



ARBEITSBERICHT NAB 24-14

Geological long-term evolution:
Climate and Iceflow

September 2024

**Nagra | National Cooperative for the
Disposal of Radioactive Waste**

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KEYWORDS

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Nagra Arbeitsberichte ("Working Reports") present the results of work in progress that have not necessarily been subject to a comprehensive review. They are intended to provide rapid dissemination of current information.

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Abstract

This report summarises past and possible future climates and glacier conditions in Northern Switzerland. It provides the geoscientific basis to assess the impact of future climate change on erosion (Nagra 2024a), on transport in the geosphere (Nagra 2024b, 2024e) and biosphere (Nagra 2024d) and ultimately serves the comprehensive description of the post-closure evolution of a geological repository (Nagra 2024c and references therein). The report focuses on glacial climate conditions and related foreland glaciations.

The climate of the Quaternary showed a cyclicity of glacials and interglacials with a ~ 41 -kyr cycle before the Mid Pleistocene Transition (MPT). After the MPT, which occurred between 1.25 and 0.7 Myr ago, the cyclicity changed to predominantly ~ 100 kyr. In the past, multiple glacial advances have occurred from the Alps into the Alpine Foreland in Switzerland. Of particular interest are the last glacial maximum (LGM), which peaked around ~ 24 ka ago, and the most extensive glaciation (MEG) which occurred sometime between 600 and 400 ka ago, with ice extents reaching the siting regions (LGM) or even further north to the Black Forest (MEG).

While past climates can be studied using proxy data, the determination of future climate conditions, especially for long timespans, can only be achieved using global climate simulations. The global climate simulations used here cover the next one million years and also the last ~ 800 kyr so that results of past simulated climates can be compared against known data e.g. from ice cores or pollen assemblages from sediments. The comparison of simulated past climates with proxy data shows that past climates are well reproduced by the simulations and, hence, future climate projections can be regarded as meaningful. Results from the global, long-term climate simulations show that, depending on emission scenarios of greenhouse gases, climate in the future will be significantly warmer than today and the onset of the next glaciation will be postponed. Even using preindustrial atmospheric greenhouse gas concentrations, the next glacial conditions will only peak at ~ 110 kyr with simulated global total ice-volumes below those of the LGM. The higher the estimated future emission of anthropogenic greenhouse gases, the more the next glacial inception will be postponed and, hence, the longer the current interglacial climate conditions will persist. In the most extreme emission scenario, the next glaciation is simulated to only peak after ~ 850 kyr.

Glaciations are of particular importance with respect to repository long-term evolution because the advance and retreat of glaciers in the foreland will alter the landscape in Northern Switzerland and may reduce the overburden of the repository by glacial erosion. Ice-flow simulations have been performed, using temperatures and precipitation of downscaled global climate simulations as inputs. Results of these simulations indicate temperate basal conditions below much of the glacier advancing onto the Alpine foreland during the LGM enabling sliding and, thus, providing conditions conducive to glacial erosion. However, the results also indicate a poor fit between the sliding distance, a proxy for glacial erosion by moving ice, and the location of existing overdeepened valleys. Calculation of the time-integrated flow accumulation area based on maps of hydraulic potential, a proxy for erosion by subglacial water flow, yields zones with significant subglacial water flow that favour enhanced sediment removal and increased fluvial erosive action. This therefore shows the importance of subglacial water flow in the formation of overdeepenings in the distal Swiss lowlands.

Geological field evidence and results of the ice-flow simulations show that two of the three possible repository sites were covered by ice during the last glacial maximum. It is therefore likely that, given glacial climate conditions, repository sites will in the future again be covered by glaciers, should they reach the lowlands of Northern Switzerland. Ice thickness of $\sim 200 - 400$ m and a duration of ice occupation of $\sim 1'000 - 3'000$ yr in the siting regions were simulated and are expected for future foreland glaciations.

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List of Acronyms

AIF	Alpine Ice Field
AMOC	Atlantic Meridional Overturning Circulation
a.s.l.	above sea level
CESM	Community Earth System Model
CFC	Chlorofluorocarbon
DEM	Digital Elevation Model
DGR	Deep Geological Repository
ESM	Earth System Model
EMIC	Earth System Model of Intermediate Complexity
GI	Glacial Index
HE	Heinrich Event
HLW	High-Level Waste
IPCC	Intergovernmental Panel on Climate Change
IQR	Interquartile Range
LGM	Last Glacial Maximum
L/ILW	Low- and Intermediate-Level Waste
MEG	Most Extensive Glaciation
MIS	Marine Isotope Stage
MPT	Mid-Pleistocene Transition
NHIS	Northern Hemisphere Ice Sheet
PDD	Positive Degree-Day
PgC	Petagram Carbon
PI	Pre-Industrial
PISM	Parallel Ice Sheet Model
RCM	Regional Climate Model
SIA	Shallow Ice Approximation
SRTM	Shuttle Radar Topographic Mission
SSA	Shallow Shelf Approximation
WRF	Weather Research and Forecasting

1 Introduction

This report summarises past and possible future climate and glacier conditions in Northern Switzerland. It provides the geoscientific basis to assess the impact of future climate change on erosion (Nagra 2024a), on transport in the geosphere (Nagra 2024b, 2024e) and in the biosphere (Nagra 2024d) which is used for the analysis of scenarios of long-term evolution of a geological repository (Nagra 2024c). Because of the potential for rapid and deep erosion during glacial periods, the report has a strong focus on glacial climate conditions and related foreland glaciations.

The report is organised as follows:

- Chapter 1 introduces the topic of climate evolution in the framework of the long-term evolution of a geological repository in Northern Switzerland. It also discusses the relationship between climate and geomorphologic processes with a focus on glacial erosion. Furthermore, the key factors driving climate evolution are summarised.
- Chapter 2: Because the Quaternary period (~ 2.58 Ma until today) forms the basis for estimations of the (deep) future climate, this chapter gives an integrated summary of the Quaternary climate. Here, it is shown that glacial conditions are most relevant with respect to scenarios for deep erosion. Accordingly, it introduces the link between climate, climate modelling, and ice flow modelling to assess erosion proxies in the long-term evolution. It also includes results of past climate simulations and compares them to known data from Quaternary climates.
- Chapter 3 builds on the argumentation from Chapter 1 using results introduced in Chapter 2. Accordingly, it applies results from regional climate modelling to ice-flow modelling and provides a characterisation of glacier conditions in Northern Switzerland. Proxies for erosion by moving ice and subglacial water flow are introduced and a sensitivity analysis is provided to allow a discussion of the main climatic and glaciological effects on the erosion potential.
- Chapter 4 summarises the climate models derived for the future one million years. A brief description of the climate states is added.
- The report ends with a conclusion and summary in Chapter 5.

1.1 Relevance of climate evolution for the long-term evolution of a geological repository

The geology and geomorphology around the repository will be subject to climatic forcing that drives erosional processes. Erosion, in general, reduces the repository overburden and might – over very long timescales – lead to decompaction effects and reduction of the self-sealing capacity of the host rock. To ensure the stability of the geological barrier over the period of consideration for high-level nuclear waste of one million years, such climate-driven processes need to be considered.

The Quaternary climate evolution is characterised by glacial/interglacial cycles. Especially, the change from a predominant ~ 41-kyr cycle to the establishment of a ~ 100-kyr cycle during the Mid-Pleistocene Transition (MPT; Clark et al. 2006) impacted Quaternary landscape evolution. This cyclicity modulates regional climate parameters such as temperature and precipitation, which in turn influence the water and sediment discharge and hydrogeological evolution. Large foreland glaciations can change sediment routing and drainage pathways or erode overdeepened valleys to considerable depths (e.g. Preusser et al. 2010, Häuselmann et al. 2007). In addition, mechanical loading of the host rock by the ice cover, as well as subglacial infiltration, might impact the hydraulic system (Lemieux et al. 2008, Neuzil 2012). Large foreland glaciations are accompanied

by extensive, possibly deep-reaching permafrost which might impact the higher regional aquifers (Malm aquifer and above; e.g. Harris et al. 2009). Note that also the biosphere model used for dose calculations depends on climate parameters.

As a benchmark, climate simulations for future climate projections must be capable of also simulating past climates. Proxy data and measurements from the Quaternary period are, hence, compared to climate simulations for the geological past. The short-term climate simulations (for the next ~ 80 yr) performed by the IPCC (e.g. (Pörtner et al. 2023)) are insufficient to discuss the long-term evolution of a deep geological repository. Long-term future climate simulations (simulations for the "deep future") have thus been carried out for this report to assess the future climate evolution including the timing of possible future glaciations.

1.2 Climate as a geomorphic process driver

Climate-driven erosion and deposition can cause significant changes of the Earth's surface if cumulated over sufficient time periods. The rates at which these changes occur may vary over several orders of magnitude at different locations (Selby 1985). Examples of such changes are erosion and sedimentation due to fluvial processes in a river system, glacial erosion causing the over-deepening of subglacial valleys or isostatic rebound due to unloading of the crust by e.g. the melting of ice sheets (e.g. SKB 2010).

While fluvial erosion is often observable in the short term (though it can continue over many thousands of years) and depends strongly on the amount and intensity of precipitation as well as on topographic factors, isostatic rebound, after an initial faster phase following deglaciation, occurs rather slowly over thousands or tens of thousands of years and depends on, among other factors, the growth and decay of ice sheets and glaciers (which in turn are determined by time-varying regional patterns of temperature and precipitation), and the rheological characteristics of the lithosphere and asthenosphere (McConnell 1968, Hagdorn 2003). Thus, recurring patterns of climate change and alterations in the landscape are closely interlinked. A change in climate may cause increased or decreased precipitation/evapotranspiration and, in more extreme cases, an increase or decrease in the volume and distribution of ice sheets, ice caps and glaciers (Hagdorn 2003, SKB 2010, Näslund et al. 2013). Both the persistence of particular climatic conditions and changes in climate with time will inevitably lead to various changes in the landscape expressed on timescales ranging from a few hours (for a single extreme weather event) to cumulative effects over many thousands or even millions of years.

In the context of a deep geological repository for radioactive waste, climate must hence be regarded as a driver for landscape evolution (see also Becker et al. 2014).

1.3 Key factors for climate evolution

Climate refers to long-term average patterns of weather conditions that occur in a particular location. Climate is determined by parameters such as temperature, precipitation, humidity, wind patterns, and other weather-related variables, which can be influenced by a variety of factors, including topography, distance from the coast, insolation or concentration of greenhouse gases. Insolation as a primary energy source, for example, refers to the amount of solar radiation that reaches the Earth at a particular geographic location. The amount of energy received is determined by Earth's orbital parameters of obliquity (Earth's axial tilt), eccentricity (how much Earth's orbit around the sun deviates from a circle) and precession (changes in the orientation of the Earth's rotational axis relative to the plane of the Earth's orbit (axial precession) and the rotation of Earth's orbit (apsidal precession)) and can be calculated accurately for the past (Fig. 2-1) and future (Fig. 4-1). The chemistry of Earth's atmosphere is also one of the key factors. The concentration of greenhouse gases in the atmosphere is influenced by the climate itself (e.g. CO₂ may be removed from the atmosphere by the growth of plants during warmer climates) and recently also by anthropogenic activities (mainly through burning of fossil fuels). Volcanic activity (because of an increase of pyroclastic material and aerosols, such as sulphates, in the atmosphere) and solar radiation (the amount of energy emitted by the sun) are also influencing factors.

Climate, however, does not respond linearly to each of the aforementioned drivers, the response is non-linear and spatially varying because multiple feedback loops, thresholds and tipping points exist. In addition, the drivers themselves are complex, for example, greenhouse gases interact with infrared radiation, heat emitted by the Earth's surface, and the atmosphere itself, leading to a warming effect, but they also interact with incoming solar radiation, leading to a cooling effect. The net balance of this effect is positive, meaning that the Earth warms as greenhouse gases are emitted into the atmosphere. Feedbacks are also complex; global warming for example has a direct impact on ice (ice sheets and sea ice) and snow cover. Ice and snow cover in general increase Earth's reflectivity (albedo) and, hence, decrease the amount of solar energy being absorbed at the Earth's surface. On long timescales, warming due to increasing concentrations of greenhouse gases leads to the melting of large ice sheets, which decreases Earth's albedo and brings large amounts of colder melt water into the oceans. The melt water reduces local ocean temperatures and changes local ocean density potentially modifying ocean circulation, while the reduced surface area of glaciers leads to a reduced albedo, causing more heating. A more detailed description of the different feedback loops in the climate system can be found e.g. in Forster et al. (2021).

It has been shown by e.g. Ganopolski et al. (2016) or Paillard (1998) that the variation of insolation and atmospheric greenhouse-gas concentration are the main drivers for Quaternary climate changes on a global scale. Long-lasting climate simulations of the past (and future) often include only a limited number of climate drivers with insolation and greenhouse gas emissions/ concentration almost always being two of them (see Lunt et al. 2017 for a summary of long-term climate simulations).

2 Quaternary climate

Understanding climates in the past, especially during the Quaternary which began approximately 2.58 million years ago and continues to the present day, is an important basis to project possible future climates. This understanding can be developed by reconstructing the climate in the geological past. However, although precise quantitative measurements of modern climate data such as temperature, precipitation and wind speed exist, these data are only directly available for the past ~ 150 years (e.g. Jones et al. 2012). Before this time, climate reconstructions mainly rely on indirect information from so-called proxies. Proxies are physical or biological representations of climates and include e.g. information from past marine or terrestrial life (pollen, tree rings, marine fossil assemblages etc.), physicochemical variations of isotope ratios (e.g. $\delta^{18}\text{O}$ in microfossils such as foraminifera) or landforms (e.g. valleys carved by glaciers, moraines, frost-created features; e.g. Bradley 2015 and references therein). The precision, accuracy and number of these proxies tends to decrease the further back in time they reach, and information from several proxies is often needed for a more robust determination of past climates at any given time. In addition, some proxies are interpreted as being indicative of climate and its changes on a hemispheric to global scale (e.g. Deuterium data from ice cores) while others are indicative of a much more regional scale (e.g. pollen assemblages; e.g. Jouzel et al. 2007).

To determine climate, and especially climate changes, in the more distant past, it has become standard practise to benchmark climate model simulations against available proxy data (e.g. Braconnot et al. 2012). The results of climate simulations (or time slices from a simulation) are compared with climates derived from proxy data which helps to understand the processes leading to the observed climates but can also be used to test hypotheses (e.g. do the assumptions lead to the observed climates defined by proxy data? e.g. Harrison et al. 2016).

This chapter gives a brief overview of the younger Quaternary climate with special focus on the last several glaciations and the application of climate simulations to past climates.

2.1 Global Quaternary climate

During the Quaternary period, multiple cycles between cold and warm climates occurred, termed glacial and interglacial periods. During glacial periods, continental ice sheets and glaciers grew to a certain maximum extent, which was different for each of the glacials (e.g. Ehlers et al. 2011, Huybers 2007). Glacials are divided into stadials and interstadials where the latter describes warmer periods within a glacial while stadials refer to the colder periods of a glacial. Interglacials are periods of warmer climate with a much-reduced ice-sheet extent (note that Antarctica and Greenland frequently remained covered by ice, even during interglacial periods, e.g. Masson-Delmotte et al. 2010).

The so called Mid Pleistocene Transition (MPT; Clark et al. 2006) that occurred between ~ 1.25 to 0.7 Ma ago, describes a marked change of the cyclicity. After the MPT, the glacial/interglacial cycles typically had a period of roughly 100 kyr, while before the MPT, ~ 41-kyr cycles were prevalent (Past Interglacials Working Group of PAGES 2016 and references therein). This change of periodicity has been deduced from variations of $\delta^{18}\text{O}$ in deep-sea sediments (Clark et al. 2006, Lisiecki & Raymo 2005). Usage of $\delta^{18}\text{O}$ is based on the fact that oxygen has three naturally occurring isotopes, ^{16}O , ^{17}O and ^{18}O where the numbers refer to the atomic mass. ^{18}O is heavier than ^{16}O (because it has two more neutrons) which causes the water molecule incorporating this heavier isotope to vaporise more slowly (and condense faster) than the lighter ^{16}O . The ratio of ^{18}O to ^{16}O (commonly referred to as $\delta^{18}\text{O}$) can be used as a representative time series describing the sea level (and hence also ice volume) and (water-) temperature by combining $\delta^{18}\text{O}$ records from different locations (e.g. Ahn et al. 2017). Fig. 2-1 shows the measured $\delta^{18}\text{O}$ concentration from microorganisms in deep-sea sediments together with variations of orbital parameters of the

Earth (obliquity, eccentricity and precession), the resulting insolation (at 65° N in July), global mean temperature difference, CO₂ content and the estimated ice volume for part of the Quaternary. Some of the information (mean temperature difference and CO₂ content) was taken from ice cores drilled in Antarctica (EPICA ice dome; e.g. Jouzel et al. 2007, Lüthi et al. 2008) while other data come from a global compilation of measurements from marine sediment cores (e.g. Lisiecki & Raymo 2005). Although the information was taken from very different locations and employs very different measurement methods, the results are in general highly correlated: higher CO₂ content leads to higher temperatures, reduced total ice volume and also a reduced $\delta^{18}\text{O}$ (Fig. 2-1).

Data depicted in Fig. 2-1 was used by Lisiecki & Raymo (2005) to define so-called Marine Isotope Stages (MIS's). Generally, even numbered stages represent glacial periods while odd numbers refer to interglacial periods. Fig. 2-1 also shows that the durations and specific climate conditions during the different glacial and interglacial periods varied. In general, the climate during the glacial periods was colder than during the interglacials, leading to a growth and advance of glaciers and subsequently to lower sea levels. The transition from glacial to interglacial periods was often rapid, with changes in temperature and sea level occurring over the course of a few thousand years. The exact timing and duration of each MIS, as well as the climate conditions during each period, are the subject of ongoing research (e.g. Lisiecki & Raymo 2005, Past Interglacials Working Group of PAGES 2016 and see Section 2.4 for a summary of some regional studies).

MIS 1, which began about ~ 14 ka ago and continues to the present day, is the most recent interglacial period. Temperatures during MIS 1 were variable (± 0.5 °C compared to today) until recently (pre-industrial, PI) where a sudden increase of 1 °C and more can be observed (e.g. Marcott et al. 2013). MIS 2, which occurred from about 29 ka to 14 ka ago, was the peak of the last glacial period. In the northern hemisphere, during the so-called Last Glacial Maximum (LGM) which occurred ~ 24 ka ago, large ice sheets covered much of North America and Europe and, as a result, global sea levels were approximately 120 m lower than they are today (Spratt & Lisiecki 2016, Batchelor et al. 2019). As used in this report and worldwide, LGM refers to the last maximum glacier extent. Coldest temperatures in central Europe with the formation of deep-reaching permafrost and vegetation of a cold desert was millennia later and relate to the reduction of the Atlantic Meridional Overturning Circulation (AMOC) after the onset of melting and winter pack ice penetrating to the latitudes of the Pyrenees (see Section 2.2 for details).

Batchelor et al. (2019) used available proxies and model simulation results to infer ice extent in the northern hemisphere during the last glacial/interglacial cycles revealing pronounced differences between the different glacials. Fig. 2-2A shows the available data for the time period of MIS 6 where points represent empirical data and coloured polygons represent outlines of empirical data from different sources (see Batchelor et al. (2019) for more details). Ice extents have been combined to show the maximum ice extent in the respective time slice (e.g. Fig. 2-2B for MIS 6). Black numbers in Fig. 2-2 represent a robustness score for each region. The score is based on the availability of proxy data, where "0" means no data at all were available (and ice extent was estimated from periods with comparable configurations), while "5" means a large number of reliable data were available. The mean of these numbers is represented in the lower left of the panels which can be used as a guideline for the general robustness of the projection (see Batchelor et al. 2019 for more details). These estimations show that during the glacials, large parts of the northern hemisphere were covered by ice; especially the North American and European ice sheets were pronounced, although with different extents during the different glacials. The north-east Asian ice sheet is considerably smaller and less connected in most MIS's. Batchelor et al. (2019) attribute this to the development of the substantial North American ice sheet which changed atmospheric circulation patterns resulting in warming and the observed limited glaciation in north-east Asia.

