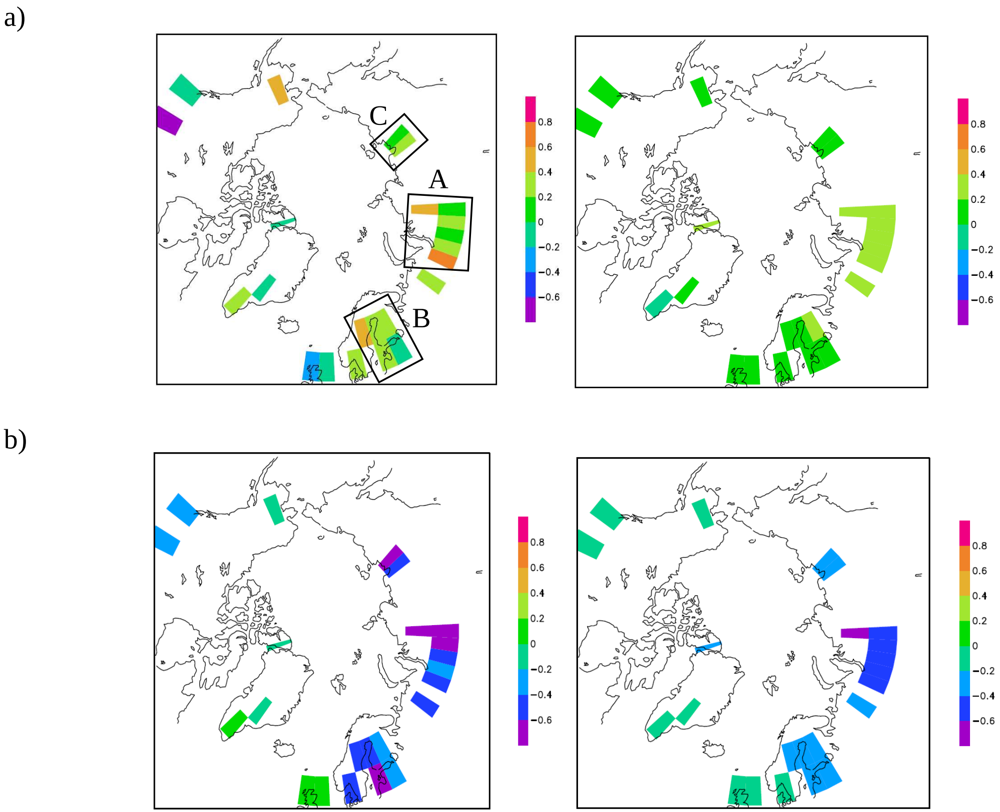

Fig. 2. Anomaly in annual mean surface temperature (˚C) during a warm and a cold period in the proxy data (left column) and the model results averaged over the 5 simulations (right column). The model results are shown only at the location where the proxies are available. **(a)** 1470–1520 and **(b)** 1600–1650. The reference period is 1600–1950 AD. The boxes in (a) correspond to the regions over which averages are performed to obtain the time series shown in Fig. 3.

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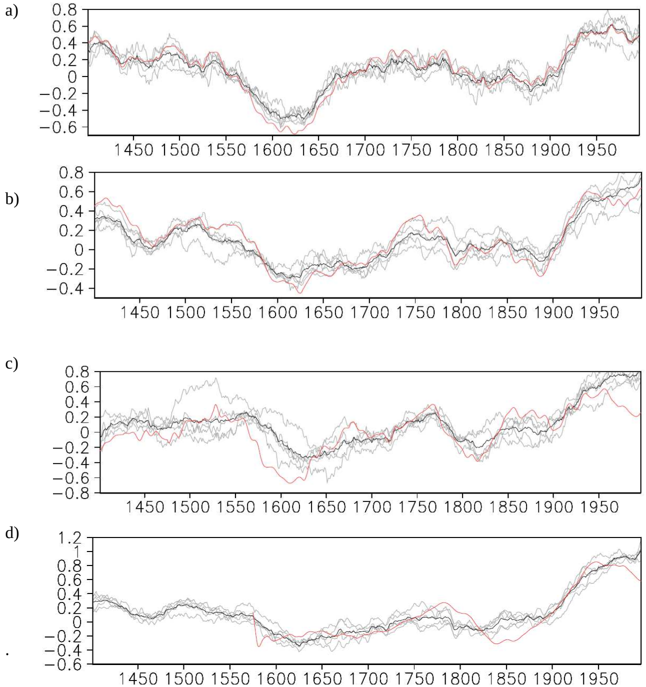

