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Impact of long-term climate change on a deep geological repository for spent nuclear fuel

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Abstract

The radioactivity of spent nuclear fuel will decay over a period of time (100,000 years or longer) in which we expect major environmental change. Climatically driven changes such as glaciation, permafrost and changes in sea level will affect the subsurface environment and must therefore be considered in performance and safety assessments.

We regard the state of climate to be determined by the climate system, comprising the atmosphere, the biosphere, the oceans, the ice sheets and the surface of the lithosphere. Climate can change as a consequence of external forcing or as a consequence of the internal dynamics. The changes in the past show a complex cyclical pattern with repetitive periodicities and magnitudes of change. On the relatively long time scale which is of concern in this report climate changes are believed to be triggered by changes in insolation due to cyclical changes in the earth's orbit around the sun. This is referred to as the Milankovitch theory or the astronomical climate theory. A brief review of our knowledge of the climate system, proxy records and observations of past climate change is presented in the report.

The most extreme departures from modern environmental conditions in Sweden have occurred during the cold, glacial cycles with ice sheets covering the whole of Sweden.

As part of SKB's palaeohydrogeological research programme a time-dependent, thermo-mechanically coupled model of ice sheet behaviour has been developed in order to simulate past fluctuations of the Scandinavian ice sheet and forecast its future.

The ice sheet model can calculate the temperature field and isostatic response of the underlying bedrock, subglacial and proglacial permafrost and the subglacial melt rates. By connecting the glaciation model with hydrogeological and rock mechanical models, the response of the subsurface to climate change can be investigated.

In this report the climate-driven environmental changes are represented as a series of successive climate-driven process domains. The process domains defined are the glacial domain, the permafrost domain and the temperate/boreal domain. Within each of these domains different regimes and subregimes have been identified, reflecting significantly different combinations of typical basic processes.

These domains, regimes and subregimes are meant to depict the prevailing conditions within a certain area during a certain time period. The evolution at a studied site can then be decribed as a series of domains, regimes and subregimes.

The climate-driven environmental processes likely to be of greatest significance for the performance of the geological barrier are:

- freezing
- loading by a glacier
- enhanced rates of groundwater flow
- changes in groundwater recharge chemistry
- changes in relative level of the sea and its salinity.

In this report the processes for each of the domains are described qualitatively and quantitatively as divided into biosphere, thermal, hydrological, mechanical and chemical conditions. Especially, the impact on the geological barrier is described.

The glaciation model has been applied to simulate the last 700,000 years and the next 200,000 years of environmental change along a flowline transect from the Norwegian coast, through Sweden and Denmark to northern Germany. The results are presented as domains and regimes in time and space and as properties such as ice thickness and basal temperature which vary continuously in time along the transect. Furthermore, site-specific climate-driven boundary conditions (ice thickness, head gradient, basal melt rate and basal temperature) have been calculated for an inland site and a Baltic coast site, respectively.

Sammanfattning

Radioaktiviteten hos det använda kärnbränslet kommer att avta under en tidsperiod (100 000 år eller längre), då vi förväntar oss stora förändringar i vår miljö. Klimatrelaterade förändringar, såsom inlandsisar, permafrostförhållanden och förändringar i havsnivå kommer att påverka förhållandena i berggrunden och måste därför behandlas i funktions- och säkerhetsanalyser.

Klimatet bestäms av klimatsystemet vars komponenter är atmosfären, biosfären, oceanerna, inlandsisar och glaciärer och ytan på litosfären. Klimatet kan ändras på grund av yttre påverkan eller genom klimatsystemets inre dynamik. Historiska klimatförändringar har visat ett komplext cykliskt mönster där cykler med olika frekvens och magnitud överlagrat varandra. I den relativt långa tidsskalan som är av intresse i denna rapport, är den gängse uppfattningen att det är variationer i solinstrålningen pga av cykliska förändringar av jordens bana kring solen som föranlett klimatförändringarna. Detta brukar kallas för Milankovitch-teorin eller den astronomiska klimatteorin. I rapporten ges en översiktlig sammanfattning av vår kunskap om klimatsystemet samt olika geologiska arkiv med information om de klimatförändringar som skett i det förflutna.

Den största avvikelsen från nuvarande förhållanden i Sverige har inträffat under tidigare kallperioder med inlandsisar som täckt hela Sverige. Som en del av SKBs paleohydrogeologiska forskningsprogram har en tidsberoende, termo-mekaniskt kopplad modell av en inlandsis utvecklats i syfte att simulera fluktuationer för en Skandinavisk inlandsis i förfluten tid samt för att förutsäga förhållandena vid ett sådant istäcke i framtiden.

Ismodellen kan prediktera temperaturfältet i och under isen, isostatisk påverkan på den underliggande berggrunden, förekomsten av subglacial och proglacial permafrost samt subglaciala smältvattenflöden. Genom att koppla denna modell till hydrogeologiska och bergmekaniska modeller kan den kompletta responsen hos berggrunden på klimatförändringar undersökas.

I denna rapport behandlas de klimatrelaterade förändringarna i miljön som en serie av successiva klimatstyrda processtillstånd. Processtillstånden som definieras är glacialt tillstånd, permafrosttillstånd och tempererat/borealt tillstånd. Inom vart och ett av dessa tillstånd har olika regimer och underregimer definierats, vilka uppvisar signifikant olika kombinationer av typiska grundläggande processer. Med dessa processtillstånd, regimer och subregimer kan de rådande förhållandena inom ett specifikt område beskrivas för en bestämd tidperiod. Utvecklingen för en plats kan beskrivas som en serie processtillstånd, regimer och subregimer.

De klimatstyrda processer som sannolikt kommer att ha störst betydelse för funktionen hos den geologiska barriären är:

- frysning
- belastning av isen
- ökade grundvattenflöden
- förändringar i grundvattenkemi
- förändringar i relativ havsnivå och havets salthalt

I rapporten beskrivs dessa processer för vart och ett av processtillstånden i första hand kvalitativt och men även kvantitativt, uppdelat på biosfär, hydrologiska, mekaniska och kemiska förhållanden. Speciellt beskrivs påverkan på den geologiska barriären.

Glaciationsmodellen har använts för att simulera förändringarna i miljön de senaste 700 000 åren och de kommande 200 000 åren längs en isflödeslinje från den norska kusten genom Sverige och Danmark ner till norra Tyskland. Resultaten presenteras i form av processtillstånd och regimer i tid och rum och i form av parametrar såsom istjocklek och marktemperatur under isen, vilka varierar kontinuerligt i tiden längs med den studerade linjen. Vidare har platsspecifika randvillkor (istjocklek, smältvattenflöde, tryckgradient och temperatur) beräknats för en plats i inlandet respektive en plats vid Östersjökusten.

Contents

1 Introduction

1.1 Background

Radioactive wastes are hazardous materials. If people are exposed to radiation, inhale or ingest radioactive particles they can get harmed. Measures must be taken to keep the waste separated from humans and environment. The separation can be achieved by burying the waste deep beneath the earth's surface in underground repositories.

The radiotoxicity of the waste is related to its radioactivity. Figure 1-1 illustrates how the radioactivity of spent nuclear fuel takes about 100,000 years of decay before it is comparable with the radioactivity of naturally occurring uranium. Underground repositories must provide safe disposal during this time period and even longer. The ability of a repository system to provide safe disposal is evaluated in performance and safety assessments. Long term performance and safety assessments must thus consider the effects of possible events and processes into the far future.

The radioactivity of spent fuel will decay over a period of time in which we expect major environmental change. During the last 100,000 years there have been major fluctuations of climate, including a glacial cycle during which ice sheets have extended from the Scandinavian mountains as far as the German Plain. There have also been major shifts of climate on timescales as short as 100 to 10,000 years. There is no reason to suppose that the next 100,000 years will not bring changes of climate of similar magnitude and frequency.

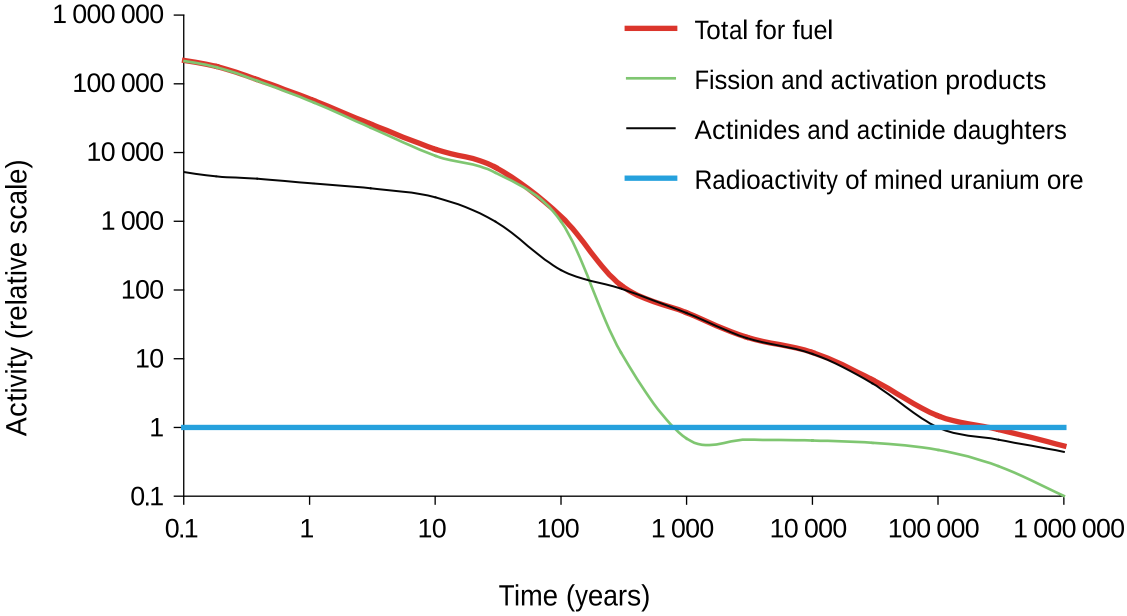


Figure 1-1. *The radioactivity of spent nuclear fuel in relation to the radioactivity in mined uranium ore /Hedin, 1997/.*

Climatically driven changes such as glaciation, permafrost and changes in sea level will affect the subsurface environment and must therefore be considered in performance and safety assessments. This report contains a summary of our current knowledge of the Quaternary climatic cycles. Climatically driven processes, which we believe may influence the performance of a repository in Sweden, are discussed. Results from the SKB palaeoclimate and palaeohydrogeology research programme /Wikberg et al, 1995/ are used to develop a practicable method whereby scenarios of future climate change can be formulated and incorporated into performance assessments.

1.2 Evaluating the impact of climatically-driven changes

The impact of climate change on repository performance can be thought of as three links in an interactive causal chain:

1. Interactions between climate and biosphere and topography/geology which determine the surface environment.
2. Interactions between surface environment and intrinsic properties of the geological barrier, which determine conditions and processes within the geological barrier.
3. Interactions between the geological barrier and the engineered barriers which influence conditions within a repository.

This report deals with link 1 and 2 in the chain. It does not evaluate impacts on individual engineered barriers or the effects on the integrated repository system.

Evaluation of the long-term future performance of a repository system in response to climate change requires a means of estimating future surface climates and their impact on subsurface environments as well as a means of establishing and testing confidence in these evaluations. Our approach depends upon an adage that "the future has happened before". The geological record contains an archive of past changes that reflect repetitive and fundamental relationships between climate, surface and subsurface environments.

- The repetitive changes of climate and surface environment, which have occurred in the immediately antecedent past, can be expected to continue into the immediate future.
- The frequency and magnitude of past climate change have shown systematic fluctuations that can be extrapolated into the future.
- Subsurface responses to surface environment can be described and modelled.
- Theoretically modelled relationships between climate and subsurface response are, in principle, amenable to testing through the approach of palaeohydrogeology, although this research is in its early stages.

These principles have formed the basis of a methodology to create scenarios of future climate change:

- Based on our knowledge of the past global changes the general, global characteristics of major climatic cycles can be described.
- Local or regional components (in this case Swedish) of global change can be inferred from correlation between global and regional changes.

- In a specific region, and for a specific regional climatic expression, there will be a characteristic environment, in which a series of processes tend to recur. We term these *climate-driven process domains*. Within each such domain, there are typical thermal, hydrological and mechanical conditions.
- A glacial cycle can be regarded as a time series of climatically driven process domains.
- Given the surface conditions the impact on the geological barrier can be computed.

Traces of past climate changes are kept for example in deep-sea and lake sediments, loess, peat bogs, stalactites and glaciers. Analysis of these so-called proxy records have exposed that climate change on earth follow a cyclic pattern, where cycles of different magnitude and frequency overlap each other. Long term climate change has been correlated with changes in insolation due to changes in the earth's orbit around the sun. As the latter can be calculated for the future, it is possible to make forecasts of future climate. Relationships between past climate changes and changes of climatically-driven environment in Europe, particularly the advance and decay of ice sheets, permit forecasts of future climate to be extended to create forecasts of future ice sheet growth and decay over Scandinavia.

- The evolution of a site can be described qualitatively as a series of climatically driven process domains.
- Thermal-hydrological-mechanical-chemical boundary conditions for the geological barrier can be identified as a discontinuous series related to domains or as a continuous, quantitatively expressed series of parameters derived from model simulations.

Appropriate tools and methods can be chosen for analysis of sub-surface consequences within the performance assessment.

1.3 Palaeohydrogeological research programme

To build confidence that assessments of future subsurface change based on climate forecasts are well founded, the capacity to reconstruct subsurface responses to past surface climate change can be tested.

The causal chain described in 1.2 can be computationally modelled. Its outcome is to predict a hydrogeological response to climate forcing. In principle, this can be tested against palaeohydrogeological reconstructions.

Palaeohydrogeology utilises observations of hydrochemical, isotopic, mineralogical and hydraulic properties of the modern groundwater system to infer its evolution in the past. It depends upon the identification of distinctive water chemistry or mineral chemistry which can be related to water masses from different sources, and whose spatial evolution can then be inferred. The waters, or their mineralogical effects, are derived from chemically distinctive and dated recharge events and the movement of waters from recharge to present state is reconstructed.

SKB has recently carried out a palaeohydrogeological research programme. The general objectives of this programme have been:

- to identify and improve the understanding of the principal climatically-driven processes that, over a time scale of 100,000 years, could affect the integrity of a deep waste disposal site, and

- to form a basis for developing climate-related scenarios for performance and safety assessment of a waste repository in the long-term future perspective.

As a major part of the research programme, a time-dependent, thermo-mechanically coupled model of ice sheet behaviour has been developed (see Appendix A and B) and a large number of model simulations have been made (see Appendix C). The model is able to predict the internal temperature and velocity fields of the ice sheet, the temperature field and the isostatic response of the underlying bedrock, subglacial and proglacial permafrost extent and the subglacial melting rate. By connecting the glaciation model to models of groundwater flow and rock mechanical impact, the response of the subsurface to climate change has been investigated. The progress made within the palaeohydrogeological programme over the last couple of years has been essential for the qualitative and quantitative descriptions of environmental effects and subsurface impact made in this report.

2 The repository system

Geological disposal is currently the most favoured strategy of handling nuclear waste. In the Swedish approach to geological disposal of spent nuclear fuel (entitled KBS-3), the spent fuel is isolated from people and the environment by a series of complementary barriers, which comprise a multibarrier system. A KBS-3 repository operates at two levels:

- **Level 1 - Isolation**
  The spent fuel is encapsulated in impervious canisters which are deposited deep in the bedrock. Isolation prevents radionuclides from entering the biosphere and coming into contact with man and environment during their decay.
- **Level 2 - Retardation**
  If the isolation is broken the radionuclides are retarded and retained by slow dissolution of the fuel, sorption and slow transport through the different barriers of the repository.

Furthermore the radionuclides are dispersed and diluted in the biosphere.

The spent fuel is contained in massive steel and copper canisters. The canisters are buried at 400 – 700 m depth in crystalline rock. To protect the canisters from mechanical and chemical strain a bentonite clay buffer surrounds them. A KBS-3 repository is believed to be able to maintain isolation for hundreds of thousands of years. Even if the isolation is broken people and environment are protected since the radionuclides are retarded and retained in the different barriers. Doses to man also depend on dispersion and dilution in the biosphere. *Figure 2-1* summarises the barriers and their safety functions.

The function of the geological barrier (The rock, *Figure 2-1*) depends on a series of key variables, through which, for each moment, the performance of the barrier can be analysed. Different processes, both internal and external to the barrier itself, affect the variables. The geological barrier interacts with adjacent barriers and the environment through thermal (T), hydrological (H), mechanical (M) and chemical (C) processes. In order to provide a comprehensive and systematic documentation of the variables and the processes affecting them they have been structured in so called T-H-M-C-diagrams /SKB 1999/. The T-H-M-C-diagram for the geological barrier is shown in Appendix D.

2.1 The geological barrier

2.1.1 Isolation

The main function of the repository is to isolate the radioactive waste. The geological barrier contributes to the isolation by;

- making the waste inaccessible
- maintaining a relatively stable environment around the engineered barriers

This report explores the way in which climatically driven processes will impact the geological barrier. The results in this report contribute to the analysis of the capacity of the geological barrier to provide a chemically and mechanically stable environment. A stable environment in this context is not one in which no changes occur, but one in which changes do not prejudice the functions of the engineered barriers on a 100,000-year timescale.

The biosphere

The biosphere is not included in the repository system. The prevailing ecosystem is crucial when estimating radiation doses to man.

Dilution conditions, capacity of recipient to buffer, store or accumulate radionuclides, and land and water use influence radiation doses to man and environment.

Radionuclides are transmitted to humans via recipients for deep groundwater and local ecosystems..

The KBS-3 Repository System

A. The rock

Isolation;

- keeps the waste separated from man and environment
- supplies a stable environment for the engineered barriers

Retard and retain;

- limited water turn over and flow and thereby long transit times
- retains radionuclides by acting as a filter and buffer

B. Bentonite buffer

Isolation;

- acts by its rheological properties as a buffer against mechanical stresses
- prevents water flow and transport of corrodants to the canister
- buffers the chemical environment around the canister

Retard and retain;

- prevents water flow and transport of radionuclides from a leaking canister
- particles and solutes are captured via filtration and sorption

C. The copper and steel canister

Isolate;

- totally isolates the spent fuel
- the copper shell makes the canister watertight
- the steel insert provides mechanical strength

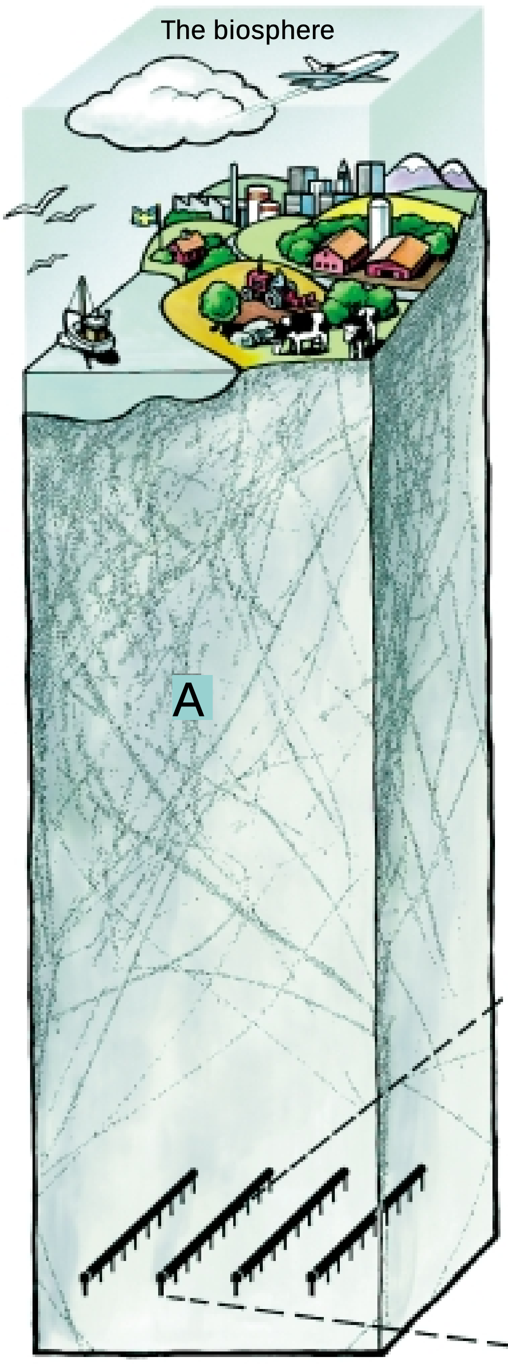
Retard and retain;

- undamaged parts of a leaking canister restricts transport of water in and out of a the canister

D. The spent fuel

Retard and retain;

- low solubility in water and low corrosion rate
- most of the radionuclides are tightly bound in the fuel structure and therefore difficult to dissolve



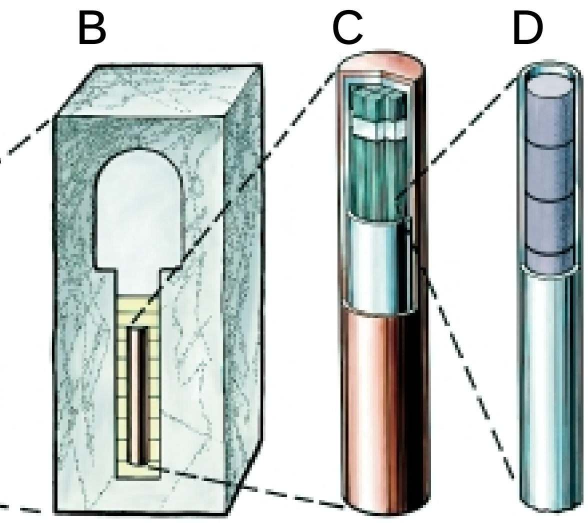


Figure 2-1. *The KBS-3 system for geological disposal. The role of the biosphere and the barriers and their function.*

Mechanical stability

Mechanical stability requires that;

- the movements on repository depth should be small
- there should not be rapid, large changes of load on the engineered barriers

It is believed that future changes in the stresses within the geological barrier can be taken up as movements in existing fractures. Studies indicate that there is a connection between fracture size and movement along the fracture /Turcotte 1992/. By adopting the construction and layout of the repository to the fracture system at the site, the properties of the geological barrier are used to provide a mechanically stable environment at the canister positions.

Chemical stability

Chemical stability requires that;

- long-term stable reducing conditions are maintained at the repository depth
- the rock mass provides an acceptable chemical environment for the buffer
- the rock mass is conducive to an acceptable chemical environment for the canister

Since water chemistry is an important determinant of chemical reactions in the geosphere, chemical stability requires a stable groundwater composition. The chemical composition of the groundwater mainly depends on the groundwater flow and the mixing of waters of different origins. Other factors affecting the groundwater chemistry are the composition of the original waters, the water/gas conducting properties of the geological barrier, and the reactions with rock and fracture minerals.

Water may be meteoric or marine in origin. Its composition may be changed through biologically mediated processes as it flows through soil layers where organic matter will tend to consume oxygen. At greater depths, reactions with rock and fracture minerals will affect its composition. The redox buffering properties of rock minerals is particularly important. At great depths one finds very salt water which is believed to be very old. This water can under certain conditions move upwards. Chemical conditions at repository depth should lie within limits that maintain chemical stability within the repository.

2.1.2 To retard and retain

If the isolation is broken, radionuclides leaking through the engineered barriers should be retarded and retained in the geological barrier. This is achieved by different and slow transport times to the surface and retention processes.

The capacity of the geological barrier to retard and retain radionuclides has a physical and a chemical part. The physical part has to do with the system of fractures and pores in the rock and groundwater movement. The chemical with the chemical properties of the groundwater, the rock and the radionuclides. The involved mechanisms are;

- *advection* – the transport of solutes solely by the movement of the groundwater
- *dispersion* – a mixing phenomenon produced by different flow velocities within and between fractures

- *diffusion* – the movement of solutes from areas of high concentration to areas of low concentration, especially important is the diffusion into the micropores of the rock
- *sorption* – a common name for processes when solutes fix on a solid phase

Variables of the geological barrier affecting these processes are conductivity, flow porosity, fracture pattern, hydraulic gradient, the contact area between water and rock, the porosity of the solid rock and the ambient chemical conditions.

These mechanisms, in combination with the repository depth and layout, contribute to long transport pathways and times, and to retention of radionuclides in the geological barrier. Long transport times permit radionuclides to decay before they reach the surface and retention processes will prevent some reaching the surface.

2.1.3 Variables and processes affected during a glacial cycle

During a glacial cycle major changes occur in the surface environment, changes that will affect the geological barrier. Performance and safety assessments should be designed to analyse the impact of these changes on barrier and repository performance and safety. The key question is whether the isolation, that is the integrity of the canister, is threatened. In case of leaking canisters it is also important to investigate how the retardation of radionuclides is affected. In chapter 4 of this report the range of impact on the surface environment and the geological barrier is described as an input to performance and safety assessments. Here the variables and processes classified in T-H-M-C terms (Appendix D) which are of importance for the function of the geological barrier are identified.

Thermal processes

The temperature is the only variable in the THMC-diagram of the geosphere that is affected by thermal processes. The temperature is a function of time and space. The temperature is affected by heat transport within the barrier, and by thermal exchange with the surface environment and the buffer/backfill.

Extremely low temperatures at repository depth could threaten the isolation of the waste if it causes freezing of the buffer material. The temperature has both direct and indirect influence on groundwater and groundwater movement. Freezing of water can cause enrichment of dissolved components in the unfrozen water, and thereby for example increase the salinity. The temperature affects the viscosity and density of the groundwater and thereby the groundwater flow. Via thermal expansion/contraction and freezing of water the temperature can cause changes of the variables fracture structure and aperture.

Hydrological processes

Variables in the THMC-diagram affected by hydrological processes are groundwater flow, groundwater pressure and gas flow. They are all functions of time and space. They interact with the surface environment and buffer/backfill through changes in water/gas movement and water/gas pressure.

The hydrological variables are very important for the retarding function of the barrier. Groundwater movement and pressure can affect the variables fracture structure and aperture and thereby, through the change of water/gas conducting ability of the rock, generate a feedback process. The movement and mixing of groundwater also influence the groundwater composition.

Mechanical processes

Variables affected by mechanical processes are fracture structure and aperture and rock stresses. The matrix porosity, that is the porosity of the solid parts of the rock, is to some extent influenced by precipitation and dissolution of minerals. The rock stresses, fracture structure and aperture vary in time and space. The internal processes rock creep, fracture formation, reactivation and thermal expansion/contraction affect them. They are also influenced by seismic activity in the surrounding rock masses, and by erosion, ice load and certain human actions on the surface. The rock stresses and fracture aperture around the tunnels and canister positions are affected by the swelling of the bentonite material in the buffer/backfill.

The fracture characteristics, rock stresses and the processes affecting them are obviously very important for the mechanical stability. The fracture structure and aperture are also very important for the groundwater movements in the rock, and thereby influence both the radionuclide transport and groundwater composition.

Chemical processes

Variables affected by chemical processes are groundwater composition, gas composition, rock mineralogy and fracture mineralogy. The chemical variables vary in time and space, sometimes to low extent and very slowly. The chemical variables are affected by a number of chemical processes, for instance precipitation and dissolution of minerals, changes in pH and redox conditions. Microbiological activity is also considered to be a chemical process. Some chemical processes depend on temperature and pressure. The chemical variables also depend on the movements and mixing of water and gas. Further more they are influenced by exchange with the buffer/backfill and by infiltration processes, vegetation and water and land use on the surface.

The variation of groundwater and gas composition, rock and fracture mineralogy determines whether the chemical conditions can be considered to be stable. They also influence the transport of radionuclides through the geosphere.

3 Relationships between climate and surface environment

3.1 Climate and climate change

The word climate originates from the Greek word "klima"; inclination, latitude. The climate is by definition a summary of weather conditions during a certain period. Climate at a specific location is usually described through statistical properties of the climatic elements, for example mean, maximum and minimum values. The most important climatic elements are air temperature, precipitation, humidity, air pressure and wind. Other parameters used to describe climate are cloudiness, fog, frost, ground temperature and sometimes the content of dust and gases in the atmosphere.

Processes in the atmosphere determine the *weather* but *climate* does not merely depend upon atmospheric phenomena. The physiography of the surface and the disposition of the continents determine the poleward energy flow paths. A very large proportion of the poleward energy flow takes place through the ocean, with ocean heat and pressure anomalies being strongly coupled to atmospheric flow. The nature of the surface (e.g. forest, grassland, ocean, snow or ice cover, sandy desert) determines the albedo, that is the extent to which solar radiation is absorbed or reflected by the surface. The continental and oceanic biosphere and processes of weathering play fundamental roles in governing the concentration of atmospheric gases, including greenhouse gases which absorb and re-emit solar radiation in such a way as to warm the lower atmosphere.

As a consequence, we now regard the state of the climate to be determined by the *climate system*, comprising the atmosphere, the biosphere, the oceans, the ice sheets and the surface of the litosphere (Figure 3-1). The climate at a specific location is largely determined by:

- The site's latitude and elevation
- The site's location in relation to the sea
- The site's location in relation to the major poleward fluxes of heat
- The radiation absorptive capacity of the surface and of atmospheric gases

Climate can change as a consequence of *external forcing* or as a consequence of *internal dynamics*, when one of the components of the climate system changes. Sources of external forcing are:

- Dissipation of internal earth energy producing shifts in earth physiography (plate movement, mountain uplift) or volcanism
- Changes in insolation

Changes in climate show a complex cyclic pattern, in which it is possible to discern repetitive periodicities and magnitudes of change. The spectrum of climate cycles is a consequence of different frequencies and amplitudes of forcing and different magnitudes of response. Earth physiography changes over timescales of 100,000 years to several million years. Changes in the earth's orbit around the sun causing variations in insolation have a periodicity in the order of 10,000 – 100,000 years. Changes in the activity of the sun show a wide spectrum of periods. Ice sheets vary over periods of 1000 – 100,000 years.

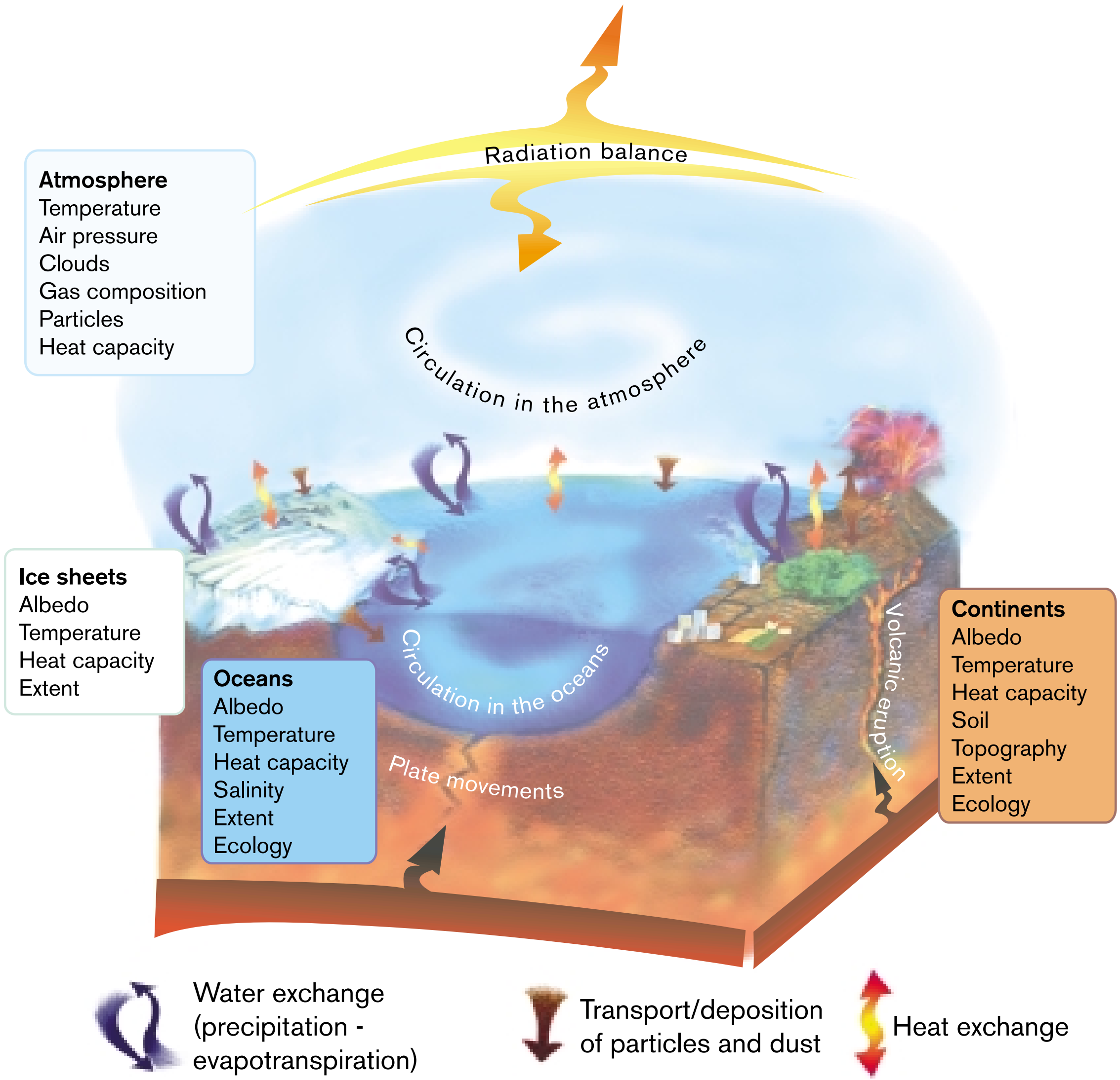


Figure 3-1. *The earth climate is formed through interactions between the different parts of the climate system.*

Deep ocean circulation changes over periods of 100 – 1000 years. The biosphere changes over periods of 1 – 1000 years while the atmosphere can reorganise itself over periods of the order of days to years.

