Holocene temperature trends in the extratropical Northern Hemisphere based on inter-model comparisons

Yurui Zhang [1 & 2], Hans Renssen [2 & 3], Heikki Seppä [1], Paul J. Valdes [4]

[1] Department of Geosciences and Geography, University of Helsinki, P.O.BOX 64, FI00014 Helsinki, Finland

[2] Department of Earth Sciences, VU University Amsterdam, De Boelelaan 1085, 1081 HV Amsterdam, The Netherlands

[3] Department of Natural Sciences and Environmental Studies, University College of Southeast Norway, 3800 Bø i Telemark, Norway

[4] School of Geographical Sciences, University of Bristol, Bristol BS8 1SS, UK

Correspondence to Yurui Zhang (yurui.zhang@helsinki.fi)

Abstract

Large uncertainties exist in Holocene climate estimates, especially during the early Holocene when large-scale reorganization occurred in the climate system. To improve our understanding on these uncertainties, we compare four Holocene simulations performed with the LOVECLIM, CCSM3, HadCM3, and FAMOUS climate models. The simulations are generally consistent on large-scale Northern Hemisphere extratropics, while the multi-simulation consistencies are heterogeneous on sub-continental scale. Consistently simulated temperature trends are found in Greenland, N Canada, NE and NW Europe and central-west Siberia. These Holocene temperatures show a pattern of an early-Holocene warming, mid-Holocene warmth and gradual decrease toward the pre-industrial in winter, and the extent of early-Holocene warming varies spatially, with 9°C warming in N Canada compared with 3°C warming in central-west Siberia. In contrast, mismatched temperatures are detected: in Alaska, the warm early-Holocene winter in LOVECLIM primarily results from strongly enhanced southerly winds induced by the ice sheets; in E Siberia, the intense summer warmth in CCSM3 is caused by large negative albedo anomalies due to overestimated snow cover at 0 ka; in the Arctic, cool winter temperatures in FAMOUS can be attributed to too extensive sea ice coverage probably due to simplified sea-ice representations. Thus the Holocene temperature trends in these regions remain inconclusive.

Keywords Holocene temperature, inter-model comparisons, extratropical Northern Hemisphere, the ice sheets and meltwater, climate sensitivity

1. Introduction

The Holocene is an important period for investigating climate variability and improving our understanding of the climate system due to the detectable changes in climate variables and the abundance of available proxy records. The early Holocene (11.5–7 ka) was a transitional phase, accompanied by reorganizations in several components of the climate system (e.g. CAPE project, 2001; Dyke et al., 2003; Shakun et al., 2012). Investigating this transient period can increase the knowledge of climate system's variabilities, in addition to provide further information on the climate history. Climate modelling is a useful tool to investigate climate change, and thus have a potential to improve our understanding of the indeterminate early-Holocene climate, especially by conducting transient simulations. Recent model simulations have either included or specifically focused on this critical time phase to investigate the climate responses to dynamic climate forcings (He, 2011; Zhang et al., 2016). However, uncertainties related to the melting of the ice sheets adds challenge in accurately simulating the climate of this period. The ice-sheet related uncertainty during the early Holocene has been tested by Zhang et al. (2016), who investigated two different freshwater scenarios in model simulations to identify a more reasonable scenarios of the early-Holocene cooling and the timing of HTM.

Uncertainties in the early-Holocene simulations are not only caused by ambiguity in climate forcings, but can also result from model-dependent variations. The performance of models and their sensitivities to given forcings are primarily determined by the physical representations of various climate processes in the models. For instance, climate sensitivity to radiative forcings (CSr) usually refers to the change in the global annual mean surface air temperature (in °C) experienced by the climate system after it has attained a new equilibrium in response to a doubling of the atmospheric CO_2 concentration (Knutti and Hegerl, 2008). Estimates of CSr vary from 2.1 to 4.7°C among the models in the Coupled Model Inter-comparison Project (CMIP5), due to various feedbacks involved in the models (Randall et al., 2007; Flato et al. 2013). In addition, it is found that CSr is greater for cold than for warm climates, and the sensitivities to changes of climate condition differ among individual studies owing to non-uniform representations of cryospheric processes or dynamical ocean feedback processes (Boer and Yu, 2003; Randall et al., 2007; Kutzbach et al., 2013). Variations among models in the estimates of key climate process or parameters, such as the climate-state-dependent CSr, add extra difficulties to simulate the early-Holocene climate.

Multi-model comparison provides an option to validate the reliability of model performances and hence to increase the confidence in simulations. Reliability of and confidence in simulations increases if similar features are observed in other independent model results; conversely, further investigations are required when multi-model simulations differ. The Paleoclimate Modeling Inter-Comparison Project (PMIP) has performed a wide range of inter-model comparisons, covering a series of periods, such as the Last Glacial Maximum (LGM, 21 ka), the Last Interglacial (LIG, 130–115 ka) and the mid-Holocene (6 ka) (e.g. Harrison et al., 1998; Braconnot et al., 2000; Lunt et al., 2013). However, the PMIP comparisons for the Holocene primarily focus on the mid-Holocene with snapshot experiments, and previous transient inter-model comparisons have only been conducted for periods shorter than the whole Holocene, such as 8–2 ka and the last millennium(e.g. Bakker et al., 2014) to avoid the uncertainty related to impact of ice sheet on the climate. Therefore, the question of how similar or different the model simulations are during the early Holocene is still unknown, and inter-model comparisons spanning to the entire Holocene are greatly demanded.

