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Arctic climate over the past millennium: Annual and seasonal responses to external forcings

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Abstract

The annual and seasonal temperatures in the Arctic over the past 1150 years are analyzed in simulations performed with the three-dimensional Earth system model of intermediate complexity LOVECLIM forced by changes in solar irradiance, volcanic activity, land use, greenhouse gas concentrations and orbital parameters. The response of the system to individual forcings for each season is examined in order to evaluate the contribution of each forcing to the seasonal contrast. For summer, our results agree relatively well with the reconstruction of Kaufman et al. (2009). Our modelling results suggest that the temperature changes during this period were characterized by large seasonal differences. In particular, while annual mean temperatures display a decreasing trend during the pre-industrial period, spring temperatures appear to rise. The variations in the Earth's orbital parameters are the main cause for those seasonal differences. Larger climate variations are simulated in autumn compared with the other seasons in response to each forcing, particularly in response to changes in greenhouse gas concentration during the industrial period and in response to land use forcing, which surprisingly has a significant impact on Arctic temperature. These contrasting changes for the different seasons also underline the need for an adequate estimate of the season represented by a proxy.

Keywords

Arctic climate, climate modelling, external forcings, past millennium, seasonal contrast, temperature evolution

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Introduction

Many studies have been devoted to the reconstruction and understanding of the annual mean, large-scale temperature changes over the past millennium, using both proxy-based reconstructions and models (e.g. Briffa et al., 2001; Crowley, 2000; González-Rouco et al., 2006; Goosse et al., 2010a; Jones and Mann, 2004; Jones et al., 2009; Mann et al., 2008, 2009; Osborn and Briffa, 2006; Osborn et al., 2006; Rutherford et al., 2005). Less attention has been paid to seasonal trends at the regional scale, except maybe for Europe (see, for instance, Goosse et al., 2006; Guiot et al., 2010; Hegerl et al., 2011; Luterbacher et al., 2004; Xoplaki et al., 2005). However, it is important to improve our knowledge of the evolution of seasonal temperatures, because they may behave very differently from the annual ones (Bauer and Claussen, 2006; Jones et al., 2003). This has been clearly shown over the last 150 years, as the instrumental records exhibit a larger warming over the Northern Hemisphere in winter than in summer (Jones et al., 2003). These differences between changes in annual and seasonal temperatures can be explained by the response of the climate system to a specific forcing which may vary from one season to another and from one region to another (Bauer and Claussen, 2006; Shindell et al., 2003; Zveryaev and Gulev, 2009). For instance, over the pre-industrial period, Shindell et al. (2003) showed that solar and volcanic forcings led to spatially and seasonally different climate responses in the Northern Hemisphere. When analyzing the average over the Northern Hemisphere in the CLIMBER model, Bauer and Claussen (2006) identified the changes in orbital parameters as responsible for an increase in seasonal differences in the temperature response over the past millennium, and the deforestation and variations in atmospheric carbon dioxide concentration as the main forcings responsible for a decrease in this difference over the past century.

None of the abovementioned studies has specifically investigated the Arctic climate. However, a deeper analysis of this region is justified by placing the rapid and large temperature variations observed in the Arctic during the last century in a wider context (e.g. McBean et al., 2005; Serreze and Francis, 2006). Additionally, a polar amplification of temperature changes is simulated in climate models driven by an increased radiative forcing, because of positive climate feedbacks involving, among other processes, albedo changes caused by the decrease in sea ice and snow coverages (e.g. Holland and Bitz, 2003; Serreze and Francis, 2006). An evaluation of model behaviour at the scale of the millennium thus appears to be of interest.

Multiproxy climate reconstructions are currently available for the Arctic region (Kaufman et al., 2009; Overpeck et al., 1997). The most recent one (Kaufman et al., 2009) consists of decadally resolved summer proxy temperature records covering the past 2000 years. According to this reconstruction, a long-term decreasing trend in summer Arctic temperatures occurs over this period, except for the last century, and is attributed to the steady reduction in summer insolation. However, because of the lack of widespread data and since most proxies do not reflect annual conditions but just the ones of the warmest months of the year, much less information is available on changes in the annual cycle

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through time. Climate model simulations are thus necessary, in complement to the proxy-based reconstructions, to help to confirm the proposed hypotheses and to improve the understanding of climate variations in this region, both for annual and seasonal means.

In this framework, the goal of this study is to document the differences in the Arctic temperature changes over the past millennium between the various seasons and to understand the causes of those differences. To do so, we analyze the annual and seasonal responses of the Arctic climate to natural and anthropogenic forcings such as solar, volcanic, astronomical, greenhouse gas and land use, in the Earth system model of intermediate complexity LOVECLIM (Goosse et al., 2010b). With a coarser spatial resolution and a simpler representation of the physical processes than in climate general circulation models, LOVECLIM has the advantage of being much faster than the latter and, consequently, of being affordable for performing the large ensembles of long simulations required here. These advantages inevitably come with some limitations, such as, for instance, a smoothed topography and a lack of a representation of stratospheric dynamics. However, the model is suitable for studying long-term climate changes at mid and high latitudes (Goosse et al., 2010b).

In this paper, first, a brief description of the model and forcings used is presented. The evolution of the simulated temperatures over the last millennium in response to the different forcings is then analyzed, a subsection being devoted to the contribution of each forcing. A final discussion of the results follows, including a brief discussion of the implications of our results for the calibration and interpretation of proxy data.

