Sensitivity Tests and Glaciation Scenarios of the Future with CLIMBER-2 — SICOPOLIS Model System for Olkiluoto

Laura Thölix, Natalia Korhonen, Ari Venäläinen, Hannele Korhonen

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ABSTRACT

This report provides climate model simulations performed at the Finnish Meteorological Institute (FMI) in support of formulating climate scenarios for the safety analysis of the Olkiluoto repository. The simulations were run with an earth system model of intermediate complexity (EMIC) CLIMBER-2–SICOPOLIS.

The sensitivity tests showed that the modelled ice volumes over the northern hemisphere (NH) and Fennoscandia are not highly sensitive to the factors and model parameters tested in this study. The ice thickness at Olkiluoto was found to be highly sensitive to the model set-up when Olkiluoto was located close to the ice sheet margin. When CLIMBER-2–SICOPOLIS simulations were compared against a land uplift model, the best fit between the two models was obtained when the time lag of relaxed asthenosphere was set to 4,000 years. A comparison to reconstruction data of the Fennoscandian ice sheet revealed that CLIMBER-2–SICOPOLIS tends to underestimate the ice extent over the Kara Sea during the earlier glaciations, but during the glaciation at 20 kyr BP, CLIMBER-2–SICOPOLIS reproduces well the maximum ice extent.

CLIMBER-2–SICOPOLIS was also used to project future glaciations in the next 200 kyr under different assumptions of atmospheric CO2 concentrations. Under natural background CO2 conditions, variations in insolation make NH glaciation possible in 10–20 kyr AP, 50–60 kyr AP, 90–100 kyr AP and 130–140 kyr AP, and at 20-30 kyr intervals after that. However, human activities have released large amount of additional carbon into the atmosphere and increased the global atmospheric CO2 concentration from ~270 ppm to close to 400 ppm within the past 250 years leading to delay the onset of the next glaciation by several tens of thousands of years. Assuming future scenarios with constant CO2 concentrations, CLIMBER-2–SICOPOLIS predicted ice at Olkiluoto within the next 144 kyr only in the simulations with CO2 equal to or lower than 300 ppm. The timings of these ice events were 50-60 kyr, 95-105 kyr, and 125-135 kyr AP, and the predicted ice thicknesses at Olkiluoto 1600-1800 m for CO2 concentration of 270 ppm, and 800-1300 m for 300 ppm. Furthermore, the simulations showed that the total Fennoscandian ice volume correlates negatively with the atmospheric CO2 concentration, and the simulation with 500 ppm of CO2 prevented all glaciation events in Fennoscandia during the next 144 kyr.

More realistic simulations with time-varying future CO2 concentrations were constructed by combining CO2 scenarios, atmospheric CO2 decline rates and CO2 from a new regression equation developed in this study. The regression equation was fitted using sea surface temperature as a predictor, and showed to be able to reproduce the historical CO2 record. Simulations conducted using the time-varying CO2 concentrations showed that the onset of a glaciation at 10–20 kyr AP is unlikely due to high atmospheric CO2 conditions. However, glaciation was predicted in ~100 kyr AP in all simulated scenarios, and in ~50 kyr AP in all but the business-as-usual scenario with continued high CO2 emissions in the coming centuries. The scenario runs predicted that ice extends Olkiluoto when the atmospheric CO2 concentration has dropped below 280 ppm. In the high-emission run the onset of significant ice formation in Olkiluoto is
delayed until ~130 kyr AP. The maximum ice thicknesses in Olkiluoto are in all the scenarios 1,800-2,000 m.

These simulations with CLIMBER-2–SICOPOLIS highlight that the future concentration pathways of atmospheric CO₂ can have large impacts on the ice formation over Olkiluoto in the next 100,000 years. However, the future emissions from both anthropogenic and natural sources are extremely challenging to foresee due to uncertainties in economical, technological and political developments as well as in the climate-change driven feedbacks in the natural carbon cycle.

**Keywords:** last glacial cycle simulation, Fennoscandian ice sheet, emission scenarios, future climate, next glaciation
Olkiluodon tulevaisuuden jäätköitymissskenaarioita ja herkkystarkastelujensa
CLIMBER-2–SICOPOLIS mallinnusohjelmalla

TIIVISTELMÄ


CLIMBER-2-SICOPOLIS:ta käytettiin myös tulevaisuuden jäätköitymisten arvioimiseen seuraavan 200,000 vuoden aikana. Näitä simulaatioita tehtiin käytäen ilmkehän CO₂-pitoisuudelle erilaisia skenaarioita. Luonnollisten tausta-CO₂-pitoisuuksien valitessa auringonsäteilyn muutokset mahdollistavat pohjoisella pallonpuloliskolla jäätköitymisen 10,000–20,000 vuoden, 50,000–60,000 vuoden, 90,000–100,000 vuoden ja 130,000–140,000 vuoden kuluttua ja tämän jälkeen 20,000–30,000 vuoden välilein. Ihmiskunnan toimet ovat kuitenkin tuottaneet ilmkehään suuria määriä hiiltä, minkä seurauksena CO₂-pitoisuus on nousut esiteollisen ajan 270 ppm:stä lähelle 400 ppm:ää viimeisen 250 vuoden aikana. Tämä tulee viivästytämään seuraavaa jäätköitymistä useilla kymmenillä tuhansilla vuosilla. Kun oletettiin tulevaisuuden CO₂-pitoisuuksien pysyvän vakion, CLIMBER-2–SICOPOLIS ennustaa jäättä Olkiluotoon seuraavan 144,000 vuoden aikana vain simulaatioissa, joissa CO₂-pitoisuus oli 300 ppm tai alempi. Jäätköitymisä tapahtui simulaatioissa nykyhetken jälkeenajina 50,000–60,000, 95,000–105,000 ja 125,000–135,000. Jään paksuus Olkiluodon alueella ylsi 1,600–1,800 metriin 270 ppm:n, ja 800–1,300 metriin 300 ppm:n CO₂-pitoisuudella.

Jäätkön muodostumista Olkiluodon ylle tarkasteltiin myös simulaatioissa, joissa tulevaisuuden CO₂-pitoisuuden arvona käytettiin yhdistelmää IPCC:n CO₂-skenaariosta, ilmkehän CO₂:n eliniästä, ja uudella tämän tutkimuksen yhteydessä kehitettyllä regressiomenetelmällä saadusta CO₂-pitoisuudesta. Regressioyhtälö muodostettiin käytäen selittäjänä merenpintalämpötilaa, ja sen onnitettiin pystyvän tuottamaan historialliset CO₂-pitoisuudet. Muuttuville CO₂-pitoisuksilla tehdyt simulaatiot osoittivat, että jäätköityminen 10,000–20,000 vuoden kuluttua on hyvin epätedennäköistä johtuen ilmkehän korkeasta CO₂-pitoisuudesta. Seuraava jääkausi ennustettiin alkavaksi noin 50,000 vuoden kuluttua muissa paitsi runsaimman CO₂-pitoisuuden skenaariota käyttävissä simulaatioissa, jossa jäätköityminen tapahtui vasta 100,000 vuoden kuluttua. Skenariosimulaatiot osoittivat, että juuri reuna voi yltää Olkiluotoon, jos CO₂-pitoisuus laskee alle 280 ppm:n. Suurin jään paksuus Olkiluodossa oli kaikissa simulaatioissa 1,800–2,000 m.
CLIMBER-2–SICOPOLIS-simulaatiot osoittavat, että tulevaisuuden CO₂-pitoisuuksilla on suuri merkitys jään muodostumiselle Olkiluodon alueella seuraavan 100,000 vuoden aikana. Tulevaisuuden päästöjä ja siten CO₂-pitoisuuksien kehittymistä on kuitenkin erittäin vaikea ennustaa johtuen epävarmuudesta taloudellisissa, teknologisissa ja poliittisissa tilanteissa. Myös ilmastonmuutoksen palautemekanismit saattavat vaikuttaa hiilen kierrtoon ilmastosysteemissä.

Avainsanat: viime jääkausisyklin simulaatio, Fennskandian mannerjäätikkö, hiilidioksidin emissio, skenaariot, seuraava jääkausi
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<th>Definition</th>
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<tbody>
<tr>
<td>AD</td>
<td>Anno Domini</td>
</tr>
<tr>
<td>AP</td>
<td>After Present</td>
</tr>
<tr>
<td>AR5</td>
<td>Fifth Assessment Report</td>
</tr>
<tr>
<td>BP</td>
<td>Before Present</td>
</tr>
<tr>
<td>CLIMBER</td>
<td>CLIMate-BiosphER model</td>
</tr>
<tr>
<td>CMIP5</td>
<td>The fifth phase of the Coupled Model Intercomparison Project</td>
</tr>
<tr>
<td>CRU</td>
<td>Climate Research Unit</td>
</tr>
<tr>
<td>EMIC</td>
<td>Earth system model of Intermediate Complexity</td>
</tr>
<tr>
<td>FMI</td>
<td>Finnish Meteorological Institute</td>
</tr>
<tr>
<td>GAM</td>
<td>Generalized Additive Model</td>
</tr>
<tr>
<td>GCM</td>
<td>General Circulation Model</td>
</tr>
<tr>
<td>Gton C</td>
<td>Gigatonnes of carbon</td>
</tr>
<tr>
<td>IPCC</td>
<td>Intergovernmental Panel on Climate Change</td>
</tr>
<tr>
<td>kyr</td>
<td>1,000 years</td>
</tr>
<tr>
<td>LGM</td>
<td>Last Glacial Maximum</td>
</tr>
<tr>
<td>LLN 2-D</td>
<td>The Louvain-la-Neuve climate model</td>
</tr>
<tr>
<td>MIS</td>
<td>Marine Isotope Stage</td>
</tr>
<tr>
<td>MPI/UW</td>
<td>Max Planck Institute for Meteorology and University of Wisconsin-Madison</td>
</tr>
<tr>
<td>MPM</td>
<td>green McGill paleoclimate model</td>
</tr>
<tr>
<td>NH</td>
<td>Northern Hemisphere</td>
</tr>
<tr>
<td>PIK</td>
<td>Potsdam Institute for Climate Impact Research</td>
</tr>
<tr>
<td>ppb</td>
<td>parts per billion (10^{-9})</td>
</tr>
<tr>
<td>ppm</td>
<td>parts per million (10^{-6})</td>
</tr>
<tr>
<td>RCA3</td>
<td>Rosby Centre Regional Climate Model</td>
</tr>
<tr>
<td>RCP</td>
<td>Representative Concentration Pathway</td>
</tr>
<tr>
<td>SEMI</td>
<td>Surface Energy Mass-balance Interface</td>
</tr>
<tr>
<td>SGU</td>
<td>Geological survey of Sweden</td>
</tr>
<tr>
<td>SICOPOLIS</td>
<td>Simulation Code for POLythermal Ice Sheets</td>
</tr>
<tr>
<td>SRES</td>
<td>Special Report on Emissions Scenarios</td>
</tr>
<tr>
<td>SST</td>
<td>Sea Surface Temperature</td>
</tr>
<tr>
<td>UVic</td>
<td>University of Victoria</td>
</tr>
<tr>
<td>VECODE</td>
<td>Vegetation Continuous Description model</td>
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</table>
1 INTRODUCTION

Posiva Oy is the responsible organization for the final disposal of spent nuclear fuel of its owners (Teollisuuden Voima and Fortum) nuclear power units. The disposal facility will be built to Olkiluoto, where the spent fuel will be placed into a repository at about 400 m depth in the bedrock. Planning this disposal repository requires careful considerations because of handling of highly radioactive material and since it takes about 100,000 years until the radioactivity of the spent fuel has declined to a level comparable with that of the uranium ore used to enrich the nuclear fuel (Chapman & McCombie, 2003). To ensure that no significant amounts of radioactivity will be released into the ground water or the biosphere, multiple technological safety barriers as well as stringent research into future scenarios of the changes in the physical environment at the Olkiluoto site are needed. Climate scenarios, accounting for the possibility of cold periods with large-scale ice sheet growth in the northern hemisphere high latitudes (often called ice ages), are a key consideration (Näslund et al. 2013).