On the relatively long timescale which is the concern of this report changes in external forcing take place as a consequence of cyclical changes in the earth's orbit around the sun. In the 1940's the astronomer Milankovitch calculated the changes in insolation due to the orbital variations and related them to long term climate change. Milankovitch suggested that changes in insolation cause ice ages on earth. Since the 1940´s the idea of orbital variations as the pacemaker of ice ages on earth has been broadly accepted even if some ambiguities still exist. It is referred to as the Milankovitch theory or the astronomical climate theory. The orbital parameters are:

- **Orbital eccentricity;**
  The earth's orbit changes from being almost circular to being elliptic. The periodicity is about 100 ka. Eccentricity is the only orbital variable that changes the total amount of solar energy received by the planet.

- **Axial obliquity;**
  The angle of tilt of the earth's axis in relation to the ecliptic varies between about 21.5° and 24.5°. The length of the cycle is about 41 ka. It affects the amount of irradiation reaching the poles and the temperature contrast between summer and winter.

- **Precession (or June 21 distance to the sun);**
  The motion of the spinning earth makes it wobble so that the axis of rotation sweeps out a cone. This process affects the season in which the Earth makes its closest approach to the sun. The distance between the earth and the sun at the Northern Hemisphere summer solstice varies with three frequencies, which in average have a periodicity of about 22 ka. This cycle affects the seasonal contrasts.

While – on the time scale with which we are concerned – Milankovitch variations are the dominant external forcing mechanism, it is the internal dynamics of the system that determine the response to this forcing. For instance the total effect of variations in eccentricity would at the surface of the earth affect the temperature by maximum 0.5°C /Crowley et al 1991 in Holmgren et al 1998/. The temperature during the last glacial maximum has been estimated to be about 5°C below the present /Jouzel et al 1993 in Holmgren

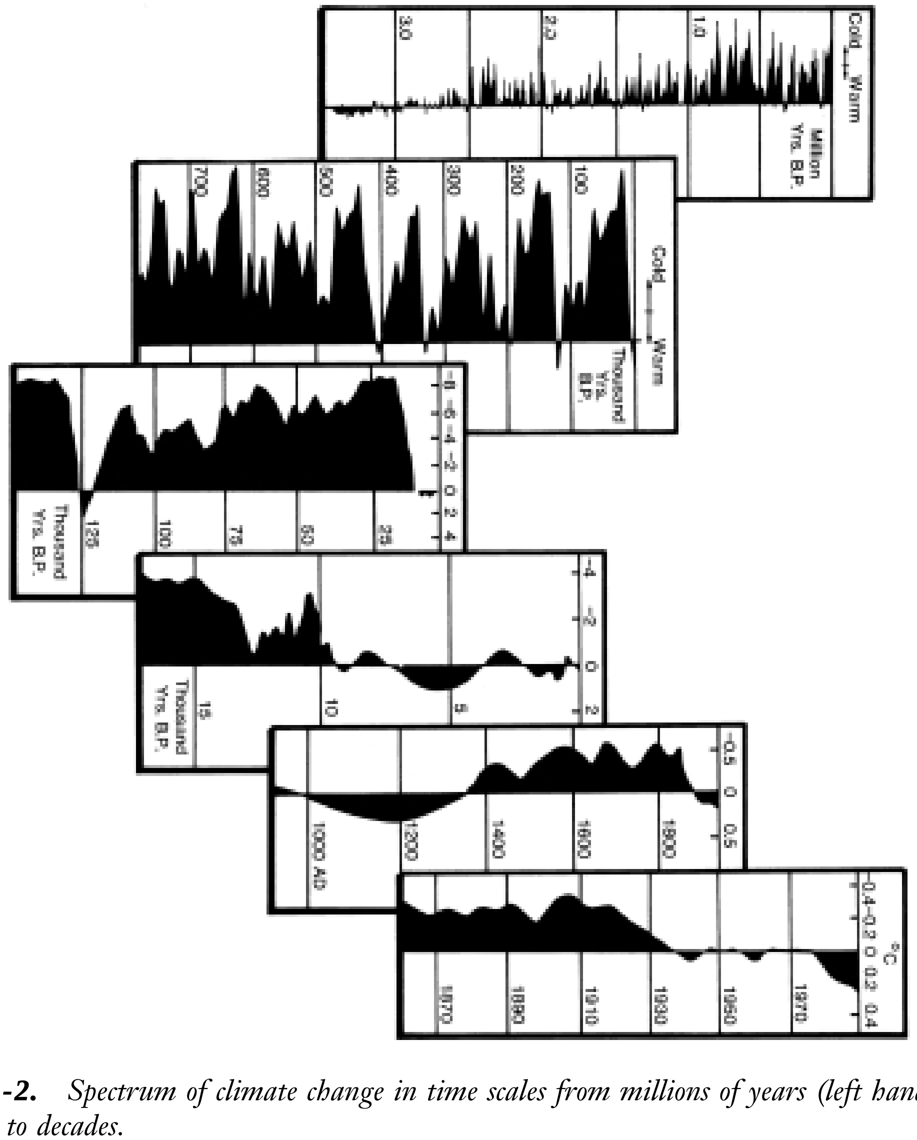


Figure 3-2. Spectrum of climate change in time scales from millions of years (left hand diagram) to decades.

et al 1998/. This suggests that positive feedback mechanisms within the climate system have amplified the solar signal to produce the observed changes in global mean temperature. Such feedback processes are numerous. For example, if Milankovitch forcing causes an ice sheet to grow, the increased albedo of the growing ice sheet will cool the atmosphere, and further enhance the rate of ice sheet growth. Figure 3-2 illustrates the spectrum of climate change, which reflects responses to all these processes.

There have been considerable advances over the last 25 years in understanding the operation of the climate system, both in characterising the details of change in the system and simulating the operation of the system computationally. This understanding of change is vital to the task of evaluating the future performance of a deep geological waste repository.

As major climate changes in Sweden are associated with the growth and melting of ice sheets a numerical model of ice sheet behaviour has been developed, in order to simulate past fluctuations of the Scandinavian ice sheet and forecast its future. The ice sheet model is described in appendix B. Climate and climate change and how to produce a climate drive for the ice sheet model is described in Appendix A.

3.2 Environmental effects of climate change

The most extreme departures from modern environmental conditions in Sweden have occurred during the cold, glacial cycles. Cooling climates during early glacial phases have led to:

- replacement of forest floras by shrub tundra
- extension of permafrost over much of the land area
- falling sea levels as glaciers and ice sheets have grown, principally in North America, but also in Eurasia
- growth of glaciers in the Scandinavian mountains

During the extremes of cold:

- growth of mountain glaciers into ice sheets which cover the whole of Sweden
- depression of the lithosphere beneath the ice sheet load
- compression and consolidation of rock and sediment masses beneath the ice sheet
- strong erosion beneath the ice sheet which disperses detritus towards its margin, mature soils are replaced by raw, unweathered mineral surfaces

During late-glacial periods of warming:

- ice sheets retreat through Sweden
- high relative sea levels occur immediately after deglaciation because of strong residual depression of the lithosphere as a memory of ice loading
- plant growth recommences

These surface environmental changes had major impacts on the water balance at the surface, the patterns of groundwater recharge and discharge, the driving heads for groundwater flow and the mechanical and transmissive properties of subsurface rock masses.

3.3 Characterisation of climate-driven process domains and regimes

The climate-driven environmental changes listed above are represented in this report as a series of successive *climate-driven process domains*. A process domain is defined as *a climatically determined setting on the earth's surface in which a series of processes habitually occur together*. The process domains identified are:

- *the glacial domain* where processes which influence subsurface conditions are dominated by the presence of glaciers
- *the permafrost domain* where the processes are predominantly influenced by the existence of permafrost
- *the temperate/boreal domain*, where processes which influence subsurface conditions are influenced by topography, precipitation, evaporation, changing coast line due to global sea level change and land depression and upheaval

Within each one of these domains, different *regimes* and *subregimes* can be identified which reflect significantly different combinations of the basic processes typical of the domain. For instance, in the glacial domain, the combination of processes in the ice divide region and the ice marginal region produce very different boundary conditions for rock mass and groundwater behaviour, and are therefore used to define different regimes. Within the ice marginal regime a further division into subregimes is required to describe the boundary conditions in detail.

The domains, regimes and subregimes are simplifications designed to facilitate tractable approaches to performance analysis. They concentrate on capturing dominant environmental processes and the key attribute of sequence. Although they may not capture all possible sequences, they do capture the modal types which observation and theory suggest exist. Viewed from the repository, domains, regimes and subregimes define the external, climatically driven conditions for groundwater flow.

The purpose of identifying process domains and regimes is to create relatively simple characterisations of the processes associated with a particular climatically determined surface environment. Our concern is to evaluate the effects of surface environments on subsurface conditions. Thus we have chosen to define specific domains and regimes by those attributes which are of most relevance to and have the greatest impact on thermo-hydro-mechanical-chemical processes in the subsurface. For instance we have not defined warm temperate and *boreal* (cold temperate) conditions as separate regimes, even though their fluctuations could be mapped through time and space. We believe transitions between temperate and boreal conditions are likely to produce only small changes in comparison with transitions between the domains and regimes.

The domains, regimes and subregimes are meant to depict the prevailing conditions within a certain area during a certain time period. They can be used as building blocks for scenarios of climate evolution. The evolution at a studied site can be described as a series of domains, regimes and subregimes. The identified domains, regimes and subregimes are listed in table 3-1, they will be described in detail in chapter 4.

Table 3-1. The different climate-driven process domains, their regimes and subregimes.

Climate-driven process domain	Regime	Subregime
Glacial		Ice divide -
	Melting zone	-
	Ice marginal	Proglacial permafrost and above marine limit
		No proglacial permafrost and above the marine limit
		Below the marine limit
Permafrost	Continuous	-
	Discontinuous	-
Temperate/boreal	Interglacial	-
	Preglacial	-
	Postglacial	Above the marine limit
		Below the marine limit

The long term climate changes of interest in this report are driven by orbital forcing including the parameters and periodicity mentioned above. The patterns of glaciations are known from a variety of geological records e.g. deep-sea sediment cores. These records show that the last 400 ka has been dominated by an about 100 ka long *glacial/interglacial cycle*. Beyond this time the 100 ka cycle appears less dominant /Holmgren et al 1998/. Each glacial/interglacial cycle includes a long cold phase, which abruptly ends with a sharp temperature increase. The long cold phase include colder and less cold periods termed *stadials* and *interstadials.* The warm periods between the long cold phases are called *inter-glacials.* They embrace about 8% of the whole glacial/interglacial cycle /Holmgren et al 1998/. Figure 3-3 shows the oscillations between cold and warm climate the last 700 ka (derived from SPECMAP – Spectral Mapping Project – data).

The extension of the different domains, regimes and subregimes vary throughout a glacial cycle. In reality there are no distinct boarders between them, either in time or in space. They are interlaced and coexist. During the most severe stadials of a glacial cycle the glacial domain covers the whole of Scandinavia and extends as far as to Poland and Northern Germany. During less cold stadials and interstadials the glacial domain can cover the north-western parts of Scandinavia while permafrost and temperate/boreal prevail to the south-east. The extension of the domains, regimes and subregimes in time and space can be calculated using the ice sheet model (Appendix B). Model output from a two dimensional simulation are shown in Figure 3-4.

it is widely believed, may drive the evolution of climate in the immediate future away from the natural trajectory. A model of the Northern Hemisphere climate has been used to investigate the impacts of different greenhouse gas scenarios /Berger et al 1996/. The model results suggest that the impact greenhouse warming will be to delay major climatic events during 10-thousands of years, but that subsequently, the changes will converge with the natural trajectory of climate.

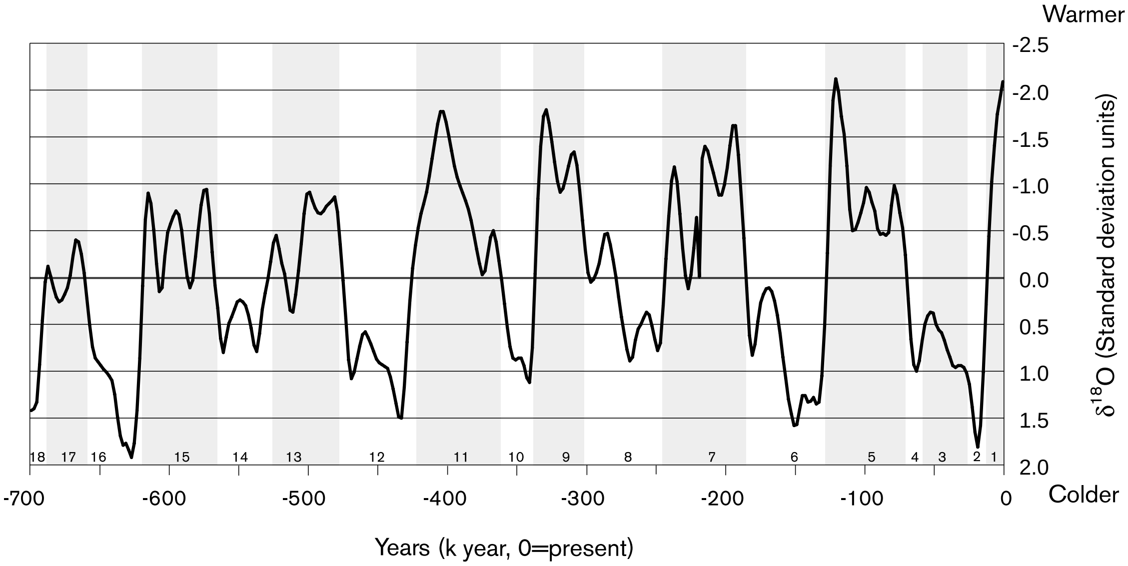


Figure 3-3. *Climate oscillations the last 700 ka as depicted in $\delta^{18}O$ variations /revised from Imbrie et al 1984/. The ratios of the stable oxygen isotopes ^{18}O and ^{16}O are expressed in the δ-notation. Variations in $\delta^{18}O$ in fossil foraminifera in deep-sea sediments reflect changes in oceanic isotopic composition. These changes are caused primarily by the waxing and waning of great ice sheets. High values mean greater content of the heavy isotope in the oceans and extended ice sheets. The grey and white fields and the figures at the bottom edge indicate warm and cold periods, called isotopic stages /according to Imbrie et al 1984/.*

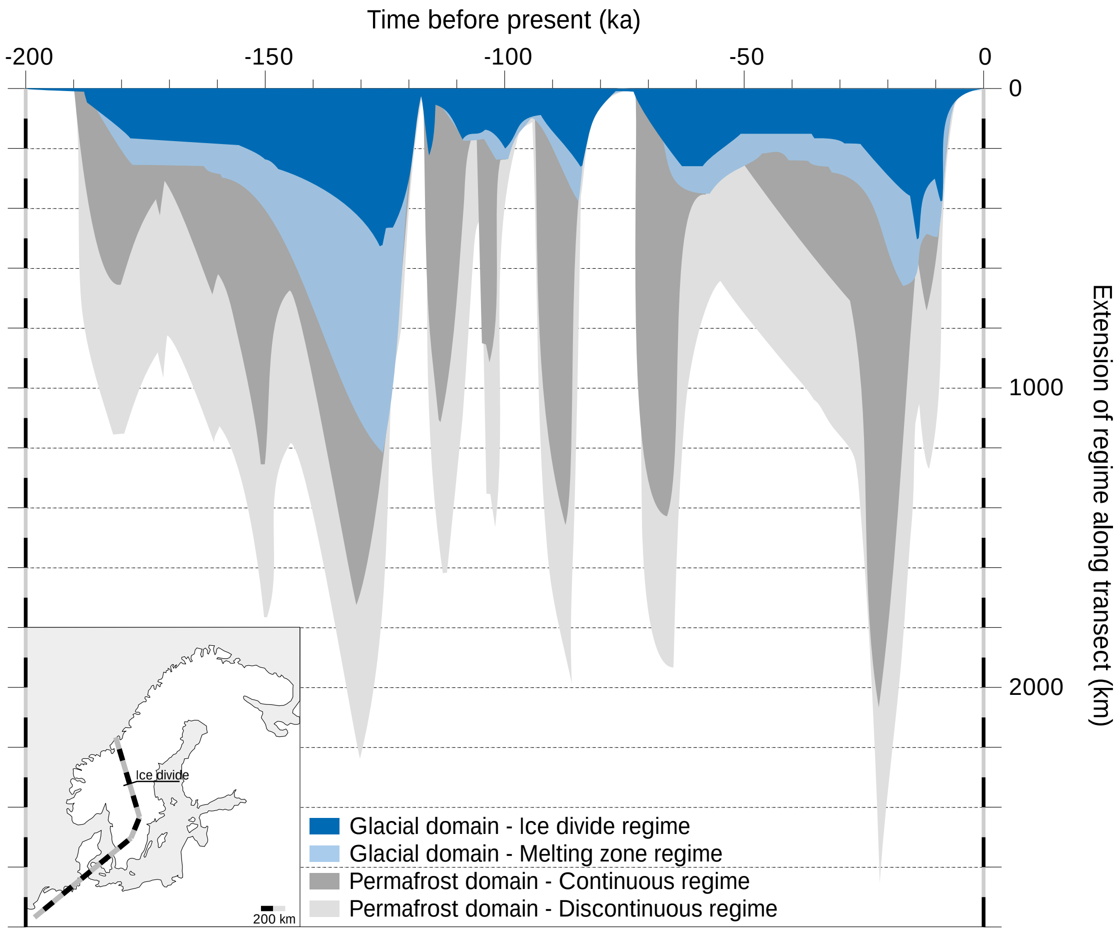


Figure 3-4. *The extension of the glacial and permafrost domains and regimes to the south of the initial ice divide zone.*

4 Surface environment and subsurface impact

In chapter 3 we presented an approach in which a climate scenario can be represented as the extension of climate-driven process domains, regimes and subregimes in time and space.

In this chapter, the environmental characteristics of each process domain, regime and subregime are described. Subsurface processes influenced by climatically driven surface environments are discussed, together with changes in the biosphere which might influence doses to humans.

The climate-driven environmental processes likely to be of greatest importance for the performance of the geological barrier are:

- freezing
- loading by a glacier
- enhanced rates of groundwater flow
- changes in groundwater recharge chemistry
- changes in the relative level of the sea and its salinity

Changes in Baltic Sea level and salinity occur as a consequence of changes in runoff and the relative shoreline. The relative shoreline depends on the global ice volume, which controls global "eustatic" sea level, and the regional state of crustal depression or uplift as a consequence of the history of ice loading and the location and elevation of the connection between the Baltic Sea and the North Sea.

We also review changes in climate of lesser magnitude, which occur in the temperate/boreal domain and which may influence performance of the repository.

The domains are now considered individually and the major climate-related conditions of importance for the repository system are described.

4.1 Temperate/boreal domain

The temperate/boreal domain is primarily characteristic of interglacials. There are no ice sheets in Sweden and no permafrost, except in mountainous areas. However, temperate/boreal conditions can also exist in the southern parts of Sweden during interstadials.

The climate in Sweden is influenced by cyclonic activity transporting oceanic, humid air, and determining the extent of maritime or continental conditions in Sweden. The Baltic Sea also buffers the temperature variations. Climate can change rapidly if the cyclon paths change.

During the current interglacial conditions, the average summer temperatures over the much of Sweden vary between 15 and 17 °C. In winter the average temperature varies from 0 °C in the southernmost Sweden down to –13 °C in the north.

Annual precipitation is between 500 to 800 mm in most parts of Sweden, except at upland locations. Maximum precipitation generally occurs during late summer and autumn, except for the southernmost part where there is a winter maximum. The variation between months is moderate. Snow constitutes about 20% of annual precipitation in southern areas, 25% in middle Sweden and up to 50% in northern Sweden. The annual runoff is between 200 and 500 mm in most parts of Sweden.

The period of snow cover is an important regulator of recharge. In the southernmost areas there is usually little frost and the dominant winter precipitation permits continuous recharge of groundwater reservoirs in winter. In the northernmost areas, frost starts early, preventing groundwater recharge during autumn, producing the main recharge period in spring and early summer.

Coastal areas are strongly affected by prevailing winds and low-high pressure changes, which influence sea level and the water-exchange. Large-scale climatic conditions also affect the freshwater discharge into the Baltic and the inflow of marine water, thus determining the salinity of surface water and the nature of the ecosystem of the Baltic Sea.

We also include in this domain the conditions prevailing during the initiation of ice sheet growth as well as the conditions directly following glaciation. For these periods, the main characteristics depend on relative sea level, such that climate can diverge strongly from the modern interglacial climate.

As ice sheets grow, water is transferred from oceans to ice sheets reducing global sea levels. The deep ocean stratigraphic record of change in O-isotope composition reflects isotopic fractionation between ice sheets and the ocean, and gives a primary record of global ice volume changes. Given that the eustatic lowering of sea level (lowering due to changing ocean water volume) at the last glacial maximum was about 120 m /Fairbanks et al 1992/, and assuming an approximately linear relationship between ocean volume and sea level, permits us to translate an oxygen isotope record into an approximate record of eustatic sea level (Figure 4-1).

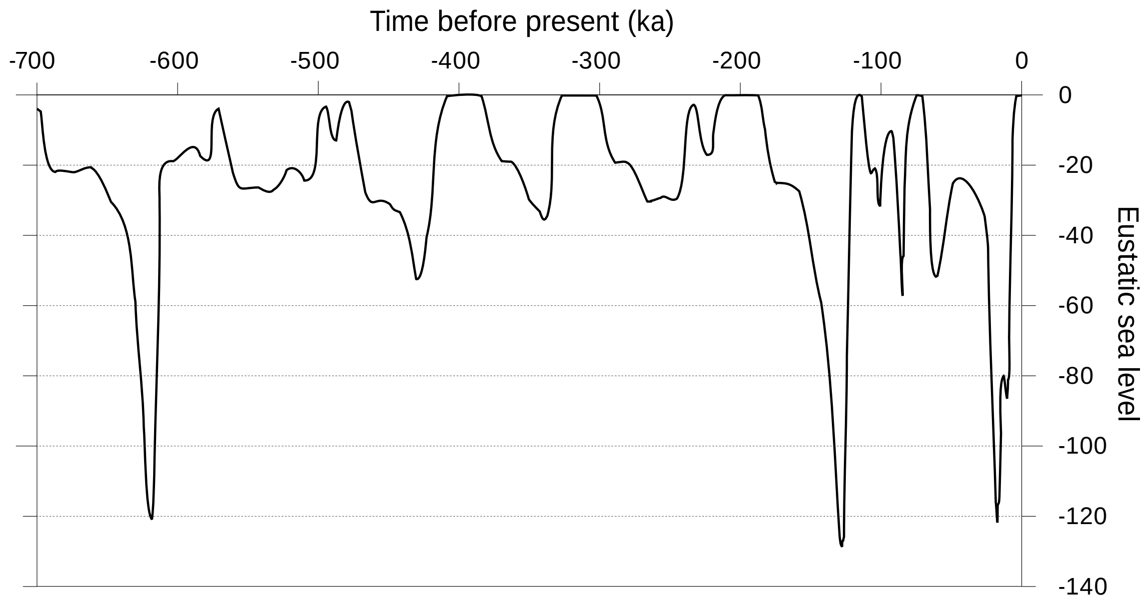


Figure 4-1. *Hindcast eustatic sea level curve for the last 700,000 years, assuming that global ice sheets vary in parallel with the European ice sheet.*

In regions, such as Sweden, where large ice sheets have grown, depression of the earth's lithosphere beneath the ice sheet load has been generally greater than eustatic sea level lowering. As a consequence, sea level relative to a fixed point on the local lithosphere surface (relative sea level) has risen during glaciation. Figure 4-2 is a schematic representation of relative sea level through a glacial cycle /Boulton 1990/.

Far from the ice sheet, the lithosphere warping component is minimal and sea level predominantly reflects global eustatic changes. Ice proximal areas that are glaciated first and deglaciated last show the largest isostatic component. At the beginning of a glacial cycle coastal sites suffer sea level lowering due to global ice growth (probably in North America or in polar areas, whose susceptibility to glaciation is greater). As the European ice sheet begins to grow however, an isostatic effect is initiated. The rate of relative sea level lowering is slowed, and ultimately, as lithosphere depression begins to exceed eustatic sea level lowering, relative sea level starts to rise (Figure 4-3).

Figure 4-4 shows a number of late-glacial and post-glacial relative sea level curves from Sweden which match the deglacial components of different idealised curves from Figure 4-2. Given the general similarity of successive modelled areal patterns of ice sheet growth and decay /Boulton et al 1992/, we suggest that the growth-phase component of the idealised curves can be assumed for actual sites, where only the deglacial sea level record is available.

The recent relative uplift as recorded by precision levellings and tide gauge data is shown in figure 4-5. This figure indicates that the southernmost part of Sweden is sinking. This submergence has been interpreted to be a result of eustatic rise, which is in the order of 1 mm/year /Påsse 1996/.

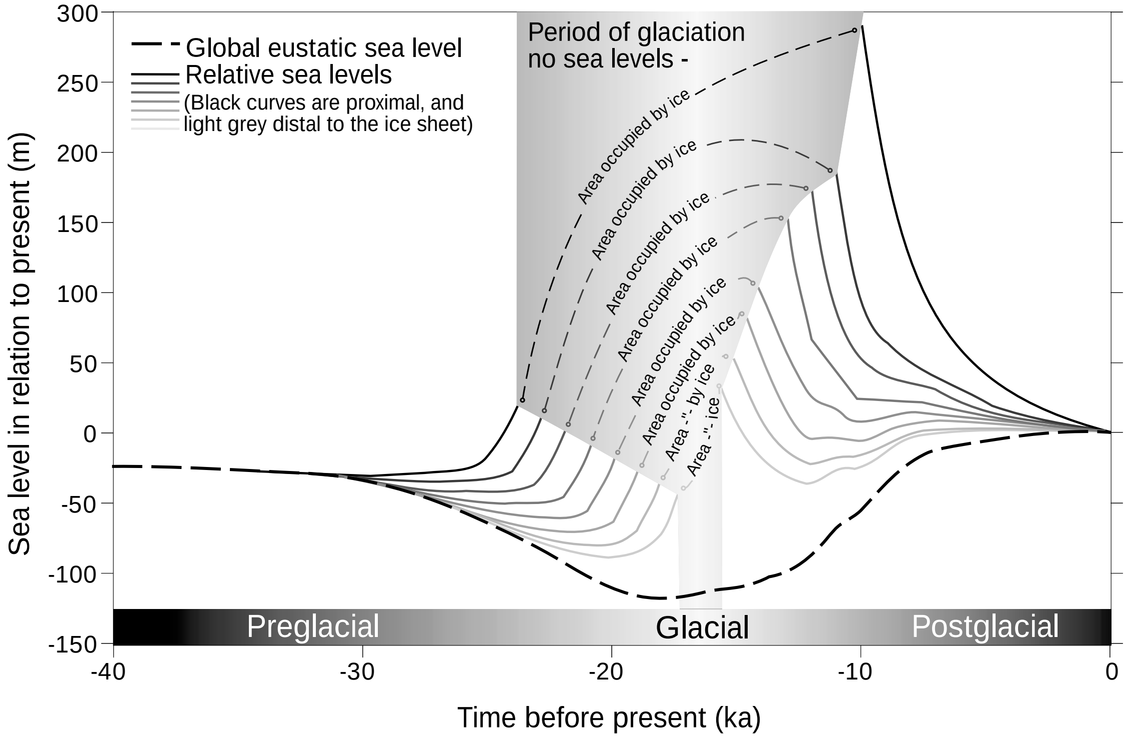


__Figure 4-2.__ The generalised relative sea level change in the vicinity of the Scandinavian ice sheet before and after the last glacial maximum. The grey area, where the curves are linked together by dashed lines, represents the period of glaciation, when the area is occupied by ice.

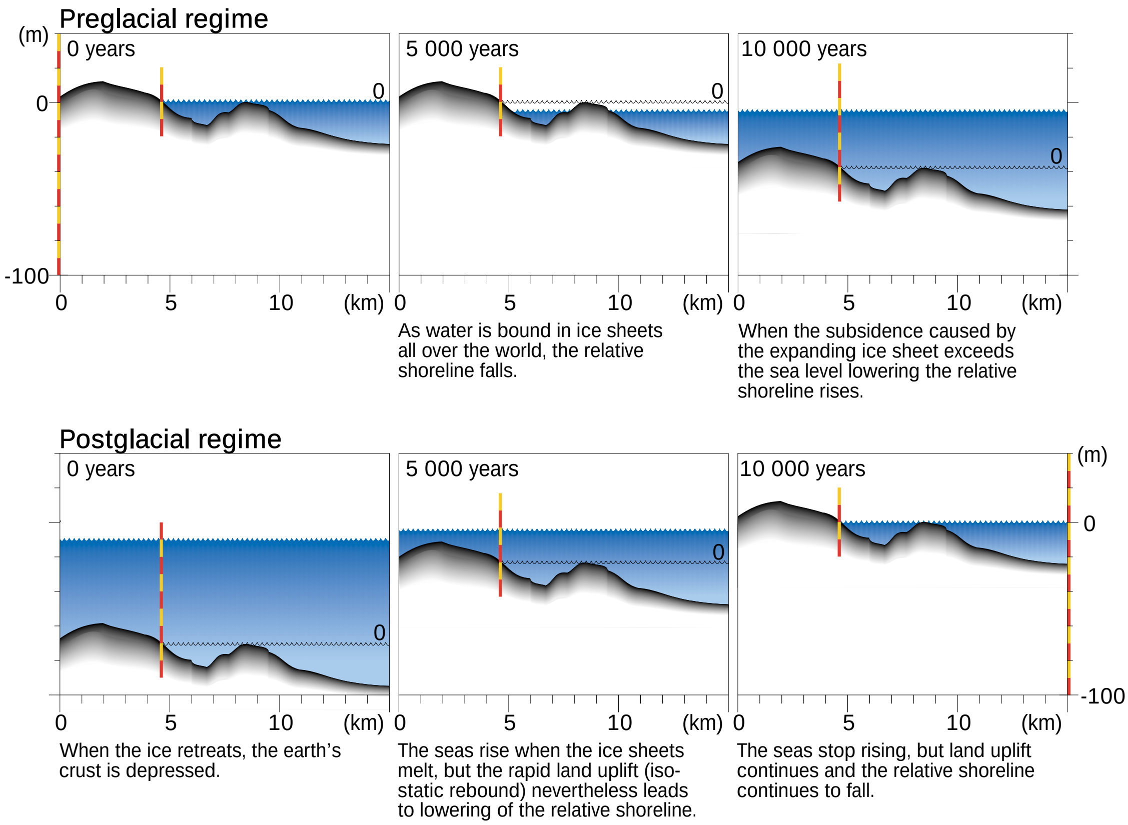


Figure 4-3. *Sketch showing the evolution of relative sea level change at a hypothetical coastal site during a) the preglacial regime and b) postglacial regime.*

For coastal areas with low surface slopes, small changes in relative sea level can produce very large changes in the position of the coastline. By use of a mathematical model of shore level displacement in Fennoscandia /Påsse 1996, Påsse 1997/, the future development can be forecasted.

4.1.1 Biosphere conditions

The natural vegetation during interglacials is dominated by forests. Deciduous species dominate in the warmer, temperate zones in the south and coniferous forest in northern areas (boreal vegetation zone). The forest type affects the soil type. In deciduous forest, cambisols (brown soils rich in mull) predominate, whilst in coniferous forests, podzols and lithosols dominate. Soil and forest type affect understorey vegetation as well as groundwater chemistry and the potential transport of radionuclides. Herbs and grasses occur in open areas and are more affected by the chemical composition and groundwater level. However, the extension of forests and different soil types can be slow. Forest requires long time periods (80 to 100 years) to establish and soil development is even slower. Thus, it can take several centuries before a change in climate results in establishment of forest with a characteristic soil profile.

The current moist climate allows rapid establishment of vegetation cover, which prevents soil erosion except in unstable dune areas. The climate encourages high productivity due to good supply of water and low temperatures, which restrict respiration.

