The sensitivity of the Late Saalian (140 ka) and LGM (21 ka) Eurasian ice sheets to sea surface conditions

Florence Colleoni · Johan Liakka · Gerhard Krinner · Martin Jakobsson · Simona Masina · Vincent Peyaud

Received: 29 December 2009 / Accepted: 14 June 2010
© Springer-Verlag 2010

Abstract This work focuses on the Late Saalian (140 ka) Eurasian ice sheets’ surface mass balance (SMB) sensitivity to changes in sea surface temperatures (SST). An Atmospheric General Circulation Model (AGCM), forced with two preexisting Last Glacial Maximum (LGM, 21 ka) SST reconstructions, is used to compute climate at 140 and 21 ka (reference glaciation). Contrary to the LGM, the ablation almost stopped at 140 ka due to the climatic cooling effect from the large ice sheet topography. Late Saalian SST are simulated using an AGCM coupled with a mixed layer ocean. Compared to the LGM, these 140 ka SST show an inter-hemispheric asymmetry caused by the larger ice-albedo feedback, cooling climate. The resulting Late Saalian ice sheet SMB is smaller due to the extensive simulated sea ice reducing the precipitation. In conclusion, SST are important for the stability and growth of the Late Saalian Eurasian ice sheet.

Keywords Sea surface temperatures · Late Saalian · Last Glacial Maximum · Eurasian ice sheet · Climate modelling · Quaternary

1 Introduction

The recent reconstructions of the Eurasian ice sheet spanning the last 250,000 years (Svendsen et al. 2004) show that during the Late Saalian glaciation, with its maximum at about 140 ka, the Eurasian ice sheet extended over nearly 11 × 10⁶ km² and reached further southward and eastward than during the Last Glacial Maximum (LGM, 21 ka) (Fig. 1). Numerical ice sheet model results suggest that the Late Saalian Eurasian ice sheet was ≈ 54% larger in area than the LGM ice sheet and at least double the thickness (Peyaud 2006; Lambeck et al. 2006; Peltier 2004). This raises the question: Why did the Late Saalian ice sheet grow so large over Eurasia?

The ice age cycles are generally thought to be ultimately driven by cyclical variations in the Earth’s orbit around the sun (e.g. Hays et al. 1976), although glacial inception results from a complex interaction between processes such as for example land-ice feedbacks, atmosphere-ice feedbacks, and ocean forcing. In comparison to the LGM, the Late Saalian period is characterized by a larger eccentricity of the Earth’s orbit (Berger and Loutre 1991) which enhances the effect of precession (Fig. 1). This difference in orbital configurations directly influences the amount of...
solar downward radiation at top of the atmosphere (Fig. 2). There were consequently cooler and longer Late Saalian springs and early summers compared to during LGM while there were at the same time shorter Late Saalian falls in the Northern Hemisphere high latitudes (Table 1). The nature of glacial maxima also depends on the evolution of the solar insolation, as a function of the Earth’s orbital parameters, prior to the glacial inception. This evolution was markedly different between the Late Saalian and the LGM (Fig. 1). Greenhouse gas concentrations seem to have been similar during these two glacial periods (Petit et al. 2001; Spahni et al. 2005) (Fig. 1) and therefore cannot directly explain the difference in ice extent and volume of the Eurasian ice sheet. Although it is possible to point out differences in orbital forcing between the Late Saalian and the LGM glacial periods, the question why the Eurasian ice sheet extended much further to the east and south during the Late Saalian remains unanswered.

Part of the answer might reside in different regional continental and oceanic conditions during the two glacial maxima such as the presence of ice dammed lakes, dust deposition on snow, changes in vegetation and in the ocean surface conditions (Krinner et al. 2004; Colleoni et al. 2009a, b). There seem not to have been any ice dammed lakes in front of the Eurasian ice sheet during LGM (Mangerud et al. 2004) while the Late Saalian ice sheet configuration suggests that rather large lakes may have developed (Colleoni et al. 2009b). The LGM river network near the Taymyr Peninsula was likely still connected to the Arctic Ocean due to the relatively small extent of the Eurasian ice sheet at this time while it was certainly not the case during the Late Saalian (Svendsen et al. 2004). According to Antarctic ice core dust concentration records, the dust transportation into the atmosphere was larger during the LGM than during the Late Saalian (Delmonte et al. 2004). In previous studies we have shown that during the Late Saalian, the presence of large ice dammed lakes and reduced dust deposition on snow directly contributed to the relative stability of the Eurasian ice sheet at 140 ka (Colleoni et al. 2009b) and that the Late Saalian steady-state modeled vegetation positively influenced the ice sheet surface mass balance (SMB) (Colleoni et al. 2009a). However, in those studies, we forced an Atmospheric General Circulation Model (AGCM) using CLIMAP (CLIMAP 1984) and Paul and Schaefer-Neth (2003) LGM sea surface temperature (SST) reconstructions since no similar datasets have been produced for the Late Saalian. This limits the interpretation of the other regional feedbacks whose amplitude and intensity can be influenced by the climatic context and the variations in ice sheet topography over continents (Kageyama et al. 2006). In this work, we investigate the influence of SST on the Late Saalian and LGM Northern Hemisphere climates and on the SMB of the Eurasian ice sheets during these periods.

Most of the previous modeling studies addressing the influence of SST and the sea-ice cover on glacial climate focused on the LGM since there is a considerably larger...
number of continental and marine geological records available to compare with the numerical simulation results. Some of these studies forced an AGCM with prescribed LGM SST reconstructions (e.g. Kageyama et al. 1999) to which a regional temperature perturbation is eventually applied to explore the impacts of different North Atlantic or tropical SSTs on the Northern Hemisphere climate (e.g. Rind 2006; Toracinta et al. 2004). In other studies, SSTs are computed using a coupled Atmosphere-Ocean GCM (e.g. Ramstein and Joussaume 1995; Dong and Valdes 1998). Within the framework of the Paleoclimate Model Intercomparison Projects (PMIP) I and II, Kageyama et al. (1999, 2006) showed that the use of prescribed and simulated SST in AGCM experiments leads to different climate-related (mostly snow fall and temperatures) oceanic and continental climate anomaly patterns of various amplitudes over the North Atlantic and Eurasia. These discrepancies are explained by the use of different models as the computed climate depends on the physics of each model and on the models own inter-annual variability partly resulting from the strong fluctuations of the sea-ice cover extent. The extent of sea ice also has a large impact on the transient atmospheric activity, which in turn influences the regional climate (Dong and Valdes 1998). Compared to present-day Northern Hemisphere sea-ice extent, the larger LGM sea-ice cover pulled the storm tracks eastward (Kageyama et al. 1999). Ruddiman and McIntyre (1979) and Hebbeln et al. (1994) show that seasonally open waters in the North Atlantic, such as suggested by the recent MARGO LGM SST reconstructions (Kucera et al. 2005a), significantly influence the SMB of the Northern Hemisphere ice sheets during the LGM.

There are a number of studies clearly showing that sea surface conditions, primarily SST, strongly influence the atmosphere due to effects on the mean oceanic heat exchange. Forcings from different sea surface conditions (prescribed or calculated) also modulate the model simulated climate feedbacks. Consequently, it is important to determine how different sea surface temperatures and sea-ice extent reconstructions can influence the surface- atmosphere interactions during the Late Saalian when addressing the underlying reasons for the large ice extent.