Model description and experimental design

The simulations analyzed here were conducted with the Earth system model of intermediate complexity LOVECLIM1.2 (Goosse et al., 2010b). This three-dimensional model includes representations of the atmosphere, the ocean and sea ice, the land surface and its vegetation, the carbon cycle and the polar ice sheets. However, the last two components are not activated in this study. The atmospheric component, ECBilt2 (Opsteegh et al., 1998), is a quasi-geostrophic model with a resolution of 5.6° in longitude and latitude and three vertical levels. The oceanic component, CLIO3 (Goosse and Fichefet, 1999), is a primitive-equation, free-surface ocean general circulation model, with a resolution of 3° in longitude and latitude, and 20 unevenly spaced vertical levels. It is coupled to a thermodynamic-dynamic sea ice model, where sea ice is assumed to behave as a two-dimensional viscous-plastic continuum for the computation of sea ice dynamics. Its representation of sensible heat storage and vertical heat conduction within the snow and ice are based on a three-layer model, and the energy budget at the bottom and top boundaries of the snow-ice cover and in leads determines the vertical and lateral growth and decay of sea ice (Fichefet and Morales Maqueda, 1997). The component representing the terrestrial vegetation is named VECODE (Brovkin et al., 2002) and simulates the annual evolution of trees, grassland and deserts, at the same resolution as ECBilt. The computed vegetation changes affect both the surface albedo, surface evaporation and water storage. LOVECLIM has been used successfully in many studies focused on recent, past or future climate changes at hemispheric and regional scales (e.g. Crespin et al., 2009; Driesschaert et al., 2007; Goosse et al., 2005, 2006; Renssen et al., 2005). More information about LOVECLIM is available at: http://www.climate.be/LOVECLIM.

All the simulations start at year AD 850 and end in AD 2000, following the experimental design of the third phase of the Paleoclimate Modelling Intercomparison Project (PMIP3) until the year AD 1850, and the fifth phase of the Coupled Model Intercomparison Project (CMIP5) afterwards. The initial conditions come from a 1000 year long, quasi-equilibrium run, using the greenhouse gas and astronomical forcings corresponding to AD 850. The simulations are driven by the forcings adopted by PMIP3 (v1.0), i.e. variations in solar irradiance, volcanic activity, orbital parameters, land use and greenhouse gas concentrations (Schmidt et al., 2011). The solar irradiance follows the reconstruction from Delaygue and Bard (2011) between AD 850 and 1609, and from Wang et al. (2005) between AD 1610 and 2000. The Earth's orbital parameters vary according to the calculations of Berger (1978). The forcing due to volcanic activity is derived from Crowley et al. (2008) and is implemented through anomalies in solar irradiance at the top of the atmosphere. The anthropogenic land use changes are based on the reconstruction of global agricultural areas and land cover of Pongratz et al. (2008) from AD 850 to 1700 and on the reconstruction of Ramankutty and Foley (1999) from AD 1700 onwards. This forcing is applied in LOVECLIM through a reduction in the area covered by trees and an increase in grassland since VECODE does not include a specific vegetation type corresponding to cropland. The evolutions of the concentration of the main greenhouse gases (CO_2, CH_4, and N_2O) are provided by Joos and Spahni (2008). For a detailed description of all these forcing reconstructions, see Schmidt et al. (2011). After AD 1850, the changes in sulfate aerosol load are taken into account through modifications in the surface albedo (Charlson et al., 1991), and the variations in tropospheric ozone concentration are included after AD 1950. In addition to the simulations including all those forcings, the contribution of each of them (with the exception of sulfate aerosol and ozone, because of our focus on the whole millennium) is evaluated in a set of experiments driven by one forcing at a time.

Each experiment set consists of an ensemble of ten simulations with identical forcing, in which the different members differ only in their initial conditions, with a small noise being added to the atmospheric streamfunction (as in Goosse et al., 2010a). The ensemble mean of these simulations provides an estimate of the response of the system to each forcing, as the influence of the natural variability simulated by the model, which differs in each member of the ensemble, is reduced by the averaging process. In our study, winter is taken as the months of January, February and March (JFM), spring as April, May and June (AMJ), summer as July, August and September (JAS), and autumn as October, November and December (OND). This choice is justified by the fact that, in the Arctic, spring starts later than at mid-latitudes, the maximum sea ice extent being observed for instance in February–March (Chapman and Walsh, 1993; Comiso and Nishio, 2008; Stroeve et al., 2007). Furthermore, this definition groups months with similar tendencies and thus gives more contrasted results between the seasons, as discussed in the next section. In the following analysis, the Arctic is defined as the region located north of 64°N.

Temperature response to different forcings

Response to greenhouse gas forcing

The climate response to changes in greenhouse gas concentrations is rather weak in the Arctic during the first centuries of the millennium (Figure 1a). In contrast, a rapid rise in surface temperature is simulated after AD 1850. The temperature difference due to the greenhouse gas forcing between the last and first decades of the 20th century for the Arctic region amounts to 1.7°C in our simulations. The corresponding value for the Northern Hemisphere is much lower (0.7°C), in accordance with the Arctic amplification of the warming.