The proxy data from ice cores and ocean sediments document that during the last 650,000 years the Earth’s climate has alternated between cold phases with extensive ice sheets (glacials) and warmer phases without large ice sheets (interglacials) with a periodicity of approximately 100,000 years (Hays et al. 1976; Clark et al. 2009). Within the glacialis, the climate typically varies between cold periods (stadials) and relatively warm periods (interstadials) that last several hundreds or thousands of years. During the cold stadials, ice sheets typically grow in extent, whereas they tend to shrink during the warmer interstadials. The Olkiluoto area has been covered by ice sheets several times during the past glaciations, the latest ending about 9500 years ago.

Studies of the Quaternary period (about 2.6 million years BP to present) provide evidence that past glacial-interglacial variations have been largely driven by the Earth’s orbital changes (Saltzman et al. 1984; Lisiecki, 2010; Huybers, 2010). The orbital theory states that glaciations are triggered by minima in summer insolation in the Northern high latitudes. As a response, the snow aggregated during the winter does not melt over the summer and can therefore gradually accumulate, generating Northern Hemisphere (NH) continental ice sheets. However, the well-known insolation variations cannot alone explain the glacial-interglacial cycles. The impacts of insolation have been amplified by several types of interactions and feedbacks in the Earth’s climate system (Jansen et al. 2007). These include foremost the responses of the carbon cycle (redistribution of carbon between the atmosphere and ocean), the hydrological cycle (ice-albedo feedback, changes in the ocean circulation) and the terrestrial biosphere (albedo effect, CO₂ fertilization effect). Studies of the Last Glacial Maximum (LGM) have estimated that a significant fraction of the global cooling that time resulted from the decreased atmospheric CO₂ concentration, shrunken vegetation cover and increased atmospheric dust content (for details, see e.g. Levis et al. 1999; Wyputta & McAvaney, 2001; Schneider von Deimling et al. 2006).

For simulating climate on a time-scale of 100,000 years, earth system models of intermediate complexity (EMICs) are a helpful tool: Unlike detailed but computationally very heavy general circulation models (GCMs) used in centennial climate scenario runs, EMICs enable orders of magnitude longer simulations with a
reasonable computational cost. On the other hand, in contrast to simple energy balance models, EMICs provide information on the basic large scale circulation features of the climate system. EMICs typically include simplified component models for the atmosphere, ocean, sea ice, land surface, terrestrial vegetation, and ice sheets, which interact under a prescribed solar forcing (insolation). While the spatial resolution of EMICs is rather low (typically from 5 to 50° in horizontal), they are complex enough to capture the essential climate processes and feedbacks over millennial timescales. It should be noted, however, that many EMICs (including the model used in this research) do not explicitly simulate full carbon cycle, and therefore carbon dioxide concentrations need to often be prescribed.

Climate scenarios in support of planning of the Olkiluoto repository have been previously made based on simulation results from several international EMIC modelling groups (hosting the models CLIMBER-2–SICOPOLIS, MPI/UW, or UVic) (Pimenoff et al. 2011; 2012). These simulations indicate that temperate climate will continue at Olkiluoto at least for the next 30,000 years (Pimenoff et al. 2011; 2012), but the insolation minima after 50,000 – 60,000 and 90,000 – 100,000 years after present hold a potential for the onset of the next glaciation (Pimenoff et al. 2011). Therefore, on the time-scale of the next 120,000 years, the repository site may experience the following climate features: interglacial and periglacial climate, ice sheet margin near Olkiluoto, ice sheet over Olkiluoto, and Olkiluoto site below sea level after glaciation due to isostatic depression.

In order to test the robustness of these previous simulations and to perform further model runs specifically designed to address the future glaciation at the Olkiluoto site, the CLIMBER-2–SICOPOLIS model system has now been installed to Finnish Meteorological Institute (FMI) computers. In this modelling system, CLIMBER-2 (Petoukhov et al. 2000; Ganopolksi et al. 2001) describes the climate – biosphere interactions and SICOPOLIS (Greve 1997; Calov et al. 2005) the evolution of the ice sheets. Comparison of the simulation results to proxy data has shown that the CLIMBER-2–SICOPOLIS model system is able to reproduce the amplitude and timing of the ice volume and temperature during the last glacial cycle (Ganopolksi et al. 2010; Pimenoff et al. 2011).

This report documents the new model simulations with CLIMBER-2–SICOPOLIS undertaken at FMI to support the formulation of climate scenarios for the safety analysis of the Olkiluoto repository. The main aims of the research were to 1) conduct sensitivity studies in order to find the optimal model configuration that can reproduce the paleoclimatological reconstructions over Fennoscandia, 2) investigate the impact of future CO₂ scenarios on the climate and onset of glaciation in the next 200,000 years, and 3) estimate the maximum potential thickness of the ice sheet that could form over Olkiluoto within this time period.
2 SENSITIVITY STUDIES

2.1 CLIMBER-2–SICOPOLIS model system

CLIMBER-2 is a climate system model of intermediate complexity that consists of modules for atmosphere, ocean, land surface processes, and terrestrial vegetation cover (Petoukhov et al. 2000). The latitudinal resolution of the atmosphere, land surface and terrestrial vegetation modules is $10^\circ$ and the longitudinal resolution roughly $51^\circ$ (Figure 1). The atmospheric module is a 2.5-dimensional statistical-dynamical model, and the vegetation model is a 2-layer soil moisture module VECODE (Brovkin et al. 1997). The ocean model is a 2-dimensional zonally-averaged 3-basin oceanic module (Atlantic, Indian and Pacific). The resolution of the ocean model in the latitudinal direction is $2.5^\circ$, and in the vertical direction there are 20 levels.

The horizontal resolution of CLIMBER-2’s atmospheric component is low compared to more sophisticated climate models (general circulation models, GCMs), which typically have resolutions of a couple of degrees. This is because the choice of the resolution is always a compromise between the model complexity and computation time; increasing the model resolution increases the computational cost of the model exponentially. Because of this, high resolution GCMs can be applied only to centennial scale simulations; on the other hand, simulations of several hundreds of thousands of years are possible with CLIMBER-2. On such long time scales, it is the average changes over several millennia that are of key interest. For predicting such changes, short term processes in the synoptic scale (of the order of weeks), which require a high atmospheric resolution, do not need to be simulated.

Previous studies have shown that the atmospheric module of CLIMBER-2 reproduces the key observed climate features reliably and also agrees well with simulations done using more complex GCMs: A comparison with present-day climate data has shown that the model successfully describes the seasonal variability of a large set of characteristics of the climate system, including radiative balance, temperature, precipitation, ocean circulation and the cryosphere (Petoukhov et al. 2000). Sensitivity experiments by Ganopolski et al. (2001) showed that the CLIMBER-2 model is able to simulate the climate response to changes in different types of forcing and boundary conditions (such as freshwater flux into the Northern Atlantic, atmospheric CO$_2$ concentration, solar insolation and vegetation cover) in reasonable agreement with the results of GCMs. A recent intercomparison of 15 EMICs showed that CLIMBER-2 reproduces the global climate features of the past millennium well (Eby et al. 2013), and lies well within the range of the other 14 EMICs.
For simulating glacial climates, the CLIMBER-2 model has been coupled with the high resolution 3-dimensional thermomechanical ice sheet model SICOPOLIS (SImulation COde for POLythermal Ice Sheets) (Greve 1997). Figure 2 shows a flow diagram of the CLIMBER-2–SICOPOLIS model system. The resolution of SICOPOLIS is 1.5° x 0.75° (lon x lat), i.e. comparable to the CLIMBER-2 ocean module, and has 20 vertical levels. The model domain covers latitudes from 21° N to 85.5° N. The ice sheet model simulates the extent and thickness, velocity, temperature, water content and age for the ice sheet. The climate and ice sheet components are coupled bi-directionally using a high resolution physically-based Surface Energy and Mass-balance Interface (SEMI) as described in detail by Calov et al. (2005). The SEMI model simulates annual surface mass balance and temperature fields on the grid of the SICOPOLIS model using daily fields of temperature, precipitation, surface height, zenith angle, short wave and long wave radiation, cloudiness and surface wind from CLIMBER-2-model. In SEMI, the albedo of ice is constant (0.4) but the albedo of snow depends on the solar zenith angle, temperature, snow age and atmospheric dust concentration (Calov et al. 2005). Also cloudiness is taken into account in albedo calculations. The climatological variables from CLIMBER-2 are bi-linearly integrated from the coarse grid to the finer grid of SICOPOLIS. It is important to notice that in simulations that extend several millennia into the future (and thus are likely to experience highly various and potentially also completely new climate conditions) a physically based downscaling, such as is done in SEMI, is better than statistical downscaling that is based solely on observed climate data.

Isostatic adjustments of the lithosphere due to changing glacial load are calculated by a local lithosphere relaxing asthenosphere model (Le Meur & Huybrechts 1996), with an adjustable relaxation time parameter. Geothermal heat flux data, for the lower boundary of the ice sheet model, has been previously interpolated from the data by Pollack et al. (1993). However, Section 2.2.1 reports sensitivity tests of the heat flux data, based on which, further simulations were performed with data from Näslund et al. (2005). In ice sheet modelling, the geothermal heat flux values used as input have significant effect on the simulated basal temperature, melt water and ice flow velocities (Greve & Hutter 1995, Näslund et al. 2005). The sensitivity of the ice sheet model to geothermal heat flux have been reported by Greve (2005).
2.2 Experimental design and results

In this study, both historical and future simulations of glaciation cycles in Fennoscandia and especially in the Olkiluoto area were undertaken. The simulated periods were from 126 kyr before present (BP) until either present day (0 kyr BP) or 200 kyr after present (AP). The output data were stored every 1,000 model years as monthly means of the output of CLIMBER-2 and as annual means of the output of SICOPOLIS. The time step used in CLIMBER-2 was one day and in SICOPOLIS one year.

