Through a glacial cycle:
simulation of the Eurasian ice sheet
dynamics during the last glaciation
Pirjo-Leena Forsström

Through a glacial cycle: simulation of the Eurasian ice sheet dynamics during the last glaciation
The Eurasian Weichselian glaciation was studied with the SICOPOLIS (Simulation Code for POLythermal Ice Sheets) ice-sheet model, which has been validated for Greenland and Antarctica in the European Ice Sheet Modelling INiTiative (EISMINT) program. Palaeoclimate Modelling Intercomparison Project's (PMIP) UK Meteorological Office Unified Model (UKMO) climate anomaly data was utilized for forcing the simulation runs. A set of simulations were completed, including runs in cold-ice mode, using different positive-degree-day (PDD) factors and modified climatic data-sets. The model set-up with present day climatology modified by a glacial index brings forth an areally correct Last Glacial Maximum (LGM) extent in the western areas, but the ice-sheet volume was too small compared to reconstructions from rebound rates. Applying modified climate data results in a similar extent as indicated by the Quaternary Environments of the Eurasian North (QUEEN) Late Weichselian ice-sheet reconstruction.

The forming ice sheets are very sensitive and respond quickly to changes in temperature and precipitation. This indicates a great sensitivity to the climatic signal, especially in the Fennoscandian sector of the ice sheet. The simulation results suggest areal and temporal variations in fast-flow areas. The fast flow areas that form are located in topographic troughs. Some of these maintain a stationary type of flow, indicating that the corresponding ice stream was of an isbræ-type (the Norwegian Channel Ice Stream). The lobate structure in the Baltic area developed as a result of mass balance flow. In the simulation results, the LGM ice sheet grew to a maximum extent in the Baltic sector without any fast flow events in the area. The fast flow at the onset areas of the Baltic Ice Stream were short events with moderate velocities, developing at the deglaciation phase. No wet terrestrial southern margin on the ice sheet was detected in the simulation results.

The peaks in freshwater discharge coincided in all the runs and the volume of the ice sheet and the climatic conditions determined the freshwater flux. Freshwater peaks can be correlated to the observed occurrence of Heinrich Events (HE) reflected in ice-rafted debris in deep sea sediment, and to some Dansgaard-Oeschger (D-O) oscillations seen in ice core record. Other connections to D-O oscillations can be drawn from the ice volume flux peaks in the western marine margin of the ice sheet. High volume fluxes preceded the freshwater peaks.
The results pinpoint that the Eurasian ice sheet was clearly active in the periods of HEs and D-Os, especially on the anomalous Heinrich Events H3 and H6. The occurrence of ice stream activity seemed to be shifting from south to north with ongoing time.

**Keywords**: Eurasian Ice Sheet, Weichselian glaciation, ice sheet model, ice streams, fresh water flux, basal temperatures.

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To my daughter Amanda


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<td>AGCM</td>
<td>Atmospheric General Circulation Model</td>
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<td>BIS</td>
<td>British Island Ice Sheet</td>
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<td>BSIS</td>
<td>Barents Sea Ice Sheet</td>
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<td>CMIP</td>
<td>Climate Modelling Intercomparision Project</td>
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<td>D-O</td>
<td>Dansgaard-Oeschger, oscillations in ice core records</td>
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<td>EISMINT</td>
<td>European Ice Sheet Modelling INiTiative</td>
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<td>ESM</td>
<td>Earth System Model</td>
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<td>EWG</td>
<td>Early Weichselian Glaciation</td>
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<td>GCM</td>
<td>General Circulation Model</td>
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<td>GIS</td>
<td>Geographical Information System</td>
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<td>GRIP</td>
<td>Greenland Ice core Project</td>
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<td>HE</td>
<td>Heinrich’s Event, seen in deep sea sediments</td>
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<td>ICE4G</td>
<td>Global ice sheet topography for the last 21 000 years</td>
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<td>IOA</td>
<td>Ice-Ocean-Athmosphere, referring to climate systems</td>
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<td>IRD</td>
<td>Ice-Rafted Debris</td>
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<td>KIS</td>
<td>Kara Ice Sheet, Kara Sea part of the Barents-Kara Ice Sheet</td>
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<td>LGM</td>
<td>Last Glacial Maximum</td>
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<td>MWG</td>
<td>Middle Weichselian Glaciation</td>
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<td>mwp-1A</td>
<td>Meltwater Pulse 1A, to Atlantic ocean</td>
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<td>mwp-1B</td>
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<td>NCIS</td>
<td>Norwegian Channel Ice Stream</td>
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<td>OIS3</td>
<td>Marine Oxygen Isotope Stage 3</td>
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<td>Positive Degree-Day Method, for melting calculations</td>
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<td>PMIP</td>
<td>Palaeoclimate Modelling Intercomparision Project</td>
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<td>QUEEN</td>
<td>Quaternary Environments of the Eurasian North</td>
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<td>SIA</td>
<td>Shallow Ice Approximation, method in numerical glaciology</td>
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<td>SIS</td>
<td>Scandinavian Ice Sheet</td>
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<tr>
<td>THC</td>
<td>Thermohaline Circulation in oceans</td>
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<td>UKMO</td>
<td>United Kingdom Meteorological Office climate model</td>
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The Earth’s climate is strongly linked to incoming solar radiation. This is controlled by Earth’s astronomical, orbital parameters (eccentricity, obliquity, precession) and atmospheric back-scattering effects. The orbital parameters are responsible for the Milankovitch cycles (95.8, 41 and 21.7 thousand years) (Milankovitch, 1941; Imbrie and Imbrie, 1979; Berger, 1988). The variations of these parameters do not significantly modify the overall annual solar radiation, but they affect the seasonal distribution (Berger, 1980).

Northern Hemisphere ice sheets have evolved with the periods of these orbital parameters (Hays et al., 1976), and thus they form the basis of climate variability. Interlinkage of processes such as ocean circulation, atmospheric circulation and glaciations leads to a complex system of evolving and changing Earth climate.

Past climate during the last 100,000 years (the Weichselian stage) has been dominated by a succession of stadials (cold phases) and interstadials (warm phases). These events are imprinted in different proxy records around the world. Large climatic variations in the Weichselian stage are concluded from ice core data (Dansgaard et al., 1993) and deep sea sediments (Keigwin and Jones, 1994).

The Pleistocene ice sheets had a profound influence on climate and global sea level by affecting atmospheric and oceanic circulation patterns and planetary albedo through changes in ice sheet extent, height and stability (Hughes, 1987, 1992, 1998; Knight 1999). Circulation effects depend on the height of the ice sheet and on temperature effects on the ice sheet area (Rind, 1987). Studies with climate models suggest several mechanisms for this influence. These include southward displacement of the winter jet stream by a high ice sheet dome, cooling over the dome and downwind of the ice sheet dome, reorganization and strengthening of storm tracks and the generation of anti-cyclones at the ice sheet surface (Manabe and Broccoli, 1985; Pollard and Thompson, 1997; Kutzbach et al., 1998; Ganopolski et al., 1998).

Marine-based ice stream instabilities might be a factor in high frequency climate variations. Particularly, the North Atlantic Heinrichs layers (sediment layers enriched in ice-rafted debris) provide a good explanation for rapid disintegration of ice streams and ice shelves (Alley and MacAyel, 1994; Andrews et al., 1994; Bond and Lotti, 1995; Hulbe et al., 2004). Experiments with dynamic ocean models suggest that the strength of the thermohaline circulation in the North Atlantic ocean is sensitive to the freshwater input (Stocker et al., 1992). This stresses the importance of the oceanic component of the system.

An important feedback factor for ice sheets and climate is the effect of proglacial
ice-dammed lakes. Simulations with an atmospheric circulation model showed that the main influence of these lakes at the southern margin of the Eurasian ice sheet was a significant reduction of melting as the ice lakes created strong regional summer cooling over the area (Krinner et al., 2004). Proglacial lakes may also have had an important role in the destabilization and surging of ice stream margins (Dredge and Cowan, 1989). The deepening ice-contact lakes strengthen ice sheet flotation and draw-down. Ice sheets are thus slow to grow and rapid to melt, slow in filling ice-marginal lakes but rapid in drainage (Clark et al., 1999).

To study these complex climate-related processes, numerical models are invaluable. It is impossible to conduct laboratory experiments to test or repeat the functioning of global climate system. However, models can never completely characterize the system they are designed to describe (Oreskes, 2000). This is due to the openness of the model. This is a function of the relationship between the complexity of the system being modelled and the model itself.

An ideal tool to study the Northern Hemisphere Ice Sheets and hence climate, would be an integrated climate-ocean-ice sheet-land processes model. The land system includes components like hydrology and permafrost. Earth System Models, the coupled models of different Earth System components, are only just evolving. These integrated systems allow more detailed characterisation of Earth’s atmosphere, ocean and continental surfaces and their actions and interactions. These system models have fundamentally insatiable computational requirements, and are still very much under development. Hence, reduced models with sensible parameterisation are widely used (Stocker, 2000). The “complexity paradox” is incorporated in the system: while we are trying to include all relevant processes in a system model, the more difficult it is to test the model (Oreskes, 2000).

In this study, the Eurasian ice sheet dynamics and involvement with the other climate components during the Weichselian will be exploited using a numerical model driven by various climatologies. The model is validated by comparisons to other models, and with calibration to known data. The goal is to reconstruct ice sheet extent, volume, freshwater flux and flow according to the glacial-geologic records. A special focus is given to the fast flowing outlets of the marine margins of the simulated ice sheets, which may play a role in Heinrich Events, and to basal conditions.

**Palaeoclimate**

A correct understanding of ice-sheet extent and flow is linked closely to our understanding of the Eurasian palaeoclimate. To obtain a reliable reconstruction of the ice sheet, we need reliable and wide-ranging palaeoclimate data. Climate-dependent natural phenomena incorporate into their structure a measure of this climate signal (Bradley, 1999). Natural archives, like deep sea sediments and tree rings, have also recorded this signal in the past. This type of data is called proxy data, and it contains a filtered climate signal. Unfortunately, continuous proxy records for climate history reconstructions are lacking. Also, direct proxies for local temperature and precipitation during the glaciation are rare. Where records do exist, they are derived from palaeobotanical evidence (Peyron et al., 1998; Tarasov et al., 1999). Regional records, from Greenland ice cores and deep-sea sediments, are also of use in bridging local observations and driving and testing palaeoclimate models. These records need to be correlated with each other to get the timing of the event studied. When dealing with proxy data, it is important to understand how these records are interlinked by various processes actualising in the cou-
pled ice-ocean-atmosphere (IOA) system (Lowe et al., 2001).

A widely used record in ice sheet modeling is the Greenland GRIP ice core $\delta^{18}O$ record (Dansgaard and others, 1993). Determining the age-depth relationship is one of the most important tasks in ice core dating (Bradley, 1999). The $\delta^{18}O$ record is considered to be valid from 90 ka ago onward (Johnsen et al., 2001). Calibrating the $\delta^{18}O$ record of snowfall and temperature is not straightforward. It is assumed, that past accumulation rates and ice-flow patterns have a coupling to the temperature history (Johnsen et al., 1995; Cuffey and Clow, 1997). A direct reconstruction of past temperatures from the GRIP borehole (Dahl-Jensen et al., 1998) suggests that the LGM temperatures were $-23 \pm 2$ K colder than at present. This is in excellent agreement with the $\delta^{18}O$ record.

Nevertheless, the GRIP event stratigraphy may not always be the most appropriate to force ice sheet simulations (Knight, 2003), as some glacial events (like the late Devensian in Ireland) have a better match to shifts seen in marine records, suggesting that the oceanic conditions are the primary driving force for glacial events in northwest Europe.

The relative sea level has changed temporally and spatially, exhibiting the glacial signal. This signal is the boundary condition in ice volume evolution and timing of past glaciations (Lambeck and Chappel, 2001). The ocean component of the IOA system seems to have had a key role in the ice sheet growth in the northwestern Barents Sea area and in western Norway (Spielhagen, 2001).

To study the Early Weichselian Glaciation, Atmospheric General Circulation Model (AGCM) simulations were conducted to study the climate-ice lake interaction (Krinner et al., 2004). The resulting cooling anomaly from present day summer temperatures was in the order of $-8$ °C in summer months. From these results, Krinner et al. concluded that the summer cooling mechanism inhibited ice sheet melting and thus delayed ice sheet decay.

Later Weichselian (from 60 ka to 25 ka BP) conditions have also been studied, for example, Pollard and Barron (2003) simulated the oxygen-isotope stage 3 (OIS3), based on nesting a high-resolution mesoscale model (RegCM2, Giorgi et al., 1993a,b) for Europe within a general circulation model (GENESIS GCM, Thompson and Pollard, 1997), and found that the North Atlantic sea-surface temperatures and the ice sheet size were the key to a very cold European continent.

Kageyama et al. (2001) have reconstructed the Last Glacial Maximum (LGM) climate using all available proxies and compared the reconstructions to the output climate data-sets from the Palaeoclimate Modelling Intercomparison Project (PMIP). One of the main goals of PMIP was to create palaeoenvironmental data sets for model evaluation. As one of the boundary conditions to simulate the LGM climate, ice sheet extent and height was defined according to Peltier (1994). This data set, called ICE4G, consist of topographical heights for ice sheets at 1 ka intervals since the LGM. Kageyama et al. (2001) noted significant differences in two main regions, southwestern Europe and north of the western Siberia. In the first case, the resulting PMIP temperatures were too warm with too much precipitation, and in the second, the resulting summer temperatures were too cold. Specifically, if these reconstructions are compared to PMIP results from the United Kingdom Meteorological Office Unified Model (UKMO) (Hewitt and Mitchell, 1997), we can see that the winter temperature anomaly in the southeast regions is in agreement with the results obtained by Tarasov et al. (1999). However, the UKMO results showed an anomaly of similar magnitude in the
northeast, whereas Tarasov et al. (1999) stated that the temperatures during the LGM were warmer. The UKMO21 PMIP results indicated an anomaly in precipitation and temperatures, as shown in Fig. 1.

Figure 1. Anomalies for precipitation (left panel, in mm/a) and temperature (right panel, in °C) at the LGM defined by UKMO21 simulation results. Areal coverage is 6040 × 2840 km² in the stereographic plane. Axis units are in km. Reproduced from Forsström and Greve, 2004.
**Eurasian glaciation**

The Weichselian glacial stage is characterized by huge ice sheets in the Northern Hemisphere (Peltier, 1994). Ice sheets that formed in Scandinavia expanded east across the northwest Russian Plains and the White Sea periphery. Ice sheets forming in the Barents and Kara Sea areas spread to south. Many reconstructions have been made of Eurasian ice-sheet extent.

An updated synthesis of the most complete reconstructions of the Eurasian glaciations has been compiled by the QUEEN (Quaternary Environment of the Eurasian North) project (Svendsen et al., 1999; Thiede et al., 2001; Svendsen et al, 2004). This synthesis states that the distribution of glaciers in the Eurasian Arctic has had great fluctuations and the largest ice sheets existed for a relatively short time period. This reconstruction deviates from earlier compilations, especially in the southern Kara Sea and the Taymyr Peninsula area, which was thought to have been ice-free during the Last Glacial Maximum (LGM) around 20 ka ago (Svendsen et al., 1999; Polyak et al., 2000; Mangerud et al., 2002, Svendsen et al., 2004).

In total, three glaciations are referred in the Eurasian area for the Weichselian glacial stage: the Early Weichselian Glaciation EWG (before 80 ka ago, maxima 95–85 ka ago and 85–72 ka ago in Scandinavia), the Middle Weichselian Glaciation MWG at 75–25 ka ago, maximum at 60 – 50 ka ago and the Last Glacial Maximum LGM (25–15 ka ago). For example, the Markhida moraines in the Pechora lowlands are interpreted as a product of the MWG. A pre-Weichselian glaciation (Late Saalian, before 140 ka ago) was thought to have been larger in this area. Evidence also points to two glaciations in southern Fennoscandia and North Sea region (Sejrup et al., 2003).

The QUEEN ice-sheet extents of the Weichselian glaciations are depicted in Fig. 2.

The Early Weichselian Glaciation (EWG) was restricted and covered only high mountainous areas in Scandinavia, including Lapland. The Barents Sea (except the Bear Island Trough area), Kara Sea, Taymyr

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Figure 2. Reconstructed extents for the Weichselian glaciations after QUEEN (Svendsen et al., 1999; Thiede et al., 2001; Svendsen et al, 2004). Dash-dotted line: Early Weichselian Glaciation, long dashed line: Middle Weichselian Glaciation and small dashed line: Last Glacial Maximum.
Peninsula and Putorana Plateau were glaciated, as well as the High Arctic Islands. At this EWG maximum, a major ice dome was located on the continental shelf in the northern Kara Sea.

The Middle Weichselian Glaciation (MWG) covered all of Fennoscandia, extending to the modern Baltic Sea south coast. This MWG was mainly an ice advance from the regrowth of the Barents-Kara Ice Sheet, with ice reaching onto the northern margin of the Eurasian mainland. The Barents Sea was glaciated, as was Kara Sea, at least partially. The glaciation reached the northern rim of the Taymyr Peninsula and the Putorana Mountains had a limited ice cover.

In the Last Glacial Maximum, the Scandinavian Ice Sheet grew to its maximum extent during the Weichselian stage, as the eastern margin was located along a morphologically distinct ice marginal zone in the northwestern Russian Plain and Arkhangelsk region. For the LGM, in the western section, the Scandinavian Ice Sheet (SIS) and the British Isles Ice Sheet (BIIS) developed, up north in the Barents Sea and the Arctic Islands the Barents ice sheet (BSIS), and to the east, the controversial Kara Ice Sheet (KIS). The Barents-Kara Ice Sheet terminated on the seafloor in the southeastern Barents Sea and on the western Kara Sea shelf. On the northern Kara Sea shelf, the ice sheet advanced eastwards to northern parts of Taymyr for a short time. Severnaya Zemlya was probably not glaciated.

The timing of the LGM across Eurasia, and it is a truly diachronous event. From the QUEEN summary (Svendsen et al., 1999; Thiede et al., 2001; Svendsen et al., 2004), the timing of the LGM is given in Table 1.

The maximum position in the west was a short event at 27–25 ka ago, with the decay starting around 22–20 ka ago (Svendsen et al., 1999; Thiede et al., 2001; Svendsen et al., 2004). The eastern SIS grew to a maximum extent at 19–17 ka ago and started then to retreat. The BSIS started to retreat at 15 ka and deglaciation was completed by 10 ka ago. In this paper, LGM is referred as the coldest time of the simulation period.

The geological records discussed here include ice sheet extent (from QUEEN data), topographic evolution (Kakkuri, 1987; Balling, 1980; Lambeck and Purcell, 2003), and ground temperature observations (Balling et al., 1990; Kukkonen and Šafanda, 1996; Kukkonen and Jœleht, 2003; Šafanda et al., 2004). A special attention is given to the areas of Norwegian Channel Ice Stream (NCIS) and Baltic Ice Streams. For palaeogeographical and geological reconstructions on these areas, the works of Sejrup et al. (2000), Sejrup et al. (2003), Houmark-Nielsen (2003) and Houmark-Nielsen and Kjær (2003) are referred. The last is based on synchronised land, sea and glacier configurations.

### Models for climate and climate-related processes

Natural phenomena are not simple to describe. A real phenomena can never be fully specified by a model, as natural systems are always open (Oreskes, 2000). Even as such, models can be complex. Models do not give exact answers but provide insight into the real phenomena. They are

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<th>Denmark and S Baltic</th>
<th>Arkhangelsk Region</th>
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<tr>
<td>appr. 27 ka ago</td>
<td>20 ka ago</td>
<td>17 ka ago</td>
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highly needed for cases when experiments can not be controlled or carried out (Irobi, 2004).

Numerical models describing chosen parts of nature are simplifications of the real phenomena, and their reliability needs to be verified and validated (Babuska et al., 2003). Numerical model verification is the process of determining if the numerical model for the mathematical description of a physical process is accurate enough. The model itself can not be fully verified (Oreskes, 2000). Models can and need to be validated by comparisons with similar models, calibration and confirmation (Irobi, 2004). For the validation problem, the mathematics of the selection is under development (Babuska et al., 2003).

In Earth Sciences, models can be crudely divided into two categories: the reduced models and the quasi-realistic models. The reduced models serve as tools to understand better the underlying processes, and to test hypotheses. However, they often fail in detailed descriptions of the phenomena. The quasi-realistic models enable the simulation of real-world systems and experimentation with them, but they can fail to produce insight and a deeper understanding into the system’s functioning. Often both types of models are needed to gain deeper insight of a phenomena (von Storch and Flöser, 2001).