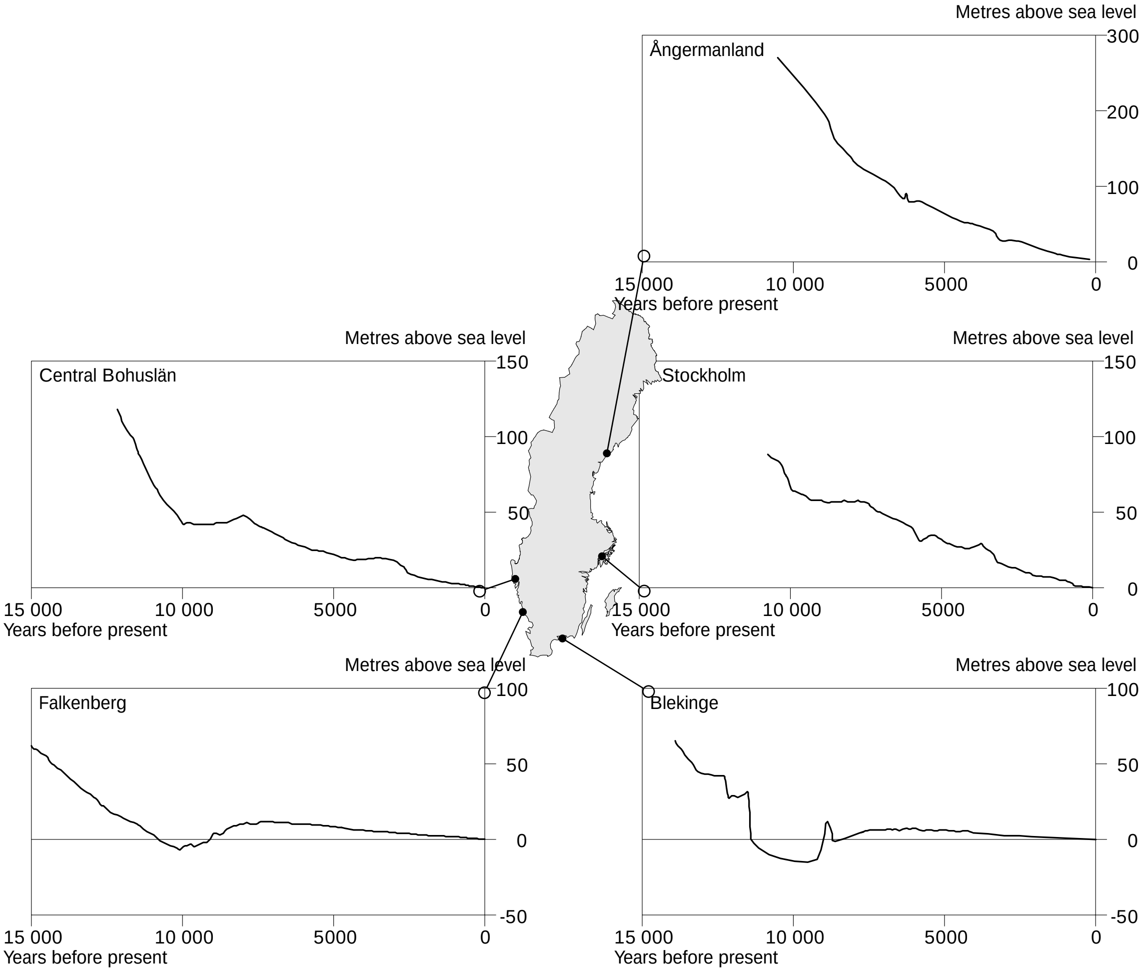


Figure 4-4. *A series of typical relative sea level curves for the deglacial phase from sites in Sweden /revised from Fredén 1994, shore level curves from Cato 1992, Påsse 1988, Hedenström et al 1999, Miller et al 1988, Björck 1979/.*

Today, wetlands constitute 21% of the total land in Sweden and are abundant where net precipitation is high. Peat bogs are abundant in areas where the rate of accumulation of organic matter is larger than the temperature dependent rate of decomposition. Fens are abundant in many areas and are mainly dependent on a flat topography.

The rich vegetation and the water-rich environment favour an abundant insect fauna during the summer half-year. During winter it is too cold for insects, whose offspring winter as eggs or pupae. High insect production as well as the production of berries and seeds is an important food resource for birds. Due to the high seasonality of these resources many birds migrate to warmer places during winter and forage there. The climate and the forest allow a high production of larger fauna such as moose, deer and rodents. Some hibernate during winter or spend their time under the snow cover.

The moist climate allows good conditions for a productive agriculture, especially in the mull rich southern parts of Sweden. However, the short seasons usually allow only one crop per year. Thus a new growing season starts in late autumn or spring and is harvested in summer or autumn. The climate is especially good for pasture, cattle and sheep farming. Generally the open sea in the southern parts permit year round fisheries. In the northern area however large-scale fishery is seasonally restricted.

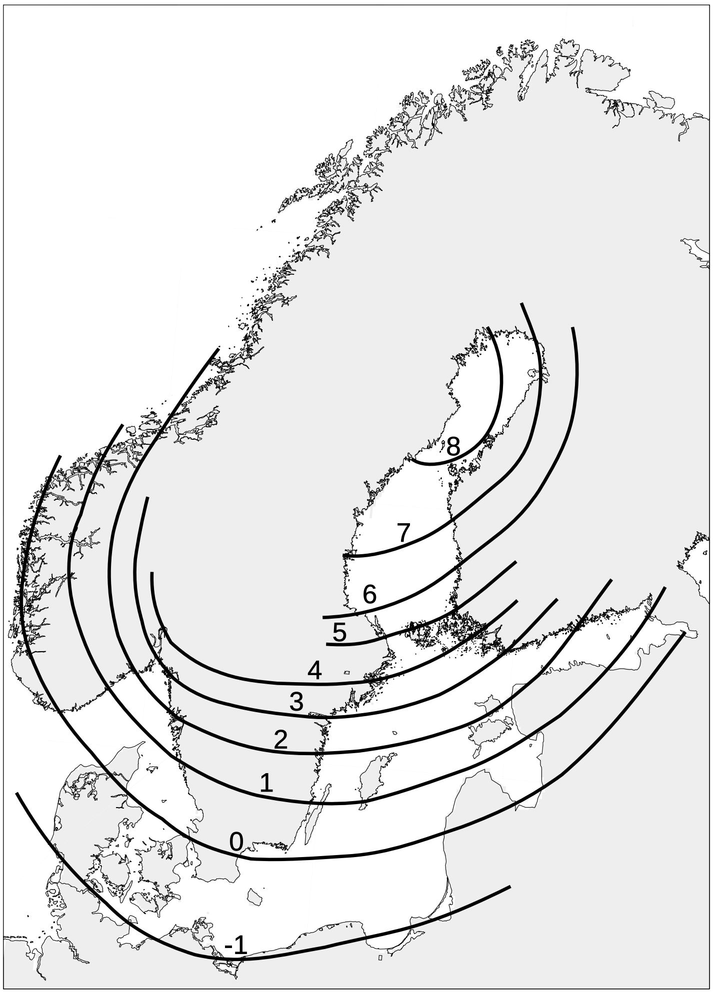


Figure 4-5. *The recent relative uplift (mm/year), recorded by precision levellings and tide gauge data /Påsse 1996 after Ekman 1996/.*

The human population is not directly limited by the climate in the temperate/boreal domain. Other factors such as soil composition, lakes and coastal areas have a greater effect on population.

The productivity of fishes and algae is influenced by water mixing which controls nutrient supply and the stability for phyto-plankton. Thus there are periods when plankton is restricted. This favours benthonic plants. The relatively cool water-temperature also favours the growth of kelp-like algae in marine environments. This in turn gives a good feeding and nursery area for fish and, in sufficiently saline areas, for lobsters and crabs.

During periods of sea level change there are rapid changes in the biosphere at coastal sites, especially if the topography is flat. At present, transgression /regression is small (in the order of mm/year) in relation to weather-induced sea level fluctuations (£1.5 m in the Baltic Sea). The biosphere close to the coast, and the pattern of colonisation of aquatic / terrestrial species in the biosphere, are more influenced by the seasonal changes than by the slow transgression / regression.

During periods of rapid transgression, aquatic species will colonise earlier terrestrial habitats. Aquatic species have short generation times and rapid dispersal in water and they are efficient in colonising the new sea areas. During periods of rapid regression, the aquatic system will become terrestrial. The colonisation of annual plants will be rapid, but forest and organic soil profile development will be slow.

The shoreline is subject to considerable erosion by waves that rapidly disperse finer sediments. It will be particularly strong during changes of sea level. Erosion of sediment can release accumulated radionuclides.

Shore-level displacement can drain straits and turn bays into lakes and vice versa. This gives rise to large changes in water turnover as well as rapid changes in salinity, which affects the structure of ecosystems being dependent on the salinity.

4.1.2 Thermal conditions

Natural variations in mean annual temperature within the temperate / boreal domain are on the order of a few degrees. During the Holocene maximum, the mean annual temperature was about 2 °C higher than today. Potential global warming, anthropogenically caused by the greenhouse gases, is believed to be on the same order of magnitude.

4.1.3 Hydrological conditions

In the temperate /boreal domain, regional groundwater flow is predominantly topographically driven but also to various degree affected by fluid density variations. Groundwater recharge is dependent on precipitation, evapotranspiration and surface run-off.

The effects of minor changes in temperature and precipitation on groundwater recharge and surface run-off have been investigated by modelling /Losjö et al 1999/. It is concluded that relatively large changes in groundwater recharge and surface run-off, especially the seasonality, can be expected also for changes in temperature of the order of ±2 °C and for decreases in precipitation of 20%. In terms of the total, yearly, deep groundwater recharge, the result of a decrease in temperature of 1°C is hardly noticeable, but an increase during summer and a decrease during winter are observed.

Relative sea level changes represent important changes in the boundary condition for groundwater flow in a coastal region (see e g /Voss et al 1993/). A rising sea level, which inundates a low-lying coastal region, will extend the region in which dense saline water may displace fresh groundwater. A falling sea level may cause flushing out and replacement of saline groundwater by freshwater.

Changes in net precipitation as well as the on-going, slow land-rise in coastal areas may cause groundwater divides to move and groundwater recharge areas turn to discharge areas and vice versa. Sea level will also affect the salinity of the Baltic Sea due to the shallow sills delimiting the Baltic Sea (–18m). Thus, the Baltic Sea is a freshwater lake when the sea-level is 18m lower than present. The evolution of the groundwater system at Äspö during the last 10000 years has been studied by means of numerical modelling /Svensson 1999a/. This work has included variations in sea water salinity, relative sea level change and freshwater recharge. Figure 4-6 illustrates the hydraulic evolution in a coastal location.

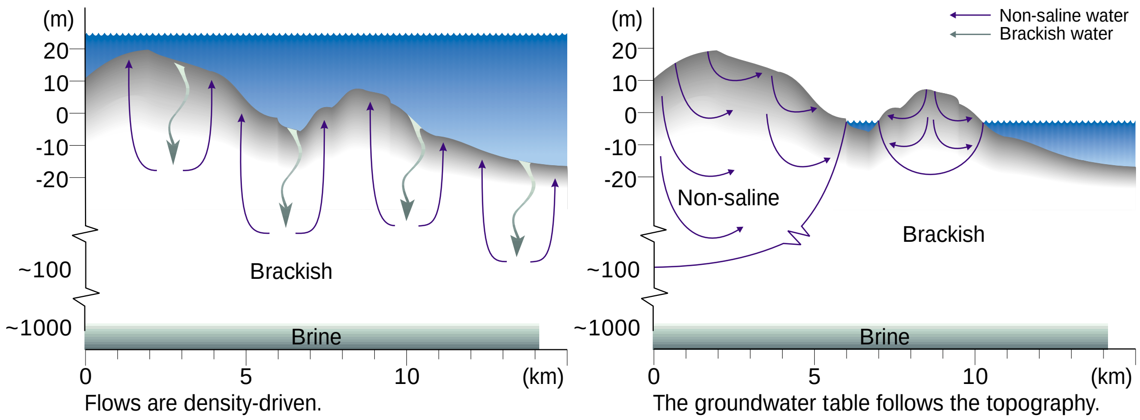


***Figure 4-6.** Schematic figure of water exchange in a coastal location in conjunction with regression.*

4.1.4 Mechanical conditions

Current seismic activity in Sweden appears to be related to a combination of post-glacial rebound and plate movements ("ridge push") /Muir Wood 1993/. The relative importance of each of these mechanisms is debated. Displacements associated with single earthquakes of different magnitude have been investigated in literature studies and model simulations /LaPointe et al 1997/. Earthquakes occurring in the temperate / boreal domain are likely to be of small magnitude. Directly following deglaciation earthquakes of greater magnitude can be expected (see glacial domain).

4.1.5 Chemical conditions

The distinctive feature of periods when land is covered by sea is recharge of heavy, saline marine water into the system. The location and orientation of fractures will play an important role in determine the trajectories of saline water in displacing freshwater masses. During the later stages of a deglaciation, damming of a freshwater Baltic Sea is possible, such that infiltrated water becomes more dilute with a lower density. The consequences of events within this domain may be found as characteristic palaeohydrochemical signatures of groundwater.

The chemical composition of groundwater is changed due to changes in the recharge of meteoric water. The modelling carried out for the Äspö region /Svensson 1999a/ (see section 4.1.3) has simulated the evolution of groundwater composition with respect to origin of the water (meteoric, fresh lake water, brackish or salt sea water or a very old deep brine). Results of this initial study were not in total agreement with the results of water chemical analyses. However, modelling neither accounted for chemical reactions nor mixing of the water types.

4.1.6 Identified regimes

Critical attributes of the temperate/boreal domain are:

- temperature
- precipitation
- direction of sea level change

- rate of sea level change
- salinity of sea/lake water

Three different regimes with different combinations of boundary conditions for groundwater flow and rock mass behaviour have been identified. They are the interglacial regime, the preglacial regime and the postglacial regime. The postglacial regime is subdivided into two sub-regimes, depending on whether the location is above or below the marine limit. In the following sections, the subsurface impact is described individually for each of the regimes.

4.1.7 Interglacial regime - impact on the geological barrier

Thermal impact

Differences in air temperature on the order of a few degrees, as described in section 4.1.2, affect the depth of ground frost to some extent, but do not have any significant impact on the thermal conditions and processes of the geological barrier.

Hydrological impact

Most of the climate-related processes within this regime are restricted to a zone close to the ground surface. Changes in precipitation and air temperature affect groundwater recharge /Losjö et al 1999/. The on-going slow land rise in coastal areas changes the locations and extent of recharge and discharge areas and hence the transport times and pathways.

The climate-related changes at repository depth are expected to be small and can thus be included in estimates of general uncertainty.

Mechanical impact

The effects of the minor earthquakes that are expected during this regime are very small at repository depths.

Chemical impact

The chemical composition of groundwater is changed due to changes in the recharge of meteoric water, which is dependent on precipitation, vegetation cover and temperature but also on groundwater level. However, at repository depths the hydrochemical conditions are believed to be unaffected.

4.1.8 Preglacial regime - impact on the geological barrier

This regime is concerned with the period of ice build-up during which there is lowering of sea level from near modern values The early part of sea level evolution follows the global eustatic curve and the later part includes, in Sweden, an increasing isostatic component. Modal curves shown in Figure 4-3 are considered to represent the range of rates of sea level change represented in Sweden under conditions in which the all or most of the country is ultimately glaciated.

Hydrological and chemical impact

Conditions in this regime are predominantly influenced by sea level fall. As the eustatic lowering of the oceans takes place, localities previously covered by water will rise above the sea level. This means that marine water recharge is replaced by meteoric recharge in coastal regions.

However, in areas proximal to the ice sheet, isostatic depression caused by a growing ice sheet will be larger than the eustatic fall. This will result in a rising relative sea level and marine water recharge.

4.1.9 Postglacial regime - impact on the geological barrier

In this regime there is a fall in relative sea level from an elevation above modern sea level, as a consequence of the decay of the glacio-isostatic component of shore displacement, or a fall and rise from a level below modern during deglaciation. It is represented by the characteristic curves shown in Figure 4-3. Geographic position is important for the rate of sea level change (Figure 4-2, 4-3) and hence the hydrological impact. Zones near to the ice sheet's maximum extent but which were never covered by ice, are still effected by isostatic depression. As a result, sea level lowering is less than at sites which reflect the full global eustatic component. As the ice maximum position is approached, there may be a relative sea level regression during the early part of deglaciation. In those areas, which were covered by the ice sheet, zones that have suffered least isostatic depression are exposed first and have the lowest marine limit.

Hydrological and chemical impact

Sea level change produces the major impact. The conditions at a specific locality are changed with time due to interacting effects of the large amount of meltwater originating from the ice sheet, the isostatic land uplift and the global eustatic rise. Sites below the marine limit suffer a change from glacial meltwater recharge to marine water recharge which may rapidly affect the conditions at great depth. On the west coast, the salinity can be expected to be fairly constant, whereas the east coast is subject to large variations (fresh-brackish-saline) /Andersson 1998/.

The subregime *Above the marine limit* is not affected by the different stages of the Baltic Sea. However, there is still an isostatic land uplift which may be taken into account.

4.2 Permafrost domain

This domain comprises areas of permafrost, which do not extend unbroken up to the ice sheet margin. This distinguishes it from permafrost which does continue unbroken to the ice sheet margin where sub-permafrost groundwater flow is glacially driven.

Permafrost in Sweden is unlikely to be sufficiently thick that it is insensitive to the thermal effects of large surface water bodies. It will not therefore extend for any significant distance beneath the sea and is assumed to be broken by any large lake (> 0.5 km width) or a major river. Two different permafrost zones are normally distinguished (figure 4-7):

- **continuous permafrost;**
  where, apart from beneath large lakes, rivers and arms of the sea, the permafrost is unbroken;
- **discontinuous permafrost;**
  consists of isolated permafrost masses rather than a continuous sheet. It occurs where conditions for permafrost would be generally satisfied if it were not for the warming effect of surface water bodies and slope segments of warm aspect.

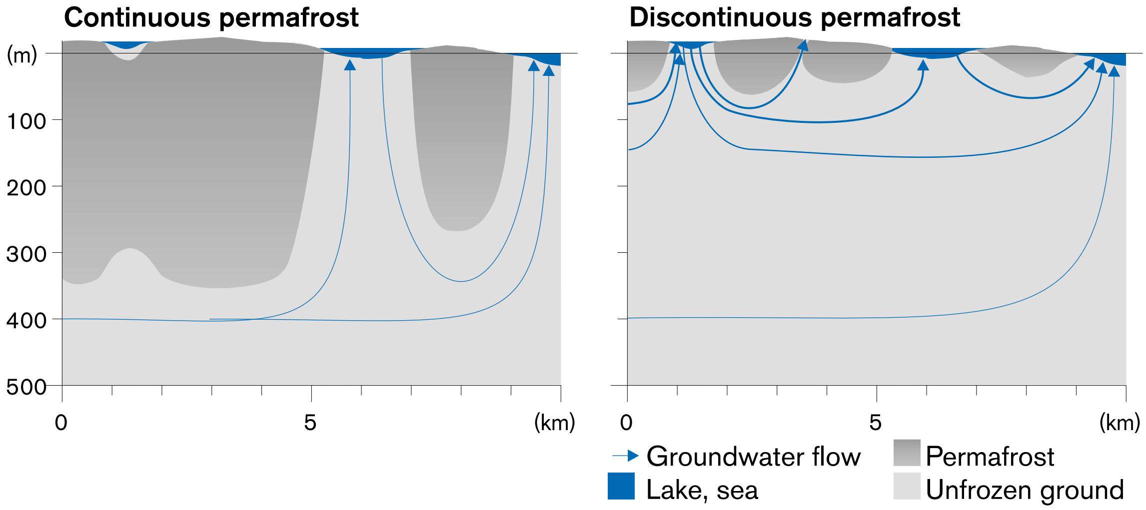


Figure 4-7. *Illustration of groundwater flow conditions in continuous and discontinuous periglacial permafrost.*

In general, continuous permafrost gives way to discontinuous permafrost where the mean annual ground temperature just below the level of no seasonal variation in temperature is –5 °C. The permafrost thickness in this area is estimated to be about 50m. Sporadic permafrost occurs where conditions for permafrost are not generally satisfied, but where it is sufficiently cold to develop permafrost patches at sites of cold aspect.

Although permafrost can have significant permeability /van Everdingen 1987/, particularly where it is fractured, the permafrost body itself is assumed to be impermeable.

There is no direct geological evidence of past permafrost depth in Sweden. When estimating permafrost depth and spatial extension we have to rely on the results of modelling to a great extent.

The climate within the permafrost domain is characterised by low temperature and low precipitation, due to low evaporation transporting water to the atmosphere. The absolute amounts of precipitation are difficult to assess, but values less than 50% of current ones seem reasonable /e g French 1996/.

4.2.1 Biosphere conditions

The permafrost domain is characterised by the tundra ecosystem. The low evaporation encourages accumulation of surface water, which is unable to seep into the ground because of underlying permafrost. During the summer, extensive areas are waterlogged, but the amount of peat formed is negligible because plant productivity is low. The tundra is devoid of forests and the vegetation mainly consists of herbs, shrubs and bushes. In raised dryer sites, lichens can be found, whilst on wet ground mosses dominate. The growing season is short and several species have adapted to flower and set buds in different years. Most plants develop thick roots that serve as storage. No plants produce berries for the dispersal of seeds.

The major part of the vertebrate fauna of the tundra migrates south during winter. Birds, which are abundant during summer, migrate long distance to subtropical areas. During the arctic summer they thrive on mosquitoes. Lemmings spend most of their life beneath the isolating snow-cover. Reindeers utilise lichens in snow-free areas in winter. Otherwise they migrate to forested areas.

Periglacial environments are characterised by intense wind action due to the lack of significant vegetation. Even if there is a snow-cover of 50 cm during winter, raised parts are blown free of snow. Sediments are transported and redeposited as dunes.

Taliks, i.e. unfrozen windows in the permafrost region under lakes or rivers, are potential sites where animals and humans can settle.

Radionuclides entering the biosphere in the permafrost domain will probably not be accumulated for long periods due to the constant disturbance of the soil. Even if they are trapped in the frost, the large amounts of water available when thawing takes place will ensure that the release is strongly diluted.

4.2.2 Thermal conditions

Permafrost is traditionally defined, on the basis of temperature, as ground that remains at or below 0 °C for at least two consecutive years. A balance between internal heat gain with depth and heat loss from the surface determines the depth to which permafrost develops.

The depth of permafrost is hence dependent on the mean annual air temperature, the thermal conductivity of the ground and the thickness of a potential snow cover. Based on modelling, the maximum depth of periglacial permafrost in Sweden during the last glaciation has been estimated to be about 300 m /Boulton et al 1992/.

Also at temperatures above the freezing point, the physical properties of the water, such as viscosity and density, will be altered, but not very significantly.

4.2.3 Hydrological conditions

The presence of permafrost and a seasonally frozen active layer restricts the infiltration of water and its recharge to groundwater systems. A very large percentage of precipitation and snowmelt is therefore likely to contribute to surface runoff. Seasonal maximum streamflow rates can be expected to be very large.

The principle impact of permafrost on a specific geological setting is to change the permeability distribution by freezing, blocking fractures and thereby changing the hydraulic geometry of the rock mass. Through sealing extensive areas of surface, permafrost will change the locations of recharge and discharge. Its impact will be to drive groundwater flow to deeper levels and to increase potential gradients where flow is significantly blocked (see figure 4-7).

4.2.4 Mechanical conditions

In the active layer, alternate freezing and thawing in soil and rock produces frost cracking, cryogenic weathering, upfreezing of stones and the creation of tundra polygons. Solifluction, i e soil movement, takes place even on the gentlest of slopes.

There are many processes disturbing the soil and also exposing it to erosion. Radionuclides that may have been accumulated in the soil during previous warmer periods will thereby be released. The released nuclides may be diluted or cause a potentially high dose.

4.2.5 Chemical conditions

The low ground temperatures associated with the permafrost will reduce reaction and dissolution rates. Because of the increased solubility of carbon dioxide at low temperature, the solubility of calcite will be increased.

The freezing process itself as well as the longer residence times for groundwater below an impermeable permafrost cap may alter the chemical composition of groundwater and result in higher salinities. Higher fluid density can produce vertical transport of saline fluid to greater depths.

In coastal areas residual saline groundwater may play a role in the dynamics of permafrost evolution, by affecting the conditions for freezing.

4.2.6 Identified regimes

The critical attributes of the permafrost domain are:

- temperature
- permafrost thickness
- permeability
- occurrence of holes and gaps in the permafrost

We identify the following two permafrost regimes for the cases where permafrost does not extend unbroken as far as the ice sheet margin;

- continuous permafrost regime
- discontinuous permafrost regime

The major differences between these two regimes are related to the permafrost thickness and permeability. The permafrost is considered to be continuous when its thickness is greater than 50m. In the discontinuous regime permafrost thickness is less than 50m. The impact on the geological barrier in these two regimes is discussed below. Since impacts in the regimes are similar, they are not described separately, only major differences are identified.

4.2.7 Impact on the geological barrier

Thermal impact

The development of periglacial permafrost is not likely to have a direct thermal impact on the repository. Calculations show that repository depths will not be affected by freezing, although temperatures will be lowered. The heat generated by the waste in the repository will theoretically act to reduce the thickness of the permafrost. At the time of the presumed next major permafrost development in Sweden, the generated temperature at the repository depth has declined significantly from its maximum values /e g Claesson et al 1996/ and the local reduction in permafrost thickness is expected to be rather small /Vallander et al 1991/.

Hydrological impact

In general, groundwater flow will be forced to greater depths than during temperate/boreal conditions. The actual depths of flow paths are governed by the thickness of the permafrost cover. The presence of continuous permafrost will also result in extended groundwater flow paths and increased transit times (see figure 4-7). Recharge and discharge are likely to be concentrated at fewer and smaller areas than under unfrozen conditions. In the continuous regime, water will be driven to greater depth and the transit times will be longer than in the discontinuous regime.

Within the permafrost, layers or bodies of unfrozen ground can occur due to local anomalies in thermal, hydrological or hydrochemical conditions. Such unfrozen parts are called taliks.

Taliks in discontinuous permafrost provide locations for recharge or discharge. A discharge point for sub-permafrost groundwaters can be associated with very large potential gradients. The flow between a location of recharge and a location of discharge, is determined by the infiltration rate, the water-conducting ability of the sub-surface and the hydraulic gradient. Groundwater flow towards holes in the permafrost can produce locally high permeabilities by creating quicksands and hydrofractures (e.g. /Boulton et al 1995a/. Open taliks provide paths through which release of radionuclides can take place /Boulton et al 1995a, McEwen et al 1991/.

Mechanical impact

The development of periglacial permafrost will not have any direct mechanical impact at the repository depth. Shallow erosion of sediments on the surface will take place. Freezing of water in fractures results in a volume expansion and may under confined conditions extend and broaden fractures. This effect can be expected to be partially permanent, since the deformation is not likely to be fully elastic.

Chemical impact

Through the freezing process itself and longer residence times for groundwater, we can expect changes in hydrochemistry, especially higher salinities. Gas hydrates (e.g. methane), frozen as clathrates, form under conditions where gas and water exist under high pressure and low temperature. As permafrost degrades, these gases will be released to the atmosphere. However, since the permafrost is not expected to reach repository depths, potential mechanical impact of clathrates is not considered important.

There is no clear hydrochemical evidence in bedrock for the past existence of permafrost in Sweden.

4.3 Glacial domain

Extensive simulation modelling has been carried out of the evolution of the Scandinavian ice sheet through the last glacial cycle /e g Boulton et al 1992/. These simulations show a well-defined thermal regime. The ice/bed interface is frozen in the ice divide region, melts in a broad zone beyond that, and is frozen again in the terminal zone of a few tens of kilometres (see figure 4-8). During advance phases in particular, proglacial permafrost represents an extension of the terminal frozen bed zone.

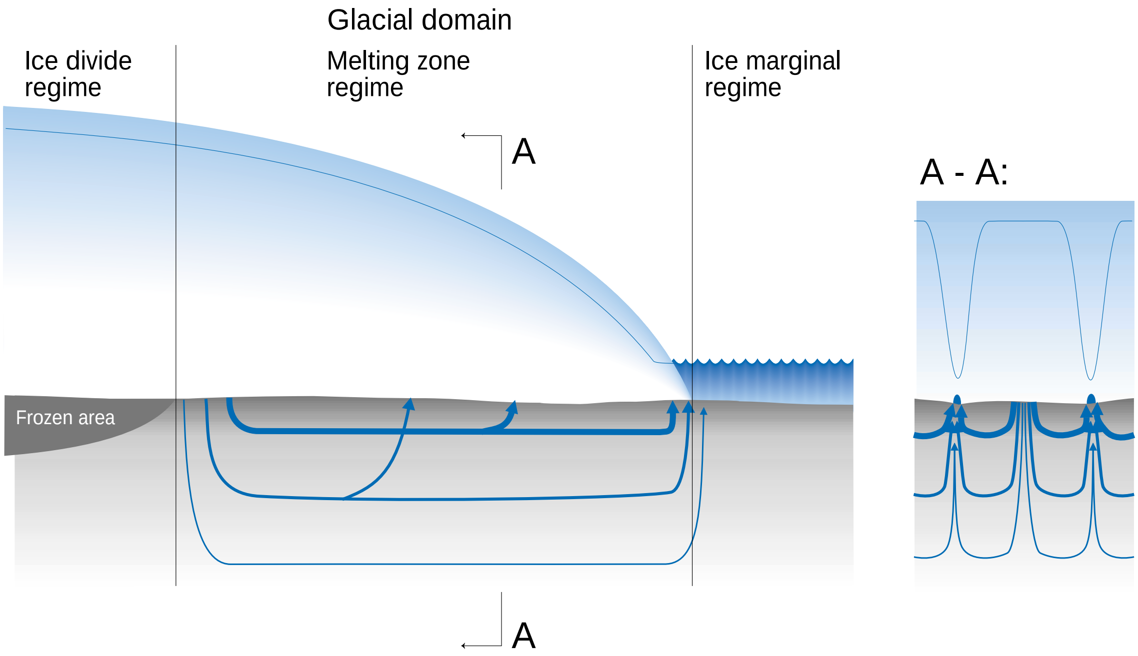


Figure 4-8. Sketch showing the glacial domain. The boundary conditions for groundwater flow are determined by the presence of the ice. Prevailing regimes and subregimes determine the more exact conditions.

During the glacial periods ice sheets extending over Sweden originate in the mountain zone along the Swedish-Norwegian border. As they grow, the ice divide tends to migrate in an easterly direction, but does not move further east than western central Sweden (see figure 4-9) or as far as the east coast in northern Sweden /Lundquist 1986/. The expansions of the sub-divide frozen bed zone have been simulated through the last glacial cycle. Its expansion along a transect through southern Sweden is shown in Figure 4-10 (see also Appendix C).

In the following text, the general conditions and process couplings within the domain are discussed as divided into biosphere, thermal, hydrological, mechanical and chemical conditions.

4.3.1 Biosphere conditions

In the glacial domain the presence of the glacier ice will prevent biosphere/geosphere contact except on nunataks. There we can expect lichens or occasional herbs, however productivity will be low and due to the position there will be no contact with potentially contaminated water. There will be microbes and algae on the ice surface. Close to the ice margin one can find vertebrates, e g birds and mammals. At marine ice margins a productive aquatic community can exist which can sustain a fish population. Human populations may exist near to an ice margin, living on fish at marine margins. However, at such coastal locations the water turnover is likely to be rapid giving high dilution rates for any glacially transported radionuclides.

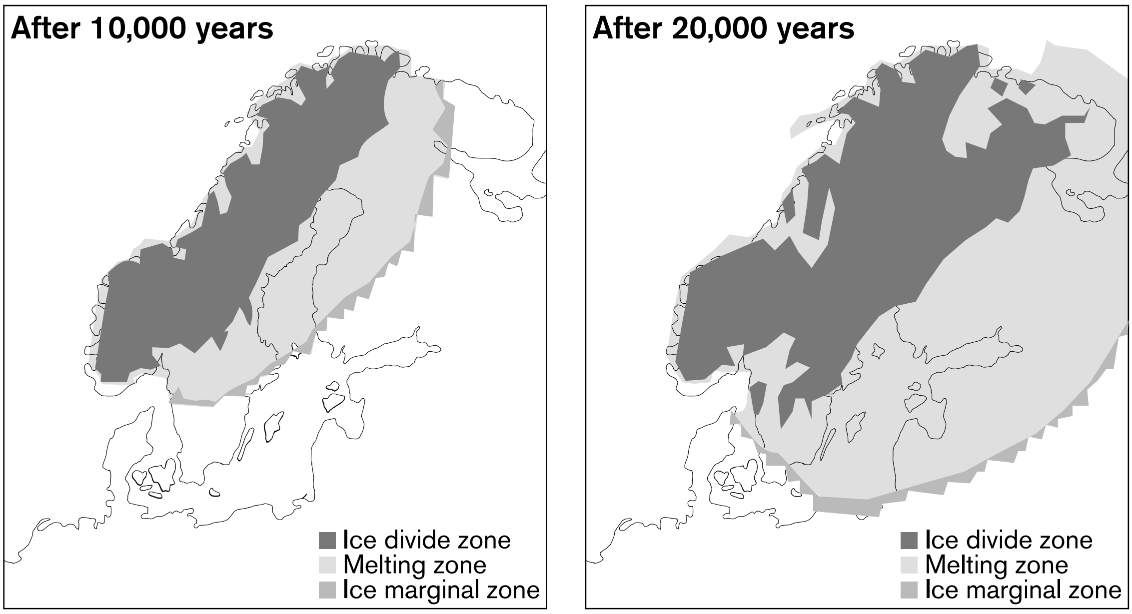


Figure 4-9. *Simulations of the areal growth of the European ice sheet showing the extent of different regimes (ice divide regime, melting zone regime and ice marginal regime) after 10,000 years and 20,000 years, respectively.*

4.3.2 Thermal conditions

The temperature within the ice sheet is a function of air temperature, geothermal heat flow, ice flow rate and accumulation of snow on the surface. When ice temperature exceeds the pressure melting point, determined by the ice thickness, then melting can occur. Under the ice divide cold ice advection leads to an unbroken sequence of permafrost development. Maximum depths of subglacial permafrost during the last glaciation were calculated to be about 400 m /Boulton et al 1992/.

4.3.3 Hydrological conditions

Two processes drive groundwater flow beneath ice sheets:

1. Consolidation of subglacial rocks due to the ice sheet load, resulting in the expulsion of water from the rocks and consequent groundwater flow. This will occur whether or not melting occurs at the ice sheet sole, but it is a transient feature. If a steady state ice sheet were resident for a long enough period of time, the underlying rocks would consolidate to equilibrium, and groundwater flow would cease.
2. Recharge to subglacial rocks by basal meltwater will generate a potential gradient and cause groundwater flow. Flow will continue as long as melting continues.

