Effects of melting ice sheets and orbital forcing on the early Holocene warming in the extratropical Northern Hemisphere

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Abstract

The early Holocene is marked by the final transition from the last deglaciation to the relatively warm Holocene. Studies have analyzed the influence of diminishing ice sheets and other forcings on the climate system during the Holocene. The climate response to forcings before 9 ka, however, remains not fully comprehended. We therefore studied how orbital and ice-sheet forcings contributed to climate change during the earliest part of the Holocene (11.5–7 ka) by employing the LOVECLIM climate model.

Our equilibrium experiments for 11.5 ka suggested a lower annual mean temperature at the onset of the Holocene than in the preindustrial era in the Northern extratropics with the exception of Alaska. The magnitude of this cool anomaly varies regionally and these spatial patterns are broadly consistent with proxy-based reconstructions. Temperatures throughout the whole year in N Canada and NW Europe for 11.5 ka were 2–5°C lower than those of the preindustrial era as the climate was strongly influenced by the cooling effect of the ice sheets, which was caused by enhanced surface albedo and ice-sheet orography. In contrast, temperatures in Alaska for all seasons for the same period were 0.5–3°C higher than the control run, which were caused by a combination of orbital forcing and stronger southerly winds that advected warm air from the South in response to prevailing high air pressure over the Laurentide Ice Sheet (LIS).

The transient experiments indicated a highly inhomogeneous early Holocene temperature warming over different regions. The climate in Alaska was constantly cooling over the whole Holocene, whereas there was an overall fast early Holocene warming in N Canada by more than 1°C ka^{-1} as a

consequence of progressive LIS decay. Simulated temperatures compared with proxy records illustrated uncertainties related to the reconstruction of ice-sheet deglaciation, which can be constrained by applying different freshwater scenarios. Our results demonstrated the variability of the climate during the early Holocene, both in terms of spatial patterns and temporal evolution.

1 Introduction

The early Holocene from 11.5 to 7 ka B.P (hereafter noted as ka) is palaeoclimatologically interesting as it represents the last transition phase from full glacial to interglacial conditions. This period is characterized by a warming trend in the Northern Hemisphere (NH) that has been registered in numerous proxy records and indicated by stacked temperature reconstructions (Shakun et al., 2012). Oxygen isotope measurements from ice cores in Greenland (Dansgaard et al., 1993; Grootes et al., 1993; Rasmussen et al., 2006; Vinther et al., 2006; 2008) and the Canadian high-Arctic (Koerner & Fisher, 1990) consistently show an increase in $\delta^{18}O$ by up to 3–5 ‰, which indicates a couple of degrees warming in the climate system (Vinther et al., 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 chironomids and macrofossils data obtained from lake sediments (Brooks & Birks, 2000; Brooks et al., 2012; Birks, 2015). In addition, this transition is recorded in other high-resolution records from further East in Eurasia, such as in the speleothems from China (Yuan et al., 2004; Wang et al., 2005). Comparable trends have been identified in marine sediment core data, such as a sea surface temperature (SST) rise in the N Atlantic reflected by the variation in $\delta^{18}O$ and planktonic foraminifera (Bond et al., 1993; Kandiano et al., 2004; Hald et al., 2007). Although these proxy records provide a general view of early Holocene warming, their detailed expression in different regions and the reasons for this spatial variation are poorly known.

The orbitally induced increase in NH June insolation was one of the main external drivers of the climate change during the last deglaciation (Berger, 1988; Denton et al., 2010; Abe-Ouchi et al., 2013; Buizert et al., 2014). This increase peaked in the earliest Holocene (Berger, 1978) and resulted in warming over large areas. However, the early Holocene was also characterized by critical adjustments in components of the climate system that further affected the temperature through various feedback mechanisms. In the cryosphere, the Laurentide ice Sheet (LIS) and Fennoscandian Ice Sheet (FIS) were melting at a fast rate and eventually demised around 6.8 ka and 10 ka, respectively (Dyke et al., 2003; Occhietti et al., 2011), which exerted multiple influences on the climate system (Renssen et al., 2009). 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 can also influence the climate through the mechanism of adjustment to the atmospheric circulation (Felzer et al., 1996; Justino & Peltier, 2005; Langen & Vinther, 2009). For instance, the orography of a large ice sheet generated a glacial anticyclone that locally tended to reduce further the temperature (Felzer et al., 1996), but may also have caused a 2–3°C warming over the North Atlantic under the Last Glacial Maximum (LGM) (Pausata et al., 2011; Hofer et al., 2012). Third, both modelling and proxy studies have found that the Atlantic Meridional Overturning Circulation (AMOC) was relatively weak during the early Holocene due to the ice-sheet melting, 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). Overall, the net effect of ice sheets on the early Holocene climate can be expected to have tempered the orbitally induced warming at the mid and high latitudes. Important adjustments in the carbon cycle occurred in the early Holocene, as evidenced by the rise in atmospheric CO_2 levels by 20–30 ppm that contributed to the warming (Schilt et al., 2010). Changes also happened in the biosphere during the early Holocene. Vegetation reconstructions revealed a northward expansion of boreal forest in the circum-Arctic region after the retreat of the ice sheets (MacDonald et al., 2000; Bigelow et al., 2003; 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 of the surface albedo and induced a positive feedback to the warming trend (Claussen et al., 2001).

The impact of all these forcings on the Holocene climate has been examined in modelling studies. The particular focus in these studies has been on the influence of the LIS and Greenland ice sheet (GIS) decays had on the climate after 9 ka relative to other climate forcings (Renssen et al., 2009; Blaschek & Renssen, 2013). Renssen et al. (2009) used transient simulations performed with the ECBilt-CLIO-VECODE model and found that the Holocene climate was sensitive to the ice sheets and that the LIS cooling effects delayed the Holocene Thermal Maximum (HTM) by up to thousands of years. Blaschek and Renssen (2013) applied a more recent version of the same model (renamed as LOVECLIM) and revealed that the GIS melting had an identifiable impact on the climate over the Nordic Sea. However, these Holocene modelling studies only started at 9 ka, implying that the climate of the early Holocene between 11.5 and 9 ka was not included in these studies. The most important challenges in simulating climate during this initial phase of the early Holocene are the inherent uncertainties in the ice-sheet forcings in terms of the ice-sheet dynamics and the related meltwater release. 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; Clark in preparation), primarily because of a large uncertainty in bulk organic sample ages and the possibility of old carbon contamination (Carlson et al., 2014; Stokes et al., 2015). Furthermore, the Younger Dryas stadial ended at 11.7 ka and may still have influenced the early Holocene climate due to the long response time of the deep ocean (Renssen et al., 2012). Therefore, the climate system's response to forcings before 9 ka, especially those of the ice sheets is poorly comprehended.

We have extended the study of Blaschek and Renssen (2013) back to 11.5 ka in the present study to explore the early Holocene climate response to these key forcings. By employing the same climate model of intermediate complexity LOVECLIM, we first analyzed the impact of forcings on the climate at 11.5 ka and subsequently investigated the influence of two ice-sheet deglaciation scenarios in transient simulations. The comparison of these different simulations enable us to disentangle how the ice sheets influenced the early-Holocene climate. More specifically, we have addressed the following research questions: 1) What were the spatial patterns of simulated temperature at the onset of the Holocene (11.5 ka)? 2) What were the roles of the forcings, especially ice-sheet decay, in shaping these features? 3) What was the spatiotemporal variability in the simulated early Holocene temperature evolution?

2 Model and experimental design

2.1 The LOVECLIM model

We conducted our simulations with version 1.2 of the three-dimensional Earth system model of intermediate complexity LOVECLIM (Goosse et al., 2010), in which the components of the atmosphere, ocean including sea ice, vegetation, ice sheets and carbon cycle are dynamically included. However, in our version, the components for the ice sheets and the carbon cycle were not activated. Therefore, the ice-sheet configuration was prescribed in our present study. The atmospheric component is the quasi-geostrophic model ECBilt that consists of three vertical layers and has T_{21} horizontal resolution (Opsteegh et al., 1998). CLIO is the ocean component which consists of a free-surface, primitive-equation oceanic general circulation model (GCM) coupled to a three-layer dynamic-thermodynamic sea-ice model (Fichefet & Maqueda, 1997). The ocean model includes 20 vertical levels and a 3°x3° latitude-longitude horizontal resolution (Goosse & Fichefet, 1999). These two core components were further coupled to the biosphere model VECODE, which simulates the dynamics of two main terrestrial plant functional types, trees and grasses, in addition to desert (Brovkin et al., 1997). More details on LOVECLIM can be found in Goosse et al. (2010).

The LOVECLIM model is a useful tool to explore the mechanisms behind climate change and it has made critical contributions to our understanding of the long-term climate change in the Holocene observed in proxy records (Renssen et al., 2005; 2006; 2010). For example, it has helped the investigations of the potential forcings behind the Younger Dryas (Renssen et al., 2002; Wiersma et al., 2006, Renssen et al., 2015), and the role of the decaying LIS and GIS in the temperature evolution over the last 9 ka (Renssen et al., 2009; Blaschek & Renssen, 2013). Moreover, the LOVECLIM model simulates a reasonable modern climate (Goosse et al., 2010). It also simulates a reasonably well meridional overturning streamfunction and reproduces a large-scale structure of atmosphere circulation, which agrees with observations and with other models (Goosse et al., 2010). In addition, the model's sensitivity to freshwater perturbation is reasonable compared to that of other models (Roche et al., 2007), and it sensitivity to a doubling of atmospheric CO_2 concentration is 2 K, which is in the lower end of estimation in of coupled general circulation models (GCMs) (Flato et al. 2013).

2.2 Prescribed forcings

We included the major climatic forcings in terms of greenhouse gases (GHG) in the atmosphere, astronomical parameters (orbital forcing or ORB) and decaying ice sheets. In all simulations, the solar constant, aerosol levels, the continental configuration and bathymetry were kept fixed at preindustrial values. We based the concentrations of CO_2, CH_4 and N_2O on ice core measurements for GHG forcing (Loulergue et al., 2008; Schilt et al., 2010). The radiative GHG forcing anomaly (relative to 0 ka) in W m^{-2}, representing the overall GHG contribution, at first showed a rapid rise with a peak of -0.3 W m^{-2} at 10 ka, which was followed by a slight decrease towards a minimum at 7 ka, and gradual increased towards 0 ka (Fig. 1). The astronomical parameters (eccentricity, obliquity and longitude of perihelion) determine the incoming solar radiation at the top of atmosphere, and were derived from Berger (1978). An example of the resulting change in insolation is shown as the anomaly for June at 65° N in Figure 1, which shows the gradual decrease over the course of the Holocene. At the beginning of the Holocene (11.5 ka), the orbitally-induced insolation anomaly in the NH was positive in summer and negative in winter (Fig. SI1). Overall, this setup of GHG and ORB forcing is in line with the PMIP3 protocol (http://pmip3.lsce.lsce.ipsl.fr), except that our simulation excluded the increase in GHG levels during the industrial era (Ruddiman, 2007).