Here, we evaluate the robustness of four different simulations covering the whole Holocene in detail by conducting inter-model comparisons and analyzing their uncertainties. These models include LOVECLIM (Zhang et al. 2016), CCSM3 (He, 2011), HadCM3 (Singarayer & Valdes, 2010), and FAMOUS (this study), and they differ in multiple critical aspects, as summarized in Table 1. Our goal is to analyze how the climate responds to dominant forcings in these different models and to evaluate how robust these responses are through comparisons of multiple Holocene transient simulations. In particular, our comparisons aim to: 1) identify the agreements and divergences among these Holocene simulations, 2) detect which aspects of the climate system in simulations cause these multi-model variations, and 3) further examine the potential origin of these inconsistencies, such as different parameterizations and biases in the models. A detailed data-model comparison is, however, beyond the scope of the present study, and will be the topic of a future publication.

2. Models and simulations

2.1 Models and prescribed climate forcings

The climate models employed in this study include the LOVECLM, FAMOUS, CCSM3 and HadCM3, which have various resolutions and complexities. Key information on these model is summarized in Table S1 and more detailed information can be found in SI. We included the major climatic forcings in terms of insolation variation on orbital scale (ORB), greenhouse gases (GHG) (Fig. S1) in the atmosphere and decaying ice sheets (ice-sheet configuration (ISC) and associated freshwater fluxes (FWF)). Other forcings, such as the solar constant and aerosol levels, were kept fixed at preindustrial values in all simulations. More detailed information on the climate forcings are given in SI.

2.2 Setup of simulations

The LOVECLIM simulation is an 11.5 kyr long transient run that was initialized from an equilibrium experiment (more details are provided in Zhang et al., 2016). The simulation was forced by the annually-varied ORB and GHG throughout the whole period. In addition, the prescribed ISC was included with a time step of 250 yr until 6.8 ka when the ice sheet eventually disappeared, and associated FWF was also applied with a stepwise time series (Fig. 1). The CCSM3 simulation was taken from the TraCE-21ka project, which is a 21-kyr long simulation forced by transient ORB, GHG, ISC and freshwater forcings. The ISC were modified every 500 yr based on the ICE-5G reconstructions. Freshwater was discharged with irregular time steps until 6 ka (He 2011). The FAMOUS simulation is also the Holocene part of a 21-kyr simulation that was forced by the transient GHG and ORB forcings, with prescribed ISC and freshwater from the melting of ice sheets. The HadCM3 results were derived from two sets of snapshot experiments at 1-kyr temporal intervals; and the GHG, ORB and ice sheet forcings were updated in each snapshot. The main reason for including these experiments is that the high spatial resolution of the HadCM3 model allows a direct test of the impact of resolution and complexity, although differences between transient and equilibrium adds complexity in the inter-comparison.. The differences between these two sets of experiments are the ice sheet configurations that were based on ICE-I5G and ICE-I6G respectively, and they were accordingly named as HadCM3-I5G and HadCM3-I6G. Given the similarity between HadCM3-I5G and HadCM3-I6G, they are generally considered as a HadCM3 simulation unless specifically indicated. Each of these snapshot experiments was initialized from a spun-up pre-industrial simulations (Singarayer and Valdes, 2009) and was run for at least 300 yr with fixed GHG and ORB forcings, of which the last 30yr average was taken as representative of the climate during the corresponding time window. Main information on these simulations are summarized in Table 1.

The model name is also used as the indicator of corresponding simulation to reduce redundancies. Temperatures presented are simulated surface temperatures, and shown as the deviations from 0 ka (the pre-industrial). To obtain the overall temperature trend throughout the Holocene, the ensemble mean was calculated by averaging all transient simulations. This implies that the HadCM3 results were not included in the ensemble and are separately shown in figures.

3. Results

3.1 Simulated temperature in the NH extratropics

Although HadCM3 and FAMOUS show a slightly cooler climate until 5 ka than CCSM3 and LOVECLIM, the simulated summer temperature trends in the NH extratropics (30–90°N) are generally consistent across the models (Fig. 2a). They reveal an early-Holocene warming, a maximum temperature at 10~7 ka, and a cooling toward 0 ka. In summer, the magnitudes of early-Holocene anomalies in these models roughly correlate to their climate sensitivities, as high sensitivity implies a large climate response to given forcings (Fig. 2b, Tab. S1). The simulated annual mean temperature shows a gradual warming of about 2°C by 4 ka, after which the temperature stays at the 0 ka level (Fig. 2b). There is a 1°C spread in simulated annual mean temperature before 5 ka, with the coolest climate in FAMOUS and warmest in LOVECLIM. An abrupt cooling at ~8.5 ka is found both in the CCSM3 and FAMOUS simulations.

In order to approximately evaluate the reliability of the simulations, the simulated results are briefly compared with proxy data. The proxy data are stacked temperatures using proxy records from the 30–90°N latitudes band (Marcott et al., 2013). This model-data comparison reveals an overall agreement, despite a slightly stronger early-Holocene warming until 9.5 ka suggested by the proxy data. In particular, the simulated summer temperatures show better agreement with proxy data than annual mean temperatures, which is consistent with the suggestion of potential seasonal biases in the biological proxy data (Lohmann et al., 2013; Liu et al., 2014).