The most sophisticated integrated models are the Earth System Models, that constitute a family of General Circulation Models (GCM’s) and other earth process models. IOA-system is a typical combined model set. For example, global coupled ocean-atmosphere general circulation models that include dynamic sea ice, simulate the physical climate system given only a small number of external boundary conditions, such as the solar “constant” and atmospheric concentrations of radiatively active gases and aerosols. These kinds of models have been used for decades in theoretical investigations of the mechanisms of climatic changes. Coupled GCMs have also been used to separate natural variability from anthropogenic effects in the climate record of the 20th century and to estimate future anthropogenic climate changes, including global warming (Houghton et al., 1997). Several coupled GCMs have been developed by different research groups and verified by an international Climate Modelling Intercomparison Project (CMIP) (Covey et al., 2003). The atmospheric and oceanic components include large-scale dynamics, convection, and (for the atmosphere) detailed treatments of solar and infrared radiation, clouds and precipitation. Processes such as atmospheric chemistry, marine biogeochemical cycles, soil microbial and trace-gas activity, net primary productivity and climate-induced changes in vegetation are included in the various models. The main goal of many ESMs is to effectively and reliably combine these existing models and to simulate future changes in these systems on 50 to 200 year timescales due to greenhouse-gas emissions and other anthropogenic forcing.

**Eurasian ice sheet models and simulations**

Ice sheet models exploit proxy and reconstructed climate data. To reconstruct the glaciations, simulations of Northern Hemisphere ice-sheet retreat using atmospheric general circulation model (AGCM) results have been carried out by Charbit et al. (2002), and by Marshall et al. (1999) for the Laurentide and Greenland ice sheets, as well as by Siegert and Dowdeswell (2004) for the Eurasian ice sheet. Charbit et al. focused on the last deglaciation, with climatologies for 21, 15, 9, 6 and 0 ka BP from AGCM climate snapshots. They are able to reproduce the last deglaciation congruent with the QUEEN reconstructions, and in
a consistent manner. Siegert and Dowdeswell modelled the time-dependent behaviour of the ice sheet, starting from the QUEEN reconstruction. From the resulting ice sheet characteristics, they are able to deduce the LGM climate.


Siegert and Marsiat (2001) used an AGCM for studies of the LGM climate. Siegert et al. (1999) used the QUEEN-project data for ice sheet extent. In their simulation, a zero level of precipitation is set across the Taymyr Peninsula and resulting reconstructions are then compared with geological evidence. From these, they selected the most likely glacial scenario and based their reconstructions on it. The predictions of glaciological output included sediments, icebergs and meltwater.

Arnold and Sharp (2002) forced their model of the Scandinavian ice sheet with eustatic sea level and climatic change, deduced from the GRIP-record. They focused on studying the resulting fast flow areas and basal water conditions. The development of localised fast flow areas, forming lobes, is strongly influenced by the evolving ice sheet topography routing the subglacial water flow.

Boulton et al. (2003) utilize the model of Boulton and Payne (1994) and developed by Hulton and Mineter (2000), and complement it with a new coupling between the ice sheet and lithosphere. The forcing for temperatures over ice sheet was based on GRIP-data, adjusted by using palaeotemperature records. The focus of the study was the streaming flow of the European ice sheet. Their study indicate that the width and location of the ice streams are strongly influenced by topography, the width being in the order of 100 km. In simulation results, muted topography stabilizes the location of ice streams. The results suggest that the ice sheet thickness can be reduced by over 1 km, if the ice sheet is relatively warm and has broad ice streams.

Zweck and Huybrechts (2003) used a simple marine-extent relationship and thus managed to reproduce the first-order processes associated with the changes in the marine margin of the Northern Hemisphere ice sheets since the LGM. They concluded that the simple parameterisation for calving, with water depth as the most crucial variable, served as a physically reasonable tool to model the marine retreat.

Siegert and Dowdeswell (2004) exploited the model discussed by Siegert et al. (1999). Among other things, they found that the ice accumulation rates across western Scandinavia and Barents Sea to be in excess of 300 mm/a. In the eastern margin of the ice sheet and central areas of the Barents Sea, the ice accumulation was very low (less than 200 mm/a). Over the Kara Sea, it was less than 100 mm/a. They also noted that the southern margin of the ice sheet had a strong surface ablation, suggesting relatively warm surface air temperatures.

**Basal conditions**

During cold climatic spells, ice-free areas were subjected to low surface temperatures. During these times, the ground temperatures underneath ice sheets can be notably warmer and more stable than outside the ice sheet (Paterson, 1994). If an ice sheet is thick and has long, stable occupation time, the temperature variations underneath the ice sheet are probably smaller compared to
values outside the ice sheet. This is supported by measurements from the super-deep holes in Siljan, western Sweden, which suggest no significant temperature gradient (Balling et al., 1990). The heat flux density data for Siljan has so far been not reported. This site was situated under the main dome of the Scandinavian ice sheet. The variation is a bit more pronounced in eastern Finland (Kukkonen and Šafanda, 1996) and large for areas in Karelia (Kukkonen et al., 1998).

Conditions at the ice sheet base are controlled by ice sheet pressure, temperature, hydrology and properties of the underlying substrate. The cold based area is stagnant, with no water, and very little movement and deformation at the base. The temperate based area can be of two different types: temperate base with temperate ice above, and temperate base with cold ice above. In the first case, the temperate base maintains favourable conditions for fast flow, and the flow draws down temperate, less viscous ice and the flow is thus maintained. In the second case, fast flow can cease due to draw-down of cold and viscous ice.

Basal temperatures at the base of late Weichselian ice sheets have been inferred by modelling (Marshall and Clark, 2002) and from measurements (Kukkonen and Jœleht, 2003). In the latter case, geothermal heat flow density in the Fennoscandian Shield and east European platform was analysed. Monte Carlo inversion was applied to determine ground surface temperatures during the Weichselian. The results suggest the lowest values for ground surface temperatures about 20 ka BP. After this, an average warming of 8.0 ± 4.5 °C in 10 ka took place.

Recently, very low glacial temperatures were reported on the rim of the Fennoscandian ice sheet in northeast Poland (Šafanda et al, 2004). From analysis of deep borehole temperature profiles, they concluded that a mean ground temperature of –10.5 °C and a more than 500 m thick permafrost layer during the last glaciation are apparent.

**Ice Streams**

An ice stream is defined by Paterson (1994) as “a region of grounded ice sheet in which ice flows much faster than in regions on either side”. This definition is emphasized by Stokes and Clark (2001) in their review article of palaeo ice streams. This definition is not tied to the continuity of the flow. Surge events are not ice streams, but rather dramatic flow accelerations happening at regular cycles.

Ice streams have a key role in the stability and dynamics of contemporary ice sheets, which also applies to the Weichselian (e.g. Denton and Hughes, 1981; Marshall et al., 2000; Dowdeswell and Siegert, 1999; Payne and Baldwin, 1999). Several mechanisms for fast ice stream flow have been observed. Modelling results also indicate that fast flow areas can develop without control of topography and soft sediments in areas of hard bed such as the Baltic Shield (Payne and Baldwin, 1999). In Antarctica, the distribution of deformable till as a proxy for palaeo ice stream configuration indicates that bathymetry was a first-order control on high velocity ice flow (Howat and Domack, 2003). Varying topography is a likely reason for fast flow and ice streams in ice sheets to occur, but it is not a preliminary cause of fast flow (e.g. Kaplan et al., 2001).

Two extreme types of fast flow in ice sheets are ice streams and isbræ, and they are the end members in a continuum of ice stream behaviour (Truffer and Echelmayer, 2003). A typical isbræ has very high driving stresses, has a steep surface slope and flows through a deep bedrock channel. Clark and Stokes (2003) defined this type as a topographic ice stream. An isbræ is a stable feature of the ice sheet and only shuts down when there no longer is a high ice flux from
the ice sheet. In contrast to isbræ, a typical ice stream is wide (approximately 10 km) and the location is not strongly controlled by bed topography. A typical ice stream also has a low surface slope and low driving stress. Stokes and Clark (2003) summarized the basic characteristics of ice streams, in the broadest sense, to be more than 20 km wide and 150 km long, with ice velocities of more than 300 m/a and highly convergent onset zones. Marine ice streams can maintain a high velocity without advancing, but a terrestrial ice stream presumably has to advance by forming lobes at the terminus (Stokes and Clark, 2001). The flow events associated with lobe formation are thought to either include fast flow or involve a uniform but slower flow.

Enormous speed variations in ice streams paced by oceanic tidal oscillations have been observed (Bindschandler et al., 2003). How this behaviour manifests in the long-term evolution of the ice streams still needs to be studied. The main questions in ice stream dynamics are what controls the location of an ice stream and what allows it to move fast and to shut down quickly. Ice sheet flow is controlled at the surface by mass balance and temperature, at the base by friction at basal contact, by lithosphere rheology, thermal conductivity and geothermal heat flux and within the ice sheet by ice rheology, thermal conductivity and friction and water content. The coupling of temperature and flow within the ice sheet is complex because ice flow depends on the ice temperature, and heat is in turn advected by the flow (Boulton et al., 2001).

Several lines of evidence indicate fast ice stream flow in palaeo ice sheets of the Northern Hemisphere. Distinct pulses of ice-rafted debris (IRD) in deep sea sediments have been interpreted to represent rapid discharge events through ice streams (Bond et al., 1992). Material entrained in flowing ice is transported forward to the area of floating ice and it settles to the seafloor when the iceberg has melted. Glacial geologic features also support the existence of ice streams, in both the North American and Eurasian ice sheets. The evidence includes topographic troughs, intense glacial scouring and streamlining of bedrock, erratic dispersal plumes, specific drumlin patterns, mega-scale glacial lineations, sedimentary and tectonic evidence of deformation of glacial sediments, hummocky topography and large till deltas (Clark and Stokes, 2003).

More than fifty palaeo ice streams have been hypothesized, and the identification of these from the geomorphological imprint of activity is not simple. The bedform record may only be related to the final stages of ice stream activity and thus not represent the flow system evolution (Stokes and Clark, 2001).

For example, the Scandinavian ice sheet dynamics is characterized by several major ice streams (Stokes and Clark, 2001; Boulton et al., 2001). The Norwegian Channel Ice Stream (NCIS) has been identified from geological data (e.g. glacially fed fans) (Sejrup et al., 1998, 2000), the Baltic Ice Stream and its sublobes have been discussed (e.g., Kleman et al., 1997) and the Finnish, the Karelian and the Archangel area Ice Streams were studied by e.g. Houmark-Nielsen et al. (2001), Larsen et al. (1999) and Lunkka et al. (2001). In the north, ice streams were located at the Saint Anna Trough, Franz Victoria Trough, Bear Island Trough and Storfjorden Trough (Sejrup et al., 2000; Knies et al., 2001).

The marine ice streams transported ice to the calving area. The fast ice sheet flow associated with large flux events will remove basal ice, and thus basal debris, by frictional melting. This basal debris entrained in calved icebergs is sedimented to the ocean as the icebergs melt. Ice stream location and vigour determines the distri-

IRD record
The marine record of the glacial-age northern ocean is punctuated by ice-rafted debris (IRD) events. Ruddiman (1997) has found that the flux of lithic grains in the North Atlantic is different for glacial and interglacial conditions. The flux is large during glacials and the pattern of IRD deposition form an IRD-belt along the 50° latitude. During glacials, the centre for deposition moves a few degrees south.

The long-term cooling trends (Bond cycles) seen in ice core records, incorporate Dansgaard-Oeschger (D-O) oscillations (Bond and Lotti, 1995; Dansgaard et al., 1982). A Bond cycle is sequence of interstadials and stadials, a gradual cooling followed by rapid warming. The cooling trend termination is supposed to have a relation to Heinrich Event (HE) (Heinrich, 1988). The HEs are associated with a temperature decrease recorded by ice core isotopes and low sea-surface temperatures in North Atlantic sediment cores (Bond et al., 1993).

The timings of HEs from Bond et al. (1993) and Elliot et al. (2002) are represented in Table 2.

Dansgaard-Oeschger temperature oscillations can be grouped as warm-cold oscil-
Figure 4. Ice Rafted Debris (IRD) events as idealised time series, after Alley and Clark, (1999).

Table 2. Average age and duration of the HEs. For HEs 1–4 in ka BP, (Elliot et al., 2002), for HEs 5–10 and all calendar ages (Bond et al., 1992, 1993). For the earlier HEs, the ages are subject to uncertainties of ±5% (McManus et al., 1994). These are marked with * in the table. The base and top values refer to HE start (base) and end (top) in ocean sediment cores from the North Atlantic.

<table>
<thead>
<tr>
<th>HE</th>
<th>Pos.</th>
<th>Age (14C ka)</th>
<th>Cal. ka</th>
</tr>
</thead>
<tbody>
<tr>
<td>H1</td>
<td>Base</td>
<td>15.1</td>
<td>16.8</td>
</tr>
<tr>
<td></td>
<td>Top</td>
<td>13.4</td>
<td></td>
</tr>
<tr>
<td>H2</td>
<td>Base</td>
<td>22.1</td>
<td>24.0</td>
</tr>
<tr>
<td></td>
<td>Top</td>
<td>20.4</td>
<td></td>
</tr>
<tr>
<td>H3</td>
<td>Base</td>
<td>27.7</td>
<td>31.0</td>
</tr>
<tr>
<td></td>
<td>Top</td>
<td>25.1</td>
<td></td>
</tr>
<tr>
<td>H4</td>
<td>Base</td>
<td>34.9</td>
<td>38.0</td>
</tr>
<tr>
<td></td>
<td>Top</td>
<td>33.9</td>
<td></td>
</tr>
<tr>
<td>H5</td>
<td></td>
<td>52.0*</td>
<td>45.0</td>
</tr>
<tr>
<td>H6</td>
<td></td>
<td>69.0*</td>
<td>60.0*</td>
</tr>
<tr>
<td>H7</td>
<td></td>
<td>71.0*</td>
<td></td>
</tr>
<tr>
<td>H8</td>
<td></td>
<td>76.0*</td>
<td></td>
</tr>
<tr>
<td>H9</td>
<td></td>
<td>85.0*</td>
<td></td>
</tr>
<tr>
<td>H10</td>
<td></td>
<td>105.0*</td>
<td></td>
</tr>
</tbody>
</table>

lations, with increasingly cold minima, until a very warm maximum. After this culmination (HE), the oscillation cycle begins again (Heinrich, 1988; Dansgaard et al., 1982, 1993; Clark et al., 1999; Alley and Clark, 1999; Stocker, 2000). The D-Os occur more frequently than the HEs and are centered in the northern Atlantic.

As the HEs and D-O oscillations are reflected as detrital events in sediment cores with varied petrology, they are associated with changing iceberg source area (Bond and Lotti, 1995) and different palaeo-ice streams. The question of lead or lag of different ice streams around the North Atlantic has been argued (e.g. Bond and Lotti, 1995; Dowdeswell et al., 1999; Grousset et al., 2000). Specifically, Grousset et al. raised the issue of the Laurentidian Ice Sheet acting as a motor behind HEs. HEs occur in cold climatic spells and they are followed by an influx of fresh water into the northern Atlantic (Hemming, 2004). The provenance of the source area for the IRD is not evident and it is easier to eliminate sources than to prove an area as the source (Hemming, 2004). Heinrich events especially seem to have a connection to the
Laurentide ice-sheet dynamics (Alley and Clark, 1999; Calov et al., 2002, Hemming, 2004).

Different HEs have different characteristics and H3 and H6 especially differ from the others. During these events, the IRD flux is only moderately increased. Studies by Grousset et al. (1993) revealed a distinctly European origin of IRD in isotopic data for Heinrich layer H3.

The timing of these events in radiocarbon years and the conversion to calendar years produces large error margins (Kitagawa and van der Plicht, 1998). The duration of the HEs varies widely, from 208 to 2280 years (Hemming, 2004).

A compilation of the D-O oscillations is given in Table 3, with the numbering applied by Bond and Lotti (1995).

Several authors consider HEs to be characterised by an abrupt increase in IRD (Heinrich 1988, Bond et al., 1992). This does not hold for sites close to the former ice sheet front (Kirby and Andrews, 1999), where the association of IRD and HEs is far from obvious (Andrews, 2000). This fine-grained, laminated structure of HE units suggests that deposition happened in combination with meltwater (Hillaire-Marcel et al., 1994) and thus the HEs are associated with the generation of massive amounts of meltwater at the bed of the ice stream (Andrews and MacLean, 2003).

The record of cores located near the Scandinavian coast show well-defined IRD-layers with sharp bottom and top boundaries (Elliot et al., 2001). These records also indicate that HEs are characterized by nearly synchronous deposits in all locations, as the timing spans about 1 ka with a duration of a few hundred years at the sites. The inter-HEs show a synchronous increase only in the Norwegian Sea area. The IRD records also exhibit a change in regime between 20 and 13 ka ago. As these time-transgressive events display no clear latitudinal pattern, they are caused by different iceberg source areas and trajectories (Elliot et al., 1998). European marine ice sheets were actively discharging icebergs in millennial-scale pulses during the glaciations and depositing IRD layers in close proximity to the ice sheet margin (Bauch et al., 2001).

The HEs are thought to have a connection to ice stream activity and ice shelf collapse (Hulbe et al., 2004). The basal debris can survive fast melting if the iceberg is capsized, either bottom to top or bottom to side. Thus the debris’ lifetime in icebergs is enhanced and it can be deposited over a wider area. Deep topographic troughs in the ice shelf base may encourage water to freeze to the ice shelf base (Bombosch and Jenkins, 1995; Robin, 1979). This is caused by meltwater plumes rising up-gradient and becoming supercooled. The total sediment content in an ice shelf is thus dependent on the local topography and on the interval over which the ice was subject to basal melting before refreezing (Hulbe, 1997). This is supported by studies of the Hudson ice stream (Andrews and MacLean, 2003).

A mechanism without surge events has also been suggested to produce centurial-millennial IRD-pulses from marine ice sheets (Hindmarsh and Jenkins, 2001). They showed that a mechanism that includes external forcing in the form of climate warming and following internal glacial response can create pulses of IRD.

Table 3. Results of D-O oscillations and timing in ka BP for core Na87-22 (Elliot et al., 2002) and in calendar ka (Bond et al., 1997).

<table>
<thead>
<tr>
<th>D-O</th>
<th>Age (ka)</th>
<th>Cal. age (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>15.9</td>
<td>19</td>
</tr>
<tr>
<td>b</td>
<td>17.3</td>
<td>21</td>
</tr>
<tr>
<td>c</td>
<td>19.0</td>
<td>22</td>
</tr>
<tr>
<td>d</td>
<td>23.3</td>
<td>25.5</td>
</tr>
<tr>
<td>e</td>
<td>24.5</td>
<td>28</td>
</tr>
<tr>
<td>f</td>
<td></td>
<td>31.3</td>
</tr>
</tbody>
</table>

*Table 3.* Results of D-O oscillations and timing in ka BP for core Na87-22 (Elliot et al., 2002) and in calendar ka (Bond et al., 1997).
These pulses occur after the climatic warming, as it takes time to break apart an ice shelf. The pulse can cease before the warming ceases, as the base of the ice stream warms and basal ice melts. The base of the ice sheet has to be frozen to incorporate debris, otherwise there can be a fresh water flux without a debris flux.

**Freshwater Flux**

There is an ongoing debate about the evolution of climatic events around the HEs and the D-O maxima (see e.g. Clark et al., 2002b). The meltwater pulses from ice sheets affecting the thermohaline circulation, temperatures and salinity are thought to be a significant factor in the climatic interactions. Numerical modelling studies indicate that the Atlantic thermohaline circulation (THC) is sensitive to freshwater input, especially in the areas of deepwater formation (Stocker and Wright, 1991; Stocker et al., 1992).

Prominent freshwater fluxes to a sensitive location inhibit deepwater formation. Melting and calving from ice sheets are notable sources for the fresh water, although the exact relationship with abrupt climate change is still unknown. Ganopolski and Rahmstorf (2001) showed that warm events, like the D-O maximums, can be triggered by a small perturbation in the freshwater flux to the Nordic Seas. Clark et al. (2001) found episodic increases in the freshwater flux to the North Atlantic that originated from routing and ice margin fluctuations. These pulses of freshwater cool the sea surface waters and cause climate cooling. Cold climatic events increase iceberg calving, thus increasing the fresh water input. Lambeck et al. (2002), ChapPELL (2002) and Yokoyama et al. (2001) suggested that sea level fluctuations of 10–15 m correlate to HEs. This represents the total volume of the Scandinavian LGM ice sheet or 15–25 % of the Laurentide LGM ice sheet (Lambeck et al., 2002). Rohling et al. (2004) suggested that Antarctic and Northern Hemisphere ice sheets had similarly timed meltwater contributions seen in glacial sea level changes in a time period of 75 to 20 ka ago. The relative volumes of northern and southern sources are still undefined. In their study, sea level changes around Heinrich events are quantified for H4 a sea level rise of 23–30 m, for H5 24–35 m and for H6 40–47 m.