We took three aspects into account concerning the ice-sheet forcing, namely: their spatial extent, their thickness and their meltwater discharge. The reconstructions of ice-sheet spatial extent are based on the dating of geological features and on the correlation of these geological datasets between different regions (Dyke et al., 2003; Svendsen et al., 2004; Putkinen & Lunkka, 2008). The FIS at 11.5 ka covered most of Fennoscandia except for southern Scandinavia and eastern Finland (Svendsen et al., 2004; Putkinen & Lunkka, 2008; Clark in preparation). The LIS occupied most of the lowland area north of the Great Lakes region and filled the whole Hudson Strait (Licciardi et al., 1999; Dyke et al., 2003; Occhietti et al., 2011). The relative thickness of LIS was up to 2000 m, and for the FIS this thickness was only about 100 m (Ganopolski et al. 2010), which is comparable with the ICE-5G reconstruction (Peltier 2004). Both the spatial extent of the ice sheets and their thickness were updated every 250 yrs in our transient experiments, and they decreased rapidly during the earliest Holocene, followed by a more gradual deglaciation rate from 8 ka onward (Fig. 2a).

We applied the meltwater release for 1200 yrs in our equilibrium experiments for 11.5 ka by adding 0.11 Sv (1 Sverdrup is 10^6 m^3/s) of freshwater at the St. Lawrence River and 0.05 Sv at Hudson Strait and Hudson River, 0.055 Sv from FIS and 0.002 Sv from GIS (Licciardi et al., 1999; Jennings et al., 2015). The total freshwater volume added to the oceans in our transient experiments was about $1.46x10^{16}$ m^3 in the first 4700 yrs (Fig. 2b), which roughly matches the estimated ice-sheet melting volume during the early Holocene (Dyke et al., 2003; Ganopolski et al., 2010; Clark in preparation). The volume of meltwater was slightly lower than the volume of the estimated 60 m sea level rise that took place during the early Holocene (Fig. 2c) (Lambeck et al., 2014), which suggests a coeval Antarctic melting contribution that is not considered here. Given the lack of a direct imprint left by meltwater on terrestrial records and hence the relatively large uncertainty, we used two versions of the freshwater flux (Fig. 2b) that represent two possible deglaciation scenarios of the GIS and FIS, named FWF-v1 and FWF-v2. The GIS FWF_v1 scenario is derived from the ICE_5G reconstruction, and FWF_v2 is based on the reconstruction of Vinther et al. (in 2009) that suggests a faster GIS thinning. The two FIS fwf scenarios are based on two estimations of the FIS melting, since the recent cosmogenic dating (FWF_v2) supports a faster melting (Clark in preparation) than previously thought (FWF_v1). However, we kept the freshwater discharge from LIS the same as in version-1, since the LIS deglaciation has been relatively well studied and we are more certain about its contribution.

2.3 Setup of experiments

We performed two types of experiments: equilibrium and transient simulations. First, a series of equilibrium experiments with boundary conditions for 11.5 ka (Table 1) were designed, of which only two simulations are presented here: OG11.5 and OGIS11.5. The OGIS11.5 experiment included ice-sheet forcing whereas no ice sheets were included in OG11.5 (Table 2). Each of these experiments was initiated from the model's default modern condition, and was run for 1200 yrs, of which the last 200 yrs of data were used for the analysis. Renssen et al. (2006) have demonstrated that a 1200-year spin-up is sufficient to reach a quasi-equilibrium in all components of the model.

The end of the 1200-yrs equilibrium run data were then taken to initialize the transient experiments that covered the last 11.5 ka. In the first transient simulation named ORBGHG, both GHG and ORB vary on an annual basis. In the second simulation OGIS_FWF-v1, the ice-sheet topography (Fig. 2a) and FWF_v1 (Fig. 2b) were additionally included. A third experiment (named OGIS_FWF-v2) was performed with the freshwater version-2 (Fig. 2c) and with the same ice-sheet topography as in the OGIS_FWF-v1 to further investigate the climate response to the relatively uncertain freshwater forcing. A pre-industrial simulation (PI) was run for 1200 yrs from the model's default with the boundary conditions that are shown in Table 1 and, similarly as for the other equilibrium experiments, the results of the last 200 yrs were used as a reference. These simulations and their forcings are summarized in Table 2. All temperature values in this study are shown as deviations from the PI simulation. The temperatures presented here are simulated surface temperature values without the environmental lapse rate corrections to the sea level temperature, which imply approximately a 0.5°C cooler bias over ice-sheet covered regions when compared with site specific proxy records.

3 Results

3.1 Equilibrium experiments at the onset of the Holocene

3.1.1 Simulation with only ORB & GHG forcings at 11.5 ka (OG11.5)

In experiment of OG11.5, summer temperatures were 2–4°C higher over most of extratropical continents than in the simulation of PI , with the maximum deviation of 5°C in the central part of continents (Fig. 4a). The warming over the oceans was about 1.5°C, and less conspicuous than that of continents. The warmer conditions were caused by the orbitally induced positive summer insolation anomaly, as all atmospheric greenhouse gas levels were lower at 11.5 ka than in the preindustrial era (Fig. 1). The most obvious feature of the simulated winter temperature anomaly was the marked contrast between high latitudes and areas more to the South (Fig. 3b). For instance, the mid-latitudes were 1.5–3°C cooler with the strongest cooling in the central continents, whereas the high-latitude Arctic was clearly warmer with a maximum up to +3°C than in the PI. This latitudinal gradient can be seen in the annual mean temperature as well (Fig. 3c). Annual mean temperatures over the Arctic were about 1–4°C higher than in the PI with slightly stronger warm climate in winter than in summer, which indicates the Arctic Ocean damping effect on a seasonal signal due to a large heat capacity. The temperatures were roughly unchanged (mostly within ±0.5°C) over the southern regions with a stronger seasonality (with warmer summers and cooler winters), which is consistent with the insolation change at 11.5 ka (Fig. 2).

3.1.2 Climate response to melting ice sheets at 11.5 ka (OGIS11.5)

Our simulation OGIS11.5 (including the impact of ice sheets) suggests a much cooler climate than that of the OG11.5 (Fig. 4d). Most notably, ice sheets induced a strong summer cooling over ice-covered areas, and reduced temperatures up to 5°C compared to the PI, with the strongest cooling at the center of the LIS. Additionally, the SST was also more than 1.5°C lower over the N Atlantic Ocean. In contrast, temperatures over the ice-free continents were mostly above the preindustrial level, but still lower than those found in the OG11.5, except for Alaska region. The OGIS11.5 simulation in winter indicated warmer central Arctic than that of in the PI, although the area with colder conditions clearly expanded in the OGIS11.5 than in the OG11.5, with anomalies fell below -5°C (Fig. 4e). Alaska was the only continental region where the winter temperatures exceeded the preindustrial values by up to +3°C. The strongest cooling effect was present in the regions covered by ice sheets, for instance more than 3°C cooler over the LIS. The simulated annual mean temperature in the OGIS11.5 clearly showed an overall lower value than in the PI due to the ice-sheet impacts (Fig. 4f). The Eurasian continent was mostly 1.5–3°C cooler, and a maximum temperature reductions of more than 5°C was simulated over the LIS. Only two areas were still warmer: Alaska, including the adjacent sector of the Arctic Ocean, and the Nordic Seas. The most distinct feature was thus a thermally contrasting pattern over North America, with simulated temperatures being around 2°C higher than those in the PI for Alaska, whereas over most of Canada the temperatures were more than 3°C lower.

3.2 Transient simulation for the Holocene

It is clear in our analysis of section 3.1 that the climate showed different responses in the following areas: the Arctic, NW Europe, N Canada, Alaska and Siberia (marked in Fig. S2). Therefore, these areas were selected for special examination and the temperature evolutions of these regions will be shown. Our major focus was on the millennial-scale temperature trends, therefore we applied a 500-yr running mean to our simulated time-series that effectively filtered out high-frequency variability.

3.2.1 Temperature evolution in the Arctic

The Arctic summer temperature in the ORBGHG continuously decreased, which resulted in a total cooling of 2°C during the Holocene (Fig. 4). The winter temperature showed an even stronger overall cooling that fell by about -3°C. Accordingly, the annual mean temperature displayed a cooling that was intermediate between that of the summer and winter seasons.

The simulation OGIS_FWF-v1 with full forcings revealed a more complicated Arctic climate evolution compared to that of the ORBGHG. The effect of ice sheets at the onset of Holocene caused the temperatures in both summer and winter to be more than 2°C lower than those indicated by ORBGHG. It was no surprise that the simulated effect of the ice sheets was larger at the beginning of the Holocene than afterwards when the deglaciation had progressed further. The final deglaciation of FIS happened at 10 ka and the corresponding deglaciation for LIS occurred at 6.8 ka. Therefore, their cooling effects no longer existed after 6.5 ka, and all three runs showed similar temperatures after that time. As a consequence, the temperature evolution curve of OGIS_FWF-v1 first showed a warming, with the peak being reached at around 7 ka when the cooling effects of the ice sheets had been counterbalanced by the insolation. This was subsequently followed by a gradual cooling that was controlled by the orbital forcing. The temperature initially had increased by 6.5 ka at rate of about 0.26, 0.21 and 0.44°C ka^{-1} for summer, winter and annual temperatures, respectively. The OGIS_FWF-v1 simulation indicated that the summer climate in the Arctic experienced a slightly faster warming at the beginning, followed by a more gradual warming toward the maximum anomaly of 1°C warmer than PI at about 7.5 ka (Fig. 4a). The simulated winter temperature stayed at a lower level that was 2°C lower than that of ORBGHG before 7 ka, which was followed by a rapid increase of about 1.5°C within a 500 yrs periods, and then reached a temperature peak of about 1.5°C warmer than in the PI (Fig. 4b). The simulated annual mean temperatures showed a relatively stable rise until 6.5 ka, which reached a maximum of about 1.5°C warmer than the PI (Fig. 4a). The simulation OGIS_FWF-v2 gave similar results for the Arctic, but had an even cooler climate before 9 ka than in OGIS_FWF-v1 with the maximum cooling of up to 0.3°C for all seasons at 10.5 ka.

3.2.2 Temperature evolution in NW Europe

The ORBGHG indicates smaller climate variability in NW Europe than in the Arctic. The temperature declined by around 1.5°C through the entire period in summer, less than 0.5°C for annual mean and rose by 0.5°C in winter (Fig.6), which implied a stronger seasonality than in the preindustrial period. This contrasts markedly with the clear cooling of climate in each season in the Arctic.

The OGIS_FWF-v1 simulation showed an overall cooler climate in NW Europe at the onset of Holocene, with the temperature anomalies of -1.5°C in summer, -3°C in winter and -2.8°C in annual mean compared to the PI. The temperatures increased from this point (11.5 ka) toward 6 ka at an overall rate of 0.28, 0.48, 0.54 °C ka^{-1} for summer, winter and annual mean, respectively. The most important feature in summer was a sharp rise of temperature from a negative anomaly (-1.5°C) to a positive one (+1°C) by 10 ka, when the first peak was reached. Subsequently, a slight cooling was noted before 8 ka, followed by another temperature increase, which led to a second warming peak at 7.4 ka. The climate in winter showed a relatively stable warming by 6.5 ka with no identifiable warm peak. The annual temperature reflected the same phases of warming as in summer, one before 10 ka and another before 7.5 ka, but without a clear early temperature peak. Temperatures in all seasons from around 7 ka followed the ORBGHG simulation. It is worth noting that the OGIS_FWF-v2 simulation produced a further cooling in summer between 11.5 and 9 ka relative to the OGIS_FWF-v1, which was also reflected in the annual mean temperature. As a result, there was only one clear thermal maximum in summer for NW Europe, which peaked at around 7.4 ka.