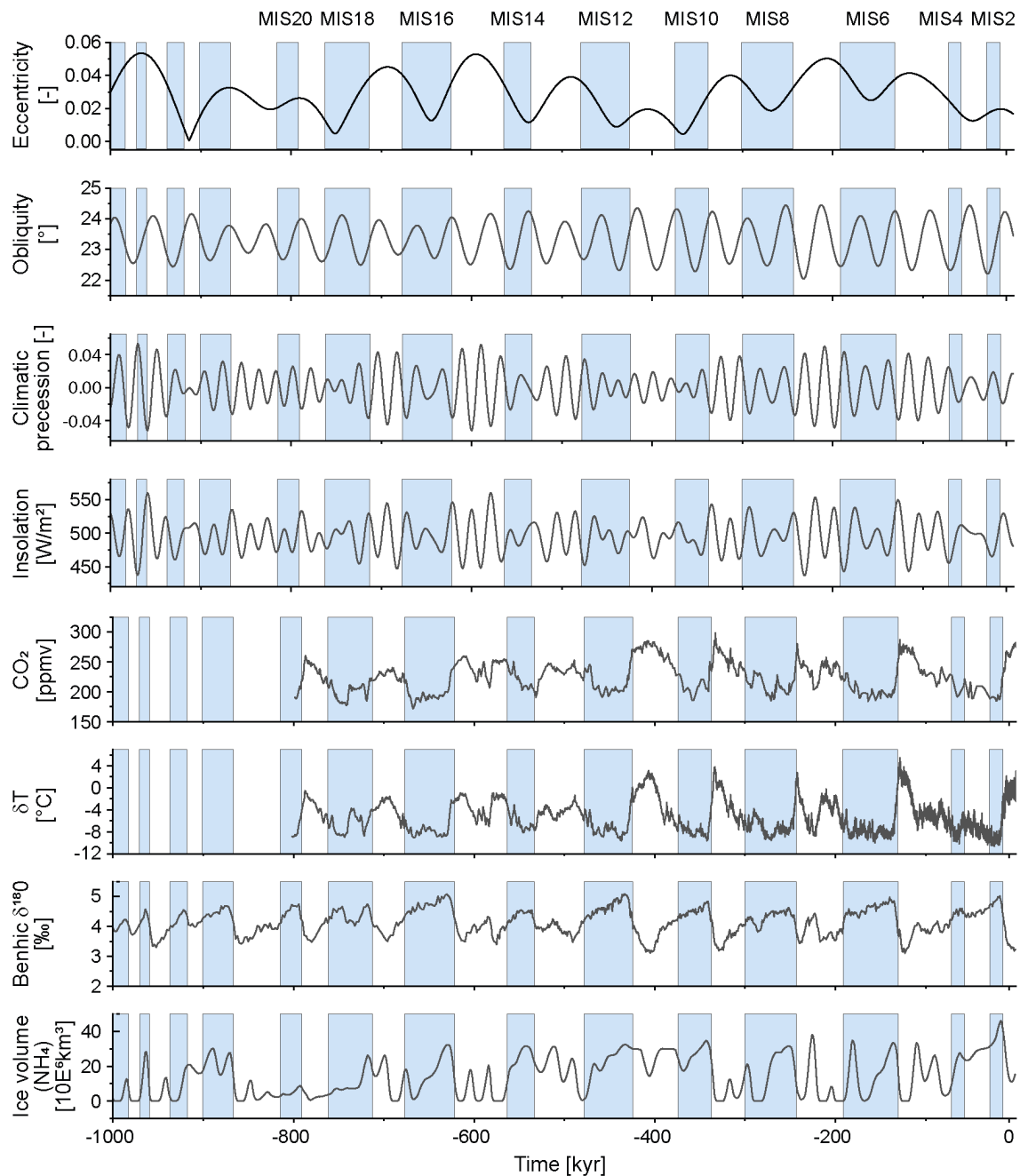


Fig. 2-1: Overview of orbital parameters (eccentricity, obliquity and precession) and resulting insolation, CO₂ concentration, temperature difference, δ¹⁸O and estimated ice volume

Insolation and CO₂ content mainly determine the general climate (e.g. temperature difference – delta T). The general climate determines the occurrence of ice sheets on the continents, which is traced by δ¹⁸O from benthic foraminifera.

Insolation and the respective parameters are given for 65°N (July) according to Berger & Loutre (1991). The CO₂ content follows the determinations from Lüthi et al. (2008). Global temperature differences are given as 100-yr averaged values compared to today and are taken from Jouzel et al. (2007). Both data are based on measurements from the EPICA ice core in Antarctica. δ¹⁸O was determined from sediment cores and can also be used to calculate a rough estimate of the northern hemisphere ice volume (Lisiecki & Raymo 2005).

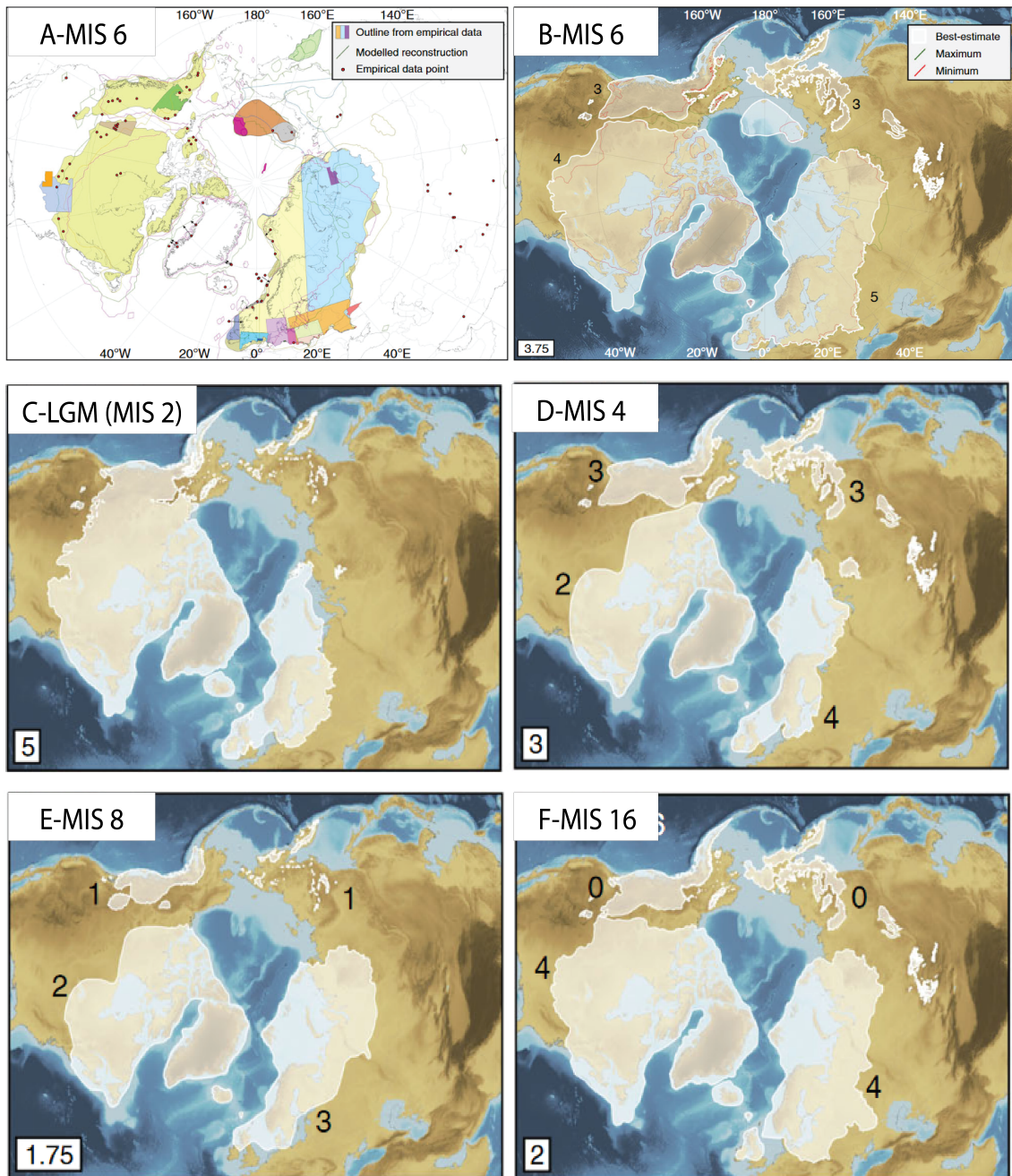


Fig. 2-2: Reconstructions of the Northern Hemisphere Ice Sheet extent during part of the Quaternary (Batchelor et al. 2019)

A) Compilation of data sources for the example of a time slice of MIS 6. B) Maximum, minimum and best-estimate reconstructions of ice-sheet extent during MIS 6 derived from the data shown in A). C) – F) Best-estimate reconstructions of Northern Hemisphere Ice Sheet extent for time slices in the Quaternary. The overall robustness score for each time slice reconstruction is shown in the bottom left corner. Black numbers are individual ice-sheet robustness scores. See text for more details.

2.2 Local Quaternary climate

Although the Quaternary climate on a global scale can be viewed as a general guideline for prevailing local climates, it is necessary to use local proxies to determine climate and climate-related processes for a specific region of interest. Information from local proxies is often based on analyses of, for example, the pollen record in (lake) sediments, or on measurements of $\delta^{18}\text{O}$ from, for example, stalagmites or stalactites in caves (e.g. Luetscher et al. 2021, Skiba et al. 2023).

One important characteristic of the Quaternary climate evolution in Switzerland are the frequent glacier advances and retreats which are also reflected in proxy data. Preusser et al. (2011) compiled, on the basis of proxy data and the sometimes complex sedimentation and erosion history, the different glacial advances and durations in Switzerland (Fig. 2-3).

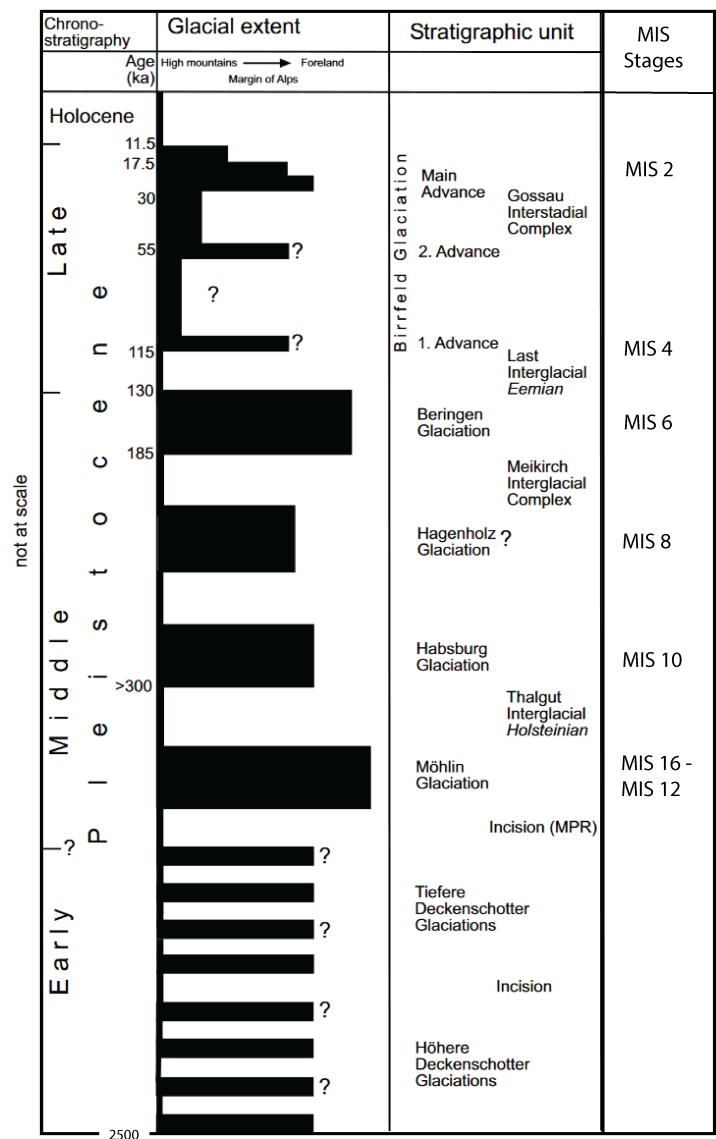


Fig. 2-3: Glaciation history of Switzerland (after Preusser et al. 2011)

The advance and retreat of glaciers and their maximum extents for the Birrfeld, Beringen, Habsburg and Möhlin Glacials in Switzerland are discussed in Preusser et al. (2011) and summarised in Nagra (2024a). The Möhlin Glaciation (Fig. 2-4) is of particular interest as glaciers reached the southern slopes of the Black Forest and it is believed to be the most extensive glaciation (MEG) of the Swiss Alpine foreland during the Quaternary (e.g. Graf 2009). The different glaciers (Rhine, Linth, Reuss and Aare – Valais glaciers) combined and formed a continuous ice cover. During the last glacial maximum (LGM), ice also reached far into the Alpine foreland, with the glaciers forming a relatively continuous ice cover, but the north-western extent was less than during the MEG. Ice extents of the Habsburg, Beringen and Birrfeld Glacials were smaller than during the Möhlin Glacial and, at least during the Habsburg and Birrfeld Glacials, the ice cover is not believed to have formed a continuous ice covers (Preusser et al. 2011).



Fig. 2-4: Estimated ice extent during the Möhlin Glaciation in Northern Switzerland

Figure from Preusser et al. (2011).

Climate in Switzerland during the glaciations was categorised as a polar climate with ice covers and tundra in ice-free terrain during the coldest phases. Polar climate with an ice cover is, next to the ice cover itself, defined by sub-zero temperatures throughout the year. Tundra climate is defined by an average temperature of the warmest month during the year between 0 °C and 10 °C. These low temperatures may also induce permafrost conditions, usually in close proximity to the ice cover. Permafrost refers to a condition where subsurface material remains at negative temperatures (frozen) throughout the year and often over very long periods of time (centuries, millennia and more). Permafrost exists today at cold high latitudes and high altitudes. The fact that deep-reaching permafrost during the past ice ages also existed in wide parts of the European lowlands was documented by the analysis of characteristic deformation features (ice wedge casts) caused by former deep-reaching frost action in loess deposits (Poser 1948, Washburn 1979, Heiri et al. 2014). During the coldest time intervals, permafrost north of the Alps was continuous, cold (mean

annual surface temperature about -5 °C with shorter-term extremes down to -10 °C; Duprat-Oualid 2019) and reached depths of about 100 to 200 m (Haeberli et al. 1984, Frenzel et al. 1992, Delisle et al. 2003, Lindgren et al. 2016). Continuous means that permafrost occurred everywhere and was only absent underneath deep (> 3 m) lakes and rivers. The implication of this is that maximum cooling in Europe during the ice age was at least 12 to 15 °C colder than pre-industrial values and, hence, much stronger than the global scale average (about 5 °C cooling; Annan et al. 2022).

Ice core and ocean floor drilling combined with terrestrial bio-indicators (especially pollen) (Duprat-Oualid 2019) provide detailed information about the sequence and interplay of solar forcing, CO₂ forcing and freshwater forcing in the Atlantic Ocean that play a key role for the considerable time difference between the maximum advance of glaciers and the maximum cooling with permafrost formation in the Alpine foreland. The advance of large piedmont glaciers in the Alpine foreland during times of decreasing solar forcing with a still wide-open Atlantic Ocean culminated around 26 to 22 ka ago (Kamleitner et al. 2023). During this time degradation of permafrost from a colder and dryer time interval around 30 to 28 ka ago took place up to altitudes of about 1'000 m a.s.l. under conditions of thick winter snow as inferred from cryospheric cave deposits in the eastern Alps (Spötl et al. 2021). Minimum atmospheric CO₂ concentrations (close to 200 ppm) were only reached around 20 to 18 ka ago, i.e. millennia later. At this time, solar forcing had already started to increase again. This led to higher summer temperatures, increased meltwater production and partial instabilities of the European and Laurentide ice sheets. The resulting massive input of freshwater (often intermixed with floating ice), first from the Baltic area and then from the Laurentide Ice Sheet, weakened the AMOC (Hodell et al. 2017, Pöppelmeier et al. 2023) – the key factor of the global ocean "conveyor belt". As a consequence, the reduction of warm water advection from the South caused strong sea surface cooling with rapid expansion of winter sea ice on the Atlantic Ocean down to about the latitude of the Pyrenees. Extremely cold and dry winters developed in the Arctic and in Europe at around 18 to 16 ka ago (Heinrich stadial 1; He et al. 2021, Naughton et al. 2023). Sea surface temperatures during this time lowered by about 10 °C and mean continental surface temperatures by about 5 °C with respect to the preceding millennia. The closing down of the Atlantic as a winter humidity source for Europe produced conditions of a cold desert there, with tundra vegetation, stagnating or retreating polythermal to cold glaciers and deep penetration of mainly periglacial permafrost, especially also due to the absence of thick insulating winter snow.

The polar climate is one endmember representing the coldest climate. From Fig. 2-1 and Fig. 2-3 it is obvious that between the retreat of a glacier (or glaciers) and the advance of new glaciers lies a variable, but substantial, amount of time during which the climate must have been warmer than during the polar climate. While temperatures gradually (often quickly) increased, climate transitioned from a polar climate to a continental and temperate climate. Continental and temperate climates are defined based on temperature, precipitation and seasonality. A continental climate has at least one month with average temperatures below 0 °C and at least one month with average temperatures above 10 °C (a modern-day analogue is e.g. climate in parts of northern and central Europe), whereas a temperate climate has at least one month above 18 °C (a modern-day analogue is e.g. climate in parts of central and southern Europe).

Within the region, several particularly well studied sites have revealed climate information in the past. However, most of them have revealed information on the climate for specific time slices only, and these mostly cover times between the LGM and today (see e.g. Davis et al., 2022 for a summary of proxy data for the LGM and references below). The reason for this is that terrestrial archives of climatic and environmental changes are mainly based on recorders such as lake sediments, stalagmites, tree rings and pollen. However, the advance and retreat of (large) glaciers readily destroys climate records by erosive processes (see Chapter 3 or Nagra 2024a) and/or climate-sensitive recorders are generally rare (e.g. caves). In the vicinity of the region of interest,

some outstanding sites that have preserved detailed climate information older than the LGM are Bergsee (Germany), Wehntal (Switzerland), la Grande Pile (France) and Fürmoos (Germany).

Becker et al. (2006), Duprat-Oualid et al. (2017) and Kämpf et al. (2022) used pollen and sediments to determine the prevailing vegetation for the last ~ 45 ka in the Bergsee area (Fig. 2-5). The pollen and sediments imply a frequent advance and retreat of glaciers during the Birrfeld Glaciation (locally termed Würm, MIS 2 and MIS 3; Fig. 2-3) with only moderate warming during the interstadials. Pollen representative of a quick succession of (boreal) forest and steppe development during MIS 3 were found. At the onset of MIS 2, vegetation quickly changes to dry grassland, a sign of increasingly colder climates. Younger (at the end of MIS 2) flora and fauna show a gradual warming towards current climates. Close to the Bergsee, Anselmetti et al. (2010) and Dehnert et al. (2012) studied drill cores from a glacially overdeepened valley, the Wehntal. The drill cores comprise a 180-kyr sedimentary succession and, hence, cover the time from MIS 6 (Beringen glaciation) until today.