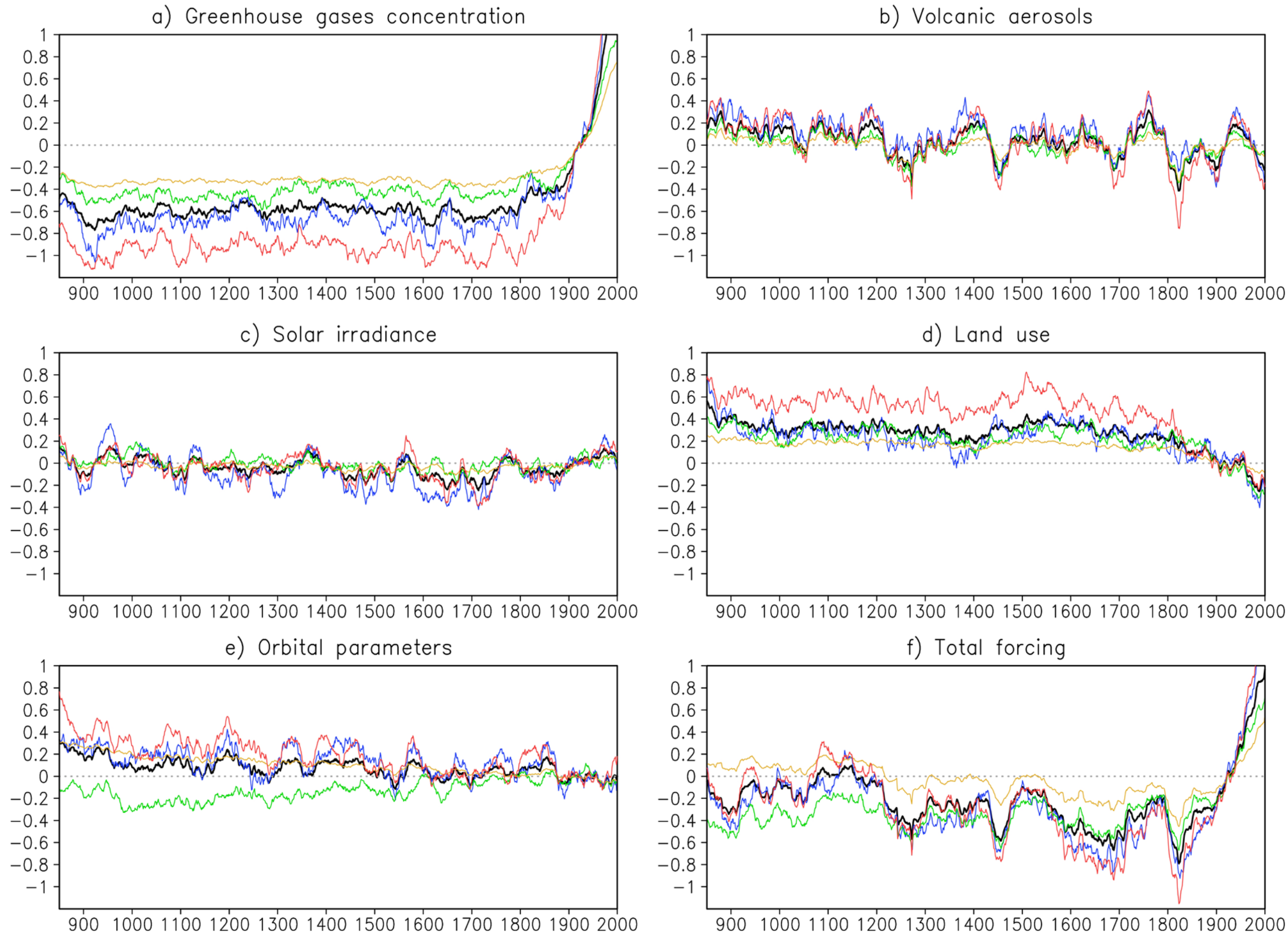

Figure 1. Anomaly in annual and seasonal mean surface temperature (°C) in the Arctic (north of 64°) over the last 1150 years as simulated by LOVECLIM in response to different forcings. Each time series represents the mean of an ensemble of ten simulations. The annual mean is displayed in black, the winter (JFM) in blue, the spring (AMJ) in green, the summer (JAS) in yellow, and the autumn (OND) in red. The reference period is AD 1850–1980. A 31year running mean has been applied to the time series. Colour figure available online.

The temperature increase varies considerably between the seasons. The maximum change occurs in autumn (2.5°C between the end and the beginning of the 20th century), and the smallest increase in summer (0.9°C). During autumn, the direct effects of the temperature–albedo feedback are relatively weak at high latitudes, because of the weak incoming solar radiation. It is likely the insulation effect of sea ice which is instead responsible for the larger temperature response to the forcing in this season compared with the others. The process leading to this has been observed and explained in other studies (e.g. Manabe et al., 1992; Vavrus et al., 2012). During summer, the amount of heat stored in the Arctic Ocean increases with the rise of greenhouse gas concentrations. This absorption of heat by the ocean is enhanced by the decrease in surface albedo resulting from the reduction in sea ice extent. However, the temperature variability and the response to the forcing are relatively low in summer, as temperatures remain mainly at the freezing point because of the melting of sea ice. In contrast, a decrease in the ice cover in summer has a large impact on the temperature in autumn and winter. During these seasons, the production of sea ice is slowed down because of the increased summer heat storage in the mixed layer of the ocean. The thinner and less extensive ice cover allows greater heat transfer from the ocean towards the cooler atmosphere and thus a large air temperature increase. This process appears valid in the response of LOVECLIM to each of the forcings, as shown in the following sections. This is confirmed by the changes in sea ice extent (average over area with at least 15% sea ice concentration), depicted in Figure 6 for the response to all forcings combined, which indicate a decrease in sea ice extent that is almost twice as large in summer as the other seasons between AD 1850 and 2000.

Response to volcanic forcing

As only a few major eruptions took place between AD 850 and 1200, the Arctic mean temperature is relatively stable during that period in our simulations driven by the volcanic forcing only (Figure 1b). For the last 800 years, a long-term cooling trend is observed, because of the higher frequency of eruptions, in addition to the abrupt temperature drops coinciding with the strong volcanic eruptions. Over the 1150 years of simulation, this cooling rate is equal to −0.016 ± 0.006°C per century (95% confidence interval for the trend).