3.2.3 Temperature evolution in N Canada

Simulated temperatures in ORBGHG for N Canada decreased by 2.2°C in summer, by 0.6°C in winter and by 1.1°C for annual average during the Holocene (Fig.7). The stronger cooling in summer than in winter reflected the strong early-Holocene seasonality, which decreased over the whole period.

The OGIS_FWF-v1 simulation described a much cooler climate in N Canada during the early Holocene than that indicated by the ORBGHG. This cooling was up to 5°C for all seasons at the onset of Holocene. The climate dramatically warmed up with an overall high rate of more than 1°C ka^{-1} in both winter and summer during the early Holocene, which was due to the impact of the decaying LIS. The early Holocene warming was however not linear because an initial phase with more rapid warming was followed by a more gradual temperature increase. In summer, this warming resulted in a thermal peak at around 7.4 ka, which was about 1.5°C warmer than in the PI. From 7.4 ka onwards, the climate experienced a gradual cooling that is very similar to that of the ORBGHG. The simulated temperatures in winter and annual mean did not show such a clear warm peak in comparison to summer. The results of OGIS_FWF-v2 only indicated marginal differences relative to OGIS_FWF-v1 for all seasons. Overall, the most significant feature of the simulated temperatures in N Canada was the strong warming that took place in the early Holocene.

3.2.4 Temperature evolution in Alaska

The ORBGHG simulation showed an overall cooling climate in Alaska for all seasons. The simulated summer and annual mean temperatures experienced a decrease of more than 2°C throughout the whole period. The winter temperature had slightly increased by 10 ka, and then stayed about 2°C higher for a period of 800 yrs, which was followed by a constant decrease toward the preindustrial era.

In contrast to other areas, both summer and winter temperatures in OGIS_FWF-v1 showed an overall cooling trend in Alaska during the entire Holocene (Fig. 7), which was slightly warmer than in our ORBGHG simulation. The OGIS_FWF-v1 simulation indicates a 2°C decline in summer temperature over the whole period with a slightly faster rate between 7 ka and 6.5 ka. The simulated winter temperature decreases by 3.5°C during the early Holocene, with two small declines at 9.5 ka and 6.7 ka. The annual temperature in the OGIS_FWF-v1 simulation reflected a 2.3°C cooling during the Holocene. The OGIS_FWF-v2 simulation produced a rather similar Alaska temperature trend to that of the OGIS_FWF-v1.

3.2.5 Temperature evolution in Siberia

The ORBGHG described an almost 2°C decline of summer temperature over Siberia during the last 11.5 ka (Fig. 8). Simulated winter temperature showed a smaller variation, as it decreased by less than 1°C, and annual mean temperature decreased by around 1°C on the course of the Holocene. The evolution of simulated temperatures in the ORBGHG simulation over Siberia was on a similar scale to that of NW Europe.

The difference of simulated Siberian temperatures between ORBGHG and OGIS_FWF-v1 varied in summer and winter. On the one hand, the simulated summer temperature in the OGIS_FWF-v1 was generally similar to that in the ORBGHG with the exception of a small warming of 0.7°C before 10 ka. On the other hand, simulated winter temperature in the OGIS_FWF-v1 was around 2°C lower than in the ORBGHG before 7 ka, followed by a rapid increase over the next 500 yrs after which it followed the ORBGHG simulation. Consequently, the simulated early Holocene warming lasted much longer in winter than in summer. The simulated Siberian temperatures evolution in OGIS_FWF-v2 generally followed that of the OGIS_FWF-v1.

4 Discussion

We will evaluate our results by briefly comparing the simulations with proxy-based reconstructions, which will be followed by an analysis of the mechanism behind the simulated temperature patterns. The impact of freshwater forcing will also be discussed based on the two fwf scenarios.

4.1 Comparison of simulations with proxy records

At the onset of the Holocene, the overall cool climate indicated by the reconstructions generally matches that of our OGIS11.5 simulation, which showed a lower annual temperature at 11.5 ka than in the PI. Climate reconstructions based on proxy data generally show a cooler early Holocene over N Europe than at the present both in the summer and winter (Heiri et al., 2104; Mauri et al., 2015). Terrestrial and ocean sediment data also suggest a cooler early Holocene climate over eastern Siberia (Klemm et al., 2013; Tarasov et al., 2013) and a slightly lower SST over the N Atlantic

Ocean (Came et al., 2007; Berner et al., 2008). Cooler conditions over the Barents Sea and Greenland are also indicated by multiple proxies (Peros et al., 2010; de Vernal et al., 2013; Vinther et al., 2008). Therefore, these proxy data agree with the simulated lower temperatures over these areas.

However, there is less agreement with proxies in places where the reconstructions are sparse. Only available pollen-based reconstructions from the western side of the Ural Mountains suggest a similar early Holocene summer (within 1°C anomaly) to that the present (Salonen et al., 2011), whereas OGIS11.5 indicates that the summer temperature was slightly higher at 11.5 ka over most areas. At the high latitudes, the sea-ice cover reconstructions serves as an indirect palaeoclimate proxy due to the scarcity of temperature records, and reveals an inconclusive temperature signals over the Canadian Arctic (de Vernal et al., 2013), whereas our simulation reflected an overall warmer climate in the west and cooler conditions in the east.

Proxies indicate significantly different climate patterns over the east and the west of N America. The later initiation and termination of HTM over N Canada imply lower temperatures during the early Holocene in the east (Kaufman et al., 2004). Whereas, the higher-than-present early Holocene temperatures over Central Beringia and Alaska are reflected by peat accumulation and by northward expansion of animal species (Kaufman et al., 2004; Jones & Yu, 2010). This thermal contrast agrees with those simulated patterns in the OGIS11.5, which indicate a warmer temperature for Alaska and a much cooler climate over Canada. However, this interpretation of high temperature was recently changed by Kaufman et al. (2016), who argued that the highest summer temperature in Alaska occurred as late as 8-6 ka. Hence our simulation agrees better with the interpretation of Kaufman et al. (2004). In general, our simulation with full forcings was able to capture the main temperature features that are also indicated in proxy-based reconstructions.

Both stacked reconstructions (Shakun et al., 2012; Marcott et al., 2013) and our simulation OGIS_FWF-v2 show that the Holocene was generally characterized by an initial warming and subsequent Holocene warm period over the NH extratropics, which indicate the broad consistency between simulation and proxy data. However, there are some disagreements related to seasonality (Fig. 9). Marcott et al. (2013) interpreted the stacked temperature reconstruction representative of the annual mean climate, whereas it shows a better agreement with our simulated summer temperature than with annual mean value (Fig. 9). One potential explanation is that the proxy records have seasonal bias toward summer conditions, as has been suggested recently for many marine-based SST reconstructions (Lohmann et al. 2013). Further region-by-region comparisons of these warming rates with proxy records are hindered by limited space.

4.2 Mechanism of climate response to forcings

It is clear from our data that the spatial patterns of climate response at the onset of the Holocene can be attributed to the variation in the dominant forcings prevailing in the different areas. Orbital scale insolation variation is one of important driving factors for the early Holocene climate. For instance, the higher temperature in Alaska could be attributed to the orbitally induced positive insolation anomaly in combination with an anomalous atmospheric circulation caused by the remnant LIS. The air descended over the cold LIS surface, which created a high surface pressure anomaly that produced a clockwise flow anomaly at the surface, as indicated by the 800 hPa Geopotential height (Fig.11c). This would induce stronger southerly winds over Alaska, which advected relatively warm air from the South. A potentially different early Holocene atmospheric circulation near the N Atlantic has also been reported by Steffensen et al. (2008) who found an abrupt transition of deuterium excess and a considerable reduction in dust concentration from the GRIP core data.

The strong influence of the ice sheets on early Holocene temperatures has been found in previous studies (Renssen et al., 2009; 2012). The simulated lower summer temperatures over N Canada and NW Europe in our OGIS11.5 simulation is the result of such ice-sheet induced cooling, which would fully overwhelm the warming effect of the positive summer insolation anomaly. The ice-sheet cooling effect can be partly explained on a local scale by the enhanced albedo over the ice sheets and by the climate's high sensitivity to albedo change (Romanova et al., 2006). Indeed, the summer surface albedo over the ice sheets is much higher (up to 0.8) than over ice-free surfaces where the values vary from only 0.1 to 0.5, depending on the vegetation type and the fractional snow cover (Fig. 10). The temperatures can be further reduced by the ice-sheet orography impact. The elevation of ice sheets introduced descending air over ice-sheet surface, which caused the cooler condition on the local scale. There is also an approximate 0.5 cooling bias induced by the lapse rate effect compared with the site-based records.

Changes in vegetation and land cover during the early Holocene contributed to climate change as well, especially over ecotonal regions. Modelling studies suggest that deforestation in boreal regions decrease the regional temperature by up to 1°C due to an increase in surface albedo and related positive feedbacks (Levis et al., 1999; Claussen et al., 2001; Liu et al., 2006). Taking Siberia as an example, the insolation-induced warming was partially offset by the overall higher summer albedo (Fig. 10) induced by the southward expansion of the tundra or/and bare ground and related feedbacks at 11.5ka, which led to only a minor warmer summer climate than PI. The albedo-related feedbacks and the smaller insolation anomalies annually counterbalanced each other at 11.5 ka, and resulted in a 0.5–2°C cooler climate. We are aware of the potential role of permafrost at high latitudes, however, the discussion of the impact of permafrost thaw is hindered by the fact that our model version did not include a dynamic permafrost module. A version of LOVECLIM that is coupled to a permafrost module (VAMPERS) is currently in development (Kitover et al. 2015), and should enable us to quantify the role of permafrost in a future study.

Meltwater release and sea-ice related changes also had a footprint in the early Holocene climate. The OGIS11.5 simulation produced a sluggish AMOC in the N Atlantic with the largest decrease being more than 3 Sv. It was also reflected in a shallower overturning circulation at 11.5 ka compared to the PI simulation period as a response to meltwater release (Fig. 12). This slowdown also coincides with the foraminifera data from the Arctic Ocean and the Fram Strait that suggest a reduced northward oceanic heat transport (Thornalley et al., 2009). The slowdown and reduced heat transport led to a slightly lower temperature at high latitudes (western Arctic Ocean) at 11.5 ka than that found in the PI. Likewise, after the meltwater fluxes of the LIS diminished around 7 ka strong intensification of the AMOC followed. This sudden intensification of AMOC would explain the rapid Arctic temperature increase that occurred at this time (Fig. 4). However, it is worth noting that the temperature decrease was not simply inversely linear with the amount of northward transport of heat since the sea-ice feedbacks further reinforce this change (Roche et al., 2007). Actually, sea-ice coverage in the OGIS11.5 was much more extensive over the Davis Strait (N Labrador Sea) than the corresponding value in the OG11.5 (Fig. 13). This extended sea-ice cover in this region was stronger than the direct cooling effect of the reduced oceanic heat transport. Such an anomaly might be explained by positive feedbacks involving sea-ice being active (Renssen et al., 2005). The Greenland Sea warming can be attributed to enhanced convective activity that releases more oceanic heat into the atmosphere. This enhanced convective activity was caused by the shift of deep water formation from the eastern Greenland Sea to the west, which was initially induced by the freshwater discharge from ice-sheet melting. The net overall response of climate reflects the impact of a combination of forcings and feedbacks, which showed a high temporal-spatial variability.