Evidence from the Irish Sea basin suggests a meltwater pulse at 19 ka ago (Clark et al., 2004). This meltwater pulse indicates heavy melting events on one or more ice sheets. This would provide a large and rapid flux of freshwater (0.25–2 Sv) in a time span from 100 to 500 years. At least two meltwater pulses appear to have occurred during the deglaciation following the LGM: the melt-water pulse IA (mwp-IA) and IB (mwp-IB) (Koc Karpuz and Jansen, 1992). These are timed at 13.5 ka BP and 9.5 ka BP, respectively. The corresponding estimated meltwater discharges from all continental ice sheets are 14000 and 9000 km$^3$ (Fairbanks, 1989). If the volume of the Scandinavian ice sheet at those dates was $1/3$ and $1/4$ of the total northern hemisphere ice volume (Peltier, 1994), then their corresponding share of SIS meltwater production were 4700 and 2300 km$^3$. A modelling study of North American ice sheets (Marshall and Clarke, 1999) suggests that the freshwater runoff was highly variable through the last glacial cycle. Drainage to the North Atlantic averages 0.356 Sv from 60 to 10 ka BP.

In a review article on Heinrich Events, Hemming (2004) estimated the Heinrich layers spatial coverage, thickness and volume and calculates the volume of freshwater needed. The minimum freshwater influx is $3.0 \times 10^4$ km$^3$ in one year. The corresponding maximum value is $1.0 \times 10^7$ km$^3$, resulting from a 500-year inflow of 14 Sv in certain conditions.
To investigate the Eurasian ice sheet mass balance and geographic extent, a thermodynamic ice sheet model is applied, with a particular interest in calving margin discharge. This is realised by the ice-sheet model SICOPOLIS (SImulation COde for POLythermal Ice Sheets; Greve, 1997a,b), run in combination with data manipulation and analysis programs. SICOPOLIS was one of the ice sheet model codes compared in the European Ice Sheet Modelling INiTiative (EISMINT) (Huybrechts and Payne, 1996; Payne et al., 2000). The EISMINT benchmarks and intercomparisons provide valuable results of the accuracy and consistency of ice sheet model codes.

SICOPOLIS is written in Fortran95, and has been ported to Aix, Irix and Linux operating systems. For the time being, versions of SICOPOLIS for the study of the ice sheets of Greenland, Antarctica, Northern Hemisphere, Eurasia and Mars north polar area exist. The Eurasian application was done by the author.

The simulation environment set-up is presented in Fig. 5.

The climate and other data is manipulated with a Geographic Information System (GIS) for projection and then written out in ASCII-format for SICOPOLIS -input. SICOPOLIS-simulations produce packed files and time series files in ASCII-

---

**Figure 5. Simulation environment setup.**
format for some variables at predefined time steps. These result files are visualised and analysed statistically either with Matlab (commercial numeric computation and visualization software) or Elmer-program, an element method code developed at CSC (Savolainen et al., 2000). Main visualisations are done by the ElmerPost program, a post-processing package of Elmer. ElmerPost includes a SICOPOLIS-special data reading routine, developed by Dr. T. Zwing.

The simulations were done on an IBM eServer Cluster 1600. The simulation time steps varied from 100 years to 0.1 years. The small time steps were necessary at time periods of rapid changes in the climate signal. A simulation of 250 000 years took about 3 hours of wall clock time. The resulting files from one run take about 4 000 MB of disk space. In total, some 400 separate simulations were performed. Statistical analysis and visualization was done on a SGI Origin 2000 -server, and the mapping with a Geographic Information System on a desktop PC.

**SICOPOLIS**

SICOPOLIS is based on the continuum-mechanical balance equations and jump conditions of mass, momentum and energy. The model treats ice as an incompressible, heat-conducting, isotropic power-law fluid with thermo-mechanical coupling due to the strong temperature dependence of the flow law,

\[
D = E A(T', \omega) \sigma^{n-1} t^D,
\]

where \( D = \text{sym grad} \ v \) is the strain-rate tensor (symmetrised gradient of the velocity \( v \)), \( t^D \) the Cauchy stress deviator, \( \sigma = [\text{tr}(t^D)/2]^{1/2} \) the effective shear stress, \( n \) the power-law exponent and \( A(T', \omega) \) the flow-rate factor, which depends exponentially on the temperature \( T' \) relative to the pressure melting point and linearly on the water-content \( \omega \) (see Greve et al., 1998). The flow-enhancement factor \( E \) is equal to unity for pure ice and can deviate from unity due to the softening or stiffening effect of impurities in the ice. Values of the physical model parameters are listed in Table 4.

SICOPOLIS is based on the shallow-ice approximation (SIA), that is, normal stress deviators and shear stresses in vertical planes are neglected. The large-scale behaviour of ice sheets is simulated well with the SIA; however, it is not valid locally in the vicinity of ice domes and close to the margin.

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gravity acceleration, ( g )</td>
<td>9.81 m s(^{-2})</td>
</tr>
<tr>
<td>Density of ice, ( \rho )</td>
<td>910 kg m(^{-3})</td>
</tr>
<tr>
<td>Power-law exponent, ( n )</td>
<td>3</td>
</tr>
<tr>
<td>Flow-enhancement factor, ( E )</td>
<td>3</td>
</tr>
<tr>
<td>Heat conductivity of ice, ( \kappa )</td>
<td>9.828 ( e^{-0.0057T[K]} ) W m(^{-1})K(^{-1})</td>
</tr>
<tr>
<td>Specific heat of ice, ( c )</td>
<td>((146.3 + 7.253 T[K]) ) J kg(^{-1})K(^{-1})</td>
</tr>
<tr>
<td>Latent heat of ice, ( L )</td>
<td>335 kJ kg(^{-1})</td>
</tr>
<tr>
<td>Clausius-Clapeyron gradient, ( \beta )</td>
<td>8.7 \cdot 10^{-4}K m(^{-1})</td>
</tr>
<tr>
<td>Geothermal heat flux, ( q_{geo} )</td>
<td>55 mW m(^{-2})</td>
</tr>
<tr>
<td>Threshold seabed elevation for glaciation, ( z_{thresh} )</td>
<td>–500 m</td>
</tr>
<tr>
<td>Isostatic time lag, ( \tau_{iso} )</td>
<td>3000 a</td>
</tr>
<tr>
<td>Asthenosphere density, ( \rho_a )</td>
<td>3300 kg m(^{-3})</td>
</tr>
<tr>
<td>Density ( \times ) specific heat of the lithosphere, ( \rho_r c_r )</td>
<td>2000 kJ m(^{-3})K(^{-1})</td>
</tr>
<tr>
<td>Heat conductivity of the lithosphere, ( \kappa_r )</td>
<td>3 W m(^{-1})K(^{-1})</td>
</tr>
</tbody>
</table>
Glaciers have, for a long time, been classified according to their thermal structure (Ahlmann, 1935). The coldest of these are the polar glaciers, which are totally below melting temperature. Sub-polar glaciers maintain melting temperature at the base but are cold elsewhere. Temperate glaciers are at a melting temperature. In addition to these, polythermal glaciers have been suggested (Fowler and Larsen, 1978; Hutter, 1982) and observed (Classen, 1977; Blatter and Kappenberger, 1988). Polythermal glaciers include both cold ice regions and temperate regions.

SICOPOLIS ice sheet model discerns cold ice, with a temperature below the pressure melting point and temperate ice, with a temperature at the pressure melting point. Temperate ice is considered as a binary mixture of ice and small amounts of water. Polythermal ice includes both types. The interface that separates cold and temperate ice in the model is monitored using a Stefan-type energy flux and mass flux matching conditions. The general assumption is that the water content in temperate zones of polythermal ice sheets is small (Hutter, 1993). Basal sliding is assumed to be zero for a cold base and for a temperate base, it is described by the Weertman-type sliding law in the form of Greve et al. (1998),

$$v_{sl} = -C_{sl} H ||\text{grad } h||^2 \text{grad } h,$$

where \(v_{sl}\) is the basal sliding velocity, \(C_{sl}\) the sliding coefficient, \(H\) the ice thickness and \(h\) the surface elevation.

SICOPOLIS can produce fast-flow due to sliding over a temperate ice base [Eq. 2]). These fast-flow features resemble ice streams, but as no specific ice stream dynamics is included, they are a coarse approximation of the real dynamics. Ice shelves are not treated explicitly. Instead, the ice sheet is allowed to spread to the continental shelf below sea level, provided that the seabed elevation is larger than a given threshold value \(z_{\text{thresh}}\) (Forsström et al., 2003).

Elsewhere, the ice sheet thickness is set to zero, which may be interpreted as a simple model for calving and might lead to underestimated calving rates. The general understanding of iceberg calving is at the moment not sufficient to thoroughly model the calving process. The essential parameters involved in the process, like iceberg flux and calving speed, are difficult to quantify and predict (Bahr, 1995). For temperate glaciers terminating in tidewater, a simple relationship between calving speed and water depth seems to be accurate (Brown et al., 1982). As the main calving margins of our study are temperate and marine, it would thus seem that the error in our simple calving law is not tremendous in this respect. Recent studies of Arctic ice caps (Dowdeswell et al., 2002) suggest that iceberg calving in modern conditions represent around 40% of the overall mass loss. The remaining mass loss is by surface melting.

The model tracks freshwater production by (i) melting and calving, (ii) water drainage due to basal melt and (iii) water drainage from temperate layers. As there is no specific calving law included, ice arriving at the coast is directly converted into freshwater. There is no distinction between water produced by surface melt (assumed to run off instantaneously) and ice loss due to calving in the model results for freshwater production. At every simulation time step, melting anywhere on the surface as well as calving at the margins is summed up to the total volume of freshwater released.

Surface melt and calving are not separated because the PDD-method tends to melt more ice than is actually available at the ice sheet margins. This in turn leads to negative ice thickness, so the ice thickness is set to zero to get around this. This is the reason why summing up all melted ice in
SICOPOLIS would largely overestimate the amount of meltwater produced. In the SICOPOLIS results, the total freshwater flux due to mass conservation equals the sum of ice accumulation over all grid points minus the sum of ice thickness changes over all grid points. This is then the ice-volume change.

The water flux into the ice sheet base is generated from melting basal ice. The causes for melting are the geothermal heat flux and basal frictional heating from sliding over the bed. The water storage capacity of temperate ice is assumed to have an upper limit, above which surplus water is drained instantly into the ground. No model for basal water flow or storage in the ground is included. This would not be an easy task, as the current knowledge about basal conditions and melting is limited. Models and some observations of bed properties (Engelhardt and Kamb, 1997ab; Hulbe and MacAyeal, 1999; Huybrechts and de Wolde, 1999) indicate that basal melting has a strong control on ice flow (see also Tulaczyk et al., 2000; Fowler and Johnson, 1995; Raymond, 2000, Blankenship et al., 1993). The best method to include basal water would be to couple the depth of basal water with the sliding laws (Johnson and Fastook, 2002).

Isostatic depression and rebound of the lithosphere due to changing ice loads is described by a local-lithosphere-relaxing-asthenosphere (LLRA) model with an isostatic time lag $\tau_{iso}$ (LeMeur and Huybrechts, 1996; Greve, 2001). In the LLRA model, an ice load at a given position causes a steady-state displacement of the lithosphere in the vertical direction. The value of the displacement is determined by the balance between ice load and the buoyance force which the lithosphere experiences. As the asthenosphere is viscous, the lithosphere assumes the displacement after a time lag. The time lag used here is 3000 years. A more detailed description of the model is given by Greve (1997a,b, 2001).

The geothermal heat flux is prescribed at the lower boundary of the lithosphere layer. The heat flux value for Precambrian shields is 42 mW m$^{-2}$, and the global mean is 55 mW m$^{-2}$ (Lee, 1970). In Pollack et al. (1993), the observed value for Proterozoic crust is 58 mW m$^{-2}$ and for Archaean 52 mW m$^{-2}$. In a number of studies (for example Calov and Hutter, 1996; Ritz et al., 1997; Huybrechts, 2002) the geothermal heat flux is assumed to be spatially and temporally constant, with values in the range of 42 to 65 mW m$^{-2}$. It was demonstrated by Greve and Hutter (1995) that this input quantity is crucial for the modelled basal temperature, and hence to ice flow velocities. The global mean was chosen for use in this study, as there is no spatially and temporally covering data for the whole study area. The effect of geothermal heat flux in SICOPOLIS simulation is reported in Greve (in press).

**Application to Eurasia**

The model computes the three-dimensional evolution of ice extent, thickness, velocity, temperature and water content in the temperate ice region and age in response to external forcing. The forcing due to climate and heat from Earth's interior, are (i) the mean annual air temperature at the ice surface, (ii) the surface mass balance, which is ice accumulation (here assumed to be snowfall) minus ablation (here assumed to be melting), (iii) the global sea level and (iv) the geothermal heat flux entering the ice mass from below. Surface processes like refreezing are accounted for. Sea level changes were derived from the SPECMAP record (Imbrie et al., 1984).

The schematic diagram for the simulation is depicted in Fig. 6.

In an earlier application of the model (Forsström et al., 2003), the geographic area was bordered by Novaya Zemlya and
Figure 6. A schematic presentation of processes involved in SICOPOLIS model.

Figure 7. The study area of SICOPOLIS simulations for Eurasia. The area is from the British Isles in the west to the Taymyr Peninsula in the east. Main geographic features referred in the text are indicated.

Urals in the east and by the North Sea area in the west. Denmark was just on the southern border and part of Greenland in the northwest. Resulting ice sheet dynamics indicated that the chosen geographic area could not satisfactorily incorporate known palaeo features.

The numerical grid applied here is 40 km × 40 km. The geographical area in these simulations extends from the northern Atlantic to the Taymyr Peninsula and covers an area of 6040 × 2840 km², with a total areal coverage of 17.1536 × 10⁶ km². The study area is shown in Fig. 7.

The domain was chosen to include both the British Isles and the eastern section of the Eurasian ice sheet. The area includes the Alps in the south, and the Urals and the highlands of Taymyr and Tunguska in the east. Greenland is not part of the area, so there is no modern ice sheet in the study area, just glaciers.

A stereographic projection with a reference latitude of 71° is used. The model domain grid spacing for the whole region is 40 km in the horizontal, leading to a total of 151 × 71 grid points. Computations are carried out on a stereographic plane with standard parallel 71°N, defined by the Cartesian coordinates x and y. The vertical coordinate z is defined positive upwards. The zero-level is the present-day reference geoid. The distortion arising from the projection is accounted for by introducing the
corresponding components of the orthogonal metric tensor in all terms with horizontal derivatives in model equations.

The vertical resolution is 21 grid points in the cold-ice region, 11 grid points in the temperate-ice region (if existing) and 11 grid points in the lithosphere (Greve, 1997a). The surface elevation data is from the ETOPO5 data-set (NOAA Product Information Catalog, 1988) and is also interpolated to the coarser grid of SICOPOLIS.

The model is initialised in an arbitrarily chosen ice-free state, 250 ka ago. The model needs approximately 50 ka to clear the effect of the initial state. Climate conditions are tied to the Greenland ice core record. Thus, the simulation runs through one complete glaciation and deglaciation before the time period of interest, the Weichselian, which started 115 000 ka BP and ended 10 000 ka BP.
The Weichselian glacial stage consists of several cooling and warming events. The description of a time-evolving climatology is essential for the simulation of ice-sheets through glacial cycles. This can be realised in several ways. The most sophisticated solution would be to connect a climate model and the ice-sheet model, but this is computationally very costly. A general solution is to use snapshots of climate, produced by climate models. This was done successfully for the Northern Hemisphere deglaciation by Charbit et al. (2002). They assumed climate to vary in time linearly between climates derived from separate climate model runs. A perturbative method of the present day climate was applied to minimize errors resulting from GCM model deficiencies. This was realised by adding the anomaly in temperature resulting from modern and LGM climate runs to the present day observed temperature.

The first attempt to parameterise past climatic conditions for Eurasian SICOPOLIS simulations was similar to the method applied by Greve and others (1999). Past temperatures were calculated from present values with a conversion formula accounting for general temperature deviations and the changes in surface elevation. However, this approach inevitably leads to a very extensive glaciation, especially in the east. The reasons for this are the differential spatial patterns of present and LGM climate conditions (Forsström et al., 2003), and it is therefore not suitable for a detailed study of Eurasian glaciation.

In this study of Eurasian Weichselian glaciations, the anomaly of temperature and precipitation, resulting from a climate model run, is used in a similar way as by Charbit et al. (2002). This anomaly is scaled by a glaciation index in which maximum glacial conditions are those at the LGM and maximum interglacial conditions are those of the present-day. This differs from Charbit et al. (2002), as they used also lapse rate and surface topographies to define past values. Their model includes thus a feedback from ice sheet height.

**Climate data**

Modern surface temperature and precipitation values are from the measured and modelled data sets by Legates and Wilmott (1990). These data sets were chosen, as these were the data sets used by Siegert et al. (2001) and Charbit et al. (2002), and the comparison of the simulation results with results from existing literature was easier. These data sets have been gathered from the Global Historical Climatology Network (Vose et al., 1992) and station records from the years 1950 throughout 1996. The source resolution is \(0.5° \times 0.5°\), and the data is interpolated for each season with inverse-distance weighting to the 40-km grid of SICOPOLIS.
For the LGM climate conditions, temperature, precipitation, and elevation values derived from the UKMO LGM simulations (Hewitt and Mitchell, 1997) are used. The UKMO21 dataset was chosen from comparisons of resulting ice sheet mass budgets in the Palaeoclimate Modeling Intercomparison Project (PMIP) simulations (Pollard and PMIP Participating Groups, 2000).

The UKMO data-set was closest to QUEEN project results, with main ice masses in the west and only slight glaciation in the east. The UKMO21 values for temperature and precipitation in these results have a minima in the northeast, which is an excellent match for the evaluated LGM conditions. The coarse atmospheric model output was again interpolated with inverse-distance weighting to the model grid.

Forcing method

The seasonal temperature and precipitation distributions are interpolated linearly between the present (Legates and Wilmott, 1990) and LGM (UKMO21) anomaly values with a glaciation index \( g(t) \). This index scales the GRIP \( \delta^{18}O \) record (Dansgaard and others, 1993) to represent LGM \( (g = 1) \) and present \( (g = 0) \) conditions (Fig. 8).

A simple approach is to interpolate seasonal climate variables linearly between the present and LGM conditions. Forsström et al. (2003) used a glaciation index \( g(t) \), that scales the GRIP \( \delta^{18}O \) record (Dansgaard et al., 1993) to represent LGM \( (g = 1) \) and present \( (g = 0) \) conditions (Fig. 8). Here, this approach is refined by defining anomalies

\[
\begin{align*}
    f_{\text{anom}}(x, y) &= f_{\text{LGM}}(x, y) - f_{\text{present}}(x, y),
\end{align*}
\]

where \( x \) and \( y \) are Cartesian coordinates which span the stereographic plane, \( t \) is the

Figure 8. Glacial index \( g(t) \) as a function of time. This index scales the GRIP \( \delta^{18}O \) record (Dansgaard et al., 1993) to represent glacial \( (g = 1) \) and present \( (g = 0) \) conditions.
time, \( f \) is the surface temperature or precipitation (Fig. 1) and \( f_{\text{present}}(x,y) \) and \( f_{\text{LGM}}(x,y) \) are the spatially resolved data for the two time-slices, respectively. Anomalies, with respect to present day conditions, are used to eliminate systematic errors present in GCM simulations. This approach has been applied in Forsström and Greve, (2004).

The temperature or precipitation at any time \( t \) is then calculated based on data-sets for modern precipitation and temperature,

\[
f(x,y,t) = f_{\text{present}}(x,y) + g(t) f_{\text{anom}}(x,y),
\]

where \( f_{\text{present}}(x,y) \) and \( f_{\text{anom}}(x,y) \) are the spatially resolved data for the two time-slices, respectively.