Since CLIMBER-2–SICOPOLIS modelling system was newly installed on FMI computers for the current study, a set of simulations was performed that enabled a comparison of the model performance to the runs conducted at PIK (Potsdam Institute for Climate Impact Research) which were previously used in the Olkiluoto studies of Pimenoff et al. (2011). Identical simulation set-ups were found to produce closely comparable results on the two machines. Furthermore, a test of two spin-up times (126 kyr and 400 kyr) showed that the shorter spin-up can be reliably used, thus significantly decreasing the computational cost of the model runs.

To investigate how the model results react to varying the values of uncertain physical parameters in the model, a set of sensitivity tests was performed. The tested parameters were chosen in collaboration with Posiva based on new data available for comparisons (Ganopolski et al. 2010, Näslund et al. 2005, Pässe, 2001) or on previously published studies on factors that could have significant impact on the modelled ice sheet onset and development. The sensitivity runs conducted are listed in Table 1 and further details about them are given in the sections below.
Table 1. List of simulations for the sensitivity tests.

<table>
<thead>
<tr>
<th>Name of the simulation</th>
<th>Initialization time (kyr BP)</th>
<th>Goal of the simulation</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>126</td>
<td>base line simulation</td>
</tr>
<tr>
<td>HEATFLX</td>
<td>126</td>
<td>effect of geothermal heat flux</td>
</tr>
<tr>
<td>TIMELAG</td>
<td>126</td>
<td>effect of time lag parameterisations</td>
</tr>
<tr>
<td>SLIDE</td>
<td>126</td>
<td>effect of sliding parameterisation</td>
</tr>
</tbody>
</table>

2.2.1 Effect of geothermal heat flux

The geothermal heat flux is an important boundary condition for ice sheet modelling since it strongly contributes to whether the ice sheet base is cold-based or wet-based, and therefore affects the stability of the ice sheet. Geothermal and borehole measurements have demonstrated that the geothermal heat flux shows significant spatial variation (Pollack et al., 1993) and therefore spatially resolved, measurement based datasets for the heat flux are required for ice sheet modelling. In our simulations, data derived from Pollack et al. (1993) and Lee (1970) was used in the baseline run (CTRL), following the approach of Ganopolski et al. (2010). The Pollack et al. (1993) dataset is based on borehole data and has therefore quite poor spatial resolution. More recently, Näslund et al. (2005) utilized γ-emission measurements from radioactively decaying nuclides in the lithosphere in Sweden and Finland to calculate a spatially much more detailed geothermal heat flux field for these two countries (i.e. the approximate core area of the Scandinavian ice sheet during the Last Glacial Maximum). Here a sensitivity test (run HEATFLX) was preformed which uses a combination of the Näslund data for Finland and Sweden, and borehole-based heat flux data (from Pollack and others, ftp://ftp.ngdc.noaa.gov/Solid_Earth/Global_Heatflow) for the surrounding areas.

Figure 3 illustrates the heat fluxes from these two datasets. Overall, the datasets give consistent heat flux fields but some differences are also seen (e.g. over France and the Arctic Ocean). In the Fennoscandian region, the heat flux in Southern Finland and especially in the west coast of Finland (where Olkiluoto is located) is more than 10 mWm⁻² higher in the Näslund dataset. On the other hand, in Eastern Finland the heat flux is about 20 mWm⁻² lower in the Näslund dataset.
Figure 3. Geothermal heat flux (mWm$^{-2}$) over Europe according to a) Pollack et al. (1993) and Lee (1970) and b) Näslund et al. (2005) in SICOPOLIS grid.

Figure 4 demonstrates how altering the geothermal heat flux dataset within the CLIMBER-2–SICOPOLIS model system affects the simulated time series of ice volume in the NH and Fennoscandia as well as the ice thickness in Olkiluoto. For the NH mean ice volume the difference between CTRL and HEATFLX runs are in general minor (Figure 4a). However, the HEATFLX run predicts slightly smaller ice volume around 250 kyr and 50 kyr BP. The glacial time periods are consistent between the model runs. Good agreement between the two runs is evident also in Figure 4b which shows the ice sheet volume over Fennoscandia. However, the HEATFLX run predicts larger maximum ice volumes during some of the glaciation peaks.

Figure 4c compares the ice and bedrock height in the SICOPOLIS grid containing Olkiluoto between the two simulations. While the two simulations agree very well for most of the simulated period, there is a clear discrepancy around year 270 kyr BP when ice melts completely in the HEATFLX run but the ice thickness exceeds 1200 m in the CTRL run. This is because at this time the Olkiluoto site is located close to the ice sheet margin, and even small changes in the key model parameters can affect the exact location of the ice sheet edge. This is depicted in Figure 5a, which shows the average and standard deviation of the ice thickness from the nine model grid points closest to Olkiluoto as well as separately the Olkiluoto grid point values. Around 265 kyr BP, both model runs predict ice-covered and ice-free grids in the vicinity of Olkiluoto, although on average the ice thickness is clearly higher in the CTRL run. This example illustrates that when local conditions close to the ice sheet margin are of interest, one should not focus on a single model grid point but rather also look at the predictions for the neighbouring model grid points. (Near Olkiluoto the gridsize in the CLIMBER-2–SICOPOLIS model is about 80 km x 80 km.) Furthermore, Figure 5 shows that in the CTRL run 264 kyr BP, the Olkiluoto site is under ice, whereas in the HEATFLX run the ice sheet has retreated further north and left Olkiluoto without ice. Another difference between the two model runs for the Olkiluoto site is that before the last glacial, around
35 kyr BP, there is more ice in the HEATFLX run (Figure 4c). A map of the ice coverage for this time period (i.e. 35 kyr BP) in Figure 6 shows that the HEATFLX run gives overall a slightly larger coverage. The figure also illustrates that Olkiluoto is again located close to the ice margin.

Based on this sensitivity test, it is concluded that Näslund et al. (2005) data is well suited for CLIMBER-2–SICOPOLIS modelling. Since it has much better spatial resolution than the original Pollack et al. (1993) data, all simulations from Section 2.3.1 onwards were run using the Näslund heat flux data.
Figure 4. Effect of the geothermal heat flux on a) the simulated ice volume of the Northern Hemisphere, b) simulated ice volume of the Fennoscandian ice sheet, c) simulated ice height (solid line) and bedrock height (dashed line) over the nearest grid point of Olkiluoto. Red lines show HEATFLX run with Näslund geothermal heat flux data and black lines CTRL run with Pollack geothermal heat flux data.
Figure 5. a) Average and standard deviation of the ice sheet thickness in the nine nearest gridpoints to Olkiluoto between 300 and 200 kyr BP from HEATFLX (red line, orange error bars) and CTRL (black line, grey error bars) runs and the Olkiluoto grid point values from HEATFLX (green line) and CTRL (blue line) runs. b) The placement of the ice sheet over Fennoscandia at 264 kyr BP. Olkiluoto is shown with a green dot.
Figure 6. Map of NH ice sheet at 35 kyr BP. Red lines show HEATFLX run with Näslund geothermal heat flux data and black lines CTRL run with Pollack geothermal heat flux data. Olkiluoto is shown with a green dot.

2.2.2 Effect of time lag of the relaxed asthenosphere

In SICOPOLIS, the isostatic depression and rebound of the lithosphere due to changing ice load is modelled with a local lithosphere relaxing asthenosphere (LLRA) approach. This approach assumes that the ice load causes a steady-state displacement of the lithosphere in the vertical direction only locally (Greve and Blatter, 2009). The viscous properties of the underlying asthenosphere means that the displacement of the lithosphere cannot take place immediately, and the relaxing asthenosphere approach parameterizes this time lag by a single time constant. In reality, however, the elasticity of the lithosphere results also in non-local displacements. Le Meur & Huybrechts (1996) have shown that the LLRA approach can capture the time-dependent behaviour of the total ice volume and mean bedrock elevation well. On the other hand, the assumption of purely local displacement can cause locally higher uplift rates than more complex approaches (Le Meur & Huybrechts, 1996).

In the simulations by Ganopolski et al. (2010) the time lag of the relaxed asthenosphere was set to 3,000 years (Calov et al. 2005), and also doubling of the time lag was tested. To investigate the sensitivity of the simulation results to the assumed value of this delay, additional runs were performed with the time lag set to 2,000 years, 4,000 years, 5,000 years and 6,000 years. Based on earlier studies (e.g., Ganopolski et al. 2010), these values can be considered realistic in the light of the current uncertainty range.

Figure 7 shows the simulated ice volumes in the NH and in Fennoscandia (panels a and b) as well as the ice thickness in Olkiluoto (panel c). Overall, the simulation results are not highly sensitive to the value of the time lag parameter. The ice volume tends to
increase with increasing time lag (Figures 7a and 7b). Furthermore, the generation of ice in the beginning of a glaciation period is slower when the time lag is longer. This is because a longer time lag causes the bedrock height to remain below the sea surface for a longer period of time. The heat flux from the oceans increases and slows down the generation of the ice sheet.

For ice height in Olkiluoto, the picture is more complicated (Figure 7c). During interglacials, the bed rock height is the larger, the shorter the time lag is. This is because with small time lags the bed rises faster after the ice has melted. Similarly, the bedrock height decreases faster with smaller time lag values during glaciation. Large differences between the sensitivity simulations are seen around 55 kyr BP, when ice remains over Olkiluoto in the runs with long time lags, but melts completely in runs with short lags. This is also reflected strongly in the bedrock height. Significant differences are seen also around 35 kyr, when the Olkiluoto grid point is close to the ice sheet margin (Figure 8).
Figure 7. Effect of the time lag of the relaxed asthenosphere on a) the simulated ice volume of the Northern Hemisphere, b) simulated ice volume of the Fennoscandian ice sheet and c) simulated ice height (solid line) and bedrock height (dashed line) over the nearest grid point of Olkiluoto. Magenta line is 2,000 years time lag, red 3,000 years, black 4,000 years, blue 5,000 years and turquoise 6,000 years.
Figure 8. Map of the simulated ice border in NH at 35 kyr BP. Magenta line is 2,000 years time lag, red 3,000 years, black 4,000 years, blue 5,000 years and turquoise 6,000 years. Olkiluoto is shown with a green dot.

Overall, these sensitivity simulations show that the predicted total ice sheet volumes and ice thickness in Olkiluoto are not highly sensitive to the time lag of the relaxed asthenosphere. However, the value of the time lag can affect the simulated local ice heights when the site of interest is located close to the ice sheet margin. In order to choose the best-fit value of the time lag to be used in further simulations, we compared our sensitivity test results against land uplift data from a study of Pohjola et al. (2012), which investigated the past topography in Fennoscandia from 14,000 to 500 BP. The uplift data in their study was obtained from a semi-empirical land uplift model of Pässe (2001), which is based on data from Geological Survey of Sweden (SGU). The parameters of the land uplift model were refined in Pohjola et al. (2012) with lake and mire isolation data and new estimates of ice recession time, Moho depth and eustatic sea level rise. The data points (lake and mire isolations, archaeological findings) were not evenly distributed in the modelling area, which may have influenced the results of the Pohjola et al. (2012) study. Here, the land uplift model data is interpolated to the same grid with CLIMBER-2–SICOPOLIS model for the comparison.