4.3 Early Holocene warming and climate-ocean system response to freshwater

The simulation of OGIS_FWF-v2 produced a stronger cooling in the Arctic and NW Europe than those found in the OGIS_FWF-v1 before 9 ka with a stronger temperature reduction of around 10 ka. 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, which reduced convective activity (Renssen et al., 2010; Blaschek & Renssen, 2013). Indeed, this reduction led to a further slight reduction of the northward heat transport by the Atlantic Ocean, which was associated with a weakened AMOC (by 1-2 Sv): this in turn produced a slightly stronger cooling at 10 ka (Fig. S3) and a sea-ice expansion over the Denmark Strait (Fig. 14 & SI4).

The meltwater flux freshened the surface ocean in the deep-water formation area, which led to a slowdown in the ocean circulation. However, the efficiency of this effect is determined by multiple aspects. The most important factor is the maximum flux of meltwater that was added to the ocean, while the total freshwater amount had only a second-order effect (Roche et al., 2007). Numerous investigations on the behavior of the coupled atmosphere-ocean system suggest that the applying of freshwater will not lead to a disruption of the North Atlantic Deep Water production (NADW) as long as a certain threshold is not crossed (Ganopolski et al., 1998; Rahmstorf et al., 2005). Apart from the intensity and duration, the ocean circulation response to freshwater also depends on the location where this freshwater is released. For instance, it is more sensitive to the release of freshwater in the eastern Norwegian Sea than at the St. Lawrence River outlet since the former is closer to the main site with NADW formation (Roche et al., 2010). This is consistent with a previous study by Blaschek & Renssen in (2103), who found that freshwater from the GIS did have a tangible impact on Nordic seas, even though the total amount was minor. This location-depended sensitivity could also partially explain why the AMOC was weaker in the OGIS_FWF_v2 simulation than in OGIS_FWF-v1.

The OGIS_FWF-v1 indicated two peaks in the temperature evolution over NW Europe, at around 10 and 7 ka. High temperatures at 7 ka are recorded in proxy-based reconstructions as well. However, no warm peak at 10 ka was observed in pollen-based reconstructions, which suggests a cooler climate prevailed at 10 ka than in present-day Europe (Mauri et al., 2015). In contrast to the climate simulated in the OGIS_FWF-v1 simulation, the simulation with updated freshwater (OGIS_FWF-v2) produced a warming trend that is consistent with a highest temperature around 7 ka. Moreover, the OGIS_FWF-v1 produced a temperature decrease between two peaks whereas the proxies indicated a rapid temperature increase at the beginning followed by a more gradual warming (Brooks et al., 2012). Therefore, from the viewpoint of the temperature evolution in NW Europe, the OGIS_FWF-v2 represented a more realistic climate than OGIS_FWF-v1 did, which implied that the existing uncertainties in the reconstructions of ice-sheet dynamics can be evaluated by applying different freshwater scenarios.

5 Conclusions

We performed both equilibrium and transient simulations by employing the LOVECLIM climate model to explore the spatial patterns of the climate response to forcings at the onset of the Holocene and the temperature evolution over the last 11.5 ka. We focused on three research questions in our analysis, which are outlined below with the main finding:

1) What were the spatial patterns of simulated temperature at the onset of the Holocene?

The temperatures at 11.5 ka were regionally heterogeneous compared with those of PI, which are shown as a range of annually negative anomalies over many areas but which were higher in Alaska. The climate in eastern N Canada and NW Europe was much cooler than those in other regions with temperature anomalies of -2 to -5°C relative to 0ka throughout the year. The climate over the N Labrador Sea and the N Atlantic was also 0.5–3°C cooler. The temperatures in Siberia were 0.5–3°C and 1.5–3°C lower in winter and annually, and the summer temperature showed only a small deviation (between +0.5 to -1.5°C) compared to 0 ka. The summer temperature anomaly in the eastern Arctic Ocean was also small (between ±-0.5°C), and annual temperatures were 0.5–2°C lower. In contrast to cooler conditions in other areas, the temperatures in Alaska were 1.5–3°C higher than the pre-industrial period for all seasons.

2) What were the roles of forcings, especially ice-sheet decay, in shaping these features?

The ice-sheet cooling effect in N Canada and NW Europe overwhelmed the warming impact of the positive insolation anomaly, which caused the relatively cold climate at 11.5 ka. In particular, the enhanced surface albedo over the ice sheets and the orographic effect were important for promoting this ice-sheet related cold conditions. The cooler climate over the N Labrador Sea and the N Atlantic was related to both reduced northward heat transport and activated sea-ice feedbacks which were initiated by weakened ocean circulation. A small summer temperature anomaly was found in Siberia, where the positive insolation anomaly was partially offset by the cooling effect of the higher albedo associated with the relatively extensive tundra cover in the early Holocene. The overall lower winter and annual temperatures at 11.5 ka can be attributed to both vegetation-related albedo feedbacks and to the relatively small negative insolation deviation compared to the preindustrial level over central Siberia.

The dominant factors for eastern Arctic Ocean climate were the amount of northward heat transport associated with the strength of ocean circulation and the orbitally forced insolation variation. The annual mean temperature at 11.5 ka was lower than at 0ka because the cooling effect of a reduced northward oceanic heat transport (induced by weakened ocean circulation) was larger than the insolation-induced warming. During summer, these two factors were of similar magnitude and temperatures were similar to those of the preindustrial era. Temperatures in Alaska were higher for all seasons in response to the dominant positive insolation anomaly and the enhanced southerly winds induced by the LIS, which advected relatively warm air from the South. Therefore, this regional heterogeneity is the result of the climate response to a range of dominant forcings and feedbacks.

3) What was the spatiotemporal variability in the simulated early Holocene evolution?

The Holocene temperature evolution and early Holocene warming were also geographically varying. In Alaska, the climate was constantly cooling throughout the Holocene due to the decreasing insolation and atmospheric circulation variability. In contrast, N Canada experienced a strong warming with an overall warming rate over 1°C ka-1 and this warming lasted until 7 ka. Although in NW Europe, the Arctic and Siberia different forcings and mechanisms played different roles, the overall warming effect was similar for these regions, with a rate of around 0.5°C ka^{-1}. In addition, the comparison of early Holocene temperatures over NW Europe with proxy records suggest that the OGIS_FWF-v2 represented a more realistic climate condition than the OGIS_FWF-v1 does, and imply that the uncertainties with regard to the ice-sheet decay can be constrained by applying different deglaciation scenarios. Overall, our results demonstrated a large spatial variability in the climate response to diverse forcings and feedbacks, both for the early Holocene temperature distribution and for the early Holocene warming.

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Table 1. Boundary conditions for 11.5 ka and pre-industrial (PI)

	11.5 ka			PI [a]		
Greenhouse gases (GHG)	CO_2 253 ppm	CH_4 511 ppb	N_2O 245 ppb	CO_2 280 ppm	CH_4 760 ppb	N_2O 270 ppb
Orbital parameters (ORB)	ecc 0.019572	obl 24.179 °	lon of perih 270.209 °	ecc 0.016724	obl 23.446 °	lon of perih 102.040 °
Ice sheets (relative to present)	Size $69.2*10^5$ km^2	Max thickness 2331 m	Meltwater flux 220 mSv	–	–	–

PI[a]: GHG for 1750 AD; ORB for 1950 AD.

Table 2. Experiments and corresponding setup

Equilibrium		Transient	
Name	Forcing	Name	Forcing
OG11.5	ORB+GHG	ORBGHG	ORB+GHG
OGIS11.5	ORB+GHG+IS+FWF	OGIS_FWF_v1	ORB+GHG+IS+FWF_v1
		OGIS_FWF_v2	ORB+GHG+IS+FWF_v2

Figure captions

Fig. 1. Evolution of greenhouse gas concentrations (GHG) shown as the radiative forcing's deviation from the pre-industrial level (with solid lines corresponding to the left axis), and June insolation at 65 °N derived from orbital configuration (with red line and the axis on the right).

Fig. 2. The prescribed ice-sheet forcing during the early Holocene. (a) Variation in ice-sheet extent (km^2) displayed as the black lines with the axis on the left and their maximum thickness (m) indicated by the green lines with the axis on the right. A relatively minor change in GIS is not shown due to its small scale. (b) Two freshwater flux scenarios (in mSv), FWF-v1 (dashed lines) and FWF-v2 (solid lines). (c) Total meltwater discharge in equivalent sea level (m).

Fig. 3. Simulated temperatures for 11.5 ka, shown as deviation from the PI. Left column shows the simulation with only GHG and ORB forcings (OG11.5). For the right column, the ice-sheet forcing is included (OGIS11.5). Upper, middle and lower panels present summer, winter and annual mean temperatures, respectively.

Fig. 4. Simulated temperature evolution, shown as the anomalies compared to the PI, since the early Holocene at high latitudes (North of 70°). The red line indicates the simulation with only ORB and GHG forcings (ORBGHG), while the black and green lines represent the simulations OGIS_FWF-v1 and OGIS_FWF-v2, in which the ice-sheet forcing are included. (a), (b) and (c) panels represent the summer, winter and annual values respectively. The slope indicates the overall warming rate and is based on the least squares regression over the period from the 11.5 to 6 ka, as from 6 ka the temperature start to decrease. It is only a general estimation, thus uncertainty ranges are not provided. The warmest peak is marked by shaded bar and represents the simulated peak during which the temperature was over 1°C higher than in PI.

Fig. 5. Simulated temperatures, shown as the anomalies compared to the PI, during the Holocene in Northwestern Europe (5 °W–34 °E, 58 °N–69 °N). The caption is the same as in fig. 4.

Fig. 6. Simulated temperatures, shown as the anomalies compared to the PI, during the Holocene in Northern Canada (120 °W–55 °W, 50 °N–69 °N). The caption is the same as in fig. 4.

Fig. 7. Simulated temperatures, shown as the anomalies compared to the PI, during the Holocene in Alaska (170 °W–120°W, 58 °N–74 °N). The caption is the same as in fig. 4.

Fig. 8. Simulated temperatures, shown as the anomalies compared to the PI, since the early Holocene in Siberia (62–145° E, 58–74 °N). The caption is the same as in fig. 4, except that the warming rate slope is indicated for a shorter period.

Fig. 9. Model-data comparison over the latitudinal band of 30-90 °N, shown as a deviation from the PI. Black lines represent the stacked temperature reconstruction with 1δ uncertainty (grey band). Red and green lines indicate the simulated summer and annual temperatures.

Fig. 10. Summer surface albedo in the extratropical Northern Hemisphere. (a), (b) and (c) represent the control run (PI), the simulations without ice sheets (OG11.5) and with ice sheets (OGIS11.5) respectively.