Conversion from seasonal precipitation \( P \) to seasonal snowfall (solid precipitation) \( S \) is done with the empirical relation by Marsiat (1994),

\[
S = P \times \begin{cases} 0, & T_s \geq 7^\circ C, \\ (7^\circ C - T_s) / 17^\circ C, & -10^\circ C \leq T_s \leq 7^\circ C, \\ 1, & T_s \leq -10^\circ C, \end{cases}
\]

where \( T_s \) is the seasonal surface temperature. Mean annual air temperature is the mean of seasonal temperatures.

Climatic forcing is completed by the degree-day method for surface melting in the form presented by Reeh (1991). The ablation model is based on the degree-day method and accounts for the daily and annual temperature cycle, a different degree-day factor for ice and snow melting and superimposed ice formation. The number of degree days is calculated by determining the average departure of the mean temperature above a reference temperature (0 °C) on each day and finding the mean for several days.

The parameters used here are: degree-day factor for snow melt \( \beta_{\text{snow}} = 3 \) mm w.e. d\(^{-1}\) °C\(^{-1}\), degree-day factor for ice melt \( \beta_{\text{ice}} = 12 \) mm w.e. d\(^{-1}\) °C\(^{-1}\), saturation factor for the formation of superimposed ice \( P_{\text{max}} = 0.6 \), standard deviation of short-term, statistical air-temperature fluctuations \( \sigma_{\text{stat}} = 5 \) °C. The sea level changes are derived from the SPECMAP record (Imbrie and others, 1984) and the global-mean value \( q_{\text{geo}} = 55 \) mW m\(^{-2}\) is used for the geothermal heat flux.
The basic forcing method described earlier is in the following applied with UKMO PMIP data (simulation #2), in cold ice mode (simulation #3), with modified PDD-factors (simulations #4 and #5) and with the modified climate anomaly runs (simulation #6). The study with sinusoidal forcing (simulation #1) was constructed otherwise, as will be described in detail in the next section.

Several other simulations were conducted to test the modelled ice sheet sensitivity. These included simulations with enhanced sliding and added viscosity.

Sinusoidal forcing

To study the effect of periodical forcing on the Eurasian Ice Sheet, a sinusoidal temperature forcing is applied. The peak temperature deviation of the forcing was 10 °C and the period was 20 ka. The forcing is described by

\[ \Delta T_{ma}(t) = 10 \, ^\circ C \times (\cos(2\pi t_{Mil}/t_{Mil}) - 1), \]  

where \( \Delta T_{ma} \) is the difference in mean annual temperature, \( t_{Mil} \) is the period, and \( t \) is the time of interest.

The simulations were done in cold ice mode. Sea level was kept constant at today’s value. Accumulation was coupled linearly to the temperature deviation.

PMIP FORCING

Here, the anomaly in climate forcing is defined from the previously mentioned UKMO PMIP simulation results by

\[ f_{anom}(x,y) = f_{UKMO}^{LGM}(x,y) - f_{UKMO}^{present}(x,y), \]  

where \( x \) and \( y \) are Cartesian coordinates which span the stereographic plane, \( t \) is the time, \( f \) is the surface temperature or precipitation (Fig. 1) and \( f_{UKMO}^{LGM}(x,y) \) and \( f_{UKMO}^{present}(x,y) \) are the spatially resolved data for the two time-slices, respectively.

The temperature or precipitation at any time \( t \) is then calculated based on the Legates and Wilmott (L&W) data-sets for modern precipitation and temperature,

\[ f(x,y,t) = f_{present}^{L&W}(x,y) + g(t) f_{anom}(x,y), \]  

where \( f_{present}^{L&W}(x,y) \) and \( f_{anom}(x,y) \) are the spatially resolved data for the two time-slices, respectively. The temperature and precipitation data is calculated in a similar way for all seasons.

This approach is referred to as the PMIP forcing in the following text.

Also the UKMO climate model 6 ka results could have been used as an additional reference state, but as the Eurasian ice sheets had melted before that time period, the usefulness of this approach would be questionable. The inclusion of the 6 ka results might even increase the deglaciation.
Here, the best glacial index zero-reference level is the present-day climate as direct measurements of the climate parameters are available.

COLD ICE MODE
Cold ice is ice with temperature below the pressure melting point. In principle, cold ice can be treated as a viscous, heat-conducting, incompressible one-component fluid.

The polar glaciers often consist solely of cold ice. Also, the large ice sheets on Earth contain mainly cold ice. The responses of the cold and temperate ice are different, as the flow parameter in Eq. 1 increases rapidly as ice warms. A comparison of results from cold ice mode and polythermal mode allows us to deduce the effect of polythermal ice on the total dynamic of the ice sheet. Basal temperatures can, of course, reach the pressure-melting temperature and melt the ice, thus allowing the onset of basal sliding.

This approach is referred to as the cold ice mode in the following text. Simulation setup in cold ice mode was otherwise similar to the PMIP forcing setup described earlier.

MODIFIED PDD FACTORS
The PDD method is based on empirical relations between positive degree days and observed melting. For the calculation, three phases are needed. First, one has to derive the expressions for the annual mean and mean July air temperatures, secondly, the positive degree days must be calculated and climate variability must be accounted for with a stochastic term, and thirdly, the ablation needs to be estimated (Van der Veen, 1999).

The observations come mainly from a series of measurements in Greenland (e.g. Reeh, 1991). Critics of these observational values claim that the time series are too short and are spatially unrepresentative. PDD calculations are sensitive to variations in air temperature. A small increase in climate variability can have large effects on meltwater production (van der Veen, 1999).

Various authors use a range of degree-day factors. Braithwaite (1995) asserted that at low temperatures, even values of $\beta_{ice} = 20$ mm w.e. d$^{-1}$ °C$^{-1}$ are realistic. For snow, he estimated the ablation factor to be in the range of $\beta_{snow} = 3...5$ mm w.e. d$^{-1}$ °C$^{-1}$

Hagen et al. (1998) used the values $\beta_{snow} = 2.9$ mm w.e. d$^{-1}$ °C$^{-1}$ and $\beta_{ice} = 4.6$ mm w.e. d$^{-1}$ °C$^{-1}$ for Svalbard, and $\beta_{snow} = 4.1$ mm w.e. d$^{-1}$ °C$^{-1}$ and $\beta_{ice} = 7$ mm w.e. d$^{-1}$ °C$^{-1}$ for Norway.

The model sensitivity to $\beta_{ice}$ was tested with extreme values of 5 mm w.e. d$^{-1}$ °C$^{-1}$ to 12 mm w.e. d$^{-1}$ °C$^{-1}$. The snowmelt factor is not varied due to its smaller influence on computed melt rates.

MODIFIED LGM PRECIPITATION AND TEMPERATURE ANOMALIES
Prior experience and simulation runs suggest that our basic forcing may lead to an incorrect estimate of ice mass and ice-sheet extent in the eastern part of the study area with PMIP UKMO data. This is because PMIP runs use the ice sheet configuration of ICE4G by Peltier (1994). ICE4G is estimated to be too extensive in the eastern sector of the Eurasian ice sheet. This is stated for example by the QUEEN project results for the Eurasian ice sheet extent.

AGCM simulations indicate that at the LGM, atmospheric circulation over the Atlantic Ocean had two branches (Marsiat and Valdes, 2000). A very weak flux of heat entered the Eurasian continent. In the Kara Sea region, an anticyclonic air-flow system isolated Taymyr Peninsula and Laptev Sea from the Barents Sea and Fennoscandian climate system. This leads to much lower precipitation rates in the eastern section than at present day. QUEEN-based Gen-
eral Circulation Model LGM runs produce precipitation-rate anomalies of -200 mm a$^{-1}$ in the northeast and -400 mm a$^{-1}$ in the southeast part (Siegert and Marsiat, 2001). Not having complete GCM results based on the QUEEN extent yet, the above-defined precipitation anomalies were modified instead.

The comparison of proxy data and numerical simulations has several problems (Renssen and Osborn, 2003). One technique to unify these is called DATUN (Data Assimilation Through Upscaling and Nudging) (Jones and Widmann, 2002). This method nudges the climate in an atmospheric model towards a state that has been reconstructed from palaeodata. Standard techniques for assimilation of palaeodata are not yet available.

In the following, two modifications are made to the PMIP climate anomalies. At first, the precipitation anomaly in the eastern sector of the model domain was set to a value between zero and -400 mm, depending on the season (Fig. 9). Second, the LGM temperature anomalies according to Kageyama et al. (2001) were applied as modifications to the PMIP anomaly. This data is presented on a coarser grid and convolution filtering was applied to the interpolated values for smoothing. Anomalies with the coldest temperatures were assigned to the winter months (December, January, February), and those with the

![Winter, Spring, Summer, Autumn precipitation anomalies](image.png)

Figure 9. Anomalies for precipitation (in mm/) at the LGM, with the UKMO PMIP data modified in the eastern part of the model domain. Areal coverage is 6040 × 2840 km$^2$ in the stereographic plane. Axis units are in km. Reproduced from Forsström and Greve, 2004.
The warmest temperatures were assigned to the summer months (June, July, August). Spring (March, April, May) and autumn (September, October, November) values were interpolated from the winter and summer as

\[ f_{\text{anom}} = f_{\text{anom}} = \frac{(f_{\text{anom}} + f_{\text{anom}})}{2}. \]

The modified temperature and precipitation anomalies are depicted in Fig. 10.

This method is referred to as modified LGM anomaly in the following text.

Figure 10. Anomalies for temperature (in °C) at the LGM modified from results by Kageyama et al. (2001). Areal coverage is 6040 × 2840 km² in the stereographic plane. Axis units are in km. Reproduced from Forsström and Greve, 2004.
Errors in forcing
With the anomaly method, errors in computed present and past climate parameters are assumed to cancel each other (Charbit et al., 2002). Other sources of error include processes like feedback effects due to changed ice sheet topography, the calving law and unified ablation coefficients. The degree-day model, which relates the melting rate to the air-temperature excess above the melting point, does not account for the influence of wind speed, albedo and cloud cover. Using an energy-balance model (Braithwaite, 1995) would improve the surface melting calculations.

Prior experience with different data sets suggests that the Fennoscandian Ice Sheet model is very sensitive to even small changes in climate parameters, highlighting the quality of climate data applied (Forsström et al., 2003). An example of this is the effect of snow-melt factors, seen in Fig. 17.

To evaluate our model sensitivity and to validate the application of the model, several runs were conducted to study the influence of parameterisations and forcing data sets. In these runs, we studied the effect of different climate data sets, cold and polythermal mode and different PDD factors, described in detail below.
Ice sheets resulting from different simulation runs differ in extent and volume. In the following, simulation results are studied by comparing the variation in time of ice volume, ice thickness, sea level equivalent, ice-covered area, area covered by temperate ice, freshwater production due to melting and calving, volume of temperate ice and maximum thickness of temperate ice layer.

Table 5 gives an overview of the different simulations.

**Sinusoidal forcing (simulation #1)**

In a previous study (Greve, 1995), an EISMINT-case simulation of SICOPOLIS with sinusoidal forcing of 20 ka, 40 ka and 100 ka was run. Results from these runs showed that the total ice volume grows to a typical saddle form. Starting from temperature maximum, the ice volume grows quickly to the first volume maximum as the temperature falls. After this maximum, the volume declines slightly, and starts to grow again after temperature minimum. The main maximum of the ice volume is reached in the halfway between the temperature minimum and maximum.

This phenomenon is connected to the ice-covered area. The increase in ice volume simultaneously with decreasing temperatures is not the result of increasing ice sheet thickness but rather an effect caused by the expanding coverage of the ice sheet. The maximum elevation of the ice sheet actually decreases throughout this phase. Ice volume increases as long as the spatial extent reaches the maximum continental margin positions. This can be seen from the result plots in the truncated maximums for ice sheet extent. The timing of these truncated maximums falls in the time of ice volume double maximum. After the continental margin position is reached, the ice sheet is prohibited from further expansion and the ice sheet declines slightly.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Set-up</th>
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<tbody>
<tr>
<td>#1</td>
<td>Sinusoidal forcing</td>
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<tr>
<td>#2</td>
<td>PMIP forcing</td>
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<tr>
<td>#3</td>
<td>cold-ice mode</td>
</tr>
<tr>
<td>#4</td>
<td>$\beta_{\text{ice}}$ reduced to 5 mm w.e. d$^{-1}$°C$^{-1}$</td>
</tr>
<tr>
<td>#5</td>
<td>$\beta_{\text{ice}}$ increased to 12 mm w.e. d$^{-1}$°C$^{-1}$</td>
</tr>
<tr>
<td>#6</td>
<td>Modified LGM anomaly based on Kageyama et al. (2001)</td>
</tr>
</tbody>
</table>
Only after the culmination of temperature minimum, the ice volume starts to grow again as the precipitation increases with increasing temperatures. The turning point for the main volume maximum is reached when the temperature is so high that the melting exceeds accumulation. After this, the ice volume shrinks quickly and at the time of maximal temperatures, the ice volume is minimal.

The values describing temperate ice history (volume of temperate ice, area covered by temperate ice and maximum altitude of temperate ice layer) show a component of fast changes that become more pronounced with shorter time periods of the sinusoidal forcing. This high-frequency oscillation is an effect due to truncation, noise from the natural oscillation of the system.

This exercise was repeated in this study for the Eurasian area. The results for sinusoidal forcing of 40 ka are depicted in Fig. 11.

The effect of more complex geometry (the Eurasian topography) is mirrored in the simulation results, as they show clearly the influence of the initial state for the first 50 ka. From this on, the results agree with the EISMINT case, except for the volume maximums. Only some of them show the saddle form. As the saddle form was a result of the areal expansion, here the topography is more limiting to the ice sheet expansion than in the EISMINT case.

Figure 11. Sinusoidal forcing results for simulation# (sinusoidal forcing of 40 ka) as functions of time: glacial index, sea level, ice volume, maximum ice thickness, sea-level equivalent, ice-covered area, area covered by temperate ice, freshwater production due to melting and calving, volume of temperate ice, maximum thickness of temperate ice layer.
The PMIP forcing (simulation # 2)

The simulated results of the PMIP forcing indicate a first Eurasian glaciation at 60 ka ago. There is a local glaciation at 90 ka ago, centred on Novaya Zemlya, Franz Josef Land, Taymyr and Severnaya Zemlya. The resulting ice sheet configurations for the EWG is depicted in Fig. 12. It is very modest compared to the EWG in the QUEEN results.

The first real glaciation in the simulation results, centred around 70–60 ka ago, was initiated in Scandinavia. The central dome is located in the mountains. The British Isles are not glaciated, but the Taymyr area and upper arctic Russian mainland are glaciated. Franz Josef Land, Svalbard and Novaya Zemlya have local ice caps and Severnaya Zemlya is also glaciated. The Fennoscandian ice sheet covers the Finnish and Baltic areas and extends deep into the Russian plains. The Barents Sea is not totally glaciated. As the MWG ice sheet retreats, the Fennoscandian ice sheets disappears first. The High Arctic ice caps are still growing at this stage and reach into central areas of the Barents Sea. This ice cap remains over Franz Josef Land through the remaining Weichselian glaciation, spreading occasionally deeper into the Barents Sea. The MWG has a double maximum, the first concentrated over western Scandinavia and the second in the eastern

Figure 12. Ice sheet configuration at the EWG in the PMIP forcing simulation results.
and High Arctic areas. The resulting ice sheet configurations for the MWG are depicted in Fig. 13.

The PMIP forcing configuration produces a relatively expansive LGM ice sheet in the east while relatively little glacial ice forms in the western-most region of the model domain. The northern Arctic islands are covered by extensive ice masses and the Yamal and Taymyr Peninsulas are glaciated. The Barents Sea Ice Sheet extends in the east to the Novaya Zemlya coast, with a maximum surface elevation of 2 km. The ice sheet has a maximum areal extent at 22 ka ago, when the lowlands to the south of the main ice domes are covered by a few hundred meters of ice. The BIIS appears as a modest few hundred meters of ice, covering Scotland and north Ireland for some 1000 years around the LGM. The resulting ice sheet configurations for the LGM are depicted in Fig. 14.

At the LGM, the Barents Sea Ice Sheet does not cover the whole Barents Sea area. For example, the Barents Sea north of the Kola Peninsula is ice-free. This is due to the precipitation anomaly defined by the PMIP output: the central areas of the ice sheet are dry. The precipitation is forced with the glacial index, but the real change in precipita-
tion in the area is apparently not that simple. The simulated SIS is two-domed. One dome is centred in the Norwegian highlands and the other in Finland. The maximum surface elevation for both is about 2 km.

The main ice sheet domes grew between the LGM and 18 ka ago, even as the spatial extent decreased. After that time, the SIS and southeastern sector rapidly disintegrated. The cooling at 12 ka ago caused a short, modest glaciation of Scandinavian areas. Thereafter, the SIS disappears and only the BSIS remains. By 9 ka ago, all ice disappeared from the area.

The effects of model-simulated fast ice flow are readily apparent in the ice sheet topography. Relatively rapid draw-down produces troughs. The troughs form in regions identified as sites of palaeo ice-stream flow. This is discussed in detail in Forsström et al. (2003).

Figure 15 summaries some of the simulated parameters during the evolution of the Eurasian glaciation. The ice volume varies along with the glaciation index. Before the LGM, there are smaller peaks in volume around 60 and 35 ka ago. Near the LGM, ice-sheet volume peaks at 22 ka ago ($7.3 \times 10^6$ km$^3$) and 19 ka ago ($8.1 \times 10^6$ km$^3$). From that time on, the ice sheet declines, with temporary hiatuses at 16–14 ka ago and 12–11 ka ago. The maximum
Figure 15. Forcings and results for simulation #2 (PMIP forcing) as functions of time: glacial index, sea level, ice volume, maximum ice thickness, ice-covered area, freshwater production due to melting and calving, volume of temperate ice, maximum thickness of the temperate layer.

Volume at 19–18 ka ago is an exception to the otherwise synchronous trends in ice volume and glaciation index; the ice volume grows as temperatures are on the rise. This growth is concentrated on the domes mentioned earlier and is probably caused by increased precipitation accompanying climate warming. Deglaciation starts in Scandinavia and the SIS is the first ice sheet to disappear. The BSIS disintegration is sluggish, until all ice has vanished by 9.0 ka ago.

The volume of temperate ice is, in general, closely connected to the total volume of the ice sheet. The volume of temperate ice shows several peaks in the simulation: at 35 ka ago (75 x 10³ km³), 27 ka ago (90 x 10³ km³), 23–22 ka ago (140 x 10³ km³), 20–19 ka ago (60 x 10³ km³) and 12 ka ago (15 x 10³ km³). The peak at 19–18 ka ago is unusual, as it is much smaller than expected from the total volume. This is presumably caused by the large surface accumulation on the ice domes, leading to suppression of temperatures. Larger spatial extent coincides with larger areas of temperate ice at the base. This is evident as the flow is enhanced due to the existence of temperate ice and water at the base.

The combined freshwater production due to melting and calving vary in step with the climate forcing. There are peaks at 34–35 ka ago (1340 x 10³ km³/a), 23–26 ka ago (7340 x 10³ km³/a), 21 ka ago (1380
× 10³km³a⁻¹), wider peaks at 31–32 ka ago (4000 × 10³km³a⁻¹), 29 ka ago (1420 × 10³km³a⁻¹), 17–18 ka ago (3300 × 10³km³a⁻¹), 14 ka ago (2450 × 10³km³a⁻¹) and finally a small peak at around 11 ka ago.

The spatial extent of the ice sheet changes continually through the simulation. Maximums in the ice-covered area occur at 35 ka, 27–25 ka and 22–21 ka ago (8.5 × 10⁶ km²). The maximum volume at 19–18 ka ago is accompanied by a smaller spatial extent (6 × 10⁶km²). Ice-cover maximum with thinner ice peaks at the LGM and the volume growth after that is presumably caused by increasing precipitation as the temperature of air rises.

Not presented in the summarising figure is the fact that the maximum ice-sheet thickness of 3.3 km occurs at 17.3 ka ago, but values above 3.0 km are reached from 19.3 ka ago n. The ice sheet gradually declines and the final disintegration starts around 10 ka ago.

**COLD-ICE MODE (SIMULATION # 3)**

The importance of including polythermal ice processes is tested by re-running the PMIP forcing set-up in cold ice only model. In this case, the Stefan-type conditions at the transitionsurface between cold ice and temperate ice (which occurs typically in thin layers at the base) are ignored. Where heat balance would yield temperatures above the melting point, the ice temperature is reset to the pressure melting point. The results for the cold ice mode are summarized in Fig. 16.