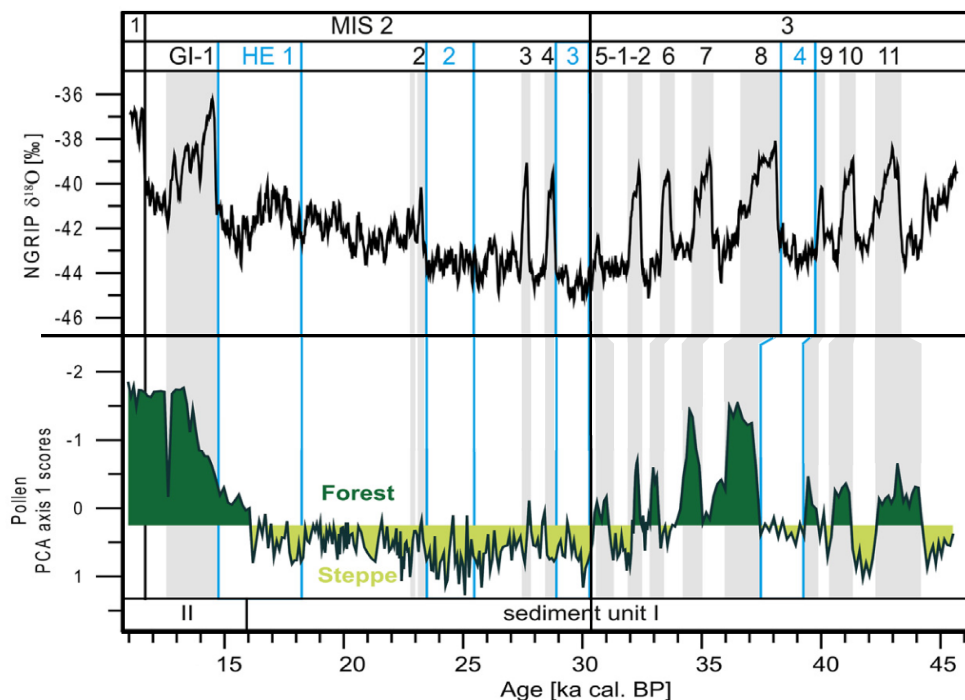


Fig. 2-5: Vegetation in southern Germany (lower panel) and $\delta^{18}\text{O}$ from Greenland (upper panel) (after Kämpf et al. 2022)

Bergsee proxy record from 46 – 11 ka cal. BP (bottom) and North Greenland Ice Core Project (NGRIP) event stratigraphy (GI 1 to 11, 100-yr mean; Rasmussen et al. 2014) and marine isotope stratigraphy (MIS 1 – 3; Lisiecki & Raymo 2005) (top). Grey bars indicate correlations of interstadial conditions in the Bergsee pollen record and Greenland and blue framed bars pronounced stadials related to the formation of Heinrich events (HE) in the North Atlantic (Hemming 2004).

The warmest climates between the LGM and the MPT are most likely indicated by peat deposits during MIS 5 (Anselmetti et al. 2010). Today, peat deposits are mainly found in polar (tundra), continental and colder temperate climates (Xu et al. 2017). A similar result was found in the pollen record from La Grande Pile (France). Here, three distinct warm temperate intervals were identified by Woillard (1979) and more recently by Rousseau et al. (2007). Rousseau et al. (2007) estimated mean annual temperatures of up to 15 °C (between 125 and 120 ka), current mean

annual temperatures at the location of the borehole in La Grande Pile are around 10 °C (Rousseau et al. 2007), in lowland Switzerland 8 – 12 °C (Meteoswiss), so several degrees lower than what was estimated for MIS 5. Bolland et al. (2022) inferred maximum summer temperatures at the location of Fürmoos (Germany) from a drill core covering at least the time from MIS 6 until today. Again, the highest temperatures were found between 125 and 120 ka (MIS 5) reaching up to 16 °C. Similar results were found in speleothems from caves in Switzerland (Skiba et al. 2023, Wilcox et al. 2020).

The examples mentioned above show that between the frequent glacier advances, climate changed from polar to temperate climates with mean temperatures even exceeding today's mean temperatures by several degrees for a considerable amount of time. Glaciations lasted tens of thousands of years although this does not mean that all of Switzerland was covered by ice the whole time.

2.3 Global Quaternary climate simulations

The general cyclicity of the climate introduced in Section 2.1 is also the basis for the climate simulations performed as part of this report. It is highly likely that orbital forcing will continue to have an influence on climate in the future, although the fact that CO₂ concentrations will be, at least for some time, outside the range of those experienced during the Quaternary (Lord et al. 2016) means that the nature of the climatic response to that orbital forcing is highly uncertain. Climate is a spatially heterogeneous, highly non-linear phenomenon with multiple feedback loops and tipping points. Simulated future climate scenarios are only significant if the results of past climate simulations are comparable to the known observed past climate, and only if the future driving conditions, such as CO₂ concentrations, are accurate or, at the very least, if the uncertainties in those driving forces are well accounted for.

Simulations of climate (and respective changes) must take into account all relevant processes on a global scale. Currently, it is not possible to achieve this, in part because the different processes are not sufficiently well understood and in part because they are too complex to be included in detail on such a scale (i.e. we do not have the computational resource to model them). Hence, long-term climate simulations have to be performed with a reduced level of complexity, which may vary depending on the necessary scale in space and time. In long-term global climate simulations, climate processes are simplified primarily because currently computers are not fast enough for highly detailed simulations. Existing state-of-the-art Earth System Models (ESM) are computationally too expensive to routinely perform simulations beyond a few hundred years. Most future climate simulations used in the IPCC are of the order one hundred years (e.g. Randall et al. 2007). On longer timescales, only simple models or models of intermediate complexity (EMICs) can be used on a global scale (e.g. Claussen et al. 2002). One approach to overcome the problem of increased computational costs for modelling increasingly long timespans is to combine a hierarchy of models so that the trade-off between precision of results (compared to known empirical data) and computational time is within an acceptable range. These global long-term climate simulations can then be downscaled at particular time slices, providing boundary conditions for much more complex ESMs in a nested approach.

For long-term climate simulations, a key boundary condition for the simulations is the insolation. In this case, insolation is often defined as the amount of energy received by the Earth at a particular latitude range and during a particular season, and can be calculated by the well-known, and predictable, astronomical parameters of obliquity, eccentricity and precession (e.g. Berger 1978, Berger & Loutre 1991; see also Fig. 2-1). For the prediction of the long-term climate evolution, the emissions and resulting concentrations of (anthropogenic) greenhouse gases also play a major role. According to the IPCC (Pörtner et al. 2023) and many others, the global atmospheric concentrations of greenhouse gases have increased due to human activities. Forster et al. (2021) and Lord et al. (2016) summarise the lifespan of greenhouse gases and their contribution. According to them, the residence time of CO₂, the most significant man-made green-

house gas, in the atmosphere is difficult to determine because several processes actively remove CO₂. An estimated 60 to 80% of the released CO₂ dissolves into the ocean over a period of 20 – 200 years. The remainder is removed by slower processes such as chemical weathering, which may take up to 100 kyr or more. This means that, once released, CO₂ impacts the climate for very long timespans. N₂O, another common anthropogenic greenhouse gas, is destroyed in the stratosphere and removed from the atmosphere in around 100 yr. Methane, CH₄, is mostly removed from the atmosphere by chemical reactions which roughly take only a decade or so. However, natural emissions of both of these greenhouse gases increase with temperature, and so will remain elevated as long as CO₂ (and therefore temperature) is also elevated. Other greenhouse gases such as compounds containing chlorine and/or fluorine exist in a large number of different chemical species, each of which can last in the atmosphere for a specific length of time – from less than a year to many thousands of years. The IPCC has published a comprehensive list of the atmospheric lifetime of the various chlorofluorocarbons (CFC) and other greenhouse gases (Forster et al. 2021).

Numerous previous simulations have been performed to investigate future climate changes. The majority of these studies have been performed for the immediate future using ESMs, that is simulations run until the year 2100 (e.g. Masson-Delmotte et al. 2021) using very complex models. On slightly longer time scales, large numbers of studies are also available, often starting with the LGM (e.g. Ganopolski & Brovkin 2017, Kageyama et al. 2021) but mostly with an already reduced complexity. On even longer timescales, the number of available studies quickly decreases. Simulations are often run using ESMs or conceptual models and often focus on the onset of the next glaciation taking only a limited number of drivers into account, mainly CO₂ content and/or insolation (e.g. Berger et al. 2003, Tzedakis et al. 2012, Ganopolski et al. 2016). On timescales up to 1 Mio yr or more, only a few simulations exist (e.g. BIOCLIM 2001, Berger & Loutre 2002, Archer & Ganopolski 2005, Talento & Ganopolski 2021, see also Lunt et al. 2017 and references therein).

Simulations of future climate evolution generally agree that temperature is very likely to continue to increase in the 21st century and beyond, depending on the CO₂ emission scenario (e.g. Friedlingstein et al. 2006, Collins et al. 2013). The next glacial inception is generally projected to occur approximately at 50 kyr under pre-industrial conditions (e.g. Berger & Loutre 2002, Texier et al. 2003, Ganopolski et al. 2016). However, given the amount of current greenhouse gas concentrations in the atmosphere and the projected emission scenarios in the future, it appears very likely that mean global temperatures remain elevated above natural values (e.g. Archer 2005, Colbourn et al. 2015, Lord et al. 2016). Previous studies also agree that this increased temperature will postpone the onset of the next glaciation for a long time, depending on the emission scenario (e.g. Berger & Loutre 2002, Archer & Ganopolski 2005, Ganopolski et al. 2016).

The global long-term climate simulations discussed in this report focus on the two main drivers, insolation and carbon emission (Archer & Ganopolski 2005, Ganopolski et al. 2016, Talento & Ganopolski 2021). The model consists of a system of three coupled non-linear differential equations representing physical mechanisms relevant for the evolution of the climate – ice sheet – carbon cycle system on timescales longer than thousands of years (Talento & Ganopolski 2021). Many simulations with slightly varying parametrisations and assumption were performed (Talento & Ganopolski 2021). In this report, only the resulting best estimate is further discussed because it shows the best fit to palaeo-reconstructions of climate based on measured data (e.g. δ¹⁸O from sediments and ice cores) and to other complex models for the last 800 kyr (Fig. 2-6). The last 800 kyr were selected because currently this is the only period of time for which reconstructions of the atmospheric CO₂ concentration based on proxy data are available, and it was dominated by long glacial cycles (100-kyr cyclicality) which are expected to continue in the future (depending on anthropogenic greenhouse-gas emissions, see Chapter 4).

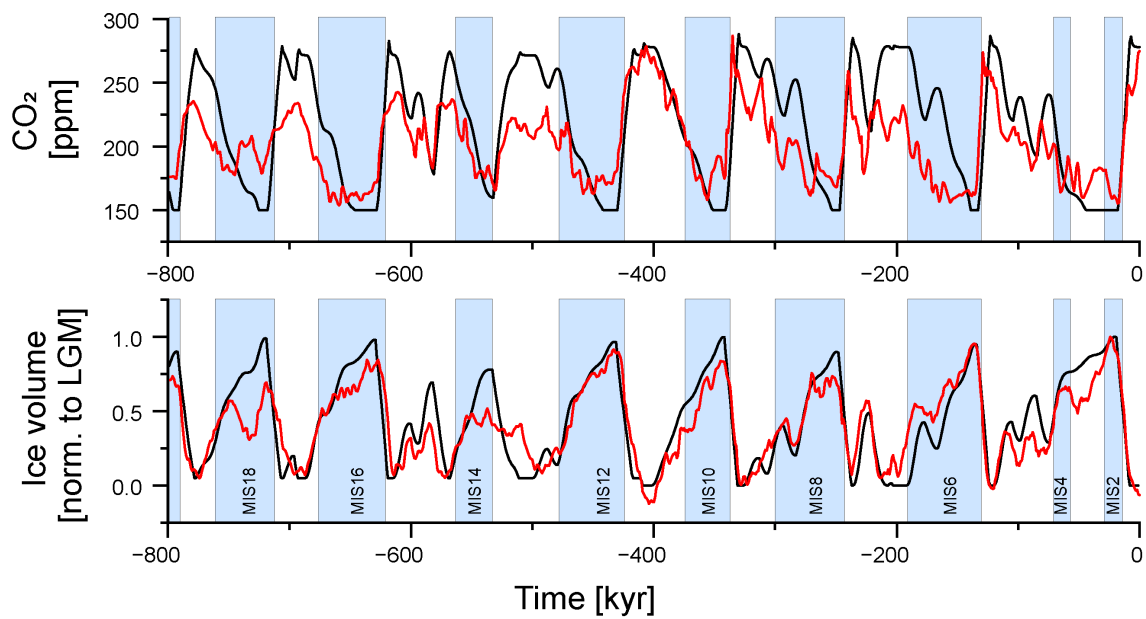


Fig. 2-6: Comparison of climate model results and palaeoclimatic records

Top: Atmospheric CO₂ concentration (ppm) from simulation (black; Talento & Ganopolski 2021) and from palaeoclimatic record (red; Lüthi et al. 2008).

Bottom: Global landmass ice volume (normalised to LGM) from simulations (black; Talento & Ganopolski 2021) and from reconstruction (red; Spratt & Lisiecki 2016).

Fig. 2-6 shows that the performed simulation of past Quaternary climate correlates very well with known proxy data. Both CO₂ as well as global ice volume after ~ 500 kyr can be simulated with low uncertainties. Before ~ 500 kyr, the model overestimates the CO₂ concentration and ice heights (see Talento & Ganopolski 2021 for more details). However, the model succeeds in capturing the timing of the glacial/interglacial cycles and hence can be used for downscaling to more local climate data (see next section and Chapter 3) as well as for future climate projections (see Chapter 4).

2.4 Local Quaternary climate simulations

Global scale climate simulation results performed for this study can be downscaled to a much higher resolution using a nested approach with more complex climate models. One challenge is to deduce high-resolution information from low-resolution variables using a dynamical downscaling approach. A nested model approach was applied where the global-scale simulations were used as input for a (much) more complex Earth System Model (ESM) which in turn is used as input for a Regional Climate Model (RCM), resulting in a resolution of ~ 4 km² for pre-selected time slices in the past. The Community Earth System Model (CESM) (Hurrell et al. 2013) is used to perform simulations for key periods (referred to as "states") during the late Quaternary (Holocene; LGM, MIS 4, MIS 6, MIS 8 and MIS 16) and the Pre-Industrial (PI). The climate forcing of the states consists of changes in orbital parameters, concentration of greenhouse gases, and prescribed reconstructed Northern Hemisphere Ice Sheet (NHIS) configurations.

The ESM simulations are the basis for the next step in the model chain, the dynamical downscaling using the Weather Research and Forecasting (WRF) model (Skamarock et al. 2008, Powers et al. 2017). For the different glacial periods, the RCM requires information on the top surface elevation of the Alpine Ice Field (AIF) at the time of the considered state, which is not available a priori. Therefore, the LGM ice-surface topography derived from an earlier modelling work (Seguinot et al. 2018) is used as ice-surface topography input for the model over the Alpine region for all states except for PI and MIS 16 for which the present-day surface topography and a more extensive AIF are taken, respectively. The climate model chain provides physically consistent high-resolution values of temperature and precipitation fields at daily timescales over 10 years for each targeted climate state. The selected time slices were based on a combination of available empirical data (e.g. LGM) and on the severity of the glaciation resulting in high-resolution climate data for the periods of the PI, LGM, MIS 4, MIS 6, MIS 8 and MIS 16. PI and MIS 4 were mainly selected to be able to verify results with known measured data (PI) and recent existing simulations (MIS 4), thus benchmarking the performed downscaling against independent data. MIS 6 and MIS 8 represent different states of orbital forcing presumably leading to different severities of glaciation in the area of interest (Swiss Alps and the Alpine foreland). MIS 16 is a glacial period of particular interest. It belongs to the two most severe dips in the oxygen isotopic ratio of benthic foraminifera (Fig. 2-1) during the past 800 kyr and globally contained one of the largest glacial volumes of the Quaternary (Barbante et al. 2010, see also Fig. 2-2). Boundary conditions for the simulation of MIS 16 are different from the other simulations, mainly because no proxy-based ice-sheet reconstructions are available for this period. Hence, the Northern Hemisphere Ice-Sheet distribution was taken from simulations using an EMIC (CLIMBER X; Willeit et al. 2022). As a benchmark, simulations of the LGM from CLIMBER X and the RCM were performed. The main difference between the two simulations is largely in the distribution (and thickness) of ice sheets resulting in temperature and precipitation differences where the ice sheets are different. Hence, results for MIS 16 presented below are not directly comparable with results from the other presented simulations but can still be used on a qualitative basis.

The regional-scale glacial time-slice simulations for LGM, MIS 4, MIS 6, MIS 8 and MIS 16 show a rather wide range of different climate states which can be met during glacial times. The general characteristics of a rather cold and dry climate on a global scale is found for all the assessed states. However, the strong response of the atmospheric circulation to the ice sheets in the northern hemisphere and orbital forcing induces specific regional patterns so that the general cold and dry climate characteristic is not found in all regions and seasons.

To illustrate the more general statements above, a comparison of mean values (averages over 10 years) of temperature and precipitation over the area of Switzerland can be used (Fig. 2-7, Fig. 2-8 and Tab. 2-1).

Under pre-industrial conditions, a mean temperature of 8.3 °C was simulated. The difference in temperature between the coldest and the warmest month is 17.5 °C. The yearly mean precipitation is 4.39 mm/day. The difference between the driest and the wettest month, a measure of the variability within a year, is 2.74 mm/day.

The simulated LGM is, on average, much colder than the PI leading to a mean temperature of only -1 °C with a reduced difference of 15.7 °C between the coldest and warmest month. It is also dryer than the PI showing a mean precipitation of 3.35 mm/day and a slightly reduced difference in precipitation of 2.65 mm/day over northeastern Switzerland.

The MIS 4 simulation shows a mean temperature of -1.5 °C and a temperature difference of 20.3 °C, thus a colder and stronger difference than the LGM, with warmer summers and slightly colder winters. The mean precipitation is 3.2 mm/day and the difference between the driest and wettest month is 2.53 mm/day, only slightly reduced with respect to the LGM.

Tab. 2-1: Mean modelled temperature and precipitation for selected time slices of the local scale WRF simulations

Mean temperature and mean precipitation represent a 10-yr mean over Switzerland, difference indicates the maximum difference between warmest and coldest or wettest and driest month, both indicating seasonality.

	Mean T	Difference T	Mean precipitation	Difference precipitation
	[°C]	[°C]	[mm/d]	[mm/d]
PI	8.3	17.5	4.39	2.74
LGM (MIS 2)	-1	15.7	3.35	2.65
MIS 4	-1.5	20.3	3.2	2.53
MIS 6	-1.3	18.4	3.13	2.97
MIS 8	-0.3	22	3.44	3.5
MIS 16	-4.5	16.1	5.34	5.29

The simulated MIS 6 behaves similarly to the LGM (slightly colder (-1.3 °C) and dryer (3.13 mm/day) but shows a stronger temperature and precipitation difference (18.4 °C and 2.97 mm/day).

The MIS 8 simulation is warmer (mean temperature of -0.3 °C) with a larger difference between the warmest and coldest months (22 °C) compared to results from the LGM, MIS 4 and MIS 6. MIS 8 is also wetter (mean precipitation of 3.44 mm/day) compared to LGM, MIS 4 and MIS 6 and shows the largest difference of 3.5 mm/day between the wettest and driest months.