The recharge of groundwater is dependent on the interaction between meltwater fluxes, hydraulic heads (strongly governed by the surface elevation of the ice sheet) and the geometry of hydraulic flow paths at the ice/bed interface.

The hydraulic properties of the rock mass are affected by the stress situation. As described in section 4.3.4, large stress changes occur in the glacial domain, which means that changes in permeability anisotropy and magnitude are expected.

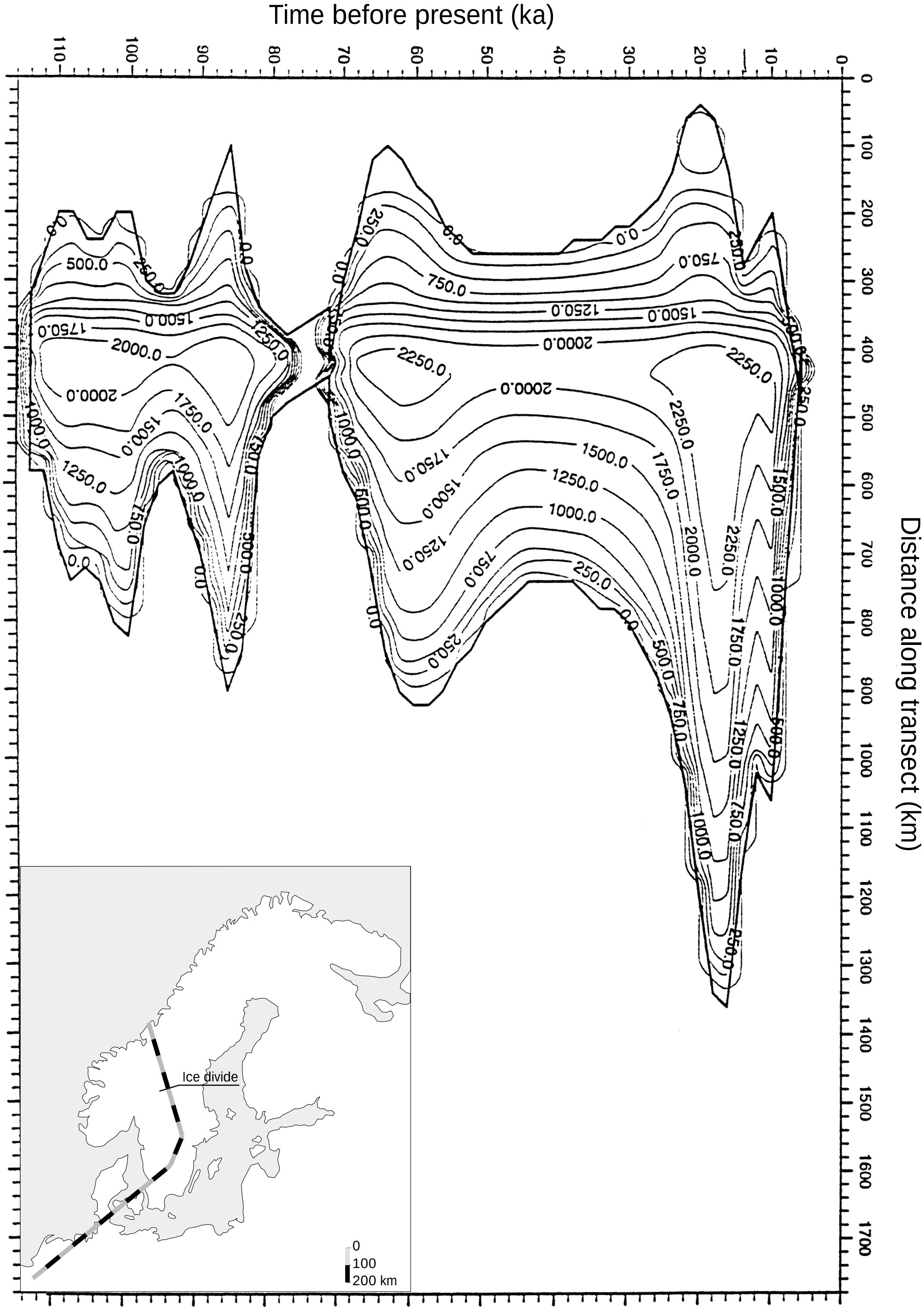


Figure 4-10. *Simulation of the behaviour of the European ice sheet during the last glacial cycle along the transect shown in the left corner. Ice sheet thickness in time and space.*

4.3.4 Mechanical conditions

An ice sheet imposes a load on the underlying bedrock, which leads to downwarping beneath the loading and upwarping (forebulge) on the margins of the depression (figure 4-11). The maximum crustal downwarping beneath the Weichselian ice sheet may have been on the order of 800 – 900 m. A theoretical analysis of displacements and flexural stresses induced by a load corresponding to an ice sheet has recently been presented /Rehbinder et al 1998/.

Both the total and effective in-situ stresses are affected by a glacier advancing over a site, through the interaction of load and fluid pressure. The total stress results from the summation of all gravity forces. However, the effective stress (i e the total stress reduced by the water pressure) will dictate the mechanical behaviour and properties of rock fractures, and hence, also their water-conducting ability.

We expect stress axes to rotate during the change from non-glacial to glacial phases. In a non-glacial phase, we expect the largest principal stress to be horizontal at shallow depth /Stephansson 1993/. As an advancing ice sheet overruns the site, vertical stresses will increase much more than the horizontal stresses. Thus we expect different mechanical response for differently oriented fractures (figure 4-12).

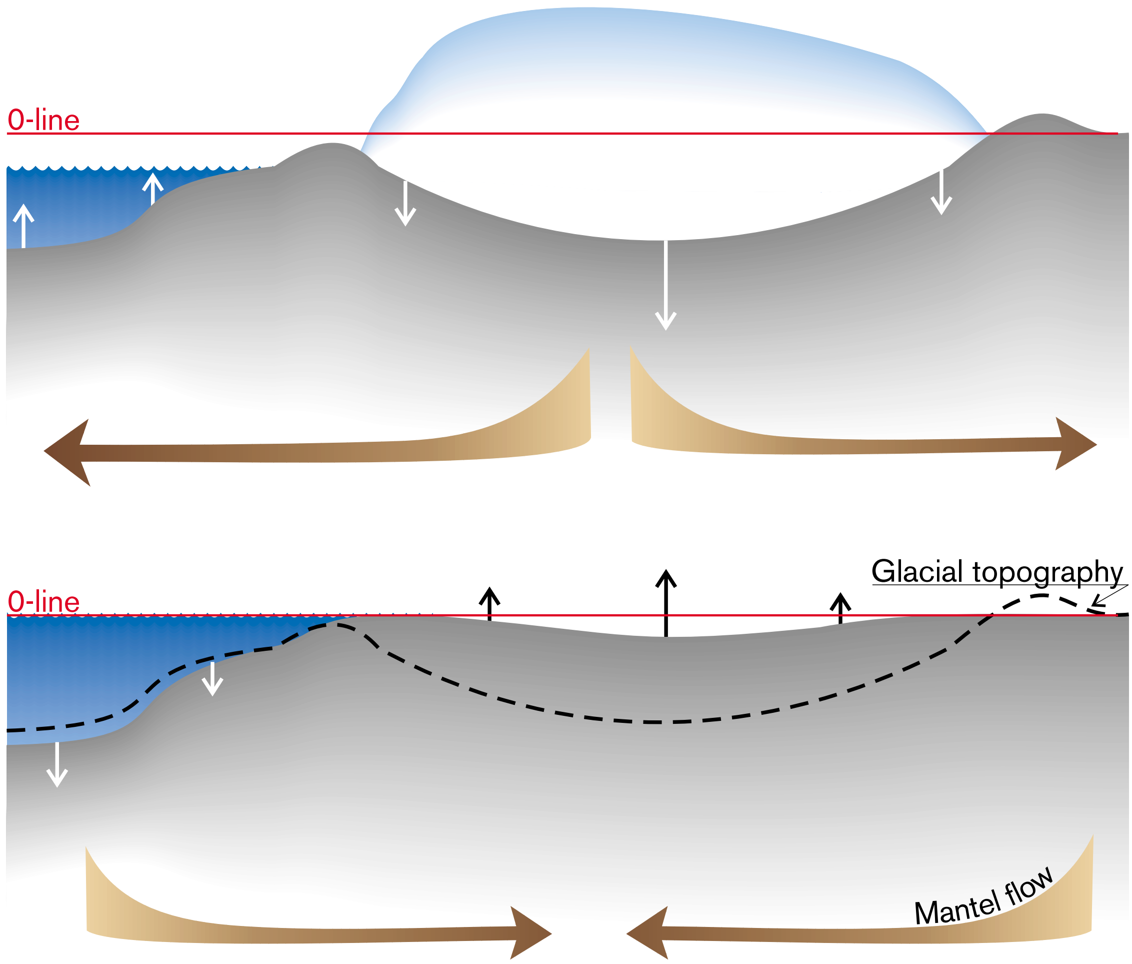


Figure 4-11. *Illustration of the mechanical response of the lithosphere and the viscous flow underneath the lithosphere under glacial conditions /from Morén et al 1999/.*

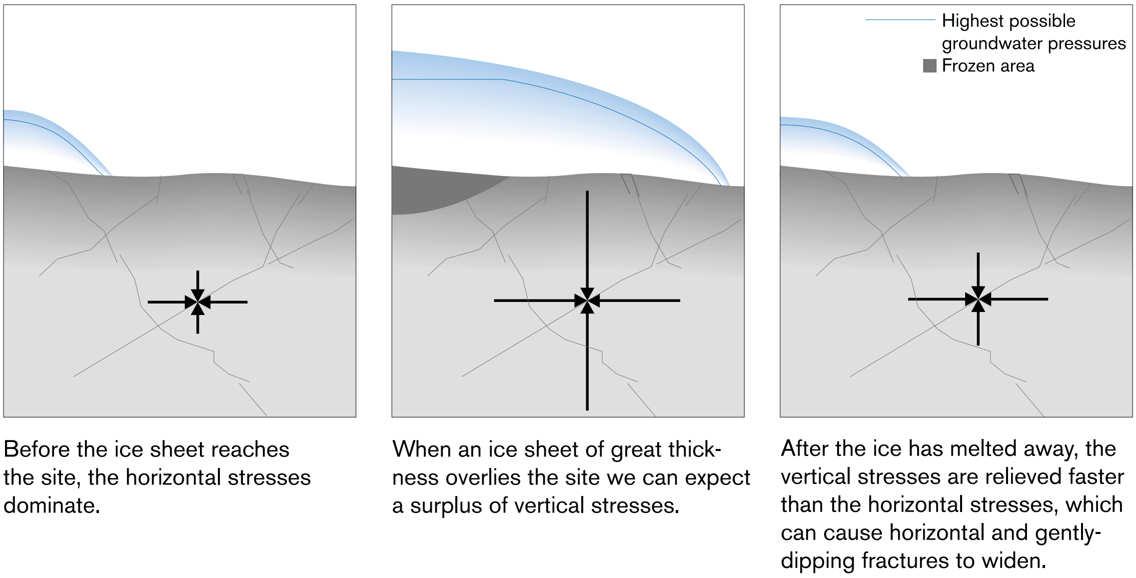


Figure 4-12. *Rock stress situations during a sequence of glacial advance and retreat.*

Stresses generated beneath ice sheets may produce fractures in the bedrock /Boulton et al 1995a, Boulton et al 1999/. It is anticipated that failure initially takes place along existing fractures /Rosengren et al 1990, Israelsson et al 1992/, but opening and propagation of new fractures by hydrofracturing may locally be possible. The fracture shear strength is also altered since it is a function of the effective normal stress acting across the fracture – a decreased effective stress results in a reduction in fracture shear strength.

During deglaciation we expect drastically decreased vertical stress and increased deviatoric stresses, which can lead to opening of gently dipping fractures and shear displacements /Leijon et al 1992/, see figure 4-12.

The changing orientation of the largest principal stress from horizontal under non-glacial conditions to vertical beneath the ice sheet can produce a change in permeability anisotropy and magnitude, depending upon the orientations of available fractures. In general, the increase in stress beneath the ice sheet will act to reduce the transmissivity of fractures and fracture zones. However, the increased deviatoric stresses can potentially induce shearing of critically oriented fractures which can lead to enhanced fracture apertures. Moreover, the same effect may result from hydraulic shearing and jacking due to locally high fluid pressures.

Rock mass permeability may be considered as time-dependent and self-organising /Boulton et al 1995b/. High groundwater pressures occur when low permeabilities mobilise potential gradients in groundwater up to the magnitude of the limiting potential gradient set by the ice sheet surface. Hydrofracturing or hydraulic jacking occurs when groundwater pressures exceed the value of the minimum stress. The tensile failures produced increase the permeability of the rock or sediment and thereby reduce potential gradients and groundwater pressures. These processes will therefore continue until increasing fracture permeability leads to a decrease in water pressure below the value of the minimum stress.

It can be expected that loading by the glacier will act to suppress seismic activity. On deglaciation, we would expect stored strain energy to be released to produce enhanced seismicity. A special study of deformation and seismicity during different stages of the Weichselian glaciation in southern Sweden /Muir Wood 1993/ concluded that rapid or differential unloading is the only process involved in the deglaciation that is likely to have significantly raised the seismicity levels and reactivated faults with critical orientations.

Post-glacial fault movements have been studied by SKB in the Lansjärv area in northern Sweden /Bäckblom et al 1989, Stanfors et al 1993/. One of the major conclusions of the investigations was that pre-existing structures have been reactivated as a result of a combination of plate tectonics and deglaciation.

Seismic activity and its consequences for the repository system is analysed separately in the performance assessment. A description of seismicity is therefore not included in this report although seismic events can be induced due to climate-driven environmental changes. A methodology to estimate the magnitude of shear displacements along fractures for earthquakes of different magnitude has been presented /LaPointe et al 1997/.

4.3.5 Chemical conditions

The chemistry of non-glacial recharge water is strongly influenced by its movement through soils, where organic material in a reduced state tends to strip dissolved oxygen from the water. The subglacial condition is quite different. Ice which has formed on the surface of an ice sheet and then flowed through the ice sheet to melt at its base contains numerous air bubbles, which are trapped within the ice. Meltwater from this ice, recharged into groundwater, is fresh and highly oxygenated. Dissolved oxygen is not stripped from the waters, as soil will generally have been removed by glacial erosion at the ice/bed interface. The chemistry of glacial waters is also characterised by distinctive isotopic compositions (e.g. ^{18}O and ^{14}C), dependent upon whether the water permeates through the ice or not.

We also expect that during glacier advance and retreat over a site, the pressure from the ice and the changed flow geometries result in changes in the vertical position of the saline / freshwater interface /Svensson 1999b/ (see figure 4-13).

As in the permafrost domain, the freezing process may increase the salinities of fluids in the subglacial permafrost. Methane hydrates (clathrates) may be formed due to freezing. The increased pressure as a result of the ice sheet load will ensure that this can occur at rather high temperatures.

4.3.6 Identified regimes

The critical attributes of the glacial domain are:

- temperature
- permafrost thickness
- permeability
- hydraulic head
- ice load
- meltwater flux
- recharge chemistry

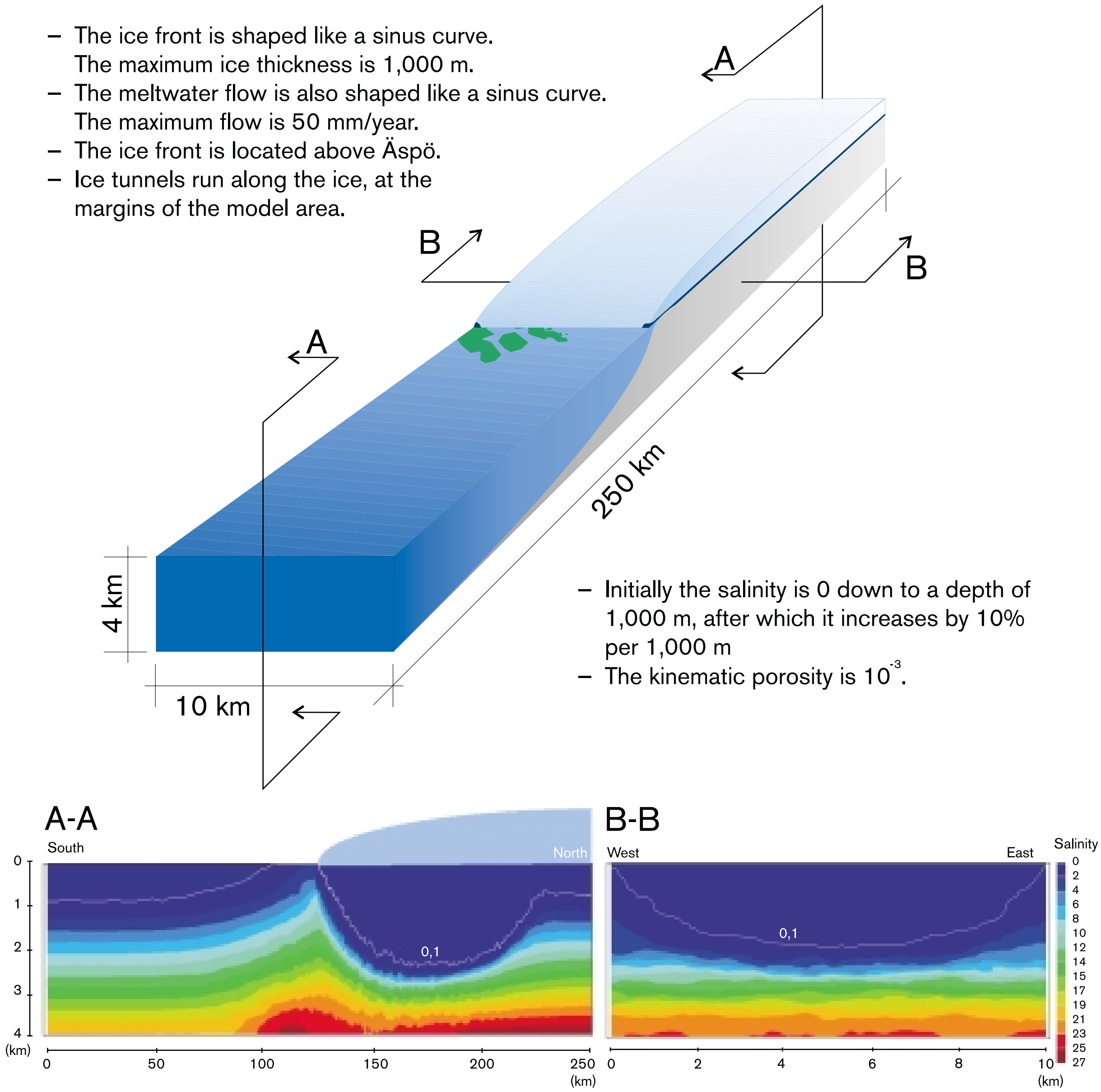


Figure 4-13. *Illustration of the changed vertical position of the saline /freshwater interface during glacial advance/retreat /Svensson 1999b/.*

Three different regimes having significantly different combinations of basic processes and boundary conditions for groundwater flow and rock mass behaviour have been identified (figure 4-8). They are the ice divide regime, the melting zone regime and the ice marginal regime. The attributes listed above diverge significantly in the different regimes. The subsurface impact is described for each of the regimes in the following sections.

The ice marginal regime is subdivided into three sub-regimes depending on whether the proglacial zone is above or below sea level (or a lake level) and, if above sea/lake level, whether permafrost is developed or not (see figure 4-14).

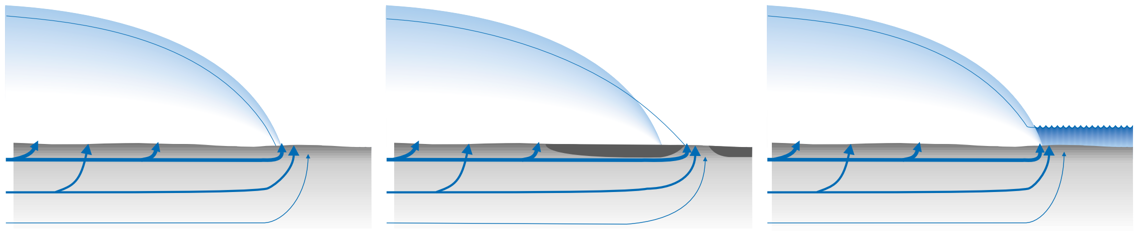


Figure 4-14. *Ice marginal regimes. The ice sheet terminates on land without proglacial permafrost. Left) The ice sheet terminates on land that has proglacial permafrost. Middle) There is no permafrost. Right) The ice sheet terminates in the sea.*

Ice sheet terminates on land with proglacial permafrost.

The proglacial zone is above the marine limit and there is proglacial permafrost. The zone extends 5 km beyond the outer edge of continuous permafrost.

Ice sheet terminates on land without proglacial permafrost.

The proglacial zone is above the marine limit and there is no proglacial permafrost immediately beyond the ice sheet. Water will flow very strongly upwards towards the free surface. Effective pressures in the terminal zone will be zero or near zero. The regime extends 5 km beyond the ice margin.

Ice sheet terminates in the sea.

The proglacial zone of the ice sheet lies below the marine limit and the retreating ice sheet terminates in water. The zone extends 5 km beyond the ice margin.

4.3.7 Ice divide regime - impact on the geological barrier

Thermal impact

Permafrost can normally be expected to reach down to a few hundred metres depth beneath the ice divide and affect the groundwater flow geometry. The depth of freezing may change in response to changes in the ice sheet with an accompanying phase change. As mentioned above, model results suggest that freezing at repository depths will not occur /Claesson et al 1996, Vallander et al 1991/.

Hydrological impact

Modelling (figures 4-9 and 4-10) suggests that the bed beneath the broad region of the ice divide is frozen. In this region, there is no melting and, in the absence of a proximal zone of melting, no groundwater flow. As a consequence there will be no head gradient. The groundwater head surface will be horizontal and defined by the head at the proximal limit of the zone of melting (Figure 4-8). Water pressures will be lower than ice pressures and effective pressures will be high. In this zone therefore, rock permeability will be decreased because of increased effective pressure, and permeability will change through a glacial cycle of changing ice thickness. As ice thickness increases during ice sheet growth, rock masses will be compressed and as a consequence transient groundwater flow out of this zone will occur even though there is no melting. During ice sheet decay, rock mass expansion will draw water into the zone. Although the overall effect may be small, there will therefore be a transient head gradient rather than the horizontal groundwater head surface shown in Figure 4-8.

Mechanical impact

The ice load implies that the rock mass is strongly depressed (downwarping). However, this occurs over a large region and will not result in any significant movements along individual fractures or fracture zones. Relatively large effective pressures will develop in this regime resulting in an increase in the rock stresses at repository depth. In general, we expect that the maximum principal stress will be vertical. The difference between the additions in vertical and horizontal stress will result in development of relatively large deviatoric stresses, which potentially can induce shearing of critically oriented fracture zones. The induced stress redistribution can be expected to be rather complex for complex patterns of fracture zones and thus difficult to predict in detail. This is also valid for effective stresses, especially if fluid pressures are high.

During ice sheet retreat the reduction in effective pressure on the surface may release strain energy to cause dilation of fractures.

Large hydraulic heads imply a considerable hydrostatic pressure in the subsurface.

Chemical impact

Freezing in the permafrost layer beneath the ice sheet divide can result in higher groundwater salinity. The groundwater below the permafrost is expected to be more or less stagnant. The saline / fresh water interface may be expected to move vertically as an ice sheet advances and retreats across a site /Svensson 1999b/. Mixing of the deep saline water with other, more shallow waters will potentially occur.

4.3.8 Melting zone regime - impact on the geological barrier

Thermal impact

The glacier is warm-based in this area and melting takes place.

Hydrological impact

Meltwater produced at the bed of the ice sheet in the melting zone cannot accumulate there, but must be discharged towards the ice sheet terminus. The slopes of the ice sheet surface provide the pressure drive for the flow of water melting from the bed of the ice sheet. Although basal melting rates are low, of order less than 0.1 m/y, the horizontal flowline is long, of order 10^5 -10^6 m, giving two-dimensional fluxes of order 10^4 – 10^5 m^2/y.

Water produced by melting at the sole of the ice sheet will first be pressed into the ground and discharged as groundwater. The pressure surface represented by the ice sheet surface gives the limiting head gradient for subglacial water flow. When the head, which drives groundwater flow, rises to this level, it is assumed that excess water forms conduits at the ice/bed interface and is drained through them. The large-scale transmissivity of most rock sequences in Sweden is inadequate to maintain fluxes of the magnitude referred to above. It is concluded therefore that a minor part of the flux is discharged by groundwater flow and the major part is discharged through conduits at the ice/bed interface.

A theory which explores the subglacial drainage systems in three dimensions has been presented /Boulton et al 2001/. The large-scale geometry of channel distribution (see figure 4-15a) is a product of the pattern of basal meltwater production and the transmissive properties of the bed. These subglacial tunnels are the main agents of long-distance water transport to the ice sheet margin. The drawdown of the hydraulic head (figure 4-15b) ensures that the groundwater flow is predominantly transverse to ice flow. On the basis of these principles, modelling of ice tunnel flow and groundwater flow towards tunnels for the Äspö area has been conducted /Svensson 1999b/.

The water, which flows in conduits, is assumed ultimately to give rise to eskers. The Swedish esker spacing (normally 5–20 km) gives the spacing between conduits and can also be used to estimate typical basal melting rates.

If disregarding the local effects of ice tunnels, effective pressures at the ice / bed interface will be near zero, and as a consequence, we assume that, in contrast to the conditions in the ice divide zone, rock permeability will not be very different from the unglaciated state. However, there may be a difference in permeability anisotropy between glacial and non-glacial conditions.

Mechanical impact

In the zone of melting, effective pressures at the ice/bed interface will be zero or near zero. We expect the largest principal stress to be vertical beneath the melting zone.

Water pressures equivalent to the elevation of the ice sheet imply a considerable hydrostatic load in the subsurface.

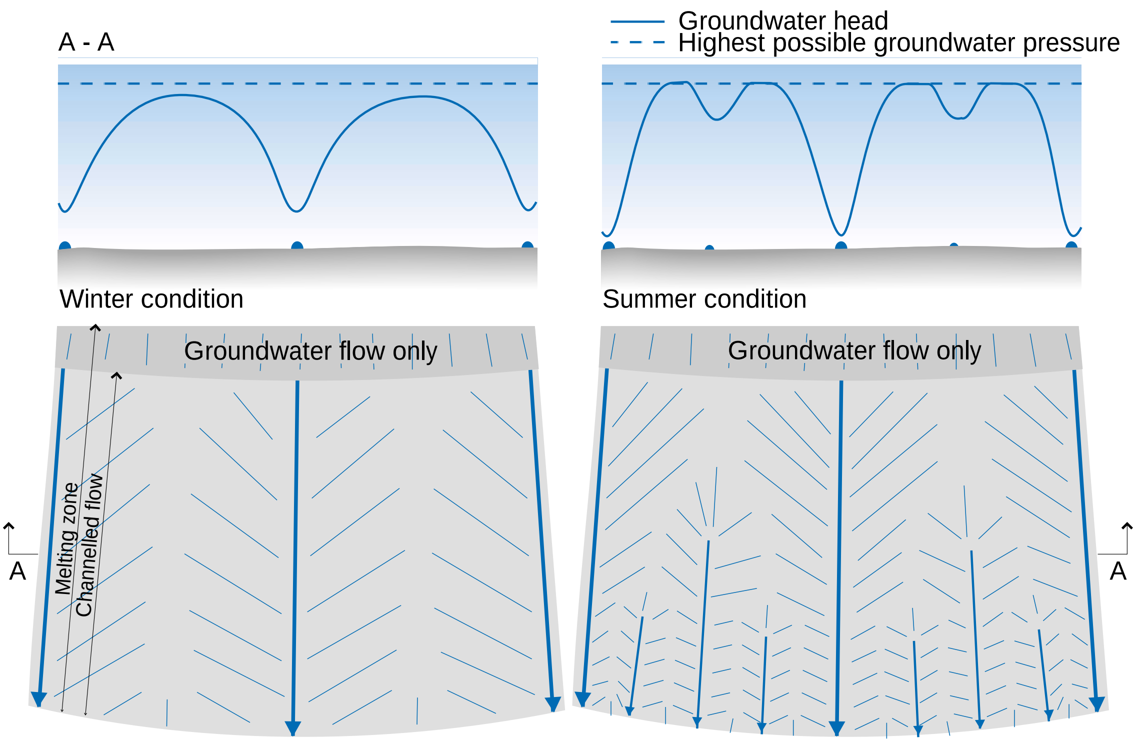


Figure 4-15. *Top) Illustration of the subglacial drainage system in the melting zone Bottom) Illustration of the drawdown of pressure along subglacial tunnels /Boulton et al 2001/.*

Pre-existing vertical fractures tend to be hydraulically jacked open subglacially where the maximum principal stress is predominantly vertical /Boulton 1995/. The fractures would be filled by forcing overlying sediment into them as a consequence of seepage forces generated by the downward potential gradients associated with infiltration of basal meltwater. As a glacier retreats, horizontal fractures begin to open under hydraulic jacking conditions, as the maximum principal stress becomes horizontal. Sediment is injected along them to form the horizontal components of sediment filled fracture networks. This is in accordance with observations made at Forsmark, where vertical and horizontal fractures filled with surface-derived sediments have been found /e g Pusch et al 1990, Talbot 1990/.

Chemical impact

The meltwaters produced at the base of the ice sheet are highly oxygenated. The large pressure heads may imply that such waters can penetrate to large depths through conductive vertical zones. Evidence for the presence of glacial waters to depths of at least 400 m exists at many sites . Results from a modelling study of reactive transport /Guimera et al 1999/ suggest, however, that despite the glacial origin, ground waters will remain anoxic due to the presence of iron acting as oxidant sinks.

Preliminary hydrogeological buoyancy modelling /Svensson 1996/ suggest theoretical penetration to depths of about 2 km. However, it is clear that the presence of physicochemical barriers such as density variations and subhorizontal fracture zones may inhibit the penetration.

Simulations of the effects of ice tunnel flow /Svensson 1999b/ indicate that the salinity distributions are locally affected also at rather great depths since the low pressures cause an uplift of salt water.

4.3.9 Ice marginal regime - impact on the geological barrier

Thermal impact

In the cases without proglacial permafrost there is no thermal impact on the geological barrier. In the case with proglacial permafrost, the permafrost is continuous and the thermal effects are comparable to those described in that process domain.

Hydrological impact

In general, in the terminal, frozen zone of the ice sheet, the absence of melting will produce groundwater flowlines parallel to the ice sheet sole. Since the bed is frozen, meltwater originating from a position beneath the glacier is forced to rise strongly to the surface in the area into the proglacial zone (see figure 4-13).

In the subregime with proglacial permafrost, the groundwater discharge flowing from beneath the ice sheet must be maintained beneath the permafrost, and will require to be driven by a head gradient identical to that beneath the terminal part of the glacier if transmissivity is constant (Figure 4-14b). As a consequence, buoyant overpressures will be developed in the terminal zone and beneath the permafrost. Groundwater will rise strongly to the surface immediately beyond the outer edge of the continuous permafrost or through taliks breaking the permafrost.

In the subregime where the ice sheet terminates in water, the groundwater discharge is driven by a head gradient given by the difference between ice sheet elevation and sea (or lake) level.

Mechanical impact

We can expect a rather complex sequence of stress changes in this regime during ice retreat, which is likely to affect both the stability and water-conducting ability of the rock mass.

The existence of permafrost in the proglacial zone can, in combination with large hydraulic gradients, generate excess fluid pressures and produce buoyant forces which will dilate intergranular voids and fractures and suck in water. A condition is created whereby hydraulic jacking of horizontal joints can occur so as to produce sediment filled joints as described in section 4.3.8. The conditions for hydromechanical instability in the bedrock close to the ice front have been analytically investigated /Lindblom 1997/. It was found that hydraulic jacking phenomena could theoretically be expected at great depths and that hydraulic shearing may also occur.

Chemical impact

In this regime, water originating from beneath the glacier is forced to the surface beyond the edge of the ice sheet or permafrost. The chemistry of this discharge water is dependent on the transit times and from what depth water is forced upwards. For the subregime where the ice terminates in the sea, there will be mixing of discharge water with seawater.

5 Evaluating the impact of future climate change in performance assessment

5.1 Scenarios

Scenarios deal with uncertainties of the future by using alternative outcomes based upon what is judged to be credible considering available knowledge. Performance and safety assessment scenarios for climatically-driven effects on deep repositories should include:

a) Climatically-driven processes with potential impact on the repository system.

b) Interactions between these processes and the repository system.

c) Influence of the interactions on barrier performance and function.

d) Influence of changed barrier performance on overall system performance and safety.

This report contains item a) and b). The mainly qualitative descriptions are supported by quantitative results from model simulations. The changes within the geological barrier are also treated (part of item c) but the impact on the geological barrier performance and function is not analysed. Analysis of impacts on barrier performance and function is part of the performance and safety assessment itself. The results described in this report are meant to support performance assessment.

The performance assessment requires answers to the questions:

1. Will there be climatically-driven processes likely to impact the repository system?
2. What will be the nature and magnitude of these processes?
3. When will the changes occur?
4. How long will they last?

Questions 1) and 2) have been addressed in previous chapters through an analysis of climate-driven domains and regimes and their impact on the geological barrier. But to define scenarios the sequence of domains and regimes in time and their duration also needs to be addressed.