Were the Late Saalian SSTs different from the LGM ones? This question is currently difficult to answer because there are few SST estimates based on marine sediment units from Marine Isotope Stage (MIS) 6, which correspond to the time period of the Late Saalian. Some of these estimates are listed in Table 2 together with the LGM SST estimates allowing a direct comparison for each coring site. From this table, it is not obvious whether the temperatures of the Late Saalian and the LGM surface oceans were significantly different.

In summary, to understand how SST could have influenced the SMB of the Late Saalian Eurasian ice sheet, we first test the effects of two different prescribed LGM SST reconstructions on the Late Saalian climate by forcing an AGCM model with these two datasets. The same runs are

### Table 1

<table>
<thead>
<tr>
<th></th>
<th>140 ka</th>
<th>21 ka</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter</td>
<td>90</td>
<td>93</td>
</tr>
<tr>
<td>Spring</td>
<td>97</td>
<td>89</td>
</tr>
<tr>
<td>Summer</td>
<td>92</td>
<td>90</td>
</tr>
<tr>
<td>Fall</td>
<td>86</td>
<td>94</td>
</tr>
</tbody>
</table>
performed for the LGM to estimate the Eurasian ice sheet sensitivity to the oceanic surface conditions. Then we derive Late Saalian SST through a modeling experiment. As a first step, we use a simplified AGCM coupled to an ocean mixed layer model to simulate the Late Saalian SST (the LGM SST are simulated as well as a control experiment). As a second step, we estimate the influence of those simulated Late Saalian SST on the surface mass balance of the Late Saalian Eurasian ice sheet and on the Northern Hemisphere climate to finally understand if these simulated oceanic surface conditions contribute to the stability of this ice sheet over the continent.

2 Methods and settings

This work is divided into two distinct parts:

1. the sensitivity of the Late Saalian and LGM Eurasian ice sheets with respect to the prescribed LGM SST reconstructions by CLIMAP (1984) and Paul and Schaefer-Neth (2003).

2. the sensitivity of the Late Saalian Eurasian ice sheet to simulated SST

Concerning the first topic, the Late Saalian and the LGM climates are explored and compared through 21 year snap-shots generated using the LMDZ4 AGCM (see description below, Hourdin et al. 2006). For the second topic, Late Saalian SSTs are computed using the Planet Simulator AGCM mixed layer model (Fraedrich et al. 2005). The Late Saalian climate is computed by the LMDZ4 using these simulated Late Saalian SST. These results are then compared with the Late Saalian climate previously obtained forcing the LMDZ4 with Paul and Schaefer-Neth (2003) LGM SST. The outline of this procedure is presented in Fig. 3.

2.1 Late Saalian atmospheric simulations: the LMDZ4 model

Twenty-one year snap-shots are performed using the LMDZ4 atmospheric general circulation model (Hourdin et al. 2006) which takes into account the climatic impact of

Table 2 Measured SST from various sources for both LGM (21 ka) and Late Saalian (140 ka)

<table>
<thead>
<tr>
<th>Ocean</th>
<th>SST (°C)</th>
<th>Season</th>
<th>Method</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>21 ka</td>
<td>140 ka</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Atl.</td>
<td>12.5</td>
<td>13.8</td>
<td>Annu.</td>
<td>Alk.</td>
</tr>
<tr>
<td>Atl.</td>
<td>21.5</td>
<td>21.5</td>
<td>Annu.</td>
<td>Alk.</td>
</tr>
<tr>
<td>Atl.</td>
<td>18.0</td>
<td>18.0</td>
<td>Annu.</td>
<td>Alk.</td>
</tr>
<tr>
<td>Atl.</td>
<td>9.8</td>
<td>11.5</td>
<td>Annu.</td>
<td>Alk.</td>
</tr>
<tr>
<td>Atl.</td>
<td>16.5</td>
<td>17.0</td>
<td>Annu.</td>
<td>Alk.</td>
</tr>
<tr>
<td>Atl.</td>
<td>23.0</td>
<td>24.0</td>
<td>Annu.</td>
<td>Alk.</td>
</tr>
<tr>
<td>Pacif.</td>
<td>22.5</td>
<td>22.5</td>
<td>Annu.</td>
<td>Mg/Ca</td>
</tr>
<tr>
<td>Pacif.</td>
<td>26.0</td>
<td>26.0</td>
<td>Annu.</td>
<td>Mg/Ca</td>
</tr>
<tr>
<td>Pacif.</td>
<td>19.0</td>
<td>19.0</td>
<td>Annu.</td>
<td>Mg/Ca</td>
</tr>
<tr>
<td>Pacif.</td>
<td>22.0</td>
<td>21.25</td>
<td>Annu.</td>
<td>Mg/Ca</td>
</tr>
<tr>
<td>Pacif.</td>
<td>21.0</td>
<td>22.0</td>
<td>Annu.</td>
<td>44Ca/40Ca</td>
</tr>
<tr>
<td>Ant.</td>
<td>6.6</td>
<td>7.0</td>
<td>Annu.</td>
<td>Mg/Ca</td>
</tr>
<tr>
<td>Atl.</td>
<td>5.0</td>
<td>8.0</td>
<td>Sum.</td>
<td>foram</td>
</tr>
<tr>
<td>Atl.</td>
<td>3.5</td>
<td>3.0</td>
<td>Sum.</td>
<td>Mg/Ca</td>
</tr>
<tr>
<td>Atl.</td>
<td>11.75</td>
<td>12.5</td>
<td>Sum.</td>
<td>mod. ana.</td>
</tr>
<tr>
<td>Pacif.</td>
<td>27.4</td>
<td>28</td>
<td>Sum.</td>
<td>plane. trans.</td>
</tr>
<tr>
<td>Pacif.</td>
<td>12.5</td>
<td>11</td>
<td>Sum.</td>
<td>mod. ana.</td>
</tr>
<tr>
<td>Pacif.</td>
<td>4.0</td>
<td>3.5</td>
<td>Sum.</td>
<td>mod. ana.</td>
</tr>
<tr>
<td>Pacif.</td>
<td>23.0</td>
<td>23.0</td>
<td>Win.</td>
<td>plane. trans.</td>
</tr>
<tr>
<td>Pacif.</td>
<td>9.0</td>
<td>8.0</td>
<td>Win.</td>
<td>mod. ana.</td>
</tr>
<tr>
<td>Atl.</td>
<td>11.0</td>
<td>13.0</td>
<td>Win.</td>
<td>plane. trans.</td>
</tr>
<tr>
<td>Atl.</td>
<td>0.0</td>
<td>0.0</td>
<td>Win.</td>
<td>Mg/Ca</td>
</tr>
</tbody>
</table>

Data are classified according to annual values (Annu.), summer values (Sum.) and winter values (Win.). Data have been measured and inverted according different methods: alkenone (Alk.), Mg/Ca ratio, foraminifera.
The model is run with 96 grid cells and with 19 vertical layers. The horizontal resolution is irregular, varying from 100 km in Eurasia (at 65°N/60°E) to 550 km elsewhere.

### 2.1.1 LGM boundary conditions

In the LGM experiments, we use the ICE-5G ice sheet extent and topography provided by Peltier (2004), assuming that the differences between the Late Saalian glacial maximum and LGM geographies far from our region of interest are sufficiently small to not critically influence our results. The Eurasian Late Saalian maximum ice sheet topography was computed with the GRISLI ice model (Ritz et al. 2001; Peyaud 2006; Peyaud et al. 2007) using the reconstructed ice limits from Svendsen et al. (2004) and taking into account lithospheric deflection (Fig. 1). Eustatic sea-level is set at 110 m below present-day sea-level according to Shackleton et al. (1990) and Peyaud (2006).

