Changes in cryosphere, represented as glacial-interglacial cycles, are among the most prominent changes occurred in the Earth system throughout the geological history. The Early Holocene experienced the final melt of the Laurentide Ice Sheet and Fennoscandia Ice Sheet and related multiple transitions in various components of the climate system. This period therefore deserves specific attention for both better understanding the climate history, and for understanding the current global change given its similarity as to the ongoing transition.

By employing numerical climate model and proxy data, this work analysed the effects of dynamic ice sheets on climate in Northern Hemisphere during the early Holocene. The effects of ice sheets included the enhanced surface albedo, topography effect of the ice sheets and the influence of meltwater on the ocean-atmosphere system. These processes exerted the great heterogeneous patterns on climate.

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Holocene temperature trends in the Northern Hemisphere extratropics

YURUI ZHANG

UNIVERSITY OF HELSINKI
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YURUI ZHANG

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Abstract

As the latest epoch of the Earth's history, the Holocene is commonly defined as the last 11.7 ka BP (hereafter referred to as ka) and represents a new phase, encompassing the time span of human civilization. The last deglaciation lasted well into the Holocene, implying that the early Holocene was characterized by a large-scale reorganization with transitions in various components of the climate system. Studying the Holocene can provide insights into how the climate system functions, apart from the theoretical contributions to climate history itself.

We first conducted sets of simulations with different combinations of climate forcings for 11.5 ka and for the entire Holocene to investigate the response of the climate–ocean system to the main climate forcings. In particular, two possible freshwater flux (FWF) scenarios were further tested considering the relatively large uncertainty in reconstructed ice-sheet melting. Moreover, we compared four Holocene simulations performed with the LOVECLIM, CCSM3, FAMOUS and HadCM3 models by identifying the regions where the multi-model simulations are consistent and where they are not, and analysing the reasons at the two levels (of the models' variables and of the model principles and physics) where mismatches were found. After this, these multi-model simulations were systematically compared with data-based reconstructions in five regions of the Northern Hemisphere (NH) extratropics, namely Fennoscandia, Greenland, North Canada, Alaska and high-latitude Siberia. Potential uncertainty sources were also analysed in both model simulations and proxy data, and the most probable climate histories were identified with the aid of additional evidence when available. Additionally, the contribution of climate change, together with forest fires and human population size, to the variation in Holocene vegetation cover in Fennoscandia was assessed by employing the variation partitioning method.

With effects of climate forcings, including variations in orbital-scale insolation (ORB), melting of the ice sheets and changes in greenhouse gas (GHG) concentrations, the climate shows spatial heterogeneity both at 11.5 ka and over the course of the Holocene. At 11.5 ka, the positive summer ORB forcing overwhelms the minor negative GHG anomaly and causes a higher summer temperatures of 2–4 °C in the extratropical continents than at 0 ka. The ice-sheet forcings primarily induce climatic cooling, and the underlying mechanisms include enhanced surface albedo over ice sheets, anomalous atmospheric circulation, reduced the Atlantic Meridional Overturning Circulation (AMOC) and relevant feedbacks. In particular, the most distinct feature is a thermally contrasting pattern over North America, with simulated temperatures being around 2 °C higher than those at 0 ka for Alaska, whereas over most of Canada, temperatures are more than 3 °C lower. The geographical variability of simulated temperatures is also reflected in Holocene temperature evolution, especially during the early Holocene, as constant Holocene cooling in Alaska contrasts with strong early-Holocene warming (warming rate over 1 °C kyr^{-1}) in

northern Canada. The early-Holocene climate is sensitive to the FWF forcings and a brief comparison with proxy records suggests that our updated FWF (FWF-v2, with a larger FWF release from the Greenland ice sheet and a faster FWF from the Fennoscandian Ice sheet (FIS)) represents a more realistic Holocene temperature scenario regarding the early-Holocene warming and Holocene temperature maximum (HTM).

Comparison of multiple simulations suggests that the multi-model differences are spatially heterogeneous, despite overall consistent temperatures in the NH extratropics as a whole. On the one hand, reasonably consistent temperature trends (a temporal pattern with the early-Holocene warming, following a warm period and a gradual decrease toward 0 ka) are found over the regions where the climate is strongly influenced by the ice sheets, including Greenland, N Canada, N Europe and central-West Siberia. On the other hand, large inter-model variation exists in the regions over which the ice sheet effects on the climate were relatively weak via indirect influences, such as in Alaska, the Arctic, and E Siberia. In these three regions, the signals of multi-model simulations during the early Holocene are incompatible, especially in winter, when both positive and negative early-Holocene anomalies are suggested by different models. These divergent temperatures can be attributed to inconsistent responses of model variables. Southerly winds, surface albedo and sea ice can result in divergent temperature trends across models in Alaska, Siberia and the Arctic. Further comparisons reveal that divergent responses in these climate variables across the models can be partially caused by model differences (e.g. different model physics and resolution). For instance, the newly adopted formulation of the turbulent transfer coefficient in CCSM3 causes an overestimated albedo over Siberia at 0 ka, which leads to a stronger early-Holocene warmth than in other models. Moreover, the relatively simplified sea ice representation in FAMOUS probably leads to overestimated sea ice cover in the Arctic Ocean. The coarse vertical resolution in LOVECLIM might also introduce strong responses in atmospheric circulation over Alaska. From the perspective of climate features, the transient feature of the early-Holocene climate driven by the retreating ice sheets also influences the inter-model comparisons, as this transient feature induces a large degree of uncertainty into the FWF forcing.

Comparisons of multiple model results with compiled proxy data at the sub-continental scale of NH high latitudes (i.e. Fennoscandia, Greenland, north Canada, Alaska and Siberia) reveal regionally-dependent consistencies in Holocene temperatures. In Fennoscandia, simulations and pollen data suggest a summer warming of 2 °C by 8 ka, although this is less expressed in chironomid data. In Canada, an early-Holocene warming of 4 °C in summer is suggested by both the simulations and pollen results. In Greenland, the magnitude of early-Holocene warming of annual mean ranges from 6 °C in simulations to 8 °C in $\delta^{18}O$-based temperatures. By contrast, simulated and reconstructed summer temperatures are mismatched in Alaska. Pollen data suggest 4 °C early-Holocene warming, while the simulations indicate 2 °C Holocene cooling, and chironomid data show a stable trend. Meanwhile, a high frequency of Alaskan peatland initiation before 9 ka can either reflect a high temperature, high soil moisture content or large seasonality. In high-latitude Siberia, simulations and proxy data depict high Holocene temperatures, although these signals are noisy owing to a large spread in the simulations and to a difference between pollen and chironomid results. On the whole, these comparisons of multi-model simulations with proxy reconstructions further confirm the Holocene climate evolution patterns in Fennoscandia, Greenland and North Canada. This implies that the Holocene temperatures in these regions have been relatively well established, with a reasonable representation of Holocene climate in the multiple simulations and a plausible explanation for the underlying mechanisms. However, the Holocene climate history and underlying mechanisms in the regions of Siberia and Alaska remain inconclusive.

Variation partitioning revealed that climate was the main driver of vegetation dynamics in Fennoscandia during the Holocene as a whole and before the onset of farming. Forest fires and population size had relatively small contributions to vegetation change. However, the size of the human population became a more important driver of variation in vegetation composition than climate during the agricultural period, which can be estimated to have begun at 7–6 ka in Sweden and 4–3 ka in Finland. There is a clear region-dependent pattern of change caused by the human population: the impact of human activities on vegetation dynamics was notably higher in south Sweden and southwest Finland, where land use was more intensive, in comparison with central Sweden and southeast Finland.

This thesis investigates the climate responses to the main forcings during the Holocene through various approaches, which has potential implications for the interactions between ice sheets and the climate, the Holocene climate history and current global change. The atmosphere-ocean system was sensitive to the FWF forcing during the early Holocene, implying that existing uncertainties in reconstructions of ice-sheet dynamics can be constrained by applying different freshwater scenarios via a comparison with proxy data. The Holocene climate history in most of the Northern Hemisphere extratropics is relatively well established, especially in regions that were strongly influenced by ice sheets. The implications of our investigation (on the transient early-Holocene) for the current global change are twofold. First, regional heterogeneity of the climate responses implies that regional differences should be taken into account when adapting to the current global change. Second, apart from the different scenarios of GHG forcing, inter-model comparison would be a good option to reduce model-dependency in estimation of the future climate.

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By doing this research during the last four years, I finally understand why a PhD experience is often seen as a metaphor of an adventurous journey. Like many journeys, my PhD life possesses all elements of an adventure, including happiness, surprises, worries, anxiety, and all kinds of memorable moments. Luckily, I received countless kind helps and supports from many kind people, which enables me to approach the end of this journey. This journey is enormously valuable for me because it not only leads to academic progresses, but also means the self-improvements and personal development. At this moment, I am excited about to finish it, and also want to express my sincere thanks to all of those who made this special journey to happen.

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“知之為知之，不知為不知，是知也”

—《论语·为政》，孔子

When you know a thing, to hold that you know it; and when you do not know a thing, to allow that you do not know it - this is fundamental/primary knowledge.

—Confucius

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List of original publications

This thesis is based on the following publications:

I Zhang, Y., Renssen, H., Seppä H.2016. Effects of melting ice sheets and orbital forcing on the early Holocene warming in the extratropical Northern Hemisphere, *Clim. Past,* 12,. 1119–1135.

II Zhang, Y., Renssen, H., Seppä, H., Valdes, P. J. Holocene temperature trends in the extratropical Northern Hemisphere based on inter-model comparisons. Submitted to *Journal of Quaternary Science*.

III Zhang, Y., Renssen, H., Seppä, H., Valdes, P. J. 2017. Holocene temperature evolution in the Northern Hemisphere high latitudes – data-model comparisons, *Quaternary Science Reviews,* 173, 101–113.

IV Kuosmanen, N., Marquer, L., Tallavaara, M., Molinari, C., Zhang, Y., Alenius, T., Edinborough, K., Pesonen, P., Reitalu, T., Renssen, H., Trondman, AK. Seppä, H. The role of climate, forest fires and human population size in Holocene vegetation dynamics in Fennoscandia. Under revision to *Journa of vegetation Science.*

The publications are referred to in the text by their roman numerals.

Author's contribution to the publication

I The study was designed by H. Renssen and H. Seppä. The experiments were conducted by Y. Zhang with aids of H. Renssen, and the results were jointly interpreted by Y. Zhang and H. Renssen. Y. Zhang was responsible for preparing the manuscript, with the comments and contributions from H. Renssen and H. Seppä.

II The study was designed by H. Renssen, Y. Zhang and H. Seppä. The analysis was conducted by Y. Zhang with advices from H. Renssen, and P.J. Valdes provided FAMOUS and HadCM3 simulations data. Y. Zhang had the main responsibility of preparing the manuscript, which was commented on and edited by other co-authors.

III The study was designed by H. Renssen, Y. Zhang and H. Seppä. The analysis was conducted by Y. Zhang with advices from H. Renssen and H Seppä. Y. Zhang was responsible for preparing the manuscript, which was commented on and edited by other co-authors.

IV The study was designed by N. Kuosmanen, L. Marquer and H. Seppä. The analysis was conducted by N. Kuosmanen and L. Marquer. Y. Zhang provided the climate simulation data and related interpretation. N. Kuosmanen had the main responsibility of preparing the manuscript, which was commented on and edited by other co-authors.

Abbreviations

AMOC	**A**tlantic **M**eridional **O**verturing **C**irculation
AOGCMs	**A**tmosphere-**O**cean **G**lobal **C**limate **M**odels
CCSM3	**C**ommunity **C**limate **S**ystem **M**odel version 3
EMICs	**E**arth system **M**odels of **I**ntermediate **C**omplexity
FAMOUS	**FA**st **M**et **O**ffice/UK **U**niversities **S**imulator
FIS	**F**ennoscandian **I**ce **S**heet
FWF	**F**resh**w**ater **F**lux (Meltwater flux)
GHG	**G**reen**h**ouse **g**ases
HadCM3	**Had**ley Center **C**limate **M**odel
HTM	**H**olocene **T**emperature **M**aximum
LGM	**L**ast **G**lacial **M**aximum
LIG	**L**ast **I**nter**g**lacial
LIS	**L**aurentide **I**ce **S**heet
LOVECLIM	**LO**CH-**VE**CODE-**CLI**O-AGISM **M**odel
MAT	**M**odern **A**nalog **T**echnique
MH	**M**id-**H**olocene
NH	**N**orthern **H**emisphere
ORB	**Orb**ital
PFTs	**P**lant **F**unctional **T**ypes
REVEALS	**R**egional **E**stimates of **VE**getation **A**bundance from **L**arge **S**ites model
WAPLS	**W**eighted **A**veraging **P**artial **L**east **S**quare regression and calibration

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1 Introduction

Reliable projections require a comprehensive and thorough understanding of the climate system (such as climate variability), as these climate projections are generated by climate models that were built using our understanding of the climate system. Climate changes in the past extend the range of the observed climate, providing an extra constraint for climate models and increasing confidence in the predicted results. Both warmer (e.g. Last Interglacial (LIG, 130–115 ka)) and cooler climates (e.g. Last Glacial Maximum (LGM, ~21 ka)) than the present can be found in geological history, which provides an opportunity to examine the response of the climate system under climate states that differ from the present. For similar reasons, periods characterized by marked climate change, such as the last deglaciation, may shed light on current global changes.