5.2 Forecasting climate change

The existence of long term climate change and periods of major glaciation has been known since the turn of the 1900s century. Physical evidence of a long series of ice ages was obtained in the middle of this century owing to improved deep-sea coring technique and the use of advanced mass spectrometry analyses /Holmgren et al 1998/. During the last decades information about long term climate change has been collected from loess /Maher et al 1995/, deep-sea sediments /e g Imbrie et al 1984; Martinson et al 1987; Raymo et al 1997/, lacustrine sediments /e g Guiot et al 1989/ and glaciers /e g Johnsen et al 1995; Jouzel et al 1993; Petit et al 1999/. All these records show a cyclical pattern of long cold periods with major glaciations, interrupted by shorter warm interglacial periods. Short-term variations are superimposed on the long-term fluctuations.

The seasonal distribution and amount of solar radiation reaching the earth also vary in a cyclical pattern. In the 1940s the Serbian astronomer and mathematician Milankovitch suggested that long-term climate changes are driven by changes in insolation due to changes in the earth orbit around the sun. Later the variations between warmer and colder climate seen in various proxy records have been correlated to changes in the orbital parameters /Hays et al 1976; Imbrie et al 1983/.

The changes in solar radiation reaching the earth do not explain the range of climate change derived from proxy records. Approximately the same changes in orbital parameters have during different periods of earth history given a different climate response. During the Quaternary even though variations in radiation affect the hemispheres differently there is a global synchrony in timing of major changes. The conclusion from these observations is that the major and sometimes abrupt changes in earth climate are partly an effect of internal feedback processes. The feedback processes can both amplify or soften the external forcing /Holmgren et al 1998/.

There is a general opinion that the change in solar radiation due to changes in the orbital parameters is the driver of long-term climate changes during the Quaternary. The response of the earth climate system and the importance of involved feedback processes are though not fully understood and debated. The ice sheets of the Northern Hemisphere are believed to be an important factor.

Past climate conditions can be reconstructed from different kinds of geological and biological records. The variations – both past and future – of the orbital parameters can be calculated /Berger 1978; Berger et al 1991/. To predict future climate we need to correlate the observed climate changes to the orbital variations. This requires modelling of the causality between the driver and the climate change.

The earth's climate system is very complex. There are processes, feedback mechanisms and interdependencies that are not fully understood, furthermore the system is believed to include some chaotic aspects. Consequently predictions of future climate are bound to contain uncertainties. We believe though that if models can depict past conditions they can also be used to create scenarios of future evolution. If the models can reproduce the conditions that used to be, we believe they can predict a future that may be.

5.3 To formulate scenarios from domains and regimes

In previous chapters changing temperature, precipitation and sea level, occurrence of permafrost and growth/decline of glaciers were identified as climate-driven changes with potential impact on a deep geological repository. Thermal, hydrological, mechanical and chemical boundary conditions of importance were discussed for different climate-driven process domains and regimes. For the identified key parameters future trends are not specifically predictable and the probability of a particular outcome cannot be determined. Instead we suggest that geological precedent can be used to generate possible future scenarios of surface environments. Such scenarios consist of a sequence of climate-driven domains (temperate/boreal, permafrost, glacial), which have the potential to influence processes and states in the repository system. The scenario describes a time dependent variation of surface environments. The thermal, hydrological, mechanical and chemical key properties can be described either as continuous variation determined by modelling or as discontinuous series of arbitrarily fixed conditions.

From correlation between global time series of palaeoenvironmental change and records of regional change it is possible to reconstruct the sequence of domains which have occupied areas of Sweden. This can be done with confidence for the last glacial cycle, and with a lesser confidence through several past glacial cycles. Fitting models to these geologically inferred sequences, permits us to reconstruct the past sequence of regimes within the principle domains. Such sequences of past regimes and domains can be regarded as scenarios that have been fulfilled. Reconstructions of the past permit us to test if our models are able adequately to capture significant boundary conditions and their interaction with the geological barrier, and to form templates for scenarios of future change. Scenarios whose logic and structure are based on relatively well understood patterns of past change form a clear basis for the creation of scenarios about future sequences of domains and regimes.

The future can be seen as a continuation of past trends and processes whose evolution will be driven by processes similar to those of the immediately antecedent past, and which will contain a memory of that past. Models of climate evolution which have been tested against their capacity to simulate past change and which permit the sequence of regimes in time to be reconstructed, can also be used to forecast future change and to form a framework for future regimes. This approach permits a semi-quantitative analysis to be performed.

5.4 The ice sheet model

The growth and decay of the Scandinavian ice sheet has played a key role in determining the surface environment in Sweden during the geologically recent past. Modelling the fluctuations of the sheet and the associated permafrost belts is necessary in reconstructing past regimes and thereby quantifying the climate-driven boundary conditions needed for analysis of sub-surface evolution. It is equally important in forecasting environmental regimes in scenarios of future change. As a tool to reconstruct the past and to create scenarios of the future a numerical ice sheet model has been developed.

Input data to the ice sheet model are:

- Mass balance pattern
- Sea level air temperature (SLAT) as a function of time
- Equilibrium line altitude (ELA) as a function of time
- Calving at the marine margin
- Geothermal flux and temperature field within the underlying lithosphere
- Form of the lithosphere surface
- Rheology of the lithosphere and diffusivity of the asthenosphere
- A basal sliding function

SLAT, ELA and mass balance pattern are the climate descriptors required to drive the ice sheet model. They are discussed in Appendix A: Climate and climate change. Remaining input and the model itself are presented in Appendix B: The ice sheet model. The model comprises five separate components:

- ice sheet form;
- internal velocity field of the ice sheet;

- internal temperature field of the ice sheet;
- temperature field in the underlying bedrock;
- isostatic response of the underlying bedrock

As the flow of ice is non-linearly dependent upon its temperature, temperature and flow are coupled within the model. The model computes the changes in ice sheet flow, form, extent, internal temperature and melt water recharge in response to changes in mass balance and temperature on its surface. Furthermore the permafrost depth is calculated. The computation of lithosphere flexure as a consequence of ice loading permits relative sea levels in the vicinity of the ice sheet to be estimated.

The ice sheet model has been used to simulate past glaciations along a transect through Sweden. The results are presented in Appendix C: Results from the ice sheet model. The model can be tested by simulating some aspects of the geological evidence of past glaciations.

The ice sheet model has primarily been used as a research tool to examine the behaviour of a Scandinavian ice sheet and the interactions between ice sheet and subsurface. The ice sheet model generates continuously varying boundary conditions for subsurface processes. To examine the impact of glaciation on groundwater flow, groundwater chemistry and rock stresses, the ice sheet model has been used to drive models of hydrogeological and rock-mechanical response in the sub-surface /Boulton et al 1995a; Boulton et 1995b/. Results from simulations of the Scandinavian ice sheet have also been used in an analysis of groundwater flow beneath ice sheets including a theory for the formation of eskers /Boulton et al 1999/.

5.5 Involved uncertainties

The major uncertainties in creating scenarios as described above are in forecasting climate and in the boundary conditions and parameter values in simulation modelling.

There is a common opinion that the variation in orbital parameters is the driving force of long-term climate change. The orbital parameters can be predicted quite well /Berger et al 1991/. We therefore believe that our models can capture the periodicity of change quite well, while the involved uncertainties are greater when predicting the magnitude and exact timing of change.

A problem in extrapolating forecasts from the geological record of past change is that the latter shows "change points", where the frequency spectrum of the record changes. Long oceanic records show that prior to 700–800 ka, a 40 ka cycle of lesser amplitude than the subsequent 100ka cycles dominated the pattern of global change. Computations of Milankovitch forcing beyond 700–800 ka have failed to reveal any systematic contrasts with the later period, and have led to the conclusion that internal dynamic behaviour within the climate system has conditioned a different response to similar patterns of external forcing at different times. It has been suggested /Saltzman et al 1990/ that this reflects a progressive decrease in the atmospheric concentration of CO_2. Another explanation is that Cenozoic mountain uplift has changed the gross configuration of the earth's surface so as to change the circulation of the atmosphere and the oceans /Ruddiman et al 1990/. It is assumed that the changed circulation effects the efficiency with which excess equatorial heat is transported poleward sufficiently to change the response of the climate system to changes in incoming solar radiation. Although internal forcings change gradually, the response of the climate system switches suddenly at critical thresholds. If such change points occurred in the past, they can occur in the future, and are a source of uncertainty.

The approach taken in this report is restricted to forecasts of "natural" climate evolution. It ignores the anthropogenic impact on climate through the burning of fossil fuels which, it is widely believed, may drive the evolution of climate in the immediate future away from the natural trajectory. A model of the Northern Hemisphere climate has been used to investigate the impacts of different greenhouse gas scenarios /Berger et al 1996/. The model results suggest that the impact greenhouse warming will be to delay major climatic events during 10-thousands of years, but that subsequently, the changes will converge with the natural trajectory of climate.

If long term climate change is a chaotic process, the specific state of the system at a given time in the future is inherently unpredictable. We cannot assign a probability that climate will be in a specific state at some time in the future. This arises because of the non-linearity of the climate system such that small differences in initial conditions between two evolutions may produce large differences in outcome.

Although our knowledge of the earth complex climate system has improved over the last decades, it is not possible to quantify the uncertainties involved with predictions of future climate conditions.

Uncertainties in modelling arise from physical and mathematical approximations, from scale problems and from parameter uncertainties. This is true both for the ice sheet model and the models used to evaluate the impact on the geological barrier. Below some of the uncertainties involved with the ice sheet model are commented.

Physical and mathematical uncertainties

Impacts on ice sheet model output of the low slope approximation, of ignoring longitudinal stresses, of using a flowline rather than a three- dimensional flow model, of progressive ice fabric evolution and consequent time dependent changes in ice rheology, etc have to some extent been explored through sensitivity tests /e g Boulton et al 1992; Boulton et al 1995/. A full analysis of their impact on environmental forecasts would require a substantial programme of research. We assume however, that the effective tuning of the ice sheet model so that it responds in a way which is compatible with the geological evidence of ice sheet fluctuation in Europe is able to contain some of these uncertainties. Testing the results of hindcasts gives confidence that model behaviour is reasonably realistic through its capacity to simulate unparameterised behaviour.

Parameter uncertainties

Parameters such as those in the ice flow law and basal sliding law are derived from best estimates in glaciological experiments. The history matching process is assumed to contain parameter uncertainty.

Problems of scale

An important potential problem is the possibility that small-scale properties of ice flow over a highly variable bed may control aspects of large-scale evolution of the ice sheet. This is primarily a computational problem, and is currently being addressed, as more computing power becomes available.

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Climate and climate change

1 The climate system

The climate is by definition a summary of weather conditions during a certain period. Climate is usually described through statistical properties of the climatic elements, for example mean, maximum and minimum values. The most important climatic elements are air temperature, precipitation, humidity, air pressure and wind.

Processes in the atmosphere determine the *weather* but *climate* does not merely depend upon atmospheric phenomena. The physiography of the surface and the disposition of the continents determine the poleward energy flow paths. A very large proportion of the poleward energy flow takes place through the ocean, with ocean heat and pressure anomalies being strongly coupled to atmospheric flow. The nature of the surface (e.g. forest, grassland, ocean, snow or ice cover, sandy desert) determines the albedo, that is the extent to which solar radiation is absorbed or reflected by the surface. The continental and oceanic biosphere and processes of weathering play fundamental roles in governing the concentration of atmospheric gases, including greenhouse gases which absorb and re-emit solar radiation in such a way as to warm the lower atmosphere. As a consequence, we now regard the state of the climate to be determined by the *climate system*, comprising the atmosphere, the biosphere, the oceans, the ice sheets, and the surface of the lithosphere (Figure 3-1 main text).

Radiation from the sun is the primary energy source driving the climate system /Bogren et al 1998; Holland et al 1999/. The global energy balance is determined by the balance between incoming solar radiation and the amount of radiation that is reflected and emitted from the earth. Clouds and aerosols in the atmosphere or the earth surface reflects 30–40% of the solar radiation, and ozone, water vapour and other gases in the atmosphere absorb 15–25%. The remaining 40–50% is absorbed by, and heats the earth surface. The heated earth emits long-wave radiation. This radiation is highly affected by the gas content in the atmosphere. While the atmosphere is more or less open for the incoming short-wave solar radiation (0.15–4 μm), the long-wave radiation (4–100 μm) can only leave the system within the so-called atmospheric window (8–14 μm). Gases in the atmosphere, mainly water vapour and carbon dioxide, absorb the long-wave radiation. Because of this the atmosphere gives rise to a greenhouse effect, which rises the temperature on earth.

The amount of radiation that is absorbed by the earth surface and thus the heating is affected by the surface albedo (the quota between reflected and incoming radiation). For land areas the albedo can vary from 5 to 45%. Dark areas, for instance woodland, have low albedo, while light areas like desert sand have higher albedo. Newly fallen snow has a very high albedo, 75–95%. The major part of the earth surface is covered by water. Thus the albedo of the surface of water is of great importance. The occurrence of waves and the angle of incidence of the solar radiation affect the albedo. Normally the albedo of the surface of water is 5–10%, but it can be up to 50% at low altitudes of the sun. Short-wave solar radiation penetrates into the water. Because of this and the mixing of water the energy from the absorbed radiation is distributed over a larger volume than energy absorbed by land areas. Furthermore the high specific heat of water makes the oceans effective reservoirs for the earth's heat energy.

The fact that the radiation energy reaching the earth surface depends on the altitude of the sun and thus varies with latitude, and that different surfaces, mainly land and water, absorbs radiation energy to different extent give rise to an uneven heating of the earth. The climate is a result of this uneven heating and the heat transfer mechanisms to which it gives rise. Energy flow takes places in the atmosphere and in the oceans, gravity forces affect the flow. Transport of warm air towards the poles and cold air from them – sensible heat flow in the atmosphere – contribute to about a third of the energy flow. About the same amount of energy is transported via ocean currents. The remaining third is transported via latent flow, that is water evaporates in tropical areas and the heat of vaporisation is released when the water vapour condenses at higher latitudes.

The energy budget at the surface depends on the gas composition in the atmosphere and the albedo of the surface. Furthermore heat transfer in the atmosphere and the oceans affects the energy budget. Physical, chemical and biological processes affecting the climate are strongly coupled. Heat and water exchange and also exchange of particles, between and within the different parts of the climate system impact the climatic processes. The climate system is a combination of deterministic behaviour and unpredictable chaotic fluctuations.

1.1 Climate change

The radiation from the sun is the primary source of energy for the earth's climate system. Variation of the solar radiation is consequently a cause of climate change. Variations in insolation have two different sources. The first is related to variations in the earth orbit around the sun, which mainly affect the geographical distribution, but also the amount, of radiation that reaches the earth. The other source of variability is changes in the solar irradiation. Another source of external forcing on the climate system is dissipation of internal earth energy producing volcanism or shifts in earth physiography (plate movement, mountain uplift).

A change in internal dynamics of the climate system is another source of climate change. The concentration of different gases in the atmosphere affects the heat balance and the meteorological processes and thereby climate. Important for the climate are also biological processes, the albedo and the dynamics of ocean currents and ice sheets.

The most important gases affecting the climate are water vapour, carbondioxide, ozone, methane, nitrous oxide and chlorofluorcarbons (CFCs). Stratospherical ozone is of major importance for the absorption of ultraviolet radiation from the sun, and of great importance for life on earth. Water vapour is the most important greenhouse gas, it contributes to 60% of the natural greenhouse effect /Bogren et al 1998/. The content of water vapour in the atmosphere also affects the occurrence of clouds. Clouds reflect incoming solar radiation back to space. They also emit long-wave radiation to space and back to the earth. Clouds close the atmospheric window through which long-wave radiation emitted from the earth surface can leave the climate system. Carbon dioxide is the second most important greenhouse gas, it contributes to 26% of the natural greenhouse effect /Bogren et al 1998/. The occurrence of aerosols in the atmosphere plays an important role in the climate system, they absorb, reflect and scatter solar and terrestrial radiation and influence the formation of clouds.

The surface albedo is also of great importance for the heat balance. Dark areas like woodland and water can reflect as little as 5–10% of the incoming radiation while light areas as snow and ice can reflect up to 95%.

The temperature and salinity of the water affect the dynamics of ocean currents. Ice sheets affect the climate in several ways. They have high albedo, they contribute to lower sea levels, and they impact ocean currents by affecting the salinity.

The components of the climate system interact via physical, chemical and biological processes. The response times of the components of the climate system to an external forcing differ significantly /Holland et al 1999; Henderson-Sellers 1996/:

- Atmosphere: from less than a day up to a few years,
- Biosphere: from years to several hundred years,
- Oceans: from a week to several thousand years,
- Ice sheets: thousand to several hundred thousand years,
- The surface of the lithosphere: from millions to hundred of million years.

Furthermore the response of the climate system to a change can be amplified or reduced by feedback processes. A positive feedback process amplifies a change and decreases the stability of the system. Negative feedback processes increase the stability. Some important feedback processes are:

- Ice-albedo:
  When the climate gets warmer ice and snow melts and the albedo of the surface increases which causes a positive feedback due to a warming of the earth surface and the atmosphere.
- Temperature-water vapour:
  When the temperature rises the atmosphere can keep more water vapour, which causes a positive feedback because of an enhanced greenhouse effect.
- Temperature-carbondioxide:
  The content of carbondioxide in the atmosphere is determined by the content of solved carbondioxide in the oceans. The solubility decreases with increasing water temperature. Increasing temperature in the oceans will thus cause increased content of carbondioxide in the atmosphere and an enhanced greenhouse effect.
- Temperature-cloudiness:
  Increased temperature and humidity in the atmosphere cause increasing cloudiness. Increased cloudiness causes increased reflection of incoming solar radiation and increased reemission and blocking of long-wave radiation from the earth. If the effects on the long-wave radiation exceed the increased reflection a positive feedback is generated, otherwise the feedback process will be negative. If the feedback process will be positive or negative depends on whether the clouds are high or low. (Changes in the content of aerosols in the atmosphere also affect the cloudiness.)
- Vegetation-albedo:
  Woodland has lower albedo than deserts. The vegetation is of great importance for the albedo. The feedback processes can be either positive or negative.

The result of different magnitude and time scale of external forcing, different response times of the components of the climate system, feedback processes and the complexity of the climate system is a rich spectrum of climate change.

1.2 Past climate

During the past millions of years the earth's climate has been characterised by global cold periods when continental ice sheets and glaciers have extended. The cold periods have been interrupted by shorter warm periods with a climate similar to the current. The cold periods are termed glacials and the warm periods are termed interglacials. The glacials contain colder and warmer stages called stadials and interstadials, respectively. During the past 900,000 years or so, a cyclic pattern with approximately 100,000-year-long glacials, abruptly terminated by a transition to a warm interglacial climate, is repeated. The stadials towards the end of the glacials tend to be the coldest. They end suddenly in a rapid transition to interglacial conditions.

The changes between glacial and interglacial conditions are related to variations of the orbital parameters. The orbital parameters are eccentricity, obliquity and precession and their variation primarily change the intensity of the seasonal cycle. Changes in eccentricity – the form of the earth's orbit changes from being almost circular to being elliptic – have a periodicity of about 100,000 years. Changes in eccentricity affect the amount of solar energy received by the earth. Changes in obliquity – the tilt of the earth axis – have a periodicity of about 41,000 years. Increased tilt causes a larger contrast between summer and winter. The precession – the motion of the spinning earth makes it wobble so that the axis of rotation sweeps out a cone. Precession together with the rotation around the sun affects the season in which the earth makes its closest approach to the sun. The precession has a periodicity of about 22,000 years. It is amplified by large eccentricity and affects the seasonal contrast.

There is a general consensus that long-term climate changes (10,000–100,000 years) are triggered by variations of the orbital parameters. During the last 700,000–900,000 years the 100,000 years long glacial cycle mentioned above dominates the climate change. The spectrum of climate variations also has large amplitudes for periodicities correlated to variations in obliquity and precession. The dominance of the 41,000-year cycle related to changes in obliquity can be seen for the last 6 million years /Holland et al 1999/. The 100,000-year climate cycle has only existed the last 900,000 years, and the role of the variation in eccentricity as the forcing factor of the glacial cycle has been questioned /Holmgren et al 1998/.

Abrupt millennial-scale changes in climate, termed Dansgaard-Oeschger cycles, have punctuated the last glacial period (about 10,000–100,000 years ago) but not the following interglacial period (Holocene, the past 10,000 years) /Holland et al 1999, Holmgren et al 1998/. The events start with an abrupt warming of Greenland by 5–15°C over a few decades. Then a gradual cooling over several hundred years follow, ending with an abrupt final reduction of temperature back to cold stadial conditions. Several such successively cooler cycles are bundled together. In the beginning and end of these bundled cycles, every 7,000–10,000 years, during the cold stages of the Dansgaard-Oeschger cycles large discharges of icebergs from the Laurentide and European ice sheet, so-called Heinrich events, took place /Holland et al 1999/. Abrupt millennial-scale changes have also been seen during the previous interglacial, the Eem, which was slighly warmer than the current, the Holocene.

These large and abrupt changes are believed to involve large changes in the ocean heat transport. The Atlantic thermohaline circulation is a significant contributor to the regional heat budget over the North Atlantic region. Furthermore it is sensitive to changes in the salinity in the North Atlantic and shows a non-linear behaviour with thresholds for transitions between different circulation modes. The Atlantic thermohaline circulation is believed to play a crucial role for millennial-scale climate changes.

Deep-sea thermohaline circulation is driven by differences in temperature and salinity. In the North Atlantic cold air cools the surface water, simultaneously freezing increases the salinity. The dense, cold, salty water plummets to the bottom of the ocean, a so-called sink is created. As the polar water sinks warmer water is drawn in from the south creating a current flowing across the Atlantic from the north to the south. The current is very sensitive to changes in the salinity of the northern Atlantic. Input of freshwater from calving ice sheets during glacial periods moves the sink southwards and the circulation gradually declines, finally it reaches a point where the cooling is not sufficient to create a sink and the current is shut off, causing a cold climate in the Atlantic region. During interglacial periods increased temperature could reduce the cooling and cause a similar effect. North Atlantic deep-water formation, Dansgaard-Oeschger cycles and Heinrich events are illustrated in Figure A1-1.

Ocean circulation is believed to be of major importance for millennial-scale climate change. But it is currently not known what triggers the changes. During glacial periods ice sheet dynamics may cause calving of icebergs and trigger the perturbations of the thermohaline circulation. Changes of temperature occur simultaneously with changes of the ocean circulation, but it is not known if they trigger the perturbations of the thermohaline circulation or vice versa.

The long-term and millennial-scale climate changes are overlapped by changes of shorter periodicity. These changes of smaller amplitude may be triggered by changes in the solar irradiation, volcanic eruptions or internal dynamics of the climate system.

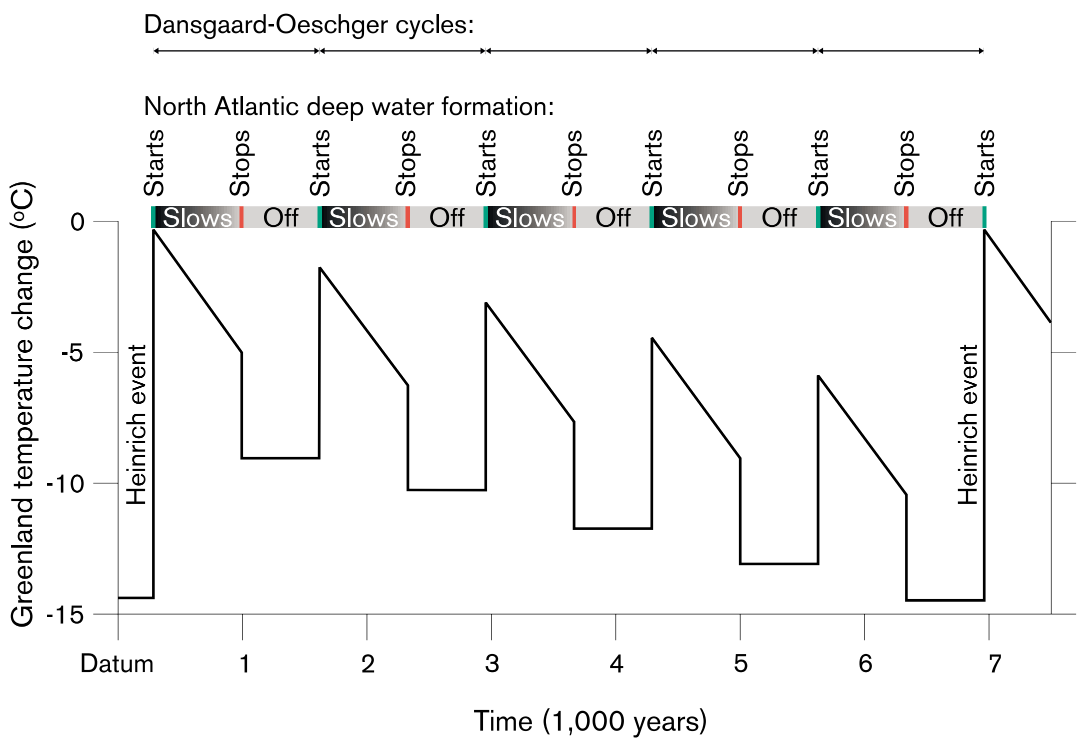

Figure A1-1. *Several successively cooler Dansgaard-Oescheger cycles are bundled together. In the beginning and ending of these bundled cycles Heinrich events occur. Dansgaard-Oescheger cycles start with a rapid warming believed to be associated with the start of North Atlantic deep-water formation. A shut off of the North Atlantic deep-water formation is believed to cause the abrupt cooling at the end of the cycles /revised from A-22/.*

Human activities affect the content of carbondioxide, methane and aerosols in the atmosphere. Land use changes the surface albedo. Today there is a consensus that the warming of the climate that has been observed during the last decades at least partly is caused by human activities. The observed global warming is in the order of tenth of degrees, but anthropologically caused climate change has been calculated to be able to generate global warming of as mush as 1.5–4.5°C /Henderson-Sellers 1996/. This can be compared to the changes in global temperature during a glacial cycle which has been estimated to be about 6°C /Houghton et al 1990/. Human induced global warming may influence the climate for several 10,000 years /Berger et al 1996/.

1.3 Measuring and modelling climate

1.3.1 Climate data

Our knowledge of past climate is based on measurements of climatic variables. Observations of temperature, precipitation and other climate variables began in western Europe during the late 17th and early 18th century and gradually spread to the remaining land areas of the world /Holland et al 1999/. For information of the climate before the 18th century we must interpret proxy data recorded in different natural features such as trees, ice sheets, lake and ocean sediments, shoreline terraces and other morphological features. The transformation of the measured proxy data to climate variables requires models, which often are based on empirical observations under present day conditions /Holmgren et al 1998/.

Historical climate observations were mainly taken to monitor the weather. Considered as climate data they are unfortunately often of poor quality. Changes in instrumentation, relocation and changes in the surrounding environment as well as natural climate variation affect the recorded data. This must be taken into account when evaluating the observed climate changes. Another drawback is that the observations are poorly distributed over the world. They are mainly from land areas on the Northern Hemisphere, and there are very few in situ observations over the oceans. There are though millions of merchant ship observations. Thanks to major efforts the last decades to assemble these data, quality controlled marine data sets extending back to the final decades of the 19th century exist along well-travelled ship tracks /Holland et al 1999/.

The launching of meteorological satellites have made global climate monitoring possible. Satellite time series span the last 20–30 years. The satellite observations are central for the monitoring of sea surface temperature (SST) and precipitation. Merchant ship data do not include precipitation, which means there are practically no pre-satellite precipitation data for the oceans. Other climate variables derived from satellite measurements are content of water vapour in the atmosphere, clouds, radiation, surface and atmospheric temperature, snow cover, vegetation and sea level. Satellite observations are very important for the understanding of current climate dynamics and its variations /Holland et al 1999/.

The abundance of different animal and plant species depends on climate. Key species or groups of species (assemblages) can be associated with broad climate zones. The varying occurrence of fossil plankton in deep-sea sediments, diatoms in lake or marine sediments and pollen in peat bog or lake sediments infer climatic variations. To transfer the occurrence of species and assemblages to climate variables it is necessary to derive a transfer function. The transfer functions are calibrated to recent temperatures etc. In reality the relation between a climate variable and the occurrence of a species is affected by several factors. These factors may be correlated to each other and may be of different importance depending on the overall state of the climate system. This fact, the methods to derive transfer functions, and methods to measure the occurrence of species affect the derived climatic data /Holland et al 1999/.

Another approach to derive past climate variables is to measure the abundance of different isotopes of an element in for instance deep-sea cores or ice cores. Most elements occur in nature as a mixture of stable isotopes. The isotopes have different thermodynamic properties and thus differ slightly in their chemical and physical behaviour. Natural processes may entail separation of stable isotopes. An example of this is that the lighter oxygen isotope, ^{16}O, evaporates more easily than the heavier, ^{18}O. In cold climate ^{16}O is stored in snow and ice in ice sheets and there is an abundance of ^{18}O in the oceans. The $^{18}O/^{16}O$ ratio in oceans consequently reflects continental ice volume changes. There is also a relationship between the content of the heavy isotope in precipitation and temperature. The $^{18}O/^{16}O$ ratio in ice cores reflects changes in temperature. The isotopic variations are small, and generally expressed in per mil as:

$$\delta^{18}O = \frac{(^{18}O/^{16}O)_{sample}}{(^{18}O/^{16}O)_{ref}} - 1$$

where $^{18}O/^{16}O_{ref}$ is a reference value. For sea water it is the mean isotopic composition of the present world ocean and is called SMOW (Standard Mean Ocean Water).

The $^{18}O/^{16}O$ ratio in fossil plankton found in deep-sea cores is an important category of proxy record. Such records have been available the last 50 years, 1955 Emeliani /Emeliani 1955/ presented the first oxygen isotopic analysis of planktonic and benthic foraminifera in deep-sea cores /Holland et al 1999/. Deep-sea cores were the first physical evidence of a long series of ice ages /Holmgren et al 1998/. They have been very important for the development of the astronomical climate theory, that is the theory that long-term climate changes (10,000–100,000 years) are triggered by variations of the orbital parameters /Imbrie et al 1984/. Most deep-sea cores cover a time span of several 100,000 years. Recently a core covering the last 6 million years was provided /Shackleton et al 1995; Shackleton et al 1995 in Holland et al 1999/.

Ice cores are another very important source of information of past climate. Cores from Greenland and Antarctica cover the last 200,000 years /Holmgren et al 1998/. Recently data from a core from Antarctica covering the past 420,000 years was presented /Petit et al 1999/. Variations of $d^{18}O$ in ice cores infer past temperatures. Ice cores also contain other information of importance for past climate, for instance the content of gases in the atmosphere and the occurrence of volcanic eruptions. The above mentioned core from Antarctica for instance tells us that the current content of carbondioxide in the atmosphere is extremely high in a 420,000-year time perspective.

Lake sediments, loess sequences and stalagmites are other sources of long-term climatic change. For the past thousands of years trees constitute an important source of information. In Scandinavia there are no long continuous records providing a time series of climate change. Such records are only available for the past 10,000 years or so, that is since the ending of the last glaciation. The ice sheets, which have covered Scandinavia on several occasions during the past, have almost made it impossible to find long records of climate change in Scandinavia.

A correct dating of the climate changes is of great importance for the understanding of the climate system and the causes of climate change. Methods of dating have improved the last decades, but uncertainties still remain. They grow larger the further back in time we go. Radioactive dating methods are applicable for the past 350,000 years. Radiocarbon dating is possible for the past 40,000 years and samples containing remnants of living organisms. The content of ^{230}Th and ^{231}Pa in sea sediments or in carbonate shells can be used for dating 350,000 years back in time /Holland et al 1999/. Lithological and geophysical observations can be used for longer time scales and for correlation of time scales in different records. Layers of volcanic material can for instance be identified to originate from specific eruptions, and the events can be dated. The periodic reversal of the earth magnetic field can also be identified and dated. For sediments a linear relation between core lengths and age is then assumed.

In order to reproduce past climate states, climate data from all over the earth are required. The traces of past climate found in a proxy is a result of global and regional climate but it depicts the local conditions at the site of the record. For studies of past climate conditions information from various sources must be collected, interpreted and dated. Changes in insolation can be dated with relatively high fidelity and are believed to trigger climate change. A correct dating is necessary to confirm that changes in solar radiation triggers climate change, and if so to determine the lag between changes in insolation and climate.

1.3.2 Climate models

During the last 40 years or so climatology has been transformed from a descriptive science to a branch of science involving studies of coupled dynamic processes. New development in meteorology after the end of the second world war led to the conclusion that it is necessary to analyse and understand the dynamical processes behind the daily weather before we can understand climate /Holland et al 1999/. Since then atmospheric models predicting the weather with high fidelity have been developed. A prerequisite for this development is the access to observations of climate parameters and powerful computers.

Climate has substantial impact on conditions of life and the economic well-being of the nations of the world. To understand the climate system and from that understanding be able to predict future climate is a goal for mankind. Examples of natural phenomena affecting many nations are El Niño events and variation of the intensity of the Indian Monsoon. The awareness of that increasing emission of greenhouse gases, stratospheric ozone depletion, deforestation and other human activities may have major impact on the earth's climate has made understanding of the climate system a very important issue for the society.

Arrhenius made in 1896 the first attempt to make a climate change prediction by the means of modelling. He developed a model for the surface-atmosphere radiation budget, and calculated the effect on climate from changes in the carbondioxide concentration in the atmosphere /Holland et al 1999/. The kind of one-dimensional models Arrhenius, and others after him used, do not include the climate dynamics. In order to investigate the dynamics of the climate system coupled models where an atmospheric component is coupled to models of the ocean and the land surface are required. Such models are generally referred to as general circulation models or global climate models (GCM). The development of GCM models is related to the development of weather prediction models. The atmospheric component of GCM models is essentially the same as in weather prediction models /Henderson-Sellers et al 1996; Åberg 1996/.