During this period, ice dammed lakes formed along the southern margins of the Eurasian ice sheet (Colleoni et al. 2009b), dust transportation within the atmosphere was reduced compared to that of the LGM (Delmonte et al. 2004; Winckler et al. 2008) and vegetation might have been different from that of Crowley (1995) prescribed as first approximation in our simulations. The impact of these parameters on the regional climate and on the SMB of the Eurasian ice sheet has been further addressed in Colleoni et al. (2009a, b). In this present work orbital parameters have been set according to Berger and Loutre (1991) and greenhouse gas (GHG) concentrations (Fig. 1) have been specified according to Vostok and EPICA Dome C ice core concentration records (Petit et al. 2001; Spahni et al. 2005).

### 2.1.2 Late Saalian boundary conditions

With the exception for in the Eurasian region, we used the LGM global ice topography and ice sheet extents provided by Peltier (2004), assuming that the differences between the Late Saalian glacial maximum and LGM geographies far from our region of interest are sufficiently small to not critically influence our results. The Eurasian Late Saalian maximum ice sheet topography was computed with the GRISLI ice model (Ritz et al. 2001; Peyaud 2006; Peyaud et al. 2007) using the reconstructed ice limits from Svendsen et al. (2004) and taking into account lithospheric deflection (Fig. 1). Eustatic sea-level is set at 110 m below present-day sea-level according to Shackleton et al. (1990) and Peyaud (2006).

During this period, ice dammed lakes formed along the southern margins of the Eurasian ice sheet (Colleoni et al. 2009b), dust transportation within the atmosphere was reduced compared to that of the LGM (Delmonte et al. 2004; Winckler et al. 2008) and vegetation might have been different from that of Crowley (1995) prescribed as first approximation in our simulations. The impact of these parameters on the regional climate and on the SMB of the Eurasian ice sheet has been further addressed in Colleoni et al. (2009a, b). In this present work orbital parameters have been set according to Berger and Loutre (1991) and greenhouse gas (GHG) concentrations (Fig. 1) have been specified according to Vostok and EPICA Dome C ice core concentration records (Petit et al. 2001; Spahni et al. 2005).

### 2.2 Simulating the Late Saalian SST: the Planet Simulator model

In order to compute the Late Saalian surface ocean conditions, we use the Planet Simulator (PLASIM) general circulation model of intermediate complexity (Fraedrich et al. 2005) because the LMDZ4 does not include an oceanic mixed layer. The atmospheric component of PLASIM is an AGCM, which is based on the moist primitive equations representing conservation of momentum, mass and energy. The equations are solved on a Gaussian grid in the horizontal direction, corresponding to a T42 resolution (128 × 64) in the horizontal direction, and on 10 vertical sigma levels.

The AGCM is interactively coupled to a mixed layer ocean (mixed layer depth is set to 50 m by default) and to a

---

open water surfaces and dust concentration in snow (Krinner 2003; Krinner et al. 2006). The model is run with 96 × 72 grid cells and with 19 vertical layers. The horizontal resolution is irregular, varying from 100 km in Eurasia (at 65°N/60°E) to 550 km elsewhere.

![Diagram](image)

**Fig. 3** Scheme of the simulations carried out using Planet Simulator (Fraedrich et al. 2005) to compute the Late Saalian SSTs and using the LMDZ4 (Hourdin et al. 2006) to obtain a final Late Saalian climate. Methods and boundary conditions for each stage are detailed in Sects. 2.2 and 2.1.2.
simple thermodynamic sea-ice model. In this model, the oceanic heat transport is monthly prescribed and is parameterized according to:

$$Q_{c2} = \rho_w c_w h_{\text{mix}} \frac{T_{\text{mix}} - T_{\text{mix}}}{\tau_T}$$

where $c_w$ is the specific heat capacity of sea water, $\rho_w$ corresponds to the density of sea water, $h_{\text{mix}}$ stands for the mixed layer depth, $T_{\text{mix}}$ is the temperature of the mixed layer and $\tau_T$ represents the time scale at which $T_{\text{mix}}$ is relaxed to its climatological value $T_{\text{mix},c}$ ($\tau_T = 50$ days). This method makes it possible to calculate SST and has been previously used by Romanova et al. (2005, 2006) to investigate the effect of ocean heat fluxes on glacial climates.

In the absence of any available Late Saalian sea surface reconstructions, the oceanic heat transports are calculated using Paul and Schaefer-Neth (2003) LGM SST reconstructions. To extract the monthly oceanic heat fluxes, we performed a LGM control simulation of 21 years forcing PLASIM with Paul and Schaefer-Neth (2003) LGM sea surface conditions. We choose to base the calculation on Paul and Schaefer-Neth (2003) LGM SST because this reconstruction is in closer agreement with the temperatures estimated from analysis of marine sediment cores than the LGM CLIMAP reconstruction (CLIMAP 1984). LGM zonally averaged ocean heat fluxes are plotted together with the present-day ocean heat fluxes (Rayner et al. 2003) for comparison in Fig. 4b. In this LGM control run, we use ICE-5G ice topography (Peltier 2004), vegetation is prescribed as the fraction of forest extracted from the Crowley (1995) LGM vegetation map, the CO$_2$ concentration is set to 194 ppm (Petit et al. 2001) and the orbital parameters are calculated for 21 ka according to Berger and Loutre (1991).

To validate the method and prevent model deviation, we computed the LGM SST (referred to as PS21) forcing PLASIM (50 model-years) with the monthly LGM ocean heat fluxes extracted from the previous LGM control simulation. The difference between the simulated LGM SST and the Paul and Schaefer-Neth (2003) LGM SST reconstruction shows no significant anomaly except for in the Indian ocean, off the Australian coasts (Fig. 4a). The sea-ice cover is almost similar to that of Paul and Schaefer-Neth (2003) (Fig. 5).

To simulate the Late Saalian surface oceanic conditions (referred to as PS140), PLASIM is forced using the calculated monthly LGM ocean flux corrections. Orbital parameters are adjusted to 140 ka, CO$_2$ concentration is set to 192 ppm (Petit et al. 2001). The simulation is performed for 50 model-years, sea surface equilibrium is reached after 25 model-years and the last 15 years are used for analysis. The various SST reconstructions used in this work are summarized in Table 3.