1.1 The climate system, its variation and palaeoclimate change

1.1.1 The climate system and mechanisms underlying climate change

Earth's climate is a dynamic characterization of the climate system and is described by geologists as both an agent and a feature of land-surface change. The climate system includes the atmosphere, hydrosphere (ocean), biosphere (vegetation), cryosphere and land surface, with various levels of nested sub-systems (Fig. 1), as defined by the Global Atmospheric Research Programme (GARP) of the World Meteorological Organization in 1975. Subsequently, the essential role of the interactions between these components and associated processes in the climate system have attracted increasing attention, as reflected in the definition of the climate system given by the United Nations Framework Convention on Climate Change (FCCC) in 1992. The interactions of heat, energy and momentum between these components occur through physical, chemical and biological processes and feedbacks. For instance, the heat exchanges between the ocean and atmosphere components, mainly through latent heat and associated with the hydrological cycle, influence the heat distribution. Meanwhile, albedo-associated feedbacks link the components of atmosphere and cryosphere together and influence the radiation budget of the climate system. These multiple components of the climate system and their interactions determine climatology as an interdisciplinary field, of which physical processes are controlling the Earth's climate.

The climate system is a dynamic system, with exchanges of radiant energy, mass, water and, to a lesser extent, carbon and nitrogen among different components. The fluxes of these factors are vectors, as they involve the movement of some quantity from one place to another, and the direction of flow is also crucial to balancing the system (e.g. Trenberth & Stepanial 2004). The net fluxes differ considerably as a function of the time period considered, which results in a dynamic climate over given regions. Climate forcings can exert impacts on these fluxes at various time scales, and thus influence climate conditions. The various climate forcings fall into two types with reference to the climate system itself: external and internal forcings (McGuffie & Henderson-Sellers 2005). External forcings tends to push the climate system to a more stable state that is in equilibrium with the new force, while the forcing itself is not influenced by this forcing. Orbital-scale variation in insolation is one of these external forcings and provides an important explanation for the temporal climate variations of the late Quaternary (e.g. Berger 1978). The astronomical parameters of the Earth's orbit, including eccentricity, obliquity and precession, determine the solar radiation received at the top of the atmosphere and its distribution over latitudes (Berger 1978). The periodic changes in these parameters primarily explain the periodic alternation of glacial and interglacial periods throughout geological history, part of which are explained by the Milankovitch theory. By contrast, internal climate forcings involve a variety of processes occurring within the climate system, implying that these processes not only impact on the climate state, but they can also be influenced by the climate state (McGuffie & Henderson-Sellers 2005). For instance, changes in ocean circulation influence the climate locally and globally by adjusting heat and energy transportation, and the strength (changes) of ocean circulations also partially depends on spatial patterns of temperature and atmospheric circulation.

However, the boundary between the external and internal forcings is changeable in certain circumstances. For instance, variations in

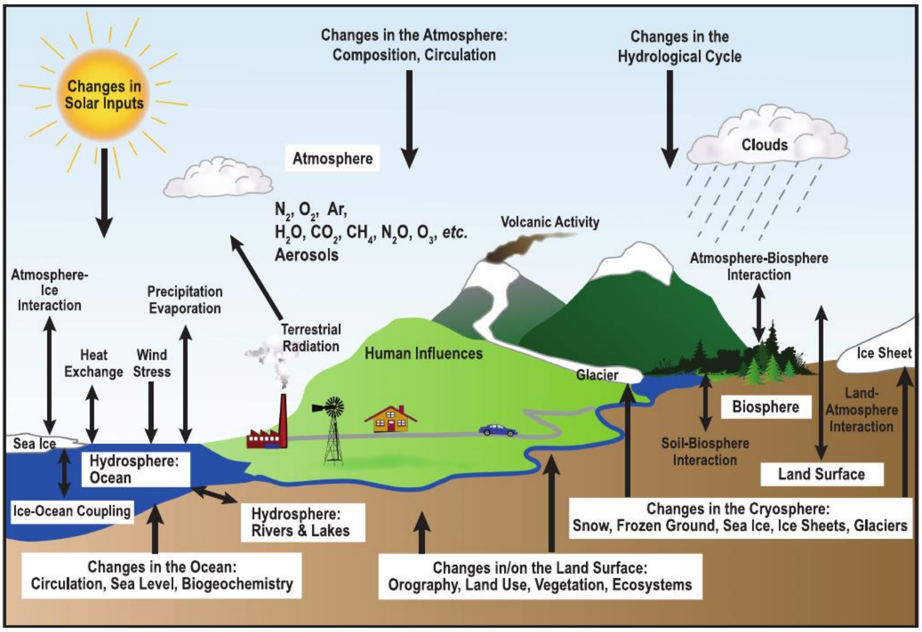

Figure 1. A schematic illustration of the components and interactions in the climate system (adapted from Le Treut et al. 2007).

atmospheric CO_2 influence the climate through the well-known greenhouse effect and are controlled by carbon fluxes between different components of the climate system. The changing CO_2 concentration can be both an external and an external forcing, depending on the time scale of interest. On short time scales (shorter than the millennial scale), CO_2 variations in the atmosphere are an external forcing, which serve as an agent of climate change and are not directly affected by the climate at this time scale. Meanwhile, on a long geological time scale (a million years or longer), variation in CO_2 can also be an internal forcing, as carbon storage in the deep ocean and the atmosphere are linked through the long-term carbon cycle (e.g. Boyle & Keigwin, 1985). Temperatures play an important role in this cycle, such as through their influence on the strength of the oceanic circulation, which impacts on CO_2 exchange between the oceans and atmosphere, and temperatures also affect carbon storage in the biosphere, some of which can be ultimately deposited in oceans through geological movements. Similarly, anthropogenic-related changes in atmospheric CO_2 concentrations should be regarded as internal processes when considering these changes on a geological timescale. However, these anthropogenic-related changes can also act as external forcings for climate change on decadal or centennial timescales, because the burning of fossil fuels injects extra greenhouse gases and sulphate aerosols into the atmosphere, and humans change land surface properties through agricultural activities, which exerts critical impacts on climate. However, the intensification of these changes is barely influenced by the climate on this short timescale.

Together with these internal and external climate forcings, feedbacks also make important contributions to the variability of the climate system. Feedback occurs when the response of a system to an input further alters the input variable of the system through the involved processes. Through this loop, feedbacks can either amplify forcing-induced changes through positive responses, or dampen the forcing-based changes via negative reaction mechanisms. For instance, albedo-related feedbacks due to changes in the land ice and sea ice have played an essential role in glacial and interglacial transitions by influencing the radiation budget of the Earth (Deblonde et al. 1992; Bintanja et al. 2002). The high albedo of ice both on the land and in the oceans reduces the amount of solar radiation absorbed, which facilitates the formation of ice and triggers positive feedbacks (Curry et al. 1995). In the climate system, some original changes even trigger both negative and positive changes at the same time, and the net effect can therefore either lead to an enhanced or a reduced response, depending on which one is dominant. For instance, an enhanced amount of water vapour in the atmosphere can both increase and decrease the original change in water vapour. On the one hand, more water vapour in the atmosphere leads to a warming climate through the greenhouse effect of water vapour, and increases the evaporation of surface water, which introduces more water molecules into the atmosphere. On the other hand, increased water vapour in the atmosphere accelerates cloud formations, thus reflecting more sunlight back into space and lowering the temperature, which ultimately limits water molecular movement into the atmosphere. Therefore, the final effect of water vapour on the climate depends on the net effect of these two opposite processes, which strongly depends on the vertical distribution of clouds and can overall be positive in the tropics (Inamdar & Ramanathan 1998; Mcguffie & Henderson-sellers 2005). The distinctions between climate forcings and feedbacks are flexible, as they depend on the timescales of processes and the period of consideration. Commonly, a process is regarded as a forcing (boundary condition) when its operating time is much longer than the time scale of interest, which can vary from years to decades (such as feedbacks related to snow, sea ice, the upper ocean, dust and aerosols) to millions of years (such as weathering and the evolution of vegetation) (Palaeosens Project members, 2012). Similarly, the process serves as a feedback when its timescale is shorter than the time period of interest.

1.1.2 Palaeoclimate change and the Holocene

With the effects of all these external and internal forcings and feedbacks, the regional and local climate varies spatially, which is a function of the geographic location (e.g. general latitudinal patterns with positive gradients from low latitudes toward high latitudes). Meanwhile, the climate system as a whole is also temporally dynamics and has fluctuated in a distinctive way over the history of the Earth, owing to various scales of temporal variation in these factors. Natural variation in climate is not simply the sum of the effects of these factors, but is the result of non-linear responses in linked climatic components to these changes. Throughout the history of the Earth, the temperature has undergone considerable changes. As recorded by multiple proxies, the climate history is punctuated by multiple glacial and interglacial cycles. For example, temporal variation in $\delta^{18}O$ in benthic foraminifera has shown cyclic changes with an amplitude of about 2‰ during the last 600 ka (EPICA Community Members, 2004; Lisiecki et al. 2005). These changes imply that temperatures fluctuated by more than 3–4 °C (Johnsen et al. 2001; Jouzel et al. 2003). Combining different proxies gives an overview of the temperature history of the Earth. For instance, from the last interglacial (LIG, ~130 ka) to the last glacial maximum (LGM, ~21 ka), the temperature dropped by 5–6 °C (CAPE Project, 2006). Following the LGM, the climate warmed up with ongoing deglaciation, although this was accompanied by abrupt cooling events (Hughen et al. 1996; Peltier 2004; Shakun et al. 2012).

During the last deglaciation, the climate system experienced a transition phase that lasted well into the Holocene. As the latest epoch of the Earth history, the Holocene refers to the last 11.7 ka and represents a relatively warm climate in comparison with preceding periods. The Holocene embraces the time span of human civilization, such as the evolution of agriculture with the domestication of plants and animals and its spatial expansion (Gupta 2004). The land surface has became progressively covered by forests as the ice retreated (e.g. MacDonald et al. 2000; CAPE project, 2001). In the late Holocene, the forest cover decreased again as mankind's demand for timber and agricultural land grew (e.g. Kaplan et al. 2009). In particular, the transition in the climate system during the early Holocene is analogous to currently ongoing global change in terms of the overall warming trend of global temperatures and transient features of the climate system, even though they have different causes. In addition, according to Vaughan et al. (2013), the total volume of ice currently existing on land is equivalent to a rise of more than 60 m in the sea level, including the Antarctic (58 m) and Greenland (7 m) ice sheets. This is similar to the volume of melting ice during the early Holocene (Lambeck et al. 2014). In addition to these similar magnitudes of ice, the mechanisms behind the melting ice sheets are also the same, such as the effects on albedo, atmospheric circulation and perturbation of meltwater, although these ice sheets are situated in very different geographical locations. Therefore, studies on the Holocene can accumulate information on how the climate system functions in response to changing climate forcings, thus providing insights into the current global change.

1.2 The early-Holocene transition and Holocene climate forcings

1.2.1 The early-Holocene transition in the climate system

During the last deglaciation, the climate system experienced a major reorganization that lasted well into the Holocene, for which relatively abundant proxy records are available. In particular, the early part of the Holocene (11.5–7 ka) was characterized by transitions in various components of the climate system (Fig. 2). In the cryosphere, the Laurentide Ice Sheet (LIS) and Fennoscandian Ice Sheet (FIS) continuously waned until they finally disappeared around 10 ka and 6.8 ka, respectively (Dyke et al. 2003; Occhietti et al. 2011; Cuzzone et al. 2016). At the onset of the Holocene, the LIS and FIS partially covered N America and Fennoscandia, with a maximum thickness of up to almost 2 km and 200 m, respectively as evidenced by multiple lines of geological data (Peltier 2004; Ganopolski et al. 2010). The total amount of melt water from these ice sheets was equivalent to a sea-level rise of up to 60 m over the course of the Holocene (Lambeck et al. 2014). The presence of the decaying ice sheets also induced temporal-spatial heterogeneity in the early-Holocene climate, as these retreating ice sheets exerted multiple influences on the climate system (Renssen et al. 2009).