Figure 16. Like Fig. 15, but for simulation #3 (cold ice mode) as functions of time: glacial index, sea level, ice volume, maximum ice thickness, ice-covered area, freshwater production due to melting and calving, volume of temperate ice, maximum thickness of the temperate layer.
The main features of the polythermal and cold ice modes are very similar. The ice-sheet volume tends to be slightly larger in the cold ice mode until the deglaciation phase. The same pattern applies to spatial extent. The implication is that the main restrictions to the ice sheet growth are topography and climate forcing. The freshwater production peaks occur in the same time periods as in the PMIP forcing approach and with nearly identical values.

**Different PDD factors**

(simulations # 4 and # 5)

Here model runs are compared with the two extreme degree factors for ice, $\beta_{\text{ice}} = 5$ mm w.e. d$^{-1}$ °C$^{-1}$ and 12 mm w.e. d$^{-1}$ °C$^{-1}$. The results are displayed in Fig. 17.

The volumetric and area features of the ice sheet are similar to the PMIP forcing run, with similar timing of freshwater production due to melting and calving. The peak magnitudes are a little lower than the maximum ablation values. It seems that the SIS is very sensitive to the PDD factors during the deglaciation phase.

The ice sheets resulting from our models are very sensitive to variations in ablation. Simulation #4 (low value of $\beta_{\text{ice}}$) strongly influences SIS and BIIS growth and decay, leading to a glaciation history that more resembles the QUEEN project outcome. The BIIS is still very modest, but

![Figure 17. Ice volume and ice-covered area as functions of time for simulations #2 ($\beta_{\text{ice}} = 7$ w.e. d$^{-1}$°C$^{-1}$, solid), #4 ($\beta_{\text{ice}} = 5$ mm w.e. d$^{-1}$°C$^{-1}$ dashed) and #5 ($\beta_{\text{ice}} = 12$ mm w.e. d$^{-1}$°C$^{-1}$dash-dotted).](image-url)
persists for some thousands of years near the LGM. The BSIS ice dome is slightly higher than for the primary model. Simulation #5 (high value of $\beta_{\text{ice}}$) causes a thinner BSIS, which is in accordance with the QUEEN results. The SIS is overly small in extent and volume.

From Fig. 17, one can conclude that a reduction of the degree-day factor has a stronger impact on ice sheet volume than does an increase. The implication is that in western areas, reducing ablation is the key to a more extensive glaciation.

**Modified LGM anomaly (simulation # 6)**

Both modifications in climatology improve some aspect of the simulated ice sheet. The reduced LGM precipitation yields ice-sheet configuration in good general agreement with the QUEEN reconstruction.

In the simulation results, the EWG starts from the High Arctic Island around 96 ka ago. The main ice masses develop over Scandinavia, and the Barents Sea remains ice-free. The ice masses disappear from

Figure 18. The simulated ice sheet extent at the EWG, resulting from the modified LGM anomaly forcing. The height scale is included in the picture. Maximum ice sheet height in the Fennoscandia is approximately 1.5 km.
Scandinavia around 90 ka ago, to reappear again with a maximal extent at 86 ka ago. Modest ice masses (300–500 m thick) are present on the Eurasian coastal mainland, and also on the Taymyr Peninsula. Both of these ice volume maximums are centred on the Scandinavian mountains. This is also what Sejrup et al. (2003) have suggested. After both of these maximums, an ice sheet with a thickness of 500 m spreads to Finnish lowlands.

The MWG is, in our simulation, the result of a series of heavily glaciated states, punctuated by warming events. The maximal ice volume is reached at 61 ka ago. This glaciation starts from the Franz Josef Land ice cap and is fortified by extensive Scandinavian glaciation. The Barents Sea remains open in the southern central areas, along with the eastern parts of the Kara Sea. The ice sheet height is around 2 km at the maximum, and the central domes are located on the British Isles (around 1.5 km), in southern Norway (2 km), in southern Finland and the Baltic (2 km) and Franz Josef Land (1.5 km). The ice sheet covers the coastal Russian mainland with an ice sheet of 1 km maximum. The Putorana Plateau has a local thin ice cap. At a later phase of this glaciation, the Barents Sea ice

Figure 19. The simulated ice sheet extent and height at the MWG, resulting from the modified LGM anomaly forcing. The height scale is included in the picture. The maximum ice sheet height in the Fennoscandia is approximately 2 km.
sheets grows, leaving only a small central part open. At this stage, ice also more heavily covers the southern Kara Sea and the Russian mainland with one ice tongue nearly reaching the Urals. The Barents Sea Ice Sheet is approximately 2 km thick at this stage. As the ice sheet starts to thin over Fennoscandia, it is still growing in the Barents area.

At the LGM, the main ice domes form over the Barents Sea and Scandinavia. The volume of the ice sheet as well as the maximum ice-sheet thickness stay lower than estimated by several authors (Elverhøi et al., 1993; Lambeck, 1995; Peltier (1994)). The Barents Sea remains partially open and the BSIS is limited to the east coast of Novaya Zemlya. Ice sheet extent at 22 ka ago is shown in Fig. 21. Other model parameters are summarized in Fig. 20. The pulses of freshwater have the same timing as in the PMIP forcing approach. The flux is of course smaller, as the area and volume are smaller.

Simulation # 6 (modified LGM anomaly) produces an ice sheet geometry that agrees very well with the QUEEN data for the LGM. The extent is shown in Fig. 22. In this simulation, the extent and volume of both the SIS and the BSIS have a maximum around 22 ka ago. The elusive BSIS reaches significant size in this model run, originating from 26 ka ago, and having a maximum surface elevation of a 2 km at 21.5 ka ago.
The LGM time period is characterized by high mass fluxes in western sites of palaeo ice streams. A considerable mass flux prevails for the NCIS area, Norwegian coast, Storfjorden and Bear Island trough. At 17 ka ago, the mass flux peaks in the Storfjorden and Bear Island region. This is also the time when the British Isles Ice Sheet has disintegrated. The mass flux continues in the sites of palaeo ice streams until 14 ka, when the SIS has completely disintegrated. Peaks in meltwater production correspond to both climate and ice-sheet discharge events. The freshwater flux was continuously high from 20–15 ka ago. The high peak between 15–14 ka ago was caused by the melting SIS. At 13 ka ago, the SIS starts to grow again, as does the BSIS. However, the SIS retreats rapidly as the climate warms and disappears about 11 ka ago, which is not in accordance with observations. The BSIS disintegrates at 8 ka ago.

Comparison of the PMIP forcing and modified LGM anomaly results

The resulting variables from PMIP forcing simulations indicate several glaciations during the Weichselian stage.
The total volume has two significant rises in the Weichselian, namely at 65–55 ka ago (appr. $5 \times 10^6$ km$^3$) and at LGM (appr. $8 \times 10^6$ km$^3$). The first is the Middle Weichselian Glaciation in these runs and it is a double-peak with a minimum around 60 ka ago. After this glaciation, there is a limited, sole peak in volume at 35 ka ago. The LGM has its largest volume at 22 ka ago.

The total volume curve peaks markedly twice in the modified LGM anomaly runs in the last 120 ka. There are two modest peaks (total volume appr. $0.5–0.7 \times 10^6$ km$^3$) at 90–80 ka ago. In these simulations, this is during the first Weichselian glaciation. The first large volume peak is around 65–55 ka ago, with the total volume being then $6–8 \times 10^6$ km$^3$. This is the middle Weichselian glaciation. The form of this maximum is a double-peak with a minimum of $4 \times 10^6$ km$^3$ at 60 ka ago. After this, there are two larger spikes in volume, developing after rapid climate cooling events at 45 ka ago ($3 \times 10^6$ km$^3$) and at 35 ka ago ($5 \times 10^6$ km$^3$). The final cooling phase culminating during the LGM starts at around 30 ka ago. At 22 ka ago, the volume is the largest during the Weichselian stage, $10.5 \times 10^6$ km$^3$. After this, the ice volume rapidly diminishes.

In the PMIP forcing run, the temperate ice has several enormous peaks during the middle Weichselian glaciation, with values of $100–150 \times 10^6$ km$^3$. The ice sheet was predominantly temperate. The LGM temperate ice volume peaks are at maximum only $80 \times 10^6$ km$^3$. In the Kageyama-style runs, the volume of temperate ice has the largest peak ($500 \times 10^6$ km$^3$) during the Weichselian at 65–55 ka ago. This peak is wide compared to LGM temperate ice volume peaks. The Eurasian ice sheet forming 65–55 ka ago was, in simulation runs, a temperate-based ice sheet. This could explain the rapid spreading of the ice sheet. Also, the first temperate ice volume peaks at 90–80 ka ago are higher than one would proportionally expect.

The freshwater production in the PMIP forcing run has the first modest freshwater fluxes at 75 ka ($2.0 \times 10^3$ km$^3$/a) and at 60 ka ago ($1.5 \times 10^3$ km$^3$/a). After this, there are several peaks with the largest one at 23–26 ka ago, but these are discussed in detail later.

In the modified LGM anomaly runs, the freshwater production has the first peak at 75 ka ago, the next at 68–69 ka ago, which is the largest during the whole Weichselian stage, with a freshwater flux of $2.2 \times 10^3$ km$^3$/a.
km$^3$/a. After this, there are several peaks, which are discussed in detail later.

The spatial extent resulting from our PMIP forcing runs is noticeable during the middle Weichselian (appr. $66 \times 10^6$ km$^2$) and around the LGM from 40–20 ka ago ($5.5–7.5 \times 10^6$ km$^2$). The large fraction of temperate ice gives rise to a large spatial extent. The spatial extent from the modified LGM anomaly runs results in the first Weichselian glaciation of the same order as during the Younger Dryas approximately 12 ka ago. The spatial extent of the Middle Weichselian glaciation is large, about $5.5 \times 10^6$ km$^2$, while at the LGM it is $6.5 \times 10^6$ km$^2$. This is probably due to the high volume of temperate, fast flowing ice.

The spatial extent of temperate ice follows the main features of the total ice volume curve, as does the basal melt curve. The only exception is in the modified LGM anomaly runs, with a peak in the basal melting at 25 ka ago. The basal melting is concentrated in the NCIS upstream and on the western marine margins of the ice sheet. The drainage from the temperate layer peaks in the PMIP forcing run at the Middle Weichselian (4.5 km$^3$/a) and at the LGM (5 km$^3$/a), and for the modified LGM anomaly runs, only during the LGM with a double-spike on the order of 80 km$^3$/a. The maximum height of the temperate layer varies constantly during the Weichselian. In the PMIP forcing runs, the largest peak occurs in the Middle Weichselian glaciation with a value of approximately 1500 m. In the modified LGM anomaly runs, there were four outstanding peaks with values of 1500 m or more at 90–80 ka, with the highest peaks at 57–55 ka, 35–25 ka and 12 ka ago.

The freshwater production curve has similarly timed peaks in all runs, but magnitudes vary. Ice-stream activity also varies among models.
Model validation

For complex interacting models, incorporating a full range of physical processes and parameters, it is difficult to say if the model configuration is factually correct (Oreskes, 2001). This is the complexity paradox: the truer the model, the more difficult it is to show that it is so.

Validation is the process of evaluating if a mathematical model of a physical event describes the actual physical event sufficiently well (Babuska et al., 2003). The validation of a model means that the model does not contain a detectable flaw and is thus internally consistent (Irabi et al., 2003).

Model verification in the context of Babuska et al. (2003) was done for SICOPOLIS within the European Ice Sheet Modelling INItiative, EISMINT (Huybrects and Payne, 1996; Payne et al., 2000). The purpose of EISMINT was the intercomparison of ice-sheet models, firstly in how well individual models simulate the basic geophysical variables (like ice sheet thickness, ice velocity and temperature) and secondly, how accurate and efficient are the numerical techniques applied in the participating models. The tests were conducted for a large set of simulations, comprising from fully described boundary conditions and fixed parameters and process models to application to real ice sheets (Greenland and Antarctica) with individual model formulation.

Detailed results are in Huybrechts and Payne (1996) and Payne et al. (2000). The model validation for the Greenland and Antarctica for SICOPOLIS was done within EISMINT, but also in Greve (1997b) and Greve et al. (1998).

Here, the validation of SICOPOLIS to the Eurasian glaciation is done by calibration, comparisons and confirmation in terms of matching observations.

The calibration is done by comparing the effect of sinusoidal temperature forcing and constant sea level forcing to the complex forcing with GCM data. This is done by testing the dependency of model variables in section 6.1.

To validate the model further, comparisons to other similar models for the Eurasian ice sheet and comparisons to geological records are done. The simulation results are compared to geological records that include ice sheet extent (from QUEEN data), topographic evolution (Kakkuri, 1987; Balling, 1980; Lambeck and Purcell, 2003), and ground temperature observations (Šafanda et al., 2004; Balling et al., 1990; Kukkonen and Šafanda, 1996; Kukkonen and Jøleht, 2003). A special attention is given to the areas of Norwegian Channel Ice Stream (NCIS) and Baltic Ice Streams. For palaeogeographical and geological reconstructions on these areas, the works of Sejrup et al. (2000), Sejrup et al. (2003), Houmark-Nielsen (2003) and Hou-
mark-Nielsen and Kjær (2003) are referred. The last is based on synchronised land, sea and glacier configurations.

After this validation procedure, suggestive conclusions about other variables in the model can be made. Still, a simulation is always fiction, never a fact. The simulation results for ice stream activity and freshwater fluxes are treated here as suggestive conclusions.

**Statistical analysis of simulation results**

The ice sheets evolve due to complex interaction of basal and ice sheet topography, ice flow and temperature. In existing ice sheets, it is sometimes impossible to discern the cause from the effect.

An important question is the dependency of model variables. The dependency between two or more variables can be tested in several ways. Statistical methods provide several tools to evaluate the role of random variations, to find connections and to estimate the mathematical form for the dependency between the variables. The cause and effect can not explicitly be proven with these methods.

Tests were conducted by exploiting the combined tools for variance analysis and regression analysis, namely covariance analysis. The aim of this analysis was to eliminate the effect of random variations. Covariance can also have negative values; positive covariance indicates that variables are increasing simultaneously, and negative covariance indicates that while the other variable increases, the other decreases. Covariance is thus a measure for linear dependency between two variables.

In our covariance analysis, the only covariance found was with the variable itself. In conclusion and not surprisingly, no clear linear dependencies were found. This is reasonable, as the physics of the ice sheet manifests in complexly interacting variables.

To test the intensity of the dependency between two variables, coefficient of correlation is used. This method demands the two variables to have a combined bivariate normal distribution, as it was in our case.

In Table 6, the significant correlation coefficient values for the PMIP forcing runs, the sinusoidal forcing runs and the cold ice simulations are listed. These were calculated with a 95% confidence interval from Pearson correlation, with the corrcoef-function of the standard MATLAB.

In this table, the ice volume and sea level have the highest correlation coefficients. These two interplay clearly in the results, as they should from a physical point of view. The ice-covered area and total volume have correlation coefficients close to 1. This depicts the growth of ice sheet as accumulating ice spreads and polythermal effects enhance this expansion. Strong negative correlation coefficients are between the ice-covered area and g-index. The warming climate affects the thin marginal areas, thus making the ice sheet less extensive. The results for sinusoidal forcing are a bit different, because parameters like sea level were constant and the run was in cold ice mode. The resulting main factors are the volume and area. The reasons for this are the constant sea level and lack of small-scale variability in climate forcing in the sinusoidal run and the missing polythermal ice effects in cold ice mode.
Table 6. The correlation coefficients for variables resulting from the PMIP forcing, the sinusoidal forcing and the cold ice mode simulations. Only the variables with significant correlation coefficient are listed. The symbols in the table are: $V_{fw}$ is freshwater volume, $V_g$ is ice sheet volume, $Z_{sle}$ is sea level, $A_{tb}$ is area covered with temperate base, $V_{temp}$ is the volume of temperate ice, $V_{bm}$ is volume of basal melt, $V_{tld}$ is drainage from temperate layer, $H_{max}$ is maximum ice sheet height, $A_{ib}$ is total ice-covered area, $g$-index is glaciation index, and $H_{tmax}$ is maximum height of temperate ice layer.

<table>
<thead>
<tr>
<th>Variables</th>
<th>PMIP forcing</th>
<th>Sinusoidal</th>
<th>Cold ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>$V_{fw}$, $V_g$</td>
<td>0.3342</td>
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<td>0.3327</td>
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<td>0.7798</td>
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<tr>
<td>$V_{fw}$, $V_{bm}$</td>
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</table>
In our simulations, the main ice volume in the Late Weichselian is concentrated in the Scandinavian and Barents area. The sea level minimum coincides with the ice sheet volume maximum. The maximum ice sheet volume is reached at 19–18 ka ago and there are small peaks at 35 and 27 ka ago. Nevertheless, the maximum volume does not coincide with the maximum extent. In all the model simulations, maximum extent is reached near the LGM and maximum volume about a thousand years afterwards. The ratio of areal extent to the total ice sheet volume is depicted in Fig. 23. The ratio is at high before LGM, as the land gets glaciated. As the ice sheet grows in volume, there is no growth in the area. Notable areal expansion happens at 23.3, 22 ka and 15.5 ka ago. These appear as peaks also in the ratio.

Figure 23. The ratio of ice sheet extent and ice sheet volume for the modified LGM anomaly results at 24–14 ka ago.
After the LGM, the dome in Finland starts to grow as the western section shrinks. The dome is largest around 19.5 ka ago. The centre of the dome shifts slowly towards the west, and at 18 ka ago, it has a centre over the Gulf of Bothnia. Comparing this to the different timings for the LGM by QUEEN, our model simulations captured very well this feature with both the PMIP forcing set-up (simulation # 2) and the modified LGM anomaly (simulation # 6). The latter simulations nicely reproduced the dynamics of different local LGMs, and showed coherent, repetitive advances at a millennial time scale.

The CLIMAP data for Last Glacial Maximum conditions (CLIMAP Project Members, 1976) estimated the ice-equivalent sea level for the combined ice sheets of Scandinavian and Barents area to be at least 20.0 meters and at the most 34.0 meters. The corresponding figures for the EWG, MGG and LGM resulting from our simulations are the following:

Notable is also the two EWG glaciations in the modified climate anomaly results. This is suggested in Sejrup et al. (2003) for southern Fennoscandia and North Sea region. This is exactly where the simulation produces the two glaciations.

Comparisons with other models

SICOPOLIS simulations for the Eurasian ice sheet can be compared with simulations by Zweck and Huybrechts (2003), Charbit et al. (2002), Arnold and Sharp (2002), and Siegert and Dowdeswell (2004).

In this study, the simulated glacial-maximum ice-sheet volumes shrink shortly after 18 ka ago in all model runs. Deglaciation proceeds in three stages, with stable periods at 17–15 ka ago and 13–12 ka ago. After 10 ka ago, the ice sheets are gone.

This simulated deglaciation can be compared to simulations done by Charbit et al. (2002) and Siegert et al. (1999). Compared to the results in Charbit et al. (2002), both the PMIP forcing and modified LGM anomaly simulations underestimate ice volume, by ca. $2 \times 10^6$ km$^3$, but show similarities in deglaciation features with their GRIP-driven simulations. The stable periods are more pronounced in our simulation results. This may be due to the late volume maximum in their results, at 15 ka ago. Their final disintegration takes place at 7 ka ago and compared to our runs, shows better agreement with observational data. This may be due to the bigger ice volume and the inclusion of AGCM results in the forcing for 15, 9 and 6 ka ago, contrary to the linear interpolation used in SICOPOLIS runs. To improve the simulation of the deglaciation chronology, GCM temperature and precipitation anomalies for 15 and 9 ka ago should be used as additional time-slices for pinning the climate interpolation in SICOPOLIS ice sheet studies.

Arnold and Sharp (2002) detected lobes developing, with a main lobe over southern Norway and Sweden. The lobate pattern in southern and eastern areas vary over time.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>90–80 ka</th>
<th>65–55 ka</th>
<th>LGM</th>
</tr>
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<tbody>
<tr>
<td>PMIP forcing</td>
<td>0.5 m</td>
<td>12 m</td>
<td>17.5 m</td>
</tr>
<tr>
<td>cold ice</td>
<td>0.5 m</td>
<td>12 m</td>
<td>18 m</td>
</tr>
<tr>
<td>Modified LGM anomaly</td>
<td>1.5 m</td>
<td>16 m</td>
<td>23 m</td>
</tr>
<tr>
<td>CLIMAP</td>
<td>–</td>
<td>–</td>
<td>20–34 m</td>
</tr>
</tbody>
</table>

Table 7. Ice-equivalent sea level resulting from PMIP forcing, cold ice mode and modified LGM anomaly simulations for the EWG, the MWG and the LGM.
During the LGM, a lobe develops over the Baltic areas, and spreads to the east in a few thousand years. In our simulation results, this is also the main feature, but our ice flow history in the Baltic area consists of moderate velocity ice advances. Arnold and Sharp obtained clear flow onset areas leading to stable fast flow in the Baltic area, while in the simulation results of this study, a more complex pattern appears with no constant fast flow in the area. Especially, the flow in our model is slower (the maximum is 300 m/a) than in the results by Arnold and Sharp (2002), (400–600 m/a). The basal conditions are also different as the cold based area in their runs extends to the Baltic Sea basin and deep south to the ice sheet margin at the LGM. In the PMIP forcing and modified LGM anomaly simulations, temperate basal conditions develop in the Baltic Ice Sheet marginal areas. As in their results, areas of moderate velocity develop independently from the forcing applied and in similar locations, in both simulation results.