Figure 9 shows the bedrock height, land uplift and uplift speed in the two nearest gridpoints from Olkiluoto. The western grid point (left panel) is under the sea level throughout the simulation, while the eastern point rises above the sea level about 7 kyr BP. The top panels of Figure 9 show that the modelled bedrock height agrees well with the data, especially when the time lag in CLIMBER-2–SICOPOLIS is set to 3,000 or 4,000 years. In the middle panels of Figure 9 the land uplift has been calculated with respect to the present day bed rock height. The model run with 4,000 year time lag fits
best with the data. In terms of the land uplift speed, shown in the bottom panels of Figure 9, it is not clear which value of the time lag results in the best match with the semi-empirical model. This is because CLIMBER-2–SICOPOLIS predicts a smooth evolution of the uplift speed, but the semi-empirical model shows fast changes in the speed after ice has just left the area. Based on these comparisons, the time lag of 4,000 years was chosen for the rest of the simulations in this report.

Using the CLIMBER-2–SICOPOLIS output with this optimized time lag, the land height distributions predicted by CLIMBER-2-SICOPOLIS and Land uplift model at different historical times were compared. The results of this comparison are shown in Figure 10. In the figure, white denote areas under ice sheet, black sea areas and grey land areas. The orange lines denote the coastline at present day for comparison. The black lines show the simulated bedrock height in meters. Comparison of the left and right panels shows that the horizontal resolution in CLIMBER-2–SICOPOLIS model is quite coarse, and thus some of the finer features e.g. along the coastline cannot be captured. However, the overall good match indicates that the optimized time lag value can be considered representative throughout the Fennoscandian region.
Figure 9. Bedrock height (a and b), land uplift (c and d) and uplift speed (e and f) at lon=21, lat=61.5 (left) and at lon=21.75, lat=61.5 (right) in case of different time lag coefficients. Turquoise line indicates time lag of 2,000 years, black line 3,000 years, green line 4,000 years, blue line 5,000 years and red time lag 6,000 years. The magenta line is from the Land uplift model. All the lines start after the ice have melted from the grid point. Since the SICOPOLIS data is obtained on a much coarser resolution than the semi-empirical land uplift model data, the mean and the standard deviation in the area of the corresponding SICOPOLIS grid point is shown for the semi-empirical model data.
Figure 10. Land height over Fennoscandia in the Land uplift model (Påsse, 2001) (left) and in CLIMBER-2–SICOPOLIS model (right). Orange line denotes present coastline in the model.
2.2.3 Effect of sliding parameterisations

Sliding processes play an important role in ice-sheet dynamics (Calov et al. 2002). In SICOPOLIS, basal sliding happens only if the base of the ice sheet is at melting point. On the other hand, Ganopolski et al. (2010) showed that for SICOPOLIS the modelled ice sheet evolution is quite sensitive to the bottom sliding parameter. Therefore, the sensitivity of the model to the bottom sliding parameter was tested by multiplying its value by 4/3 (i.e. 1/3 increase) and in a separate simulation by 2/3 (i.e. 1/3 decrease). SICOPOLIS has three ground types (rock, terrestrial sediments and marine sediments) and all the types have their own sliding parameters \((10^{-6} \text{ mm yr}^{-1} \text{ Pa}^{-1}, 5\times10^{-6} \text{ mm yr}^{-1} \text{ Pa}^{-1}\text{ and } 50\times10^{-6} \text{ mm yr}^{-1} \text{ Pa}^{-1}, \text{ respectively})\). In the sensitivity simulations all these three parameters were modified by the same coefficient. The sediment types used in this study in Europe can be seen in Figure 11.

![Figure 11. Sediment types in Europe.](image)
Figure 12 shows again the simulated ice volumes in NH and Fennoscandia and ice thickness over Olkiluoto as well as a map of ice sheet extent at 35 kyr BP. Compared to the other sensitivity runs (Sections 2.2.1–2.2.2), the differences between the simulations are large. Figure 12 shows that the ice volume decreases consistently with an increasing sliding coefficient. This is because for large coefficient values the modelled ice sheet is more mobile and slides faster to the ocean. Therefore, much of the new ice forms on ice free bedrock or on thinner ice sheets. For the NH ice volume, the differences between the individual simulations are 3-8 million km$^3$ (Figure 12a), whereas over Fennoscandia they are 0.5-1 million km$^3$ (Figure 12b). For the Olkiluoto ice height the differences are small in the beginning of the simulation, but around year 60 kyr BP it can be seen that ice remains in Olkiluoto only in the simulation with the smallest sliding parameter (Figure 12c). Consequently, the bedrock height also remains lower in this simulation. Furthermore, there is significant ice melting around 35 kyr BP with the two higher sliding parameter values, but very little with the lowest value. The map in Figure 13 shows that the choice of the sliding parameter impacts the extent of the ice sheet especially over southern Alaska and over Fennoscandia.

Based on these sensitivity runs it is concluded that the model simulates the timings of glaciations consistently for all the tested sliding parameters. However, the uncertainty in the sliding parameter does affect the simulated ice sheet extent and volume. Because Olkiluoto is often located close to the ice sheet margin, the choice of the sliding parameter can affect whether the model predicts an ice sheet over Olkiluoto.
Figure 12. Effect of the sliding parameterization on the a) simulated ice volume of the Northern Hemisphere, b) simulated ice volume of the Fennoscandian ice sheet and c) simulated ice height (solid line) and bedrock height (dashed line) over the nearest grid point of Olkiluoto. Red lines are for the CTRL run, black SLIDE+ with increased sliding and blue SLIDE- with reduced sliding. Olkiluoto is shown with a green dot.
Figure 13. Map of the simulated ice border in NH at 35 kyr BP. Red lines are for the CTRL run, black SLIDE+ with increased sliding and blue SLIDE- with reduced sliding. Olkiluoto is shown with a green dot.

2.3 Comparison with the reconstructions of the Fennoscandian ice sheet

Next, the simulated Fennoscandian ice volume and Olkiluoto ice thickness were compared to reconstruction data (Svendsen et al. 2004) in order to evaluate whether the CLIMBER-2–SICOPOLIS model is able to reliably reproduce ice occurrence over the Olkiluoto site during 126 kyr BP and present time. Svendsen et al. (2004) derived from reconstructions that the ice sheet around 100-80 kyr BP did not reach Southern Finland, but around years 65-55 kyr BP and 25-15 kyr BP it did.

The simulated ice sheet extent at three different ice sheet maximums during the past 90 kyr is compared against the Svendsen et al. (2004) paleoclimatic reconstruction of Northern Eurasia in Figure 14. According to the reconstruction, the ice sheet maximums are approximately in 90 kyr BP, 60 kyr BP and 20 kyr BP, during MIS 5b, MIS 4 and MIS 2 stadials, respectively. The corresponding largest areal ice sheet coverage is achieved in CLIMBER-2–SICOPOLIS 93 kyr, 67 kyr and 20 kyr BP. It should be noted that exact timing of the ice sheet maximums for the reconstructions is very challenging (especially the further back in time one goes), and thus slight deviations in the timing between the model and reconstructions are expected. Figure 14 presents the simulated maximum ice extent during the periods 85-95 kyr, 55-65 kyr and 15-25 kyr BP. Regarding the Svendsen et al. (2004) reconstruction data, it should be noted that the maximum extent during each glaciation is not necessarily attained at the same time in all regions.
Panels a and b in Figure 14 show that in 90 kyr BP, the model predicts a clearly smaller ice sheet extent than the reconstruction. In the north-east over the Kara Sea the ice sheet extends much further in the reconstruction than in the model simulation. Furthermore, the model predicts ice over Southern Scandinavia, whereas the reconstruction does not. The agreement between the model and the reconstruction clearly improves for the 60 kyr BP glacial maximum (panels c and d). However, the modelled ice sheet extent over the Kara Sea is still smaller than in the reconstruction. On the other hand, the model predicts larger ice sheets over parts of Northern Siberia and over the Northern Sea than the reconstruction. During the latest glacial maximum at 20 kyr BP, the model and reconstruction are in good agreement (panels e and f) over Fennoscandia, although there are still differences over the Kara Sea. In the model simulations Olkiluoto is under ice during all three glacial maxima; however, this is not the case in the reconstruction during the first glacial in 90 kyr BP. It should be noted that each glaciation cycle destroys some of the signs of previous glacial extents in the nature, and therefore the reconstructions are most reliable for the latest glacial maximum. Therefore, the good agreement between the model and the reconstruction 20 kyr BP gives us confidence in the performance of the CLIMBER-2–SICOPOLIS modelling system.

The right panels in the Figure 14 show also results from different sensitivity runs performed in the previous sections. The red curve denotes the HEATFLX run with Näslund geothermal heat flux (Section 2.2.1), the blue curve is from SLIDE run, for decreased sliding parameter (original value multiplied with 2/3, see Section 2.2.3) and the black curve from TIMELAG6000 run denotes long time lag (6,000 years) in the land uplift (Section 2.2.2). The differences between these three runs are small and confined to the ice sheet edges. Olkiluoto is covered with ice in all the simulations during all three glacial maxima. Based on this comparison, and earlier comparisons presented in previous sections, it is concluded that the ice sheet coverage simulated by CLIMBER-2–SICOPOLIS is not highly sensitive to the tested model parameters, and thus small uncertainties in their exact values do not affect the robustness of the model results.
Figure 14. Ice sheet in Fennoscandia according to reconstructions (left) and model results (right) (Figures are from Svendsen et al. 2004). The lines in the right panels are the maximum ice sheet extent between mentioned years from HEATFLX run (red), decreased SLIDE run (blue) and TIMELAG6000 run (black).

Figure 15. a) Ice sheet of North America according to reconstructions (Figure from Dyke & Prest (1987)) and b) Ice sheet of North America, Alaska and Greenland from CLIMBER-2-SICOPOLIS simulation. Both panels are from 18 kyr BP.
The simulated ice sheet extent and height was compared also against reconstructions of the North American ice sheet (NAIS). Figure 15 a shows the ice margin of NAIS from Dyke and Prest (1987) and Figure 15 b in CLIMBER-2-SICOPOLIS in 18 kyr BP. It should be noted that the reconstruction deliberately excludes the ice sheets in Alaska and Greenland, whereas they are shown for the model results. The ice sheet of CLIMBER-2-SICOPOLIS is about the same size than in the reconstruction. Also the ice surface height is comparable to the simulations of Marshall et al. (2002) (not shown).

2.4 Conditions over Olkiluoto

Due to the very low resolution in the atmospheric module of CLIMBER-2, statistical downscaling was utilized to obtain temperature and precipitation values at Olkiluoto in the same resolution as SICOPOLIS results. The statistical downscaling method was based on Generalized Additive Model (GAM)-type regression model (Wood, 2006), which can account for both linear and non-linear effects. The GAM applied in this study has been introduced in detail in Pimenoff et al. (2011) and Korhonen et al. (2014), and its construction is discussed below only briefly.