The early-Holocene transition in the atmosphere was recorded in multi-proxy records. Oxygen isotope measurements from ice cores in Greenland reveal an increase in $\delta^{18}O$ by up to 3–5‰, which indicates an early-Holocene warming (Dansgaard et al. 1993; Rasmussen et al. 2006; Vinther et al. 2006; 2009). Moreover, this early-Holocene warming is also registered in biological proxies. For example, a 4~5 °C warming in western and northern Europe is indicated by chironomid and macrofossil data obtained from lake sediments (Brooks & Birks 2000; Brooks et al. 2012; Birks 2015).

Proxy-based vegetation reconstructions also suggest a transition in the biosphere during the early Holocene. The biomass of vegetation tends to increase after the retreat of ice sheets, even though the absolute magnitudes of changes are spatially heterogeneous. Vegetation reconstructions have revealed a northward expansion of boreal forest in the circum-Arctic region by 9–8 ka (MacDonald et al. 2000; CAPE project, 2001; Fang et al. 2013). This expansion of boreal forest into regions that were not previously vegetated or were covered by tundra caused a reduction in the surface albedo and induced a positive feedback in the warming trend (Claussen et al. 2001). Therefore, these interactions between the climate and vegetation composition also had important influences on the transient climate during the early Holocene.

The ocean, one of the main constituents of the climate system, also experienced a critical transition, as indicated by marine sediment core data and numerical modelling. Firstly, multiple proxies have suggested that considerable changes in sea surface temperatures (SSTs) occurred during the early Holocene. For instance, the variation in stable isotopes ($\delta^{18}O$ and $\delta^{13}C$) measured from planktonic foraminifera (Fig. 2) reflects a rise in SSTs in the North Atlantic (Bond et al. 1993; Kandiano et al. 2004; Hald et al. 2007). Moreover, both proxy (e.g. $^{231}Pa/^{230}Th$ measurements on sediment cores) and modelling studies have found that the Atlantic Meridional Overturning Circulation (AMOC) was relatively

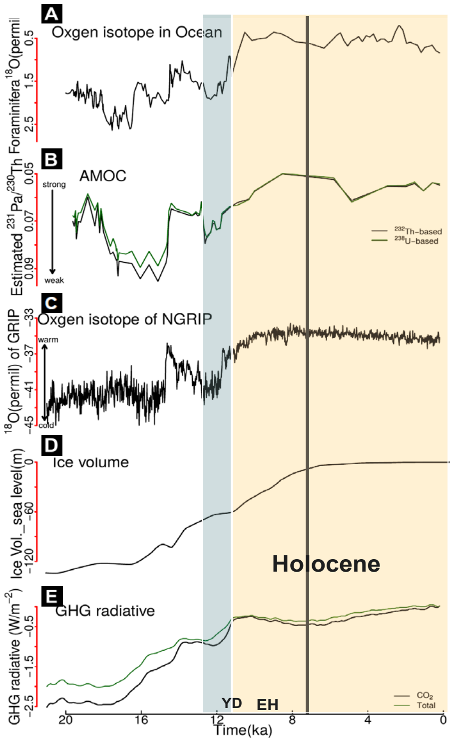

Figure 2. The Holocene period and early Holocene transition in the climate system. (A) Planktonic foraminifera $\delta^{18}O$ at the core (OCE326-GGC5) from the subtropical Atlantic, representing the near-surface temperature and/or local salinity (McManus et al. 2004). (B) Estimated $^{231}Pa/^{230}Th$, indicating the strength of AMOC (McManus et al. 2004); (C) $\delta^{18}O$ of NGRIP (Vinther et al. 2006); (D) Ice volume locked in the land, shown as the equivalent sea level (m) (Lambeck et al. 2014); (E) Radiative forcing of greenhouse gases, shown as CO_2 and total GHG (CO_2, CH_4 and N_2O) (Luthi et al. 2008). YD and EH indicate the Younger Dyras event and the earth-Holocene transition..

weak during the early Holocene due to ice-sheet melting (McManus et al. 2004; He et al. 2010; Blaschek & Renssen 2013), which led to reduced northward heat transport and extended sea-ice cover (Renssen et al. 2010; Roche et al. 2010; Thornalley et al. 2011, 2013).

Important adjustments in the carbon cycle occurred in the early Holocene. The carbon cycle describes the exchange of carbon between the carbon pools of the biosphere, pedosphere, geosphere, hydrosphere and atmosphere, implying that carbon can be reused through biogeochemical processes. Changes in the global biomass and ocean temperatures and salinities control the exchanges between the main global carbon reservoirs (the oceans, terrestrial biosphere and atmosphere). CO_2 is a well-mixed atmospheric trace gas responding to carbon flux variations in both the Southern and Northern hemispheres. During the early Holocene, for example, atmospheric CO_2 rose by 20–30 ppm that is equivalent to 0.3 W m^{-2} increase in radiative forcing (Fig. 2) (Luthi et al. 2008). CO_2 reconstructions based on stomatal indices suggest even larger variabilities (Wagner et al. 1999; Jessen et al. 2007). The reorganization of the ocean circulation during the early Holocene (Jessen et al. 2007) has been suggested as the main cause of these CO_2 changes, such as through influences on the deep water upwelling in the Southern Hemisphere, enhancing the ocean-to-atmosphere CO_2 flux.

These transient climatic components are actively interconnected by multiple processes (Fig. 1). For instance, the dynamic early-Holocene climate could have impacted on CO_2 concentrations in the atmosphere via the carbon cycle, and CO_2 concentrations have inverse influences on the climate via the well-known greenhouse effect. Meanwhile, ice sheets served as both causes and effects of changes in the climate system. Dynamic ice sheets are sensitive to climate changes, which also exert multiple impacts on the atmosphere–ocean system. Investigation of the climate response to decaying ice sheets during the early Holocene could shed light on these interactions.

1.2.2 Main forcing of the Holocene climate

Orbital-scale solar insolation (ORB) has played a role in the climate history of the Earth (Berger 1988). During the Holocene, orbital-induced summer insolation has showed a decreasing trend at mid- and high latitudes (Berger 1978; Denton et al. 2010). Meanwhile, the melting FIS and LIS also exerted in important influence on Holocene climate evolution until they eventually disappeared around 10 and 6.8 ka, respectively which further adjusted the spatial patterns of the climate (Dyke et al. 2003; Renssen et al. 2009; Blaschek and Renssen 2013; Cuzzone et al. 2016). First, the surface albedo was much higher over the ice sheets compared to ice-free surfaces, which resulted in relatively low temperatures. Second, the ice-sheet topography could have also influenced the climate through the adjustment of atmospheric circulation (Felzer et al. 1996; Justino & Peltier 2005; Langen & Vinther 2009). A large-scale ice sheet could have generated a glacial anticyclone, which could have locally reduced the temperature over the ice sheet (Felzer et al. 1996), but it may also have caused warming by 2–3 °C over the North Atlantic during the LGM (Pausata et al. 2011; Hofer et al. 2012). Third, it has been found that melting ice sheets caused a weakening of the maximum AMOC during the early Holocene, which led to reduced northward heat transport and extended the sea-ice cover, and thus influenced the high-latitude climate (Renssen et al. 2010; Thornalley et al. 2011, 2013). Overall, the net effect of ice sheets on the early-Holocene climate can be expected to have tempered the orbitally-induced warming at mid- and high latitudes. Additionally, radiative forcing due to variation in the concentration of greenhouse gases (GHG) in the atmosphere has also affected the Holocene temperatures. From the onset of the Holocene, the total effect of GHG dominated by CO_2 showed a rapid rise by 10 ka, reaching a local peak of -0.3 W m^{-2} relative to the pre-industrial period, which was followed by a slight decrease to a minimum at 7 ka, before gradually increasing towards 0 ka (Loulergue et al. 2008; Joos & Spahni 2008; Schilt et al. 2010).

1.2.3 Previous studies on the Holocene climate and remaining problems

The impact of these forcings on the Holocene climate has been partially examined in previous modelling studies. Renssen et al. (2009) found that the Holocene climate was sensitive to the ice sheets and that LIS cooling effects delayed the Holocene Thermal Maximum (HTM) by up to thousands of years. Blaschek and Renssen (2013) later revealed that melting of the Greenland ice sheet (GIS) had an identifiable impact on the climate over the Nordic Sea. However, the focus in these studies has been on the period after 9 ka by investigating the influence of the decay of the LIS and GIS on the climate relative to other climate forcings (Renssen et al. 2009; Blaschek & Renssen, 2013). The most important challenges in simulating climate during the initial phase of the early Holocene are the inherent uncertainties in the ice-sheet forcings in terms of the ice-sheet dynamics and particularly the associated meltwater release. The ocean–atmosphere system is sensitive to freshwater release, including the total volume, the location, timing and rate (Roche et al. 2007). Different approaches appear to suggest slightly different rates of ice-sheet melting. For instance, recent deglaciation studies based on cosmogenic exposure dating indicate slightly older ages of deglaciation in some regions than suggested by radiocarbon dating data (Carlson et al. 2014; Cuzzone et al. 2016), primarily because of large uncertainty in the bulk organic sample ages and the possibility of old carbon contamination (Stokes et al. 2010; Carlson et al. 2014). Furthermore, the Younger Dryas stadial ended at 11.7 ka (Fig. 2), and may still have influenced the early- Holocene climate due to the long response time of the deep ocean (Renssen et al. 2012).

Apart from these modelling studies, the mid-Holocene (MH, 6 ka) has been relatively well investigated by the Palaeoclimate Modeling Inter-comparison Project (PMIP) in terms of inter-model comparisons and model–data comparisons, as summarized in Table 1. In order to evaluate the uncertainties of Holocene simulations regarding model-dependent variations, studies have compared multiple simulations with a focus on snapshot experiments for the mid-Holocene (Harrison et al. 1998; Braconnot et al. 2000, 2007). Other currently available transient inter-model comparisons have examined periods shorter than the whole Holocene, such as 8–2 ka and the last 2 kyr (Bothe et al. 2013; Eby et al. 2013; Bakker et al. 2014). Meanwhile, PMIP has also conducted several model–data comparisons to investigate the different processes, with a focus on the mid-Holocene as well (e.g. Masson et al. 1999; Bonfils et al. 2004; Brewer et al. 2007; Zhang et al. 2010; Jiang et al. 2013). Liu et al. (2014) have also recently compared model results with proxy-based reconstructions to investigate the contradiction of Holocene temperature trends (i.e. the reconstructed cooling and the simulated warming during the Holocene on large-scale climate change, such as over 30–60°N and 60–90°N). However, it remains unclear how similar or different model simulations of the early Holocene climate are, and how these results compare with proxy-based datasets also remains unknown. Nevertheless, this is important to examine in detail, as the early

Table 1. Inter-model and model-data comparison studies included in PMIP

Ref	Period	Topic
Braconnot 2007	MH & LGM	Comparison of PMIP1 & PMIP2
Gladstone 2005	MH	NAO & SLP (compared the present day)
Kageyama 2013	MH & LGM	Comparison of IPSL_CM4 & IPSL_CM5A
Harrison 1998	MH	Vegetation and climate (T & P)
Braconnot 2000	MH & LGM	African monsson (inter-model)
Bothe 2013	The last 1 kyr	Inter-model comparison
Eby 2013	The last 1 kyr	Inter-model comparison of EMICs
Bakker 2014	MH	Multi-model-data comparison
Masson 1999	MH	Model-data comparison in Europe
Bonfils 2004,Brewer 2007	MH	Data-model comparison in Europe

Holocene provides an analogous context for the current global warming

1.3 The research questions of the thesis

The topic of the thesis is the Holocene temperature history, and different approaches are adopted in the relevant publications, as illustrated in Figure 4. These publications are arranged in the order in which they address the following research questions.

- How does the Holocene climate in the NH extratropics respond to the main forcings? (Publication I)

This general question is addressed from both spatial and temporal perspectives. For the spatial aspect, the question of how the climate in equilibrium experiments responds to the main forcing at the onset of the Holocene is a key issue. This issue can be further disentangled as follows: a) What are the main spatial characteristics of temperature at the onset of Holocene under different climate forcings, especially for ice-sheet decay? b) What would the possible mechanisms be? The climate forcings show various spatial patterns during the Holocene. Thus, answering question a) can enable us to disentangle the impact of these key forcings on the Holocene climate. Question b) can guide us to analyse the response of various climate variables, such as surface albedo, atmospheric circulation and AMOC strength.