A cooling is observed after a major eruption in all seasons. No significant seasonal contrast is observed over the last millennium in our simulations, except a larger temperature response in autumn than for other seasons. For instance, in some periods with intense volcanic activity, such as between AD 1745–1775 and AD 1810–1840, during autumn the temperature drops by up to 1.2°C (averaged over a 31 year period), while in summer the difference reaches only 0.45°C. The end of the last century also presents large seasonal differences with a drop in autumn and winter temperatures of almost 0.6°C, but with five times smaller changes in summer and spring. These seasonal differences cannot be explained by the seasonal variability of the forcing itself, since on average it is the strongest during winter and spring and the weakest in autumn. Here again, it is suggested that it is the insulation effect of sea ice which leads to this larger response, as explained in the previous section.

Response to solar forcing

Changes in irradiance due to variations in solar activity have little influence on the temperature evolution in the Arctic during the last millennium in our simulations. The temperature response (Figure 1c) is relatively weak, with a small long-term cooling trend until AD 1850 (−0.013 ± 0.006°C per century) and a small warming during the last 150 years (0.14 ± 0.08°C per century). Decadal-to-centennial fluctuations correspond roughly to positive and negative anomalies in solar activity, suggesting a simple, quasi-linear response to the forcing (after applying a 31 year running mean, the correlation between the solar forcing and the ensemble mean temperature response to this forcing in the Arctic is 0.57). These numbers should, however, be taken with caution. Because of the small signal, the standard deviation of the temperature response to solar forcing in the ensemble mean reaches 0.08°C, compared with 0.06°C in a control experiment without forcing. Ten ensemble members is too few in this case to precisely assess the contribution of solar forcing compared with a run without forcing, but we can confidently state that it is weak.

Whilst the Arctic receives little to no incoming solar radiation during winter, summer is characterized by high amount of incoming solar radiation because of the long period when the Sun is above the horizon. Nevertheless, as observed in Figure 1c, the seasonal responses to solar forcing do not display large differences. This weak seasonal contrast is due to the low absorption of solar radiation by the surface, even during summer (because of the high albedo of sea ice and snow) and to a memory effect related to sea ice, explained in section 'Response to greenhouse gas forcing': a summer warming induces a decrease in ice thickness, leading to larger oceanic heat fluxes towards the atmosphere in autumn and winter, and thus to a surface temperature increase during these seasons, although no direct effect of solar radiation is expected at high latitudes.

We must caution that the solar forcing selected in this study, and this is also the case for the other alternative solar reconstructions proposed by PMIP3 (v1.0, Schmidt et al., 2011), has a substantially smaller amplitude compared with some reconstructions used previously to drive models over the last millennium. This choice is justified from our present-day understanding of solar physics (e.g. Foukal et al., 2006). However, uncertainties remain large. Using an alternative reconstruction displaying larger variations, such as the one of Shapiro et al. (2011) included in those proposed in the version 1.1 of PMIP3 forcings (Schmidt et al., 2012), would lead to a more significant contribution of solar forcing to temperature changes during the past millennium. A new set of simulations would be required to estimate the influence of this forcing. However, we can infer from the quasi-linear behaviour of the temperature response to the solar forcing that this response will not change much qualitatively, but its magnitude would be much larger, as the Shapiro et al. (2011) reconstruction presents a TSI amplitude variance one order of magnitude larger than the reconstruction used in this study.

Response to land use changes

Surprisingly, the temperature evolution in the Arctic is strongly influenced by the deforestation taking place at lower latitudes. The land use forcing produces a significant cooling that reaches an annual mean of almost 0.6°C over the last four centuries (Figure 1d). This forcing leads to different magnitudes of temperature change for the various seasons. The largest cooling is observed in autumn, where it reaches almost 1°C since AD 1600. The cooling is substantially weaker during summer, reaching only 0.3°C.

This strong cooling is investigated in more detail by depicting the geographical distribution of the temperature response to deforestation (Figure 2). The land use changes at mid-latitudes lead to a cooling in the entire Arctic region in winter, spring and autumn, when comparing the periods AD 1950–2000 and AD 1550–1600. In summer, the signal is less clear, with still an overall cooling, but also a warming in some regions (Siberia and Canada). No significant change in either atmospheric or oceanic circulations is noticed in our simulations (not shown). Therefore, these temperature anomalies must be explained by radiative and thermodynamical effects rather than dynamical ones.

Deforestation has an impact both on the surface albedo and evaporation in LOVECLIM. The first effect induces a cooling, as the albedo of forests is lower than that of grass or crops. This difference in albedo becomes larger when snow covers the deforested areas. Moreover, the initial cooling associated with deforestation is responsible for a delayed melting of the snow, thus leading to an additional increase in surface albedo and a subsequent cooling (Figure 3b, c). The impact of changes in albedo is thus mostly visible during spring. As discussed above, the autumn and winter coolings are a consequence of the changes occurring during the other seasons (insulating effect of the sea ice), since little to no solar radiation reaches the surface during large parts of these seasons.

In summer, the reduced evapotranspiration, and hence the reduced surface latent heat flux (Figure 3d), due to the decreasing number of trees in some regions, warms up the surface locally. However, the temperature trend in the Arctic during this season remains negative over the last 400 years, because of the cooling during the other seasons. In the Arctic, the sea ice concentration increases by up to 10% in some regions in summer. The Arctic Ocean is thus more insulated from the cooler atmosphere, and the surface cooling is reinforced, mainly in autumn and winter, when the cooling in the centre of the Arctic is very large. This also increases the albedo and the amount of heat needed to melt the more extensive ice cover in summer. The net effect in this season is a cooling that overwhelms the influence of the slight warming at mid-latitudes because of the lower latent heat fluxes.