The GAM was generated by fitting statistical relationships between the low resolution CLIMBER-2–SICOPOLIS model data and either observational data or high-resolution regional model data. Temperature and precipitation data for Fennoscandia for recent climate (1961-1990) were taken from CRU (Mitchell & Jones 2005). For two historical periods (glaciation at 21 kyr and permafrost at 44 kyr BP), the climate data were obtained from the regional climate model RCA3 (Kjellström et al. 2005). Thus with the constructed GAM, it is possible to downscale the monthly mean temperatures and total precipitation produced by CLIMBER-2 model for the entire last glacial cycle. The explanatory variables used were large-scale temperature and precipitation from CLIMBER-2, and elevation, distance to ice sheet, slope direction and slope angle from SICOPOLIS. The surface temperatures produced with the fitted GAMs have been shown to be in good agreement with the observed and modeled surface temperatures with an error in present-day climate about -0.1 °C, in permafrost climate about 0.5 °C and in glacial climate about 0.3 °C (Pimenoff et al. 2011, Appendix 2). For the total precipitation GAMs overestimate the present-day climate by 11% and the permafrost climate by 4%. In the glacial climate the annual precipitation is underestimated by 19% (Pimenoff et al. 2011, Appendix 2).

For this study, the GAM was applied for downscaling precipitation and temperature from a 126,000 year simulation with the optimized CLIMBER-2–SICOPOLIS modelling system. The main interest of this research is the Olkiluoto site; however, since the model resolution is relatively poor and (as seen before in Section 2.2) even small uncertainties in the model configuration can affect e.g. the exact location of the ice sheet margin, more reliable estimates of the possible range of values can be obtained by looking at regional averages, rather than only the model grid point of interest. Therefore, the analysis concentrates on the nine nearest gridpoints to Olkiluoto (longitude between 21 and 24 °E and latitude between 60.75 and 62.25 °N), and their average values and standard deviations are presented.

Figure 16 shows a) annual mean temperature, b) annual total precipitation and c) ice sheet thickness averaged over nine nearest gridpoints to Olkiluoto. The error bars in the
figure depict two standard deviation anomalies about the mean value. SICOPOLIS surface temperature (blue line) originates from CLIMBER-2 climate model and has been scaled to the SICOPOLIS grid by the SEMI interface. Downscaled temperature (black line) in Olkiluoto has been calculated using the GAM fitting. In the figure, there are also the CRU observational and RCA3 regional climate model values for temperature for the time slices over which the GAM was fitted. Figure 15a shows that during interglacials SICOPOLIS surface temperatures are 2-10° C colder in the Olkiluoto area than the GAM 2-metre air temperatures; however, during glacial the agreement between the two methods is good. On the other hand, the GAM temperatures are in good agreement with the CRU observation and RCA3 regional modeling results in the present-day climate, but in the glacial climate 21 kyr BP and in the permafrost climate 44 kyr BP the temperatures differ. Figure 16b shows the downscaled total annual precipitation and its deviation in the Olkiluoto area. Precipitation values from observations and the regional model are again shown with red dots. In this case, the GAM captures the present-day and permafrost climates well, but in the glacial climate when the annual mean precipitation is the highest, the GAM underestimates the regional model result by 10%. Note, however, that accounting for the deviation among the nine model grid points (represented by the error bars) clearly improves the match with the regional model data also in the permafrost period.

Figure 16c presents the ice thickness and its variations in the Olkiluoto area. The CLIMBER-2–SICOPOLIS system gives the ice thickness directly in the SICOPOLIS resolution, and thus no downscaling was required. The maximum simulated ice thickness (averaged over 9 grid points) was about 2 km and it was reached in ~27 kyr BP. It is noteworthy that during all glaciation periods, the deviation in ice thickness between the 9 investigated grid points is large: it reaches up to 1500 m around 35 kyr BP and it varies between 700-800 m at other times. Accounting for this deviation, the largest predicted ice sheet thickness for the Olkiluoto area is 2.3 km in 27 kyr BP. As stated above, the largest variation between the 9 model grid points occurs around year 35 kyr BP, which is the same time that showed the biggest variation between the sensitivity studies in Sections 2.2.1–2.2.3. This again confirms the earlier conclusion that the poor resolution of EMICs is problematic during time periods when the site of interest is close to the ice sheet margin.
Figure 16. a) Annual mean temperature, b) annual total precipitation and c) ice thickness averaged around Olkiluoto grid point. Black lines are down scaled results, blue from SICOPOLIS and red cots are from CRU and RCA3. Ice thickness is from SICOPOLIS. Grey errorbars show two standard deviations within the averaged area.

Figure 17a compares the simulated ice thickness in Olkiluoto to the maximum and average Fennoscandian ice thicknesses during different Weichselian stages from year 126 kyr BP until present. The highest simulated ice thickness over Fennoscandia reaches 3.3 km around 18 kyr BP, while at the same time the Olkiluoto grid point value is only about 1.9 km. Similar behaviour is seen during all simulated glaciation maxima, i.e. the ice sheet thickness at Olkiluoto is 800 to 1400 m lower than the Fennoscandian maximum. The ice thickness remains lower than during Late Saalian glaciation.
(Lambeck et al. 2006) when it reaches 4000 m. Figure 17b shows maps of ice thickness during some of the past glaciation maxima: 116 kyr BP, 70 kyr BP and 18 kyr BP. In all these cases, CLIMBER-2–SICOPOLIS predicts the maximum ice in Norway and Northern Finland, i.e. north of Olkiluoto. This implies that a relatively small shift southwards in the modelled glaciation center could cause the model to predict much thicker ice sheet also over Olkiluoto, but not thicker than 3.3 km.

Figure 17. a) Ice thickness in Olkiluoto gridpoint (black), average ice thickness in Fennoscandia (green) and the maximum in the Fennoscandian area (red). b) Ice sheet thickness (m) in the map at 116 kyr BP, 70 kyr BP and 18 kyr BP.
3 GLACIATION SCENARIOS OF THE FUTURE

3.1 Background

Consideration of future glacial cycles is of key importance when planning disposal of spent nuclear fuel in the bedrock at high-latitude sites, such as Olkiluoto. Aspects of concern include e.g., the mechanisms of penetration to and distribution of ice sheet meltwater in the bedrock. While such aspects cannot be directly studied with climate models, climate scenarios of the future glaciation cycles are needed for permafrost and groundwater modelling. Therefore, the CLIMBER-2–SICOPOLIS modelling system, which was evaluated and optimized in Chapter 2, was used to examine the climate and ice conditions over Fennoscandia 144 kyr into the future.

The three main factors controlling the global mean climate are insolation, planetary albedo and strength of the greenhouse effect (mainly dominated by the concentrations of water vapour and CO₂), and substantial changes in any of these three factors can impact the climate. It is also possible that several of these factors change parallel in time and thus either strengthen or counteract each other’s climatic impacts. For example, according to the orbital theory of glaciation (Milankovitch, 1941), initiation of ice sheet growth over North America and Eurasia is triggered by a sufficient drop in the summer-time solar insolation. The growth of the ice sheet, on the other hand, decreases the planetary albedo causing a larger fraction of solar radiation to be reflected back to space, and thus further decreasing the surface temperature. Analyses of ice core data have also shown that the atmospheric CO₂ concentration has varied nearly synchronously with temperature over the past hundreds of thousands of years (Petit et al. 1999), thus providing a strong cooling feedback during the glacial periods. The controlling mechanisms of CO₂ decline during glaciations remain unquantified, but it is likely that changes in ocean circulation, biological and carbonate oceanic pumps, and CO₂ solubility into ocean water all play a role (Goosse et al., 2008-2010).

As discussed above, the onset of glaciation is primarily determined by sufficient drop in summer-time insolation. However, the magnitude of the required drop, i.e. the threshold value of insolation, depends on the atmospheric CO₂ concentration (Archer & Ganopolski, 2005). Figure 18 shows schematically how the EMIC simulated onset of the next glaciation depends on the solar insolation and CO₂ concentration: a NH solar insolation minimum triggers a glacial inception, while high CO₂ concentration can delay the glacial inception. The higher the CO₂ concentration is, the deeper minimum in insolation is required to trigger glaciation.

Since the dawn of the industrialization in the 18th century, human activities have released into the atmosphere about 535 ± 70 Gton of carbon, mainly from fossil fuels but also from cement production and land use changes (Le Quéré et al. 2014). This has caused the atmospheric CO₂ concentration to increase from about 280 ppm in the preindustrial time to close to 400 ppm. This increase is considered the major contributor to the observed global mean temperature rise of 0.85 ± 0.2 °C during the last 150 years (IPCC, 2014). The Earth’s climate is projected to undergo further warming in the next centuries, as the anthropogenic emissions continue to raise the atmospheric CO₂ concentration.
However, future atmospheric concentration of CO$_2$ on millennial time scales is highly uncertain. This is mainly because of two reasons: First, future CO$_2$ concentrations depend heavily on future anthropogenic emissions, which are affected by a complex – and to a large extent unpredictable – interplay of technological advances, economic growth and international treaties. Second, the natural feedback processes, which control the exchange of carbon between the atmosphere, ocean, vegetation and soil under different climate conditions, remain insufficiently quantified (Frank et al. 2010). Examples of potential future feedbacks include decline of Amazonian vegetation carbon storage due to warming and drying, increase of photosynthesis and thus vegetation carbon storage in cooler climates as a response increased CO$_2$ concentration, and decrease of soil carbon storage due to increased respiration under high temperatures (Cox et al. 2000). Furthermore, methane release from ocean clathrates (Archer et al. 2009) and CO$_2$ release from permafrost soils (Dutta et al. 2006) upon climate warming can have big impacts on atmospheric CO$_2$ concentrations. On glacial timescales, the impact of CO$_2$ from both natural and anthropogenic sources depends also on the timing of the emissions: it has been estimated that of the CO$_2$ emitted 10-15% remains in the atmosphere after 10 kyr and approximately 7% after 100 kyr, in absence of natural CO$_2$ forcing such as glacial inception (Archer, 2005).