**Fig. 4** Mean annual SSTs anomaly between the simulated LGM SST PS21 (LGM control run) and the LGM SST reconstruction of Paul and Schaefer-Neth (2003) (P03). Meridional oceanic heat fluxes computed by Planet Simulator for both present-day (black) and LGM (red, using P03)

2.3 Atmospheric General Circulation Model Simulations

In total, five LMDZ4 snap-shots of 21 years each have been carried out. P03$_{140}$ and CLIM$_{140}$ (140 refers to the Late Saalian glacial maximum) have been forced using Paul and Schaefer-Neth (2003) (referred to as P03) and CLIMAP (1984) (referred to as CLIMAP) LGM sea surface conditions respectively (Table 3 and 4). To allow for comparison, similar experiments have been performed for the LGM period (P03$_{21}$ and CLIM$_{21}$). Since the Paul and Schaefer-Neth (2003) reconstruction agrees better with geological data compared to the CLIMAP global reconstruction, P03$_{140}$ and P03$_{21}$ will be considered as the reference simulations for the Late Saalian and the LGM respectively and all impacts will be expressed as anomalies to them (differences or ratio). Finally, to estimate the impact of the simulated Late Saalian SST on the Late Saalian Eurasian ice sheet, we simulated the climate at 140 ka (FULL$_{140}$), using the LMDZ4 with the prescribed PS140 simulated SST. Since these Late Saalian SSTs have been generated using the computed ocean heat fluxes from the LGM Paul and Schaefer-Neth (2003) reconstruction, FULL$_{140}$ is further compared to P03$_{140}$.
The annual surface mass balance (SMB) of the ice sheet is evaluated according to the temperature index method of Ohmura and Reeh (1996). In this method, surface ablation is diagnosed when mean summer surface air temperature ($T_{JJA}$), recalculated on a fine resolution grid with altitude-correction, exceeds a prescribed threshold of $-1.8^\circ C$. The total ablation (in kg m$^{-2}$ year$^{-1}$) is then calculated as:

$$A = 514^\circ C^{-1} \times T_{JJA} + 930$$  \hspace{1cm} (2)

where $T_{JJA}$ is in $^\circ C$. We downscale the GCM surface air temperature to a 20-km grid by applying a 5.5 $^\circ C$ km$^{-1}$ summer temperature correction to account for altitude changes (Krinner and Genthon 1999; Abe-Ouchi et al. 2007). The surface mass balance $B$ is then calculated as:

$$B = P_S - E - A$$  \hspace{1cm} (3)

where $P_S$ and $E$ represent solid precipitation and evaporation, respectively. The mean annual SMB is calculated for each simulation using the components listed in Table 5.

Student $t$ tests have been performed between P03 and CLIM, and P03 and PS 140 to better assess the effective impact of sea surface conditions on the Northern Hemisphere climate. Since the simulated datasets are available for 20 years each, the degree of freedom is 38 calculated from $N_1 - N_2 - 2$ and results will be accepted if probability is lower than 5%.

### 3 Results

In this section we compare the prescribed P03 and CLIMAP SSTs by analyzing their climatic impacts during the Late Saalian (CLIM$_{140} - P03_{140}$) and then during the LGM (CLIM$_{21} - P03_{21}$). Following this analysis we present the simulated surface ocean reconstruction PS140 and evaluate its impact on the Late Saalian Eurasian climate by comparing simulations FULL$_{140}$ with P03$_{140}$.

#### 3.1 CLIMAP SST versus Paul and Schaefer-Neth (2003) SST

The LGM sea surface temperatures P03 and CLIMAP are generally similar except for in the North Atlantic and in the tropical and equatorial regions. During boreal winter...
both winter and summer (Fig. 6c, d). Between the two reconstructions can be observed during sea-ice cover induces a cooling of & open in P03 southward of Iceland in the east and from 60°N while this area is & colder than P03. Along the tropical and equatorial African shorelines large differences of ≈4 to 8°C between the two reconstructions can be observed during both winter and summer (Fig. 6c, d).

In the North Atlantic high latitudes, the large CLIMAP sea-ice cover induces a cooling of ≈2 to 5°C during both winter and summer. During winter, the CLIMAP sea-ice covers the entire North Atlantic from 45°N while this area is open in P03 southward of Iceland in the east and from 60°N in the west (Fig. 5). The CLIMAP sea-ice limit retreats northward in summer but still covers a large part of the North Atlantic while in P03, the southern portion of the Norwegian Sea becomes open. In the North Pacific, sea-ice remains confined to the Bering Sea and its extent is similar during both winter and summer in P03 while in CLIMAP, the annual cycle of sea-ice extent is more pronounced (Fig. 5).

### 3.1.1 The Late Saalian glacial maximum: comparing CLIM140 and P03140

Compared to P03, the most extensive CLIMAP sea-ice extent causes a large negative temperature anomaly of about 20 to 25°C in the North Atlantic south of Iceland (Fig. 7c). This lowers the temperature over the Arctic Ocean by about 5°C and affects the 500 hPa geopotential height which shows a strong negative anomaly of ≈50 to 110 m over this area (Fig. 7a). Storm tracks are generally stronger where the meridional gradients of surface temperatures are the strongest. In CLIM140, the storm tracks are consequently strengthened along the sea-ice edge (≈45°N). This tends to reduce the Icelandic low located further to the North and causes an increase in sea-level pressure (SLP) by about 8 mb (Fig. 7b). This effect has been documented by Simmonds and Wu (1993) in the present-day Southern Hemisphere where cyclones become weaker over the sea-ice area. An effect of the sea-ice cover is that evaporation almost ceases and, thus, the precipitation drops by approximately 40 to 60% over the North Atlantic (Fig. 7c). Since the CLIMAP sea-ice reaches further South, evaporation and precipitation are concentrated along the sea-ice edge (Fig. 7c–e). The storm tracks activity is intensified and shifted further to East (Fig. 8a, b) compared to P03140. This phenomenon has previously been shown by Kageyama et al (1999) in which the simulated LGM extensive sea-ice cover pulls the storm tracks eastward. An additional result from the regional cooling caused by the CLIMAP sea-ice cover (Fig. 7c) is that precipitation is decreased with about 50% over Central Eurasia. However, the more extensive sea-ice leads to a larger snow mass in Siberia in CLIM140 (Fig. 7f).

SMB is more positive along the southern margins of the Eurasian (300 kg m⁻² year⁻¹) when CLIMAP SST reconstruction is prescribed (Fig. 9b) due to an increase in precipitation by about 30 to 70% over these areas (Fig. 7e). On the contrary, the Atlantic margin exhibits negative SMB values (≈−200 kg m⁻² year⁻¹) due to a reduction in precipitation over the Arctic Ocean also affecting this margin (Fig. 7e). The mean annual SMB is positive in both the CLIM140 and the P03140 simulations due to a small ablation and evaporation (Table 5). The difference between CLIM140 (239 kg m⁻² year⁻¹) and P03140 (225 kg m⁻²...
year\(^{-1}\)) is statistically significant at 5%, but not very strong.

3.1.2 The Last Glacial Maximum: comparing CLIM\(_{21}\) and P03\(_{21}\)

During the LGM, the climatic impact caused by the CLIMAP sea-ice cover shows a pattern similar to the Late Saalian. Compared to P03\(_{21}\), CLIM\(_{21}\) presents a negative anomaly of about 5 to 25°C over the North Atlantic. This negative temperature anomaly spreads over Eurasia and causes a cooling of \(\approx 5^\circ\)C (Fig. 10c). This in turn influences the 500 hPa geopotential height which is characterized by a negative anomaly of about 60 m over this region. As explained previously, storm tracks are generally stronger where the meridional gradients of surface temperatures are the strongest. Consequently, the storm tracks in CLIM\(_{21}\) are strengthened similarly to CLIM\(_{140}\) at the sea-ice edge (\(\approx 45^\circ\)N). This reduces the Icelandic depression located northward of the edge and causes an increase in SLP by approximately 8 mb (Fig. 10b). Large sea-ice cover almost stops evaporation. This results in decreased precipitation by \(\approx 60\%\) over the North Atlantic (Fig. 10c, e). As for CLIM\(_{140}\), evaporation and precipitation are increased at the sea-ice edge in CLIM\(_{21}\) and the storm tracks activity is intensified and shifted southward (Fig. 6c, d). Contrary to the Late Saalian, the LGM Siberian snow height is almost similar in the CLIM\(_{21}\) and P03\(_{21}\) simulations (Fig. 10f).