From the temporal perspective (in transient simulations), the specific sub-questions include the following: a) What are the temporal trends of Holocene temperatures over different regions? b) How much do different forcings contribute to these temperature trends? c) Which FWF scenarios provide more a plausible Holocene temperature evolution regarding the early-Holocene warming and timing of the HTM? With these temporally varied forcings, answering question a) provides a basis for further analysis. By comparing the simulations with and without the ice-sheet forcings, question b) can be addressed. Answering question c) can provide insight into the Holocene temperature history and improve understanding of climate variability.

- To what extent are multiple simulations (performed by different models) consistent? (Publication II)

This general question can also be divided into three levels. First, what are the consistencies and divergences among these Holocene simulations (namely LOVECLIM, CCSM3, HadCM3 and FAMOUS)? In other words, to what degree and over which reigions are these Holocene simulations consistent and divergent. Second, what climate variables in models cause these inter-model divergences? In particular, detection of the variables in simulations causing these multi-model variations from the model response perspective can facilitate to the identification of their direct reasons when divergences are found. Third, where do these discrepancies originate? This implies tracking down of the original sources of these inconsistencies from the models themselves, such as different or even biased model principles.

- To what extent are model simulations and proxy data consistent? (Publication III)

There are three sub-questions under this broad question. First, to what degree do proxy-based reconstructions agree with model results? Expressly, we identify the regions where data and model results are consistent and where there is model mismatch with proxy data at the sub-continental scale of NH high latitudes. Second, what are the potential sources of uncertainty in simulated temperatures and proxy-based reconstructions? The answer to this question can provide insights into the interpretation of proxy results and validation of simulations. Third, what are the most probable Holocene temperature trends? Specifically, we conclude on the most probable temperature trends over these regions, with aid of other available evidence that is independent of the proxy data used.

- How much does climate change contribute to Holocene vegetation dynamics? (Publication IV)

Estimating the contributions of various potential factors to Holocene vegetation dynamics in S Fennoscandia be can divided into two aspects. Firstly, how much do different factors contribute to the Holocene vegetation dynamics and what are the main drivers of vegetation dynamics? Explicitly, we quantitatively evaluate the contribution of potential factors in sub-regions of S Fennoscandia (i.e. including South Fennoscandia, S Sweden, C Sweden and S Finland). Secondly, are there any changes in dominant patterns over different periods? Alternatively, are these patterns different before and after the onset of farming activity?

2 Methods to study Holocene climate

2.1 The LOVECLIM climate model

With the rapid development of computing power, climate models have become useful tools to investigate climate changes on various timescales. Climate models are mathematical representations of our understanding of the climate system, including movements of heat and mass within components of the climate system and also interactions between different components (e.g. via physical processes). Various (e.g. dynamical) equations are essential elements of climate models, conducting model simulations thus means solving these equations with given forcings and boundary conditions. The testing of different scenarios, termed sensitivity test, allows us to explore the plausible mechanisms behind climate changes and to analyse the temporal-spatial variability.

We used the LOVECLIM climate model, a three-dimensional Earth system model, to investigate the Holocene climate history. LOVECLIM stands for the five coupled sub-systems, namely LOCH, VECODE, ECBILT, CLIO and AGISM models, representing the carbon cycle, vegetation, atmosphere, ice sheets and the ocean components, respectively (Fig. 3). The atmospheric component, ECBILT, is governed by the quasi-geostrophic potential vorticity equation (Opsteegh et al. 1998), with three vertical layers (at 800, 500 and 200 hPa) and T21 (~5.6°) horizontal resolution. Processes are simplified at various degrees. For instance, the humidity in the atmosphere is determined by the total precipitable water content between the surface and 500 hPa. The diabatic heating due to radiative fluxes, the release of latent heat, and the exchange of sensible heat with the surface is parameterized. The land surface model is also part of ECBILT, including the moisture budget, snow cover and the surface albedo. The surface albedo is a function of the fraction of the grid box covered by ocean, sea ice, trees, desert and grass. The ocean module is represented by CLIO, which couples a 3-layer thermodynamic–dynamic sea-ice model with an ocean general circulation model (GCM). The ocean GCM has a free-surface with 20 vertical layers and 3° × 3° (lat×lon) resolution (Goosse & Fichefet, 1999). The albedo formulation in the sea-ice model considers the state of the surface (frozen or melted) and the thickness of the snow and ice covers. The momentum balance of sea ice results from dynamical interaction with the atmosphere and the ocean within a two-dimensional continuum frame. The coupled ECBILT-CLIO model further includes a terrestrial vegetation component that is represented by VECODE (Vegetation Continuous Description model). The VECODE model consists of three sub-modules: a vegetation structure model that calculates plant functional type fractions (PFTs), a biogeochemical model that estimates net primary productivity (NPP) and a vegetation dynamics model. With these sub-structures, dynamic vegetation cover is obtained in terms of the PFT (such as desert, trees and grasses) in each land grid cell, according to their climate variables (Brovkin et al. 1997). The cryosphere component is the three-dimensional thermo-mechanical AGISM (Antarctica and Greenland Ice-Sheet Model) (Huybrechts, 2002). AGISM simulates dynamic ice sheets by taking into account the ice flow, the solid Earth response and the mass balance at the ice-sheet surface and ice–ocean interface. The terrestrial carbon cycle on land is represented in VECODE by dynamic carbon pools on an annual time step. The oceanic carbon cycle is represented by LOCH (Liege Ocean Carbon at Heteronomous), a three-dimensional model that simulates carbon exchange between the atmosphere and ocean, and also the biological pump processes (Mouchet et al. 1996). However, it is important to note that LOCH and AGISM were not activated in our version of LOVECLIM. Accordingly, the ice-sheet configuration and greenhouse gases are prescribed based on reconstructions, as explained in Figure 3. A more detailed description of LOVECLIM can be found in Goosse et al. (2010a).

The reasons for performing Holocene simulations with the LOVECLIM climate model are multiple, mainly including its reasonable performances for the present-day climate, its capability to simulate climate change in the past and its computational efficiency, which allows several thousands of years of transient simulation be performed in a reasonable time. The performance of LOVECLIM has been systematically examined by Goosse et al. (2010a), who found that the model simulates the modern climate reasonably well, despite some minor biases in the tropics. To be more specific, the key variables of the climate system in the simulation generally agree with the observations. The model can reproduce the main characteristics of the observed surface temperature distribution. For instance, the LOVECLIM results are in good agreement with the observed location of the 0 °C isotherm, even though the 25 °C isotherm is located too far away from the equator with an overestimated temperature. LOVECLIM simulates reasonably well the large-scale pattern of near-surface circulation. The key patterns of pressure fields in the model roughly agree with observations, such as a decrease of 800 hPa geopotential height with latitude in the N Atlantic and N Pacific, despite underestimated gradients around the Aleutian low. The sea ice extent in LOVECLIM agrees more closely with observations in the N Pacific than in the N Atlantic, since the seasonal change produced by the model is too weak in the Baffin Bay and the Labrador Sea. Finally, the strength of the ocean circulation in LOVECLIM shows relatively good agreement with data-based estimates and with other models (Ganachaud & Wunsch 2000; Goosse et al. 2010a). For instance, the maximum AMOC strength reaches 22 Sv (Sverdrup = 1×10^6 m s^{-1}), with deep convection occurring both in the Greenland-Norwegian Sea and the Labrador Sea. These values are in relatively good agreement with the data-based estimates and at the high end of the values given by other models (Ganachaud and Wunsch 2000; Gregory et al. 2005; Goosse et al. 2010a).

LOVECLIM is capable of simulating plausible climates in the past, which is helpful to explain the climate variability discovered in proxy records. For instance, it has been used in PMIP2 and PMIP3 studies to investigate the climate evolution during the LIG (Bakker et al. 2014; Loutre et al. 2014), LGM (Roche et al. 2012), and the Holocene (Renssen et al. 2009, 2012; Blaschek & Renssen 2013). Climate sensitivity to radiative forcing in LOVECLIM is about 2 °C (i.e. a sea surface temperatures increase by 2 °C after 1000 yr with a doubling of the CO_2 concentration, according to Goosse et al. 2010a), which is at the lower end of GCM estimates (Flato et al. 2013). The sensitivity of the AMOC in LOVECLIM to perturbations of idealized freshwater fluxes (FWF) agrees with other models. For instance, in response to a 0.1 Sv freshwater release in the N Atlantic, the AMOC deceases by about 20–30%. This agrees well with the ensemble mean which includes several AOGCMs and EMICs (Stouffer et al. 2006).

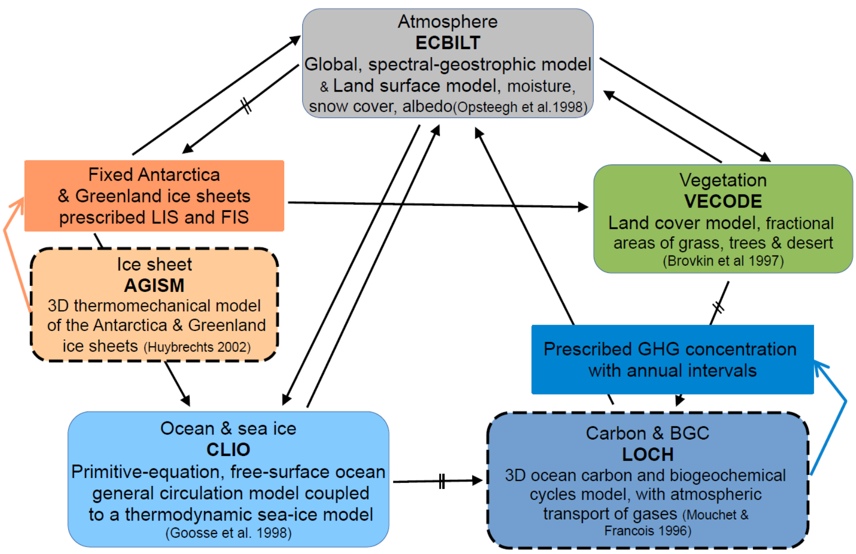

Figure 3. The climate components of the LOVECLIM model (adapted from Goosse et al. 2010a). Filled arrows depict the interactions between these components. Dashed borders and crossed-out arrows indicate inactive components and associated interactions, and open colour arrows lead to corresponding procedures in the present study.

2.2 Sensitivity experiments

Sensitivity experiments are a set of simulations performed with different forcings and/or boundary conditions relative to a control experiment. Systematic comparison of these sensitivity experiments can improve our understanding of the variability of the climate system and can accumulate our knowledge on the sensitivity of the climate to these changes in forcings and boundary conditions. This understanding is essential for the development of climate theory. The sensitivity experiments have huge potential to explore climate change in the past when no observations are available. Uncertainties in model-based palaeoclimate estimations are partially caused by uncertain forcings, which can be evaluated through sensitivity test. The first step is to perform sensitivity experiments with all possible forcing-scenarios. After this, simulated variables can be compared with evidence that is independent of the forcing data used in order to determine the most likely scenario. For instance, previous studies have investigated model responses to varying forcings, such as GHG and ORB, under the different benchmarking periods of the LGM (representing the latest glaciation) and MH (a warm period with abundant proxy data), and explored the potential role of different processes and feedbacks (e.g., Hewitt & Mitchell 1997; Kutzbach & Liu 1997; Weaver et al. 1998; Claquin et al. 2003). Comparison of these experiments provides insights into the potential mechanisms behind various responses and into functioning of the climate system under different conditions. Another good example of sensitivity experiments is related to multiple FWF scenarios. Numerous studies have examined the sensitivity of the atmosphere-ocean system to different FWF scenarios, improving understanding of the feedbacks and underlying mechanisms of abrupt climate change events that punctuated the warming trend of the last deglaciation (Teller et al. 2002; Roche et al. 2007).

In Publication I, we conducted sets of simulations with different combinations of climate forcings, as outlined in Figure 4. First, two equilibrium experiments for 11.5 ka, namely OG11.5 and OGIS11.5, were performed with identical ORBGHG forcings, and additionally including the effect of melting ice sheets in the latter experiment (OGIS11.5). The spatial patterns of the early-Holocene climate and the underlying mechanisms were analysed by comparing different responses of associated variables (e.g. surface albedo, atmospheric circulation and AMOC) in these two experiments. Moreover, the temporal features of Holocene temperature were tracked by analysing transient Holocene simulations. Similarly to the equilibrium experiments, the first transient simulation, namely ORBGHG, was forced only by the corresponding ORBGHG forcing. Another two sets of simulations were conducted by further including the ice sheets. These two simulations, namely OGIS_FWF-v1 and OGIS_FWF-v2, tested two possible FWF scenarios, considering the large uncertainties in FWF discharges. The differences between them were a larger FWF release from the Greenland ice sheet and a faster FWF from the FIS in version 2 than in version 1 (Fig. 3 of Publication I). By comparing these simulations, the associated processes and feedbacks regarding the responses of the ocean–atmosphere system to these forcings and how these responses ultimately influenced the Holocene temperature history were investigated.