Strong spatial patterns of simulated temperatures are found when zoomed into regional scale. To further illustrate the regional climate response to relevant forcings, two periods, 11.5 ka and 6 ka, were selected as specific time windows. These two time windows either represent the period when the ice sheets played important roles, or serve as a benchmarking epoch (*i.e.* in PMIPs). Simulated temperatures generally show negative anomalies at 11.5 ka, with magnitude of -1 to -5°C in annual mean temperatures, except in Beringia where positive temperature anomalies are found in LOVECLIM and CCSM3(Fig. 3). The temperatures at 6 ka show a latitudinal pattern. At the mid-latitudes, simulations suggest similar or slightly lower temperature than at 0 ka. At the high-latitudes, annual mean temperatures show model dependencies, with 1 to 3°C warmer in CCSM3 and LOVECLIM contrast with similar or -0.5°C cooler climate in FAMOUS and HadCM3. This spatial pattern of simulated temperature seems agree with previous data-model comparison in N Europe (Brewer et al., 2007). The model dependencies might imply model biases, which has been demonstrated by Jiang et al. (2012) who found that climate models under-estimated the Mid-Holocene climate in China. In addition, the differences between the present and previous studies can be partly explained by the snapshot or transient simulations. Another significant feature of climate during the early Holocene is that the multi-model consistencies are regionally heterogeneous and generally larger in winter than in summer. Target regions were further selected according to the spatial pattern of climate response to dominant forcings (Fig. S2), and the Holocene temperature trends over these regions were analyzed. Based on their consistencies, these selected regions were divided into two groups, representing consistent and inconsistent areas. The first group is formed by Greenland, N Canada, NE Europe, NW Europe and central-west Siberia, as all models give similar early-Holocene temperature trends. The second group with mismatched temperatures includes Alaska, Arctic and E Siberia, as opposite temperature trends (especially in winter) across the models are found.

3.2 Temperature over the regions with good inter-model agreements

The simulations show overall good agreements over N Canada, NW Europe, NE Europe, Greenland and central-west Siberia. The ensemble mean temperatures generally rise from the cold initial state till 6~7 ka, followed by gradual decrease to the 0 ka level, with exception of summer temperature in NE Europe and central-west Siberia. As indicated by the ensemble mean, the magnitudes of this cool early Holocene vary among regions (Fig. 4). In particular, a considerably cool early Holocene climate is found in N Canada, with 5°C lower ensemble mean in summer and 10°C lower in winter; whereas only a minor cooling is present in NE Europe and central-west Siberia, with about 4°C cooling in winter and 1–3°C warming in summer. Temperatures in Greenland and NW Europe show intermediate values, with a cooling of 2–3°C in summer and around 8°C in winter.

Only minor inter-model variations are found in Greenland and NE Europe within the range of 3°C during the early Holocene. In Greenland, all simulations indicate about 3°C cooler summer conditions at 11.5 ka in comparison with 0 ka, followed by a rise to 1°C at around 7.5 ka. In winter, all simulations suggest low temperature at 11.5 ka with about -8°C anomaly of the ensemble mean temperature, notwithstanding the temperature spread between individual simulations up to 5°C. In NE Europe, the simulations suggest a warming of 1°C in summer and cooling of 4°C in winter at 11.5 ka, with a 2°C multi-simulation spread until 9 ka. From 11.5 ka onward, simulated winter temperatures slowly rise to the preindustrial level, while in summer the temperature anomalies show a gradual decrease of about 1°C. Over central-west Siberia, the temperatures in winter show 3–4°C warming trend during the Holocene, while the simulations indicate 3°C cooling summer. Within the frame of overall consistent trends, slightly different temperatures are found in N Canada and NW Europe during the winter. In N Canada, the small early-Holocene temperature anomalies in LOVECLIM might be related to the use of a fixed modern land-sea mask throughout the simulation, as which implies underestimation of the early-Holocene albedo over the Hudson Bay. The jump of simulated winter temperature in FAMOUS at 8 ka might be related to the spike in FWF forcing (Fig.1). In NW Europe, the magnitudes of the cooling generally increase in the order of LOVECLIM, CCSM3, HadCM3 and FAMOUS. One exception is that winter temperature anomalies in HadCM3 and FAMOUS rapidly rise to a distinct peak of +2°C at around 9~10 ka, leading to high temperatures at round 9 ka. The warm peaks in FAMOUS and HadCM3 are mainly caused by a response of sea ice to the opening of the Bering Strait (Fig. S4). Overall, these roughly consistent patterns demonstrate that over these sub-regions, forced millennial climate change exceeds internal climate variability during the early Holocene.