Given the large uncertainties in the future CO$_2$ concentrations, climate scenarios investigating the timing and extent of future glaciations need to cover the whole range of feasible concentrations. Previous modelling studies (Loutre & Berger, 2000; Texier et al. 2003; Archer & Ganopolski, 2005; Cochelin et al. 2006) have estimated that there are three periods during the next 100,000 years with a potential for ice sheet formation: around 10-20 kyr after present (AP), 50-60 kyr AP, and 90-100 kyr AP. These periods coincide with the Northern Hemisphere summer insolation minima. However, many studies also suggest that glacial inception will most likely not start during the next 50,000 years if the atmospheric CO$_2$ concentration remains higher than the typical pre-industrial value of ~280 ppm (Loutre & Berger, 2000; Texier et al. 2003; Archer & Ganopolski, 2005; Cochelin et al. 2006; Herrero et al. 2014). The present day CO$_2$ concentration is much higher than this and is projected to stay high for thousands of years (Archer, 2005); therefore, glacial interception within the next 50 kyr seems unlikely if anthropogenic greenhouse gas emissions are not rapidly reduced in the next few decades. After that, the first possible time periods for glaciation occur during the NH insolation minima in 50–60 kyr AP and 90–100 kyr AP. However, it is possible that sustained high atmospheric greenhouse gas concentrations can delay the onset of the next glaciation also beyond these time periods. Cochelin et al. (2006) concluded that CO$_2$ concentration of 300 ppm may push the next glacial inception beyond the next 100 kyr (Cochelin et al. 2006). Using the total carbon release as a metric, Archer and Ganopolski (2005) showed that carbon release of 1,000 Gton (approximately double that of anthropogenic emissions up till now) would prevent glaciation for the next 130 kyr. With a carbon release of 5,000 Gton C glaciation might not occur within the next 500 kyr. In their study, the CO2 decline rate after the anthropogenic emission peak was simulated based on results from a carbon cycle model of Archer (2005). On the other hand, using simple relaxation models, Herrero et al. (2014) assumed total anthropogenic emissions of approximately 1300 Gton C and concluded that such carbon emissions will push the next glacial maximum to 105 kyr AP.
The subsequent sections detail the future climate runs made with CLIMBER-2–SICOPOLIS modelling system to investigate the future glaciation cycles in Olkiluoto. Idealized scenarios with constant CO₂ concentration were performed to enable comparison to previous modelling studies (see above) as well as more realistic scenarios using CO₂ emission scenarios resulting in time-varying CO₂ concentrations.

As always with modelling of complex highly non-linear systems, it must be remembered that even the best of numerical models are an imperfect representation of the climate system. Confidence in model performance can be obtained by evaluating the simulations of the past climates against historical records (as was done in Chapter 2); however, even a good agreement between model results and historical climate does not guarantee a truthful prediction of the future. This is because no climate model can incorporate a perfect physical representation of all interactions and feedbacks within the climate system, both due to computational cost and gaps in current theoretical understanding. Despite this, well-evaluated climate models are the best available tool to project scenarios of future forcing factors into plausible future climate conditions. As such, they can significantly limit the set of possible future states, and thereby support the planning of multimillennium-scale projects.

Figure 18. The role of solar insolation and atmospheric CO₂ concentration in the next glacial inception. a) Stadial-interstadial timing depending on the CO₂ concentration and b) June solar insolation at 65°N. Picture taken from Pimenoff et al. (2011).
3.2 Constant CO₂ scenarios

In the first set of experiments, the CLIMBER-2–SICOPOLIS model system was set to climate simulations from 126 kyr BP to 144 kyr into the future. The historical part of the simulation was forced with Vostok ice core CO₂ equilibrium concentrations (Petit et al. 1999) following Ganopolski et al. (2010). The future CO₂ concentrations were set to different constant values: 270 ppm, 300 ppm, 340 ppm, 380 ppm, 400 ppm and 500 ppm. As such, these constant CO₂ concentration simulations are highly idealized, but they are useful in quantifying a threshold value above which future glaciations in Fennoscandia and/or in Olkiluoto are delayed. Constant concentration simulations are also common in previous literature, (e.g. Archer and Ganopolski, (2005) and Cochelin et al. (2006)), and thus these experiments enable us to put the CLIMBER-2–SICOPOLIS results into context with earlier studies.

Results from the constant CO₂ simulations are depicted in Figure 18. The upper panel shows the time series of the Fennoscandian ice volume and lower panel Olkiluoto ice thickness under the different concentration scenarios. Figure 18a shows that the Fennoscandian ice volume is strongly correlated with CO₂ concentration. The higher the concentration, the smaller the peak ice volume and also the shorter the duration of the stadials. Furthermore, CO₂ concentration of 500 ppm of CO₂ prevents all glaciation events over Fennoscandia for the next 144 kyr.

Figure 19b shows that the ice sheet reaches Olkiluoto at the glacial inception, if the CO₂ concentration is 300 ppm or less. As for the Fennoscandian ice volume, the peak ice gets thicker and the duration of the ice cover longer with decreasing CO₂ concentrations. With the preindustrial CO₂ concentration of 270 ppm, the maximum ice thickness is of the same magnitude (1600 to 1800 m) as during the simulated glaciations in the past 126 kyr. Increase of CO₂ to 300 ppm reduces the predicted maximum ice thickness by about 30-50%.
Figure 19. Fennoscandian ice volume (a) and Olkiluoto ice thickness (b). The simulated ice values with constant CO₂ concentrations are in purple, black, blue, light blue, green and grey curves and match concentrations 270 ppm, 300 ppm, 340 ppm, 380 ppm, 400 ppm and 500 ppm respectively.

The simulated spatial extent of the ice sheet under the different constant CO₂ forcings is shown in Figures 20-22 at three different times, i.e. 20 kyr AP, 55 kyr AP and 100 kyr AP, respectively. In 20 kyr AP (Figure 20), Olkiluoto is outside the ice sheet in all runs; however, in the runs with the lowest CO₂ concentrations (270 ppm and 300 ppm) the site is very close to the ice sheet margin. Given the uncertainties in predicting the exact location of the margin (as demonstrated in Chapter 2), it cannot be ruled out that ice could cover Olkiluoto in 20 kyr AP if the atmospheric CO₂ concentration dropped close
to its preindustrial value. For CO\textsubscript{2} concentrations equal to or higher than 340 ppm the ice sheet growth is mostly limited to the Scandinavian mountains and latitudes above 65° N. Consistent with Figure 19, ice has almost completely vanished from Fennoscandia when the CO\textsubscript{2} concentration is 500 ppm.

Figure 21 shows the ice sheet extent for the second potential future glaciation in 55 kyr AP. Due to a larger drop than at 20 kyr AP (see Figure 18) in the summer time NH insolation, the ice sheet is now larger and the ice sheet margin reaches Olkiluoto even with 340 ppm of CO\textsubscript{2}. Note, however, that at this CO\textsubscript{2} concentration, the model predicts only a very shallow ice sheet over Olkiluoto, as is evident from Figure 19b. Again, it should be remembered that the model prediction of the exact location of the ice sheet margin is uncertain, and thus a much thicker ice sheet could also cover Olkiluoto under 340 ppm of CO\textsubscript{2}. As in 20 kyr AP, CO\textsubscript{2} concentration of 500 ppm does not produce ice sheet to the Fennoscandian area. In 100 kyr AP (Figure 22), the simulated ice sheet extents are very similar to those in 55 kyr AP (Figure 21), and will therefore not be discussed further.

As expected, the constant CO\textsubscript{2} concentration runs agree closely with the corresponding simulations presented in Pimenoff et al. (2011) for 280 ppm and 400 ppm. Comparison to other previous studies is difficult due to highly different model approaches and CO\textsubscript{2} concentrations used. In fact, the authors are not aware of any other study that would be directly comparable to simulations presented here. Cochelin et al. (2006) performed 100 kyr simulations into the future with green MPM model, which covers the latitudes between 75° S and 75° N. They find that for constant CO\textsubscript{2} concentrations of 280 ppm or higher, the start of the next glaciation is postponed to around 50 kyr BP. However, their model predicts continuous ice volume growth beyond 100 kyr BP, which is likely to be unrealistic given the changes in the insolation during this period. Loutre and Berger (2000) used the LLN 2-D model, which simulates only 2 spatial dimensions (i.e. altitude and latitude), to predict that under constant 290 ppm CO\textsubscript{2} concentration there is no glaciation in 20 kyr AP and that the NH ice volume reaches a maximum of one million km\textsuperscript{3} around 60 kyr AP and 110 kyr AP. The 3-dimensional CLIMBER-2–SICOPOLIS simulation with CO\textsubscript{2} concentration of 300 ppm produces clearly more ice than LLN 2-D: First, the CLIMBER-2 simulation produces a baseline NH ice volume of 4 million km\textsuperscript{3} during interglacials. Furthermore, it predicts small glaciation already in 20 kyr AP (NH ice volume of 8 million km\textsuperscript{3}) and large glaciation peaks with over 10 million km\textsuperscript{3} of ice around 60 kyr and 110 kyr AP.

Based on the simulations in this chapter, it is concluded that ice formation at Olkiluoto is unlikely if the CO\textsubscript{2} concentration is 340 ppm or higher. While direct comparison to earlier studies is difficult, the simulations conducted here confirm that the next NH glaciation can be postponed beyond 20 kyr, or even further, if the atmospheric CO\textsubscript{2} concentration remains high.
Figure 20. Modelled ice thickness with a constant CO$_2$ concentration at 20 kyr AP.
Figure 21. Same as in Figure 20, but at 55 kyr AP.
Figure 22. Same as in Figure 20, but at 100 kyr AP.
3.3 Varying CO₂ scenarios

3.3.1 Defining of the future CO₂ emission scenarios

As outlined in Section 3.1, the atmospheric CO₂ concentrations have increased significantly since the preindustrial time, and are projected to keep increasing also in the future. While exact prediction of the future CO₂ concentrations is impossible (due to unknown anthropogenic emissions in future and uncertainties in our understanding of the carbon cycle feedbacks), several CO₂ emission and concentration scenarios have been developed to enable future climate simulations in support of planning of climate mitigation and adaptation. These scenarios try to account for the plausible range of development pathways in terms of e.g. technological advances, global economic equality, and climate treaties in the coming centuries.

Representative Concentration Pathways (RCP) were developed for the Coupled Model Intercomparison Project 5 (CMIP5) that provided the core climate projections for the Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5) (IPCC, 2014). The four RCPs are named after their approximate total radiative forcing in or shortly after year 2100 with respect to preindustrial time (Figure 23). This total forcing accounts not only for increased CO₂ emissions but for all anthropogenic impacts on the climate system, e.g., also other greenhouse gases, aerosol particles and land use change. RCP8.5 (i.e. radiative forcing of ~8.5 W/m²) represents a high-end range of non-climate policy scenarios and results in atmospheric CO₂ concentrations above 900 ppm by year 2100. At the other end, RCP2.6 represents the low-end range of scenarios and assumed strongly declining CO₂ emissions after year 2020 and negative emissions (e.g. by carbon capture and storage) by end of the century. In this scenario, the CO₂ concentration peaks around year 2050 and declines to ~420 ppm by 2100. The two intermediate RCPs represent middle-range scenarios where the CO₂ emissions peak around (RCP4.5) or slightly after (RCP6.0) year 2050 and then slowly decline. In these two RCPs, the CO₂ concentrations in year 2100 are approximately 540 ppm and 670 ppm, respectively. Further information about the RCPs can be found in van Vuuren et al. (2011) and about their extension to year 2500 in Meinshausen et al. (2011).