The higher SMB values along the southern Eurasian ice sheet margin in CLIM\(_{21}\) result from cooler temperatures spreading over Eurasia (Fig. 9b). As for the Late Saalian, the ice sheet margin near the Atlantic shows more negative SMB in CLIM\(_{21}\) than in P03\(_{21}\) due to the CLIMAP sea-ice cooling effect. However, this negative SMB anomaly extends farther into the ice sheet during the LGM. This contributes to strongly reduce evaporation from the ice sheet. The LGM mean annual SMB in CLIM\(_{21}\) (248 kg m\(^{-2}\) year\(^{-1}\)) is significantly higher than that of P03\(_{21}\) (208 kg m\(^{-2}\) year\(^{-1}\), Table 5).

3.2 The Late Saalian surface ocean

In the previous section, we have shown that the Late Saalian Eurasian ice sheet is less sensitive to changes in SST than its LGM counterpart. The much higher Late Saalian ice sheet confines most of the circulation anomalies caused by the variation in the sea-ice extent into the North Atlantic and the Arctic Ocean while they spread over Eurasia during the LGM. This reduces the Late Saalian Eurasian ice sheet sensitivity to the changes in SST. But how will the Eurasian Late Saalian ice sheet react in
AGCM experiments if we infer simulated sea surface temperatures in equilibrium with the more extreme Late Saalian boundary conditions? Will the simulated Late Saalian SST have the same impact on the Northern Hemisphere climate as when using the LGM SST CLIMAP (1984) and Paul and Schaefer-Neth (2003)?

Fig. 7 Late Saalian climate anomalies between CLIM140 and P03140 over the Northern Hemisphere (30°–90°N). Differences are calculated as (CLIM140–P03140) for: a geopotential height at 500 hPa (m), b sea-level pressure (SLP in mb), c air temperature at 2 m (t2m in °C); d, e precipitation and evaporation anomalies expressed as the ratio (CLIM140/P03140) and f snow height (kg m$^{-2}$). Refer to Table 4 for details about the settings of the simulations.
In the following section, we present the results from the numerical experiment carried out using PLASIM (AGCM + ocean mixed layer). Comparison between LGM and Late Saalian sea surface temperatures will be performed using PS21, which correspond to the simulated LGM SST control run (Table 3). The experimental setup has been previously described in Sect. 2.2.

3.2.1 Simulated sea surface conditions

The Late Saalian SST derived from the PS140 PLASIM simulation are generally in good agreement with the mean annual temperatures estimated from the marine sediment cores reported in Table 2. A comparison between the sediment core data and the simulated summer SSTs is displayed in Fig. 11. The methods used to estimate SST from marine sediment records commonly hold a bias towards summer temperatures (Goni et al. 2001). For this reason we show the comparison for summer. In the tropical and equatorial Atlantic, PS140 matches Late Saalian data well. In our study, the North Atlantic and North Pacific constitute key areas due to their impacts on the Eurasian ice sheets. In the North Atlantic, our reconstruction is colder than suggested by the paleo-temperatures from marine sediment cores (Fig. 11). The sea-ice edge reaches as far South as to northern Spain and northwestern Portugal during winter (Fig. 5a) while it retreats a bit northward during summer, opening the ocean (Fig. 5b). On the contrary in the North Pacific, the only evidence displayed in Fig. 11 is in good agreement with the simulated SST. It should be noted that the very few core data available with SST estimates for MIS 6 in northern high to mid latitudes are not enough to confirm the temperature/sea-ice trend in the North Pacific, where the sea-ice edge reaches Northern Canada during both winter and summer (Fig. 5a) while it retreats a bit northward during summer, opening the ocean (Fig. 5b). On the contrary in the North Pacific, the only evidence displayed in Fig. 11 is in good agreement with the simulated SST. It should be noted that the very few core data available with SST estimates for MIS 6 in northern high to mid latitudes are not enough to confirm the temperature/sea-ice trend in the North Pacific, where the sea-ice edge reaches Northern Canada during both winter and summer (Fig. 5a) while it retreats a bit northward during summer, opening the ocean (Fig. 5b). On the contrary in the North Pacific, the only evidence displayed in Fig. 11 is in good agreement with the simulated SST.

Fig. 8 Northern Hemisphere variance of band pass filtered (3:7) winter sea-level pressure (mb) following Kageyama et al. (1999) for the Late Saalian (a, b) and the LGM (c, d) for simulations forced with CLIMAP and P03 SSTs. Black thick lines stand for the ice sheet contours of both periods.
Late Saalian glaciation than during the LGM (Crowley 1981). In summary, the sparse data available point to that PLASIM may underestimate sea surface temperatures in the North Atlantic and there might have been slightly more open conditions over the Northern Hemisphere during winter than simulated. For the North Pacific, the authors are not aware of any references which can confirm the large sea-ice extent simulated in our reconstruction. The mismatch observed between the marine evidence summarized in Table 2 and the simulated PS140 SST may be due to the limitation of the ocean mixed layer dynamics. In PLASIM, the ocean heat transport is limited to the oceanic surface layer, which is set to a depth of 50 m (this parameter is further discussed below), and SSTs are equilibrated only with the input from atmospheric fluxes (see Sect. 2.2). Without any deep ocean dynamics, the mixed layer does not redistribute efficiently the excess and/or lack of oceanic heat over the latitudes and may strengthen the bias between the marine evidence and our simulated reconstructions.

The difference between PS21 and PS140 presents a pronounced asymmetric sea surface temperature distribution between the Northern and the Southern Hemispheres during Fig. 9 Surface mass balance (SMB) of the Eurasian ice sheet reference simulations P03_{21} (a), P03_{140} (c) and SMB anomalies between the reference simulations and CLIM_{21} (b) and CLIM_{140} (d); e same as for c but for simulation FULL_{140} (with simulated Late Saalian SSTs) and f corresponds SMB anomalies between FULL_{140} and the reference Late Saalian simulation P03_{140}. Values indicated on the figures correspond to the mean annual surface mass balance of each simulations summarized in Table 5.
both winter (DJF) and summer (JJA) (Fig. 12c, d). The equilibrium state of PS140 is globally warmer than PS21 by about 4°C in the Southern Hemisphere and cooler by about 4°C in the Northern Hemisphere. A zonal band of 8°C anomaly appears in the North Atlantic and in the North Pacific at ≈40°N. This anomaly corresponds to the difference between the sea-ice cover southward extent in PS140 and PS21 (Fig. 5). In PS21, the North Atlantic is open during both winter and summer while this is not the case in PS140, where the North Atlantic