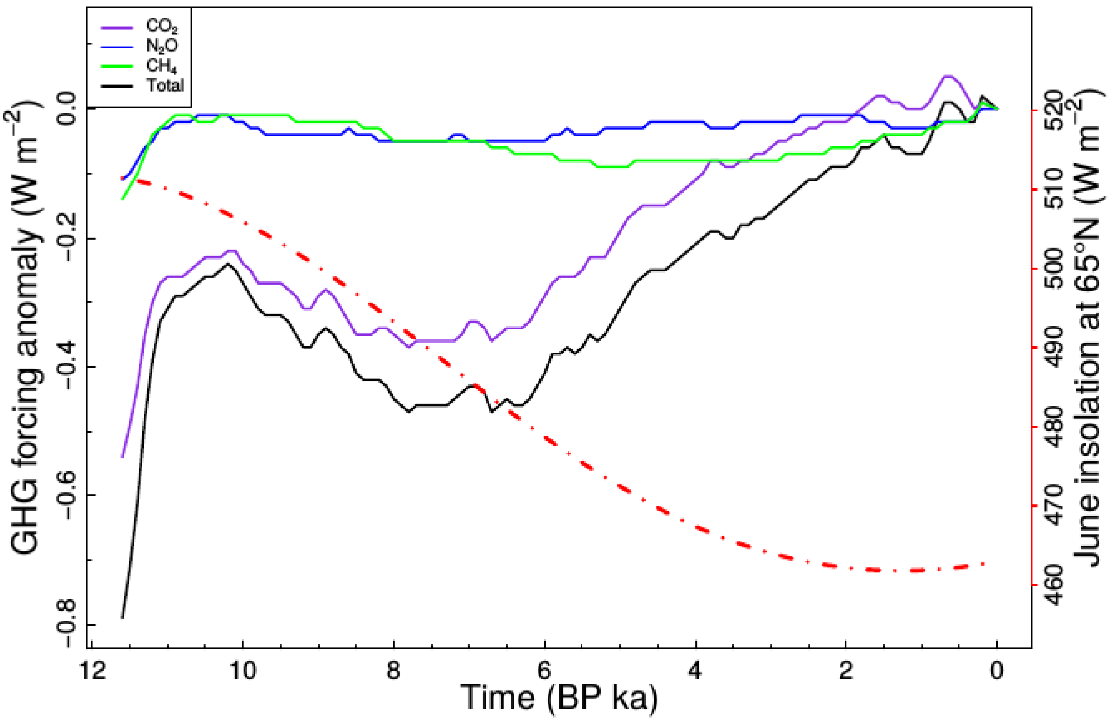
Fig. 11. Geopotential height (m) at 800 hPa in the extratropical Northern Hemisphere. (a) shows the control condition PI, (b) and (c) are the simulations OG11.5 and OGIS11.5.

Fig. 12. Meridional overturning streamfunction (Sv) in the Atlantic Ocean Basin. (a), (b) and (c) indicate the control run (PI), the simulation OG11.5 and OGIS11.5 respectively. On the left hand side, depth is indicated in meters. Positive values indicate a clockwise circulation.

Fig. 13. Minimum sea-ice thickness (m) in September for PI (a), OG11.5 (b) and OGIS11.5 (c).

Fig. 14. Response of the ocean variables (shown as a 100 yrs average) to forcings during the Holocene. (a) Maximum meridional overturning streamfunction (Sv) in the North Atlantic. (b) Sea-ice area (10^{12} m^2) in the Northern Hemisphere. The red line indicates the ORBGHG simulation. The black and green lines reflect results of the simulation OGIS_FWF-v1 and OGIS_FWF-v2.

Figure 1



CO_2
N_2O
CH_4
Total
GHG forcing anomaly (W m^{-2})
0.0
−0.2
−0.4
−0.6
−0.8
June insolation at 65°N (W m^{-2})
520
510
500
490
480
470
460
12
10
8
6
4
2
0
Time (BP ka)

Figure 2

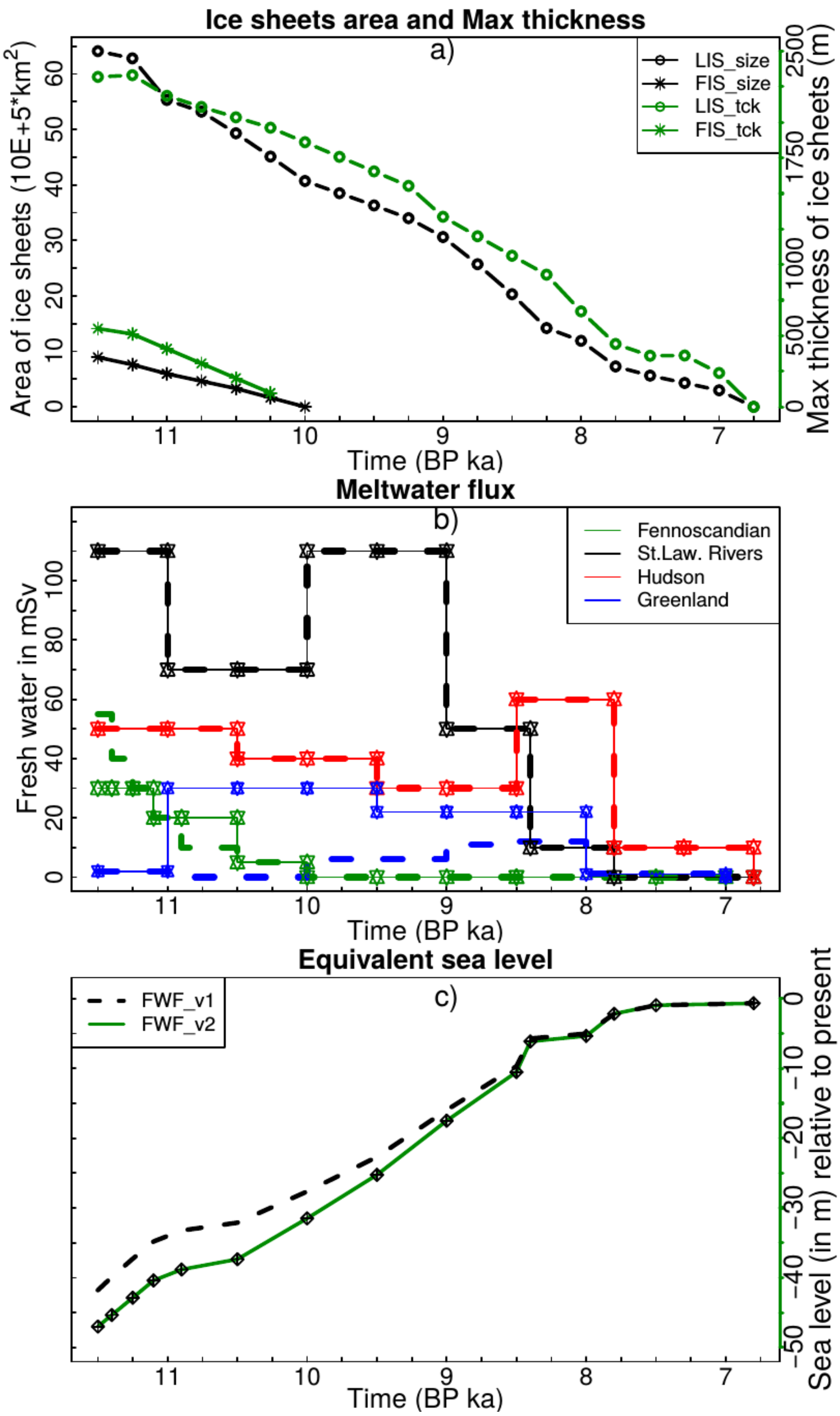

Ice sheets area and Max thickness
a)
LIS_size
FIS_size
LIS_tck
FIS_tck
Area of ice sheets (10E+5*km2)
Max thickness of ice sheets (m)
Time (BP ka)
Meltwater flux
b)
Fennoscandian
St.Law. Rivers
Hudson
Greenland
Fresh water in mSv
Time (BP ka)
Equivalent sea level
c)
FWF_v1
FWF_v2
Sea level (in m) relative to present
Time (BP ka)

Figure 3

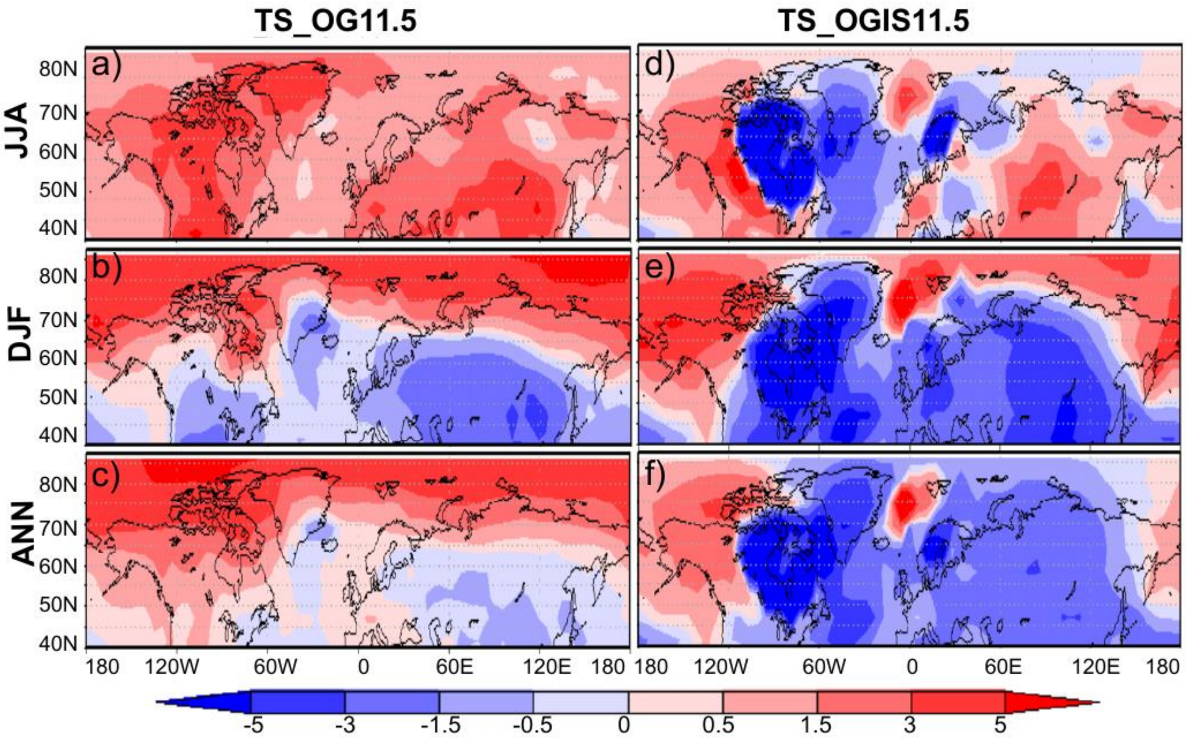
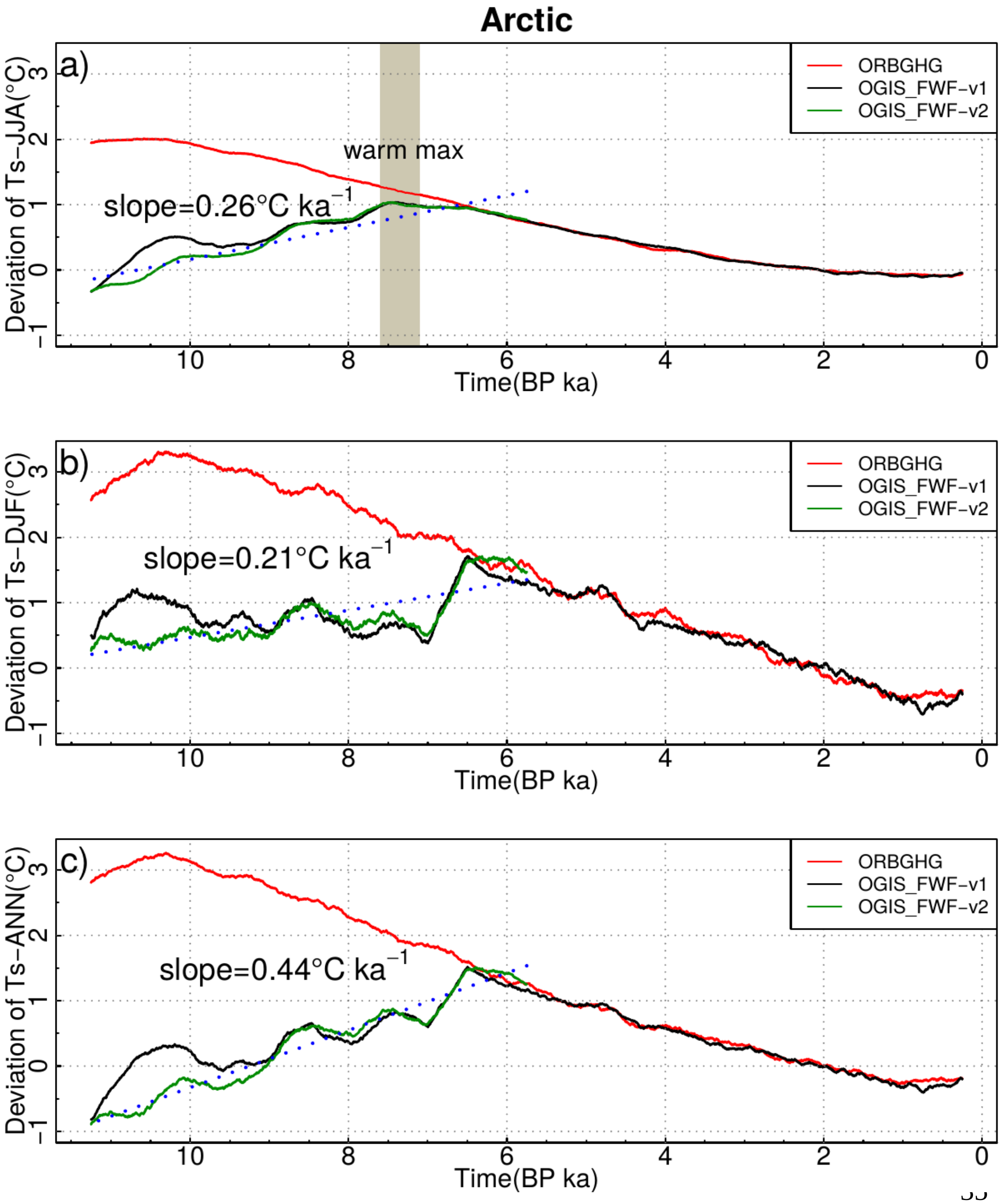