3.3 Temperatures over the regions with less multi-model consistency

In Alaska, Arctic, and E Siberia, simulated temperatures show poor consistency (Fig. 5), especially in winter when both positive and negative early-Holocene anomalies are simulated by the different models. In Alaska, the spread of the simulated winter temperatures at 11.5 ka ranges from 2°C warmer in LOVECLIM to 4–6°C cooler in FAMOUS and HadCM3 in comparison with 0 ka. This distinct multi-model variation in winter is thus up to 8°C, which is considerably larger than in summer when the inter-model variation is only 1°C. Over the Arctic, the discrepancies between the simulations also primarily exist in winter when they are up to 8°C. At 11.5 ka, the winter temperature anomaly is slightly above 0°C in LOVECLIM, while more than 8°C cooling is produced in FAMOUS. Nevertheless, the ensemble mean temperature suggests 1°C cooling in summer and 4°C warming in winter throughout the Holocene. Relatively large multi-simulation differences are found over E Siberia in both summer and winter, reaching up to 3°C at the onset of the Holocene. The simulations show decreases in summer temperatures over E Siberia throughout the Holocene, with the largest decrease (more than 4°C) in CCSM3. This large variation in summer is primarily caused by exceedingly warm early-Holocene conditions in CCSM3. In winter, over 2°C cooling is simulated by LOVECLIM during the Holocene contrasting with up to 5°C warming in FAMOUS. The ensemble mean declines by almost 4°C in summer, but generally rises by 2°C in winter despite a small drop at ~8.5 ka.

In general, although the simulations generally agree on temperatures over the large-scale NH extratropics, some regions show better inter-model agreements than others when zooming into the regional scale. The mismatches among the simulations are large during the early Holocene, and can be outlined as follows: 1) warm winter climate over Alaska in LOVECLIM in comparison with various degrees of cooling in other models; 2) large negative temperature deviations (from 0 ka) over the Arctic in FAMOUS contrasting with slightly positive values in LOVECLIM; 3) a stronger summer warming over E Siberia in CCSM3 than in other simulations, and a warm winter in LOVECLIM over E Siberia contrasting with the cool climate in HadCM3.

4. Discussion

Comparisons of multi-model simulations provide an opportunity to evaluate the performance of climate models in simulating climate response to radiative forcings and other boundary conditions. The following discussion will start from the above results with a focus on the regions where the simulated temperatures are different, and the causes of these mismatches will be investigated at two levels. We will firstly try to identify the direct causes of these inter-model discrepancies via a diagnosis of various climate variables. Subsequently, the potential origin of these divergent climate variables will be examined.

4.1 Divergent climate variables lead to mismatched temperatures

4.1.1 Mismatched Alaskan winter temperature

The relatively warm early Holocene in LOVECLIM results from enhanced southerly winds induced by the LIS which bring warm air from the south. This enhanced southerly winds can be diagnosed by examining the anomalous atmospheric circulation over the ice sheets at 11.5 ka, in comparison with the ice free condition at 0 ka. The atmospheric circulations are indicated through geopotential height fields, which reflect anomalous geopotential to standard gravity at mean sea level, with high values representing high pressure near the surface. Although there are largely similar responses of geopotential heights to the existence of the LIS, such as enhanced values over the LIS, the magnitude of these anomalies differs across individual models (Fig. 6). At 11.5 ka, winter geopotential heights in LOVECLIM and FAMOUS are up to 50 gpm higher over the center of the LIS than at 0 ka, which is larger than the enhancement of 30 gpm in HadCM3. The LOVECLIM and HadCM3 simulations have similar spatial patterns of Northern Annular Mode with a lower value over the polar region than in the surrounding areas, but with larger anomalies of geopotential height in LOVECLIM than in HadCM3. In order to further analyse this anomalous geopotential height, the geopotential height differences between the LIS (50–75°N, 65–110°W) and the N Pacific (35–45°N, 120–180°W) were calculated. These results are further standardized to changes relative to the 0 ka condition. This standardized difference of geopotential height in LOVECLIM was up to 70% at 11.5 ka, which declines with time toward 0 ka (Fig. S3). The FAMOUS and HadCM3 also show a similar decreasing trend in geopotential height changes during the Holocene, but with a smaller magnitude than in LOVECLIM. A similarly anomalous atmospheric circulation, with a comparable amplitude of changes as in HadCM3, is indicated by the surface pressure anomaly in CCSM3 (figure is not shown). Therefore, these different anomalies in geopotential height fields among individual models can lead to divergent climate over the marginal regions of the ice sheet, over which this differential signal is weak and even a minor divergence is visible. In Alaska (to the west of the LIS), the intense warm climate during the early Holocene in LOVECLIM is primarily caused by the strong gradients of geopotential height that induce southerly winds, bring warm air from the south and increase the local temperature in Alaska. This effect lasted until the final disappearance of the LIS at 6.8 ka, after which the small decreasing trend in temperature can be associated with a sea-ice cooling effect on coastal regions of northern Alaska (Fig. 7). Compared with the HadCM3 simulation, lower temperatures in FAMOUS can be explained by more extended sea ice in this model (Jones et al., 2005), and a stronger anticyclone over Alaska (Fig. 6).

Previous studies have analysed the effect of ice sheet on atmospheric circulation under the LGM condition. The experiment performed by the Polar MM5 atmospheric model has shown that 500-hPa geopotential height over the LIS increased 260 gpm in January and 70 gpm in summer (Bromwich et al., 2004; 2005). The intensified atmospheric baroclinicity induced by the ice sheets is suggested as one of the primary mechanisms behind the atmospheric circulation changes under the LGM boundary conditions (Bromwich et al., 2005). Another explanation for these effects is that the air flow can be deflected or split around an anticyclone over the LIS (e.g. Bromwich et al., 2004). Sensitivity experiments of the atmospheric circulation response to idealized circular mountains reveal that the orography effects of ice sheet is highly depended on the scale of the ice sheet (Yu and Hartmann, 1995). Given the scale of the ice sheets in our experiments, the ice-sheet effects on atmospheric circulation during the early Holocene are much smaller in comparison with the LGM, as shown in our simulations.