Fig. 10 Same as for Fig. 7 but for the LGM period: CLIM21−P0321 (see Table 4)
remains partly covered even during summer (Fig. 5). In PS140, the sea-ice limit reaches the North of Spain ($40^\circ$N), which is the southernmost sea-ice cover of all the reconstructions we use in this study (CLIMAP sea-ice cover stops in Brittany during winter and off Ireland during summer). In the North Pacific, the negative sea surface temperature anomaly is caused by the sea-ice cover that extends more southward in PS140 than in PS21 (sea-ice is confined to the Bering Sea). Due to the similar sea-ice extent during both winter and summer in PS140, seasonality is strongly reduced compared to PS21.
To understand the origin of the pronounced inter-hemispheric asymmetry observed in Fig. 12, we test whether it results from the exceptional Late Saalian Eurasian ice sheet topography, which largely affects the atmospheric circulation in the Northern Hemisphere, rather than from the extreme orbital forcing amplifying the precession effect during this period. We thus performed two more PLASIM simulations, using the ocean mixed layer to equilibrate SST (see Sect. 2.2). The outcomes are compared to PS21 in the same manner as for PS140 in Fig. 12c and d. In the first simulation PS140_orb21, we isolated the topographic effect on SST using the 140 ka boundary conditions and the Late Saalian Eurasian ice sheet and setting the orbital configuration to the 21 ka (Table 3). In the second simulation PS140_t21, we isolated the effect of the orbital parameters using the 140 ka boundary conditions and orbital configuration together with the full LGM ICE-5G ice topography, even over Eurasia. Comparison between PS140_orb21 and PS21 (Fig. 13a) shows a large mean annual inter-hemispheric SST asymmetry: colder Northern Hemisphere SST (about $-6^\circ$C) and warmer Southern Hemisphere SST (about $+4^\circ$C) when using the huge Late Saalian Eurasian topography. On the contrary, comparison between PS140_t21 and PS21 (Fig. 13b) shows that the orbital parameters do not have any significant effect on mean annual SST. If we now compare these results with the inter-hemispheric SST asymmetry visible between PS140 and PS21 in Fig. 12, the amplitude of this asymmetry is similar to that observed between PS140_orb21 and PS21. We can therefore conclude that this SST asymmetry only results from the interactions between the large and high Late Saalian Eurasian ice sheet and the atmospheric circulation. The extreme Late Saalian orbital forcing, compared to the LGM, does not affect the simulated sea surface temperatures. This has not been previously addressed in the literature and will be the subject of future investigations.

Fig. 13 a Difference between the simulated SST PS21 (LGM SST control run) and simulated 140 ka SST PS140_orb21 accounting for the LGM orbital parameters (see Table 3). b Difference between PS21 and simulated 140 ka SST PS140_t21 accounting for the LGM ice sheets topography (see Table 3). c, d simulated 140 ka sea-ice cover for the PS140 (green), the PS140_t21 (red) and for the PS140_orb21 (blue)
3.2.2 Impact of the simulated SST on the Late Saalian climate

In this section, we evaluate the impact of the simulated Late Saalian SSTs on the Northern Hemisphere climate and on the SMB of the Eurasian ice sheet. We forced the LMDZ4 with the simulated SSTs reconstruction and obtained a 21-year Late Saalian climate (FULL-140, see Table 4).

The cooling of the ocean surface causes an annual drop in air temperature of about \( \approx 10^\circ C \) over the entire hemisphere. This large decrease in temperature induces mean annual anticyclonic conditions over the Northern Hemisphere (Fig. 14b, c).

In both North Atlantic and North Pacific, the large southward sea-ice extent (Fig. 5) causes a negative temperature anomaly of \( \approx 10 \) to \( 20^\circ C \) (Fig. 14c). This increases sea-level pressure by \( \approx 8 \) mb over the North Atlantic and the North Pacific where sea-ice is more extended. This anticyclonic activity contributes to a large reduction of precipitation of about 60% and evaporation over this area by more than 90% (Fig. 14d, f). The large sea-ice extent also globally affects the 500 hPa geopotential height over the Northern Hemisphere (\( \approx -60 \) m anomaly) and large negative anomalies of \( \approx 120 \) m are located over the North Atlantic and the Bering Sea. These circulation changes intensify the storm tracks activity whose North Atlantic center is almost similar to that in CLIM140 but larger and more intense than in P03140 (Figs. 8a, b, 14e).

In FULL-140, a permanent snow cover develops over Siberia and is more extended than in P03140 (Fig. 15). However, in FULL-140, moisture fluxes over Siberia normally coming from the North Pacific are reduced compared to those of P03140 (not shown) due the more extended sea-ice cover in the North Pacific in PS140 (Fig. 5). Therefore, the larger snow cover in FULL-140 only results from the global cooling generated by the simulated SSTs which reduces the melting of the snow during spring and summer months. This can directly contribute to the eastward development of the Late Saalian Eurasian ice sheet while during the LGM no permanent snow cover could develop due to a larger amount of dust deposition on snow (Kriinner et al. 2006). Blocking the moisture fluxes coming from the Atlantic as well as a cold and dry climate (Siegert and Marsiat 2001) also favor the stability of this Eurasian ice sheet.

Because of the global cooling caused by cooler SSTs and greater sea-ice extent, there is no more ablation occurring over the ice sheet (Fig. 9e). The SMB is more positive over the western part of the ice sheet since the global cooling reduces the ablation caused by the heat fluxes coming from the East Atlantic (Fig. 9f). On the contrary, due to the strong negative temperature anomaly dramatically reducing precipitation over Siberia, the eastern part of the ice sheet exhibits a far more negative SMB compared to P03140. Similarly, the Atlantic margin shows a more negative SMB value because of the large reduction of the moisture fluxes which cannot provide this area with precipitation as in P03140. Mean annual surface mass balance is approximately 204 kg m\(^{-2}\) year\(^{-1}\) (Table 5). Compared to P03140, the evaporation has been reduced by 10 kg m\(^{-2}\) year\(^{-1}\) and snow fall by 52 kg m\(^{-2}\) year\(^{-1}\). In the case of FULL-140, the lack of accumulation prevails on the absence of ablation and decreases SMB.

4 Discussion

What causes the Eurasian ice sheet to be more than 75% larger and twice as thick during the Late Saalian glaciation compared to the LGM is not known. In this work we investigate whether or not the ocean surface conditions, more precisely SSTs, could have influenced the Eurasian ice sheets extent and volume. The few available marine sediment cores where SST has been estimated for both the LGM and the Late Saalian time periods suggest quite similar temperatures (Table 2). But these few data points, widely spread out in the vast World Ocean, do not allow us to conclude that the Late Saalian and the LGM SST were globally similar and do not rule out that SST could have played a critical role in explaining the difference in Eurasian ice volume between the LGM and the Late Saalian. Therefore, we specifically addressed, using the PLASIM general circulation model of intermediate complexity, whether the global sea surface conditions may have been different during the Late Saalian compared to the LGM and if the oceanic forcings could explain the difference in ice topography. We ultimately evaluated the sensitivity of the surface mass balance to the various prescribed and simulated sea surface conditions.

Three sets of experiments have been performed:

- Prescribing LGM Paul and Schaefer-Neth (2003) SST for both glacial periods,
- Prescribing LGM CLIMAP (1984) SST for both glacial periods,
- Prescribing the simulated 140 ka (peak of the Late Saalian) SST to simulate the Late Saalian climate.

4.1 CLIMAP versus Paul and Schaefer-Neth (2003)

Changing sea surface conditions towards colder SST, i.e. from P03 to CLIMAP LGM reconstructions, causes large negative geopotential anomalies during both Late Saalian.
Fig. 14 Late Saalian mean annual climate anomalies between FULL140 and P03140 over the Northern Hemisphere (30°–90°N). Difference are calculated as (FULL140–P03140) for: a geopotential height at 500 hPa (m), b sea-level pressure (SLP in mb), c air temperature at 2 meters (t2m in °C); d, f precipitation and evaporation anomalies expressed as the ratio (FULL140/P03140) and f winter storm tracks activity (refer to Table 4 for details about the settings of the simulations).
and LGM over the North Atlantic due to a larger southward sea-ice extent. Compared to P03, the CLIMAP SST generates a global cooling leading to reduced ablation along the southern margins of the Eurasian ice sheet and thus to a more positive SMB. However, the amplitude and the intensity of the climate anomalies over the North Atlantic and Eurasia are strongly dependent on the Eurasian ice sheet topography. In all the Late Saalian simulations, its high orography systematically blocks climatic impacts due to changed SST and sea-ice extent over the North Atlantic and Arctic Ocean and prevents them from spreading over Eurasia. This strengthens the contrast between ocean and continent, and also intensifies the North Atlantic storm tracks during the Late Saalian time period (Fig. 8). During the LGM, the lower ice topography allows for an open circulation over Eurasia and over the ice sheet itself.