2.3 Inter-model comparisons

Uncertainties in simulated palaeoclimates can sometimes be rather large, especially for transitional climate periods such as the early Holocene. The magnitude of the uncertainty in estimations of paleoclimate depends on not only on the forcings used in the simulation, but also on the model uncertainty, such as the physical principles, complexity and resolution (Hawkins & Sutton 2009). Different models may have slightly different sensitivity estimates under certain climate conditions, as various feedbacks may be involved at different spatial and temporal scales. Multi-model comparisons provide the option to validate the reliability of our model performance and increase the confidence in simulations. The reliability of the model and simulations will increase if independent models consistently indicate the same or similar results. Otherwise, varying model performances or even biases will be implied.

PMIP has performed a wide range of inter-

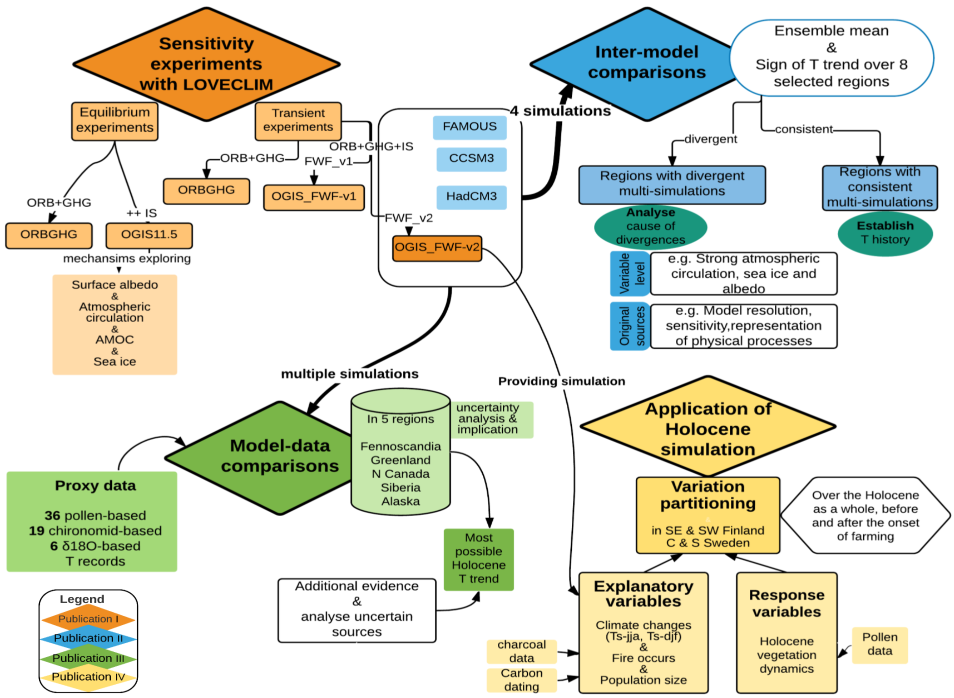

Figure 4. Schematic illustration of the approaches employed in the thesis. The four different colour tones illustrate the main procedures applied in the four publications, with orange representing Publication I, blue indicating Publication II, green depicts Publication III and yellow denoting Publication IV.

model comparisons that cover different periods with various climate forcings or processes. Stage 1 of PMIP was designed to test the climate models (atmosphere component) in the context of the LGM by investigating the climate response to large ice sheets, cool oceans and lower GHG concentrations, and also to investigate vegetation distribution during the MH (Harrison et al. 1998; Braconnot et al. 2000). Stage 2 applied coupled ocean–atmosphere and ocean–atmosphere–vegetation models to examine the role of oceans and vegetation or the land surface (Harrison et al. 2003; Braconnot et al. 2007; Jiang et al. 2008; Wohlfahrt et al. 2008). Stage 3 was extended to include the LIG and the later Holocene (Braconnot et al. 2012) and transient simulations (130–125 ka & 8–2 ka) (Bakker et al. 2014). Overall, PMIP inter-model comparisons have mainly been conducted for certain periods with snapshot experiments, and the currently available transient inter-model comparisons mainly focus on relatively warm periods (e.g. the last 1 kyr and the LIG (Bothe et al. 2013; Eby et al. 2013; Bakker et al. 2014). For the Holocene, the current multi-model comparisons studies, however, are limited to the MH by analyses of equilibrium simulations (Braconnot et al. 2012) and short transient simulations during the last 2 ka (Bothe et al. 2013; Eby et al. 2013). This implies that multi-model comparisons for the early-Holocene climate remain unexamined.

To evaluate simulations of the Holocene, we compared four simulations performed with LOVECLIM, CCSM3, FAMOUS and HadCM3 in Publication II, as explained in Figure 4. More information on these models and associated simulations is provided in Tables 1 and S1 of Publication II. The simulations were firstly compared in the time windows of 11.5 and 6 ka in order to determine the basic spatial pattern. To track the temporal variability of the simulations, we compared the transient temperature trends among these simulations over an array of regions to identify the regions where the multi-model simulations are consistent and where they are not. Furthermore, we diagnosed the models' variables when mismatches were found, to determine the potential causes of these inconsistencies. After this, these multi-model discrepancies were analysed at a deeper level (e.g. why models overestimated or underestimated these variables from the perspective of the model principles) by further referring to relevant studies by others.

2.4 Proxy records and model-data comparisons

Proxy-based reconstruction is another well-developed approach for investigating the palaeoclimate, in addition to numerical model simulations. Climate parameters (e.g. temperature) can be reconstructed based on various proxies (e.g. pollen and chironomids) through transfer functions that are based on the relationship between the climate and biological datasets. Transfer-function-based reconstructions facilitate the conversion of past fossil assemblages to quantitative temperature, precipitation and other climate parameters, assisting direct comparison with model results. Considerable improvements have been achieved with this proxy-based method. In particular, an abundance of climate records are available for the Holocene (e.g. as collected in Shakun et al. 2012; Marcott et al. 2013; Sundqvist et al. 2014), which provides multiple lines of evidence for the climate history. However, proxy data still contain inherent uncertainties or even biases. Comprehensive comparisons of proxy-based reconstructions with model simulations provide opportunities to detect and investigate these inherent uncertainties, and thus are considered to facilitate better climatic interpretations of proxy results. Meanwhile, although climate models are useful for piecing together separate of proxy records and understanding the mechanisms behind climate change, the development of climate models still requires proxy data under certain circumstances. In particular, proxy data are needed to validate the climate models at an early developmental stage (Braconnot et al. 2012) and to evaluate the simulations when multiple models perform differently. This combination can add value to our understanding of climate mechanisms through reciprocal validation of these two approaches, as the principles of model simulation and proxy-based reconstruction are different and independent.

Recent progress on proxy-based reconstructions and newly established databases provided the basis for a systematic investigation of Holocene temperature evolution. For instance, based on the Holocene proxy database (Sundqvist et al. 2014), temperature changes in the north Atlantic–Fennoscandia (Sejrup et al. 2016), Alaska (Kaufman et al. 2016), the Canadian Arctic and Greenland (Briner et al. 2016) have recently been examined. Studies have also compared model results and proxy data (e.g. Liu et al. 2014 who focused on large-scale climate change), but comparisons between multi-model simulations and proxy data on a detailed sub-continental scale have not yet been conducted. Meanwhile, considering the transient features and high uncertainty of the early-Holocene climate, comparisons of climate model results with the proxy data at a sub-continental scale are needed to evaluate Holocene simulations and to improve our understanding of Holocene climate evolution.

Publication III presents a model–data comparison (i.e. comparison between composite reconstructions and multiple model simulations) over different regions for Holocene temperature trends, as illustrated in Figure 4. We first selected 36 pollen-based and 19 chironomid-based temperature records, and six $\delta^{18}O$ records of ice cores from the NH high latitudes with consideration of the location, time frame, temporal resolution and dating interval (as explained in Publication III). These individual proxy records were compiled into five composite regional reconstructions for Fennoscandia, Greenland, north Canada, Alaska and high-latitude Siberia. We compared these composite reconstructions with the model simulations (ensemble mean and spread of multi-simulation) over different regions in terms of Holocene temperature trends. In addition, potential uncertainty sources were also analysed from both model simulations and proxy data perspectives, and then used to identify the most plausible climate history of different regions with the aid of additional evidence when available.

2.5 Variation partitioning

Variation partitioning is a statistical method to partition the variation of a given response variable into a set of explanatory variables (descriptors) whose interaction often results in an overlaid effect (Borcard et al. 1992). The method was originally used to determine how much variation in species abundance data can be simultaneously related to the explanatory variables of environmental and spatial components (Borcard et al. 1992). Recently, it has been used in palaeoecology to investigate vegetation history. For instance, it has been employed to investigate the relative contribution of the climate, forest fires and human population size to variation in Holocene vegetation dynamics (e.g. Reitalu et al. 2013; Kuosmanen et al. 2015, 2016).

With the final melting of the FIS by around 10 ka, the development of vegetation development in Fennoscandia was one of the critical processes that occurred over the course of the Holocene, which is relevant for multiple factors. In Publication IV, we use the statistical method of variation partitioning to assess the relative importance of climate, forest fires and human population size on the variation in actual vegetation cover of different PFTs (Fig. 4). The analysis uses PFTs derived from the REVEALS (Regional Estimates of VEgetation Abundance from Large Sites model) dataset as response matrix, and climate, human population size and forest fires as explanatory variables. The REVEALS model (Sugita 2007) was used to reconstruct the regional plant abundances from pollen percentage data, which reduces biases in pollen analysis caused by inter-taxonomic differences in pollen productivity, dispersal and depositional characteristics, as well as basin size. The climate data were from the LOVECLIM simulation forced by the OGIS_FWF-v2 scenario, which is a realistic scenario regarding early-Holocene warming and timing of the HTM. The frequency of forest fires was reconstructed based on sedimentary charcoal records from 19 lakes, which provide a signal of Holocene biomass burning across the study region. The human population size variable was derived from the ^{14}C dating results of archaeological findings obtained from the archeological sites. Variation partitioning analysis with climate, human population size and fire as explanatory variables was first performed for the whole study period in all regions (S & C Sweden, SW & SE Finland). To further track the transient changes in the contributions of these factors, variation partitioning was performed in two sub-periods, before and after the onset of farming, which are regionally dependent. A moving window approach (Reitalu et al. 2013) was applied to S Sweden and SE Finland to assess how the relative importance of climate and human population size changed through time.

3 Summary of the original publications

This thesis focused on the Holocene climate history by employing various methods in different publications. The overall goal was to obtain a better understanding of ice sheet–ocean–atmosphere processes during the early Holocene transition and thus a reliable estimation of Holocene climate, as that has potential implications for ongoing climate change. Publication I examines the effects of different forcings on the climate and tests two FWF scenarios with the LOVECLIM model, while Publication II compares four sets of model results to evaluate the reliability of Holocene climate simulations. In Publication III, we use proxy data (including pollen, chironomids and $\delta^{18}O$ in Greenland) to further evaluate the simulations and analyse the most likely Holocene climate history over five regions at high latitudes. Finally, Publication IV investigates the role of climate together with human disturbance and forest fires in Holocene vegetation dynamics with variation partitioning statistical analysis. The main findings are summarized below by addressing the research questions presented in the introduction.

3.1 Effects of the ice sheets on Holocene climate change in NH extratropics (Publication I)

The main question addressed in Publication I is 'How does the Holocene climate in the NH extratropics respond to the main forcings?'. To answer this question, the spatial pattern and temporal characteristics of temperature were analysed in climate simulations performed with the LOVECLIM model.