Response to astronomical forcing

The variations in the Earth's orbital parameters over the last millennium are associated with a 20-day shift in the perihelion, but also changes in eccentricity and obliquity (Berger et al., 1993; Schmidt et al., 2011). This forcing induces negligible changes at the hemispheric scale on an annual average, but its effect can be more important for specific months at particular latitudes (Bauer and Claussen, 2006). In our simulations, contrasted temperature trends for the different seasons are observed. Indeed, the temperature response to the astronomical forcing is characterized by a positive trend during the spring, contrary to the other seasons which display negative trends (Figure 1e). While the annual mean temperature in the Arctic decreases by about 0.15°C during the last millennium, the spring temperature experiences a rise of about 0.25°C.

Figure 4 displays the insolation difference at 75°N between AD 1900–2000 and AD 850–950 for the different months of the year along with the temperature difference in the Arctic for the same periods. During the first months of the year, an increase in insolation is observed, with the largest change occurring in April (1.5 W/m^2). The anomaly becomes negative after May, reaches its lowest value in July (−4 W/m^2) and remains negative throughout the entire summer season. The region north of 75°N receives no solar radiation during November, December and January. The temperature response follows the forcing, but with a time-lag of one to two months, reflecting the thermal inertia of the system. The spring months then exhibit a warming, which reaches a

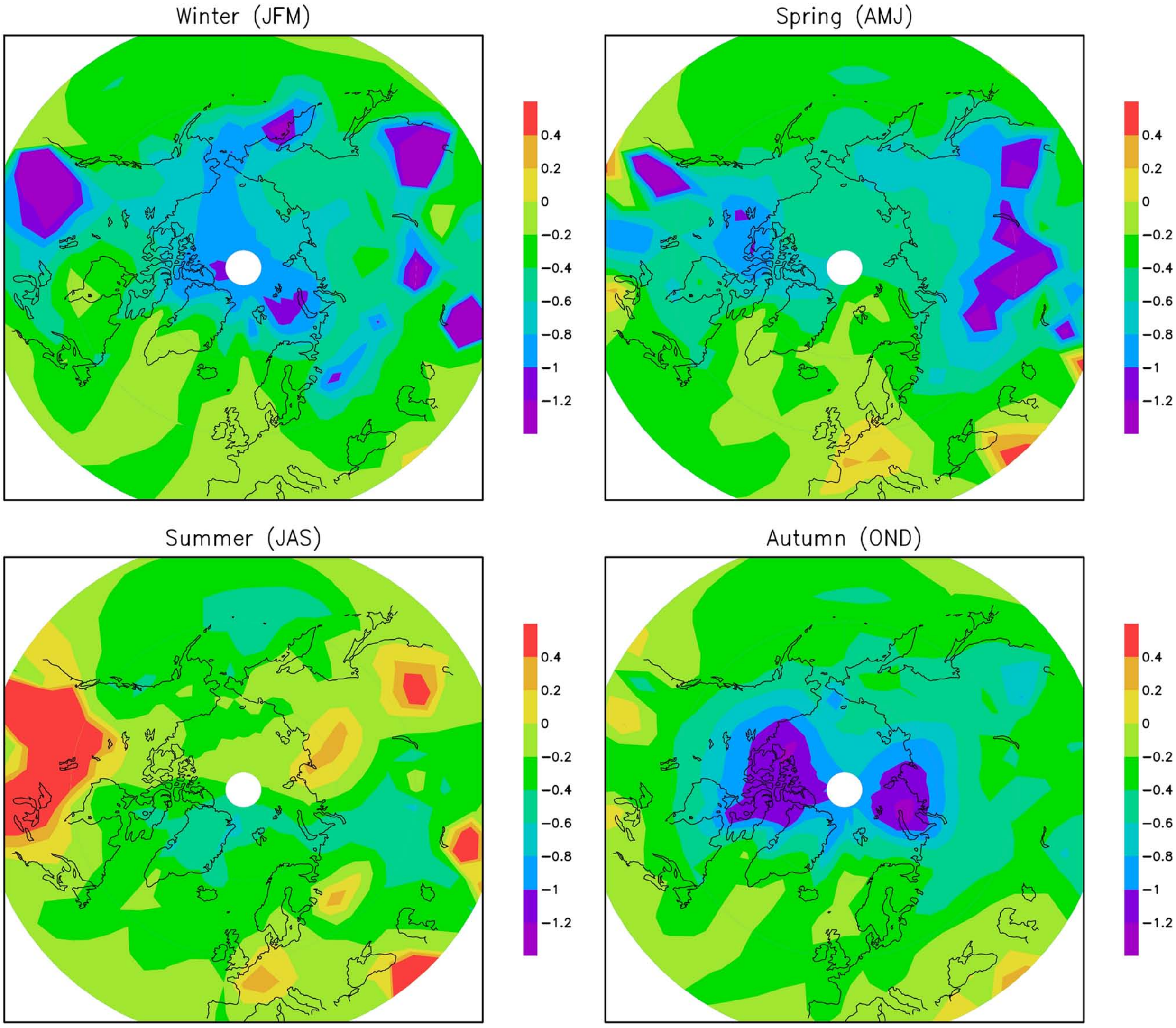

Figure 2. Surface temperature difference (°C) in the Arctic between AD 1950–2000 and 1550–1600 for the four seasons simulated by LOVECLIM, driven only by the land use forcing.

maximum in May. The rest of the year displays a cooling, with the highest negative temperature anomaly occurring at the beginning of the autumn. The temperature difference between the months of May and September reaches up to 0.8°C.