The processes controlling climate cover a wide range of time and space scales. The strategy to handle different space scales in modelling has been to use descriptions in simplified physical terms of small-scale processes. The grid size of the atmospheric component of a GCM model is typically 100*100*1 km (length*width*height). Subgrid scale processes are described in physical terms, calculated for each three-dimensional box in the GCM model, averaged and expressed as variables that can be handled by the model. Usually the variables are calculated for vertical columns within each box. This method to handle the wide range of scales is called parametrization /Holland et al 1999; Henderson-Sellers 1996; Åberg 1996/.

Handling different time scales is more complex, one strategy is to adopt the model complexity to the studied time scale. The determination of a suitable model complexity is subjective and related to the purpose of each model study. Bengtsson /in A-20/ suggests four different time scales termed *weather*, *interannual*, *decadal to centennial* and *palaeo*. "Weather" comprises very short time scales from hours to months, and requires high-resolution models with the emphasis on the atmospheric component. The main problem area is high-resolution process studies, for example the development and evaluation of parameterization schemes. "Interannual" – embracing months to tens of years – requires high-resolution coupled models including atmosphere, land surfaces and upper ocean processes. Such models are suitable for studies of climate anomalies such as El Niño events. "Decadal to centennial" – covering time scales from 1 to 1000 years – includes studies of low frequency climate fluctuations and requires fully coupled models including deep ocean processes. The model resolution is suggested to be moderate. To simulate "palaeo" phenomena – on the time scale from hundreds up to millions of years – such as ice ages, all the components of the climate system, as well as the external forcing should be included. In order to achieve this a modest model resolution is recommended.

1.4 Climate scenarios and performance assessment

For the assessment of the performance and safety of a deep geological repository for spent nuclear fuel we are mainly concerned with climate changes causing transitions between different climate driven process domains (see main text section 3.3). These changes are believed to be long-term changes triggered by variations in insolation due to variation of the earth's orbital parameters. To study these climate changes models relating orbital forcing to the growth and decay of ice sheets are required.

Models intended to support the astronomical climate theory developed in the 1980s, for instance /Imbrie et al 1980; Kukla et al 1981/, bypass the complex climate system and directly correlate orbital parameters to proxy data. These models reconstruct past changes reasonably well. As the orbital parameters can be calculated for the future, the models can also be used to create scenarios of coming ice ages. A model developed during the 1990s at the Lowaine-la-Neuve University – the LLN two-dimensional model – includes a two dimensional model of the northern hemisphere climate system coupled to an ice sheet model /Berger et al 1996; Gallée et al 1991; Gallé et al 1992/. This model can be used to study the importance of different components of the climate system for the transitions between glacial and interglacial conditions. In this report results from a model relating a set of climatic variables to the growth and decay of the Scandinavian ice sheet is utilised.

Variations in insolation caused by variations of the orbital parameters are believed to trigger global long-term climate changes. But it must be emphasised that the response of the climate system to such changes in external forcing depends on the state of the climate system. This in turn depends on the complex dynamics accounted for in section 1.1 Climate change. Simulations using the LLN-model mentioned above indicates that human induced global warming may perturb the pattern of glacial cycles /Berger et al 1996/.

Models designed to simulate long-term climate changes do generally not capture millennial-scale climate changes such as Dansgaard-Oeschger cycles. In reality the long-term climate evolution is overlapped by climate changes of shorter frequency. Due to this the development of an ice sheet most probably consists of several phases of advance and retreat. Millennial-scale changes may also very well cause transition between temperate/boreal and permafrost conditions in Scandinavia.

Lack of knowledge and capabilities to describe and simulate the climate system currently make predictions of future climate impossible. If the astronomical climate theory is valid it can be applied to generate simplified scenarios of long-term climate changes. Such scenarios may give a reasonably correct picture of the frequency of long-term changes, while the range of the climate changes includes great uncertainty. Furthermore such scenarios do not capture the dynamics and variability of the climate evolution. In spite of the uncertainties we do believe scenarios, both of past and future evolution, are valuable tools when assessing the performance and safety of a deep geological repository for spent nuclear fuel.

2 Production of a climate drive for the ice sheet model

In order to use the ice sheet model (Appendix B) to simulate the fluctuations of the Scandinavian ice sheet through past glacial cycles, it is necessary to generate a climatic drive. Although, as mentioned above, there are a number of long records of climate change in other parts of the world, there are no such long records in any part of Scandinavia. An estimation of Scandinavian climate must be created from records from elsewhere.

We seek to evaluate long term changes that might effect the performance of a deep geological repository for spent nuclear fuel. As a means of understanding what these changes might be, we need to reconstruct the pattern of climatic and environmental change through the recent geological past and if possible extrapolate the changes into the future. The time periods we want to study are 100,000 years or longer and the frequencies of change 1,000 to 100,000 years.

2.1 Climate input to the ice sheet model

The climate descriptors required to drive the ice sheet model (Appendix B) are:

- A mass balance pattern
- A time series of variations of the equilibrium line altitude (ELA)
- A time series of sea level air temperatures (SLAT)

Ice sheets will grow if snowfall accumulation on the ice sheet surface exceeds any melting or other loss of ice. In the ice sheet model the net balance of snowfall and melt is represented as a single quantity, the *net surface mass balance rate*. This is assumed to be a simple function of the altitude of the ice sheet surface relative to the *Equilibrium Line Altitude* (ELA). The ELA is the notional altitude where the surface mass balance is zero. Below this altitude the mass balance is negative, that is there is more melting than snowfall over the course of year. Above the ELA a net positive mass balance occurs. The slope of the mass balance curve tends to be steeper in wetter-warmer climates and shallower in drier-colder ones. Steeper gradients reflect higher snowfall rates combined with higher melt rates at lower elevations. Empirically based relationships between mass balance and ice sheet altitudes are shown in Figure A2-1.

For each model run a mass balance pattern is selected and an initial ELA is defined for each point in the model. The present day ELA is calculated to be inclined to the south and east (parameterisation /Oerlemans 1981/ and data /Charlesworth 1957/). A time series of ELA perturbations away from the starting values forces the modelled ice sheet to grow or decrease. If the ELA is lowered the zone of positive mass balance extends to lower elevations and causes the ice sheet to grow. Raising the ELA produces the opposite effect. The same ELA perturbation is applied to each point in the model.

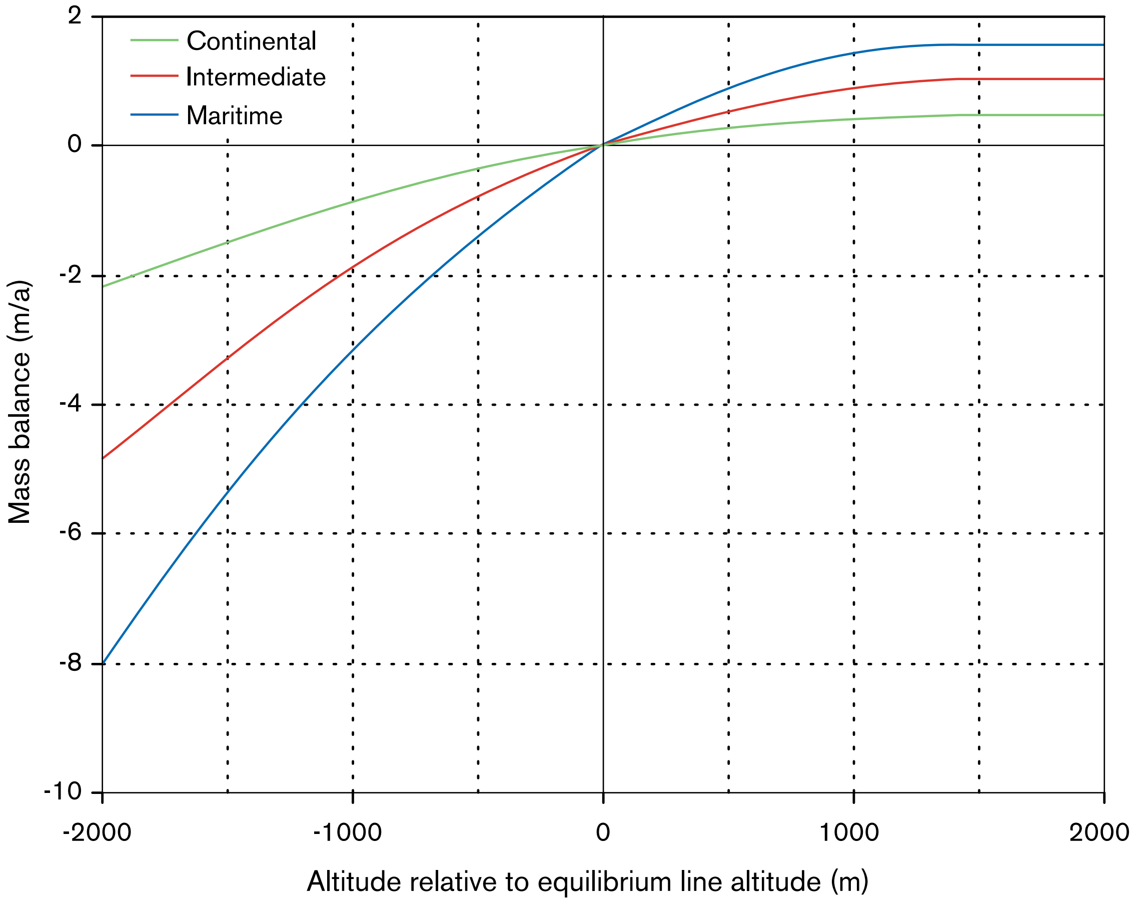

Figure A2-1. *Empirically based mass balance patterns showing continental and maritime extremes /Boulton et al 1984/.*

The *sea level air temperature* (SLAT) is used as a means of determining the mean annual temperature on the ice sheet surface, and is a boundary condition for the computation of heat conduction within the ice sheet. However, advection is a very important process in an ice sheet, and the advection of cold ice masses into the body of the ice sheet plays a major role in determining internal temperatures, which in turn influence ice rheology.

For each model run an initial sea level air temperature is defined. The temperature for each point in the model is calculated assuming a vertical lapse rate. A time series of sea level air temperature variations then generates a time/space temperature field on the ice sheet surface. Sensitivity tests /Boulton et al 1995/ have showed that ice sheet fluctuations are relatively insensitive to the absolute values of SLAT, but they are sensitive to variation of the lapse rate. An assumed lapse rate of 10 °C km^{-1} /Orvig 1970/, typical of rates over modern ice sheets is usually used in the simulations.

The time series of SLAT and ELA perturbations for a fixed mass balance distribution determine the fluctuations of the ice sheet in time. In some simulations, the mass balance distribution is also permitted to vary so as to examine contrasts between marine and continental climatic conditions.

There are no proxies of past ELA fluctuations. In order to generate ELA variations from records of climate change a linear relation is assumed between ELA and palaeotemperatures, so that:

$$\Delta\lambda = f_1(T - f_2) \qquad \text{(A-1)}$$

where:

$\Delta\lambda$ = change in ELA

T = temperature

f_1 and f_2 = constants

Changing the values of f1 and f2 alters the response of the ELA to a defined series of temperature changes and the sensitivity of the ice sheet to climate change. Whether or not sliding occurs at the ice/bed interface strongly effects the response of the ice sheet to climate forcing.

2.2 Using the ice sheet model to analyse palaeoclimate records

The fluctuations of the landward margins of the European and North American ice sheets can be regarded as a smoothed long-term proxy record of climate change. If numerical models of ice sheet dynamics are capable of simulating the response of ice sheets to climatic change, they can also be used to translate the geological evidence of ice sheet fluctuation into indices of palaeoclimate.

Boulton et al. used the ice sheet model (Appendix B) to investigate whether proxy climate sequences are compatible with evidence of contemporary ice sheet fluctuation /Boulton et al 1995/. Ice sheet fluctuations along a flowline from western Norway to northern Germany was simulated using palaeotemperature proxies as input. The results were compared to a geological reconstruction of the European ice sheet. The model was also used in a quasi-inverse mode, such that ice sheet behaviour was prescribed and the model was used to infer characteristics of the climates that could have driven the ice sheet.

Proxy climate sequences from La Grande Pile in France /Guiot et al 1989/ and the North Atlantic (core SU 90-39) were used as time series of temperature change. The ice sheet fluctuations were calculated using different values of the constants f_1 and f_2 in equation A-1. Simulations used different mass balance patterns reflecting more maritime and more continental climatic regimes, and presence and absence of sliding at the ice/bed interface. Forward modelling, in which the ice sheet was driven by proxy palaeoclimate patterns, produced simulations of ice sheet margin fluctuation. The model results were compared with Mangerud's reconstruction of the European ice sheet through the last glacial cycle /Mangerud 1991/. In the quasi inverse simulations, the ELA forcing functions which were required to generate ice sheet fluctuations comparable to Mangerud's reconstruction were determined. Multiple simulations using both non-sliding and sliding conditions were run. The results of the study are summarised in Figure A2-2.

Driving the model using temperature data derived from $\delta^{18}O$-variations in the North Atlantic as input resulted in an ice sheet that did not respond as rapidly as the real ice sheet. Driving the model using palaeo-temperatures from the pollen record from La Grande Pile failed to produce large enough ice sheet extensions at the last glacial maximum and in the period 70–60 ka. In both cases the ice sheet fails to disappear completely during Holocene. The Atlantic $\delta^{18}O$ record produces, on the whole, a better reconstruction than the French pollen record.

In the inverse simulations abrupt rises of ELA are required during the early Holocene in order for the ice sheet to disappear during the Holocene. Relatively recent geological data now suggests that the Holocene climatic optimum did indeed occur during the early Holocene. Ice sheet surface climate shifts must lead the glacial maxima by several thousands of years.

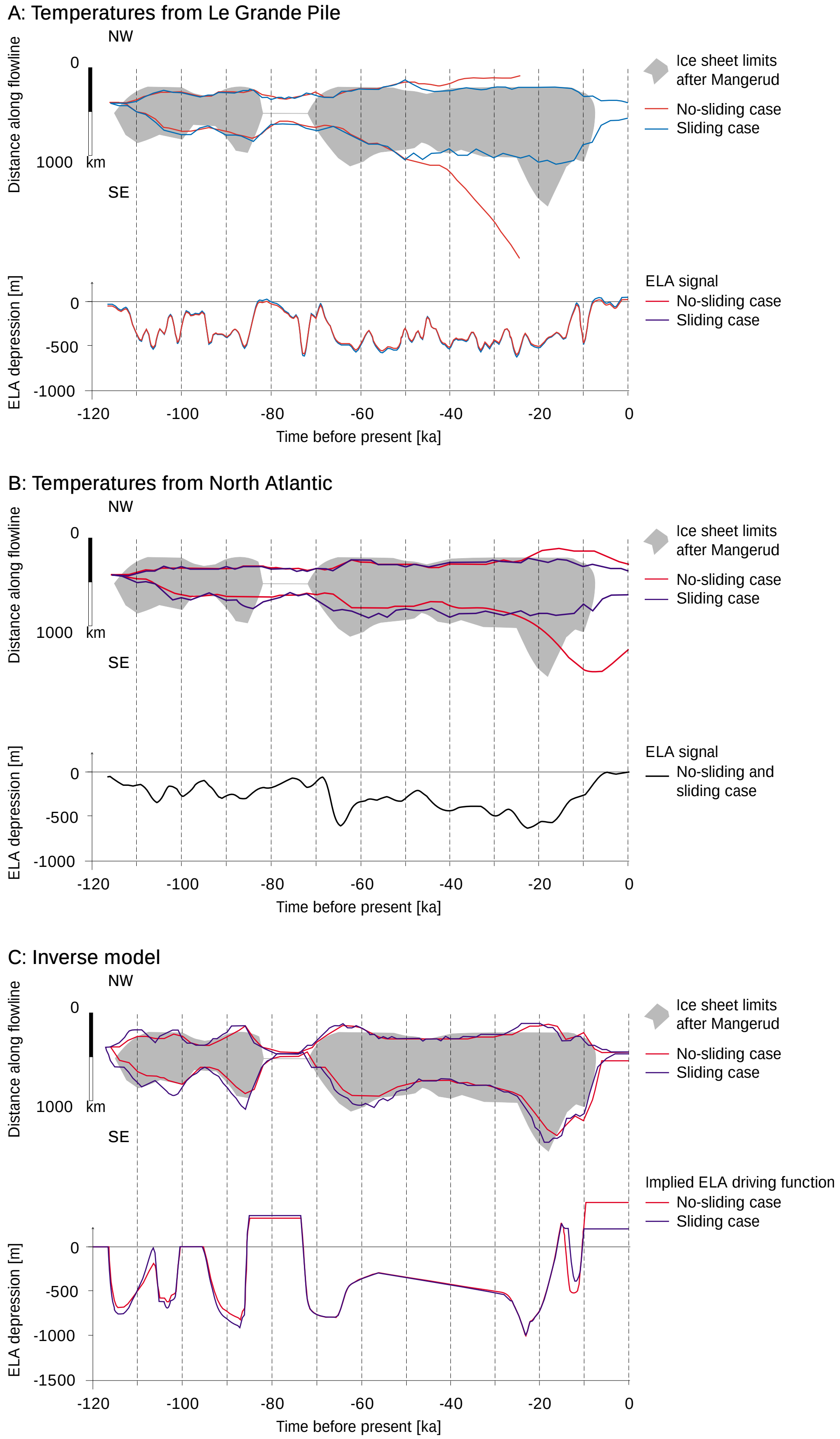

Figure A2-2. *Simulated ice sheet fluctuations compared to Mangerud's geological reconstruction. A) ELA variations based on temperatures derived from pollen data, La Grande Pile, France. B) Using temperature estimates derived from δ^{18}O-variations in the North Atlantic as input. C) ELA variations generated via inverse model simulations. (Revised from /Boulton et al 1995/)*

As can be seen in Figure A2-2 sliding ice sheets are less sensitive to changes to colder conditions than non-sliding ice sheets. This is because sliding ice sheets have a lower surface profile. A sliding ice sheet requires a stronger climate forcing – a greater lowering of the ELA – to achieve the same rate of build-up as a non-sliding ice sheet. The converse occurs during decay as a lower profile gives rise to a more negative mass balance.

Discrepancies between the simulated ice sheet driven by palaeotemperature proxies and Mangerud's geological reconstruction may be explained by several causes:

- Errors in the geological reconstructions, which depend on very sparse stratigraphic evidence.
- Errors in interpretation of palaeotemperature proxies, e.g. inertia in floral and faunal responses to climate change, and long migration times in comparison to the speed of climate change.
- Large climatic gradients near to ice sheet margins, which make proxies far from the ice sheet a poor guide to ice sheet climate.
- Ultimate ice sheet extent at certain parts of its growth can be highly sensitive to mass rate of build up and the form of the evolving ice sheet margin.
- Inappropriate model formulation or application.

2.3 Long term hindcasts and forecasts

The climate drive for long term hindcasts and forecasts are based on the SPECMAP (Spectral Mapping Project) record /Imbrie et al 1984/ and calculated earth orbital parameters. The SPECMAP record consists of $\delta^{18}O$ variations from five deep-sea cores plotted on a time scale derived from the variation of the orbital parameters. The record is essentially a reflection of global climate and ice volume change, and extends back to 780 ka before present, Figure A2-3. The great length of this record permits a statistical correlation to be made with orbital parameters over a period of many glacial cycles.

The work of Kukla et al and Berger et al. /Kukla et al 1981; Berger et al 1981/ were used to introduce the orbital parameters into the analysis. Kukla and others present an astronomical climate index (ACLIN) /Kukla et al 1981/ which predicts the gross climate state as a function of the three orbital parameters. The index is derived from the observation that interglacials are associated with high obliquity and eccentricity and a perihelion that coincident with the autumnal equinox. The empirical formula reads:

$$\alpha_{t1} = \left| \frac{\omega_{t1} - 180}{90} \right| \quad + \quad \varepsilon_{t2} - 22 \quad + \quad 500 e_{t1}^{2}$$

Where α_{t1} is the index, w_{t1} is the longitude of the perihelion[1] and e_{t1} the eccentricity at time t_1, and ε_{t2} is the obliquity at time t_2. The time t_2 is defined by $\omega_{t2} = \omega_{t1} - 90$ and $|t_1 - t_2|$ = min and corresponds to a time lag of ~5,000 – 6,000 years.

Berger and others /Berger et al 1981/ have derived relations between variations in insolation at different latitudes, and times of the year and climate change expressed as $\delta^{18}O$ variations in deep-sea sediments. In one of their regression models climate at one particular year is considered to be a function of both monthly insolations during the year and the climate during the previous 3,000 years. The system memory of the past 3,000 years is included as the $\delta^{18}O$ value 3,000 years before the actual year.

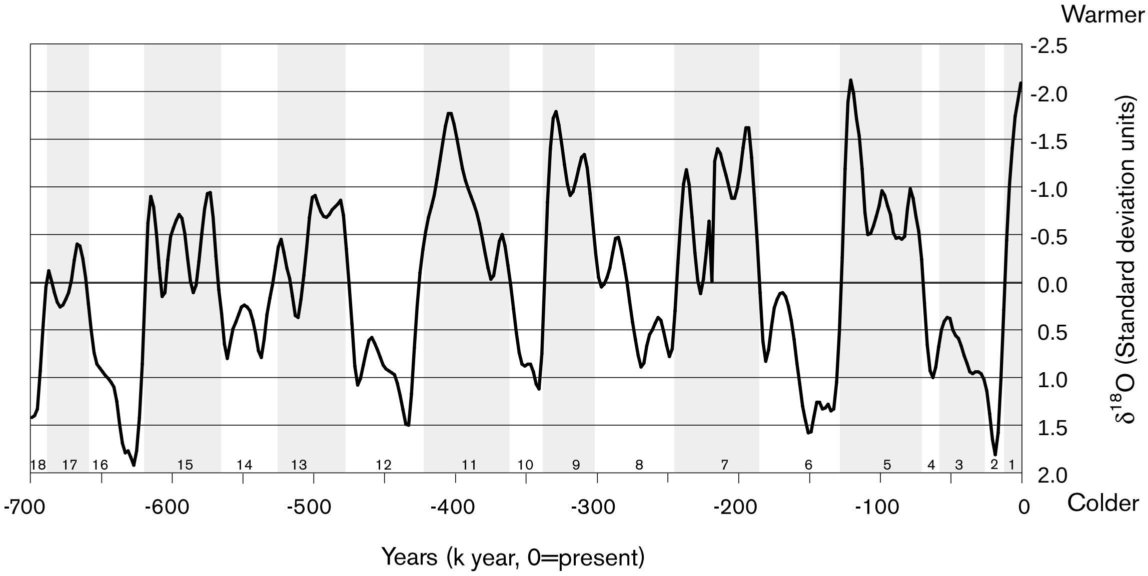

Figure A2-3. *The SPECMAP record: the climate oscillations the last 700 ka as depicted in $\delta^{18}O$ variations /revised from A-10 /.*

2.3.1 Long term hindcasts

The SPECMAP is a record of global ice volume, and although we expect the periodicities of local climate change to reflect global periodicities, we expect that amplitudes of change will vary strongly from place to place. SPECMAP cannot therefore be used directly as a reflection of European climate. This can however be done through the intermediary of local European palaeoclimate records.

An excellent record of local, European climate change has been established for the last glacial cycle from three Maar lake sites in central and eastern France /Guiot et al 1989; De Beaulieu et al 1991; Guiot et al 1992/. The Maar lake records depict variations in temperature and precipitation derived from analyses of the pollen distributions in the lake sediments. They cover the past 140 ka. When creating long term hindcasts pollen-derived temperature and precipitation records from the three Maar lake sites (Lac du Bouchet, Les Echets and La Grande Pile) were used as a proxy for European climate. In addition an averaged, regional pattern of climate change has been constructed, which is smoother and easier to model than the original data. The Maar lake records were normalised by removing the mean value and dividing by their standard deviation. The variations, given as number of standard deviations, were then averaged.

The shorter time span of the Maar records makes it statistically unacceptable to correlate them with orbital parameters, e.g. the periodicity of eccentricity is comparable to the total time span of the record. It is however possible to make simple linear correlations between the shorter Maar lake pollen records and overlapping part of the longer SPECMAP record.

Linear correlations were constructed between the shorter Maar lake pollen records and the longer SPECMAP record. The regression variables were the Maar lake records. The orbital parameters were included as explanatory variables based on the results of Kukla et al. and Berger et al. /Kukla et al 1981; Berger et al 1981/. The values of the orbital parameters were calculated using data from /Berger et al 1991/. Three SPECMAP terms were also included in the regression. This gave the following set of explanatory variables (t = an arbitrary time):

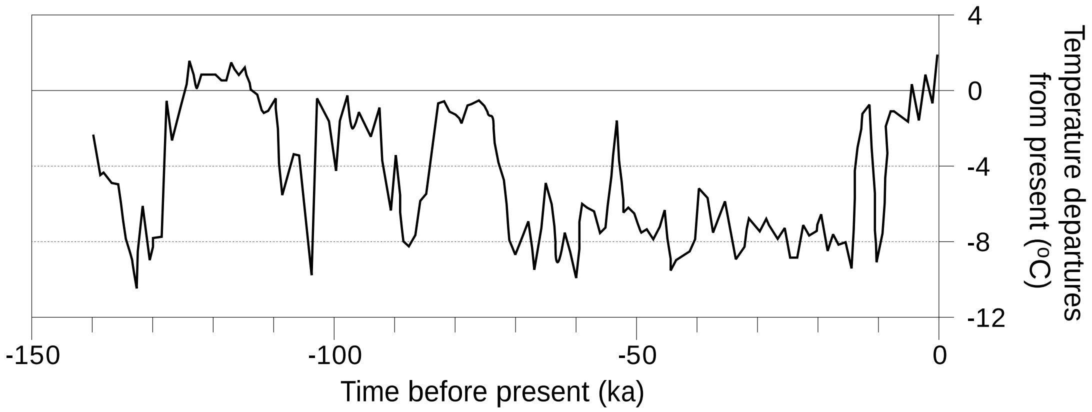

Figure A2-4. *Temperature variations at three French sites, derived from pollen distributions in lake sediments.*

x_1 = $|(\omega_t - 180)/90|$, the precession variable from ACLIN

x_2 = $\varepsilon_{t-6000} - 22$, the obliquity variable from ACLIN lagged by 6,000 years

x_3 = e^2, the eccentricity variable from ACLIN

x_4 = July insolation at 65°N in Wm^{-2}

x_5 = $SPECMAP_t$, the SPECMAP value at the time t

x_6 = $SPECMAP_{t-1000}$, the SPECMAP value at the time t – 1,000 years

x_7 = $SPECMAP_{t-2000}$, the SPECMAP value at the time t – 2,000 years

By using the SPECMAP record in this way, as the spine for our regression, a reasonable linear model of the non-linear climate changes can be produced. A backward elimination technique was used to eliminate unnecessary variables. The regression parameters for the temperature models are presented in table A2-1 and a comparison between the models and the records is shown in Figure A2-5.

Table 2-1. Regression parameters from the correlation between the Maar lake records, SPECMAP and the orbital parameters.

Variable	Grande Pile	Bouchet	Les Echets	Composite
Precession	2.29	26.6	14.8	0.258
Obliquity		14.2		0.125
Eccentricity	-5832	25247	9013	159
July insolation	-0.354	-0.551	-0.576	-0.00394
$SPECMAP_t$	-9.6	-124	-49.4	-0.589
$SPECMAP_{t-1000}$	-36.6	113		
$SPECMAP_{t-2000}$	24.2		28.8	
Constant	107	199	163	1.13
Regression coefficient, R^2	60.5	68.1	69.0	74.2

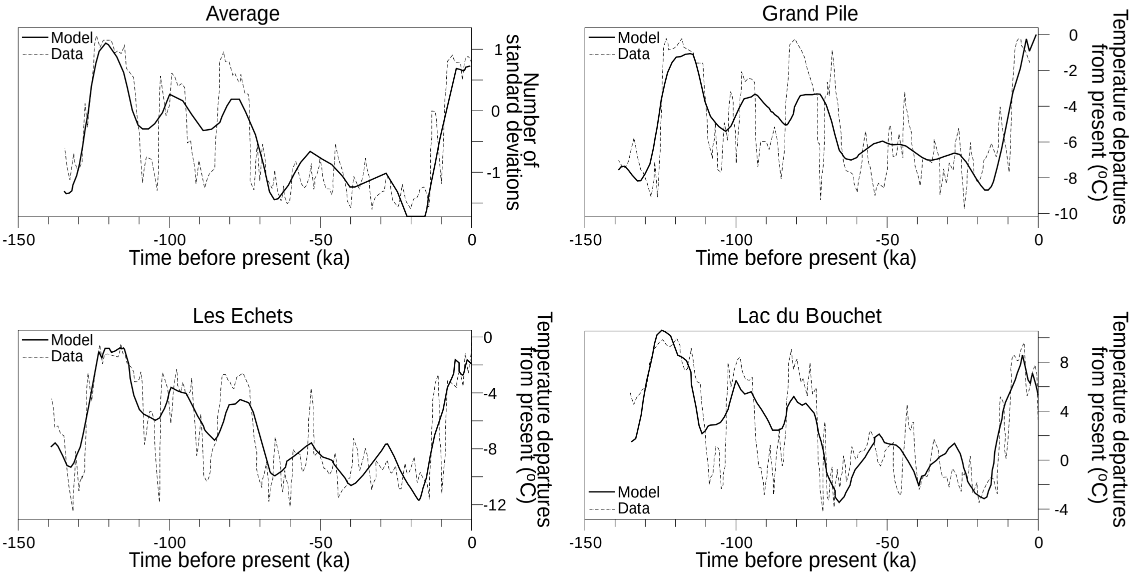

Figure A2-5. *The climate variations according to the models compared to the palaeo-temperature records.*

The models (Figure 2-5) do not capture short lived changes such as the Younger Dryas event (about 12 ka ago) or the full scale of temperature variation during the early Weichsel (about 115 – 70 ka ago, isotope stages 5a – 5d). The periodicity of change however is well reflected.

The model can now be used to hindcast the French climate variations back to 700 ka before present. Unfortunately the French maar lake sequences only cover a single glacial cycle. The model cannot be tested against a period not included in the regression. Further drawbacks are that the used linear regression technique does not permit a probabilistic envelope to be drawn around the hindcast series, and that neither the SPECMAP nor the French records include estimations of errors.

2.3.2 Forecasts

The orbital parameters can be computed for the future time. Consequently with access to a future SPECMAP record the above models could be used to forecast European climate as represented by the Maar lake record. To obtain a relation between the SPECMAP record and the orbital parameters linear regression was carried out between the d^{18}O variations and the orbital parameters. The choice of explanatory variables was based on /Kukla et al 1981; Berger et al 1981/ they were (t = an arbitrary time):

$x_1 = |(\omega_t - 180)/90|$, the precession variable from ACLIN

$x_2 = \varepsilon_{t-6000} - 22$, the obliquity variable from ACLIN lagged by 6,000 years

$x_3 = e^2$, the eccentricity variable from ACLIN

x_4 = July insolation at 65°N in Wm^{-2}

$x_5 = \text{SPECMAP}_{t-3000}$, the SPECMAP value at the time t – 3,000

The model found was:

$$O = 5.61 + 0.103x_1 - 0.035x_2 - 43.9x_3 - 0.0128x_4 + 0.934x_5$$

where O is the modelled normalised $\delta^{18}O$ variation. The regression coefficient value (R^2 = 94.5 %) indicates strong correlation between modelled and real values. The regression was derived from the period 400 – 100 ka before present. As a test it was then used to predict the last 100 ka, which it does very satisfactorily, Figure A2-6. Based on calculations of future orbital parameters a future SPECMAP series was generated. Note that one of the regression variables (x_5 = $SPECMAP_{t-3000}$) must be estimated in the extrapolation into the future. The modelled future SPECMAP values was then used together with the Maar lake models to generate future Maar lake temperatures.

2.4 Production of a Scandinavian climate record

In order to drive the ice sheet model a time series of Scandinavian temperature changes is required. Differences in January and July temperatures between the French Maar records and southern Sweden during glacial, interstadial and interglacial periods are derived from the Atlas of Palaeoclimates /Frenzel et al 1992/. The Maar lake palaeotemperatures were transformed into a southern Swedish record by applying mean annual temperature differences between central France and southern Sweden for glacial, interstadial and interglacial periods from the Atlas of Palaeoclimates. The resultant palaeotemperature curve for southern Sweden is shown in Figure A2-7 together with the temperature hindcasts from the Maar lakes.

The temperature hindcast can now be used together with the assumed linear relation between ELA and palaeotemperatures (equation A-1) to generate an ELA input to the ice sheet model. By applying different values of the constants f_1 and f_2 the modelled ice sheet fluctuations can be matched against geological evidences of ice sheet extension. In many studies the ice sheet model is matched against the last glacial maximum.