Although the simulations of MIS 16 are based on different boundary conditions, the qualitative differences of MIS 16 to the other simulation results are pronounced. In general, it is much colder (-4.5 °C) with a temperature difference of 16.1 °C and high precipitation (5.34 mm/day) throughout the year. It also has a large difference in precipitation (5.29 mm/d) with much dryer winters than summers. The generally low temperatures and high precipitation during MIS 16 might be expected to provide better conditions for ice sheet and glacier growth than the other glacials.

Overall, the different simulations suggest that the interplay between mean temperature changes and precipitation changes as well as their seasonalities and variabilities are important to provide conditions for a glacier advance onto the Alpine foreland in the region of northeastern Switzerland. The glacier advance also has the effect of increasing surface elevation at the locations where e.g. glaciers fill Alpine valleys or (larger) lobes advance into the foreland. This is visible in Fig. 2-7 where the major differences compared to the PI are mainly at locations of large elevation changes due to ice (known as environmental lapse rate effect).

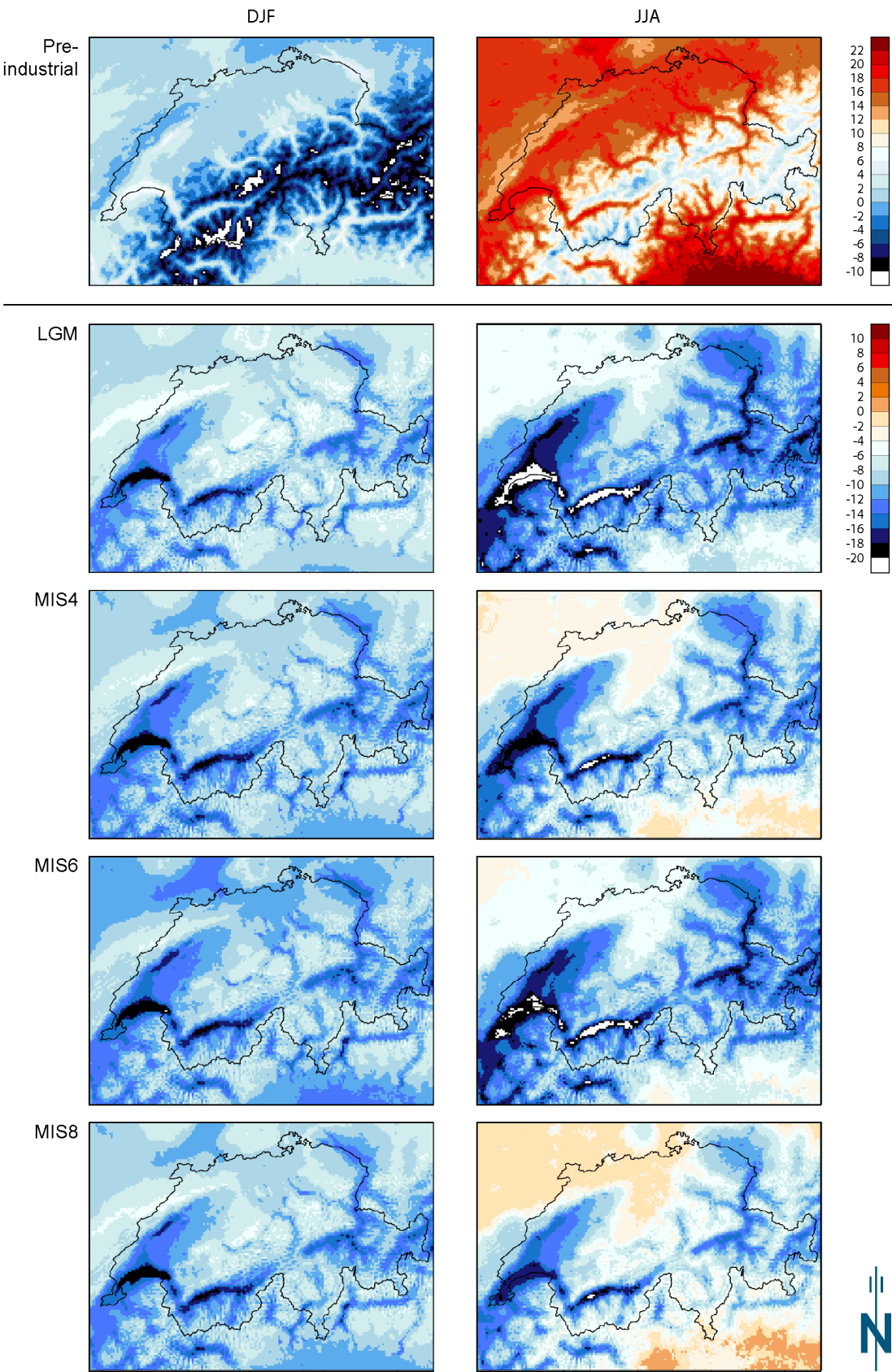
For precipitation (Fig. 2-8), the situation is more heterogeneous, with both seasons characterised by differences not of the same sign and in different locations. MIS 8 for example shows a large increase of precipitation during winter in the south while it shows a comparably large decrease of precipitation during summer compared to PI conditions. A comparable effect can be seen during the LGM and MIS 4 although here the summer months are much wetter over the Southern Alps than during MIS 8. For MIS 6, precipitation during the summer is comparable to the LGM while the winter months are exceptionally dry. MIS 16 shows a very different behaviour where the summer mostly shows a much higher precipitation than the PI while the winter months are either

drier or wetter, depending on location. These differing, somewhat irregular simulation results cannot be explained with thermodynamics alone (as cooling would generally suggest a reduction of precipitation) but must be attributed to dynamical effects (e.g. changes in the atmospheric circulation).

In addition to the inherent uncertainties in proxy data, the constantly evolving landscape destroys many of the proxies over time, or makes a precise analysis difficult, especially with respect to timing; proxy information for periods before the LGM is therefore often sparse and ambiguous.

Russo et al. (2024) used data compiled from pollen by Davis et al. (2022) to compare the climate simulations of the LGM shown in Fig. 2-7, Fig. 2-8 and Fig. 2-9. They conclude that model results are in good agreement with evidence from the proxies, in particular for temperatures. For precipitation, they found that agreement between the simulations and the proxy data is less compared to temperature. In particular, the simulations seem to overestimate precipitation during winter over Western Europe (present-day France and the Alps) and underestimate precipitation during summer over Eastern Europe (particularly the Balkan, outside of the modelling area shown in Fig. 2-7 to Fig. 2-9), compared to the pollen reconstructions from Davis et al. (2021).

Fig. 2-9 shows the classified climate states associated with the simulations of the pre-industrial period and MIS 2 according to the Köppen classification scheme (Köppen 1918). While during the pre-industrial period most of Switzerland experienced a temperate climate (except for the Alps), during MIS 2 most of Switzerland saw a polar climate and never exceeded cold, continental climatic conditions.



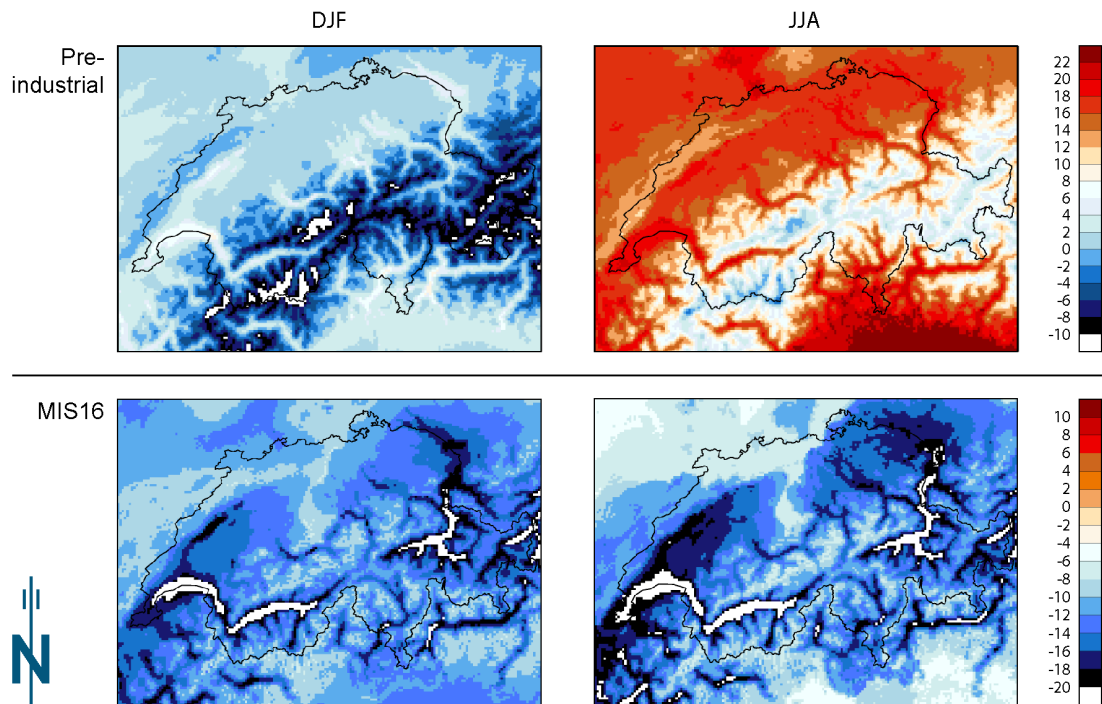
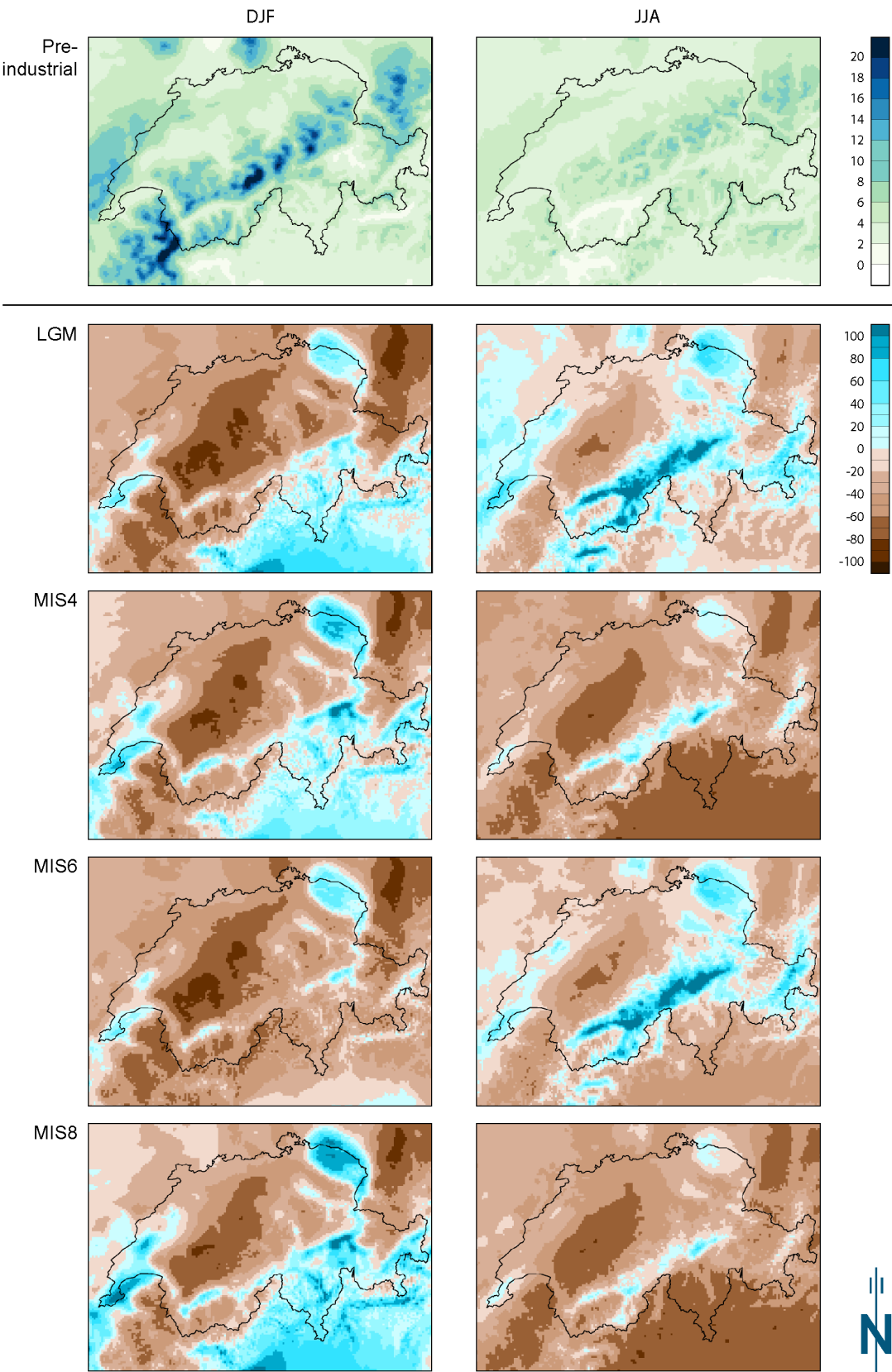


Fig. 2-7: Mean modelled temperatures during the selected periods in winter (left; December, January, February) and summer (right; June, July, August)

Concerning the pre-industrial period, temperatures are absolute (°C), for the LGM and MIS stages, temperatures are given as differences to pre-industrial. The black line outlines Switzerland. DJF: December, January, February. JJA: June, July, August.



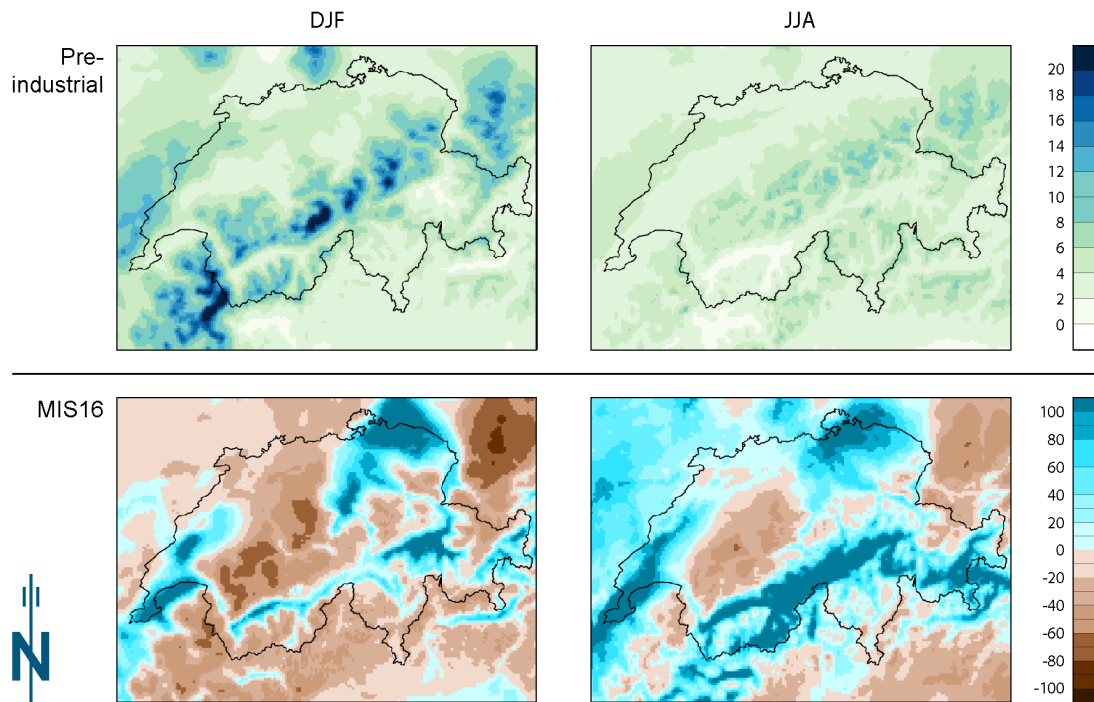


Fig. 2-8: Mean modelled precipitation during the selected glacial in winter (left; December, January, February) and summer (right; June, July, August)

Concerning the pre-industrial period, precipitation is given as absolute values of mm/d. For the LGM and MIS stages, precipitation is given as differences in percent to the pre-industrial. The black line outlines Switzerland. DJF: December, January, February. JJA: June, July, August.

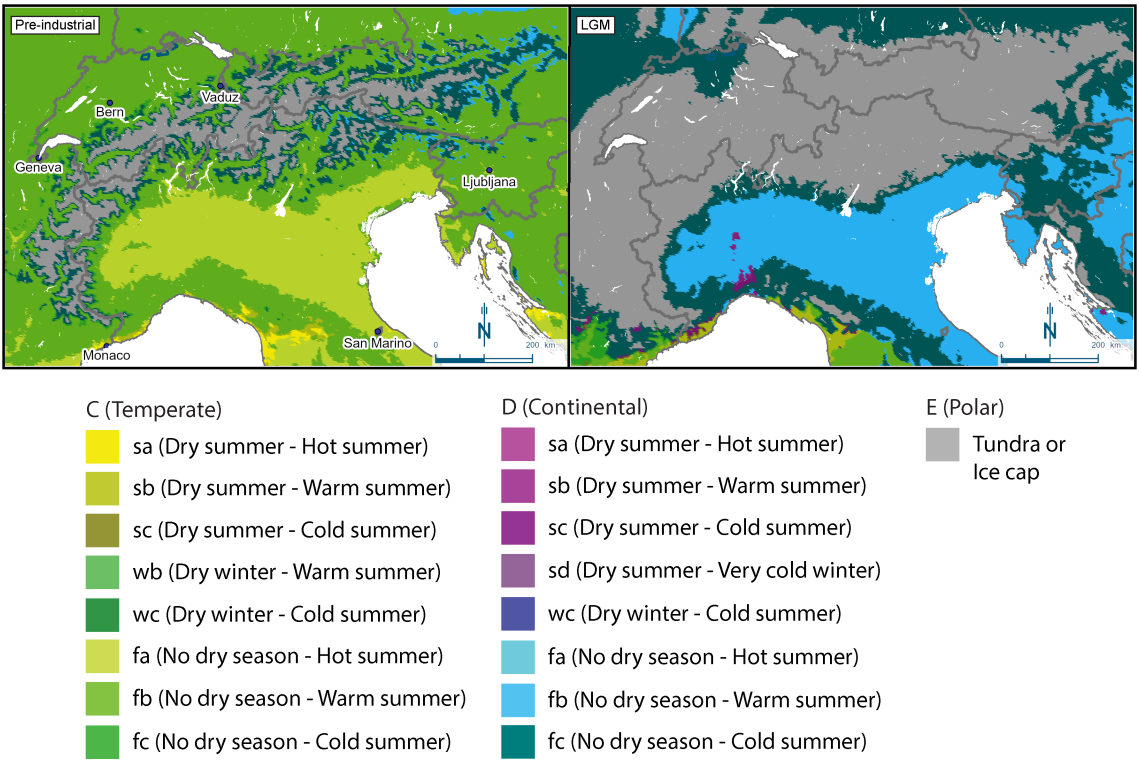


Fig. 2-9: Simulated pre-industrial and MIS climates classified according to the Köppen classification

The climate classifications follow the Köppen classification scheme (Köppen 1918).

For the interstadials, available proxy data for Northern Switzerland shows mainly polar (tundra) and continental climates with short periods within a temperate climate (MIS 5; Fig. 2-9 and Section 2.2). During the glacials, polar climates (ice cover or tundra) were likely to have dominated in Switzerland and neighbouring regions.