Bauer and Claussen (2006) also showed that the astronomical forcing plays a role in their climate simulations over the last millennium. However, they document a temperature difference of 0.1°C only between winter (DJF) and summer (JJA) for the land areas located between 30 and 70°N. For this region, LOVECLIM displays a similar result. This is smaller than the result for the Arctic region in our simulations and clearly indicates that the astronomical forcing plays a more important role over the last millennium in the Arctic than at lower latitudes. This is mostly due to the fact that modifications in obliquity have a stronger influence at high, compared with low, latitudes (Berger et al., 1993) and to the positive feedbacks amplifying the changes.

Response to all forcings

In response to all forcings combined, the annual mean Arctic surface temperature (Figure 1f) decreases slowly during the last millennium after a warm period around the 11th and 12th centuries, which is often referred to as the 'Medieval Warm Period'. The impact of large volcanic eruptions is clear during the mid-13th, mid-15th, late-17th and early-19th centuries. During the industrial period, the warming due to the increase in greenhouse gas concentrations is attenuated by the cooling effect resulting from land cover changes (and sulfate aerosol loads changes that are not studied here). The climatic response to all forcings corresponds more or less to the sum of the contributions of each individual forcing. Indeed, the RMSE between the response to all the forcings combined and the sum of the responses to each single forcing, after applying a 31 year running mean, is equal to 0.15°C. If we use this linearity, we can estimate the relative contribution of the different forcings to the cooling trend on an annual mean over the period AD 900–1850. It amounts to 35 ± 18% for the volcanic forcing, 28 ± 12% for the astronomical forcing, 27 ± 12% for the solar forcing and 20 ± 12% for the land use forcing, while the trend of the greenhouse gas forcing is positive.

The low-frequency temperature evolution in the Arctic over the last millennium has similarities with that simulated by many models at the hemispheric scale and in the different latitude bands of the Northern Hemisphere (e.g. Bauer and Claussen, 2006; Crowley, 2000; González-Rouco et al., 2006; Goosse et al., 2010a; Osborn and Briffa, 2006; Osborn et al., 2006). However, its amplitude is significantly higher in the Arctic region because of the existing feedbacks related to snow and sea ice, pointing to an Arctic amplification of climate changes. The seasonal contrast is also much more pronounced in the Arctic than at the hemispheric scale in our simulations. The astronomical forcing seems

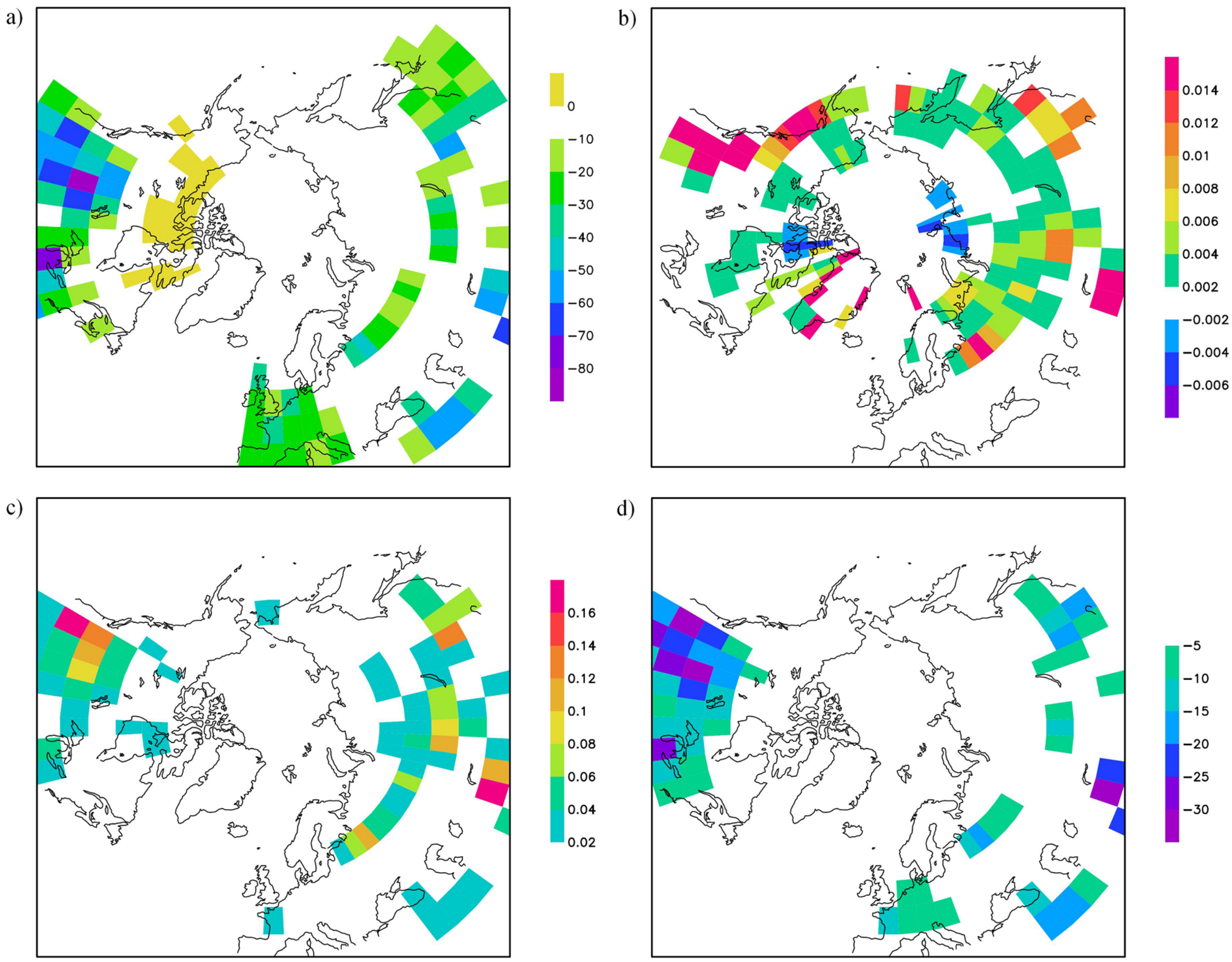

Figure 3. Difference in (a) annual mean tree fraction (%), (b) snow depth over land (m) in spring, (c) surface albedo in spring and (d) surface latent heat flux (W/m^2) in summer between AD 1950–2000 and AD 1550–1600 as simulated by LOVECLIM driven only by the land use forcing.