4.1.2 Mismatched winter temperature over Arctic

Over the Arctic, the inter-model divergences of winter temperature during the early Holocene are associated with the different sea ice across the individual simulations. Although it is difficult to establish if the sea-ice changes are a cause or effect of Arctic temperatures, sea ice plays a critical role in the climate system through sea-ice related feedbacks. Firstly, the areal extent of sea ice determines to which extend the albedo-related feedback occurs, because the albedo of sea ice is typically up to 0.5~0.6 and significantly higher than in open ocean at the high latitudes. Furthermore, the thickness of sea ice influences the amount of heat that is released from the relatively warm ocean to the cold atmosphere (Renssen et al., 2005; Holland et al., 2006). In these four simulations, the early-Holocene sea ice anomaly is overall positive compared with 0 ka (Fig. 7). However, the absolute sea ice area and the magnitude of anomalous sea-ice cover (between 11.5 ka and 0 ka) varies with the models. At 11.5 ka, the simulated NH sea ice in March is in the order of 10^{12} m^2 and is decreasing by the order of FAMOUS, CCSM3, HadCM3 and LOVECLIM. The largest change of NH sea ice in FAMOUS is up to 32×10^{12} m^2, which is two times larger than the minimum change in HadCM3 and LOVECLIM (Fig. S4). The magnitudes of anomalous sea ice area follow the same order as in the absolute sea ice cover, with the largest anomaly in FAMOUS (Fig. S4). It is well-known that accurately simulating sea ice in coupled climate models is challenging because of the high complexity of sea ice in both spatial and temporal dimensions, which can also be seen from the wide spread of simulated sea ice even under pre-industrial conditions. At 0 ka, the NH sea-ice area in these models varies from 13.5 $\times 10^{12}$ m^2 to 24×10^{12} m^2 (Fig. S4). Studies have suggested that the NH sea ice was overestimated in CCSM3, FAMOUS and HadCM3 with different degrees and spatial patterns (Gordon et al., 2000; Jones et al., 2005; Bryan et al., 2006). For instance, CCSM3 overestimated sea ice in the Labrador Sea, while HadCM3 and FAMOUS simulated more sea ice in the Barents Sea (Gordon et al., 2000; Jones et al., 2005; Bryan et al., 2006). Compared with these total NH sea ice areas, simulating a consistent anomalous spatial distribution of sea ice is even more difficult, as revealed by various sea ice patterns across these simulations. In particular, thickness anomalies of sea ice over the Greenland Sea in HadCM3 are larger than in LOVECLIM. Therefore, the cooler climate in HadCM3 than in LOVECLIM is partially related to the insulation of thick sea ice in HadCM3 at 11.5 ka. The simulated strong winter cooling at 11.5 ka in FAMOUS could be related to enhanced albedo feedback due to extensive sea ice coverage. It is also noticeable that considerable changes of NH sea ice area in FAMOUS occurred between 9 and 8 ka (Fig. S4), which results from the opening of the Bering Strait and explains why temperature substantially changed around that time. The AMOC is another associated factor impacting temperatures in the Arctic, especially contributing to climate changes associated with millennial scale events, with more detailed discussion on spatial (Fig S5) and Holocene trend (Fig. S6) are presented in SI.

A brief comparison with proxy-based sea ice reconstructions broadly supports this simulated extended sea ice extent during the early Holocene. Proxy-based sea-ice reconstructions show an overall decreased tendency in sea ice extent throughout the Holocene, such as in the circum Arctic areas. The magnitudes of sea ice extension are spatially heterogeneous (de Vernal et al., 2013). In particular, dinocyst assemblages suggest positively anomalous early-Holocene sea ice over the regions which were connected to Atlantic, such as the Labrador Sea and Greenland Sea (de Vernal et al., 2013), which roughly agrees with larger early-Holocene sea ice cover in the simulations.

4.1.3 Mismatched temperature in east (E) Siberia

As a main contributor to mismatches in multi-simulation temperatures, the intense early-Holocene warmth in CCSM3 primarily results from its large negative albedo anomaly (more than -0.2 compared to 0 ka) in E Siberia. The climate is known to be highly sensitive to surface albedo changes (Romanova et al., 2006), and thus inter-model variations in albedo could cause differences in simulated temperatures. The surface albedo anomalies between the early Holocene and the preindustrial are spatially heterogeneous. On the one hand, the anomalous albedos over the ice sheets are roughly similar. For instance, an enhanced surface albedo of up to 0.6 at 11.5 ka over the LIS is consistently found in all simulations (Fig. 8), despite a small exception over the Hudson Bay in LOVECLIM due to the fixed modern land-sea mask. On the other hand, large inter-model albedo anomalies are found over the regions where the surface albedo is primarily influenced by vegetation, snow cover and sea-ice cover.