Instead of emissions, the RCPs provide directly time series of atmospheric CO₂ concentrations that are straightforward to implement into the CLIMBER-2–SICOPOLIS model system. The RCPs can be used to estimate concentrations until year 2500 AD (500 AP). For the period 2,500–12,000 AD (500–10,000 AP), the CO₂ concentration were estimated based on results in Eby et al. (2009), who investigated 10,000-year-long scenarios with the Earth System Climate Model of the University of Victoria (version 2.8, UVic). Unlike CLIMBER-2, the UVic model includes a coupled treatment of the carbon cycle. The UVic consists of an ocean circulation model coupled to a dynamic-thermodynamic sea-ice model and an energy-moisture balance model of the atmosphere (Weaver et al. 2001) as well as an ocean carbon cycle model (Schmittner et al. 2008) and a land surface, terrestrial vegetation and land carbon flux model (Meissner et al. 2003). Eby et al. (2009) showed that the model is able to reproduce the recent evolution of atmospheric CO₂ and surface air temperature as well as predicts relatively well the global carbon budgets of the recent decades. The UVic model was also able to simulate most features of the observation-based air-sea CO₂ exchange.
For their 10,000-year simulations into the future, Eby et al. (2009) forced the model with historical CO₂ emissions between years 1,800 and 2,000 AD. From 2001 to 2100 AD, the CO₂ emissions were obtained by scaling the IPCC SRES A2 emission scenario (IPCC, 2007), and from 2101 to 2300 AD the emissions were let to linearly decline to zero. From 2300 AD onwards, only the natural oceanic and terrestrial sinks of CO₂ affected the atmospheric CO₂ concentration. In the different scenarios, the total carbon emissions between years 2001 and 2300 varied from 160 to 5,120 Gton. The simulated evolution of the atmospheric CO₂ concentration is depicted in Figure 24, which shows that after the concentration peak in just prior to 2300 (~300 AP), the concentrations start to decline rapidly. However, the time in which the terrestrial ecosystems and the oceans absorb most of the atmospheric carbon increases with increasing emissions: For the smallest emissions (160 Gton C) the amount of extra CO₂ remaining in the atmosphere is 30% after 500 years, and 15% after 10,000 years. For the highest emissions (5,120 Gton C) the amount of extra CO₂ remaining in the atmosphere after 500 years is 70%, and after 10,000 years 30%. During the first few centuries, the simulated carbon uptake is relatively rapid due to the increased terrestrial uptake (about 400 Gton C) and rapid surface ocean dissolution (about 900 Gton C). However, once these carbon sinks become saturated, the rest of the uptake is due to the slow processes of deep-ocean transport and carbonate dissolution.

The CO₂ time series from Eby et al. (2009) simulations (illustrated in Figure 24) were used to estimate the rate of CO₂ decline as a function of time and peak CO₂ concentration. These decline rates were then applied to the concentrations obtained from the four RCP scenarios for year 2500 AD. The Eby et al. (2009) simulations covered only the first 10,000 years AP, and thus could not be used beyond 12,000 AD (i.e. 10,000 years AP). From that year onwards, a regression equation developed in this study, and described in detail below, was used.
Figure 23. Total radiative forcing (anthropogenic plus natural) for RCPs,—supporting the original names of the four pathways as there is a close match between peaking, stabilization and 2100 levels for RCP2.6 (called as well RCP3-PD), RCP4.5 & RCP6, as well as RCP8.5, respectively. Figure from Meinshausen et al., (2011). Blue is RCP2.6, orange RCP4.5, grey RCP6.0 and red RCP8.5.

Figure 24. Temporal evolution of the atmospheric CO₂ concentration according to Eby et al. (2009) UVic experiments. Black shows emission of 160 Gton C, red 640 Gton C, green 1,280 Gton C, blue 2,560 Gton C, light blue 3,840 Gton C and magenta 5,120 Gton C.
CO\textsubscript{2} is a potent greenhouse gas, and changes in its atmospheric concentration are known to impact the global mean temperature. On the other hand, the global temperature also affects the CO\textsubscript{2} concentration through impacts on the carbon cycle. Measurements on ice cores have shown that the CO\textsubscript{2} concentration has varied in step with glacial/interglacial cycles, so that during warm interglacials the concentration is typically ~280 ppm and during cold glacials ~180-200 ppm (Barnola et al. 1987; Petit et al. 1999). However, the driving force behind this CO\textsubscript{2} variation remains unidentified (Jansen et al. 2007). The good correlation between the global mean temperature and the atmospheric CO\textsubscript{2} concentration makes it possible to seek a regression equation for CO\textsubscript{2} using sea surface temperature (SST) as a predictor.

Figure 25 shows the CO\textsubscript{2} concentration (red) derived from Vostok (Petit et al.1999) ice core and the simulated SST (black) from 400 kyr BP onwards. It is evident that SST and CO\textsubscript{2} have a strong positive correlation. Using this data, a regression model was derived for the CO\textsubscript{2} concentration as a function of the simulated SST. The regression equation is

\[
\text{CO}_2 = 34.2 \times (\text{SST} - 16.2) + 201.0
\]  

The result of the regression equation (1) is shown in the Figure 26 with the Vostok CO\textsubscript{2} concentration in the past years simulations. The regression model gives good agreement with the CO\textsubscript{2} values from the Vostok ice core. Encouraged by this, it was applied in calculation of CO\textsubscript{2} concentrations in future scenario simulations with CLIMBER-2–SICOPOLIS. However, since the regression equation was derived for the CO\textsubscript{2} concentration range of 150-300 ppm, it could be used only once the CO\textsubscript{2} level had declined below 300 ppm. Therefore, immediately after 10,000 AP (end of period when decline rate was determined based on Eby et al. (2009) simulation results), the CO\textsubscript{2} decline rate was assumed to be no more than 5 ppm per millennium. This rate was estimated from Figure 24 of this report and Table 5 presented in Eby et al. (2009).

As a summary, CLIMBER-2–SICOPOLIS model simulations of the next 200 kyr were undertaken using the following CO\textsubscript{2} concentration scenarios: For the first 500 years, the CO\textsubscript{2} concentrations were taken from the extended RCP scenarios outlined in Meinshausen et al. (2011). For simulation years 500–10,000, the rate of CO\textsubscript{2} decline was taken from the simulations of Eby et al. (2009). After 10,000 years, the CO\textsubscript{2} concentration was calculated from Equation (1); however, the decline rate was limited at maximum to 5 ppm per 1,000 years. The results from these simulations are discussed in the following section.
Figure 25. Simulated SST (black) and Vostok CO₂ concentration (red) in model simulation from 400 kyr BP to present time.

Figure 26. CO₂ from regression equation (Equation 1) (black) and from Vostok (green) in model simulation from 400 kyr BP to present time.
3.3.2 Results

Figure 27 shows the CO$_2$ concentrations (orange, red, magenta and purple lines) and the Fennoscandian ice volume (black, blue, turquoise and green lines) in CLIMBER-2–SICOPOLIS simulations with time-varying future CO$_2$ concentrations. Separate simulations following each of the four RCP scenarios for years 2,000 – 2,500 are presented. During the first 50, 100, 150 and 250 years of the future simulations, the CO$_2$ concentration increases rapidly and peaks at 456, 583, 809, and 2642 ppm for RCP2.6, RCP4.5, RCP6.0, and RCP8.5, respectively (Meinshausen et al. 2011). After the peak, the concentration decreases fast due to rapid uptake of CO$_2$ in the oceans and terrestrial ecosystems. In 120 kyr AP, the CO$_2$ concentration has declined to almost the same level in all the model runs. The arrows in Figure 27 indicate the approaches that have been used to define the CO$_2$ concentration in different time periods after year 2500 (see Section 3.3.1 for details).

The onset and the magnitude of the Fennoscandian glaciation events in the next 120 kyr are strongly dependent on the CO$_2$ emission pathway of the next 500 years (as represented by the different RCPs). For the high-end scenario RCP8.5, there is no glaciation before 100 kyr AP. For the other RCPs, glaciation takes place already around 20 kyr but the volume of the ice sheet remains clearly lower in the intermediate scenarios RCP4.5 and RCP6.0 than in the low-end scenario RCP2.6. Based on these simulations, the onset of glaciation requires the CO$_2$ concentration to decline to approximately 300 ppm. After 120 kyr, when the CO$_2$ concentration has declined to the same level in all the simulations, there is no difference between the simulations in terms of Fennoscandian ice volume.
Figure 27. Fennoscandian ice volumes with different CO₂ concentrations. Orange, red, magenta and purple lines are CO₂ scenarios, black, blue, light blue and green lines Fennoscandian ice volumes. The arrows indicate the approaches which have been used to determine the CO₂ concentration in different time periods (see Section 3.3.1).

Figure 28 illustrates the time series of Olkiluoto ice thickness from present to 200 kyr AP in the four simulations. The atmospheric CO₂ concentration is again shown for comparison. The model predicts that the ice extends far enough south to reach Olkiluoto when the atmospheric CO₂ concentration has dropped below about 280 ppm. RCP2.6 is the only scenario that predicts glaciation in Olkiluoto already in ~20 kyr AP. On the other hand, in the intermediate scenario runs RCP4.5 and RCP6.0, the ice sheet covers Olkiluoto for the first time around 50 kyr AP. In RCP8.5 run, the onset of significant ice formation is delayed until ~130 kyr AP. It is noteworthy, that after 50 kyr AP the durations of the glaciation events at Olkiluoto are the same in all the low-end and intermediate RCP scenario runs.
Figure 28. Olkiluoto ice thicknesses with different CO$_2$ concentrations. Orange, red, magenta and purple lines are CO$_2$ scenarios, blue, green and black lines Olkiluoto ice thicknesses.

Figure 29 shows maps of the Fennoscandian ice thickness from the four model simulations. Each row illustrates the situation during a specific glaciation event (20 kyr, 55 kyr, 100 kyr and 190 kyr AP), and each column shows the corresponding ice extent and height according to one of the CO$_2$ scenarios. As already seen in Figures 27 and 28, the first glaciation in 20 kyr AP is highly sensitive to the atmospheric CO$_2$ concentration. In the intermediate scenarios RCP4.5 and RCP6.0, the ice sheet remains small and Olkiluoto is outside the ice margin. In the high-end scenario RCP8.5, the Fennoscandian ice sheet disappears completely. The next two glaciations are stronger: In 55 kyr AP, the ice sheet covers the whole of Fennoscandia in the low-end and intermediate scenarios, while there is only very little ice in the Scandinavian mountains in the high-end scenario RCP8.5. In 100 kyr AP, ice covers all or most of Fennoscandia in all scenario runs; however, the extent and maximum thickness in RCP8.5 is clearly lower than in the other three RCP runs. By 190 kyr, the differences in the simulated CO$_2$ concentrations have disappeared and the four runs result in almost identical ice thickness maps.