3 From climate to ice flow and erosion

Glaciers are dynamic stores of solid water that constantly exchange mass and energy with the atmosphere, the hydrosphere and the lithosphere. The growth of a glacier is dictated by its climate-driven mass balance, the net accumulation of snow minus the mass loss by melting (plus in places calving). At the most basic level, glaciers thicken and advance when the overall mass balance (accumulation minus mass loss) is positive, and thin and retreat when the mass balance is negative. Similarly, glacier temperature is a function of the inputs and outputs of energy at the surface and the bed and controls the rate of ice motion through deformation and sliding. Fluctuations in glacier mass balance and temperature are therefore strongly coupled responses to changes in climate. In addition to ice extent (growth and decay), the mass balance also dictates the flux of ice and meltwater through the glacial system (i.e. the transfer of mass from accumulation to ablation zones) and the average ice thickness (Benn & Evans 2010). The ice flux and ice thickness in turn control the conditions at the glacier bed, primarily basal temperature and water pressure which determine the rate of basal ice motion, dictating the capacity of ice to do geomorphic work on the bed by glacial erosion.

This chapter presents the reconstruction of the evolution of the ice coverage over the Alps during former glaciations by multi-physical coupled climate and glacier transient modelling. The analysis follows the work of Cohen et al. (2018), Seguinot et al. (2018) and Imhof (2021) but uses the modelled local palaeoclimate of Chapter 2 instead of a distortion of present-day climate to mimic palaeo conditions for forcing the ice-flow simulations. This approach yields an improved match between LGM-related features such as maximum ice extent, ice thickness and ice flow with field observations (Jouvet et al. 2023). The LGM ice-flow simulations (Section 3.2) are identical to those described in Jouvet et al. (2023).

After a description of the climate and glacier models in Section 3.1, Section 3.2 first presents the simulated Alpine Ice Field (AIF) for the past glacial cycle and then compares it to those for previous glaciations. This is followed by a characterisation of glacier conditions in view of its implications for glacial erosion. Based on this characterisation, the potential for erosion either directly by the sliding glacier or by subglacially flowing water is estimated in Section 3.3. Finally, Section 3.4 reports sensitivity analyses to assess the influence of key climatic and non-climatic model parameters on the simulation results.

3.1 Description of model chain

To simulate the AIF and the Rhine glacier system over palaeo timescales with a spatial resolution sufficiently high to capture the complex topography of the Alpine mountain range, a series of climate states is first targeted by combining global Earth System and Regional Climate Models (see Chapter 2). Then, a transient climate is generated using a Glacial Index (GI) approach (Section 3.1.1). Because the high temporal resolution of the RCM results is not needed for glacier modelling, decadal monthly averages of daily values of temperature and precipitation are calculated for each period. Additionally, the monthly standard deviation of the mean daily temperatures is estimated to keep annual and daily variability of the original temperature data. Finally, the transient climate is used to force two independent ice-flow models, the Parallel Ice Sheet Model (PISM; Section 3.1.2) and the "full" non-linear Stokes Ice Flow Model Elmer/Ice (Section 3.1.3, Fig. 3-1). These two models do not solve directly for glacial erosion. Rather, a potential for erosion is estimated a posteriori based on the transient ice flow dynamics (sliding speed, subglacial water flow) (see Sections 3.2 and 3.3).

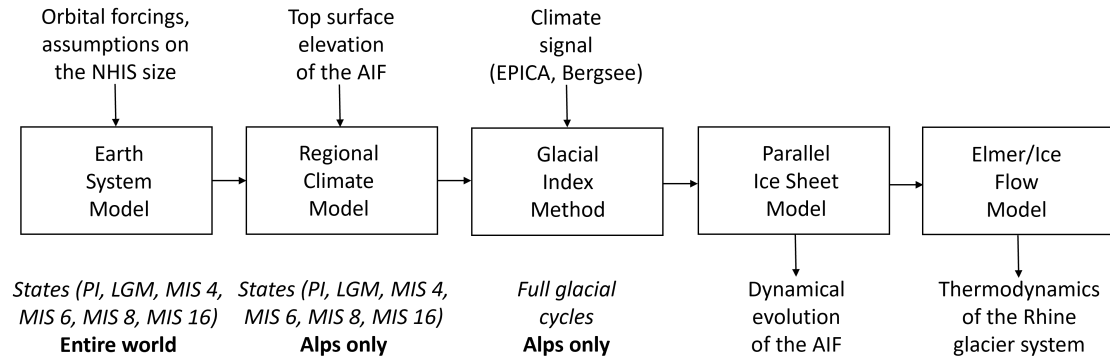


Fig. 3-1: Coupled climate – glacier model chain

The climate modelling using an Earth system model and a regional climate model, and their results, are detailed in Chapter 2.

3.1.1 Glacial Index (GI) approach

Transient ice-flow modelling requires a continuous climate over the time that spans the glacial states (LGM, MIS 4, MIS 6, MIS 8 and MIS 16) and the Pre-Industrial (PI) period. To interpolate the climate data between the states in a physically realistic way, a Glacial Index (GI) approach (Sutter et al. 2019) is adopted. A GI function is defined that maps time to a scalar between 0 and 1 with two extreme states: one state with nearly no ice over the Alps corresponding to $GI = 0$ and one maximum state in terms of glacier extent at the targeted glacial climate state (LGM, MIS 4, MIS 6, MIS 8, MIS 16) corresponding to $GI = 1$. The climate consists of a set of variables: mean temperature, temperature variability (defined by the standard deviation), mean precipitation, and lapse rate (rate at which air temperature falls with increasing altitude). The GI function is built by linearly rescaling a climate proxy signal between 0 and 1 in such way that GI is close to 1 at the glacial state and close to 0 at the PI (Jouvet et al. 2023). Here, the Antarctica EPICA temperature anomaly signal (Jouzel et al. 2007) is used, which is available for the last 800 kyr. The resulting GI function is shown in Fig. 3-2. For sensitivity analysis, a local pollen-based signal from the Bergsee, southern Germany (Duprat-Oualid et al. 2017), which is available from 45 to 15 ka BP, is also used (see Section 2.2). More details on the GI approach can be found in Jouvet et al. (2023).

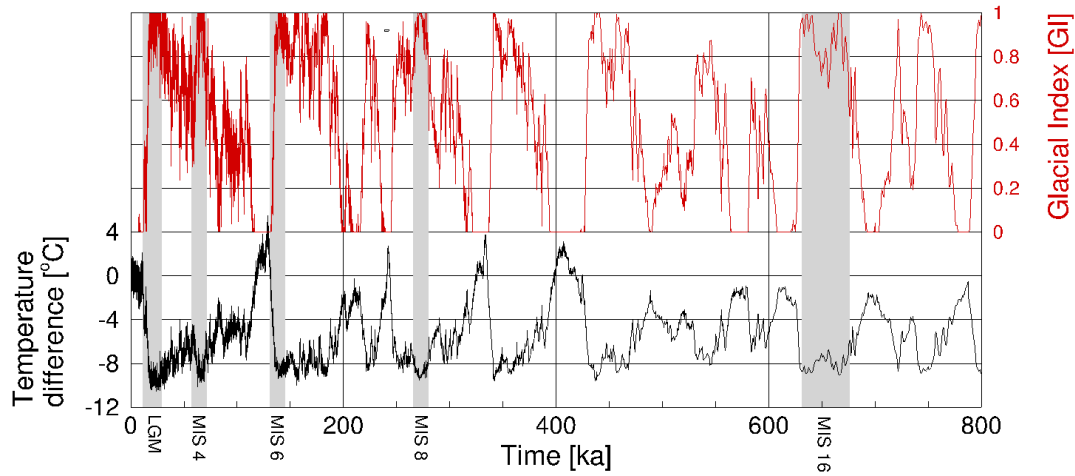


Fig. 3-2: Time evolution of the Glacial Index (red) based on the EPICA signal (Jouzel et al. 2007) for climate forcing (black)

3.1.2 Parallel Ice Sheet Model (PISM)

The Parallel Ice Sheet Model (PISM) (Bueler & Brown 2009, Winkelmann et al. 2011, Jouvet et al. 2023) jointly models the ice-thickness evolution, the ice thermodynamics, the surface mass balance, as well as the deformation of the lithosphere given initial conditions and a driving climate. (i) To model the dynamical motion of ice, PISM uses a linear combination of low order approximations, namely the Shallow Ice Approximation (SIA) for the vertical shear (i.e. motion by ice deformation) and Shallow Shelf Approximation (SSA) for the longitudinal ice extension (i.e. motion by ice sliding). In the SSA, the basal velocity and the basal shear stress are related non-linearly with the Mohr-Coulomb sliding law (Cuffey & Paterson 2010). (ii) PISM is a polythermal model, i.e. it solves jointly the ice dynamics together with the ice-temperature field in three dimensions using an enthalpy formulation (Aschwanden et al. 2012). The ice enthalpy (or the temperature and the water content) impacts both the ice softness, as well as the basal motion. Additionally, the dynamics of ice influences the evolution of the enthalpy field. (iii) PISM uses a combined snow accumulation / Positive Degree-Day (PDD) model (Hock 2003) to compute the surface mass balance from temperature and precipitation fields. On the one hand, surface accumulation is equal to solid precipitation when temperature is below 0 °C. On the other hand, surface ablation is computed proportionally to the number of PDD. (iv) PISM includes an Earth deformation model which combines a layered elastic spherical Earth with a viscous half-space by an elastic plate lithosphere.

3.1.3 Full-Stokes Ice Flow Model (Elmer/Ice)

Elmer/Ice (Gagliardini et al. 2013, Cohen et al. 2018) computes the transient three-dimensional ice velocities, temperatures, and pressures as well as the elevation of the ice surface based on the fully coupled thermo-mechanical equations by solving the non-linear Stokes flow equation together with the heat equation and the mass balance equation at the ice surface. As in PISM, the mass balance is based on the Positive Degree-Day (PDD) model (Hock 2003) that computes ice melt as a function of air temperature. At the ice-bed interface, ice can slide when water is present, which occurs when the temperature at the bed equals the melting temperature (so-called warm-based bed). The sliding speed is linked to the basal shear stress. Stokes flow is more accurate than the Shallow Ice and Shallow Shelf Approximations (SIA and SSA) in zones in which gradients in stresses are significant, like in complex transection glaciers found in Alpine settings, and where sliding is also significant (Ryser et al. 2014, Kirchner et al. 2016). Because of computational cost, however, transient simulations are limited to several thousand years at most and to smaller model domains, i.e. only parts of the AIF such as the Rhine glacier system. This is in contrast to the hybrid approach adopted by PISM (Section 3.1.2) that is a trade-off between mechanical accuracy and computational cost and, thus, permits to simulate long time scales (thousands of years over glacial cycles) for the entire AIF.

3.2 Characterisation of glacier conditions in Northern Switzerland

The objective of the coupled climate and glacier transient modelling is to assess the effects of ice-flow dynamics and climate on the conditions at the Zürich Nordost (ZNO), Nördlich Lägern (NL) and Jura Ost (JO) siting regions during previous glacial cycles and implications for glacial erosion. This assessment is based on quantitative criteria that incorporate the physical processes upon which glacial erosion depends (e.g. ice thickness, timing and duration of ice occupation, basal ice temperature, basal sliding velocity and subglacial water drainage routing). These criteria can be derived from the numerical reconstruction of the ice surface geometry, flow dynamics and temperature field of the AIF (Fig. 3-3, Fig. 3-4) and, in particular, the Rhine glacier system (Fig. 3-8) during past glaciations.

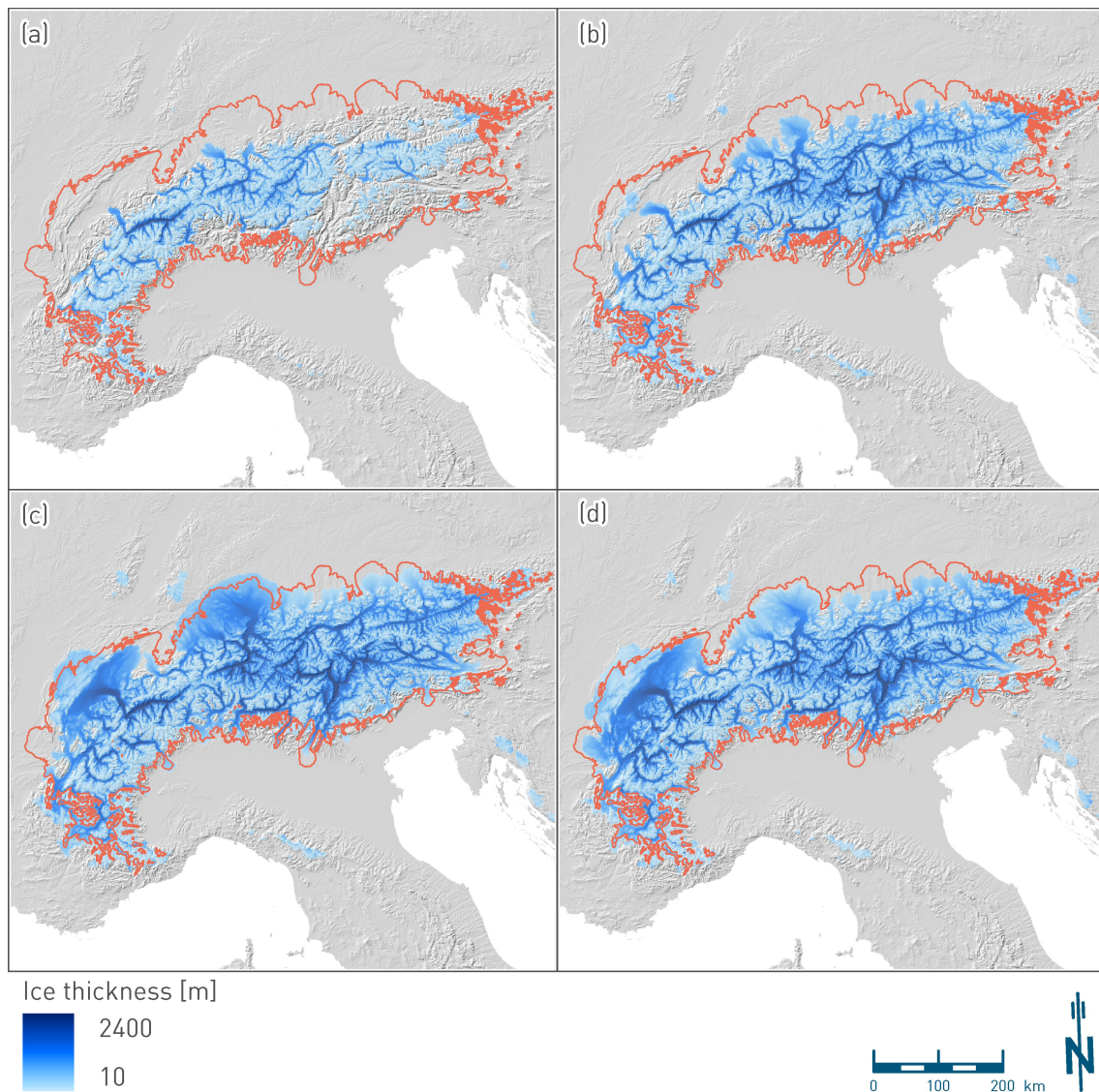


Fig. 3-3: Snapshots of the modelled evolution of ice thickness and extent of the AIF during the last glacial cycle

(a) 37'450 a BP. (b) 28'700 a BP. (c) 24'700 a BP. (d) 23'700 a BP. The red lines denote the LGM outline modified after Ehlers et al. (2011). Note that the Rhone glacier reached its maximum extent ~ 1'000 yr later than the Rhine glacier (see also caption of Fig. 3-4).

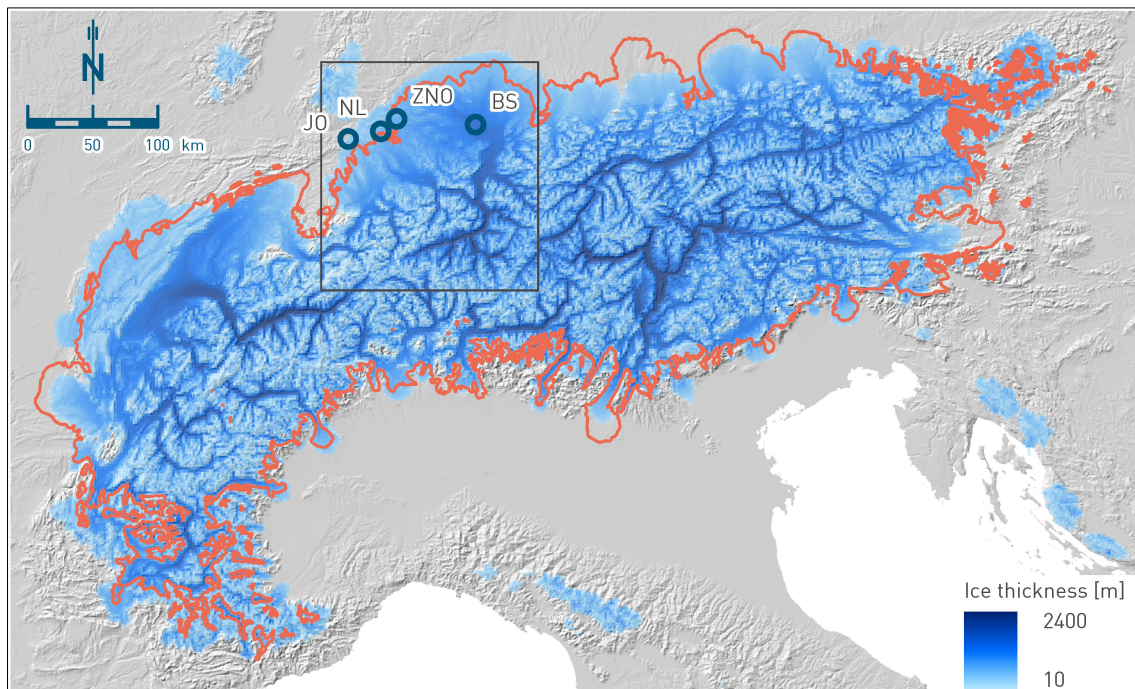


Fig. 3-4: Modelled maximum ice thickness and extent of the AIF during MIS 2 and geomorphologically reconstructed LGM outline modified after Ehlers et al. (2011) (red line)

Note that different glacier lobes reached their maximum ice thicknesses and extents at different times because fluctuations are controlled by the size of catchments and the resulting inertia of each glacier (Fig. 3-3). The blue circles (JO: Jura Ost, NL: Nördlich Lägern, ZNO: Zürich Nordost, BS: Bodensee) indicate locations where ice thickness, duration of ice occupation and sliding distance were evaluated (Fig. 3-6, Fig. 3-7 and Fig. 3-10). The box shows the Rhine glacier system (outline of Fig. 3-8).

As basal topography for the model simulations, the publicly available NASA Shuttle Radar Topographic Mission (SRTM, <http://srtm.csi.cigar.org/>) Digital Elevation Model (DEM), re-sampled at 2 km resolution and with major lakes and present-day glaciers (Farinotti et al. 2019) removed is used.

To reproduce geomorphologically reconstructed glacier outlines (Fig. 3-4), melt parameters in the ice flow model were tuned, and optimal PDD factors are found to be close to standard mean values from the literature (Jouvet et al. 2023). Modelled maximum east-west AIF extent during the LGM matches the geomorphological reconstruction of Ehlers et al. (2011) fairly well (Fig. 3-4). Furthermore, the results show that the flow dynamics of individual modelled glaciers agrees with several independent key geological data, including moraine-based maximum reconstructed glacial extents (see Fig. 9 in Jouvet et al. 2023), known ice transfluences and trajectories of erratic boulders of known origin and deposition (Jouvet et al. 2023).