Fig. 3. **(a)** Time series of the anomaly in annual mean surface temperature (°C) over the last 600 years for the region in the box A on Fig. 2. The black line is the mean over the 5 model simulations, the red line is the average over the 6 proxy data contained in box A on Fig. 2, and the grey lines are the results of the 5 different model simulations. **(b)** Same as (a) for the mean over box B on Fig. 2 (5 proxies). **(c)** Same as (a) for the mean over box C on Fig. 2 (2 proxies). **(d)** Anomaly of annual mean surface temperature in the Arctic for the last 600 years. The Arctic area corresponds to the mean over all longitudes between 64° N and 80° N. The red curve is the reconstruction of Overpeck et al. (1997). The reference period is 1600–1950 AD. A 51-year running mean has been applied to all time series.

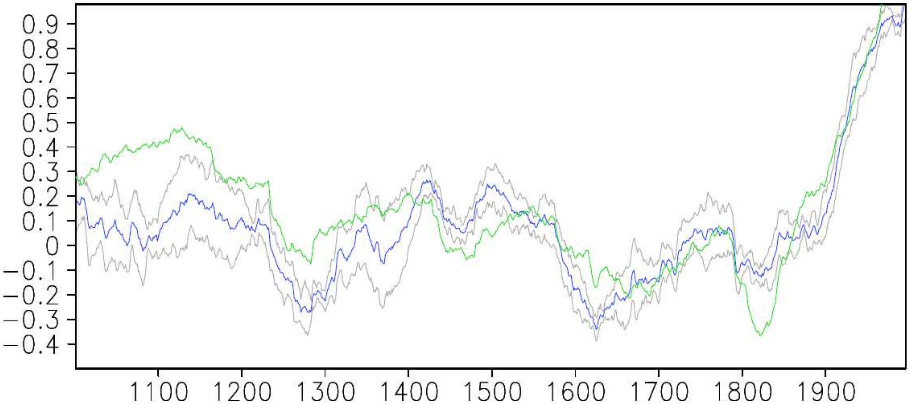

Fig. 4. Anomaly in annual mean surface temperature (°C) in the Arctic over the past millennium. The blue line is the average over the 5 model simulations performed with data assimilation, and the grey lines are the mean plus and minus one standard deviation of the ensemble. The green curve is the mean of an ensemble of 10 simulations made without data assimilation. A 51-year running mean has been applied to the time series. The reference period is 1600–1950 AD.

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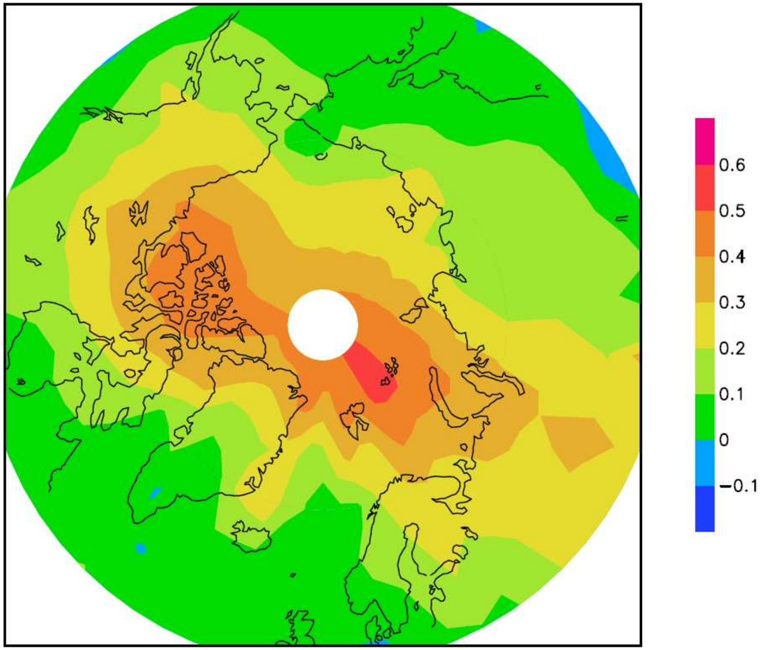

Fig. 5. Anomaly in annual mean surface temperature (˚C) over the late 15th century warm period for the model results averaged over the 5 simulations with data assimilation (mean over the period 1470–1520 AD). The reference period is 1600–1950 AD.