Figures 30 and 31 show for RCP4.5 and RCP8.5, respectively, the time series of a) annual temperature, b) annual total precipitation and c) ice thickness around Olkiluoto. The figures depict the two-meter temperature and surface precipitation from CLIMBER-2 in the model gridpoint containing Olkiluoto, as well as from the statistical downscaling method GAM (introduced in Section 2.4) averaged over nine nearest gridpoints to Olkiluoto. Since the downscaling provides the climate variables in the same grid as SICOPOLIS, it gives a much better spatial resolution than CLIMBER-2.
and thus the downscaled values can be valuable in further applied modelling work for the safety analysis for the Olkiluoto repository. The error bars in the figures depict two standard deviation anomalies about the mean value. Panel a shows for comparison the surface temperature from SICOPOLIS averaged over nine nearest gridpoints around Olkiluoto. The ice thickness in panel c is also taken from SICOPOLIS.

Figure 30a shows that in the RCP4.5 run GAM gives nearly the same air temperature values as CLIMBER-2 when Olkiluoto is ice-free but is clearly colder when there is ice at the site (ice thickness in Olkiluoto region is shown in Figure 30c). The SICOPOLIS temperature is lowest of the three, but as it is the annual mean surface temperature, it is not directly comparable to the 2-metre temperatures from CLIMBER-2 and GAM. On the other hand, precipitation given by the GAM downscaling (Figure 30b) is higher compared to CLIMBER-2 when there is no ice, but lower during glacial periods. These substantial differences between GAM and CLIMBER-2 show the added value from downscaling for other modelling applications (e.g. ice sheet, permafrost) related to the Olkiluoto safety analysis.

It should be remembered, however, that caution must be used when applying GAM to future conditions that deviate from the climate data used to derive the GAM model (Korhonen et al. 2014). For example, RCP8.5 run (Figure 31) produces at times warmer and wetter climate compared to the GAM training data, and therefore the current GAM may not be suitable to downscale this RCP scenario. While applying GAM to the RCP8.5 run leads to reasonable looking temperature and precipitation time series (Figure 30a and b), there are also clear differences compared to the RCP4.5 run whose climate is within the GAM training data (Figure 30): for example, in the RCP8.5 run the downscaled GAM temperature is almost always several degrees lower than the CLIMBER-2 temperature, whereas in RCP4.5 the two deviate mostly only during the glacial periods. Whether the climate features from downscaling RCP8.5 are realistic or an artefact from using the GAM outside its validity range requires further investigation.

Based on these simulations the following is concluded: First, the amount of carbon that human activities release into the atmosphere in the next few centuries can impact the onset and spatial extent of future glaciations tens of thousands of years from now. Therefore, uncertainties in these future emissions seriously restrict our capability to predict the future glaciatic cycles. Second, while the constant CO$_2$ concentration simulations (Section 3.2) are useful for investigating the general features of the future glaciaation responses to CO$_2$, they do not give fully consistent results with the more realistic time-varying CO$_2$ simulations (Section 3.3.2). For example, in 100 kyr AP the simulation with constant CO$_2$ concentration of 300 ppm gives an ice thickness of ~800 m over Olkiluoto (Figure 20b); however, the RCP8.5 simulation predicts only ~400 m of ice over Olkiluoto at this time although the atmospheric CO$_2$ concentration has dropped to approximately 280 ppm (Figure 29). Therefore, simulations accounting for realistic carbon cycle in the climate system are needed.

The simulations presented here are one of the first attempts to estimate the impact of time-varying CO$_2$ concentrations on future glaciations on the time scales of several hundreds of thousands of years. Recently, Herrero et al. (2014) used two simple relaxation models to predict the time evolution of the global ice volume, atmospheric
CO₂ concentration and the extent of the Antarctic ice sheet. According to the relaxation models, these three variables tend exponentially to their reference states with characteristic times that depend on the model used. The two models were calibrated using historical data. Using this approach, Herrero et al. (2014) find a slower decline of atmospheric CO₂ concentration after year 50 kyr AP (~1 ppm/millennium) than used above (5 ppm/millennium). This suggests that sensitivity tests of the decline rate could be useful in the CLIMBER-2–SICOPOLIS system.

Figure 29. Maps of Fennoscandian ice thickness with different CO₂ scenarios at 20 kyr, 55 kyr and 100 kyr AP and 190 kyr AP.
Figure 30. Results from simulation where CO₂ concentration follows RCP45. a) Olkiluoto annual 2 meter temperature and standard deviation from CLIMBER-2 (green) and GAM (black) and SICOPOLIS surface temperature (blue). b) Annual total precipitation from CLIMBER-2 (green) and GAM (black). c) Ice thickness from SICOPOLIS.
Figure 31. Same as Figure 30 but CO₂ concentration follows RCP85 scenario.
4 SUMMARY AND CONCLUSIONS

This report summarizes the climate model simulations performed at the Finnish Meteorological Institute (FMI) in support of formulating climate scenarios for the safety analysis of the Olkiluoto repository. The simulations were run with an earth system model of intermediate complexity (EMIC) CLIMBER-2–SICOPOLIS recently installed at the FMI. Previous studies have shown that despite its low spatial resolution, the CLIMBER-2–SICOPOLIS modelling system is complex enough to capture the essential climate processes and feedback over millennial timescales. The work builds on the report of Pimenoff et al. 2011 but utilizes an updated CLIMBER-2 – SICOPOLIS model system that is run in-house at FMI.

First, several sensitivity tests were undertaken to investigate the robustness of the model results to changes in the calculation platform, model initialization time, geothermal heat flux, time lag of the relaxed asthenosphere, and ice sheet bottom sliding parameter. The sensitivity tests showed that the modelled ice volumes over the northern hemisphere (NH) and Fennoscandia are not highly sensitive to the factors and model parameters tested in this study. The largest effect was found when the sliding parameter was multiplied by 4/3 or 2/3; in these simulations the NH and Fennoscandian ice sheet volumes deviated at most by about 23 and 15%, respectively, from the baseline run. On the other hand, the ice thickness at Olkiluoto was found to be highly sensitive to the model set-up when Olkiluoto was located close to the ice sheet margin. The low resolution of the EMIC, together with uncertainties in some of the model parameter values, means that the exact location of the ice sheet edge cannot be simulated reliably. In other words, reasonably small changes in the model parameter values may cause large changes in the predicted ice properties at sites close to the ice sheet margin.

Next, the CLIMBER-2–SICOPOLIS simulations of the historical period were compared against a detailed semiempirical land uplift model of Påsse (2001). The best fit between the two models was obtained when the time lag of relaxed asthenosphere was set to 4,000 years, instead of the previously used 3,000 years. A comparison to reconstruction data of the Fennoscandian ice sheet revealed that CLIMBER-2–SICOPOLIS tends to underestimate the ice extent over the Kara Sea during the glaciations 90 kyr and 60 kyr BP. Furthermore, during the former glaciation, the model predicts the ice sheet to extend over Southern Finland and Sweden, which is in contrast with the reconstructions. However, CLIMBER-2–SICOPOLIS reproduces well the maximum ice extent of the last glacial maximum ~20 kyr BP, for which the reconstructions are the most reliable. This gives us confidence in the performance of the modelling system.

Finally, CLIMBER-2–SICOPOLIS was used to project future glaciations in the next 200 kyr under different assumptions of atmospheric CO₂ concentrations. Under natural background CO₂ conditions, variations in insolation make NH glaciation possible in 10–20 kyr AP, 50–60 kyr AP, 90–100 kyr AP and 130–140 kyr AP, and at 20-30 kyr intervals after that. However, human activities have released large amount of additional carbon into the atmosphere since the industrial revolution and, as result, increased the atmospheric CO₂ concentration from ~270 ppm to close to 400 ppm within the past 250 years. The CLIMBER-2–SICOPOLIS simulations support earlier findings that sustained
elevated CO₂ concentrations will delay the onset of the next glaciation by several tens of thousands of years.

Assuming idealized future scenarios with constant CO₂ concentrations, CLIMBER-2–SICOPOLIS predicted ice at Olkiluoto within the next 144 kyr only in the simulations with CO₂ equal to or lower than 300 ppm. The timings of these ice events were 50-60 kyr, 95-105 kyr, and 125-135 kyr AP, and the predicted ice thicknesses at Olkiluoto 1,600-1,800 m for CO₂ concentration of 270 ppm, and 800-1300 m for 300 ppm. Furthermore, the simulations showed that the total Fennoscandian ice volume correlates negatively with the atmospheric CO₂ concentration, and the simulation with 500 ppm of CO₂ prevented all glaciation events in Fennoscandia during the next 144 kyr.

More realistic simulations with time-varying future CO₂ concentrations were constructed by combining previously defined Representative Concentration Pathways (RCPs) of CO₂ concentration for years 2,000-2,500 AD, atmospheric CO₂ decline rates from a modelling study of Eby et al. (2009) for years 2,500-10,000 AD, and from 10,000 AD onwards CO₂ concentration was calculated using a new regression equation developed in this study; however, the decline rate was limited to 5 ppm per 1,000 years at maximum. The regression equation was motivated by ice core data that show a clear positive correlation between atmospheric CO₂ concentration and global mean temperature over the past 400 kyr. A regression equation was fitted for CO₂ using the slow-varying sea surface temperature (SST) as a predictor, and showed it to be able to reproduce the historical CO₂ record with good accuracy.

Using the time-varying CO₂ concentrations, a set of 200,000-year-long simulations into the future were conducted. The simulations showed that the onset of a glaciation at 10–20 kyr AP is unlikely due to high atmospheric CO₂ conditions. For this time period, significant increase in Fennoscandian ice volume was simulated only for the RCP2.6 scenario, which assumes strongly declining CO₂ emissions already after year 2020 AD and negative emissions by the end of the 21st century. On the other hand, glaciation was predicted in ~100 kyr AP in all simulated scenarios, and in ~50 kyr AP in all but RCP8.5 scenario. RCP8.5 is a scenario with no climate policy and therefore with strongly increasing greenhouse gas emissions throughout the 21st century. The scenario runs predicted that ice extends far enough south to reach Olkiluoto when the atmospheric CO₂ concentration has dropped below about 280 ppm. Therefore, the low-emission scenario RCP2.6 predicts glaciation in Olkiluoto already ~20 kyr AP, while the two intermediate-emission scenarios RCP4.5 and RCP6.0 show ice sheet cover at the site for the first time ~50 kyr AP. In the high-emission run the onset of significant ice formation is delayed until ~130 kyr AP. The maximum ice thicknesses in Olkiluoto are in all the scenarios between 1,800 and 2,000 m, except for the ~100 kyr AP glaciation in the RCP8.5 scenario, which predicts a short-duration 400-m thick ice pulse in Olkiluoto.

The simulations with CLIMBER-2–SICOPOLIS highlight that the future concentration pathways of atmospheric CO₂ can have large impacts on the ice formation over Olkiluoto in the next 100,000 years. However, the future emissions from both anthropogenic and natural sources are extremely challenging to foresee due to uncertainties in economical, technological and political developments as well as in the climate-change driven feedbacks in the natural carbon cycle. These uncertainties limit
the capability to predict the future glaciation cycles. The best solution to this limitation is to use a variety of CO$_2$ concentration scenarios, together with a multi-model approach in which the scenarios are simulated with several independent EMICs.
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