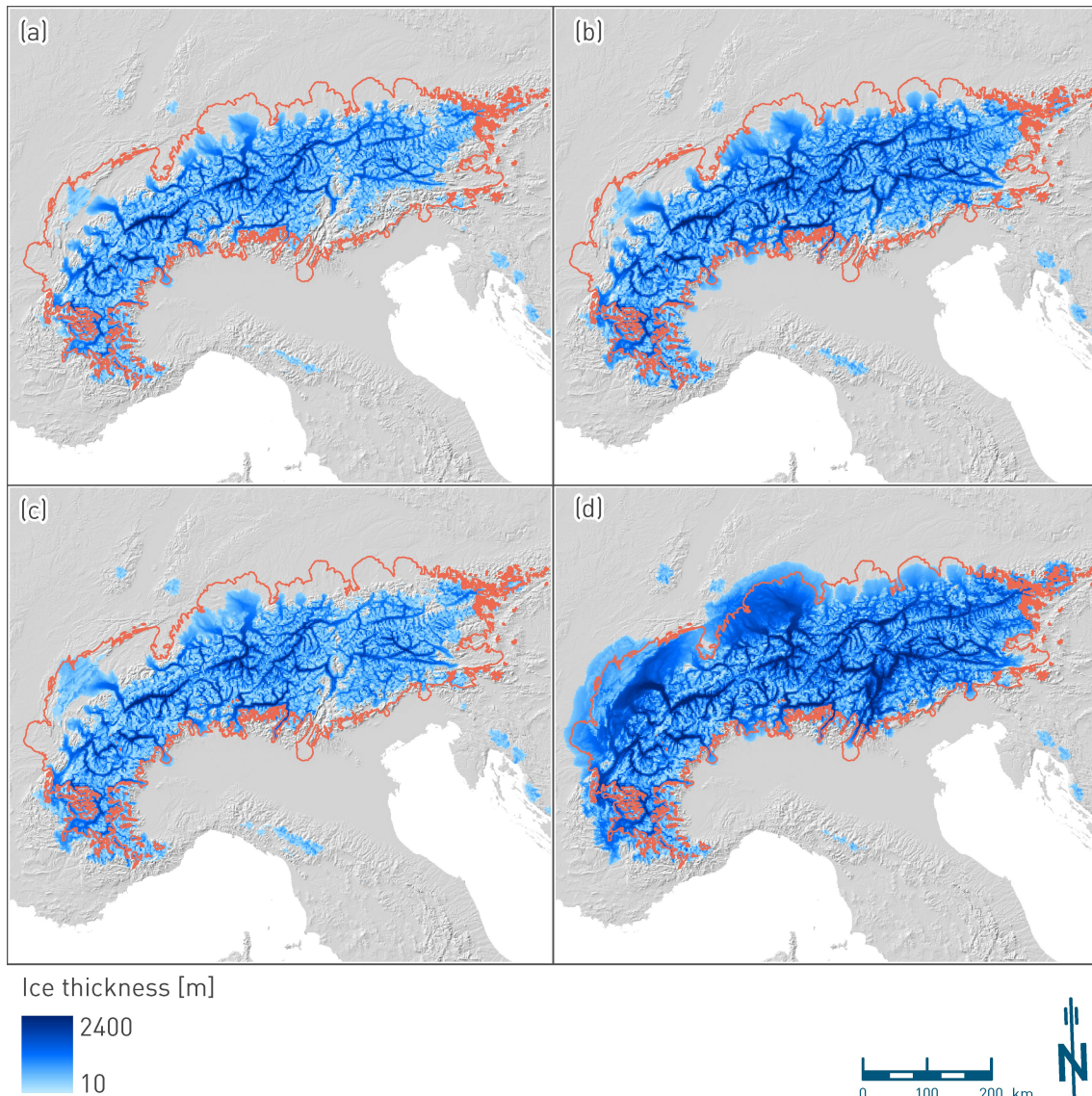


Fig. 3-5: Modelled maximum ice thicknesses and extents during (a) MIS 4, (b) MIS 6, (c) MIS 8 and (d) MIS 16 (compared to those of MIS 2 in Fig. 3-4)

The red lines denote the LGM outline modified after Ehlers et al. (2011).

Similar to the LGM, climate conditions during MIS 4, MIS 6 and MIS 8 were colder and dryer than at the PI, however, with specific regional and seasonal differences (see Section 2.4). Using these climate states, the modelled AIF during MIS 4, MIS 6 and MIS 8 was in general less extensive than during the LGM (Fig. 3-5a – c, Fig. 3-4). In particular, the largest glaciers (e.g. Rhone glacier) were significantly smaller in comparison to the LGM. In contrast, due to colder and wetter climate conditions during MIS 16 than during the LGM the extent of the AIF was larger and was made to match that of the MEG (Schlüchter & Kelly 2000; Fig. 3-5d) using a different PDD model scaling.

The modelled ice extent during MIS 6 (Fig. 3-5b) disagrees with geomorphologically reconstructed ice extents (Preusser et al. 2011) that suggest that, at least in Northern Switzerland, glaciers advanced further into the foreland during MIS 6 than during the LGM (Fig. 2-3). Most

influential for the mass balance and thus growth of a glacier is (i) the winter precipitation that determines the accumulation by snow fall and (ii) the summer temperature that controls the mass loss by ablation. While simulated climate conditions during MIS 4, MIS 6 and MIS 8 were, on average, colder than during the LGM (Tab. 2-1), a detailed analysis of the simulation results reveals that the summers during MIS 4, MIS 6 and MIS 8 were actually warmer than during the LGM (Fig. 2-7). Furthermore, the simulation results also show that during MIS 6 there was less winter precipitation than during the LGM (Fig. 2-8). In combination, the simulated warmer summers and dryer winters during MIS 6 in comparison to the LGM explain, at least in part, why the modelled ice extent during MIS 6 was smaller than that during the LGM (Fig. 3-5b, Fig. 3-4).

Fig. 3-6 shows the transient evolution of the ice thickness over 32 kyr around the LGM, for four different locations in Northern Switzerland (shown as circles in Fig. 3-4), and for different simulations that were carried out as part of a sensitivity study (Section 3.4). The black line (labelled Ref) depicts the simulation shown in Fig. 3-3. The modelled Rhine glacier underwent two advances at ~ 25 kyr and ~ 21 kyr which covered the ZNO and NL siting regions with several hundred metres thick ice while the JO siting region essentially remained ice-free. This is also reflected in the boxplots (Fig. 3-7) from which a modelled maximum ice thickness of $\sim 200 - 400$ m and a duration of ice occupation of $\sim 1'000 - 3'000$ yr in the ZNO and NL siting regions can be deduced.

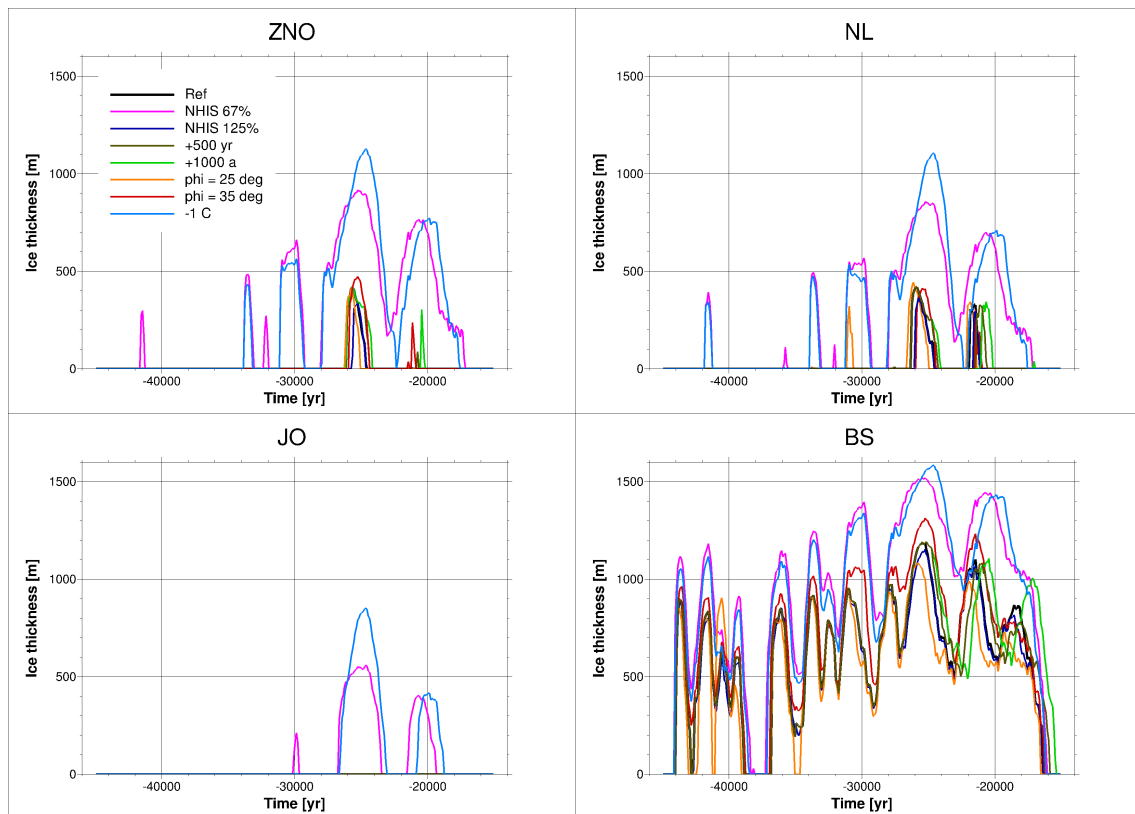


Fig. 3-6: Modelled transient evolution of the ice thickness around the LGM in the (top left) Zürich Nordost (ZNO), (top right) Nördlich Lägern (NL), (bottom left) Jura Ost (JO) siting regions and (bottom right) for the Bodensee (BS) region for different simulations with parameters varied as indicated in the legend

See Fig. 3-4 for the locations where the ice thickness was evaluated and Section 3.4 for details of the sensitivity analysis.

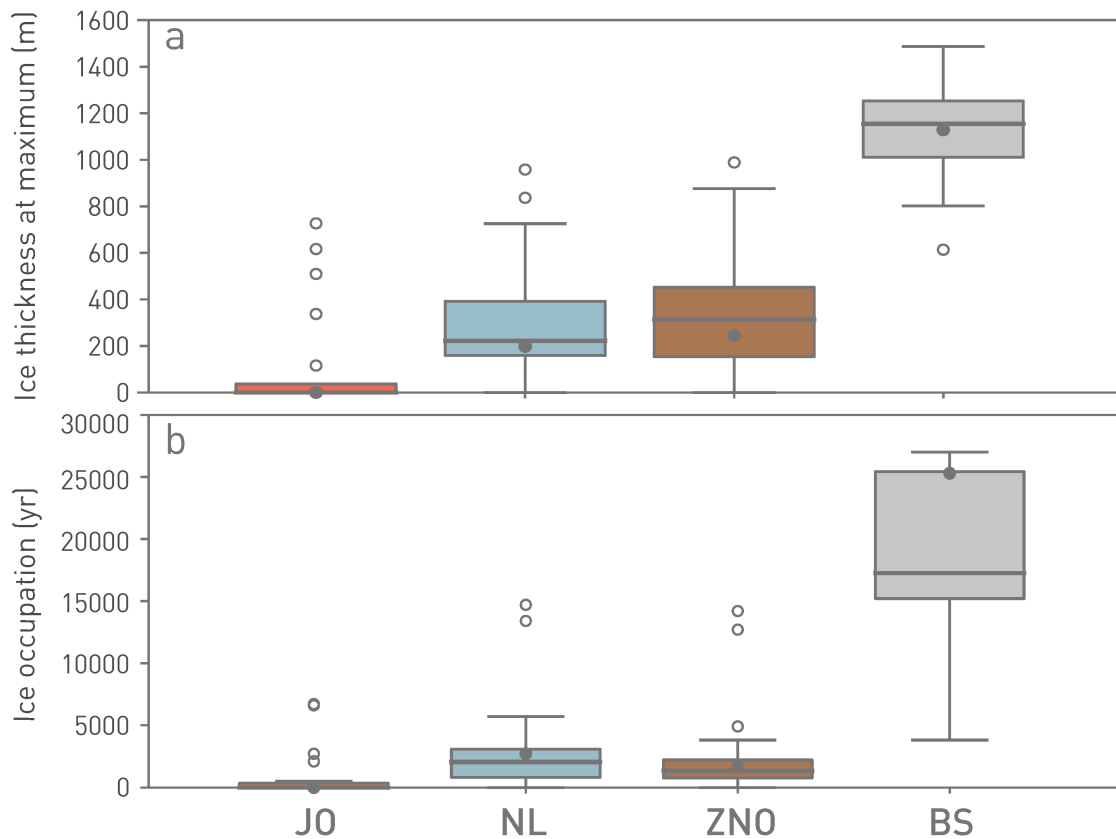


Fig. 3-7: Boxplots of modelled (a) maximum ice thickness and (b) duration of ice occupation during the LGM in the Jura Ost (JO), Nördlich Lägern (NL) and Zürich Nordost (ZNO) siting regions as well as for the Bodensee (BS) region

The boxes depict data between the first (25th percentile) and third quartile (75th percentile) known as the interquartile range (IQR) whereas the orange line in the box shows the median value (50th percentile). Black solid points indicate the result from the reference values (see also Fig. 3-10). From above the upper quartile, a distance of 1.5 times the IQR is measured out and a whisker is drawn up to the largest point from the dataset that falls within this distance. Similarly, a distance of 1.5 times the IQR is measured out below the lower quartile and a whisker is drawn up to the lowest point from the dataset that falls within this distance. All other points are plotted as outliers. Data is based on 19 simulations that were carried out as part of the sensitivity analysis described in Section 3.4. See Fig. 3-4 for the locations where ice thickness and duration of ice occupation were evaluated.

Basal temperature and sliding speed are two key variables that describe basal conditions relevant for glacial erosion (Fig. 3-8). Glacial erosion occurs where ice is warm based (not frozen to the bed), which enables basal sliding and thus abrasion. Furthermore, where meltwater is able to flow along the bed, this promotes rapid sliding and stimulates quarrying (plucking) or may directly lead to subglacial meltwater erosion. Simulations for the LGM indicate that basal ice reached the pressure melting point over much of the piedmont lobes of the Rhine glacier system, and also in the glacial valleys that fed these lobes (Fig. 3-8a). Bedrock highs beneath the lobes and Alpine valleys at high elevations in the accumulation zone remained cold based. The Rhine glacier system was thus polythermal.

In the lobes, despite low surface slopes and low basal shear stresses, sliding dictated the main fluxes of ice, which closely followed bedrock topography. Ice was channelled in between bedrock highs along troughs, some of which coincided with glacially eroded overdeepenings (Fig. 3-8b). These sliding conditions may have favoured glacial erosion by abrasion and quarrying (also referred to as plucking, a mechanism that erodes rocks by freezing and thawing of water in cracks below a glacier).

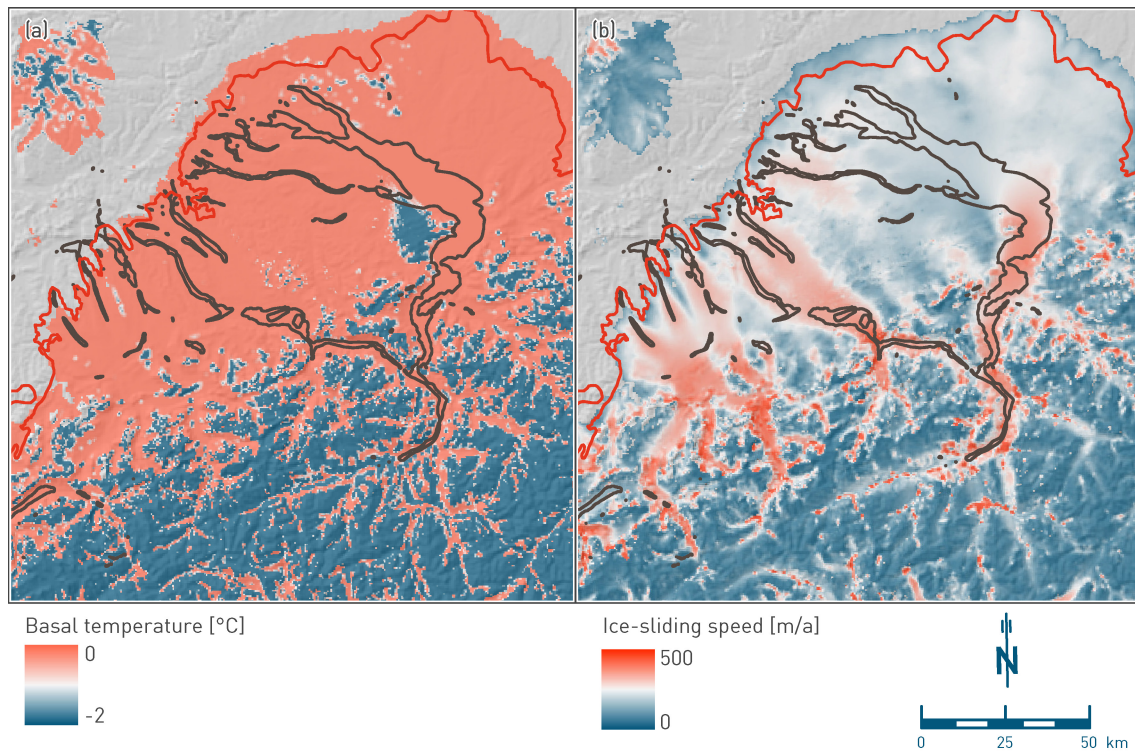


Fig. 3-8: The Rhine glacier system at the LGM

(a) Basal temperature. (b) Ice-sliding speed. See Fig. 3-4 for location. The brown lines denote the outlines of existing overdeepenings. The LGM outline modified after Ehlers et al. (2011) is shown as red lines.

3.3 Estimation of erosion potential

Based on the glaciological quantities derived in Section 3.2 (e.g. ice-surface geometry in Fig. 3-3 and Fig. 3-4, basal sliding velocity in Fig. 3-8b and areas of warm-based conditions in Fig. 3-8a), the potential for erosion is estimated.

3.3.1 Erosion by moving ice

Assuming that the glacial erosion rate scales with the basal sliding speed, the sliding distance calculated by time-integrating the sliding speed (Näslund et al. 2003, Staiger et al. 2005, Yanites & Ehlers 2016) yields a metric for glacial erosion by moving ice. Transient ice-flow modelling of the Rhine glacier system over the course of the last glacial cycle (Section 3.2) indicates that the sliding distance is greatest at the base of Alpine valleys where sliding is fastest and where ice-occupation time is longest (Fig. 3-9a). However, the sliding distance does not correlate well with the location of existing overdeepenings in the Alpine foreland (Fig. 3-9a) owing to the smaller sliding speeds near the ice margin and the short time period during which ice covered the distal foreland (Fischer et al. 2021).