In E Siberia, summer surface albedo in CCSM3 differs from other models. In CCSM3, overall Holocene albedo values are higher than 0.36, and they show a rising trend during the Holocene with a rapid increase at 3 ka; whereas other models suggest a stable Holocene trend of summer albedo with absolute values ranging from 0.15 to 0.2 (Fig. S7). This increasing Holocene albedo in CCSM3 is anti-correlated with a decreasing temperature trend, and it is clear that the 2°C decline around at 3 ka is related to an albedo rise. This negative anomalies and increased trend of simulated temperature in CCSM3 are mainly due to high albedo at 0 ka, which is up to 0.65. A further investigation on the snow cover in the simulation shows that most of E Siberia is covered by snow during the summer season, which is obviously an overestimation, as it would imply the inception of a continental ice sheet. By contrast, other simulations show low albedo (around 0.2), indicating a vegetation-covered surface, which is more realistic given the represent-day landscape. In winter, the spread of inter-model temperatures over E Siberia result from multiple factors. The different albedo response related to snow cover can partly explain the spread across the simulations, with relatively warm climate in LOVECLIM corresponding to low albedo and low temperature in FAMOUS associated with high albedo. Moreover, sea ice changes influence the temperature of the coastal Siberia. For example, the temperature bump at around 9 ka in FAMOUS and a mild increase in LOVECLIM are caused by temporal variations of sea ice (Fig. S4). Additionally, the warm winter in LOVECLIM is partially associated with the enhanced southerly winds mentioned previously.

4.2 Potential sources contributing to inter-model divergences of climate variables

4.2.1 Uncertainty of ice-sheet-related forcing

With decaying ice sheets in North America and Fennoscandia, the uncertainty in the FWF forcings during the early Holocene is mainly related to the total volume of ice melts involved, the location of discharge, and the timing of discharge, which further impacts the Holocene simulations. Different FWF-forcing scenarios and associated different AMOC responses have climate impacts through adjustments in heat transport and sea-ice related feedback (Kageyama et al., 2009; Blaschek and Renssen, 2013). The total amount of FWF can be constrained from both the ocean and land perspectives, as the FWF discharge serves as the link of water exchange between the ocean and continental ice. For instance, the fossil-coral-based estimates of far-field sea level change can reflect the total amount of FWF from the ocean perspective (Lambeck et al., 2014). From the land side, the total FWF amount can be roughly constrained by the geological indicators of ice-sheet retreat (Peltier, 2004). For the Holocene a total amount of freshwater equivalent to 60 m of sea level was released into the ocean between 11.5 and 6 ka with a large contribution from the LIS (Peltier, 2004; Lambeck et al., 2014). Sensitivity studies have further disclosed that, apart from total amount, various temporal distributions and geographical locations can induce different response in ocean circulation (Roche et al., 2010). Although ocean sediment data (e.g. detrital carbonate, ice rafted detritus) and geochemical tracers (e.g. $\delta^{18}O$, $^{87}Sr/^{86}Sr$, U/Ca) can provide certain constraints on FWF routing (Carlson et al., 2007; Jennings et al., 2015), it is uncertain how this amount of water was distributed spatially and temporally. Firstly, various magnitudes of FWF are suggested by different proxies. For instance, the geochemical tracer U/Ca suggests a slightly larger FWF discharge in St. Lawrence River between 12 and 11 ka than what the indicator $\delta^{18}O$ does, probably because additional factors, such as temperature and weathering, modulate the signal of changes in FWF (Carlson et al., 2007). Furthermore, the temporal distributions of FWF in different estimates are not identical either. For instance, the curve shapes of these proxy-based FWF estimates differ from the model-based estimates, with maximum reached at different time periods (Licciardi et al., 1999; Carlson et al., 2007), which is related to the difficulty in accurately dating samples. Additionally, approaching on a well-agreed geographical locations of FWF discharge is hindered by spatial sparseness of proxy records. Overall, the FWF is hugely uncertain during the early Holocene, especially in terms of discharge rate, location and timing. This uncertainty is also reflected in FWF differences of the four experiments discussed here, which potentially induces certain degree of inter-model divergences. To avoid the FWF related influences, it will be substantially beneficial to construct FWF protocols and apply them in all participating models for next inter-comparison project, as that would allow us to focus on dominant climate processes and feedbacks.

4.2.2 Impact of inter-model differences in climate sensitivities

In the present study, climate sensitivities are used to broadly indicate the sensitivity of the climate system to both radiative forcing and FWF forcing. The sensitivity to radiative forcing generally refers to the change in the global annual mean surface air temperature in response to doubling CO_2 (Knutti and Hegerl, 2008), representing a global average. Yet the CSr shows temporal and spatial patterns when examined in detail, as the CSr is a function of baseline climate and involves a series of processes on different timescales (Boer and Yu, 2003). In particular, multiple studies have quantitatively examined CSr for decades and have found that it varies for different climate states (Boer and Yu, 2003). The CSr generally decreases with warmer climate and increases with colder climate (Boer and Yu, 2003; Knutti and Hegerl, 2008). The feedback processes, such as those involving water vapor, lapse rate, surface albedo and clouds, can differ in strength for different climate states (Boer and Yu, 2003; Randall et al., 2007). For a simplicity, the CSr is assumed to be linearly changed over small ranges of climate differences, as vital global atmospheric feedbacks remain close to a constant with temperatures when the threshold value is not exceeded (Randall et al., 2007). Supposedly, the CSr during the early Holocene was slightly larger than that of 0 ka within this linear assumption, owning to a slightly cooler early Holocene. By contrast, the CSr at 6 ka was slightly smaller with a similar linear assumption. However, the exact change rates of CSr in response to climate states are still controversial. For instance, recent studies suggest weaker CSr enhancements than previous findings in response to the same amount of cooling (Kutzbach et al., 2013).