Scandinavian ELA variations can also be derived from the inverse modelling described in section 2.2. The calculated ELA fluctuations can be regarded as an estimate of climate conditions in the vicinity of the modelled ice sheet flowline. If there is a relation between the climate in France and the climate in the vicinity of the Scandinavian ice sheet, the above hindcast/forecast of the French climate could be used to generate an estimate of

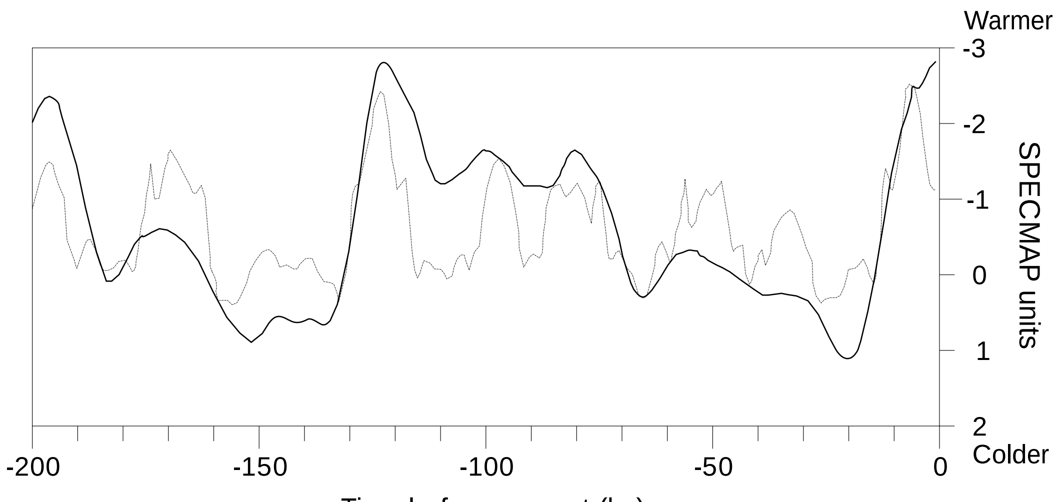

Figure A2-6. *Modelled $\delta^{18}O$-variations in comparison to the SPECMAP record.*

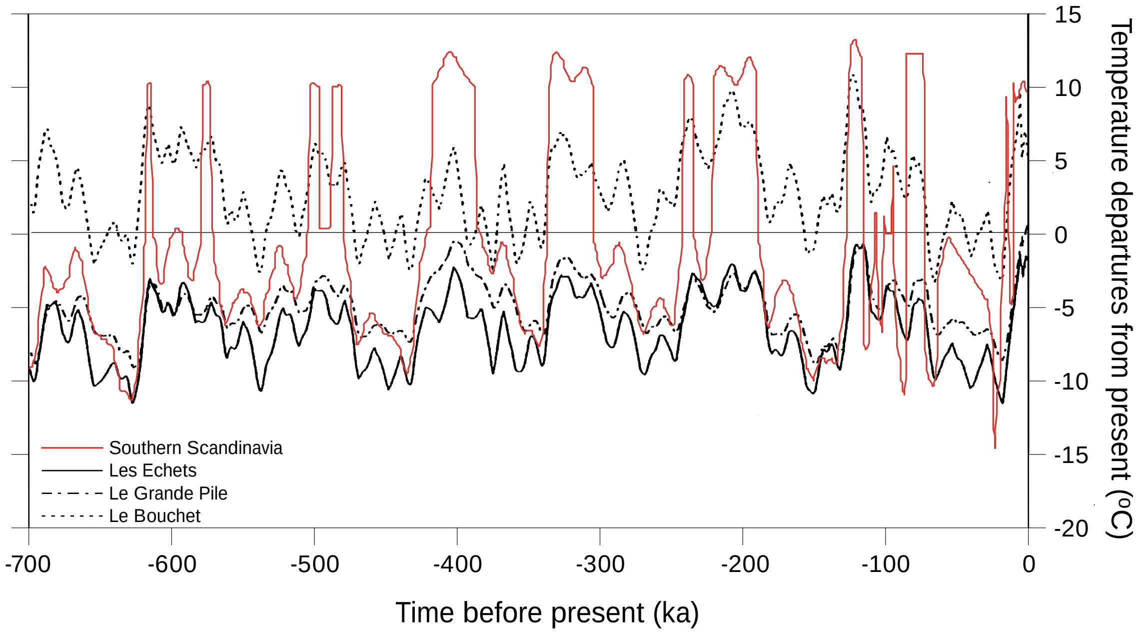

Figure A2-7. *Temperature hindcasts for the French Maar lakes and for southern Scandinavia.*

the Scandinavian climate during the same periods. A statistical correlation between the ELA variations derived from the inverse modelling and the Maar lake data would permit such hindcast/forecast.

ELA variations and Maar lake temperatures during the early Weichsel (isotope stages 5a n 5d) show changes of large amplitude which appear to have a binary form. They either have values equivalent to glacial or to interglacial magnitudes. It is thus difficult to create statistical correlations during this period. The following approach has been adopted to overcome this:

1. Estimate a treshold value in the SPECMAP record below which glacial conditions occur, and above which interglacial conditions occur. The treshold value has been estimated to –0.75 ‰, the approximate mean during isotope stages 5a –5d.
2. Construct a dichotomous time series which has a high value (2) if the SPECMAP value is greater than the treshold and a low value (–2) if the SPECMAP value is less than the treshold. This can be seen as a glacial/temperate indicator curve.
3. Estimate the relative ELA variations by averaging the indicator curve and the SPECMAP curve. Assign an ELA scale (y-axis) by matching key events. In this case the Holocene maximum and the glacial peak 60 – 70 ka before present (stage 4 minimum) were used.

The adapted ELA variations were then correlated to the Maar lake records, so that hindcasts and forecasts of Scandinavian ELA variations could be created. The derived ELA variations are shown in Figure A2-8.

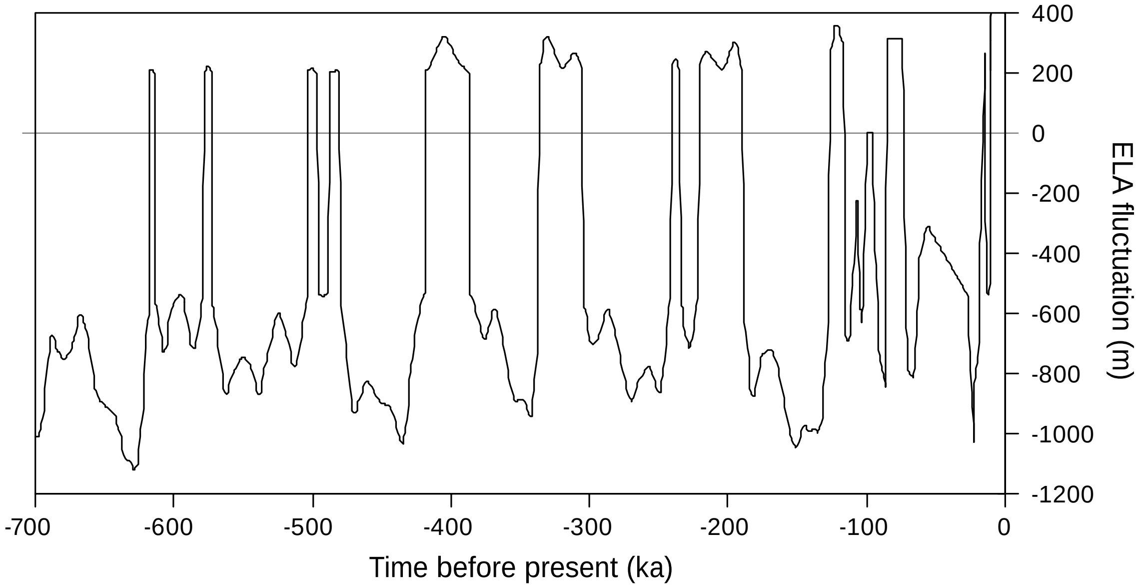

Figure A2-8. *A hindcast of the ELA fluctuations based on the results from the inverse modelling and the Maar lake records.*

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Appendix B

The ice sheet model

1 Introduction

The ice that makes up the Earth's ice sheets is the simplest of the materials, which cover a large part of the Earth's surface. The modern ice sheets of Greenland and Antarctica have predictable surface profiles, internal thermal regimes, laminar modes of flow, and simple large-scale flow patterns that reflect the relatively simple and predictable rheology of the ice that makes them up. We therefore suppose that the European and North American ice sheets of the recent past can be reconstructed using our understanding of the physical laws and theories which have been shown to predict the behaviour of modern ice sheets and other glaciers.

A numerical model of ice sheet behaviour has been utilised to simulate the expansion and decay of these former ice sheets and the associated patterns of change in relative sea level and permafrost thickness. The model is driven by changes in the elevation of the equilibrium line on an ice sheet surface (approximately the same as the permanent snow line, separating an area of accumulation from a lower altitude ablation area) and by air temperature. The ice sheet acts as a conveyor belt that transports ice from the accumulation area, where there is net accumulation of snow and ice, to its ablation area, where it is lost by melting or by calving of icebergs. The rate at which this occurs depends on the flow properties of ice, which are known, and which depend principally on ice temperature, and on the rate of accumulation. A solution of this problem yields the ice sheet surface profile. Flow of the ice also transports cold ice from a high elevation to a low elevation, and therefore changes the distribution of internal temperature. An ice sheet is thus a coupled thermo-mechanical system. The numerical model describes and predicts its behaviour when driven by given atmospheric conditions (temperature and snowline), and for an Earth's surface of given form, mechanical and thermal properties.

The model is driven by variations in the equilibrium line altitude (ELA) and the sea level air temperature (SLAT) and a mass balance pattern (see Appendix A, section 2.1). Mass balance (the difference between total annual snow accumulation and total annual ablation) is specified as a function of the difference between ELA and ice sheet surface elevation. Air temperature at the ice sheet surface is calculated using a constant lapse rate. The equations comprising the model are solved using finite difference techniques.

The model comprises five separate components

- ice sheet form;
- internal velocity field of the ice sheet;
- internal temperature field of the ice sheet;
- temperature field in the underlying bedrock;
- isostatic response of the underlying bedrock.

The overall behaviour of the model is controlled by the interactions between these components (Figure B1-1).

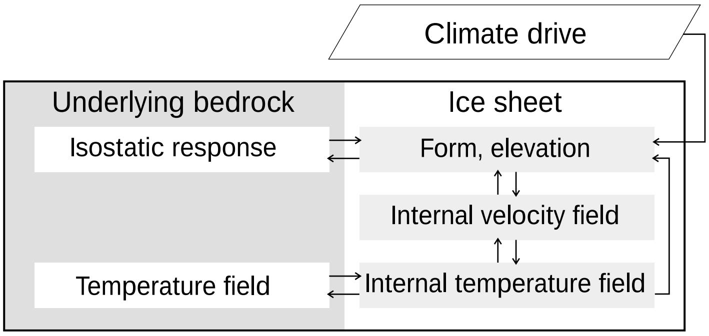

Figure B1-1. *The interactions between the components of the ice sheet model.*

2 Ice sheet form

The complex three-dimensional form of an ice sheet is best described by the distribution of ice thickness (H) over the horizontal extent of the ice sheet. It is a function of time, H(x,t), where x is the horizontal coordinate and t is time (Figure B2-1)

The function H(x,t) is estimated using the continuity equation of ice thickness /Mahaffy 1976/:

$$\frac{\delta H}{\delta t} = -\frac{\delta}{\delta x}(\bar{u}H) + b \qquad \text{(B-1)}$$

The rate of ice thickness change at any point is a function of the local mass balance b(x,t) and the product of local ice thickness and vertically-averaged horizontal ice velocity u(x,t). The latter represents the local convergence/divergence of horizontal flow within the ice sheet.

Ice flow is typically very slow and the involved accelerations are assumed to be insignificant. This implies that the forces within an ice sheet must be in balance, in particular the weight of the overlying ice must be balanced by stresses within the ice mass. Assuming that horizontal stress gradients are negligible /Nye 1959; Budd 1970; Jenssen 1970/ and that the pressure distribution within an ice mass is everywhere hydrostatic /Nye 1959/, then ice flows in simple shear and the only significant stress within the ice mass is the horizontal shear stress t_{xz}. Its direction is parallel to the ice sheet surface slope and its magnitude is:

$$\tau_{xz} = -\rho_i g(H + h - z)\frac{\delta}{\delta x}(H + h) \qquad \text{(B-2)}$$

where r_i = density of ice,
g = the acceleration due to gravity, and
H, h, x, z are defined in figure B-2.

This stress field is then related to the deformation within the ice sheet. The key to this is Glen's flow law /Glen 1955/, an empirical relationship which gives ice deformation as a function of stress.

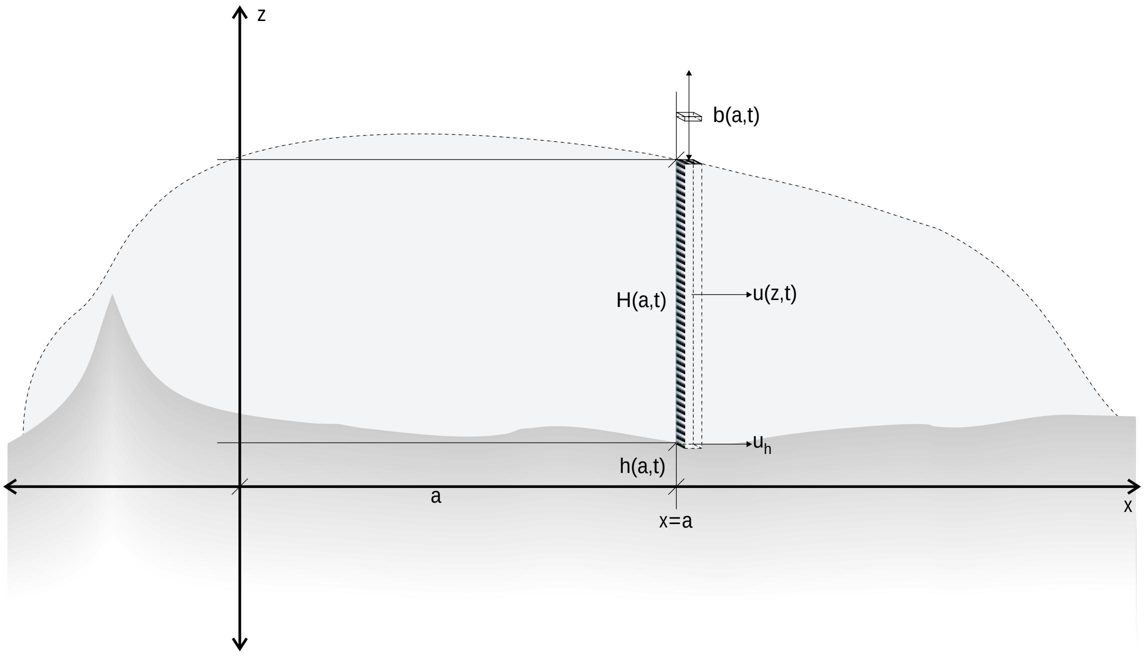

Figure B2-1. *The geometry used in the model.*

Given the simplified stress environment and the assumption that horizontal rates of change are insignificant compared to vertical ones, an expression for horizontal velocity (u(x,z,t)) anywhere within the ice mass can be derived:

$$u = -2(\rho_i g)^n \left|\frac{\delta(H+h)}{\delta x}\right|^{n-1} \left(\frac{\delta(H+h)}{\delta x}\right) \iint_h^H A(T)(H+h-z)^n dz + u_h \tag{B-3}$$

where n = power, usually 3 (Glen's flow law),
A = proportionality factor (Glen's flow law),
T = ice temperature, and
u_h = velocity at the ice/bedrock interface

The proportionality factor (A) is determined from ice temperature corrected for variations in the pressure melting point (T*). An Arrhenius equation is used /Paterson et al 1982/

$$A = a \exp\left\{\frac{-Q}{RT^*}\right\} \tag{B-4}$$

where Q = activation energy for ice creep,
R = the gas constant, and
a = multiplier

The above analysis only considers movement by internal deformation of the ice. Movement can also occur by sliding of basal ice over its bed, either through sliding over a rigid bed /Weertman 1979/ or deformation of the underlying sediment /Boulton et al 1979/. Two versions of the model have been used; one which assumes that there is no sliding and one which incorporates sliding. In the sliding model, the deformation law for a sandy subglacial till derived by Boulton and Hindmarsh /1987/; see also /Boulton et al

1998/ for review) has been used. The till was similar to the sandy tills that cover the basement areas of Sweden. The law is:

$$\dot{\varepsilon} = k\tau_b^n p_e^m \tag{B-5}$$

where t_b = shear stress at the base of the glacier,
e = the shear strain rate,
p_e = the effective pressure, and
n, m = constants.

At present, no measured sliding law for ice over a rigid substratum has been derived, although a law similar to that for deforming sediment has been used by Morland /1984/ successfully to replicate important features of ice sheet behaviour. Although the detailed behaviour of the two models (no sliding or bed deformation) varies in ways illustrated by Boulton, Hulton and Vautravers /1995/, those general characteristics of ice sheet behaviour that are likely to effect hydraulic conditions in the subsurface, are not very different. For the sake of consistency with model output in previous reports (eg /Boulton et al 1992/), we have assumed no bed sliding, even when melting occurs at the glacier bed. The consequences of this omission may not be serious because the coupling between ice sheet form and internal velocity implies that the omission of enhanced basal flow will be compensated by an increase in ice flow via internal deformation. The heat generated by internal deformation will be the same as in a sliding model, but it will not be quite so concentrated at the bed. As a consequence, the model will underestimate melting and, because of the absence of a basal process of easy *decollement* able to discharge a high ice flux with a low shear stress, it will tend to overestimate ice sheet thickness.

The vertical component of velocity is also important because it plays a part in the pattern of heat advection and distribution of temperature through an ice sheet (see the next section). Vertical velocity w(x,z,t) can be found from the calculated horizontal velocity field by invoking the conservation of mass:

$$w = -\int_h^z \frac{\delta u}{\delta x} dz + w_h \tag{B-6}$$

and integrating this equation from the bed to the ice sheet surface. This requires an estimate of vertical velocity at the bed w_h (x,t), which in the absence of sliding is:

$$w_h = \frac{\delta h}{\delta t} - S \tag{B-7}$$

where S is basal melt rate, the calculation of which is outlined in the next section.

3 Internal temperature field

Knowledge of the changing internal temperature field within the ice sheet is required because of the dependence of ice flow on temperature (equation B-4). The basis for the temperature model is the conservation of energy :

$$\frac{\delta T}{\delta t} = \frac{k_i}{\rho_i c_i}\frac{\delta^2 T}{\delta z^2} - u\frac{\delta T}{\delta x} - w\frac{\delta T}{\delta z} - \frac{g}{c_i}(H+h-z)\frac{\delta u}{\delta z}\frac{\delta}{\delta x}(H+h) \tag{B-8}$$

where c_i = specific heat capacity of ice, and
k_i = thermal conductivity of ice.

The first two terms describe heat diffusion and advection. Advection brings cold ice which formed on the ice sheet summit dome down into the body of the ice sheet, thereby cooling the ice mass as a whole. The third term describes heating produced by deformation, which produces a feedback between ice flow and temperature. Fast flowing ice produces more strain heating than slow flowing ice, and the dependence of ice deformation on temperature causes the warmer ice to flow still faster, thus generating a positive feedback /Clarke et al 1977/. This feedback is countered by the fact that faster flowing ice increases the rate of cold ice advection, which opposes the warming inherent in strain heating /Huybrechts et al 1988/.

The production of meltwater is a second source of stability within the temperature model. The melting point of ice is affected by pressure:

$$T^* = -\phi \rho_i g(H + h - z) \qquad \text{(B-9)}$$

where f is a constant.

When ice attains the pressure melting point, melting occurs. The amount of melting depends on the excess heat available /Budd et al 1971/

$$S = \frac{k_i}{L\rho_i}\left((\frac{\delta T}{\delta z})_c - (\frac{\delta T}{\delta z})_b \right) \qquad \text{(B-10)}$$

where L is the latent heat capacity of ice

Subscripts c and b refer to the temperature gradient after and before the correction for pressure melting has been made. This equation converts the excess heat available into a melt rate.

In order to integrate equation B-8 forward through time, boundary conditions, which reflect the conditions at the ice sheet surface and at the ice/bed interface, are required. The former depends on mean annual temperatures and the latter depends on the geothermal heat flux.

The geothermal heat flux, G, can be evaluated as:

$$G = k_r \frac{\delta T_r}{\delta z} \qquad \text{(B-11)}$$

where k_r = thermal conductivity of rock, and
T_r = rock temperature.

If the vertical gradient of temperature within the underlying bedrock is assumed constant at 0.013 °C/m, the basal ice is subject to a flux of 42 mW/m^2 /Haenel 1980/ which will melt ice at pressure melting point.

4 Temperature field in the underlying bed

The geothermal heat flux is determined by the vertical temperature gradient within the bed underlying the ice sheet. If the distribution of temperature within the bed changes, the geothermal heat flux must also change. The growth of an ice sheet insulates the underlying bed and causes local warming, which in turn reduces the local geothermal heat flux. The removal of the ice sheet results in an increase of the geothermal heat flux.

This process can be modelled by extending the model's temperature calculations into the bed. Work by Ritz /1987/ indicates that the top 2 km of the bed can be considered active, in the sense that its temperature evolution interacts with that of the ice sheet.

The processes governing heat flow within the bed are modelled in a simpler way than those governing heat flow within ice sheets. The vertical diffusion term is the only significant one, giving the equation for evolving bedrock temperature T_r (x,z,t) as:

$$\frac{\delta T_r}{\delta t} = \frac{k_r}{\rho_m c_r} \frac{\delta^2 T}{\delta z^2} \qquad \text{(B-12)}$$

where c_r = specific heat capacity of rock, and
r_m = density of rock material.

Equation B-12 also represents a first order approximation to the evolution of permafrost. The depth of permafrost is given by the depth of the 0 °C isotherm. The model is a first order approximation because it only considers the sensible heat budget of the bed and does not include associated latent heat fluxes.

5 Isostatic response of the underlying lithosphere

Isostatic flexure of the earth beneath an ice load is frequently analysed as a two-layer model of an elastic lithospheric plate whose properties control the shape of flexural response, overlying a viscous astenosphere whose properties control the rate of flexure. The lithosphere sinks beneath the ice sheet and is upwarped in a 'bulge' beyond it. At equilibrium, the volumetric depression of the bed is equal to the ice volume multiplied by the ratio of the asthenosphere density to ice density (approximately 0.27).

The most important feature of asthenospheric compensation is the long time period required for equilibrium, which is a reflection of the high viscosity of the asthenosphere. It has two effects. Firstly, the equilibrium depression under an ice loading is seldom attained because the residence time of an ice sheet is of the same order of magnitude as the time required for equilibrium to be established /Mörner 1971/. Secondly, asthenospheric material moving away from an area of ice loading does so very slowly and can result in transient upwarping of areas immediately beyond the ice sheet /Walcott 1970/. This leads to the formation of a forebulge, which has a different origin and behaviour to the forebulge produced by the elastic lithosphere.

A coupled isostatic model is used in which only asthenospheric flow is analysed, as the effects of lithosphere rigidity are felt mainly in the adjacent, forebulge areas. This is justified as we are primarily concerned with the ice covered area, and the general role of isostasy in ice sheet dynamics. A diffusion equation can be used to model the effect of asthenospheric flow on bedrock elevation, h(x,t):

$$\frac{\delta h}{\delta t} = D_a \frac{\delta^2}{\delta x^2}\left(h - h_0 + \frac{\rho_i}{\rho_m} H \right) \qquad \text{(B-13)}$$

where D_a = diffusivity of the asthenosphere, and
h_0 (x) = relaxed bedrock elevation.

This equation allows bedrock elevation to change as a consequence of the balance between the buoyancy of the lithosphere (the difference between relaxed and current bedrock elevation) and the imposed load (as a bedrock height equivalent). The rate at which an equilibrium is approached depends on the numerical value of the viscosity. A value of 100 km^2 /year implies that 22 500 years are needed before equilibrium is established /Huybrechts 1986/.

6 Boundary conditions for the model

The model requires four principal boundary conditions:

- mass balance boundary conditions,
- thermal boundary conditions,
- topographic and isostatic boundary conditions, and
- marine boundary conditions.

6.1 Mass balance boundary conditions

The surface of an ice sheet is divided into an accumulation area where there is net accumulation of snow and ice through the year, and an ablation area where there is a net mass loss. An equilibrium line where ablation balances accumulation separates these areas. The mass balance is a function of the difference between ice sheet elevation and equilibrium line altitude (ELA). This is described in Appendix A.

The coupling of ice thickness evolution to mass balance in the continuity equation (equation B-1) and the relation of mass balance to ice sheet surface elevation implies that a thickening ice sheet will experience progressively greater positive mass balances and its thickening rate will accelerate through time.

This is indeed a very strong feedback, which is only partially compensated in equation B-1 by the flux divergence/convergence term. The latter will tend to reduce the rate of thickening at a point by increasing the outflow of ice from that point. It is to be expected that ice sheet growth and decay rates will be very sensitive to changes in mass balance and imposed ELA.

6.2 Thermal boundary conditions

Thermal boundary conditions are required for top and bottom surfaces. The geothermal heat flux of 42 mW/m^2 is applied to the base of the 2 km thick lithosphere slab underlying the ice sheet.

The surface air temperature has a given value, and contrasts with the lower boundary condition, which is a flux term and is related to the flow of heat across that boundary. When ice is not present, this boundary condition is applied to the top of the modelled bedrock slab. In this event it is assumed that the temperature of the ground surface is 3 °C warmer than the air temperature (cf /Brown 1970/). The evolution of temperature at the ice sheet surface is described in Appendix A.

6.3 Topographic and isostatic boundary conditions

A representation of bed elevation along the flowline is required for three reasons:

(a) the mass balance and air temperature calculations use ice sheet surface elevation;

(b) the ice velocity calculation uses ice sheet surface slope as input;

(c) the marine boundary condition uses a bathymetric limit on ice sheet extent.

(a) and (b) are only important when ice sheet thickness is small, during initial growth and final decay. During most of the ice sheet's existence the great thickness of ice masks most of the initial topographic variation. (c) is important during the complete sequence of growth and decay.

A relaxed topography is required in the isostatic model (equation B-13). This represents topography along the flowline after recovery (uplift) from ice sheet loading is complete. The difference between this relaxed topography and the topography at any time during a model run can be used as a measure of the buoyancy of the bedrock. The present day topography along the flowline is used as the relaxed topography. The errors consequent upon this are probably small, for two reasons. Firstly, although southern Sweden is still rising in response to the removal of the last ice load, the rate of uplift is relatively slow compared to past rates (0.5 mm/year on average /Ekman 1991/) indicating that the area may be near isostatic equilibrium. Secondly, variations in ice thickness are expected to be both large (in excess of 2 km) and rapid (deglaciation occurred within 10 kyears /Boulton et al 1985/). Errors in the buoyancy calculation (relaxed minus current bedrock elevation) of equation B-13 are therefore minor in comparison to the fluctuations occurring in the ice load term, and a good estimate of the rate of isostatic response can therefore be expected.

6.4 Marine boundary conditions

Where an ice sheet flows into the ocean it may float locally. When this occurs the stress regime within the ice sheet ceases to be shear dominated and becomes dominated by longitudinal stresses (e g /Herterich 1987/). The complex transition between the two regimes occurs near the grounding line and has been discussed by many authors, e g /van der Veen 1987/, /Alley et al 1984/ and /Muszynski et al 1987/.

As we principally are concerned with the behaviour of the ice sheet far from the ocean, a very simple scheme is used at the marine/floating boundary. Geological evidence indicates that the western margin of the ice sheet never extended past the continental shelf /Lehman et al 1991/, and probably ended in a short shelf of floating ice. This is incorporated into the model by allowing the ice sheet to extend into water depths of up to 500 m depth but no further. Any ice flow past this point is assumed to be lost by the calving of icebergs. A boundary condition of this type is often used in models of predominantly land based ice sheets (eg /Oerlemans 1981/).

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Results of glaciation modelling

1 Introduction

The glaciation model described in Appendix B has been applied to simulate the last 700,000 and the next 200,000 years of climatically determined environmental change in Europe. The model has been applied to the flowline transect shown in Figure C1-1.

A flowline model poses problems in simulating ice sheet flow over the complex high relief topography of a high mountain area such as that of the mountain chain along the Swedish-Norwegian border. In reality, ice sheet flow across such an area is highly three-dimensional. High mountain ridges tend to block flow whilst deep valleys facilitate it. The transect which we have used crosses the mountain range at a relatively high eleva-

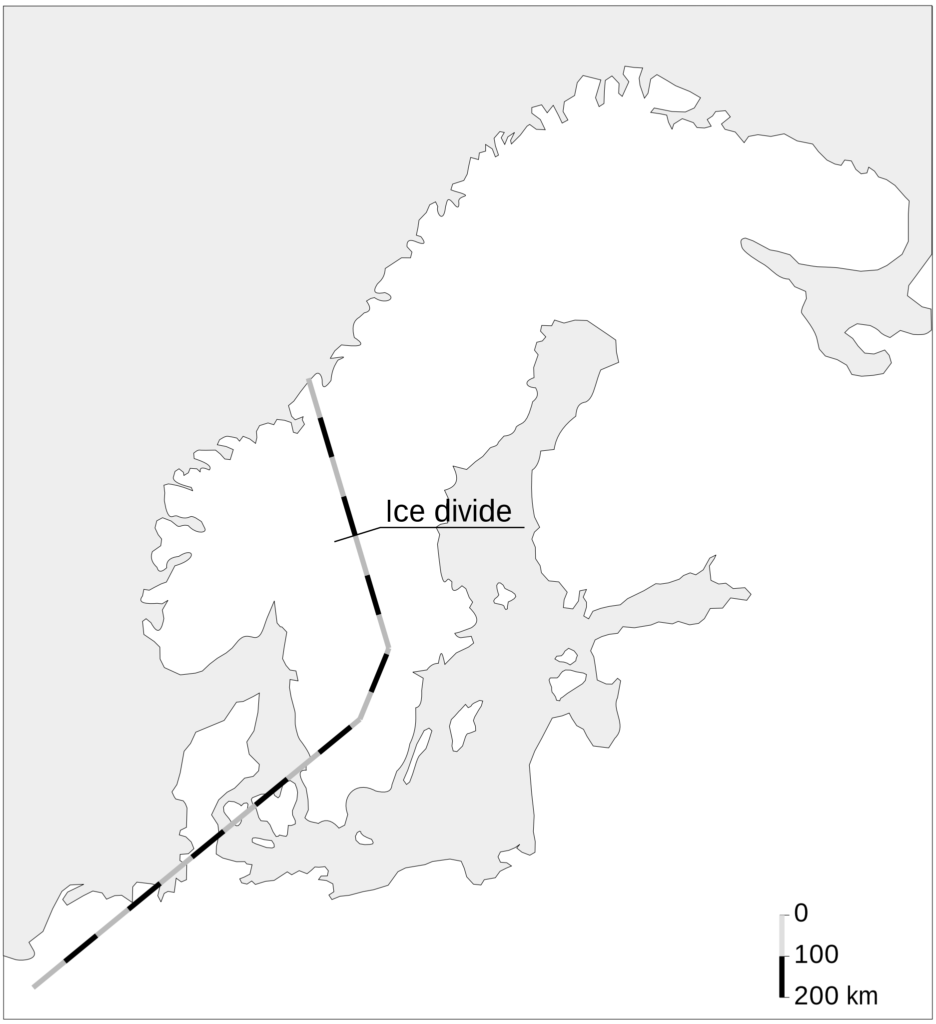

Figure C1-1. *The flowline transect used in the modelling.*

tion. As a consequence, it acts as a greater barrier to flow than it would if valleys could be included, with the result that the extension of the ice sheet to the west is much less than it would otherwise be (in its simulated Weichselian and Saalian extents for example). However, as our concern is primarily with the low relief areas east of the mountain chain, we believe that a two-dimensional approximation produces useful results, and has the benefit of simplicity in showing variations in both time and space.

The results of modelling can be presented in three ways:

- as properties which vary continuously in time along a transect;
- as a series of discontinuous domains and regimes, in which properties are constant within any one regime but change sharply at the boundary between regimes, both along the transect and through time;
- as continuous or discontinuous time sequences at a given location.

2 Simulations of the past 700,000 years

2.1 Continuous variation in time along the transect

It is difficult to force a flowline model of an ice sheet with a realistic climate model, and so we have used a series of climate parameters, these are;

- sea level air temperature (SLAT),
- equilibrium line altitude (ELA) and
- vertical mass balance distribution.

(Also see Appendix A: Climate and climate change) The hindcasts of SLAT and ELA, which have been used, are shown in Figure C2-1 and Figure C2-2 respectively, and the range of patterns of mass balance in Figure C2-3. The latter vary from distributions typical of highly continental environments to those typical of highly maritime environments.

There are a large number of potential outcomes from a model driven by three sets of variables. The deriving of temperature hindcasts are accounted for in Appendix A: Climate and climate change. It is difficult to estimate the errors in the hindcast of SLAT. The ELA and the mass balance distributions depend on vertical atmospheric temperature and moisture variations that are likely to vary strongly. The hindcast and forecast simulations presented here are not therefore comprehensive reflections of all possible outcomes, nor are they intended as precise reconstructions to match the geological evidence of ice sheet fluctuation in Europe. They are intended as illustrations of the properties of ice sheets in Europe which show patterns of temporal fluctuation similar to those indicated by the fragmentary geological record.

In the following the results from four different simulations are presented and discussed. The designations of the simulations and the used climate data are presented in Table C2-1.

The fluctuations of the ice sheet margin along the transect for simulations M and O are shown in Figure C2-4. The output from simulation M and O is presented in Figure C2-5 and Figure C2-6 show plots of ice sheet thickness and basal temperature through the last 200 ka along the extended transect for simulations M and O.