For both time periods, the Atlantic and the Arctic margins are very sensitive to any changes in SST or sea-ice extent (Fig. 5). However, since the LGM ice sheet is lower and less extended over Eurasia than during the Late Saalian, its SMB is more sensitive to the climatic conditions along the North Atlantic and Arctic margins. On the contrary, since the Late Saalian ice sheet is higher and larger, although responding similarly to the LGM ice sheet along the oceanic margins, its SMB is less sensitive because the higher elevation and the larger extent compensate for temperature fluctuations and consequently reduce the effect on ablation and sublimation (Table 5).

The differences between the mean annual SMB values are statistically significant for both glacial periods as shown by the Student t test at 5% (Table 5). Note that for the Late Saalian simulations, total precipitation and snowfall over the ice sheet are of the same order and differences are mainly caused by ablation and evaporation variations. This is not the case for the LGM simulations which show significant differences between all the components. Consequently, we can conclude that a seasonally open North Atlantic induces more heat advection toward the continent and this directly affects the ablation of the ice sheet. Similarly to Dong and Valdes (1998), we show that cooler SST do not impact significantly on precipitation but only on ablation and evaporation when using the CLIMAP and Paul and Schaefer-Neth (2003) LGM SST, regardless of the ice topography (Table 5, solid stars).

4.2 The Late Saalian surface ocean

4.2.1 Simulated oceanic conditions

In Sect. 3.2.1 we show that the Late Saalian simulated SST are colder than the temperatures derived from the marine records (Table 2) in the high latitudes of both hemispheres. This suggests that slightly more open water conditions in the North Atlantic and in the North Pacific may have prevailed in the Northern Hemisphere than those simulated in this work. The experiment using ocean heat fluxes calculated with the prescribed CLIMAP SST to compute the Late Saalian surface ocean leads to similar sea-ice extent and sea surface temperatures as when using Paul and Schaefer-Neth (2003) derived ocean heat fluxes.

Differences between P03 and the simulated Late Saalian SST PS140 result in a pronounced inter-hemispheric temperature asymmetry (Fig. 12c, d). In our reconstruction, the Northern Hemisphere surface ocean is approximately 4°C colder compared to Paul and Schaefer-Neth (2003) while the Southern Hemisphere is warmer by about 4°C. At a first glance it may seem like this asymmetry could be associated with a “seesaw” effect previously documented in literature (Broecker 1998; Stocker 1998; Ganopolski and...
Rahmstorf 2001; Clark et al. 2002) resulting from thermodaline variations in the Atlantic Ocean in turn causing heat release from one of the two hemispheres out of phase with the other. However, Planet Simulator does not simulate deep circulation and consequently, this inter-hemispheric asymmetry cannot result from such a “seesaw” effect. Therefore, we performed a sensitivity test on the impact from ice sheet topography on sea surface temperatures. This test reveals that the Late Saalian ice topography and the resulting larger albedo are mainly responsible for the observed asymmetry both in terms of spatial temperature distribution and amplitude (Fig. 13). However, this effect has not been documented in the literature so far and will be the object of future detailed investigations.

In our case, this topographic effect is strengthened by the extreme orbital configuration of the Late Saalian through the snow-albedo feedback. The Late Saalian orbital configuration is characterized by a large eccentricity, a perihelion occurring at early December (referring the vernal equinox to March 21st) implying summers near the aphelion. In this typical glacial orbital configuration, the effect of precession is strengthened by the large eccentricity which causes relatively long but cold springs reducing the melting of snow accumulated during winter. This is not the case during the LGM when the orbital configuration was almost similar to that of present-day. This leads to a reduction in the incoming solar radiation at the top of the atmosphere in the [30°N–70°N] latitudes band of 20 W m⁻² during the Late Saalian spring and early summer compared to the LGM and to an increase of 20 W m⁻² during fall (Fig. 2). In conclusion, the cooling effect resulting from simulated PS140 SSTs allows for larger snow accumulation along the ice sheet margins and especially in Siberia. Since the orbital configuration causes cold springs and summers, this snow cover becomes permanent, contributing to the large cooling of the Northern Hemisphere temperatures and helping for a larger stability of the Late Saalian Eurasian ice sheet.

However, is it possible to develop such a large and southward extending sea-ice cover in both North Atlantic and North Pacific as our simulated Late Saalian SST suggest (Fig. 5)? The effect of the ice sheet topography allows sea-ice to extend equatorward while the southmost latitude, here 40°N, reached by the sea-ice cover in both North Atlantic and North Pacific seems however limited by the 140 ka obliquity. When we set the orbital parameters to LGM and use the Late Saalian ice topography (PS140_orb21, see Table 3), the simulated sea-ice extent is almost identical to the PS140 reconstruction during both winter and summer (Fig. 13d). When we keep the 140 ka orbital forcings and use the LGM ice sheet topography (PS140_t21) the sea-ice cover does not extend as far to South as in PS140 and closely matches the LGM reconstruction P03 (Figs. 5, 13). In conclusion, the large southward Late Saalian sea-ice extent simulated in PS140 is governed by the exceptional Eurasian ice sheet topography and the larger resulting albedo in both North Atlantic and North Pacific.

If we compare the climatic implications from the large PS140 sea-ice cover with the implications using the smaller sea-ice cover from CLIMAP during the Late Saalian, the role of the North Pacific is particularly important. Indeed, the PS140 sea-ice cover is extensive over this region while this is not the case in CLIMAP. On the contrary, the sea-ice extents are broadly similar between the two reconstructions in the North Atlantic (Fig. 5). Consequently, in FULL[140] the simulated surface ocean causes a cooling of 10°C over the entire Northern Hemisphere. This is probably due to the presence of sea-ice in the North Pacific extending further to South from the Bering Sea in PS140.

In the PS140, the mixed layer depth was set to the default value of 50 m. However, during glacial periods, mixed layer could have been deeper in some areas (Shin et al. 2003). We thus performed a Late Saalian experiment, similar to PS140 but setting the mixed layer depth to 150 m. Results (not presented) show that increasing the mixed layer depth strengthens the inter-hemispheric asymmetry observed when comparing the PS21 (LGM simulated SST) by 2°C in the Northern Hemisphere and by 4°C in the Southern Hemisphere. Since the ice sheet topography effect strengthens the SST cooling, with a larger mixed layer, the cooling will be more important. This may be due to the fact that in that case, the mixed layer is deeper and consequently equilibrates more slowly with atmospheric heat fluxes due a larger thermal inertia. However in the North Atlantic, there are few differences compared to PS140 and the sea-ice cover obtained with a deeper mixed layer, extends as much southward as in PS140 during winter but retreats much northward opening almost all of the North Atlantic and the Norwegian Sea contrary to PS140 in which this area remains partially covered. In the North Pacific, the sea-ice extent is similar to the extent in PS140 and shows a reduced seasonality. Around Antarctica, the SSTs are about 6°C warmer than in PS140. This makes the sea-ice cover to retreat almost up to the present-day edge. This reconstruction is not realistic in this area since Bianchi and Gersonde (2002) showed that during MIS 6 the sea-ice cover reached 53°S during austral winters and summers. In the North Atlantic, on the contrary, the thicker mixed layer leads to a better data agreement. From this we conclude that the North Atlantic is more sensitive to the mixed layer depth leading to simulated SSTs more in agreement with the estimated SSTs derived from sediment cores. However, our exercise clearly shows that a full ocean dynamics is needed to
realistically equilibrate the SSTs in the various oceanic basins.