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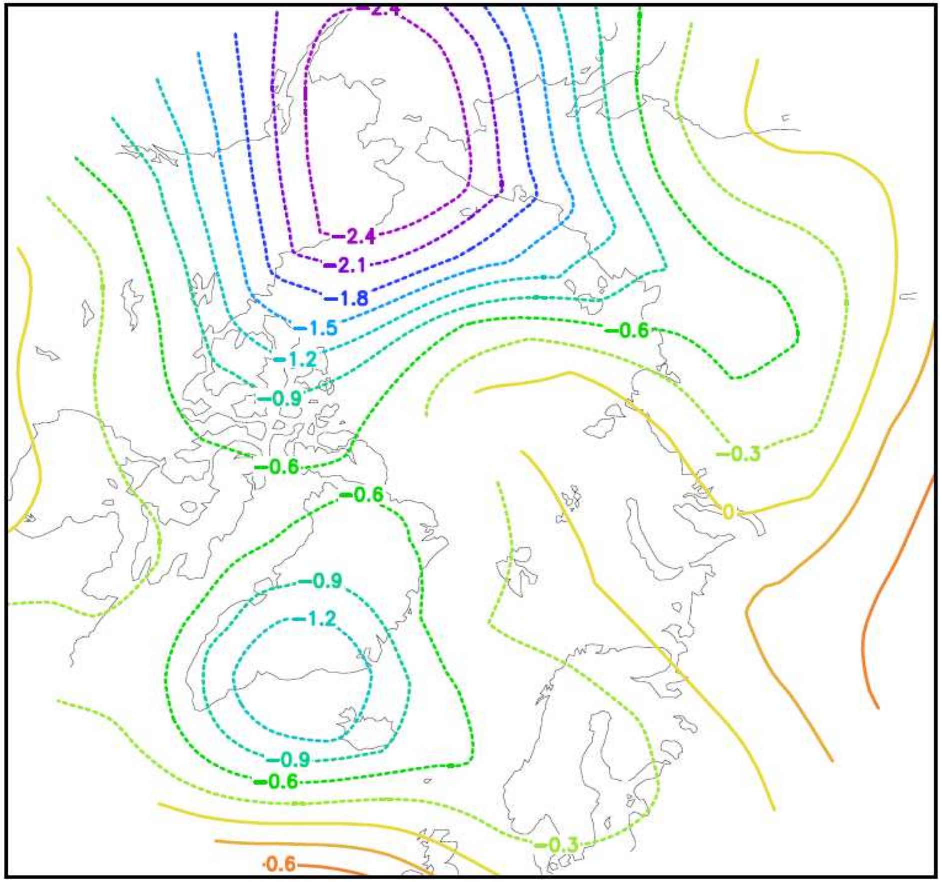

Fig. 6. Anomaly in annual mean 800 hPa geopotential height (m) over the late 15th century warm period for the model results averaged over the 5 simulations with data assimilation (mean over the period 1470–1520 AD). The reference period is 1600–1950 AD.

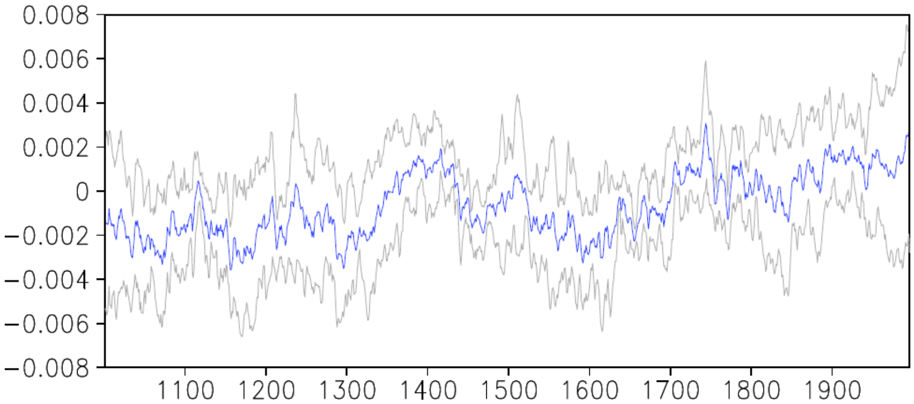

Fig. 7. Anomaly in meridional heat transport in the North Atlantic Ocean at 70° N (PW) for the average over the 5 model simulations performed with data assimilation, the grey lines are the mean plus and minus one standard deviation of the ensemble. A 51-year running mean has been applied to the time series. The reference period is 1600–1950 AD.

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