The CSr varies among individual models, ranging from 2°C in LOVECLIM to 4°C in FAMOUS (Table S1). Given this high CSr in FAMOUS and positive radiative forcings during the early Holocene, the climate in FAMOUS would be expected to be warmer than in other simulations when an identical model response to the ice sheets is assumed. However, the overall cool climate in FAMOUS seems to conflict this expectation. A plausible explanation for this paradox is that the expected warmth was overwhelmed by the ice-sheet-related cooling. Moreover, the spatial heterogeneity of CSr could outweigh this expected overall warming at certain regions and thus also potentially contributes to this conflict. For instance, Boer and Yu (2003) have revealed that the spatial patterns in CSr are partially due to local feedback processes and are reflected by the geographical distribution of sensitivity coefficient. In addition, the FAMOUS results (at 11.5ka) were obtained from a full transient simulations since the LGM, which implies that the model still had a “memory” of the preceding cold climates. Further detailed discussions on the CSr spatial pattern and sensitivity experiments deserve more effort, which is beyond the scope of the present study. The sensitivity of the climate system to freshwater forcing (CSf) varies among these models and associated influences on temperature are discussed in SI.

4.2.3 Impacts of the model physics and resolution

Model physics also contribute to the inter-models variations. By model physics, we refer to how climatic processes are represented in the model world, without considering the external radiative forcings. For instance, in CCSM3, the over-estimated albedo related to laterly adapted turbulent transfer formulations coefficient primarily contributes to multi-simulation temperature differences in E Siberia. Compared with 0 ka, the early-Holocene summer albedo in CCSM3 is reduced by more than 0.2 (Fig. 8), which primarily results from overestimated albedo at 0 ka since the value is more than 0.7. Such high albedo can be associated with the later adopted formulation of a turbulent transfer coefficient (Collins et al., 2006). According to an assessment of surface albedo (using MODIS data), this new formulation produces an extensive snow cover, because white-sky albedo in vegetated area might be insufficiently simulated and albedo increase with solar zenith angle probably is overestimated (Oleson et al., 2003).

The spatial resolution of a model determines the overall level of detail in the model representation of climate processes. The detailed representation of key physical processes such as the barrier effect of topography can improve the accuracy of the simulation. The spatial resolution, however, is limited by the computer power, especially for the simulations spanning millennial or longer time periods. Lunt et al. (2013) found that the resolution (effect) could partially explain multi-model differences, such as stronger cooling over African monsoon region in General Circulation Model (GCM) than in Earth System Model of Intermediate Complexity (EMIC), but also stated that this should be confirmed with further analysis. Some resolution-related patterns are observed when the atmosphere and ocean components are individually examined. For instance, the widely extended sea ice in FAMOUS is mainly caused by a relatively coarse spatial resolution in the ocean component (Gordon et al., 2000, Jones et al., 2005). Even though FAMOUS has similar physical and dynamical processes to HadCM3, this coarse resolution may lead of insufficient heat transport, such as in the Barents Sea, which ultimately leads to overestimated sea ice cover in that region (Gordon et al., 2000, Jones et al., 2005). This overestimated sea ice cover can cause cool Arctic climate through enhanced albedo-feedback (Renssen et al., 2005). In addition, the intense Alaskan warmth in winter in LOVECLIM might be related to its coarse vertical resolution, implying a relatively poor representation of the LIS topography. Therefore, the resolution would be a major factor when enhanced resolution is related to improvements of representation of dynamic processes. Nevertheless, to further investigate these resolution-related effects, more analyses, such as using proxy data to evaluate the simulated early-Holocene climate and applying a fully identical setup procedure, are still needed.

5. Conclusions

Transient features of the early-Holocene climate potentially introduce large uncertainties in simulated Holocene temperatures. To narrow these uncertainties and analyze the temperature trends, we compared four Holocene simulations performed with different models. The main findings are outlined as following:

1) Consistently simulated Holocene temperatures in multi-model simulations

Over the large scale of NH extratropics, the simulated temperatures are generally consistent among models with better agreements in summer than in winter, which is characterized by an early-Holocene warming, mid-Holocene maximum and gradual decrease toward 0 ka. On a regional scale, reasonably consistent temperature trends are found where climate is strongly influenced by the ice sheets, including Greenland, N Canada, N Europe and central-west Siberia. These simulated temperatures generally follow a similar pattern as stated above. Within these general patterns, the magnitude of early-Holocene warming slightly varies with regions. The strongest early-Holocene warming, up to 5°C in summer and 10°C in winter, is found in N Canada; whereas NE Europe and central-west Siberia show the least warming magnitude, with 4°C warming in winter and 1–3°C cooling in summer. An intermediate degree of warming is found in Greenland and NW Europe, with about 2°C in summer and 8°C in winter. Overall, these generally consistent temperature trends illustrate that forced climate change overwhelms the structural and parametric uncertainties, implying that the temperature trends are relatively well established in these regions.