TS_OG11.5
TS_OGIS11.5
JJA
DJF
ANN
a)
b)
c)
d)
e)
f)
80N
70N
60N
50N
40N
180
120W
60W
0
60E
120E
180
-5
-3
-1.5
-0.5
0
0.5
1.5
3
5

Figure 4


Arctic
a)
ORBGHG
OGIS_FWF-v1
OGIS_FWF-v2
warm max
slope=0.26°C ka−1
Deviation of Ts−JJA(°C)
Time(BP ka)
b)
slope=0.21°C ka−1
Deviation of Ts−DJF(°C)
c)
slope=0.44°C ka−1
Deviation of Ts−ANN(°C)

40

Figure 5

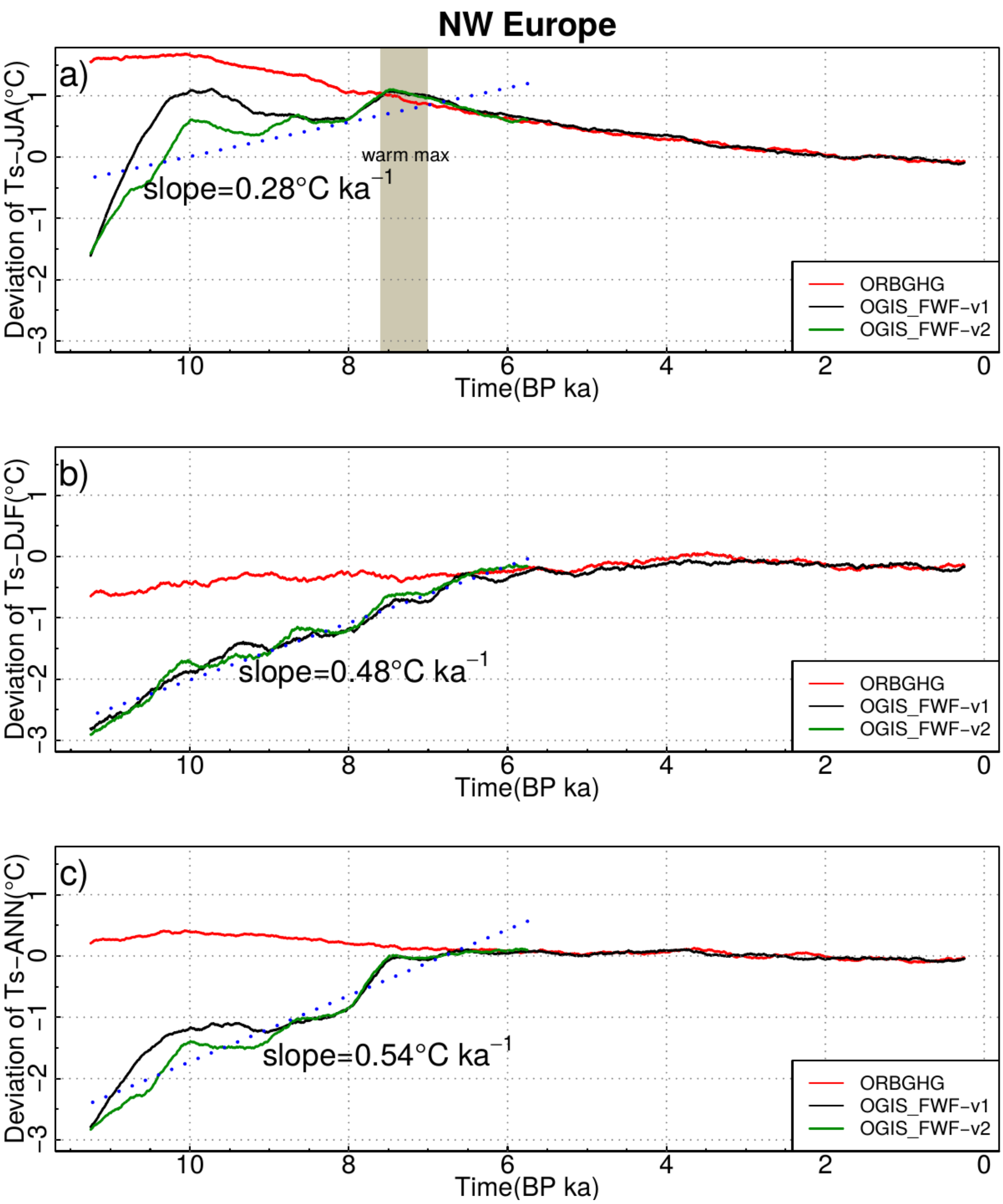
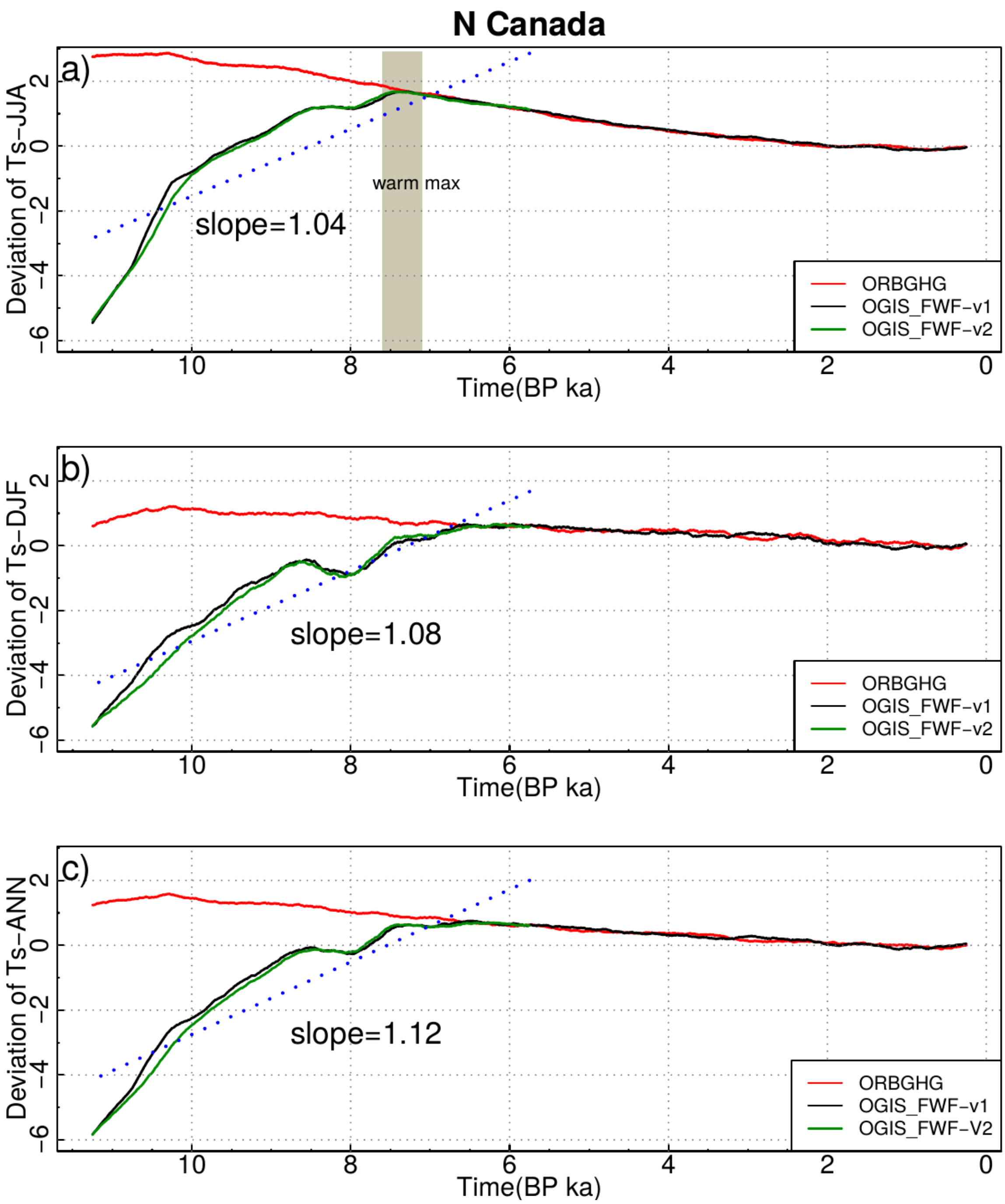
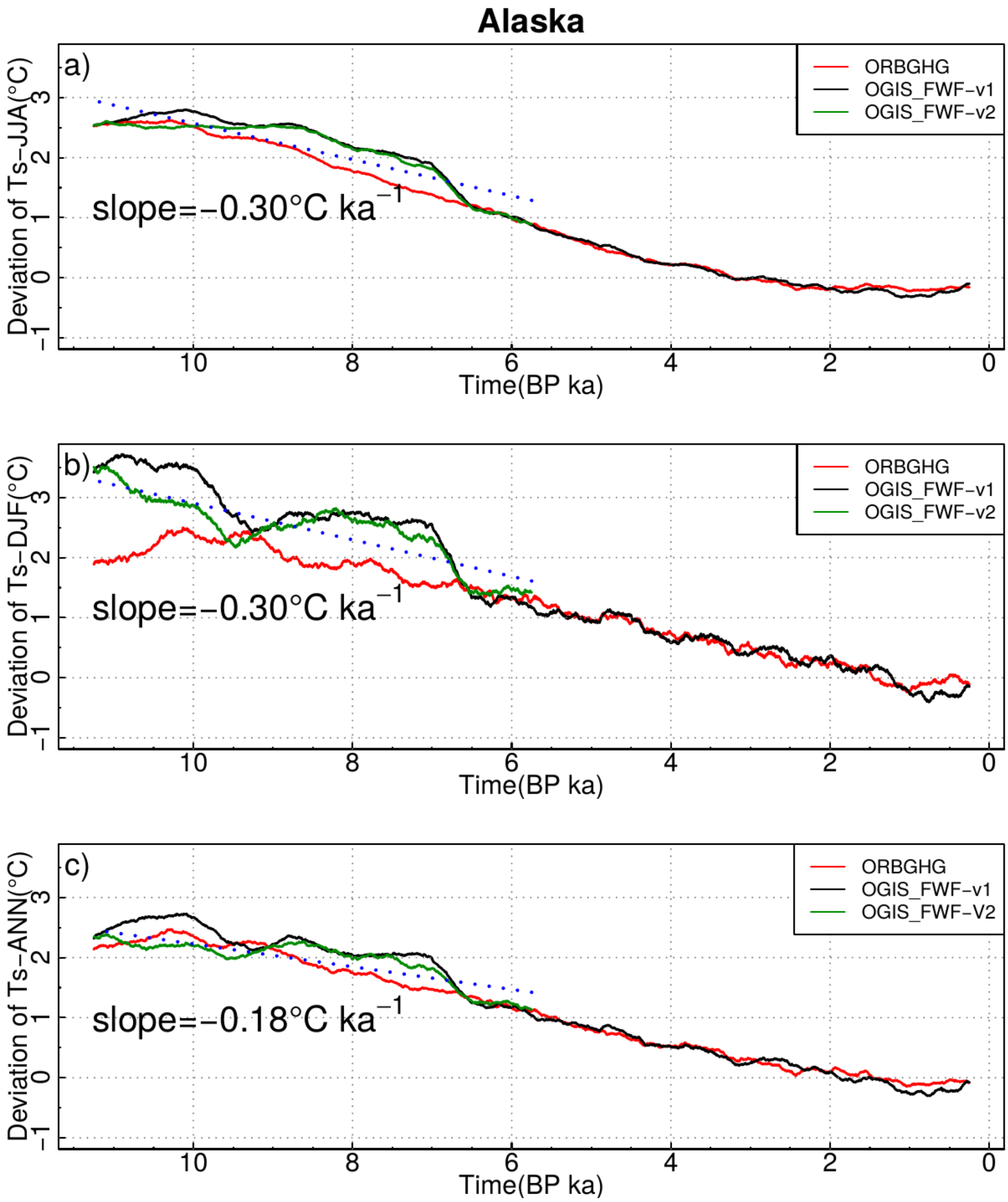