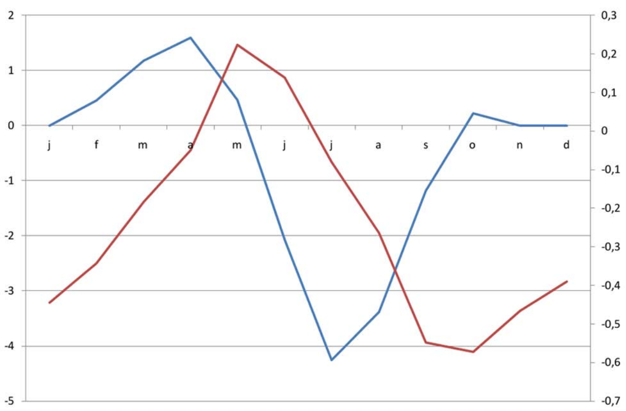

Figure 4. Insolation difference (W/m^2) at 75°N (blue line) and temperature difference (°C) in the Arctic as simulated by LOVECLIM driven only by the orbital forcing (red line) between AD 1900–2000 and AD 850–950. Colour figure available online.

to strongly contribute to the seasonal differences of the temperature evolution during the last millennium, as the opposing seasonal temperature trends (positive in spring and negative for the other seasons) observed in the response to the astronomical forcing are also simulated by the model when driven by all the forcings.

Kaufman et al. (2009) attributed the millennial-scale cooling in the Arctic to the reduction in summer (defined here as the mean of June, July and August) insolation due to the variations in orbital parameters. If we again estimate the contribution of the different forcings to the cooling trend in our simulations but now for summer (over the period AD 900–1850), we obtain a contribution of 22 ± 11% for the volcanic forcing, 57 ± 5% for the astronomical forcing, 12 ± 5% for the solar forcing, 15 ± 5% for the land use forcing and a positive trend for the greenhouse gas forcing. This confirms the role of orbital forcing proposed by Kaufman et al., but emphasizes that volcanic, land use and solar forcings also play a role in the cooling trend modelled by LOVECLIM. The time series of the model outputs and proxy-based data are depicted in Figure 5. One sees a relatively good agreement between them (even though the mean of June, July and August does not exhibit the largest cooling in our simulation, since the temperature trend in June is still positive). The low frequency variability of our simulation is lower than in the Kaufman et al. reconstruction. This might be related to the climate sensitivity of the model or to the forcing applied. Averaging over the ensemble might also play a role, as this reduces the multidecadal variability. On the other hand, the amplitude of the changes is strongly seasonally dependent and a small bias in the attribution of the signal of the proxy to a specific month in the Kaufman et al. reconstruction might also have a large impact on the model–data comparison. The reconstruction is, however, within the uncertainty range of the simulations, represented by two standard deviations of the ensemble,

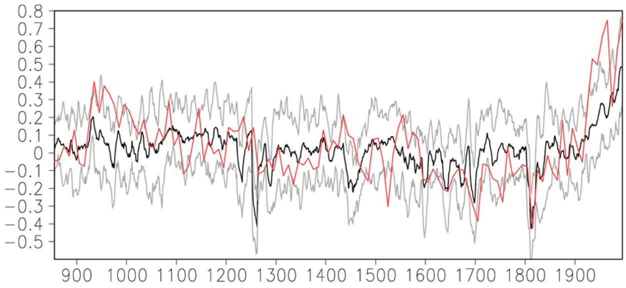

Figure 5. Anomaly in temperature (°C) averaged over the months of June, July and August in the Arctic over the last 1150 years. The black line is the mean of an ensemble of ten simulations using LOVECLIM driven by all the forcings. The grey lines are the mean plus and minus two standard deviations of the ensemble. The red line corresponds to the reconstruction of Kaufman et al. (2009). The reference period is AD 855–1855. An 11 year running mean has been applied to the model time series. Colour figure available online

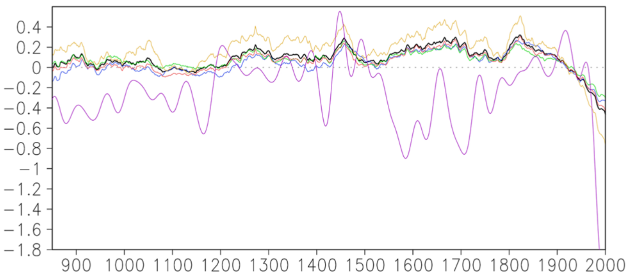

Figure 6. Anomaly in annual mean sea-ice extent (area with at least 15% sea ice concentration) in the Northern Hemisphere (10^6 km^2) over the last 1150 years as simulated by LOVECLIM in response to all forcings. Each time series represents the mean of an ensemble of ten simulations. The annual mean is displayed in black, the winter (JFM) in blue, the spring (AMJ) in green, the summer (JAS) in yellow, and the autumn (OND) in red. A 31 year running mean has been applied to the time series. The 40 year smoothed reconstructed August Arctic sea-ice extent from Kinnard et al. (2011) is displayed in purple. The reference period is AD 1850–1980. Colour figure available online.

with the exception of two particular periods. Between years AD 900 and 1000, the reconstruction shows a warming that is not simulated by the model. The model also fails in reproducing the mid-20th century warming (Goosse et al., 2010b). It thus appears that our experimental design satisfactorily captures the long-term trends but is not able to simulate these observed multidecadal fluctuations. The cause of this discrepancy will be investigated in a forthcoming study devoted to the origins of the warm periods in the Arctic.