3.1.1 Spatial patterns of anomalous temperature at the onset of the Holocene

a) The main spatial characteristics of temperature at the onset of the Holocene under different forcings, especially with the decaying ice sheets

At the onset of the Holocene, the climate shows latitudinal temperature variations in response to the ORB and GHG, and ice sheets forcing induces spatial heterogeneity in the simulated temperature (Fig. 5A). The results of OG11.5 (forced with ORBGHG) reveal that the positive ORB forcing overwhelms minor negative GHG radiative anomalies, and thus causes positive summer temperature anomalies of 2–4 °C over most of the extratropical continents relative to 0 ka, with a maximum deviation of 5 °C over the central parts of the NH continents. The simulated temperatures also show latitudinal patterns, especially in winter. A north–south gradient exists, going from a clear warm anomaly at high latitudes to areas further to the south with less warming or even cooler conditions. The OGIS11.5 simulation (forced by full forcings by additionally including the ice-sheet forcing) suggests clear spatial heterogeneity of climate responses at 11.5 ka. Ice sheets induce strong summer cooling over ice-covered areas, which contributes to these heterogeneous patterns. In particular, the most distinct feature is a thermally contrasting pattern over North America with simulated temperatures being around 2 °C higher than at 0 ka over Alaska, contrasting with more than 3 °C lower temperatures over most of Canada. Relatively small temperature anomalies can be observed in other regions, for instance slightly negative temperature anomalies in Siberia and small positive values in the Arctic Ocean.

b) The possible mechanisms behind these simulated temperatures

The maximum temperature reduction in the simulation performed with full forcings was found over the centre of the LIS. Although the ORB and GHG forcings resulted in overall higher temperature than the pre-industrial era due to overwhelming positive anomalies of the ORB, the ice-sheet forcings primarily induced cooling of the climate. The underlying mechanisms of this cooling include enhanced surface albedo over ice sheets, anomalous atmospheric circulation, reduced AMOC and relevant sea-ice feedbacks. The anomalous atmospheric circulation and AMOC can influence the regions beyond the boundary of the ice sheets, in comparison to simulations without the ice sheets. The cooler climate over the northern Labrador Sea and the North Atlantic was related to both reduced northward heat transport due to reduced AMOC strength and enhanced sea-ice feedbacks, which was induced by the freshwater from the melting ice sheets. A small summer temperature anomaly in Siberia is probably because the positive insolation anomaly was offset by the cooling effect of the high albedo associated with the relatively extensive tundra cover in the early Holocene. The relatively warm early-Holocene climate in Alaska was mainly due to increased

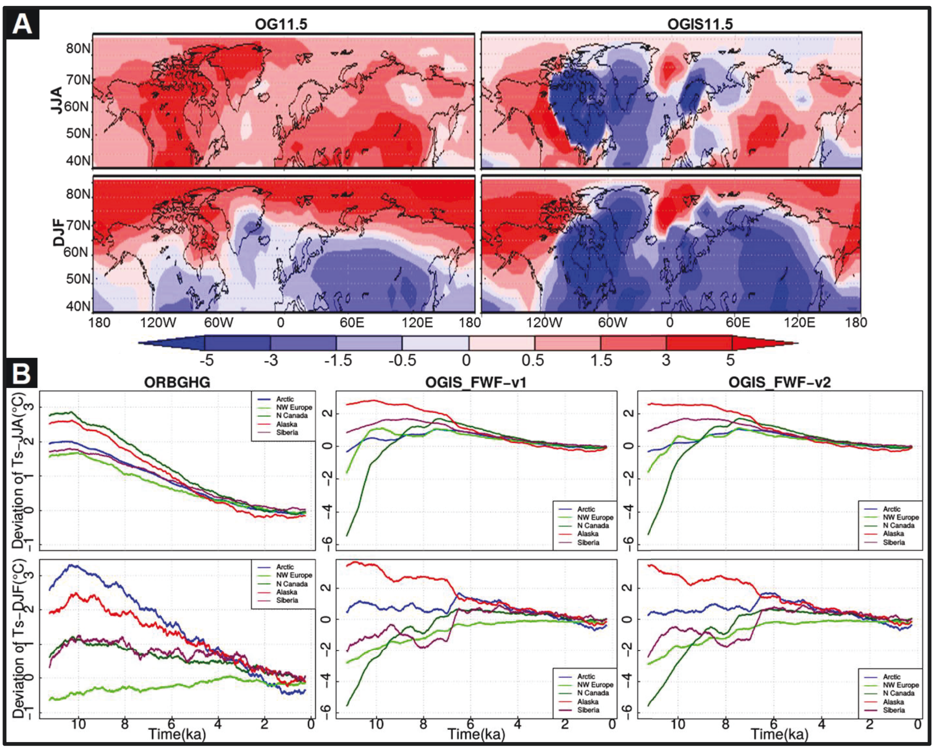

Figure 5. (A) Simulated temperature anomalies at 11.5 ka relative to 0 ka. (B) Holocene temperature evolution in simulations performed with different forcings, with colored lines indicating different regions.

southerly winds induced by the LIS. Overall, the combination of ORB, GHG and ice-sheet forcings at 11.5 ka resulted in cooling over most of regions, but over most of the Arctic Ocean the climate was warmer than preindustrial, because strong insolation anomalies overwhelm effects of a weakened AMOC induced by meltwater releases.

3.1.2 Temporal trends of Holocene temperatures

a) Temporal trends of Holocene temperatures over different regions

The Holocene temperature evolution, especially the early-Holocene temperature trends, reveals geographical variability (Fig. 5B). In Alaska, the climate has constantly cooled throughout the Holocene due to the decreasing insolation and anomalous atmospheric circulation. In contrast, northern Canada experienced strong early-Holocene warming with an overall warming rate of over 1 °C kyr^{-1}, and this warming lasted until 7 ka. Although different forcings and mechanisms played different roles in northwestern Europe, the Arctic and Siberia, the overall warming effect was similar for these regions, with a rate of around 0.5 °C kyr^{-1}.

b) Contribution of different forcings to these temperature trends

The ORB and GHG forcings lead to an overall decreased temperature trend with a larger magnitude in summer than in winter, which shows latitudinal patterns. The cooling effects of ice sheets induce the spatial heterogeneity in the Holocene temperature trends. The climate-ocean system during the early Holocene is also sensitive to FWF forcing, as indicated by the responses of the NH sea-ice area and AMOC to different FWF scenarios (Fig. 14 in Publication I). Different FWF forcings, including the intensity, duration and the location of freshwater release, can result in varied conditions in the oceans. For instance, the enhanced freshwater influx from the GIS and the redistributed meltwater from the FIS caused an alteration in the surface ocean freshening in the Nordic Seas and associated changes in temperature.

c) Which FWF scenarios provide a plausible Holocene temperature evolution regarding the early-Holocene warming and timing of the HTM?

Clear differences between two scenarios are found in NW Europe, where abundant proxy records are available. The comparison of Holocene temperatures regarding the early-Holocene warming and HTM over northwestern Europe with proxy records suggests that the updated FWF scenario (FWF-v2, with a larger FWF release from the Greenland ice sheet and a faster FWF from FIS) represents a more realistic climate. The OGIS_FWF-v1 simulation may underestimate the cooling effect of ice-sheet melting, which indicates two peaks at around 10 and 7 ka in the temperature evolution over northwestern Europe. High temperatures at 7 ka are recorded in proxy-based reconstructions, but no warm peak at 10 ka has been observed in pollen-based reconstructions (Mauri et al. 2015) in Europe. In contrast to OGIS_FWF-v1, the OGIS_FWF-v2 simulation produced a warming trend that is consistent with the highest temperature around 7 ka. Moreover, the OGIS_FWF-v1 produced a temperature decrease between these two peaks, whereas the proxies indicated a persistent temperature increase by 10 ka, followed by a more gradual warming (Brooks et al. 2012). Therefore, from the viewpoint of temperature evolution in northwestern Europe, the OGIS_FWF-v2 simulation represents a more realistic climate than OGIS_FWF-v1, which also demonstrates that the existing uncertainties in the reconstructions of ice-sheet dynamics can be evaluated by applying different freshwater scenarios.

3.2 Inter-model comparisons of the Holocene temperature trends (Publication II)

3.2.1 What are the consistencies and divergences among these Holocene simulations?

The LOVECLIM simulation OGIS_FWF-v2 was compared with similar simulations performed with CCSM3, FAMOUS and HadCM3. The multiple simulations suggest generally consistent temperatures in the NH extratropics as a whole, with better agreements in summer than in winter. On a regional scale, reasonably consistent temperature trends are found over the regions 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 of early-Holocene warming, HTM and a gradual decrease toward 0 ka (Fig. 6A). Within the above general patterns, the warming extent and rate during the early Holocene varies across these 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 lower warming magnitude among these regions, 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 the structural and parametric uncertainties across the individual models are smaller than climate forcing effects and internal climate variability in these regions.

By contrast, large inter-model variations exist in the regions where the cooling effects of the ice sheet were relatively weak, such as in Alaska, the Arctic and E Siberia, through a series of indirect influences (Fig. 6B). In particular, during the early Holocene, the signals of multi-model simulations are incompatible in winter, when both positive and negative early-Holocene anomalies are suggested by the different models. At 11.5 ka, the spread of the simulated temperatures anomalies ranges from +2 °C in LOVECLIM to -6 °C in FAMOUS in Alaska. This distinct multi-model difference in winter thus amounts up to 8 °C, which is considerably larger than in summer when the inter-model difference is only 1 °C. Over the Arctic, the discrepancies between the simulations also primarily exist in winter. At 11.5 ka, the winter temperature anomaly is slightly positive in the LOVECLIM simulation, 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. Over the course of the Holocene, the simulations show decreased trends of summer temperatures over E Siberia, with the largest decrease (more than 4 °C) in CCSM3, which contributes to the large variation among these simulations. In winter, over 2 °C Holocene cooling is simulated by LOVECLIM, contrasting with up to 5 °C warming in FAMOUS.

3.2.2 What climate variables in models cause these inter-model divergences?

Divergent temperatures in the above regions can be attributed to inconsistent responses of model variables, such as southerly winds, surface albedo and sea ice. The strong southerly winds induced by the LIS over Alaska and part of E Siberia result in an anomalously warm climate (more than 3 °C) in LOVECLIM. During the early Holocene, higher summer temperatures (1–

2 °C) over E Siberia in CCSM3 than in other models are caused by a strong difference in surface albedo (more than 20%) between 11.5 and 0 ka, which is ultimately associated with an unrealistically high albedo at 0 ka. The wide spread of simulated Arctic temperatures in winter may be attributed to inter-model differences in sea ice. Extended early-Holocene sea ice cover in the Arctic Ocean in FAMOUS 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, which contrasts with only limited sea ice changes between the 11.5 and 0 ka in CCSM3. Relatively low winter temperatures in HadCM3 over the Arctic are probably attributable to the thick sea ice at 11.5 ka, which can block the ocean heat release into the atmosphere in comparison with LOVECLIM.

3.2.3 Where do these discrepancies originate?

The multi-model comparisons reveal that divergent responses in climate variables across these simulations are partially caused by model differences (e.g. different model physics and resolution). The newly adopted formulation of a turbulent transfer coefficient in CCSM3 causes an overestimated albedo over Siberia at 0 ka (Oleson et al. 2003; Collins et al. 2006), which leads to stronger early-Holocene warmth (temperature anomalies) than in others. Moreover, the relatively simplified sea ice representation in FAMOUS probably leads to

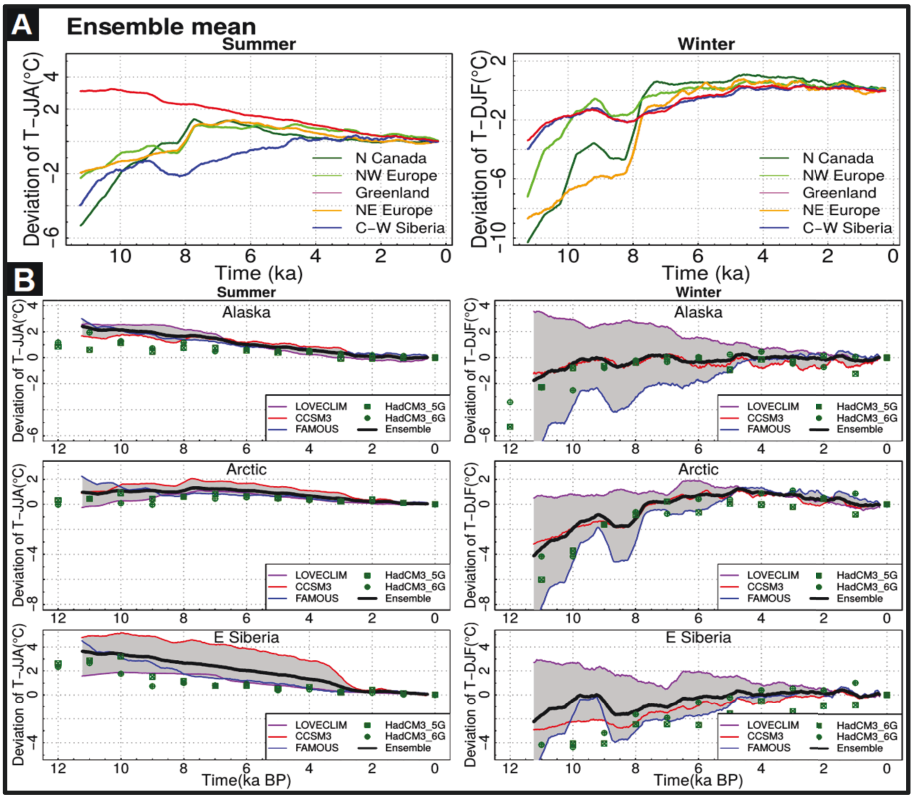

Figure 6. (A) Multi-model ensemble mean of simulated Holocene temperatures (in °C) over the regions where multiple simulations are broadly consistent. (B) Simulated temperatures over the regions where temperatures are less consistent across the simulations, with grey indicating the ensemble range and thick black lines depicting the ensemble mean.

overestimated sea ice cover in the Arctic Ocean. Furthermore, the coarse vertical resolution in LOVECLIM might introduce strong responses in atmospheric circulation over Alaska. Apart from the different models, the transient feature of the early-Holocene climate driven by the retreating ice sheets also influences the inter-model comparisons. In particular, large uncertainty exists in the FWF forcing, which has a major impact on the early-Holocene climate between 11.5 ka and 7 ka.