Zweck and Huybrechts (2003) simulated the Northern Hemisphere ice sheets and used the PMIP-data and forced it with a glacial index derived from Vostok-GRIP climate record. The resulting time-evolution of total volume and ice-covered area has similar main features as in our simulation results. They do not get any extensive EWG, and MWG has a lower volume than LGM. However, the spatial extent at WMG is proportionally larger than the corresponding volume. In the Taymyr area, results by Zweck and Huybrechts (2003) show heavy glaciation. It seems that this result is a direct cause of the PMIP-forcing applied. The most notable difference between (Zweck and Huybrechts (2003)) simulation results and the ones discussed here is the early deglaciation of the Barents Sea area not seen in our results. This is due to the defined maximum water depth in the marine-extent model, which generates the retreat. It is an overestimation compared to explicit calving-law inclusion, as in Siegert et al, (2001). A notable characteristic for the deglaciation is that the St. Anna Trough marine margin retreat precedes the Bear Island Trough retreat in all the simulation results. In the results, St. Anna Trough remains an open channel, cutting deep into the ice sheet. The BSIS is the last ice sheet to disappear in results.

Siegert and Dowdeswell (2004) simulated the last 30 ka before present. Their model is constrained to the correct QUEEN ice sheet extent. Their modelled volume of the Eurasian Ice sheet has a maximum of $5.5 \times 10^6$ km$^3$ at 16–15 ka ago. After this, the warming climate melts the ice sheet with one gently sloping step at 14–13 ka ago. The deglaciation produces a peak in iceberg calving at 14.5 ka ago, with a mass loss of $2100 \times 10^6$ km$^3$/a. This is followed by another peak at 13 ka ago, with a value of 1800 km$^3$/a. The modelled ice sheet has a single main dome over Fennoscandia. Resulting fast ice flow paths at 15 ka ago follow the topographic troughs of the western marine margins, with velocities ranging from 400 to 800 m/a. The largest velocities are located at the NCIS and Bear Island Trough areas.

**Topographic evolution**
The topographic lithosphere evolution during the Weichselian glaciations is dominated by ice dome loading, resulting in lithosphere depression. From the rebound rates that aim towards an isostatic balance, the previous load can be deduced. In Fennoscandia, the modern depression of the geoid is on the order of 8 m, and Kakkuri (1987) estimated that isostatic balance would be achieved in 7000–12000 years. Estimates for the total land uplift in the SIS dome area, Gulf of Bothnia, range from 700 to 800 meters (Balling, 1980). The most up-to-date results for the Fennoscandian rebound can be found in Lambeck...
and Purcell (2003), who predict the rate of relative sea level change due to glacial isostasy to be of order of 8.75 mm/a over northern Gulf of Bothnia.

*Resulting topography in PMIP forcing simulations*

The PMIP forcing run results also indicate gentle lowering and rebounding of the lithosphere in the North Sea area, where ice sheets sporadically reach. Coastal areas are subjected to sea level changes and are thus either below or above water. The British Isles and Scandinavia have relatively constant modern outlines, in contrast to Finland and to some extent the Baltic lowlands, which experienced marked lithosphere depression caused by the glaciations. For example, the MWG causes southern Finland to warp, forming in the southern Karelia around Lake Ladoga. This depression was largest at 55 ka ago, but it was uplifted by 51 ka ago.

The LGM ice sheet causes depression of the crust in Finland and Karelia, leaving only parts of western Finland above sea level at 17 ka ago. The depression is of the order of a few hundred meters. The final stage of uplift in the PMIP forcing simulations left Lake Ladoga and part of Lake Saimaa below sea level, but not connected to the sea. Islands like Åland and Gotland have not risen above sea level at this stage.

In the Barents Sea area, Franz Josef Land experiences heavy glaciations in the Weichselian and is thus depressed. At the MWG, it lays below sea level with a local dome above. After the MWG and LGM, areas around Franz Josef Land stay below sea level, the depression being almost 1 km. Also, the locality of Bear Island Trough grows deeper. At the end of the run, Franz Josef Land lays below sea level.

In the PMIP forcing results, the developing topographic depressions form a path to the marine deep areas. This is the North Sea in the western areas and the Polar Ocean in the North. The deepest troughs are the NCIS area, the Bear Island Trough and the Franz Victoria Trough. The Baltic Sea depression is filled with water during the Weichselian. At some stages, it is a proglacial lake, like during the LGM. At 11 ka ago, the lake expands and a marine connection via the Danish Straits is opened as late as 2 ka ago.

Today, Norwegian Channel is approximately 250 meters deep, and the onset area of the NCIS, the Skagerrak Trough, is 700 meters deep. In the PMIP forcing results, the NCIS area topography stays relatively constant through the Weichselian, with the main depth of 400 meters forming in the NCIS onset area around 24 ka ago.

*Topographic evolution in modified LGM anomaly results*

In general, the modified LGM anomaly simulations produce glaciations that have larger ice volumes in the western regions than the PMIP forcing runs produce. The evolving ice sheet resulting from the modified LGM anomaly forcing depressed southern parts of Sweden and Finland below sea level as we approach the LGM. Also, marine Baltic areas are thus depressed. The central Barents Sea area is depressed to a depth of around 1 km. At the LGM, these areas get deeper. Southern Sweden and Finland have warped to a depth of -300 meters. This depression is greatest at 18 ka ago, with a bathymetric depth of ~800 m at the Gulf of Bothnia. This depression helps the ice sheet to deglaciate quickly as sea level rises. The Younger Dryas cooling affects a heavily reduced Fennoscandian ice sheet, clearly not correctly simulated. Central and eastern Barents Sea areas stay heavily glaciated until the end of Weichselian. The Danish Straits, Skagerrak and Kattegat, the NCIS onset areas, are approximately 600 meters deep during glaciation. The Norwegian Channel is depressed to ~900 me-
ters at the LGM. The Norwegian Channel is at its widest at around 18 ka ago, with a central depth of 1 km. Nowadays, the Norwegian Channel slopes gently towards the continental margin, from a depth of ~200 meters to a depth of ~400 meters.

In the modified LGM anomaly results, the central Barents Sea area is heavily depressed from the LGM to the Holocene. The Bear Island Trough and Franz Victoria Trough are the causeways to marine depths, as is the St. Anna Trough but not to such an extent as it is shallower. As the deglaciation starts with the onset of marine transgression, the southern SIS margin becomes marine 17.7 ka ago. At 17.1 ka ago, the SIS margin is marine up to Lake Ladoga. The marine area expands until 15.5 ka ago, when the Danish straits are closed by rebound. An ice lake exists in the Baltic Sea area until 14.3 ka ago, when the marine connection opens again, this time through southern Sweden.

Comparison to another model
Howell et al. (2001) modelled the influence of glacial isostacy in the Barents Sea in Late Weichselian times. In their modelling results, the isostatic depression is centred on Svalbard, northern Scandinavia and to Novaya Zemlya at 30 ka ago. From this point on, the depression propagates to the Barents Sea centre (27 ka ago), with a depression of ~500 m. At 24 ka ago, there are 500 m deep depressions forming in northern Scandinavia and the St. Anna Trough. Over time, these depressions propagate and at 21 ka ago, the Central Bank area is depressed to ~500 m. As deglaciation starts, the isostatic loading still keeps the area depressed and at 18 ka ago, the whole central Barents Sea is depressed to a maximum of ~800 m. The temporal evolution of the crustal responses from 33 to 18 ka ago for Svalbard, Bear Island, Bear Island Trough, Central Bank and Central Deep areas differ from each other. Svalbard is depressed to a depth of over 300 meters by 28 ka ago, and remains at that level. Bear Island experiences its main depression from 26 ka ago and stays depressed to a depth of nearly 800 meters. The Bear Island Trough is first depressed to nearly 400 meters, but rebounds 28 ka ago and after that, it is depressed again to ~1200 meters. The Central Bank history is similar to the Bear Island Trough, first a small depression of 150 m and then at 28 ka ago, a main depression to ~1200 m. The Central Deep is first depressed to ~400 m and at 29 ka ago to ~1200 m.

The PMIP forcing and modified climate anomaly results for Barents Sea are represented in tables Tab. 8 and Tab. 9. Both simulations compare well to the results of Howell et al. (2000). The differences are due to different ice sheet loading, which especially in the modified LGM anomaly results is too much, because of an overly thick Barents Sea Ice Sheet.

<table>
<thead>
<tr>
<th></th>
<th>30 ka ago</th>
<th>27 ka ago</th>
<th>24 ka ago</th>
<th>21 ka ago</th>
<th>18 ka ago</th>
</tr>
</thead>
<tbody>
<tr>
<td>Svalbard</td>
<td>~100 m</td>
<td>~200 m</td>
<td>~10 m</td>
<td>~10 m</td>
<td>~10 m</td>
</tr>
<tr>
<td>Bear Island</td>
<td>~200 m</td>
<td>~300 m</td>
<td>~100 m</td>
<td>~100 m</td>
<td>~100 m</td>
</tr>
<tr>
<td>Bear Island Trough</td>
<td>~400 m</td>
<td>~500 m</td>
<td>~500 m</td>
<td>~1000 m</td>
<td>~1000 m</td>
</tr>
<tr>
<td>Central Bank</td>
<td>~400 m</td>
<td>~500 m</td>
<td>~600 m</td>
<td>~1000 m</td>
<td>~1000 m</td>
</tr>
<tr>
<td>Central Deep</td>
<td>~200 m</td>
<td>~300 m</td>
<td>~500 m</td>
<td>~500 m</td>
<td>~1000 m</td>
</tr>
</tbody>
</table>
Proglacial lakes developed on several occasions in the vicinity of the Fennoscandian Ice Sheet. In the modified LGM anomaly results, a proglacial lake resides in the Kattegat area at 38 to 37 ka ago, and again at 31 ka ago. The sea invades the area at 30 ka ago, to withdraw again and leave a proglacial lake in the Kattegat area, from 29 to 27 ka ago. After this, the ice sheet advances and only after the LGM 18.1 ka ago, the sea again reclaims the Norwegian Channel upstream areas. Seawater reaches Skagerrak at 17.9 ka ago, the southern tip of Sweden at 17.8, and the Baltic Sea areas at 17.7 ka ago. The sea connection from the Baltic Sea breaks at 15.5 ka ago as the ice sheet readvanced, but the sea invades again at 14.2 ka ago.

This proglacial lake formation after the LGM is depicted in Figures 24–28, with contemporary basal temperatures for the modified LGM anomaly simulation results. The detailed history of the NCIS and adjoining areas are discussed in connection with the fast flow events.

Table 9. The evolving isostatic depression in Barents Sea from modified LGM anomaly simulations.

<table>
<thead>
<tr>
<th>Location</th>
<th>30 ka ago</th>
<th>27 ka ago</th>
<th>24 ka ago</th>
<th>21 ka ago</th>
<th>18 ka ago</th>
</tr>
</thead>
<tbody>
<tr>
<td>Svalbard</td>
<td>−150 m</td>
<td>−200 m</td>
<td>−20 m</td>
<td>−20 m</td>
<td>0 m</td>
</tr>
<tr>
<td>Bear Island</td>
<td>−300 m</td>
<td>−300 m</td>
<td>−50 m</td>
<td>−50 m</td>
<td>−30 m</td>
</tr>
<tr>
<td>Bear Island</td>
<td>−500 m</td>
<td>−500 m</td>
<td>−600 m</td>
<td>−800 m</td>
<td>−1000 m</td>
</tr>
<tr>
<td>Trough</td>
<td>−500 m</td>
<td>−500 m</td>
<td>−800 m</td>
<td>−1000 m</td>
<td>−1200 m</td>
</tr>
<tr>
<td>Central Bank</td>
<td>−300 m</td>
<td>−300 m</td>
<td>−500 m</td>
<td>−1000 m</td>
<td>−1200 m</td>
</tr>
<tr>
<td>Central Deep</td>
<td>−300 m</td>
<td>−300 m</td>
<td>−500 m</td>
<td>−1000 m</td>
<td>−1200 m</td>
</tr>
</tbody>
</table>

Figure 24. Topography of the base (left panel) and temperature (right panel, in °C) at 18.1 ka ago, from the modified LGM anomaly simulation results.
Figure 25. Topography of the base (left panel) and temperature (right panel, in °C) at 17.9 ka ago, from the modified LGM anomaly simulation results.

Figure 26. Topography of the base (left panel) and temperature (right panel, in °C) at 17.8 ka ago, from the modified LGM anomaly simulation results.

**Basal temperature evolution**

Fast, icestream-like flow takes place in regions with a temperate base and temperate ice above. Here, of concern is the temperature evolution of the forming ice sheet in general, and especially with the outlets to the marine margins of the ice sheets.

In simulations # 2 (PMIP forcing) and # 6 (modified LGM anomaly), temperate basal conditions form coherently during Weichselian glaciations. In both simulations, the MWG glaciation gives rise to a huge percentage of temperate ice at 62 ka ago with the PMIP forcing, and at 61,5 ka
ago with the modified LGM anomaly. Near the LGM, a maximum volume of temperate ice is present at 23 ka ago in the PMIP forcing simulations, and at 26 ka ago in the modified LGM climate runs. The MWG maximums are higher than at the LGM, with the PMIP forcing even twice the LGM value. The Barents Sea is not glaciated at the MWG but glaciates at the LGM. Otherwise, the spatial extent is very similar at the MWG and at the LGM. The difference in volume in the two glaciations arises from the height of the temperate layer. These maximums in volume of temperate ice precede the maximums in total volume of ice, and are reached at the beginning of a glacial event. The reason for this is that a thin ice sheet temperature signal can prop-

Figure 27. Topography of the base (left panel) and temperature (right panel, in °C) at 17.7 ka ago, from the modified LGM anomaly simulation results.

Figure 28. Topography of the base (left panel) and temperature (right panel, in °C) at 15.5 ka ago, from the modified LGM anomaly simulation results.
agate faster. Before the MWG, the temperature signal fluctuates strongly. In simulations #2 (PMIP forcing) and #6 (modified LGM anomaly), the western BSIS has a temperate base with temperate ice above.

**Basal temperatures in PMIP forcing results**

Simulation #2 (PMIP forcing) produces large temperate basal areas with temperate ice above in the marine Norwegian marginal areas during the MWG. These especially included the NCIS and Bear Island Trough. Temperate basal conditions with cold ice above prevail in the Estonian and Karelian areas. As the ice sheet decays, temperate basal areas are gone by 56 ka ago.

The small glaciations before the LGM still produce temperate basal conditions with temperate ice above in the western Norwegian marine marginal areas and in the Frantz Victoria Trough. As the LGM approaches, glaciation is intensified by a cooling climate. The forming ice sheets in Scandinavia and the Barents Sea maintain temperate basal conditions. Both the St. Anna Trough and Franz Victoria Trough have a temperate base with temperate ice above, giving rise to high velocities. The NCIS onset area is temperate-based with temperate ice above, first at 28 to 25 ka ago and after that from 23 to –19 ka ago. The temperate base appears in the Gulf of Riga at 22 ka ago, and propagates to Estonia and Karelia. The temperate base in the area disappeared at 17 ka ago.

**Basal temperatures in modified LGM anomaly results**

In the modified LGM anomaly simulations, the western ice sheet marginal areas maintained temperatures above –1 °C, especially in the NCIS area and temporally at the Baltic Sea Basin. Similar basal conditions prevail in the Barents Sea area at Storfjorden and at the deepening of Storbanken, leading to the Bear Island Trough.

The ice sheet at Franz Josef Land remains cold throughout the LGM and warms only in the Holocene.

In the modified LGM anomaly results for the southwest sector of the SIS and also on the Norwegian coast, there was a large temperate base and temperate ice area. As a fast flow in the NCIS area sets on, for example at 22 ka ago, the temperate base area expands to the Kattegat area on the Swedish coast, around 23 ka ago.

Under the main ice dome, the highest temperatures prevail around the LGM, 21,8 ka ago. The basal temperature is then –1,5 °C in areas surrounding the Gulf of Bothnia.

The main areas for drainage from the resulting temperate layers are the NCIS onset area, the western marine margin and the Franz Victoria Trough. In these areas, the drainage is in the order of 0.3 km³/a in maximum.

The BIS has a temperate base with temperate ice above for the largest part of its existence. Basal conditions for simulation #6 at 26.7, 22.0 and 20.4 ka ago are presented in Fig. 29.

After the LGM, temperate basal conditions can be seen in the Baltic area. Contrary to the NCIS and marine ice stream temperate base areas, these Baltic temperate-based areas do not produce much basal melt water. The reason for this is that in the deglaciation phase, the ice sheet in the area is very thin and temperate ice melts from the surface. This can be seen in the volume fluxes in this area. Although our model does not keep track of the areas of freshwater production, it can be deduced from the volume fluxes. The relatively rapid draw-down allows the ice sheet to quickly recover from the volume loss of fast flow events in the NCIS.

**Basal melting**

High basal melting rates are centred on the NCIS upstream areas and on the western
Figure 29. Basal conditions before, during and after HE2 for simulation #6 (Modified LGM anomaly). The left panels depict the temperate base with temperate ice above (darker shading) and the temperate base with cold ice above (lighter shading). The right panels show the basal temperate, with the darkest values just below zero. The exact dates for these stages are 26.7, 22.0 and 20.4 ka ago. Reproduced from Forsström and Greve, 2004.

marine margins of the ice sheet. The basal melting rates are especially high around 25–22 ka ago. As the NCIS is shutting down during deglaciation, a high basal melt area exists in the south-west Norway area around 18–17 ka ago. After this, the basal melting is highest in the Bear Island Trough marginal area and at the Franz Victoria Trough. In the Baltic ice stream areas, only small and sporadic areas of basal melt formed after 18 ka, after the local maximum configuration.

Johnson and Fastook (2002) studied the interplay of basal water and sliding veloc-
ity. In their model results, water depth is greatest in regions of known palaeo ice streams, and the dynamic water model provides a positive feedback to ice stream formation. Interestingly, in their model results, the southern margin of the LGM Laurentide ice sheet is characterised by the presence of basal water. As the southern margin terminates on land, it is in this aspect similar to the terrestrial Baltic ice streams.

In our model results, in run # 6 (modified LGM anomaly), there is only modest basal melting around LGM at the southeastern Fennoscandian ice sheet margin. In comparison, the melt fraction is a 100th to a 1000th part of the melt rates in the western section of the ice sheet. The western, marine margin is characterised by high basal melting rates in well-known palaeo ice stream locations. So, in our runs it is not possible to see a wet, terrestrial Baltic ice stream, but wet marine western ice streams are readily apparent.

The difference to the LGM Laurentide southern margin arises from the latitude: the Laurentidian south margin has high temperature gradient and no permafrost. The Fennoscandian terrestrial margin was located in tundra conditions with permafrost.

**Evolution of temperate basal conditions**

The ratio of the area covered by temperate ice can be compared to the total ice-covered area ($A_{\text{temp}}/A_{\text{total}}$) to contrast the onset of fast flow with the spatial extent and basal temperature conditions. The percentage of temperate basal area compared to the total ice covered area is summed in Table 10.

The ratio is below 30 % in all runs except the modified LGM anomaly simulations, where the relation is 45 % at the LGM. The ratio is largest in the LGM.

This ratio, as well as the volume of temperate ice, is very suggestive for the sensitive behaviour of the ice sheet, especially seen in the modified LGM sheet, especially seen in the modified LGM anomaly simulations (simulation # 6).

In comparison, Marshall and Clarke (2002) reported that within 5 000 years of ice sheet inception, the warm-based area reaches 35 %, and remains almost stable for 70 ka. At the LGM, the fraction ranges from 8 to 42 %. The basal temperate area is largest at 12 ka BP, as high as 70 %. Forming new ice is usually cold-based and the base warms with time as insulation from the surface increases and the base is heated by geothermal heat flux and deformational heating.