NW Europe
a)
warm max
slope=0.28°C ka−1
Deviation of Ts−JJA(°C)
Time(BP ka)
ORBGHG
OGIS_FWF−v1
OGIS_FWF−v2
b)
slope=0.48°C ka−1
Deviation of Ts−DJF(°C)
Time(BP ka)
ORBGHG
OGIS_FWF−v1
OGIS_FWF−v2
c)
slope=0.54°C ka−1
Deviation of Ts−ANN(°C)
Time(BP ka)
ORBGHG
OGIS_FWF−v1
OGIS_FWF−v2

Figure 6


N Canada
a)
Deviation of Ts−JJA
warm max
slope=1.04
ORBGHG
OGIS_FWF−v1
OGIS_FWF−v2
Time(BP ka)
b)
Deviation of Ts−DJF
slope=1.08
ORBGHG
OGIS_FWF−v1
OGIS_FWF−v2
Time(BP ka)
c)
Deviation of Ts−ANN
slope=1.12
ORBGHG
OGIS_FWF−v1
OGIS_FWF−V2
Time(BP ka)

5

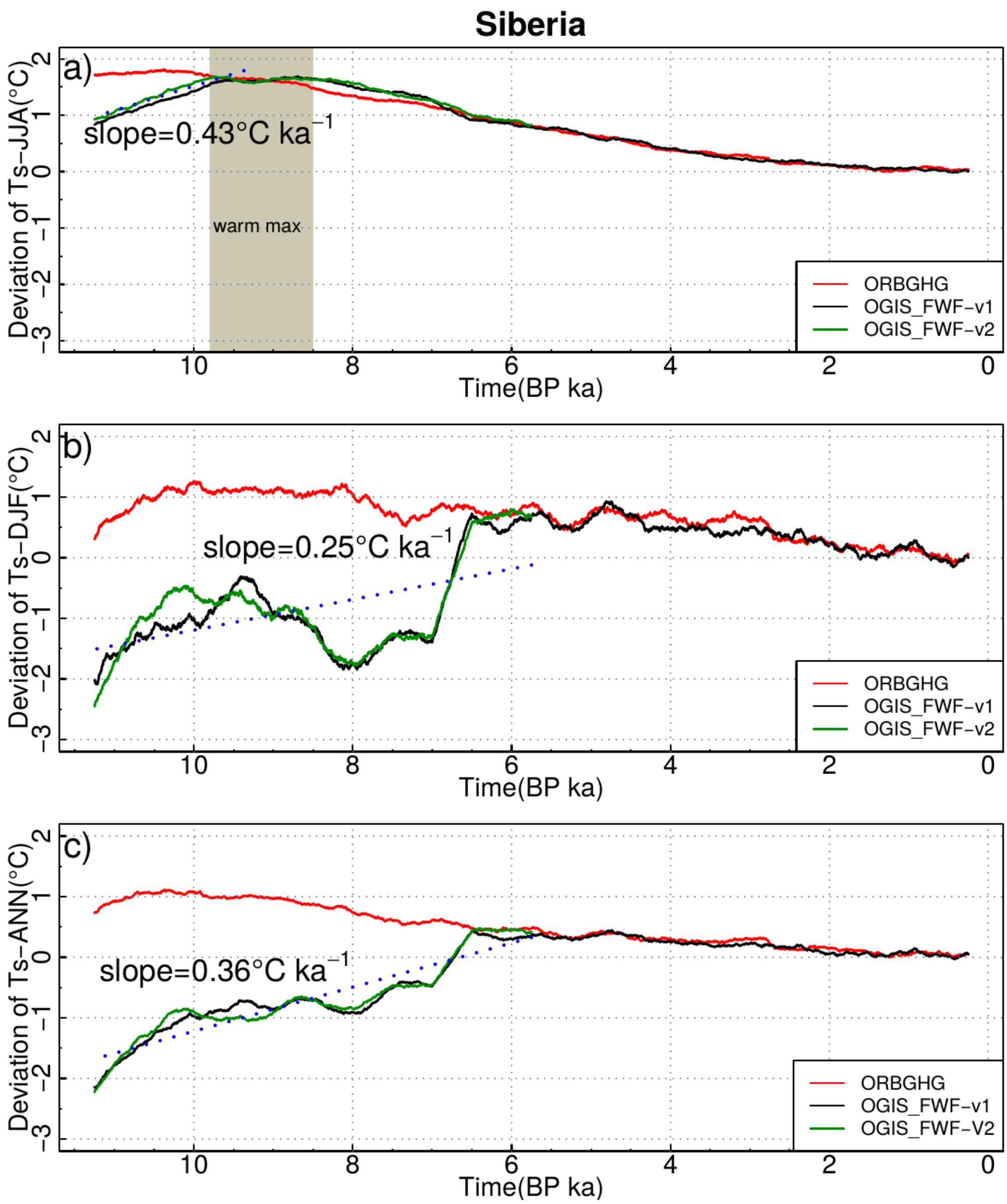
Figure 7



Alaska
a)
b)
c)
ORBGHG
OGIS_FWF-v1
OGIS_FWF-v2
OGIS_FWF-V2
slope=−0.30°C ka−1
slope=−0.30°C ka−1
slope=−0.18°C ka−1
Deviation of Ts−JJA(°C)
Deviation of Ts−DJF(°C)
Deviation of Ts−ANN(°C)
Time(BP ka)
10
8
6
4
2
0

5

Figure 8


Siberia
a)
slope=0.43°C ka−1
warm max
Deviation of Ts−JJA(°C)
Time(BP ka)
ORBGHG
OGIS_FWF−v1
OGIS_FWF−v2
b)
slope=0.25°C ka−1
Deviation of Ts−DJF(°C)
Time(BP ka)
ORBGHG
OGIS_FWF−v1
OGIS_FWF−v2
c)
slope=0.36°C ka−1
Deviation of Ts−ANN(°C)
Time(BP ka)
ORBGHG
OGIS_FWF−v1
OGIS_FWF−V2

5

Figure 9

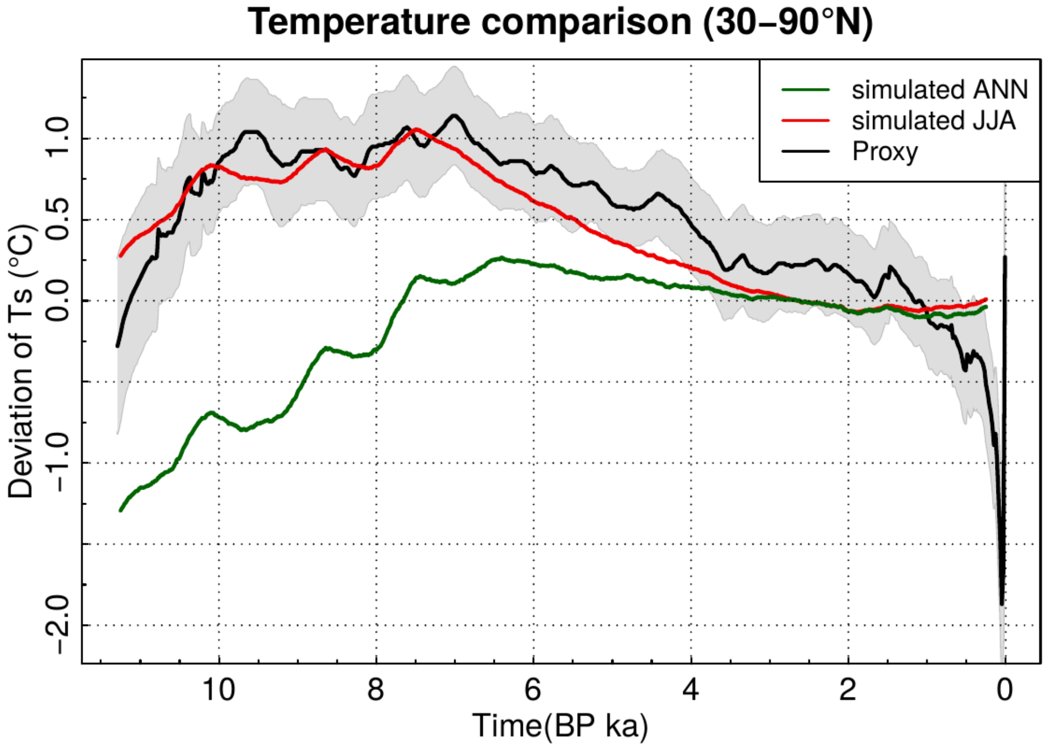

Temperature comparison (30–90°N)
simulated ANN
simulated JJA
Proxy
Deviation of Ts (°C)
1.0
0.5
0.0
−1.0
−2.0
10
8
6
4
2
0
Time(BP ka)