3.3 Model–data comparisons of Holocene temperatures at NH high latitudes (Publication III)

3.3.1 To what degree do proxy-based reconstructions agree with model results?

The consistencies between the model simulations and proxy-based reconstructions are region-dependent. In Fennoscandia (Fig. 7), the ensemble mean summer temperature of the simulations is consistent with pollen-based reconstructions, although there are minor differences between individual simulations (±0.5 °C) and proxy records (about 2 °C before 8 ka). The simulations and composite pollen-based reconstructions reveal a clear early-Holocene warming, although a stable early-Holocene temperature is indicated by chironomid data. The simulated annual temperature and δ^{18}O-based climate reconstruction in Greenland (Fig. 7) are consistent well regarding the Holocene trend. The temperature at 11.5 ka is 6 °C lower than at 0 ka, followed by warming, and reaching a maximum at ~7 ka. The only exception is that δ^{18}O data suggest a strong warming of over 8 °C by 9 ka, leading to an earlier temperature peak than in the simulations. The temperatures based on borehole measurements show a weaker cooling at the onset of the Holocene than in the δ^{18}O data, which agrees better with the simulated temperatures. However, the temperature during the HTM in borehole data was up to 2.5 °C higher than 0 ka, which slightly diminishes its agreement with the simulations. The simulations and pollen data show similar summer temperature trends over the course of the Holocene in northern Canada (Fig. 7). The ensemble mean of simulations indicates a warming of 4 °C between 11.5 and 8 ka with an inter-model spread of ~3 °C, and the compiled pollen data extending back to 10 ka suggest 1~2 °C warming until ~7 ka with a spread range of 4 °C. From 7 ka onwards, simulations and pollen data are consistent, showing a cooling temperature trend of ~1 °C toward 0 ka.

In contrast, the simulated summer temperature and proxy-based reconstruction are mismatched in Alaska with opposite signs (Fig. 7), particularly in comparisons of the simulations with pollen-based temperature reconstructions. The ensemble mean of the simulations represents a 2 °C cooling trend during the Holocene with a range of 1–2 °C across individual models, while pollen data suggest a warming of 4 °C with a range of 2 °C. Meanwhile, the chironomid-based temperatures show an almost stable level with a 2 °C variation across individual records until 9 ka. Comparisons between the simulations and proxy-based reconstructions during the Holocene are generally noisy in high-latitude Siberia (Fig. 7), despite the simulations matching the chironomid results better than pollen data. The ensemble mean of simulations indicates a 2.5 °C cooling trend, with a similar range of inter-model spread from 4 °C in CCSM3 to 0.5 °C in LOVECLIM. In spite of the large spread in the chironomid records, the median of the dataset suggests a Holocene cooling trend of 1 °C, which is slightly less than in the ensemble mean of simulations. Pollen data indicate an increasing temperature until 7 ka, reaching a distinct maximum of +2 °C, and this warmth lasts until 4 ka, after which the temperatures decrease to the 0 ka level.

3.3.2 What are the potential sources of uncertainty in simulated temperatures and proxy-based reconstructions?

1) Potential uncertainties in proxy data & implications for the interpretation of proxy data

Multiple factors contribute to the uncertainty of proxy-based reconstructions, and the dominant sources of uncertainty vary between regions. First, different proxies and transfer functions can explain an important part of the mismatch between proxy-based temperature reconstructions. During the early Holocene, pollen- and chironomid-based temperature reconstructions are different in some regions. Apart from the complexity of reconstructing early-Holocene climate change, one important issue is the difference in the representation of seasonality in these proxies. The occurrence and abundance of chironomids in lakes are strongly determined by summer air and water temperatures (Heiri et al. 2014), while the pollen data can to some extent also reflect winter temperatures, especially for plant species with low cold tolerances. In some of the pollen-based temperature records from Alaska and Canada, the Modern Analog Technique (MAT) method rather than Weighted Averaging Partial Least Square regression and calibration (WAPLS) was

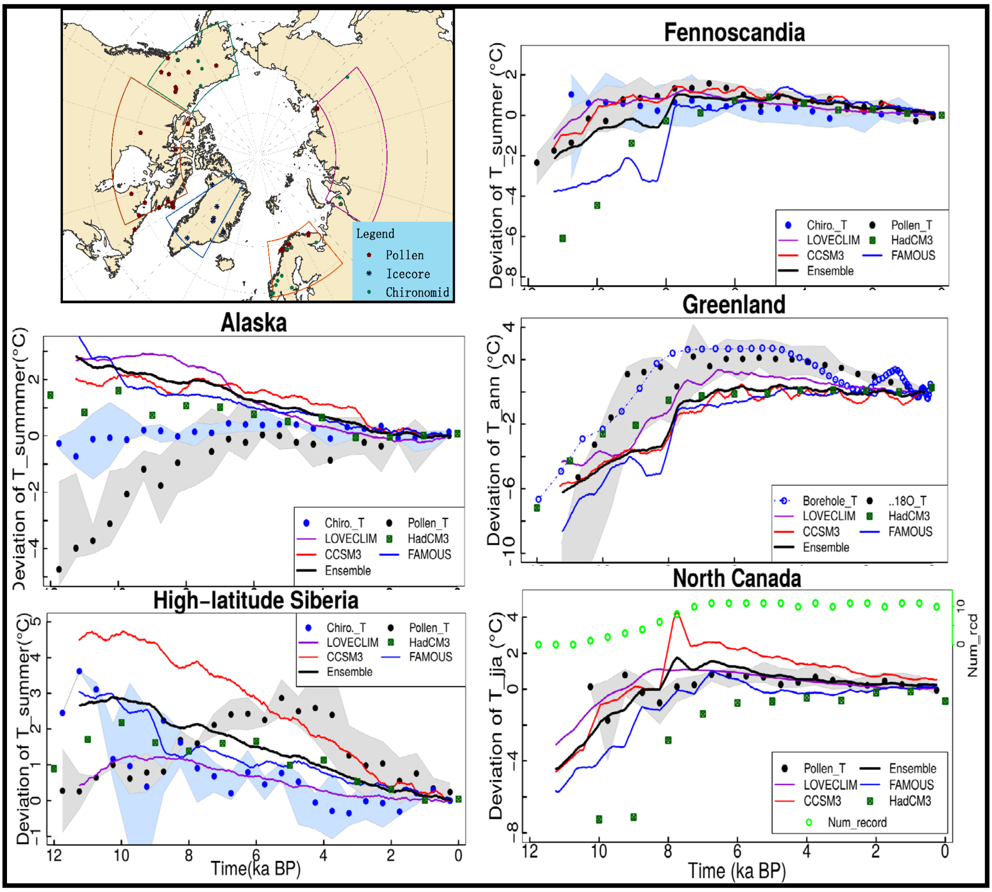

Figure 7. Locations of proxy data and Comparisons of simulated temperatures with proxy-based reconstructions in Fennoscandia, Alaska, Greenland, High-latitude Siberia and North Canada.

applied to quantitatively reconstruct the climate, which might explain some of the above model–data discrepancies. MAT and WAPLS each have their own merits and weaknesses, and one method can outperform the other for particular datasets, depending on differences in the training-set size, taxonomic diversity and complexity of the species–environment relationship (ter Braak & Juggins 1993; Telford & Birks 2005; 2009; Juggins & Birks 2012). Simple $\delta^{18}O$ conversion to temperature in Greenland may induce uncertainty, as other changes, such as atmospheric circulation (Charles et al. 1994), the seasonality of precipitation (Fisher et al. 1983) and its variation through time (Cuffey et al. 1995), affect the isotopic composition, in addition to the local environmental temperature. The borehole temperatures might include regional signals in individual records, as they are down-core measurements of the GRIP ice core (Dahl-Jensen et al. 1998).

Secondly, the representativeness of climate signals in proxies might be disturbed by other factors. Disequilibrium vegetation dynamics during the transient early Holocene (e.g. in N Canada) may influence the accuracy of pollen-based temperature records. It has been suggested that the post-glacial migration of trees to deglaciated regions may have been constrained by their limited speed dispersal and population growth rates (Birks 1986a; Webb 1986). Such a time lag in their migration history would prevent the pollen records from tracking rapid climate changes. Thus, it is possible that the considerable temperature rise indicated by the models was not fully reflected in pollen-based climate reconstruction, leading to an underestimation of the early-Holocene warming in pollen data. Non-climatic factors following the last deglaciation influence climate records derived from chironomids data. Apart from temperature, the inconstant hydrobiological and limnological factors, such as nutrient availability, trophic state, and dissolved and total organic carbon in lakes, influenced chironomid distribution and abundance, and thus potentially caused bias in the assumed relationship between chironomids and temperature (Brooks & Birks 2001; Velle et al. 2010). In Fennoscandia, the absence of early-Holocene warming in the chironomid data is probably associated with the influences of non-climatic factors. No correction for palaeo-topography was taken into account in reconstructed temperatures by assuming that the effects of ice thickness and post-glacial rebound are of the same order and they are roughly balanced out, which could induce some uncertainties in model–data comparison. The simulated temperature (e.g. in north Canada and Fennoscandia) may be influenced by palaeo-topographic changes due to the reducing thickness of the LIS and the undergoing post-glacial rebound.

Thirdly, statistical uncertainties exist in the composite reconstructions of some regions and also in selection processes. Each individual record carries some site-specific signals, which can be incorporated into compiled reconstructions when only a low number of proxy records are available. Given the extremely low number of records (only five) in Siberia, the reliability of the proxy-based reconstruction in high-latitude Siberia is statistically lower than in other regions. In addition, uncertainties related to a coarse temporal resolution might be induced in individual records when the expanded selection criteria were applied in record selection, such as in N Canada.

2) Potential uncertainties in simulations and implications for model simulations

Similarly to the proxy approach, multiple uncertainty sources exist in simulations, which also and show region-dependency. First, palaeo-topographic changes related to the retreating ice sheets and associated post-glacial rebound potentially induce uncertainties in the simulations over N Canada and Fennoscandia. Moreover, it is well known that simulating the temperature over the ice sheets is a challenge. The accuracy of climate simulation over ice sheets strongly depends on the model resolution, as a high resolution allows a detailed representation of topography and precise description of thermodynamics (Ettema et al. 2009). From this viewpoint, potential uncertainties in simulations could also be attributable to a slightly underestimated mid-Holocene warmth over Greenland in simulations. Uncertainty might also be included by taking the rectangular box of the simulations as a representative of the Greenland region, as simulated temperatures are sensitive to the southeastern coastal area. Additionally, various representations of physical processes in models might be overestimated for some parameters. In high-latitude Siberia, CCSM3 tends to simulate a high albedo bias at 0 ka in the later adopted formulation of the turbulent coefficient (Collins et al. 2006; Oleson et al. 2003). This high albedo leads to anomalously low temperatures at 0 ka, and a deceptively warm early-Holocene in Siberia (more than 4 °C warmer) in CCSM3, which is far above either pollen or chironomid-based temperature reconstructions, implying overestimated early Holocene warming in CCSM3.

3.3.3 What are the most probable Holocene temperature trends?

Additionally available Holocene data include glacier frequency data from Fennoscandia and frequency data for peatland initiation in Alaska, which can provide additional evidence to climate history. The glacier data (Nesje et al. 2009) also suggest a clear early-Holocene warming in Fennoscandia, which matches better with pollen-based reconstructions and model simulations than with the chironomid data. Therefore, multiple lines of evidence suggest a pronounced early-Holocene warming in Fennoscandia, which is followed by a gradual decrease temperature towards the preindustrial value. The high frequency of peatland initiation during the early Holocene in Alaska (Jones & Yu 2010) could result from a high temperature, high soil moisture, or large seasonality (Kaufman et al. 2004; Zona et al. 2009; Jones & Yu 2010). Accordingly, the Holocene temperature trend in Alaska is still inconclusive.

Overall, these comparisons of multi-model simulations with proxy reconstructions furtherconfirm the Holocene climate evolution patterns in Fennoscandia, Greenland and North Canada. This implies that the Holocene temperatures in these regions have been relatively well established, with a reasonable representation of the Holocene climate in the multiple simulations and a plausible explanation for the underlying mechanisms. However, the Holocene climate history and underlying mechanisms in the regions of Siberia and Alaska remain inconclusive.