The simulation results of this study differ due to the thinner Fennoscandian Ice Sheet compared to the Laurentidian Ice Sheet that maintains faster ice flow-through times. In our modified LGM anomaly simulations, the thicker Barents Ice Sheet is temperate-based in the western section.

Boulton et al. (2001) modelled the thermodynamic evolution of the European ice sheets during the Weichselian. In their results, the relation of area covered by temperate ice to the total ice covered area is about 28 % at the EWG, 22 % at the MWG and 50 % at the LGM. Also, these results are similar to the PMIP forcing and modified LGM anomaly simulation results.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>90–80 ka</th>
<th>65–55 ka</th>
<th>LGM</th>
</tr>
</thead>
<tbody>
<tr>
<td>#2</td>
<td></td>
<td>0.10–0.15</td>
<td>0.20–0.22</td>
</tr>
<tr>
<td>#3</td>
<td></td>
<td>0.12–0.14</td>
<td>0.22</td>
</tr>
<tr>
<td>#5</td>
<td></td>
<td>0.25–0.27</td>
<td></td>
</tr>
<tr>
<td>#6</td>
<td>0.30</td>
<td>0.23</td>
<td>0.40–0.45</td>
</tr>
</tbody>
</table>
Figure 30. The fraction of temperate base compared to the total ice-covered area for the modified LGM anomalies. The horizontal axis is time in ka, and the vertical axis is the ratio.

**Cold-based areas**

When the basal temperature of the ice sheet is below the pressure-melting temperature, the ice sheet is frozen to the bed and defined as *cold-based*.

In the PMIP forcing simulation results, the south-eastern sector of the ice sheet as well as in the Barents Sea area, remained cold-based throughout the Weichselian. Cold basal conditions also prevailed in the Scandic mountains, in the Kola Peninsula and in the central Baltic Sea area.

In the modified LGM anomaly simulation results, the cold-based areas are located in the Scandic Mountains ice divide area and the Kola Peninsula, in east Karelia and the eastern marginal areas of the ice sheet, in western Svalbard, Franz Josef Land and in central Novaya Zemlya. These cold-based areas are reduced with evolving times, resulting in separate cold-based areas in the Scandic mountains and in the Kola Peninsula.

**Comparisons to ground temperature observations**

The ground areas below thick ice sheets experience warmer temperatures than the areas subjected to the chilly LGM aerial temperatures. In the PMIP atmospheric temperature anomalies, northeast Poland experiences an early average LGM temperature anomaly in the order of $-10^\circ$C. The corresponding value for the modified LGM anomaly temperatures is approximately $-20^\circ$C. In both simulation results, the area is situated on the southern margin of the modelled ice sheet. The ice sheet occupies the area for a short time period around the LGM, with a thin ice cover of a few hundred meters. This is in accordance with Šafanda et al., (2004), who reported a mean ground temperature anomaly of $-18^\circ$C in the glacial period.

In Siljan, western Sweden, Balling et al., (1990) reported a lack of significant vertical variation in the ground temperature
Figure 31. The modelled lithosphere temperatures for Karelia, Outokumpu, Poland and Siljan areas mentioned in the text. The time is from 24.1 ka to 14.1 ka, and the time step 1 ka. The left panel shows the evolution of ground temperatures at depth of –500 meters, and the right panel at the lithosphere surface.
gradient. This area is in close proximity to the SIS ice divide. In PMIP forcing and modified LGM anomaly simulations, the Siljan area is either cold-based or maintains a maximum basal ice sheet temperature of \(-2\) to \(-4\) °C.

In a study of the Outokumpu area, eastern Finland, Kukkonen and Šafanda (1996) concluded from heat transfer modelling, ground surface temperature values of \(-1\) to \(-2\) °C during the Weichselian. Both in the PMIP forcing and modified LGM anomaly simulation results, this area is covered by an ice sheet in glacial times, with basal ice sheet temperatures in the range of \(-3.5\) to \(-1.5\) °C.

These results can be compared to SICOPOLIS results for lithosphere temperatures at the surface and at \(-500\) meters deep, represented in Fig. 31. These are in broad agreement with the observations mentioned above. The lithosphere surface temperatures, and thus also deeper ground temperatures, vary more in areas with sporadic ice cover.

The SICOPOLIS omits many relevant processes here; no permafrost and ground water hydrology model is included. A more detailed study of the lithosphere temperature evolution with SICOPOLIS to compare against observations and models for temperature change would provide a good validation test for the model.

Mass flux

Ice sheets are constantly adjusting to changes in climate conditions, such as precipitation and temperature. This adjustment is seen in changes in volume, spatial extent, velocities and mass fluxes. These affect the temperature dynamics. The time scale for these adjustments depends on the size of the ice sheet, being larger for a large ice mass. Here, interesting is the fast flow features and age of ice, resulting from the changes in climate conditions.

If the flow of ice adjusts quickly to the change in accumulation rate, the amplitude of the response remains small (approximately 1 m) (Van der Veen, 1999). If the flow of ice reacts very slowly to the perturbation in accumulation, the amplitude is considerably larger. This is because the increased accumulation is not transported rapidly towards the ice sheet margin.

Fast flow

In the simulation results, the largest velocities prevail in the areas that have the steepest topographic gradient.

With the PMIP forcing, the simulation results indicate the highest velocities developing at 61 ka ago (760 m/a) and at 21 ka ago (740 m/a) in the Lofoten area ice sheet margin. Other areas maintaining fast flow are the NCIS area, the Jaeren area, Franz Josef Trough and St. Anna Trough.

In the modified LGM anomaly results, high velocities prevail at the EWG in the Lofoten area at the ice sheet margin. The velocities are not high, around 600 m/a. At the MWG, the highest velocities are on the order of 1000 m/a. Areas maintaining high velocities develop at the NCIS onset area, at the Voring area and at the Bear Island Trough. In a later phase, a high velocity area develops on the BIS western margin and at the Franz Victoria Trough.

Before the LGM, the growing ice sheet has the highest velocities on the western marine marginal areas. Around the LGM, highest velocities develop at the sources of the NCIS, around 1500 m/a. This fast flow feature is very stable in the simulation results and has a high elevation and topographic gradient. In terms of ice stream type, this would be an isbræ. In conclusion, our simulations of fast flow features suggest that the NCIS was an isbræ in the LGM.

In the Baltic area, velocities do not exceed 300 m/a during any simulation time. This might also be an artefact arising from the fact that the results were visualised only for every 100 years. The ice sheet elevation
and topographic gradient stays low in the Baltic Sea area. Temperate basal conditions form occasionally at the Baltic margin, and in these time periods, velocities in the order of 200–300 m/a are maintained. In terms of ice stream type, these Baltic modest flow areas display velocities below those suggested for a typical ice stream (after the definition by Stokes and Clark, (2003)). These areas of moderate flow correlate to the known Baltic Ice Stream activity areas.

The velocity profile for a fast flow area in the Baltic at 19.5 ka ago is depicted in Fig. 32.

In the marine areas, ice velocities on the order of 600–800 m/a are seen at the western margin of the Fennoscandian and Barents Sea Ice Sheets, in the Storfjorden and Trondelag areas, at the Bear Island Trough and on the Svalbard west coast.

As the model does not include any specific ice stream model, the resulting fast flow features are only cautious approximations for flow. Anyhow, the NCIS appears as a main, steady, fast flow area for the Fennoscandian Ice Sheet. The only other fast flow area that compares to the NCIS is the Bear Island Trough Ice Stream, but only temporally.

**History of the NCIS area**

During the Weichselian interstadials, several lines of study indicate that Atlantic and subarctic Atlantic waters extended into northern Denmark, the Swedish west coast and deep into the Kattegat (Shackleton, 1987; Knudsen, 1994; Houmark-Nielsen, 1999; Lambeck et al., 2002). The NCIS area formed a narrow fjord more than 500 km long and 200–700 m deep (Rise et al., 1997).
The reconstructed ice flow data indicate that the Skagerrak Trough was a confluence zone for ice flowing in from several directions (Sejrup et al., 2003). This trough leads to the Norwegian Channel, and the North Sea Fan lies on the continental margin at the mouth of the Norwegian Channel. Sejrup et al. (2003) report increased sedimentation rates after 29 ka ago, with a peak ending close to 16 ka ago. Also, during the last 18 ka, the sediment-laden meltwater has been deflected northwards. For the Weichselian, Sejrup et al. (2000) suggested that the NCIS was active at 85 to 70 ka BP, at 50 to 36 ka BP, at 29 to 23 ka ago and from 19 to 16 ka ago. In a new study (Sejrup et al., 2003), the NCIS is thought to reach the ice margin in periods between 29 to 16 ka ago.

A palaeogeographical reconstruction by Houmark-Nielsen and Kjær (2003), based on synchronised land, sea and glacier configurations from the lithostratigraphy of tills and intertill sediments, depicts the NCIS onset area from 43 to 26 ka ago. At 43 to 36 ka ago, the ice sheets stayed on the Norwegian highlands, and the sea invaded up to the Baltic Basin. From 36 to 35 ka ago, an ice advance from southern Sweden invade into eastern Denmark. This ice margin retreats to Sweden from 35 to 32 ka ago. After this, the ice sheet advances from Norway to the Norwegian Channel and a proglacial lake forms on the southern margin. The ice sheet advances until 28 ka ago, when an ice flow event from southern Sweden and the Baltic Sea reaches eastern Denmark. At this stage, the main ice flow from Norway is channelled to the NCIS. At 24 to 22 ka ago, the ice lake disappears, and the ice sheet grows to the maximum position. At 22 to 20 ka ago, the ice margin retreats and the NCIS area is marine up to Norwegian Coast. Starting from 20 ka ago, there is an ice influx event from the Baltic Sea areas. The so-called Young Baltic Ice advance was from 18 to 16 ka ago, with rapidly flowing ice from the Baltic.

The Kattegat Ice Stream has been studied by Houmark-Nielsen (2003), who stated that on several occasions, Pleistocene glaciers penetrated deep into the Danish and German lowlands. At 31 ka ago there was an ice lake, the Kattegat Ice Lake, in front of the southern margin. Ice flow was from the Norwegian mountains and from southern Sweden towards this ice lake. Ice flow from the Norwegian mountains continued, but 29 ka ago, the ice lake was gone and ice covered northern parts of Denmark. The LGM glaciation dynamics in the area suggest a close relationship with the NCIS.

In the PMIP forcing simulation results, the NCIS is active from 80 to 70 ka ago (peak velocity in the area 250 m/a, from 40 to 35 ka ago with similar peak velocities, from 28 to 22 ka ago and from 20 to 18 ka ago. The deglaciation is too fast with both forcings and thus leads to an early shutting down of the NCIS.

In the modified LGM anomaly simulation results, the NCIS is sporadically active at 84 to 70 ka ago, 65 to 58 ka ago and from 44 to 35 ka ago. As the climate cools towards the LGM, the NCIS reaches a marginal position at 28 ka ago, and maintains high velocities until 26 ka ago. For a few ka before the LGM, the main flow direction in the Skagerrak Trough is from north to south. Over time, the flow directions in the area become more congruent, and a strong flow from east to west is evident. For example, at 24.5 ka ago, the dominant flow is from east to west, with velocities in the order of 900 m/a. At 21.4 ka ago, the highest ice velocities in the area are from south to north, with a velocity of 900 m/a. This is caused by influx of ice from Sweden and Baltic Sea. As this flow ceases, flow to the west and south is enhanced. The flow from east to west is steady from 21.5 to 19.1 ka ago. A sporadic flow event with a velocity of 1200 m/a is seen from 19.6 to 19.1 ka ago. The NCIS is at a continental margin posi-
tion from 22.1 to 20.9 and from 19.8 to 18.5 ka ago. The ice sheet extends to the Norwegian Channel from 23.4 to 18 ka ago. After this, the areas adjacent to the Norwegian Channel maintain ice flow towards the Norwegian Channel area. A dramatic change happens in the NCIS activity at 17.8 ka ago, as sea level rises with the melting ice sheets. At this date, the main flow from the Norwegian mountains shifts to the western marine coast. A fast flow area activates there, with velocities of 1000 m/a, but the flow is channelled to the mouth of the Norwegian Channel. After this, the NCIS velocities are reduced below 300 m/a. The ice sheet readvance starts at 15.3 ka ago, but the NCIS is not reactivated. The main ice flux in the area has shifted to the Norwegian west coast areas.

These results are in general agreement with the results of Sejrup et al. (2003), Houmark-Nielsen and Kjær (2003) and Houmark-Nielsen (2003). The NCIS onset in the simulation results coincides with the activity times reported. The NCIS cessation is a little early due to the overvigorous simulated deglaciation, but high ice flux still continues to the Norwegian Channel from the Norwegian mountains. A sudden shift of flow, with the main onset area shifting from the Skagerrak to the Kattegat, coincides with the uniting of the two temperate basal areas and the simultaneous sea invasion. The proglacial lake/sea history discussed in an earlier section is, in a broad context, similar to Houmark-Nielsen (2003).

Also in the modified LGM anomaly simulations, the Kattegat area has a temperate base with temperate ice above from 22.9 ka ago. The Kattegat area unites with the NCIS temperate area at 22 ka ago and the fast flow area reaches up to Kattegat.

In the results of Boulton et al., (2003), the NCIS surface velocity vectors fan strongly at the terminal. The strongest flow occurs along the axis of the Channel. Ice flow from Norway to Denmark is seen. These results are very similar to SICOPOLIS simulation results.

**History of Baltic Ice Streams**

In the PMIP forcing simulation results, the Baltic Sea has only a thin ice sheet cover. The main domes develop over the Scandinavian Mountains and Finland. No ice streams (referring to the definition of Paterson (1994)) develop in the southern Baltic area. In SW Russian Plain, Karelia and the Gulf of Finland, velocities around 100 m/a are detected at 20 ka ago. This moderate velocity zone propagates towards the centre of the eastern dome.

In the modified LGM anomaly simulation results, the Baltic basal conditions differ considerably from the western marine marginal conditions. As the LGM is nearing at 24 ka ago, the Baltic Basin area is cold-based. Only around the Gulf of Riga and in the Kattegat exists a small area with a temperate base but cold ice above. Similar small patches of temperate base with cold ice above develop near the Isle of Gotland (22.5 ka ago) and around the Gulf of Gdansk (22.4 ka ago). These areas expand and fluctuate a little, until 20.6 ka ago when there existed wide zones of temperate base with cold ice above in pro-marginal areas in the Latvian, Lithuanian and Estonian areas, as well as in the southern marine areas surrounding the tip of Sweden. At 20.2 ka ago, a temperate base with temperate ice above appears in the SW Russian Plain. The ice sheet expands and the ice above the base cools. At 19.4 ka ago, temperate areas form around the Gulf of Riga, southern Estonia, and on eastern borders of Latvia and Lithuania. These local temperate areas unite at 19.1 ka ago and form a wide front of temperate ice over the Baltic States and surrounding areas. At 18.5 ka ago, the temperate area begins to shrink and retreats towards the north. At 17.8 ka ago, the area still maintains a temperate
base but the ice above is cold. As the ice sheet decays, the margin facilitates conditions with temperate base and cold ice above on several changing locations over the Baltic Sea area. The Kattegat area joins the NCIS temperate area at 22 ka ago. The Baltic Basin around Sweden maintains temperate basal conditions with cold ice above throughout the LGM and deglaciation.

Towards the LGM, around the Baltic Sea, the ice first flows slowly towards the Baltic Central Deeps, with velocities of a few tens of meters per year. Higher velocity spots form around the Gulf of Gdansk and the Isle of Gotland at 22,5 ka ago, and a velocity of 80 m/a prevails for 200 years. After this, similar moderate velocity spots are forming in pro-marginal areas in the Latvian, Lithuanian and Estonian areas. Spots of increased velocity form and migrate into the area in question and spread towards the centre of the ice sheet over time. As the ice sheet disintegrates, these areas shift northwards. At 17,4 ka ago, one moderate velocity area is centred on the Isle of Åland and another over central and southern Finland. The maximal velocity of nearly 300 m/a in the Baltic area is reached at 19,5 ka ago in SW Russian Plain.

In southern Scandinavia, the ice sheet advances from Sweden south and west at 23 ka ago, but only very slowly (10–20 m/a). At 22,3 ka ago, velocities increase to moderate and ice starts to flow towards North Germany and Danish mainland areas. The maximum extent is very short, as the ice margin starts to retreat at 21,7 ka ago. A readvance to Denmark is seen at 15,5 ka ago.

In the simulation results, a velocity increase precedes temperate basal conditions. The localised flow enhancement due to a positive mass balance in the area creates a temperate base caused by frictional heating. The flow is temperature dependent and heat is advected by flow, demonstrated here by the expansion of temperate basal conditions and moderate flow areas in the pro-marginal areas. The Baltic expansion of the ice sheet is in our simulation results and facilitated by local, moderate velocity ice advances, which form lobes at the terrestrial advance. The modified LGM anomaly simulation results agree with Stokes and Clark (2001) who stated that lobate structures can be the result of slow sheet flow. These moderate lobate structures form on the locations hypothesised for the Baltic ice streams: the Danish Sublobe, the Oder sublobe, the Wisla sublobe and the Riga sublobe. The hypothesised Novgorod (SW Russian Plain) and Finnish Lake District ice stream locations maintain moderate ice sheet flow forming lobes. The only exception is the fast flow event in the SW Russian Plain mentioned earlier. This results in ice sheet expansion to east.

This is in conflict with Holmlund and Fastook (1993), Payne and Baldwin (1999), Boulton et al. (2001) and Arnold and Sharp (2002). Holmlund and Fastook (1993) used a numerical finite element model to study the Scandinavian ice sheets and postulated that in order to fit the geological evidence of maximum LGM extent, high flow rates throughout the Baltic Basin are necessary. Payne and Baldwin (1999) applied a thermomechanical ice sheet model to study the Scandinavian Ice Sheet and palaeo ice streams. Their model results indicated that small differences in ice flow trigger internal feedback mechanisms (creep flow) that lead to fast ice flow. Boulton et al. (2001) stated that the Baltic Sea lobes can result from either fast or slow flow, and they assumed that the first attention was the case. Boulton et al. (2003) simulate large ice stream in the Baltic Sea area. Streaming flow is seen as a balance flow for increasing downice flux in the deglaciation phase. The resulting Baltic ice streams are very wide in their simulation results (from 30 to 70 km), with velocities of 250 m/a. These broad ice
streams can reduce the ice sheet thickness by even 1 km. They suggest that the great width of these ice streams reflect a higher average internal temperature. Strong fanning towards the margin is also seen in their results.

Arnold and Sharp (2002) simulated the flow variability of the Scandinavian ice sheet and found areas of fast flow, for example, in the NCIS area and Baltic Basin. The maximal velocities are 600 m/a. These high velocities prevail at the marginal areas of the LGM, but in the NCIS area, Baltic Basin and over Finland, reach deep into ice sheet interior where the flow is slower (200 m/a).

These differences arise, in addition to differences in numerical models, from the climate forcing and resulting mass balance used. For example, in the PMIP forcing simulation results, the Fennoscandian ice sheet clearly deglaciates too fast. By setting the velocities higher in the Baltic Basin, the results might be improved to allow stronger ice inflow from the central dome areas.

Figure 33. Lobe formation event around the Gulf of Gdansk, at 22.5 and 22.4 ka ago.
Surge fans recognised in the area (Kleman et al., 1997) are related to enhanced ice flow, but in the decay stage of an ice sheet. Such enhanced ice flow is also seen in simulation results. These velocity fans consist of flow with velocities in the order of a few hundred meters per year. An example of a decay stage velocity fan is depicted in Fig. 34.

The fanning in results of Boulton et al. (2003) and the lobe formation in the simulation results of this study occur in same areas, and due to same reason: less viscous warm ice. This is also reason for the ice sheet sensitivity. As temperatures in the southern and western areas of the ice sheet a high compared to other areas, the ice warms and flows readily. This is also suggested in modified LGM anomaly results, as in the lobe formation the increase in velocity precedes temperate base formation.

**Age of ice**

The age of ice was studied for the PMIP forcing results starting from 70 ka ago to the present at a time step of thousand years and for the modified LGM anomaly results with a time step of 100 years. The maximum age of ice was set to 1000 a, and ages older than that were not accounted for.
In the PMIP forcing runs, the Barents Sea centre is not glaciated and thus the oldest ice can be found in the northern depths of the Barents Sea. The BSIS dome contains ice that is, at the oldest, some 500 years old throughout the LGM. In the source area for the NCIS, the ice reaches an age of 400 years. Also, in the Baltic Sea basin in front of Gotland Island, in the White Sea Basin and on Novaya Zemlya, the ice can survive to an age of 400 years.