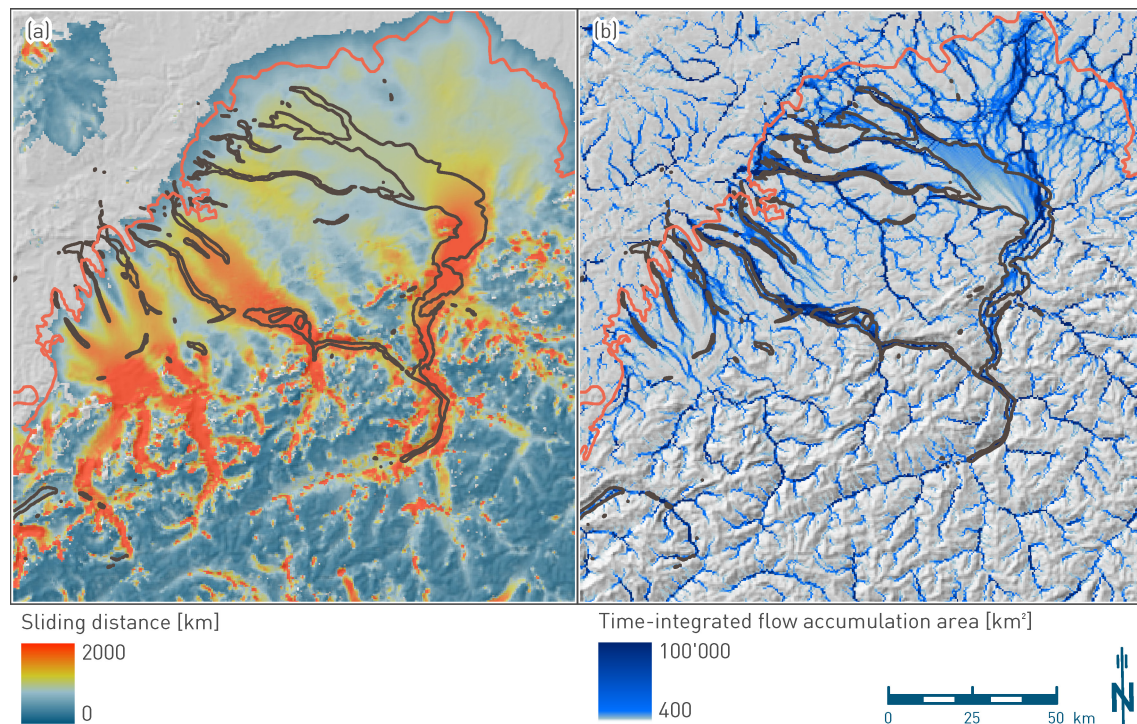


Fig. 3-9: Metrics for glacial erosion during advance and retreat of the Rhine glacier system in the foreland

(a) Sliding distance calculated by integrating the sliding speed over the course of the last glacial cycle as a proxy for glacial erosion by moving ice. (b) Time-integrated flow accumulation area over 12.8 kyr around the LGM as a proxy for glacial erosion by subglacial water flow. See Fig. 3-4 for location. The brown lines denote the outlines of existing overdeepenings. The LGM outline modified after Ehlers et al. (2011) is shown as red lines. The range of the sliding distance was restricted to 0 – 2'000 km to better show the variations in the Rhine glacier lobe.

3.3.2 Erosion by subglacial water flow

Another important process of erosion is by subglacial water flow (Alley et al. 2019, Swift et al. 2021). This may be the dominant process for deep erosion towards ice margins and for the greatest erosion depths in linearly-concentrated incisions (Fischer & Haeberli 2010). Based on computed time-dependent ice-surface elevations, maps of hydraulic potential and associated flow accumulation area are calculated to estimate the location of subglacial water drainage routes and water fluxes (Chu et al. 2016). Time integration of the flow accumulation area yields a metric for erosion by subglacial water flow. Larger values of time-integrated flow accumulation area indicate zones with significant subglacial water flow that has the potential to enhance sediment removal and increase subglacial fluvial erosion. Results from simulations of the Rhine glacier system during advance and retreat in the foreland (Section 3.2) indicate that subglacial water discharge is high and focused along glacial valleys and existing overdeepenings (Fig. 3-9b) when water pressure underneath the Rhine and Linth glacier lobes is close to the ice-overburden pressure. These conditions are necessary for subglacial water to remove basal sediments, expose fresh bedrock, and favour further glacial erosion. Knowledge of the location and fluxes of subglacial water drainage is therefore useful to understand patterns of erosion beneath ice masses that do not otherwise relate to the sliding of ice (Cohen et al. 2023).

3.4 Sensitivity analysis of model components and parameters

In additional transient simulations, the sensitivity of the calculated sliding distance to different model components and parameters is tested. The intention of the sensitivity studies is to identify the relative importance of parameters and not to provide uncertainty bounds on the simulation results. The sensitivity to components and parameters that affect the climate forcing is investigated by using different ice-flow models and climate-proxy signals or by varying the assumption on NHIS height, the temperature and precipitation at the LGM and the length of period at LGM climate conditions. Also, the sensitivity to parameters of the ice-dynamical model is investigated. A model parameter that exerts a first-order control on the ice flow, and therefore may influence the dynamic response of the AIF, is the till friction angle in the Mohr-Coulomb sliding law. The list of sensitivity cases is given in Tab. 3-1. Components or parameters have been varied individually.

Fig. 3-10 illustrates the sensitivity of the calculated sliding distance in the siting regions as well as for the Bodensee region to each model component or parameter. The sensitivity cases are ordered by impact on the sliding distance.

Tab. 3-1: List of sensitivity cases and associated parameters / components varied

The Min and Max values used in the sensitivity analysis are based on assuming reasonable uncertainty bounds associated with the model input parameters. These bounds, however, have not been calculated from models.

Temperature departure from LGM indicates the temperature change in °C from the modelled time-slice simulation at the LGM. Additional time at LGM indicates the additional number of years with constant climate forcings (precipitation and temperature) identical to the LGM values. NHIS height indicates the %-change in height of the prescribed NHIS for the time-slice simulation at the LGM.

Parameter / Component description	Units	Reference value	Min, Max / Alternative
Temperature departure from LGM	[°C]	0	-1, +1
Precipitation at LGM	[%]	100	50, 150
NHIS height	[%]	100	67, 125
Additional time at LGM	[yr]	0	0, +2'000
Friction angle	°	30	25, 35
Climate proxy signal		EPICA	Bergsee
Ice-flow model		PISM	Elmer/Ice

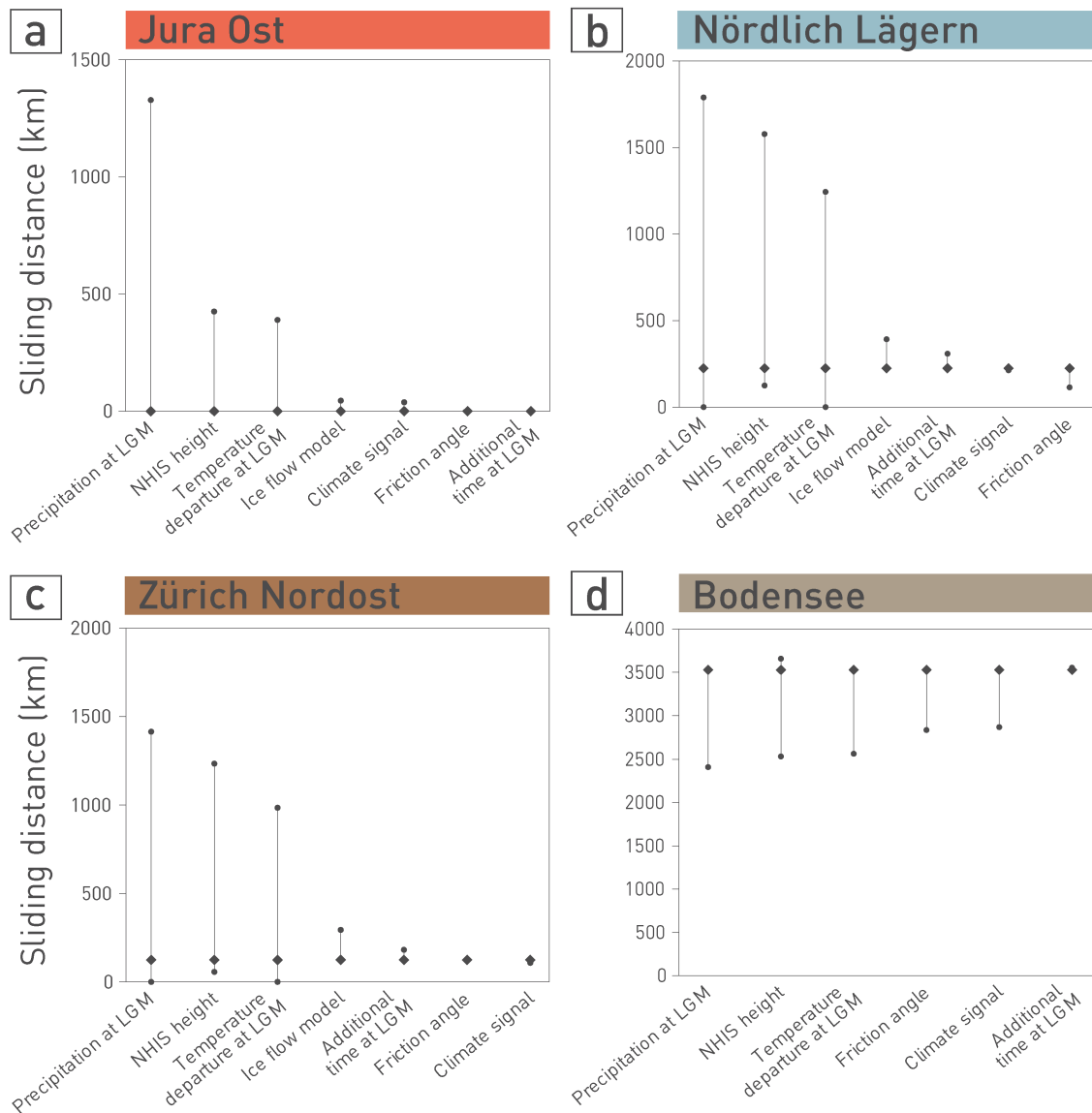


Fig. 3-10: Sliding distance in the (a) Jura Ost (JO), (b) Nördlich Lägern (NL), (c) Zürich Nordost (ZNO) siting regions as well as (d) for the Bodensee region for each sensitivity case

The results from the reference values are shown as a black diamond, with error bars representing the minimum and maximum sliding distance resulting from the Min and Max or alternative values (see Tab. 3-1 for details).

See Fig. 3-4 for the locations where the sliding distance was evaluated.

At all four locations the results are shown to be sensitive to precipitation. Increasing precipitation by 50% has a strong effect on the sliding distance for the siting regions near the ice margin, owing to increases in ice extent and thus ice occupation time there. A reduction of 50% in precipitation completely removes the ice cover at NL and ZNO.

Lower temperatures at the LGM especially enhance sliding distances in the siting regions because a colder climate increases the northern extent of the Rhine glacier lobe and thus the time of ice occupation in these regions. For the Bodensee location, a location away from the ice margin, a colder climate causes thicker ice there, which reduces the sliding speed owing to the effect of ice thickness on sliding, thus reducing the sliding distance there.

Surprisingly, additional time at LGM does not significantly increase the sliding distance. For JO, the site is still ice-free even with an additional cold period at the LGM of 2'000 years. Only slightly larger values of sliding distances are observed for NL and ZNO, while a smaller sliding distance is observed for BS. The latter is again due to thicker ice which reduces the sliding speed and, hence, the sliding distance.

Differences due to the climate signal (EPICA versus Bergsee) are either negligible (NL and ZNO) or small (JO and BS). For the Bodensee region, the Bergsee signal, with more rapid changes in climate, yields a smaller sliding distance, probably owing to a shorter ice-occupation time.

Simulation results with Elmer/Ice generally show higher values of the sliding distance than those with PISM because the latter tends to underestimate sliding at the base of thick ice (Bueler & van Pelt 2015).

Climate effects – temperature and precipitation – together with the thickness of the Northern Hemisphere Ice Sheet (NHIS) – have a much stronger influence than the friction angle (sliding coefficient) or the time at LGM conditions. Varying the friction angle has a weak overall effect but an angle of 30 degrees produces the highest sliding distances for all sites.

Concerning the sensitivity of sliding distances, the marked impact of NHIS height must relate to temperature and precipitation effects and, hence, involve redundancy with the effects from temperature and precipitation. A 100% NHIS leads to warmer temperatures and less precipitation compared to a 67% NHIS, and thus to smaller sliding distances for JO, NL and ZNO. In contrast, for BS, a 100% NHIS decreases ice thickness there, and thus increases the sliding speed and the sliding distance. Changing from a 100% to a 125% NHIS has only a minor impact on the modelled climate conditions over the Alps.

4 Future climate scenarios

The Quaternary climate during the last ~ 800 kyr was dominated by glacial/interglacial cycles of ~ 100 -kyr periodicity (see Chapter 2). Before this time (before the MPT), the climate went through comparable climatic cycles though with a ~ 41 -kyr periodicity. It is, hence, likely that the future climate evolution will also show a general cyclicity of glacial and interglacial states. In Section 2.3 it has been shown that climate simulations using the climate model detailed in Talento & Ganopolski (2021), with orbital parameters and atmospheric CO_2 content as climate forcing, can accurately reproduce climate during the last ~ 800 kyr. The climate simulations were also, with adjusted climate forcings, continued into the future. Parameters for orbital configurations can accurately be calculated for the past and future and are for example reported in Berger (1977, 1978) and Berger & Loutre (1997) (Fig. 4-1 and Fig. 2-1). The future atmospheric CO_2 content, however, cannot accurately be predicted as the additional anthropogenic emissions cannot be forecast and, hence, several emission scenarios were used to cover a range of likely scenarios.

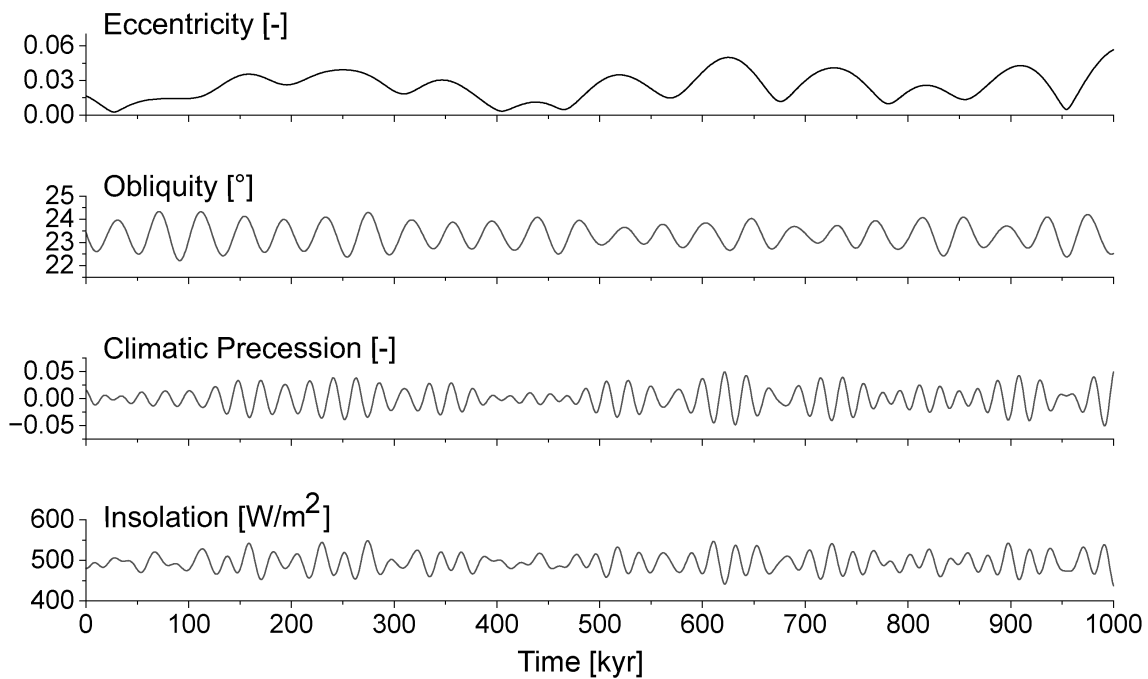


Fig. 4-1: Calculated orbital parameters and resulting insolation for the next one million years (from Berger & Loutre 1997)

Section 4.1 details the emission scenarios considered in this report. The results of the simulations using the different scenarios with focus on the onset of the next severe glaciation are discussed in Section 4.2 while Section 4.3 details the projected, continuous global climate simulations for the next one million years in terms of climates.

4.1 CO₂ emission scenarios

One of the main drivers for future climate evolution is the anthropogenic emission of greenhouse gases, mainly CO₂. Several scenarios were used to estimate the climate evolution in the future (Talentó & Ganopolski 2021; Fig. 4-2). The first scenario is a natural scenario with no anthropogenic emissions. Hence, the main driver for the climate simulations is orbital forcing. This scenario can be used to determine the impact of anthropogenic emissions and also reflects the situation that, by some kind of Carbon Dioxide Removal (CDR) by geoengineering and technological advances, all anthropogenic greenhouse gases are removed from the climate system. For the period 1750 – 2017, estimations place the fossil-fuel cumulative emissions in carbon equivalent at 660 PgC (Le Quéré et al. 2018). Projections for future cumulative emissions range between ~ 700 PgC and ~ 3'000 PgC, the former being an estimate of fossil-fuel reserves exploitable with today's technical and economic constraints and the latter an estimation considering resources that might become exploitable in the foreseeable future (McGlade & Ekins 2015). Taking into account the already achieved and future cumulative emissions estimates, three additional emission scenarios, next to the natural scenario, were used: low, intermediate and high emissions, corresponding to the instantaneous releases of 500 PgC, 1'000 PgC and 3'000 PgC, respectively.

The natural scenario would require a tremendous effort to achieve, given today's atmospheric CO₂ concentration. The 500-PgC scenario can currently be seen as a base scenario because it is close to current greenhouse gas concentrations in the atmosphere. The scenarios with even higher emissions are possible scenarios if no mitigation is implemented and current release rates continue (1'000 PgC) or a worst case where even currently uneconomic reserves are exploited (3'000 PgC).

In the simulations using the climate model described in Section 2.3, the CO₂ is instantly released at the beginning of the simulation (pulse emission). After the initial pulse, a multi-exponential decay function as proposed by Lord et al. (2016) is used to decrease CO₂ in the atmosphere, which resembles the distribution of CO₂ into the various sinks (causing an initial quick drop of CO₂ concentration; Fig. 4-2). Studies from Lord et al. (2016) and Archer et al. (2009) also have shown that CO₂ released from combustion of fossil fuels equilibrates among the various carbon reservoirs of the atmosphere, the ocean and the terrestrial biosphere on timescales of a few centuries. However, a sizeable fraction of the CO₂ (depending on the initial amount up to 20% or more) remains in the atmosphere for a long time (many thousands of years; Fig. 4-2, see also Section 2.3).