Figure 10

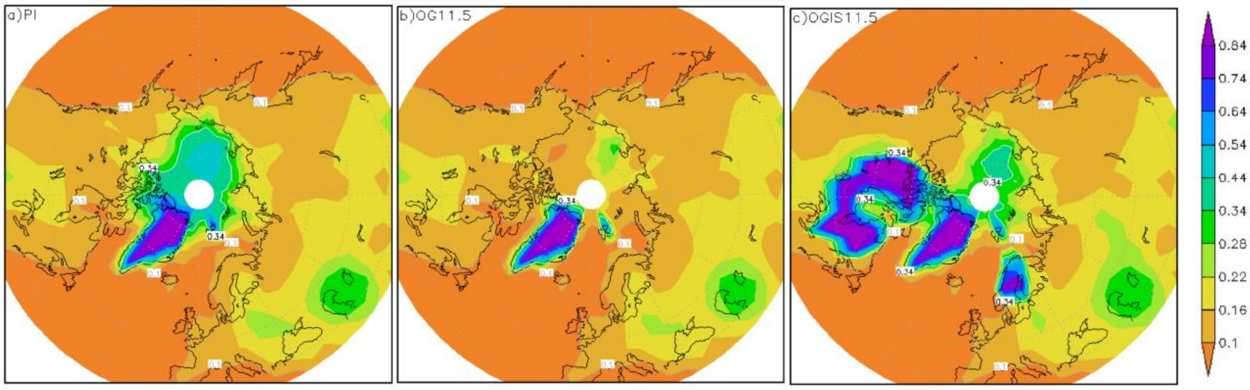


Figure 11

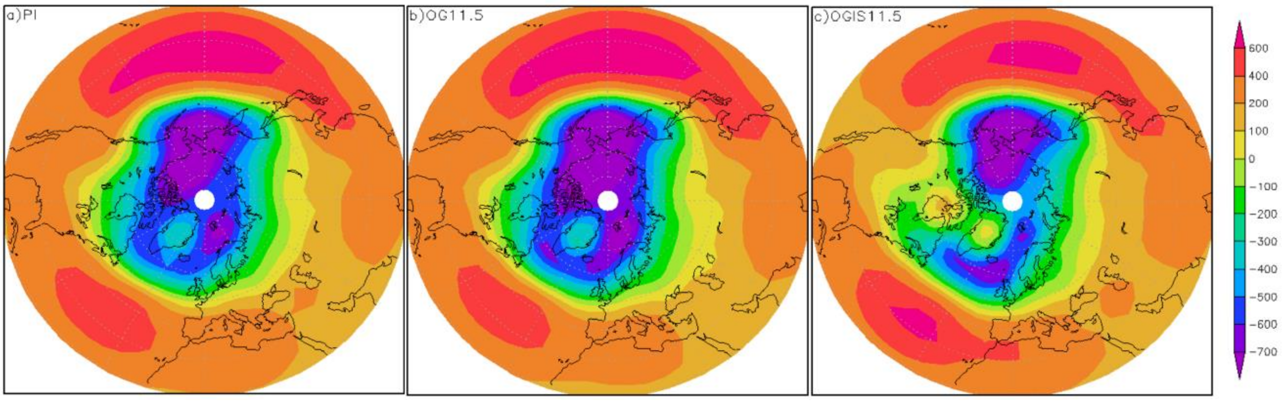

a)PI
b)OG11.5
c)OGIS11.5
600
400
200
100
0
−100
−200
−300
−400
−500
−600
−700

Figure 12

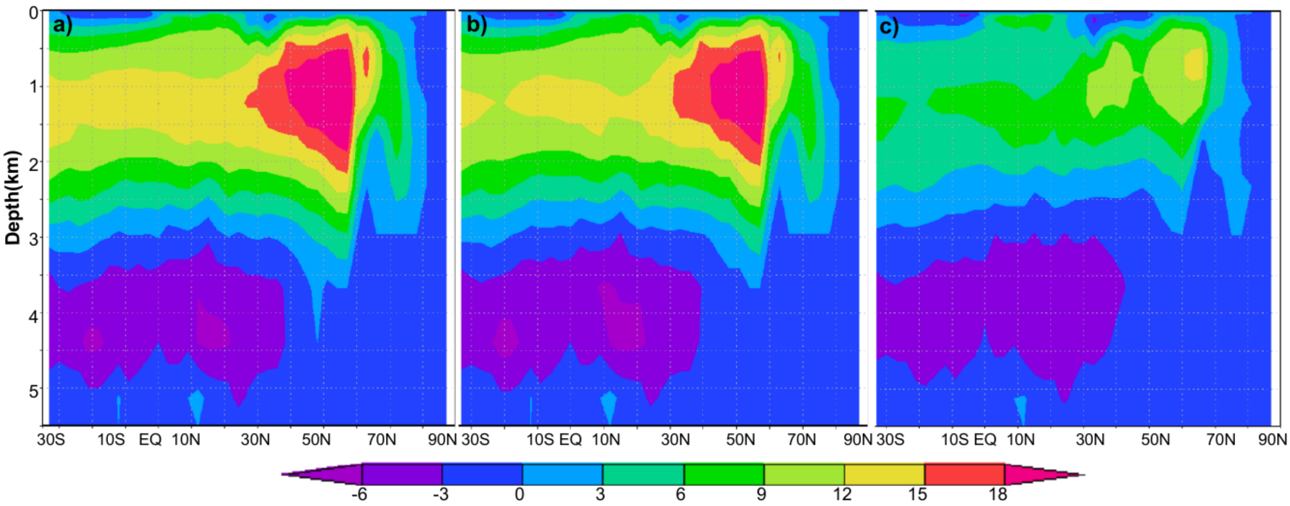

a)
b)
c)
Depth(km)
0
1
2
3
4
5
30S
10S
EQ
10N
30N
50N
70N
90N
-6
-3
0
3
6
9
12
15
18

Figure 13

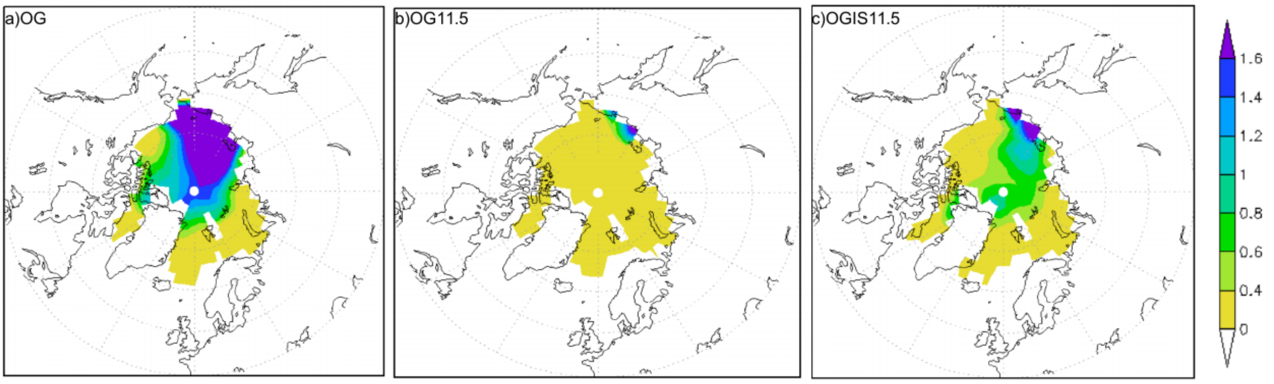
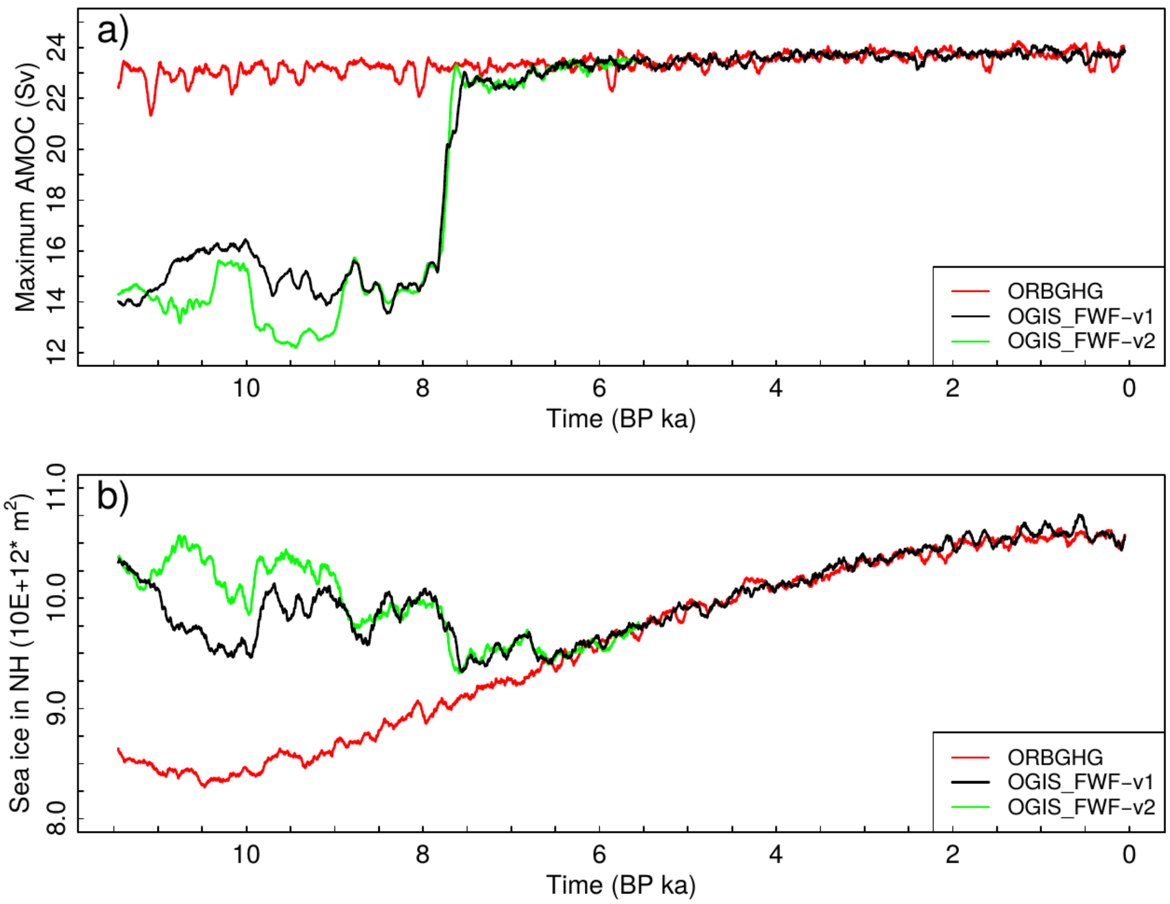


Figure 14


a)
Maximum AMOC (Sv)
12
14
16
18
20
22
24
10
8
6
4
2
0
Time (BP ka)
ORBGHG
OGIS_FWF-v1
OGIS_FWF-v2
b)
Sea ice in NH (10E+12* m^2)
8.0
9.0
10.0
11.0
10
8
6
4
2
0
Time (BP ka)
ORBGHG
OGIS_FWF-v1
OGIS_FWF-v2