There are some caveats that should be considered when evaluating our study on the sensitivity of the Late Saalian Eurasian ice sheet to changes in SST. First, we test the impacts from two SST reconstructions that were published prior to the undertaking Multiproxy approach for the Reconstruction of the Global Ocean surface MARGO (Kucera et al. 2005a). Thus, the evaluated SST datasets may hold biases compared to the more recent MARGO reconstructions leading to a different regional climate interpretation (Kageyama et al. 1999, 2006). However, Waelbroeck et al. (2009) suggest that the Paul and Schaefer-Neth (2003) LGM SST reconstruction is the best available LGM reconstruction to date for the LGM North Atlantic.

In our experiments, we assume an eustatic sea-level of about 110 m below present-day sea-level. The Laurentide ice sheet contributes ≈84 m Equivalent Sea-Level (ESL), Antarctica contributes ≈20 m ESL and our Late Saalian Eurasian ice sheet contributes ≈60 m ESL (comparable to the reconstruction of Lambeck et al. 2006). This leads to an ice volume in excess of ≈54 m ESL relatively to the total eustatic sea-level. Estimating the variation of the Antarctic ice volume since the LGM is still not easy and values range in the various models from 28 m ESL (Lambeck et al. 2003) to 7 m ESL (Ivins and James 2005). Reducing the Antarctic ice volume from 28 m ESL to 7 m ESL (21 m ESL) cannot totally compensate for our excess of 54 m ESL. As a consequence, and in any case, the Laurentide ice volume should be reduced as well to equilibrate the total eustatic sea-level. This suggests that the Late Saalian Laurentide ice sheet could have been smaller than during the LGM. This may have some consequences on the atmospheric circulation (Kageyama and Valdes 2000), on the distribution of the geopotential anomalies even over Eurasia, and on the sea-ice extent in the North Pacific.

Finally, a word of caution regarding the Late Saalian Eurasian ice sheet reconstruction used in our simulations. We have shown in this study that the huge Eurasian ice sheet, extending far into Eastern Siberia with a high surface elevation, significantly modifies circulation, cools temperatures and reduces precipitation. This Eurasian Late Saalian ice configuration (Svendsen et al. 2004) represents the maximum MIS 6 geographic extent which may not have been reached everywhere at the same time during MIS 6. In other words, there is an uncertainty regarding the Late Saalian ice sheet reconstruction and the ice sheet may have been smaller at its glacial maximum and, if so, the climatic impact may also have been different. Nevertheless, geological evidence suggests that the Late Saalian ice sheet reached a much larger extent than the LGM and the 90 ka glaciation (Svendsen et al. 2004) and marine records from the central Arctic Ocean also suggest a more extensive glaciation during MIS 6 than during MIS 2 (Polyak et al. 2001; Jakobsson et al. 2005, 2008).

5 Conclusions

In this work, we test the sensitivity of the Eurasian ice sheet to different prescribed LGM and Late Saalian sea surface conditions to determine how the oceanic forcings contribute to the huge Late Saalian Eurasian ice sheet extent and volume. We show that the choice of the prescribed SST can impact significantly and differently on the surface mass balance depending on the ice volume and extent.

- The Late Saalian Eurasian ice sheet is less sensitive to oceanic changes since its huge topography leads to compensation of regional climatic changes more than the LGM Eurasian ice sheet. We also show that SST and sea-ice impacts on regional climate are different between the Late Saalian and the LGM due to the differences in ice sheet extent and elevation.
- The surface mass balance of the Late Saalian Eurasian ice sheet is affected and using sea-ice opened oceanic conditions may increase its integrated mean annual value.
- For the first time, we simulate the Late Saalian SST. These SSTs are mostly in agreement with the marine core sediment evidence except in the North Atlantic and Pacific where sea-ice cover extends too much southward than suggested by the data.
- Compared to the LGM SST reconstructions, the simulated Late Saalian SST present a large inter-hemispheric asymmetry caused by the huge Late Saalian ice sheet topography and its albedo which affects a larger surface than during the LGM. This leads to cooler (warmer) SST in the Northern (Southern) Hemisphere than during the LGM.
- The large simulated Late Saalian sea-ice extent, reducing the oceanic moisture fluxes, combined to the Northern Hemisphere cooling causes a reduction of the surface mass balance of the Eurasian ice sheet entirely due to the reduction of accumulation.

This may explain how this ice sheet could have remained so large and stable over Eurasia during the end of MIS 6. But how did it become so large? Part of the answer may reside in the Late Saalian orbital configuration. During the Late Saalian, the incoming solar radiation during spring is 20 W m⁻² lower than during the LGM and the Late Saalian springs are longer than during the LGM. This may first have allowed a permanent snow cover to accumulate
more southward than during the LGM inducing cooler atmospheric temperatures over the Northern Hemisphere. This cold atmosphere may then have caused cooler oceanic temperatures, with a reduced seasonality, allowing sea-ice to spread equatorward. This strengthened the effect of albedo and cooled atmospheric temperatures favoring the ice sheet growth and stability over Eurasia. Finally, as we have shown in this work, the interaction between the ice sheet albedo and topography and the sea-ice cover may then have strengthened the positive feedback allowing both the sea-ice cover and the Eurasian ice sheet for reaching such unusual southward latitudes. In the case of the Late Saalian, the interaction between ocean and atmosphere might be particularly important. That is why coupled atmosphere-ocean simulations are needed to proceed with the study of this period and understand better its extreme climate.

A word of caution should be addressed about the choice of boundary conditions and the fact that the LGM climate response cannot be directly applied to other ice ages, specially when ice distribution over continents is very different. One perspective is to simulate the Late Saalian surface oceanic temperatures from a coupled climate model simulation at equilibrium with the climate of the period. In this way the SST will depend not only on air-sea fluxes but also on oceanic surface and deep circulation. One other perspective is to model surface ocean conditions for the entire Late Saalian periods (160–140 ka) which includes one important insolation peak toward 150 ka to understand the evolution of climate prior to the last glacial maximum of the Late Saalian.

Acknowledgments We would like to thank Myriam Khodri, Antje Voelker, Catherine Ritz and Thomas Crowley for their useful contributions. The authors acknowledge support by the Agence Nationale de la Recherche (project IDEGLACE), the Région Rhône Alpes (programme Explora’Doc) and the Ministère des Affaires Etrangères Français and The Bert Bolin Centre for Climate Research (Stockholm University) for their support. The climate simulations were carried out at IDRIS/CNRS and on the Mirage scientific computing platform in Grenoble (France).

References


Crowley T (1981) Temperature and circulation changes in the eastern North Atlantic during the last 150,000 years: Evidence from the planktonic foraminiferal record. Mar Micropaleontol 6(2):97–129


Delmonte B et al. (2004) EPICA Dome C Ice Cores Insoluble Dust Data


Springer
F. Colleoni et al.: The sensitivity of the Late Saalian and LGM Eurasian ice sheets