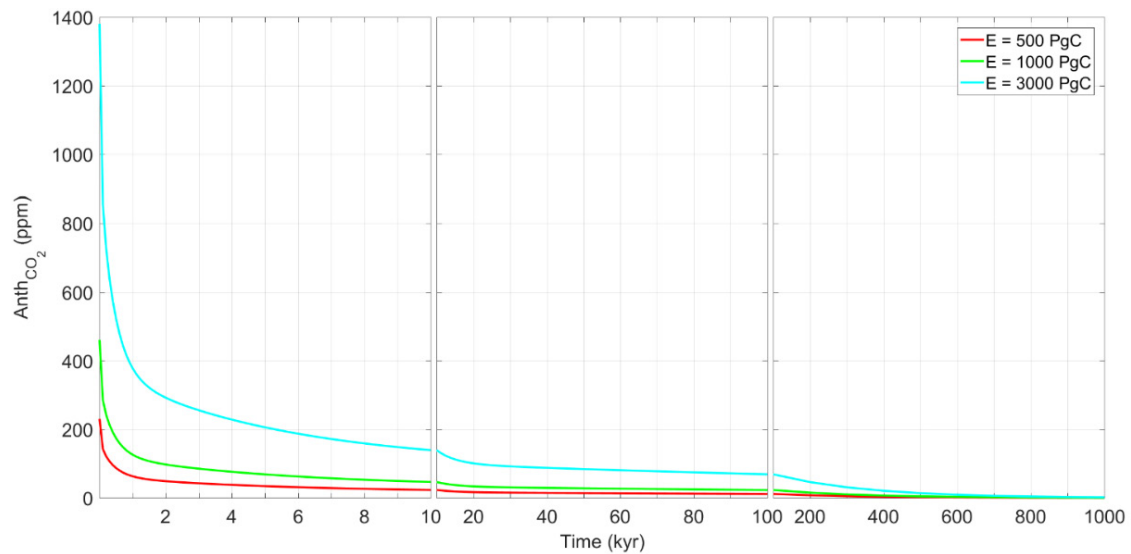


Fig. 4-2: Atmospheric CO₂ concentrations as a function of considered emission scenarios, with an instantaneous release of CO₂ at time $t = 0$ and a decay function as suggested by Lord et al. (2016)

If the duration of the emission pulse release is rather short (centuries) and is followed by zero CO₂ emission, the future long-term evolution of CO₂ depends primarily on the cumulative emission magnitude while the rate of release is only of secondary relevance (Eby et al. 2009). In other words, due to the instantaneous release, CO₂ dominates the climate evolution at the beginning while orbital forcing only has a minor influence. With decreasing CO₂, which happens quickly, orbital forcing becomes more and more dominant. Anth = Anthropogenic.

4.2 Onset of the next full glacial conditions

One relevant question with respect to future erosion scenarios is the timing of the next glaciations in the future and their severity. Since the CO₂ emissions have a large effect on the climate in general, they also play a major role in when and how severe the next glaciation will be. The global LGM ice volume is considered as a reference and all simulated ice volumes are normalised to the LGM ice volume (ice volume excludes Greenland ice). Talento & Ganopolski (2021) define the onset of a glaciation based on the simulated ice volume. If the (normalised) ice volume reaches 0.5 they consider this as a new state of full glacial conditions. If the (normalised) ice volume reaches a value of 0.8 they consider this as a major glaciation (depicted as light and dark coloured bars in Fig. 4-3). The full glacial conditions are of major interest here because it is likely that only ice volumes during full glacial conditions will reach far into the Alpine foreland.

The simulated ice volume obviously strongly depends on the emission scenario. Only after several 100 kyr, when enough CO₂ has been removed from the atmosphere, the ice volume for the different emission scenarios used here follow a comparable path, indicating a long-lasting impact of fossil-fuel CO₂ releases (Fig. 4-2 and Fig. 4-3).

In the natural scenario ($E = 0$ PgC), the next full glacial condition (where simulated ice volumes reach at least half the LGM ice volume) is simulated to peak around ~110 kyr. It will, however, not reach LGM volumes (maximum simulated ice volume is 0.84 for the first major glaciation). Cyclicity of glaciations roughly follows the 100 kyr cycle so that the second next full glacial condition is simulated after 200 kyr and so on (Fig. 4-3). Simulated ice volumes are lower than during the LGM in the natural scenario but reach values of > 0.9 several times at, and after, ~420 kyr. A small irregularity develops at 500 kyr which is even more pronounced in the low

emission scenario. Glaciations are slightly postponed compared to the general 100-kyr cyclicity. This irregularity is explained by the strong linear relationship between radiative forcing (insolation) and CO₂ concentration assumed in the simulations (see Talento & Ganopolski 2021 for a more detailed explanation). In the period from 400 – 500 kyr, insolation values (as a result of the Earth's orbital parameters) are comparatively high and show very little variability (Fig. 4-1). Thus, even when using a natural emission scenario, simulations do not result in full glacial conditions during this period.

In the low emission scenario ($E = 500$ PgC), the next full glacial condition is postponed by ~90 – 100 kyr compared to the natural scenario with a first occurrence of ice volumes > 0.5 at around 210 kyr and a peak at ~220 kyr with ice volumes of 0.8. Simulations again show roughly a 100-kyr cyclicity and a marked irregular development between 400 and 500 kyr (Fig. 4-3). After 500 kyr, glaciations follow the same trend as in the natural emission scenario ($E = 0$ PgC) because CO₂ in both scenarios after this time is very similar.

For the intermediate emission scenario ($E = 1'000$ PgC) the effect of anthropogenic CO₂ emission on the Earth system is even more long-lasting. Significantly different conditions from the natural scenario in the next half million years are indicated, with a very low probability of considerable ice growth during this time span (Fig. 4-3). Full glacial conditions only occur at ~650 kyr with simulated ice volumes of just 0.5 after which simulations indicate an evolution similar to the other, lower emission scenarios.

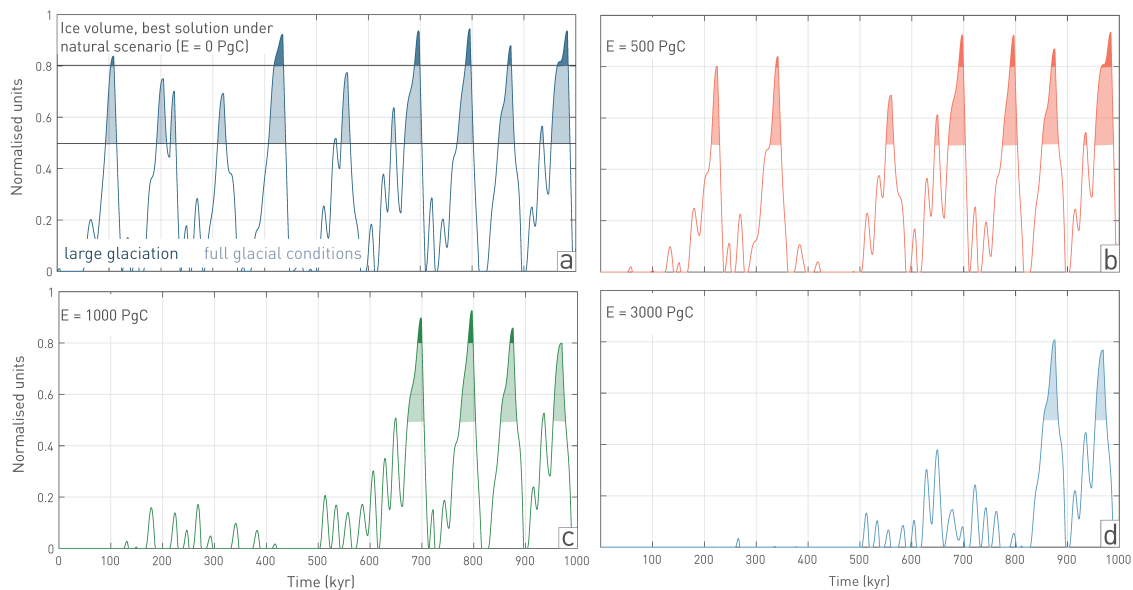


Fig. 4-3: Simulated landmass ice volume (normalised to the maximum LGM ice height) in the northern hemisphere (excluding Greenland) for the different emission scenarios (after Talento & Ganopolski 2021)

Shading depicts a glaciation (ice volume ≥ 0.5 , light colours) and a major glaciation (ice volume ≥ 0.8 , dark colours) as defined in Talento & Ganopolski (2021). Only the best-estimate scenarios are shown here. See Talento & Ganopolski (2021) for more details.

Finally, with a cumulative emission of 3'000 PgC, simulations indicate mostly ice-free conditions for the next 800 kyr (Fig. 4-3). Under this scenario, full glacial conditions are not expected to occur before 850 kyr.

The different timings for the onset of the next glaciation are in line with previously performed analyses where comparable boundary conditions were used. Williams et al. (2022) for example report the onset of the next glaciation under a natural scenario at between 50 and 100 kyr. Similar results have been found by e.g. Ganopolski et al. (2016), Archer & Ganopolski (2005) and Brandefelt et al. (2013, also references therein). All these studies agree that the higher the CO₂ concentration, the more the onset of the next glaciation is postponed.

4.3 Range of climate states expected for the next one million years

Local climate greatly depends on topography, latitudinal position (insolation), prevailing wind direction, altitude etc. It is therefore not possible to give time-dependent climate descriptions for the future climate evolution with high confidence, especially for a mountainous region like Switzerland. However, a general description of dominant climates for a glacial/interglacial cycle at the latitude of Switzerland can be given, although this necessarily must be greatly simplified. Polar, continental and temperate climates have all occurred in the past and are well documented by proxy data and simulations (see Section 2.2), so it is reasonable to assume that, for the more distant future, they are also representative for glacial/interglacial cycles. Fig. 4-4a shows modern climate regions in Europe according to the Köppen-Geiger classification scheme. Currently, polar climate is observed in higher northern latitudes, e.g. Greenland, northern Canada and northern Russia. Most summits of higher mountains also have polar climates (tundra, Fig. 4-4a), despite being in lower latitudes, due to their high elevations (e.g. Alps or Rocky Mountains). The continental climate is prevalent in intermediate northern latitudes and can be observed in large parts of Russia and northeastern Europe (Norway, Finland, Sweden etc). Currently, lowland Switzerland, or more general, central western Europe, is dominated by a temperate climate (Fig. 4-4b). Assuming high carbon emissions, Switzerland may also experience a dry climate at the height of an interglacial period in the future. A dry climate is defined as having little precipitation, most of which falls during spring and summer, and high average monthly temperatures throughout the year. Currently, parts of Spain experience temperate to dry climates. According to Beck et al. (2018), the dry climate will advance north-eastward in the next 100 years and reach the Pyrenees with arid to semi-arid conditions in most of Spain and Portugal (Fig. 4-4b). Climate estimations from proxy data in the past did not reveal a dry climate in the region of Northern Switzerland (see Section 2.2) hence, under low or natural emission scenarios, a dry climate is unlikely to occur in Switzerland during a future glacial/interglacial cycle. However, with continued carbon emissions and subsequent climate change it cannot be ruled out that, at least in the high-emission scenario, such a dry climate may occur even in Switzerland in the future.

The Köppen-Geiger classification scheme also includes a tropical climate with very warm monthly averages, low seasonal variability of temperatures and high amounts of precipitation. However, it is not considered likely that these climate conditions will, in the next one million years, be dominant in Switzerland.

During the waning of an interglacial and the beginning of a new glacial, the climate cools back to a polar climate. The exact timing of the warming and cooling is impossible to project into the future using the simulations detailed above. In the past, very short-lived climate phases (so-called Dansgaard-Oeschger events; e.g. Stocker 2000 and references therein) have been recognised. These describe a very rapid succession of cooling and warming where the cooling events lasted up to 3 kyr while the warming was extremely rapid and occurred within years or decades. A very prominent Dansgaard-Oeschger event occurred during the Younger Dryas dated at approximately 12 ka (Dansgaard et al. 1989, Velay-Vitow et al. 2024). Modelling of such short-lived events was not the focus of the long-term climate simulations performed here but it is possible that comparable events will happen in the future.

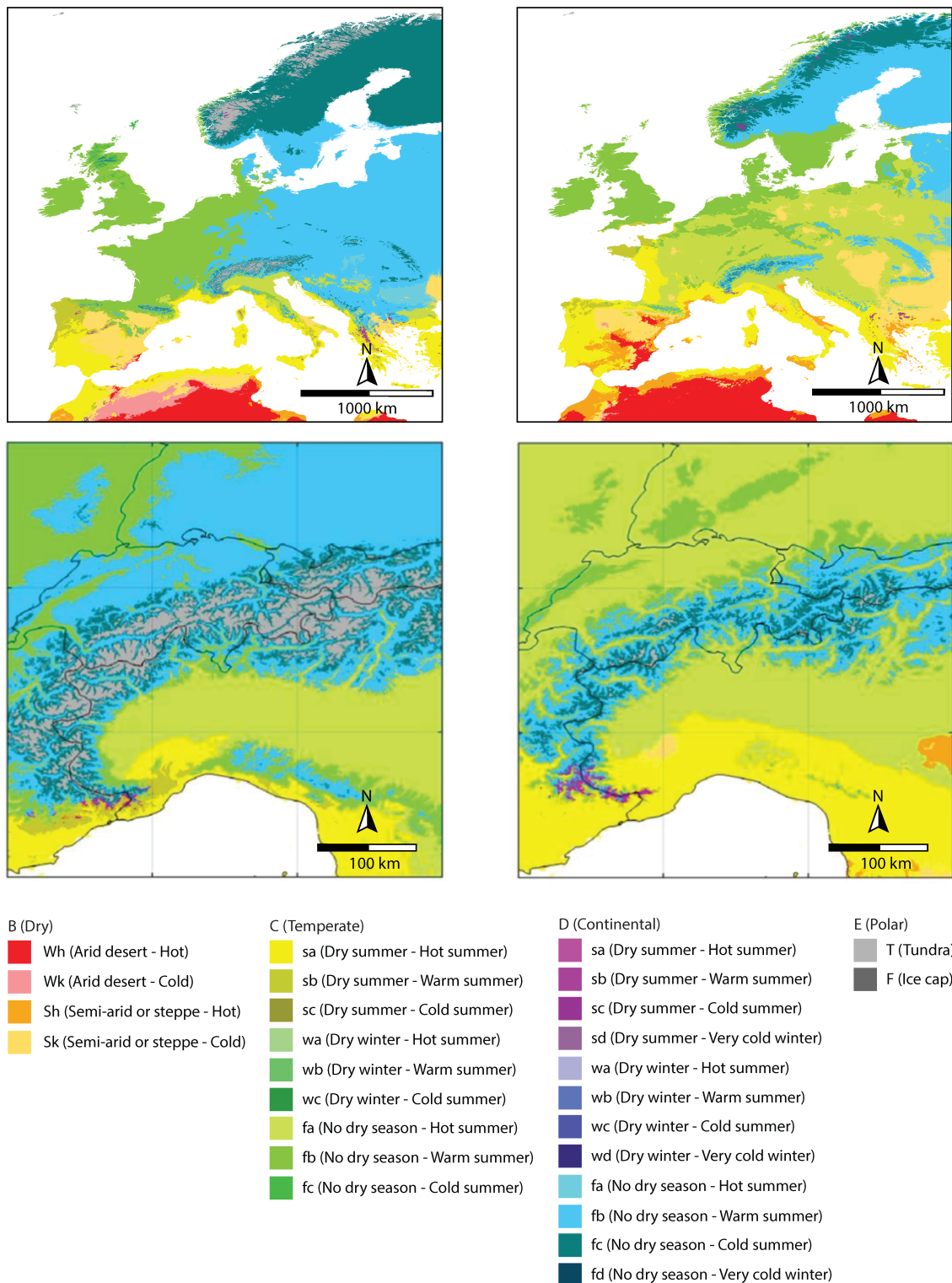


Fig. 4-4: Comparison of climate conditions at present day (left column) and at the end of the 21st century (right column) for Europe and the European Alps according to the Köppen-Geiger classification scheme

Left – Present-day results (averaged from 1980 – 2016). Right – Future results (from 2071 – 2100). Future climates were derived from absolute changes from 1980 – 2016 and projected into the future. See Beck et al. (2018) for more details.

5 Summary and conclusions

For the description of the long-term evolution of a deep geological repository, it is necessary to take into account natural processes that could negatively affect the geological transport barriers, e.g. by erosion or by changing the hydrological boundary conditions. Using climate scenarios based on proxy data and simulations, the possibility of future deep erosion of the barrier system can be explored. Special focus is placed on the timing and characterisation of future glacial climate conditions and related Alpine foreland glaciations because glacial climates have a particularly high potential to change the landscape, reduce the repository overburden and affect the hydrogeological transport system.

Understanding climates in the past, especially during the Quaternary which continues to the present day, is an important basis to simulate possible future climates. Climate simulations must be capable of capturing the glacial/interglacial cycles at least of the past ~ 800 kyr. At around ~ 800 kyr, the generally observed ~ 41 -kyr cyclicity before the Mid Pleistocene Transition changed to a ~ 100 -kyr cyclicity (see e.g. Clark et al. 2006 for more details). Here, for the global climate simulations, a process-based model taking into account only orbital forcing and atmospheric CO_2 concentration has been successfully used to simulate past (Fig. 2-6), and future (Fig. 4-3), glacial/interglacial cycles. The simulations show that the timing and severity of future glaciations strongly depends on the anthropogenic CO_2 emissions. For a natural emission scenario (no anthropogenic CO_2 emissions), the next full glaciation will only occur after $\sim 90 - 100$ kyr (Fig. 4-3). Any additional anthropogenic CO_2 emission will postpone the onset of the next glaciation into the future so that with approximately current atmospheric CO_2 concentrations, the next full glacial conditions will only occur in ~ 220 kyr. The occurrence of glacial conditions is further postponed (to ~ 650 or ~ 850 kyr) with even higher emission scenarios. The actual onset of the glaciations is obviously earlier than the peak of full glacial conditions. E.g., for the natural emission scenario, the onset of the next glaciation might already be in ~ 50 kyr while the peak of the glaciation will only be reached between ~ 90 and 100 kyr. The simulations also show the large influence of greenhouse gas emissions. Greenhouse gas emissions were introduced as pulses at the beginning of the simulation (Fig. 4-2). The greenhouse gas is then distributed into the different sinks, thereby lowering the atmospheric greenhouse gas concentration towards natural levels. On a global scale, before atmospheric concentrations are restored to natural levels, they force the climate towards warmer and, depending on geography, also drier conditions with minor contributions from the orbital parameters.

According to the modelling studies presented here and proxy data from key sites within the region, climate during the interglacials will warm from the dominantly polar climate during glaciations towards a temperate and possibly even a dry climate. The duration of each of the interglacials is highly variable and strongly depends on atmospheric CO_2 concentrations and the orbital parameters. Projections of climate into the near future from Beck et al. (2018) showed that, given that the CO_2 emissions follow the trend of the last decades, arid conditions are expected close to the Pyrenees within 100 years (Fig. 4-4). It is, hence, possible that, at least in the high CO_2 emission scenarios, a dry climate will also be prevalent at the height of an interglacial in the area of the siting regions.

Glaciers in the past have reached far into the Alpine foreland and repeatedly covered considerable parts of Northern Switzerland. As glaciers have the potential to strongly alter the landscape by erosion it is important to understand past glacier dynamics and to quantify the erosion potential. To evaluate this, the global climate simulations have been downscaled to a regional scale covering Switzerland and the Alps and climate information from these simulations have been used to analyse past glacier dynamics and their erosion potential. In accordance with proxy data, ice-flow simulations show that not all three siting regions are equally likely to be covered by ice. While

ZNO and parts of NL were frequently ice-covered, JO mostly remained ice-free during most past glacial cycles (Fig. 3-4). Ice coverage of some of the siting regions appears to have been in the order of 200 – 400 m and with only limited duration.

The simulations also show that glacial erosion is not simply a function of ice volume but is dependent on several factors, the most important being ice sliding speed and subglacial water flow. Assuming that the glacial erosion rate scales with the basal sliding speed, the sliding distance is a good proxy for glacial erosion by moving ice. Transient ice-flow modelling of the Rhine glacier system over the course of the last glacial cycle show that the sliding distance is greatest at the base of Alpine valleys where sliding is fastest and where the duration of ice occupation is long. However, the simulated locations of the highest sliding distances do not correlate well with observed locations of intensive erosion (overdeepenings) in the Alpine foreland (Fig. 3-9a). Furthermore, results from the simulated advance and retreat of the Rhine glacier system in the Alpine foreland indicate that subglacial water discharge is high and focused along glacial valleys and existing overdeepenings (Fig. 3-9b). It is therefore likely that subglacial water flow is more important for the observed deep erosion towards the ice margins and for the greatest erosion depths in linearly concentrated incisions (overdeepened valleys) than erosion by the moving ice itself.

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