In contrast, the comparison of our summer model results with the Kinnard et al. (2011) reconstruction of August Arctic sea ice extent is less satisfactory (Figure 6, model results are represented for summer but do not differ much from the August mean). The magnitude of the changes is much larger in the reconstruction throughout the whole period. The decrease in sea ice extent during the industrial period started earlier in the model than in the reconstruction and the decreasing trend is much larger in the latter. Compared with observations, the decline in summer sea ice extent in the Arctic is underestimated in LOVECLIM. The trends computed between AD 1979 and 2007 are equal to -0.056×10^6 km^2/yr in the observations and $-0.046 \pm 0.013 \times 10^6$ km^2/yr in the model (Goosse et al., 2010b). Note that the trend in the Kinnard et al. reconstruction during this very recent period is likely influenced by the use of the 40 year filter. Additionally, the reconstruction is characterized by a period with reduced sea ice extent during the late 16th and early 17th centuries (period with particularly low temperatures) which is not simulated by the model. However, before AD 1200, the reconstruction and the model agree on periods with relatively low sea ice extent and, between AD 1200 and 1450, an extensive sea ice extent is present in both of them.

Discussion and conclusions

This study aimed to improve our understanding of the evolution of the Arctic temperature during the last millennium using the Earth system model of intermediate complexity LOVECLIM. The modelled temperature response to external forcings agrees reasonably well with the reconstructed temperature of Kaufman et al. (2009). Volcanic, astronomical, greenhouse gas and, to a smaller extent, solar forcings all contribute to the simulated temperature changes over the 1150 years of simulation. More surprisingly, land use changes in mid latitudes also have a significant impact on Arctic temperatures. An Arctic amplification of the temperature changes is simulated in the responses to each of the forcings.

Our results show considerable differences between the four seasons. This seasonal contrast is mainly caused by the variations in orbital parameters of the Earth, which induce an increase of 1.5 W/m^2 in spring insolation and a decrease of 4 W/m^2 in summer insolation at 75°N during the past millennium. This astronomical forcing is larger than at lower latitudes and amplified more strongly in the Arctic by positive feedbacks involving snow and sea ice. This leads to larger differences between the seasons in that region compared with lower latitudes, with a positive long-term temperature anomaly trend during spring and a negative one during the three other seasons. During the 20th century, the larger autumn and winter warming trends are due to a stronger response to the variations in greenhouse gas concentrations during these seasons than in summer. The land use forcing has an opposite effect over this period: it leads to a larger cooling in spring, autumn and winter than in summer in the Arctic. In the study of Bauer and Claussen (2006), comparable results have been found for land areas from 30 to 70°N, except for the cooling obtained in response to land use forcing which is larger in summer than in winter in their study, in contrast to our results.

The contributions of the solar and volcanic forcings to the seasonal differences are relatively small for the past millennium in our simulations. Nevertheless, in addition to the direct radiative impact of volcanic eruptions or changes in solar irradiance, the dynamical response of the system can yield contrasting changes between the seasons in some regions, in particular over the mid-latitude continents. Indeed, the warming of the stratosphere (by absorption of both solar and terrestrial radiations) after a large volcanic eruption in the tropics is larger at low than high latitudes, leading to a strong meridional temperature gradient, especially during winter. The resulting changes in tropospheric circulation induce a winter warming over the continents, which overwhelms the direct radiative cooling effect of volcanic eruptions (Robock, 2000). Furthermore, the tropospheric circulation can also be affected by variations in solar irradiance, which influences the distribution of ozone in the stratosphere, affecting in turn its temperature and winds (e.g. Shindell et al., 2001). Because of the absence of a representation of the dynamics of the stratosphere in LOVECLIM, these effects can not be studied here (Goosse and Renssen, 2004). Since the radiative scheme used in the model is very simple, it was also not possible to include solar spectral irradiance variations (Schmidt et al., 2012), which also have a clear impact on stratospheric dynamics. These constitute limitations to our study that must be kept in mind when interpreting our results. Finally, our results clearly depend on the choice of forcings. Alternative reconstructions (such as Shapiro et al., 2011, for the solar forcing and Kaplan et al., 2011, for the land use forcing) could lead to different results, although we do not expect that they would qualitatively affect our conclusions. The response to the alternative vegetation reconstruction is particularly difficult to assess, as the Kaplan et al. reconstruction is based on a very different assumption from the reconstruction used in this study. However, we expect a noticeable impact in the Arctic of land use changes at lower latitudes.

The significant seasonal contrast in trends underlined by our simulations may have consequences for the interpretation of the reconstructions of past temperature based on proxy data. Indeed, our study indicates that some seasons are less representative of annual conditions than others. Because many proxies record changes during a specific part of the year or season (e.g. Jones and Mann, 2004), the calibration against annual temperatures may thus be biased (Briffa and Osborn, 2002; Jones et al., 2003, 2009). For the 20th century, the trends have the same sign for all seasons and the correlation between proxy records and instrumental observations may be relatively similar for all of them. Unfortunately, looking at our model simulations, it is apparent that seasonal differences have not been stationary through the past millennium, with seasonal contrast being larger at the beginning than the end of the millennium. If the modelled temperature curves are scaled over the warming of the past 100 years, the difference between summer and spring temperature anomalies at the beginning of the millennium amounts to 0.6°C in our experiments. This points out the need to carefully determine the season that most influences the proxies, as the May–June signal, for instance, is clearly different from that of July–August.

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