3.4 Vegetation dynamics and the main drivers (Publication IV)

3.4.1 Contributions of various factors to vegetation dynamics

Variation partitioning analyses with climate, human population size and fire as explanatory variables were first performed in (C & S) Sweden and (SW & SE) Finland. The results reveal that over the study period as a whole (10–4 ka in Sweden, 10–1 ka in Finland), climate explains the highest proportion of variation in PFTs in all regions (Fig. 8). Winter temperatures have a significant effect on vegetation changes in all regions, while summer temperatures are significant in all regions except for C Sweden.

In S and C Sweden, the relative importance of winter temperatures in PFT variation is higher than that of summer temperature, while in SW and SE Finland, summer temperatures explain a higher proportion of the variation (Table 2 in Publication IV). Meanwhile, the variation explained by human population size and forest fires is relatively low, when considering the Holocene as a whole study period. Apart from the contribution of climate change, human population size has a significant contribution to vegetation changes in C Sweden and S Finland, whereas forest fires have a significant effect in S Sweden and SE Finland. Generally, climate, forest fires and human population size together explain over 75% of the variation in all these regions.

Before the onset of farming (at 6 ka in Sweden and 4 ka in Finland), climate explains most of the variation in PFTs, with 13% in S Sweden, 54% in C Sweden, 25% in SW Finland and 32% in SE Finland. Although winter temperatures explain a higher proportion of variation in PFTs than summer temperatures, both climate parameters have significant effects on changes in vegetation dynamics in all regions (Fig. 8). Forest fires have a significant effect on vegetation dynamics in all regions except in C Sweden, and the highest contribution is found in SW and SE Finland, with up to 4%. The joint effects of climate and forest fires explain relatively a high proportion of the variation in vegetation, with 22% in S Sweden, 25% in SW Finland and 20% in SE Finland. The contribution of human population size to variation in vegetation is insignificant in these regions, except in SW Finland where 8% of the variation in vegetation can be explained by human population size. Overall, climate, forest fires and human population size explain at least 70% of the variation in vegetation composition in the studied regions.

After the onset of farming, human population size explains the highest amount of the variation in PFTs, with 23% in south Sweden and 22% in southwest Finland (Fig. 8). The contribution of climate to the vegetation dynamics in all regions is lower than during other periods, even though summer temperatures explain a higher proportion of variation than winter temperatures (Table 2 of Publication IV). Although the influence of forest fires on the variation in vegetation is minor, 10% of the variation in vegetation composition is explained by their joint effects with human population size in southeastern Finland. Overall, a relatively high proportion of the variation in vegetation is left unexplained after the onset of farming than in other periods in all regions.

3.4.2 Transition of the main drivers of vegetation dynamics

Variation partitioning reveals that a shift from natural (climate) driven vegetation dynamics towards human-dominated vegetation occurred during the Holocene. According to our study, Mesolithic populations did not significantly affect the vegetation dynamics in Fennoscandia, and climate was the main driver of change at that time. Neolithic populations, however, had major effects on vegetation dynamics and human population size became a more important driver of vegetation change than climate. A transition from climate- to human-driven Holocene vegetation dynamics in Fennoscandia can be estimated to have occurred at 7–6 ka in Sweden and 4–3 ka in Finland. There is a clear regional dependency of changes caused by the human population. In particular, the impact of human population size is higher in south Sweden and southwest Finland, suggesting more intense land use in these regions. The low importance of human impacts in central Sweden and southeast Finland might be related to the higher forest cover in these regions and to the smaller scale and more scattered farming compared to S Sweden and SW Finland.

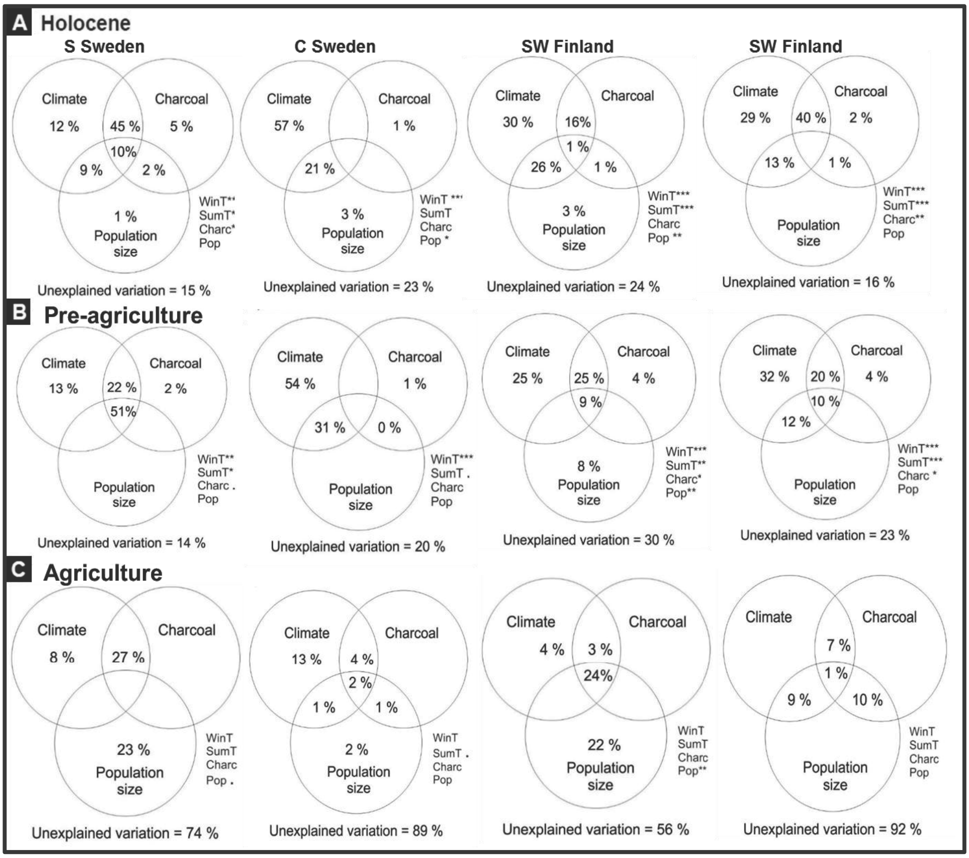

Figure 8. Relative importance of climate, forest fires and human population size to vegetation dynamics. The three panels present the results during the Holocene, pre-agricultural, and agricultural period; and from right to left representing S Sweden, C Sweden, SW Finland and SE Finland.

4 Remaining issues and outlook

Although this thesis provides insights into the transient early Holocene climate evolution by employing various approaches, several issues remain unsolved and/or require additional research. Based on current understanding, including the results of this thesis, below are listed those issues considered as deserving more attention and further investigation. However, this list is not intended to be complete

(1) On the ice sheets and their influence on the climate (focus on the early Holocene)

Dynamic ice sheets serve both causes and effects of climate evolution, with multiple glacial–interglacial cycles being experienced in geological history. On the one hand, the existence, building-up or decay of ice sheets are results of a changing climate. On the other hand, dynamic ice sheets exert multiple impacts on the climate system. The thesis investigated the complex response of the atmosphere–ocean system to melting ice sheets during the early Holocene, including their extent, topography and associated melting release with prescribed ice sheets based on ice-sheet reconstructions. In particular, responses of the climate–ocean system to different FWF scenarios were examined.

However, reconstructions of decaying ice sheets are still uncertain to some extent, especially regarding the spatial-temporal distributions of FWF. Therefore, more evidence on decaying ice sheets during the early Holocene is still highly demanded to improve our understanding of dynamic ice sheets and their responses to the environment. Meanwhile, with increasing knowledge of the interaction between the atmosphere and ice sheets, the next step would be apply the fully coupled (to a dynamical ice-sheet component) iLOVECLIM climate model (Roche et al. 2014) for simulating glacial-interglacial cycles, which can be constrained by increasingly available ice-sheet reconstructions. This coupling may provide better understanding of the interactions between dynamic ice sheets and the environment. The final goal is to disentangle the build up and decay of the ice sheets throughout the Earth's history.

(2) On inter-model comparisons

This thesis revealed that multiple factors at different levels can lead to divergences between simulations performed with different models. Different parameterizations and representations of physics in models are important sources of such divergences, which can be accessed by comparing the simulated modern climate with observations. Meanwhile, differences in experimental setups and forcings hampered our inter-model comparisons. In particular, non-uniform FWF across simulations has a major impact on the divergent early-Holocene climate (11.5–7 ka) among these simulations. Accordingly, we suggest that constructing a standardized FWF for the early Holocene would be advantageous for future inter-model comparisons.

(3) On the model–data comparisons

The model–data comparisons highlight that more work is required to reduce uncertainties from both simulation and proxy record aspects, especially in the case of the divergent temperature trends.Combinations of different proxies are expected to increase the accuracy of proxy-based reconstructions given the strengths and weaknesses of each type of proxy. However, mismatched temperature reconstructions can be yielded by different types of proxies, which challenges the integration of multiple types of proxies into one reconstruction. Systematic comparisons of different types of proxies are valuable to efficiently combine different proxy datasets. Spatial consistency between the simulations and proxy data is another critical issue given the spatial heterogeneity of the climate. Proxy data are regularly site-based and thus represent relatively small-scale climate conditions, although various spatial scales can be represented by different types of proxies and even within the same proxy. For instance, pollen data can represent vegetation composition at various spatial scales from local (stand scale) to regional, depending on the openness of the landscape and the size of the sedimentary basin (Parshall & Calcote 2001; Sugita 2007). By contrast, the simulated temperatures in a model are the average over an area covered by grid cells, representing the large-scale climate. Therefore, maintaining a consistent spatial coverage has a crucial role in model–data comparison, despite the challenges in determining the exact representative scale of a given proxy dataset.

(4) On further investigation of the transitional early-Holocene climate

The highly dynamic characteristics of the early-Holocene climate and its general warming pattern provide a potential analogue for current global change. Apart from the methods employed in this thesis (model simulations in Publication I, inter-model comparisons in Publication II and model–data comparisons in Publication III), the novel approach of data-assimilation would be another method that can be applied to investigate the Holocene climate. The data-assimilation approach makes use of proxy data in a statistical and dynamical way. Thus, compared with purely statistical approaches, data assimilation offers the potential advantage of ensuring dynamical consistency in the assessment of past climate change, despite its own limitations, such as the potential incompatibility between proxy data and model simulation due to the simplified physics of the model (Goosse et al. 2010b). This method has been applied to investigate model-data consistencies during the MH and also to Southern Hemisphere cooling from 10-8 ka (Mairesse et al. 2013; Mathiot et al. 2013). However, no study has investigated the NH extratropics climate for the whole Holocene yet. Increasingly available proxy data cover the whole Holocene provide the basis for applying data assimilation, which could ultimately improve knowledge of the early-Holocene climate.

(5) Problematic Holocene temperatures in Alaska and Siberia

According to this thesis, the Holocene temperature trends are still inconclusive in Alaska and high-latitude Siberia. In Alaska, multiple simulations are incompatible in winter: LOVECLIM suggests a decreasing Holocene temperature, while FAMOUS indicates an increasing trend. In summer, although multiple simulations consistently suggest a decreasing temperature trend, the pollen- and chironomid-based reconstructions are divergent. Additional peatland initialization data are available, but have multiple interpretations. Therefore, further investigation of the contribution of climate change to peatland initialization could provide evidence to narrow down the uncertainty of the Alaskan Holocene temperature. For high-latitude Siberia, a relatively wide spread was found in simulated temperatures in both summer and winter. Meanwhile, important issues exist in proxy-based reconstructions, such as the limited number of available records and the discrepancies between pollen-based and chironomid-based temperature reconstructions. Therefore, more proxy records are required to further examine their temperature evolution during the Holocene and differences between alternative types of proxy data. In addition, the early-Holocene climate in Beringian was influenced by various relevant changes, such as the flooding of the shelf and the connection between the Pacific and Arctic Oceans (Bartkein et al. 2015). Thus, including the potential influence of these processes expected to reduce the model-data differences.

(6) On the application of our simulations to investigate the dynamics of other parameters

With confirmation of inter-model and model-data comparisons, our simulations can reasonably represent climate variations throughout the Holocene in most the NH extratropics. Simulated Holocene temperatures have the potential to be used in analyses of the contribution of climate change to other components/parameters, such as vegetation dynamics. However, other data (e.g. on forest fires and pollen data) are typically site-specific records and represent the small-scale environment. Thus, the establishment of a downscaling procedure on our simulations would enhance its applicability in these relevant fields.

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