In the modified LGM anomaly runs, the Barents Sea old ice accumulates in the central abyss, entering the onset area of Bear Island Trough. The ice can be more than 1000 years old. Old ice also exists in the depths south of Svalbard.

At the LGM, ice in the BIS ice dome can be 600–700 years old, as it does on Novaya Zemlya. Elsewhere, the ice is younger, depicting the relatively fast flow to areas where it melts.

In reality, old ice is more cracked and contains more impurities than younger ice (Bahr, 1995). This enhances the flow and calving for older ice. Old and temperate ice is the easiest to crack and calve. As no cracking model is included, the age of ice tells of ice flow-through times and gives an indication of the locations where old ice forms.

**Freshwater production**

Freshwater is produced in our simulation runs by surface melting and calving. The melting is of course larger if the climate is warmer or the precipitation in warm areas is higher. Freshwater production varies significantly over the glacial cycle with pronounced discharge events. These peaks coincide with climate-warming events, but not all climate events produce proportionally equivalent fluxes. For example, in simulation # 2 (PMIP forcing), the warming starting at 26 ka ago produced a water flux of 7340 km$^3$/a$^1$, while a larger warming event at 35 ka ago contribute only 1340 km$^3$/a$^1$ of water.

Peaks in freshwater production resulting from the PMIP forcing, cold ice and modified LGM anomaly runs are listed in Tab. 11. From these values, estimates for calving (after Dowdeswell et al., 2002) are also calculated. The units are km$^3$/a$^1$.

Table 11. Freshwater production (combined melting and calving) peak values, and the corresponding approximation for calving after Dowdeswell et al., 2002. The simulations compared are the PMIP forcing simulations, the modified LGM anomalies runs and the cold ice mode. The values are in units of km$^3$/a.

<table>
<thead>
<tr>
<th>Timing</th>
<th>PMIP forcing calving</th>
<th>Mod. LGM anomal. freshwater calving</th>
<th>Cold ice freshwater calving</th>
</tr>
</thead>
<tbody>
<tr>
<td>11</td>
<td>1500</td>
<td>3000</td>
<td>1200</td>
</tr>
<tr>
<td>14</td>
<td>2450</td>
<td>17 500</td>
<td>7000</td>
</tr>
<tr>
<td>17–18</td>
<td>3300</td>
<td>4050</td>
<td>1620</td>
</tr>
<tr>
<td>21</td>
<td>1380</td>
<td>7000</td>
<td>2800</td>
</tr>
<tr>
<td>23–26</td>
<td>7340</td>
<td>12600</td>
<td>5040</td>
</tr>
<tr>
<td>29</td>
<td>1420</td>
<td>2930</td>
<td>5040</td>
</tr>
<tr>
<td>31–32</td>
<td>4000</td>
<td>1600</td>
<td>3360</td>
</tr>
<tr>
<td>34–35</td>
<td>1340</td>
<td>14 500</td>
<td>5800</td>
</tr>
<tr>
<td>38–39</td>
<td>2080</td>
<td>5580</td>
<td>2230</td>
</tr>
<tr>
<td>43</td>
<td>8900</td>
<td>8900</td>
<td>3560</td>
</tr>
<tr>
<td>55–56</td>
<td>3500</td>
<td>14 000</td>
<td>3560</td>
</tr>
<tr>
<td>60–61</td>
<td>1450</td>
<td>15 000</td>
<td>6000</td>
</tr>
<tr>
<td>68–69</td>
<td>1450</td>
<td>22 000</td>
<td>4400</td>
</tr>
<tr>
<td>70</td>
<td>1750</td>
<td>9000</td>
<td>8800</td>
</tr>
</tbody>
</table>
The highest freshwater fluxes are produced with the modified LGM anomaly simulations. These are depicted in Fig. 35.

These freshwater peaks correspond to warming events. The highest peaks come from warming preceded by a strong cooling event (for example at 68–69 ka ago). The gradient in cooling is more determining than the duration of the cold event. It is even more important than the amount of warming, as can be seen from the aforementioned differences for the flux events at 23–26 and 34–35 ka ago.

The combined surface melting and calving effects dominate the total freshwater production. They contribute over 99% to the total water flux. These sources also tend to produce well-defined events, lasting about 500 years. Water produced by basal melt and drainage from temperate ice contribute about 0.2% and 0.1%, respectively, to the total flux and are clearly of minor importance to the total flux.

These sources also vary over time, but produce relatively broad peaks, lasting about 2000 years for basal melt and 1000 years for drainage from temperate ice. This is to be expected, as the basal ice must warm before it can flow fast and increase calving.

By assuming an equal contribution to sea level rise for both hemispheres, then the Eurasian ice sheet should contribute approximately 1/8 of the total sea level change.

For example, the chain of events with the modified LGM anomaly results around 21 ka ago is the following: the water production by basal melt reaches the local maximum at 21.6 ka ago, the temperate-layer drainage has a maximum at 21.7 ka ago and the freshwater flux from melting and calving at 21.3 ka ago. The temperature forcing reaches a minimum at 21.9 ka ago. The pulse in basal meltwater remains high for about 2210 years.
From Table 11, freshwater fluxes are present in all simulation results at 14, 21, 23–26, 29, 31–32 and 34–35 ka ago. In the modified LGM anomaly simulations and cold ice mode, main peaks can also be seen at 55–56, 60–61 and 68–69 ka ago. The last of these peak periods is the greatest.

In comparison to other simulation runs, Siegert et al. (2001) reported high iceberg calving rates during the late Weichselian at 15 ka BP. This is the decay peak of the main ice sheet.

In the following, both Sverdrups and km$^3$a$^{-1}$ are used for easier comparison to referred studies. For the sea level rise rates mentioned in Rohling et al. (2004), the sea level rise around H$_4$ (23–30 m) is the result of an input of 264.5 to 345 Sv in 1000 years. The corresponding values are for H$_5$ (24–35 m) 276–402.5 Sv and for H$_6$ (40–47 m) 460 to 540.5 Sv.

To compare the freshwater fluxes and sea level contribution from the simulation results of this study to the literature, Table 11 is converted to annual global sea level change caused by the flux, and to Sverdrups in Table 12.

Hemming (2004) estimated the Hudson Bay freshwater input to be a fast entry of one year, 0.1 m of sea level rise and 1 Sv of freshwater flux. A slow freshwater input would have duration of 500 years and cause a sea level rise of 10–20 m and 0.15–0.3 Sv of freshwater flux. Alley and MacAyeal (1994) estimated a 0.01 sea level rise with an entry lasting 200–300 years.

One should keep in mind that the ice sheet volumes were overestimated in simulation results. Considering that the values in Tab.12 are for combined melting and calving events for the whole Eurasian ice sheet that are dispersed and routed to several seas, the calculated values still show a potential for the Fennoscandian Ice Sheet to be an active component in the climatic interplay around the HEs.

### Comparison with IRD events

In the results, some peaks in the freshwater production could be correlated to HEs (H$_1$ at 17–18, H$_2$ at 24, H$_3$ at 31, H$_4$ at 39, H$_5$ at 43 and H$_6$ at 60–61 ka ago), and to D-Os (b at 21, c at 23, d at 25–26, e at 29). Some of the D-Os are missing (like a). As men-

### Table 12. Freshwater production (combined melting and calving) approximated as sea level rise and Sv (in one year) for PMIP forcing simulations, modified LGM anomaly simulations and cold ice mode simulations. Values are in meters and in Sverdrups.

<table>
<thead>
<tr>
<th>Timing</th>
<th>PMIP forcing sea level</th>
<th>Sv</th>
<th>Mod. LGM anom. sea level</th>
<th>Sv</th>
<th>Cold ice sea level</th>
<th>Sv</th>
</tr>
</thead>
<tbody>
<tr>
<td>11</td>
<td>0.05 m</td>
<td>0.6</td>
<td>0.10 m</td>
<td>1.2</td>
<td>0.04 m</td>
<td>0.5</td>
</tr>
<tr>
<td>14</td>
<td>0.08 m</td>
<td>0.9</td>
<td>0.56 m</td>
<td>6.4</td>
<td>0.25 m</td>
<td>2.9</td>
</tr>
<tr>
<td>17–18</td>
<td>0.10 m</td>
<td>1.2</td>
<td>0.13 m</td>
<td>1.5</td>
<td>0.13 m</td>
<td>1.5</td>
</tr>
<tr>
<td>21</td>
<td>0.04 m</td>
<td>0.5</td>
<td>0.22 m</td>
<td>2.5</td>
<td>0.21 m</td>
<td>2.4</td>
</tr>
<tr>
<td>23–26</td>
<td>0.23 m</td>
<td>2.7</td>
<td>0.40 m</td>
<td>4.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>29</td>
<td>0.05 m</td>
<td>0.6</td>
<td>0.24 m</td>
<td>2.8</td>
<td>0.16 m</td>
<td>1.8</td>
</tr>
<tr>
<td>31–32</td>
<td>0.13 m</td>
<td>1.5</td>
<td>0.27 m</td>
<td>3.1</td>
<td>0.18 m</td>
<td>2.1</td>
</tr>
<tr>
<td>34–35</td>
<td>0.04 m</td>
<td>0.5</td>
<td>0.46 m</td>
<td>5.3</td>
<td>0.24 m</td>
<td>2.8</td>
</tr>
<tr>
<td>38–39</td>
<td>0.07 m</td>
<td>0.8</td>
<td>0.18 m</td>
<td>2.1</td>
<td>0.14 m</td>
<td>1.6</td>
</tr>
<tr>
<td>43</td>
<td></td>
<td></td>
<td>0.28 m</td>
<td>3.2</td>
<td>0.12 m</td>
<td>1.4</td>
</tr>
<tr>
<td>55–56</td>
<td>0.11 m</td>
<td>1.3</td>
<td>0.48 m</td>
<td>5.5</td>
<td>0.29 m</td>
<td>3.3</td>
</tr>
<tr>
<td>60–61</td>
<td>0.05 m</td>
<td>0.6</td>
<td>0.35 m</td>
<td>4.0</td>
<td>0.30 m</td>
<td>3.5</td>
</tr>
<tr>
<td>68–69</td>
<td>0.70 m</td>
<td>8.1</td>
<td>0.48 m</td>
<td>5.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>70</td>
<td>0.06 m</td>
<td>0.7</td>
<td>0.29 m</td>
<td>3.3</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 13. The duration and total volumetric contribution for HEs for simulation #6 (modified LGM anomaly).

<table>
<thead>
<tr>
<th>HE</th>
<th>Timing (ka ago)</th>
<th>Duration (a)</th>
<th>Total freshwater volume 10^3 km^3</th>
</tr>
</thead>
<tbody>
<tr>
<td>H1</td>
<td>17–18</td>
<td>300</td>
<td>11</td>
</tr>
<tr>
<td>H2</td>
<td>24</td>
<td>100</td>
<td>5.5</td>
</tr>
<tr>
<td>H3</td>
<td>31</td>
<td>200</td>
<td>11.6</td>
</tr>
<tr>
<td>H4</td>
<td>39</td>
<td>300</td>
<td>13.7</td>
</tr>
<tr>
<td>H5</td>
<td>43</td>
<td>300</td>
<td>19.4</td>
</tr>
<tr>
<td>H6</td>
<td>60–61</td>
<td>400</td>
<td>31.4</td>
</tr>
</tbody>
</table>

tioned earlier, the conversion to calendar years is not without problems (Hemming, 2004), and should be considered with caution.

The duration and total volumetric contribution of freshwater events than can be related to HEs, are represented in Tab. 13. The difference to the Tab. 11 is due to differences in the time period, rising from shorter durations of HEs compared to durations of high freshwater fluxes in the model results.

The anomalous HEs, H3 and H6, are especially prominent in the results. H6 is among the highest freshwater peaks in the results and the highest of HEs.

This agreement must be taken with some care because (i) only large calving events as a consequence of fast flow contribute to debris discharge into the ocean (whereas pure melting events do not), and (ii) debris discharge rates need not be proportional to iceberg discharge rates for all times (see, for instance, the discussion by Alley and MacAyeal (1994)). Therefore, the simulated freshwater pulses are now discussed in more detail.

First, it is evident that for all simulations, the temporal evolution of the ice-covered area is more spiky than that of the ice volume in the vicinity of the modelled freshwater pulses (Figs.15 and 35). This is so because a large area of ice-free land can turn into an area covered by a thin ice layer (and vice-versa) within a very short time when climate conditions change accordingly. In contrast, considerable volume changes, either rapid decreases due to a melting/calving event or rapid increases afterwards due to ice accumulation (snowfall) and glacial flow, naturally require more time.

In the following, the focus is on three of the model runs. Simulations # 2 (PMIP forcing) and #3 (reduced $\beta_{ice}$) yield similar scenarios. The flux peak at 26 ka results from the ice sheet reaching its maximum extent on the continental slope. The NCIS and Storfjorden are active, but large volume fluxes can be seen in the Franz Victoria Trough area. At 21 ka, the reason for the freshwater flux is the maximum extent on the continental slope, with high ice volume fluxes at the NCIS, Voring Plateau and Franz Victoria Trough. At 14 ka the SIS melts away, resulting in a large meltwater flux through Bear Island Trough. 11 ka marks the final disintegration of the BSIS and results in large volume fluxes in the Franz Victoria Trough.

For simulation # 6 (Modified LGM anomaly), the discharge history is distinctly different. The flux peak at 26 ka has the same cause as in the other two runs and the NCIS is very active. At the LGM, the ice sheet extends to the continental margin as deep as it is allowed to become glaciated. The NCIS flows fast, but ice-stream activity is moderate elsewhere. As the NCIS transports ice quickly to the coastal calving area, the western section of the ice sheet thins. This simulation thus differs from those mentioned above from 17 ka to 14 ka, which is marked with several active fast
flow areas. Volume fluxes through the Bear Island Trough remain large through this period. In the central Norwegian Voring Plateau, there is a peak in volume flux at 18–16.5 ka ago. At 14 ka the SIS melts away, with increased fluxes in central Norway, Storfjorden and the Bear Island Trough. 11 ka marks the final disintegration of BSIS. Our results indicate an active NCIS at HE3 and HE2, with the activity then shifting to northern fast flow areas. The ice stream activity precedes the fresh water flux peaks. The different temporal and spatial distributions of fast flow regions from the LGM to the Holocene are summarized in Table 15 for the simulations #2 and #6. These volume fluxes precede the freshwater peaks discussed earlier.

Elliot et al. (2001) concluded that the IRD contents in sediment cores indicate that Northern Hemisphere ice sheets send forth iceberg armadas at the same time during HEs. These records also reveal that inter-HEs operate mainly in the Nordic regions and in the Norwegian Sea, well-defined layers of high IRD-content are present close to glacier outlet areas at cold stadials. Thus, it seems that the Scandinavian Ice Sheet may have been very sensitive to climate forcing and could have responded quickly to changes in climate, like the modes of the North Atlantic Oscillation (NAO).

Dreger (1999) studied several sediment cores in the North Sea area. The cores MD95-2011 (near Voring plateau) and MD95-2012 (near Bear Island Trough) reveal high sedimentation and bulk accumulation rates in deglaciation. Around 16.7-16.6 ka, there are high values in the MD95-2011, and there is a lower peak at 18.7–18.9 ka. For the MD95-2012, there are high values from 15.9 to 15.7 ka and a lower peak value from 17.9 to 17.3 ka ago.

These simulation results are in agreement with these findings, in particular simulation # 6. The modified LGM temperature anomalies applied for this simulation are more pronounced in the Scandinavian area than the PMIP anomalies used for the other simulations, so that the temperature effect on coastal Norwegian Sea areas is enhanced. There are repetitive advances of the ice sheets, characterized by high ice volume fluxes. Fast-flow areas on the western margin transport ice to the coast but recover fast from ice volume pulses. The simulation results display freshwater fluxes from melting and calving in phase with Heinrich events. These peaks correspond to fast flow areas, with the main activity at 27 and 22 ka ago in the Norwegian Channel area and later in the Bear Island and Storfjorden region. The activity of these areas seems to be shifting from south to north from the LGM to the Holocene. Some of the freshwater pulses could correspond to Dansgaard-Oeschger oscillations, as well as the ice volume flux peaks on the western margin of the ice sheet.

In a more detailed analysis of volume flux results of Norwegian coastal areas, have similar peaks and one extra appear,
compared to the study of Dreger (1999) in the ice volume flux history of the modified LGM anomaly forcing. For the Voring Plateau, these simulated peaks occur at 18.0–17.7, 17.5–17.0 and 16.8–16.6 ka ago. In the Bear Island regions, our simulations indicate high volume fluxes from 17.8, 17.5 and from 17.2 to 14.2 ka ago.

In conclusion, the simulation results indicate that the simulated ice sheets produce freshwater events synchronously with the HEs and some of the D-Os. These events follow high volume fluxes in the marine marginal areas and ice stream activity periods. The produced volumes compare to observations (see for example Hemming, (2004)). As the ice sheet was forced with the glacial index, the comparison of lag or lead to other ice streams is not possible. However, the results pinpoint that the Eurasian ice sheet was also clearly active in the periods of HEs and D-Os, especially on the anomalous Heinrich Events H3 and H6. This indicates a great sensitivity to the climatic signal, especially in the Fennoscandian sector of the ice sheet.
Conclusions

Calibration of SICOPOLIS application to the Eurasian ice sheet is done with sinusoidal climate forcing. These calibration results do not show sensitive ice sheet behaviour. The validation of SICOPOLIS application to the Eurasian ice sheet history is done by comparisons to other models and to observations, including ice sheet extent, topographic evolution, ground temperature and ice stream reconstructions. The spatial extent and ice-stream locations of the Eurasian glaciations are simulated rather well with our PMIP forcing simulation. However, the volume of the ice sheet remains too low and as a result, deglaciation occurs too rapidly. By applying modified LGM temperature and precipitation anomalies, the results become consistent with the QUEEN reconstruction for the Eurasian ice sheet extent. To improve the simulation of the deglaciation chronology, GCM temperature and precipitation anomalies for 15 and 9 ka ago should be used as additional time-slices for pinning the climate interpolation. Also, better assimilation of palaeo proxy data in the climate simulations should be developed (like DATUN) for better representation of Eurasian climate conditions during the Weichselian.

The forming ice sheets are very sensitive and respond quickly to changes in temperature and precipitation. This indicates a great sensitivity to the climatic signal, especially in the Fennoscandian sector of the ice sheet. The reason for this sensitivity is the larger fraction of temperate ice in the modified LGM climate simulation results. This increases ice flow and ice stream activity. This is reflected in the age of ice in the simulated ice sheets. The results suggest areal and temporal variations in fast-flow areas that were already discussed by Forsström et al. (2003). The fast flow areas that form are located in topographic troughs. Some of these maintain a stationary type of flow, indicating that the corresponding ice stream was of an isbræ-type (the Norwegian Channel Ice Stream).

The simulated topography change is not controversial with observations. The simulated lithosphere temperatures agree with observed data and model results, and call attention to detailed study of the lithosphere temperature history for the Weichselian.

The lobate structure in the Baltic area develops as a result of mass balance flow of warm ice, as frictional heating warms the base and gives rise to moderate velocity flow. In the simulation results, the LGM ice sheet grows to a maximum extent in the Baltic sector without any fast flow events in the area. In the deglaciation phase, the simulation results indicate flow events at the onset areas of the hypothesised Baltic Ice Stream. Also these are short events with moderate velocities. No wet terrestrial
southern margin on the ice sheet is detected in simulation results, although the southwestern part of the ice sheet is very sensitive to the climate signal. This is demonstrated in flow events.

The peaks in freshwater discharge are coincident in all the runs, and the volume of the ice sheet and the climatic conditions determine the freshwater flux. Freshwater peaks can be correlated with the observed occurrence of Heinrich Events (HE) and to some Dansgaard-Oeschger (D-O) oscillations. Other connections to D-O oscillations can be drawn from the ice volume flux peaks in the western marine margin of the ice sheet. High volume fluxes precede the freshwater peaks. The freshwater fluxes include the combined melting and calving events for the whole Eurasian ice sheet and these are dispersed and routed to several seas. The simulated values still show a potential for the Scandinavian Ice Sheet to be an candidate for one agent in the climatic interplay around the HEs. The results pinpoint that the Eurasian ice sheet was clearly active in the periods of HEs and D-Os, especially on the anomalous Heinrich Events H3 and H6. The occurrence of ice stream activity seemed to be shifting from south to north with ongoing time.
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