2) Differences of the multi-model simulations and their direct causes

Large inter-model variations exist in Alaska, the Arctic, and E Siberia. In particular, the signals of individual model simulations are incompatible during the early Holocene. On the one hand, the strong southerly winds induced by the LIS over Alaska and part of E Siberia result in an anomalous warm climate in LOVECLIM. Higher summer temperatures (1–2°C) over E Siberia in CCSM3 than in other models are caused by strongly negative albedo anomaly between 11.5 ka and 0 ka, which ultimately is associated with a high albedo at 0 ka. On the other hand, the wide spread of simulated winter temperatures over the Arctic can be partially attributed to cold climate in FAMOUS due to its extensive sea ice cover. This extended early-Holocene sea ice cover influences the strength of the albedo-related feedback and could explain why the winter temperature in FAMOUS is 2–3°C lower than the ensemble mean.

3) Possible sources contributing to the different responses of climate variables

The multi-model comparisons reveal that varied responses in the models can be caused by the model physics, model resolution and model-dependent sensitivities. For instance, the later adopted turbulent transfer formulations in CCSM3 may cause an overestimated albedo over Siberia at 0 ka. Moreover, relatively simplified sea ice representation in FAMOUS may lead to overestimated sea ice cover. Also, the coarse vertical resolution in LOVECLIM might result in overestimated responses of atmospheric circulation to the LIS over Alaska. This inter-model comparison is partially hampered by the differences between the experimental setups and forcings, especially concerning the FWF, which has a major impact on the early-Holocene climate. Hence, using a standardized FWF for the early Holocene would be advantageous for future model inter-comparisons.

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Table 1. Main features of the setup of involved simulations

Simulation		LOVECLIM	CCSM3	HadCM3	FAMOUS
Prescribed forcing & ref.	ORB	Berger 1978	Berger 1978	Berger & Loutre (1991)	Berger & Loutre (1991)
	GHG	Loulergue et al. 2008; Schilt et al. 2010	Joos & Spahni 2008	Spahni et al. 2005; Loulergue et al. 2008	Spahni et al. 2005; Loulergue et al. 2008
	Ice sheet*	Icesheet, FWF	Icesheet, FWF	Icesheet	Icesheet, FWF
Initial condition		Eq_11.5 ka (1.2 kyr)	Tran_21 kyr	Pre-industrial snapshot	Tran_21 kyr
length_exp		11.5 kyr	21 kyr	Multiple snapshots	21 kyr
Ref. of simulation		Zhang et al. 2016	He, 2011	Singarayer & Valdes, 2010	This study

* Ice sheets includes the Laurentide Ice Sheet, Fennoscandia Ice Sheet and Greenland Ice Sheet.

Captions

Figure 1 Ice sheet related forcings during the early Holocene. (a) indicates prescribed freshwater flux (in mSv) into oceans, and (b) shows the area of ice sheets (in km^2) in the simulations.

Figure 2 Comparison of stacked proxy reconstruction with simulated summer (a) and annual mean temperature (b) in NH extratropics (over 30–90°N), shown as a deviation from 0 ka in °C. The stacked temperature reconstruction with 1σ uncertainty (dash lines) is based on Marcott et al. (2013). The proxy curve is the same in (a) and (b), although the authors interpreted it as annual mean.

Figure 3 Spatial distribution of simulated temperature anomalies (in °C) in the NH extratropics for the time windows of 11.5 ka (a) and 6 ka (b). Provided anomalies are relative to 0 ka.

Figure 4 Temperature trends (shown as anomalies from the PI in °C) over the regions where multiple simulations have good agreements, and corresponding multi-model ensemble mean (based on three transient simulations). The grey indicates the ensemble range.

Figure 5 Simulated temperatures anomalies (from the PI) over the regions where temperatures are less consistent across the simulations.

Figure 6 Anomaly in geopotential height fields (11.5 ka – 0 ka, in gpm) induced by the ice sheets at 11.5 ka, shown as values at 800 hPa in LOVECLIM, at 850 hPa in FAMOUS and HadCM3. Associated winds anomalies are indicated by the vectors.

Figure 7 Distribution of maximum sea ice (in Feburary). (a) represents the thickness of sea ice (in m) in LOVECLIM and HadCM3, and (b) shows the sea ice concentration in FAMOUS and CCSM3.

Figure 8 Surface albedo anomalies (shown as fractions) at 11.5 ka compared to 0 ka.

Lis of SI

1 Models and climate forcings

2 Discussion on AMOC, including Simulated AMOC and Impact of inter-model differences in climate sensitivities

Table S1 Summary of participated climate model

Fig. S1 Figure S1. Orbital-scale insolation and GHG related radiative forcing (in W m^{-2}) during the Holocene.

Fig. S2 Eight selected regions are marked as boxes. The background color indicates the simulated annual mean temperature in LOVECLIM at 11.5 ka.

Fig. S3 Atmospheric circulations changes induced by the topography of the LIS, shown as the differences between the LIS and N Pacific of geopotential height. Given to different vertical layers among models, the results are standardized by calculating the anomalies and percentage regarding 0 ka condition. The results are shown as 100-yr average in LOVECLIM and 1 kyr interval in FAMOUS and HadCM3.

Fig. S4 Total area of sea ice in NH (in 10e12 m^2)

Fig. S5 Meridional overturning streamfunction (in Sv) of the Atlantic Basin in the transient simulations

Fig. S6 Changes of maximum AMOC (in the box of 500-2000 m, 34S-50N, according to the definition of Hofer et al 2011; Drijfhout et al 2012) over the course of the Holocene. Results are shown as 100-yr averages.

Fig. S7 Albedo changes over the course of the Holocene