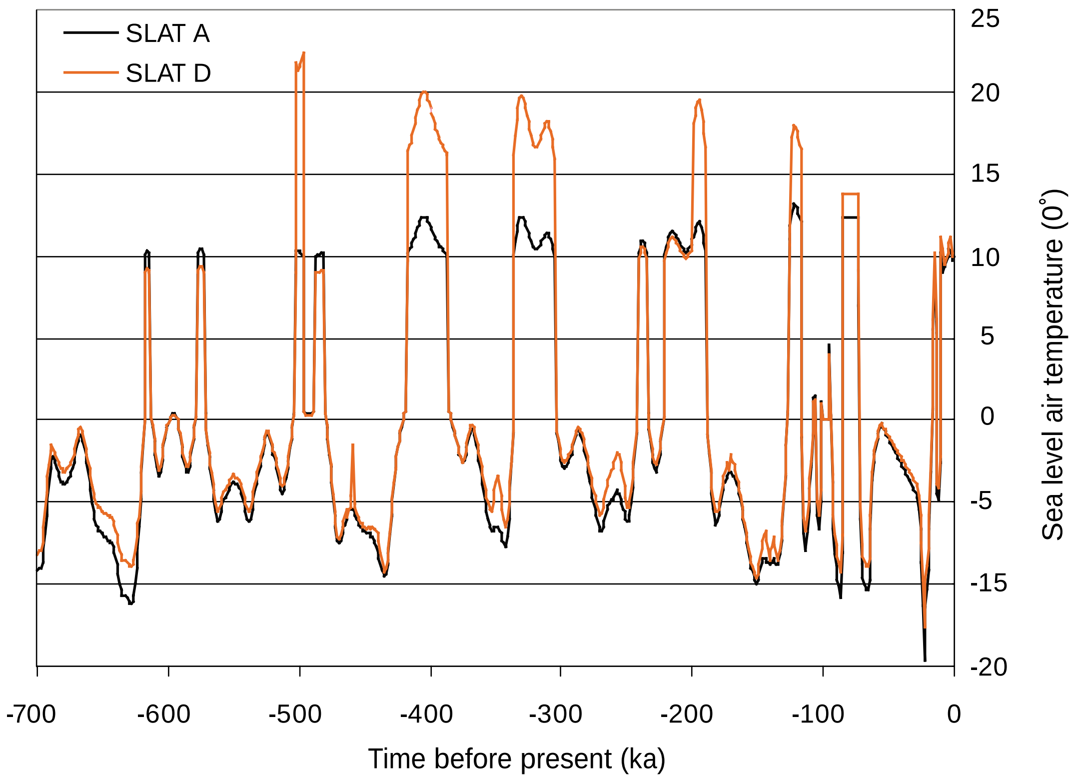

Figure C2-1. *Hindcast sea level air temperatures (SLAT) for the southern Baltic area during the last 700,000 years.*

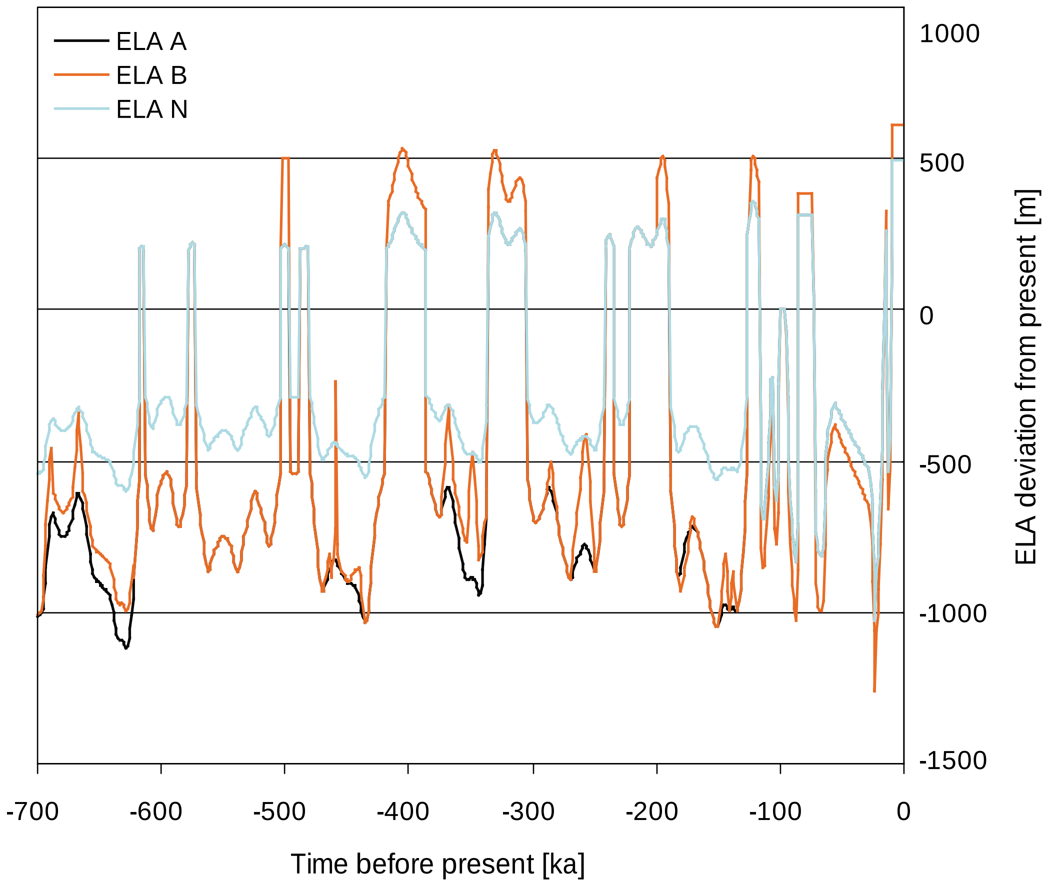

Figure C2-2. *Hindcasts of equilibrium line altitude (ELA) deviations from modern values.*

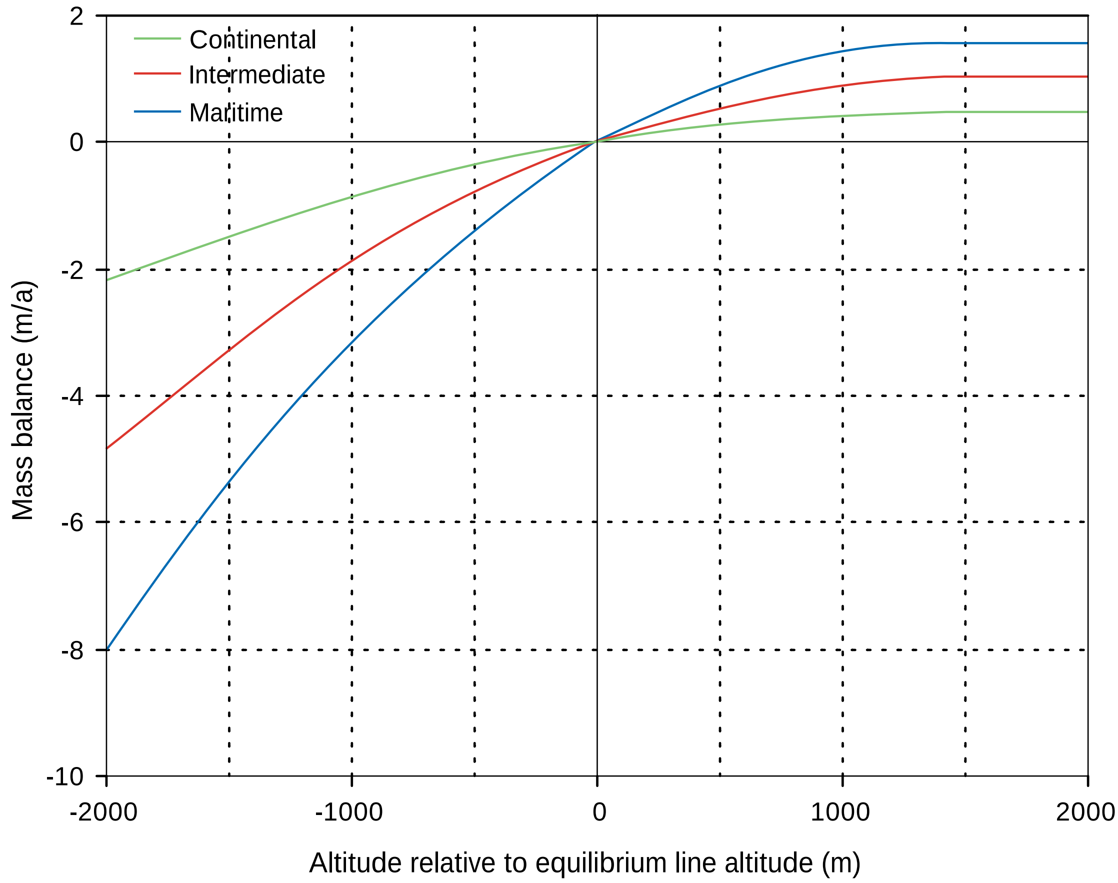

Figure C2-3. *The three empirically based mass balance patterns used in the simulations.*

Table C2-1. Designations of simulations and climate input

Simulation	SLAT	ELA	Mass balance pattern
M	A	A	Most maritime from 700,000 to 116,000 years ago, Most continental from 116,000 years ago
O	D	N	Most maritime from 700,000 to 116,000 years ago, Most continental from 116,000 years ago
S	A	B	Intermediate
N	A	Best fit ~ A	Intermediate

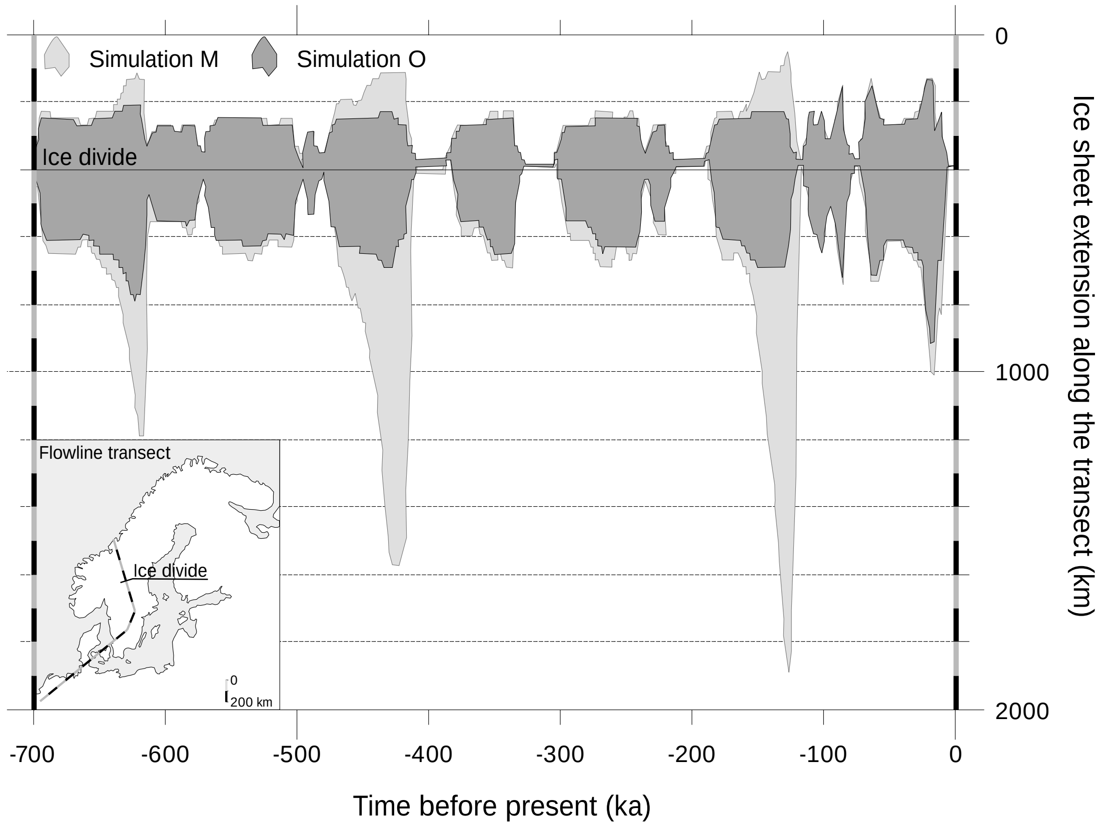

Figure C2-4. *Ice sheet variations along the transect for simulations M and O.*

The contrasts between M and O prior to 116ka reflect the impact of different amounts of ELA lowering in sustaining ice sheet growth. In simulation M, the ELA lowering is enough to drive the ice sheet to continue to expand after it has extended into the lowlands beyond the mountain region where it formed. After 116ka, the drier, more continental mass balance produces a more slowly growing ice sheet, which is forced to decay by warming before it has had an opportunity to grow to a larger size. In simulation O, a slightly higher SLAT after 116ka, ensures that the ice sheet is slightly less extensive than in the case of M. Apart from during the last glacial cycle, simulation M provides a reasonably good match with the overall patterns of maximum ice sheet extent during the last 700ka, with maxima at 130ka (Saalian), 420ka (Elsterian) and 620ka.

The interaction between climatic conditions and the form of an ice sheet surface, related to mass balance, flow and temperature, sensitively determine patterns of ice sheet growth and the potential for growth to a large size. This sensitivity is particularly marked when an ice sheet extends far from the mountain front, where a bifurcation in behaviour is often apparent. This can be seen in simulation M where the contrasts in ice extent during some periods are great even if contrasts in SLAT and ELA are not very great.

The simulations show that the ice divide does not move far from the Scandinavian Mountains during glacial cycles. In order to show variation in Sweden with more fidelity, it is convenient to fix the position of the divide and to follow ice sheet variation only to the east of the divide. Case S contains a simulation of the ice extent to the south east of the ice divide. It was set up to provide the best match with geologically inferred patterns of ice sheet extent.

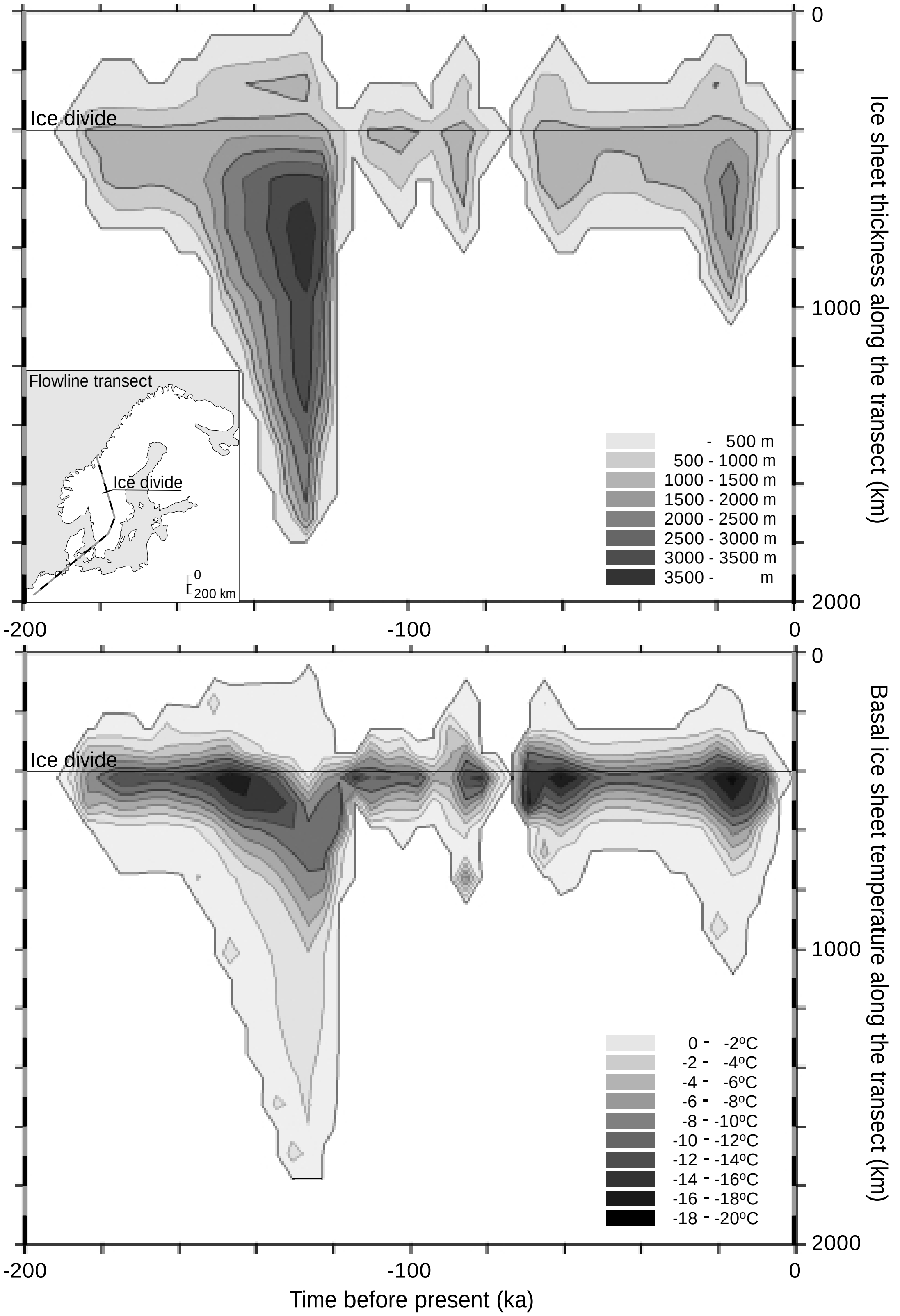

Figure C2-5. *The variation of ice sheet thickness and basal ice sheet temperature along the transect over the last 200,000 years in run M.*

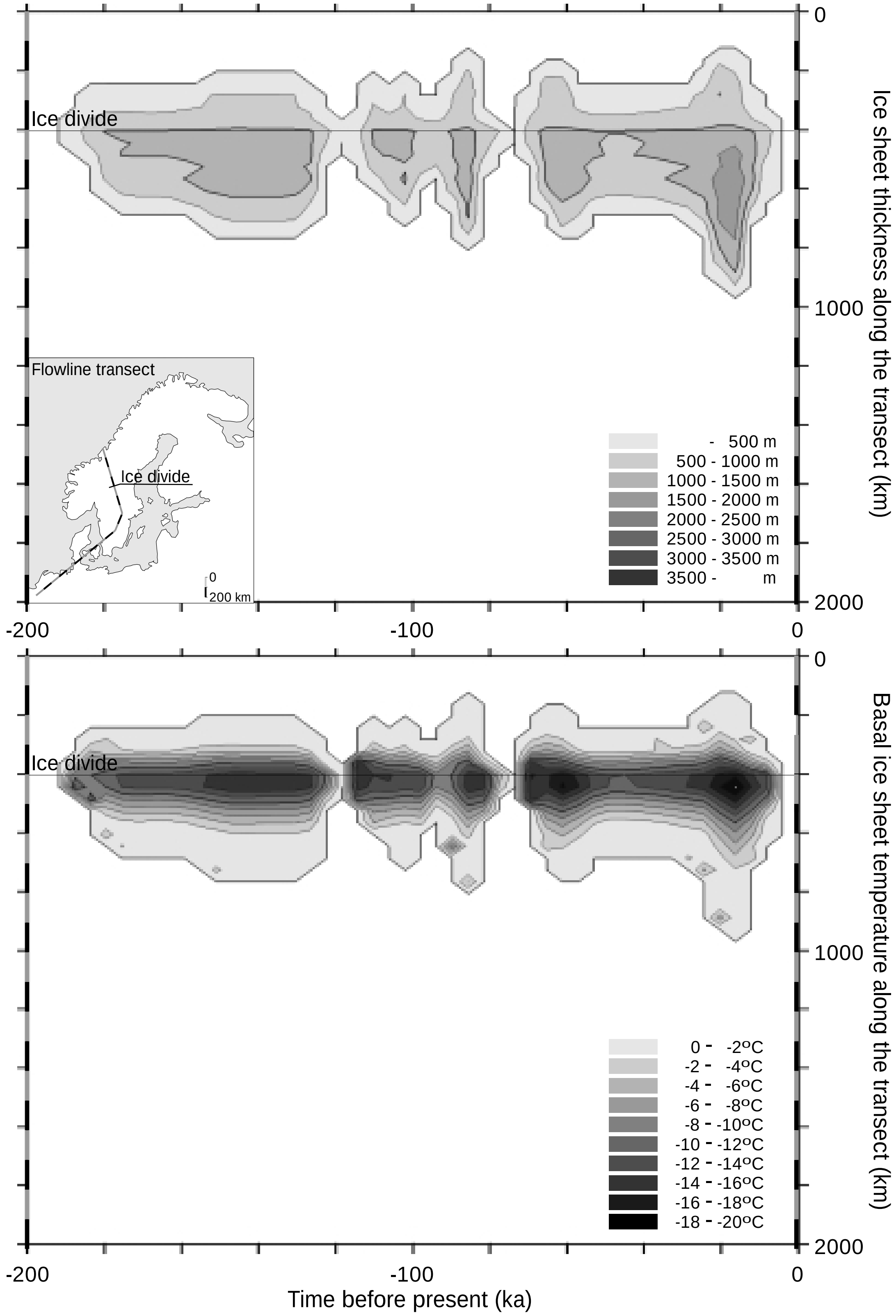

Figure C2-6. *The variation of ice sheet thickness and basal ice sheet temperature along the transect over the last 200,000 years in run O.*

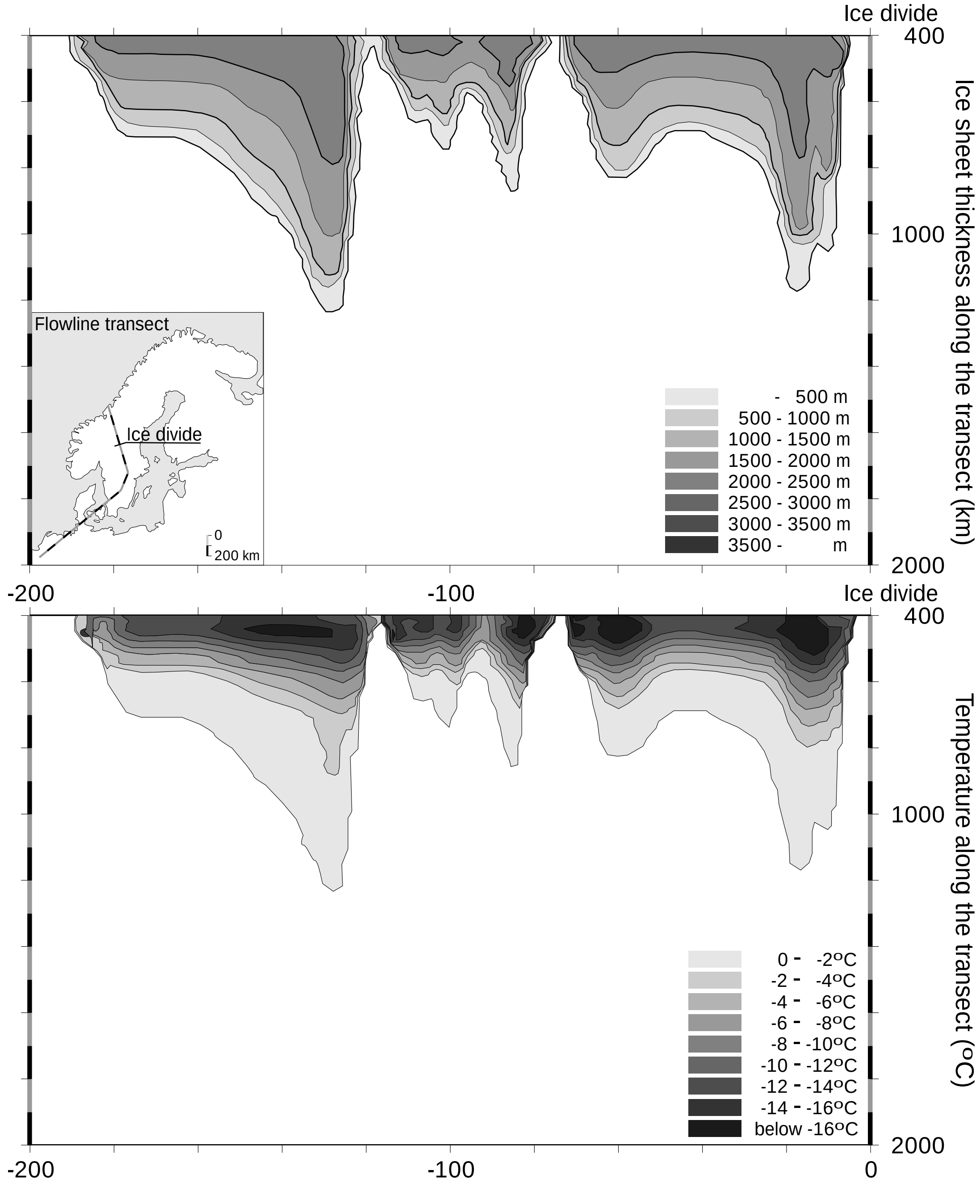

Figure C2-7. *The run S reconstruction of ice sheet thickness and basal temperatures along the transect through the last 200,000 years.*

Simulation S produces ice sheets whose extent along the transect during Weichselian and Saalian times are similar to those inferred from geology while the Elsterian extent (420 ka) is very much less than geological evidence indicates. This is because the smaller mass balance causes a slower rate of ice sheet advance in the model. Warming conditions cause the modelled ice sheet to retreat before it can achieve the extent of the analogous ice sheet in simulation M. It is possible of course that basic patterns of continentality have changed through time for similar stages of ice sheet extent. On timescales of 100,000 years these changes may be related to global tectonic changes.

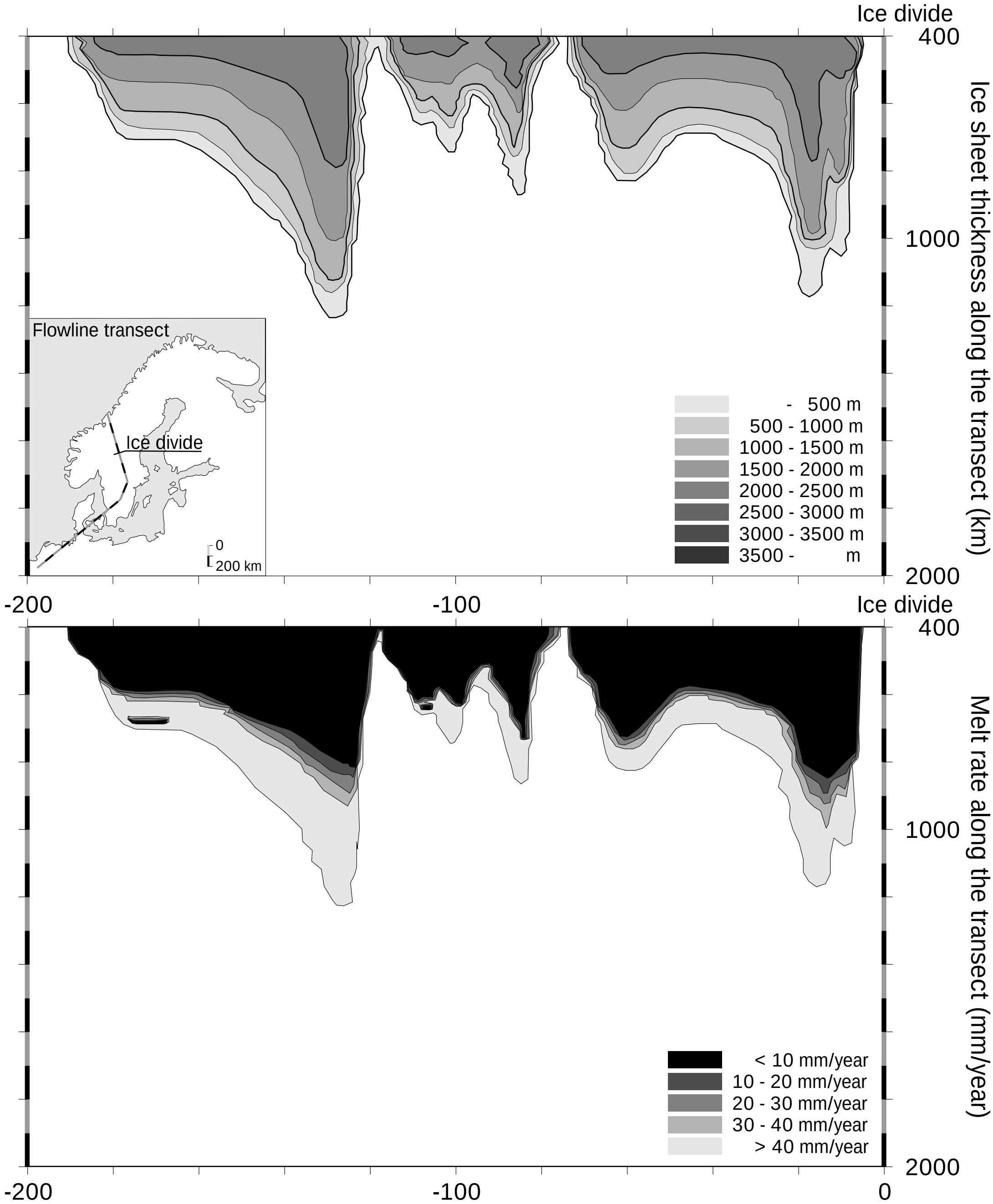

Figure C2-8. *The run S reconstruction of ice sheet thickness and melting rates at the base of the ice sheet along the transect through the last 200,000 years.*

In order to produce the best fit with geological evidence, a simulation N has been created in which the same SLAT and mass balance conditions used in simulation S have been used but the ELA has been permitted to change so as to create a best fit.

2.2 Domains and regimes in space and time along the transect

The patterns of continuous variation along the transect can be arbitrarily subdivided into a limited number of individual domains and regimes using the distinctions described in chapters 3 and 4. This has been done for run N in Figure C2-9, for which permafrost

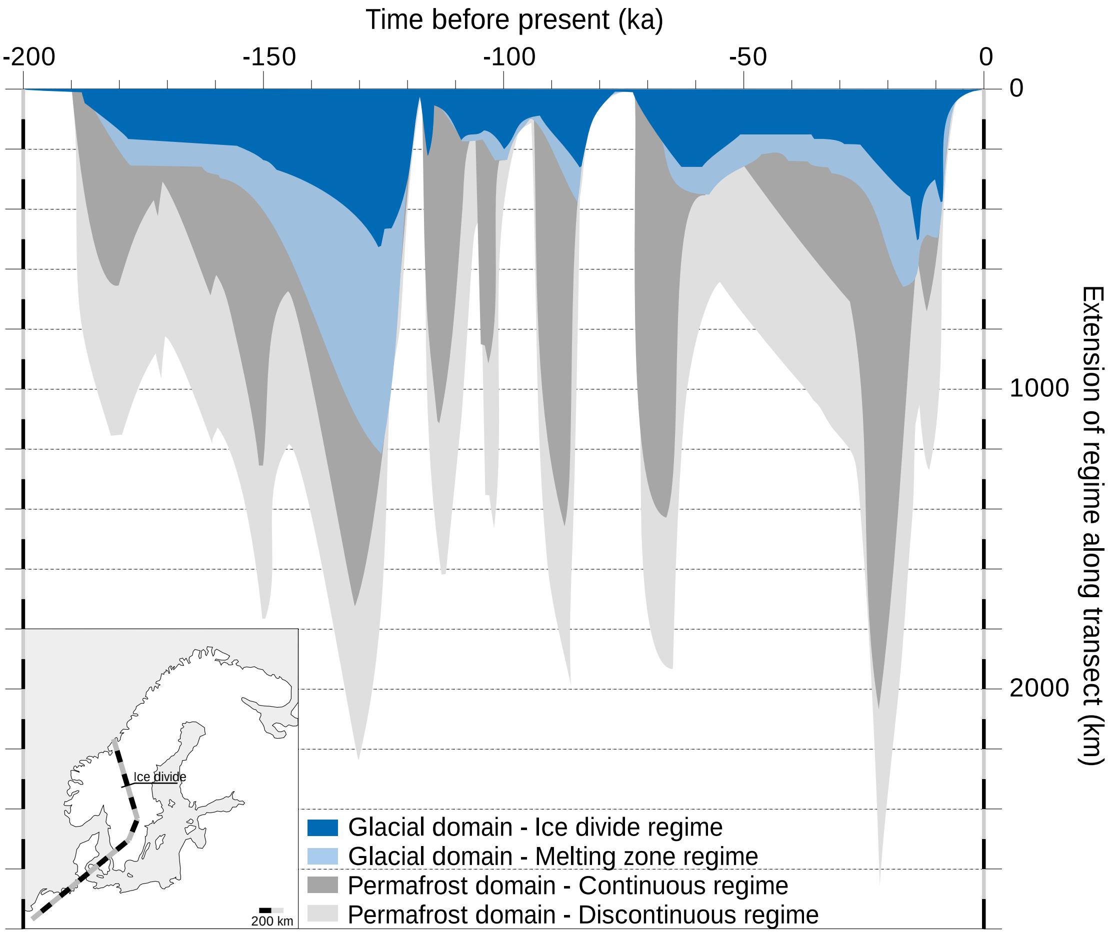

Figure C2-9. *The run N extensions of environmental domains and regimes to the south of the ice divide.*

thickness has also been calculated. The time/space field is completely occupied by glacial, permafrost and temperate/boreal domains. Along the space axis for a given time or the time axis at a given location boundary conditions do not vary within the field of a single regime, but jump abruptly to a new steady state as we cross a regime boundary. Although the extent of marine conditions is not shown, permafrost is assumed not to exist beneath marine areas. We distinguish between the following four domains and eight regimes:

- Glacial domain –
  Ice divide, melting zone and marginal regimes
- Permafrost domain –
  Continuous and discontinuous regimes
- Temperate/boreal domain –
  Interglacial, preglacial and postglacial regimes

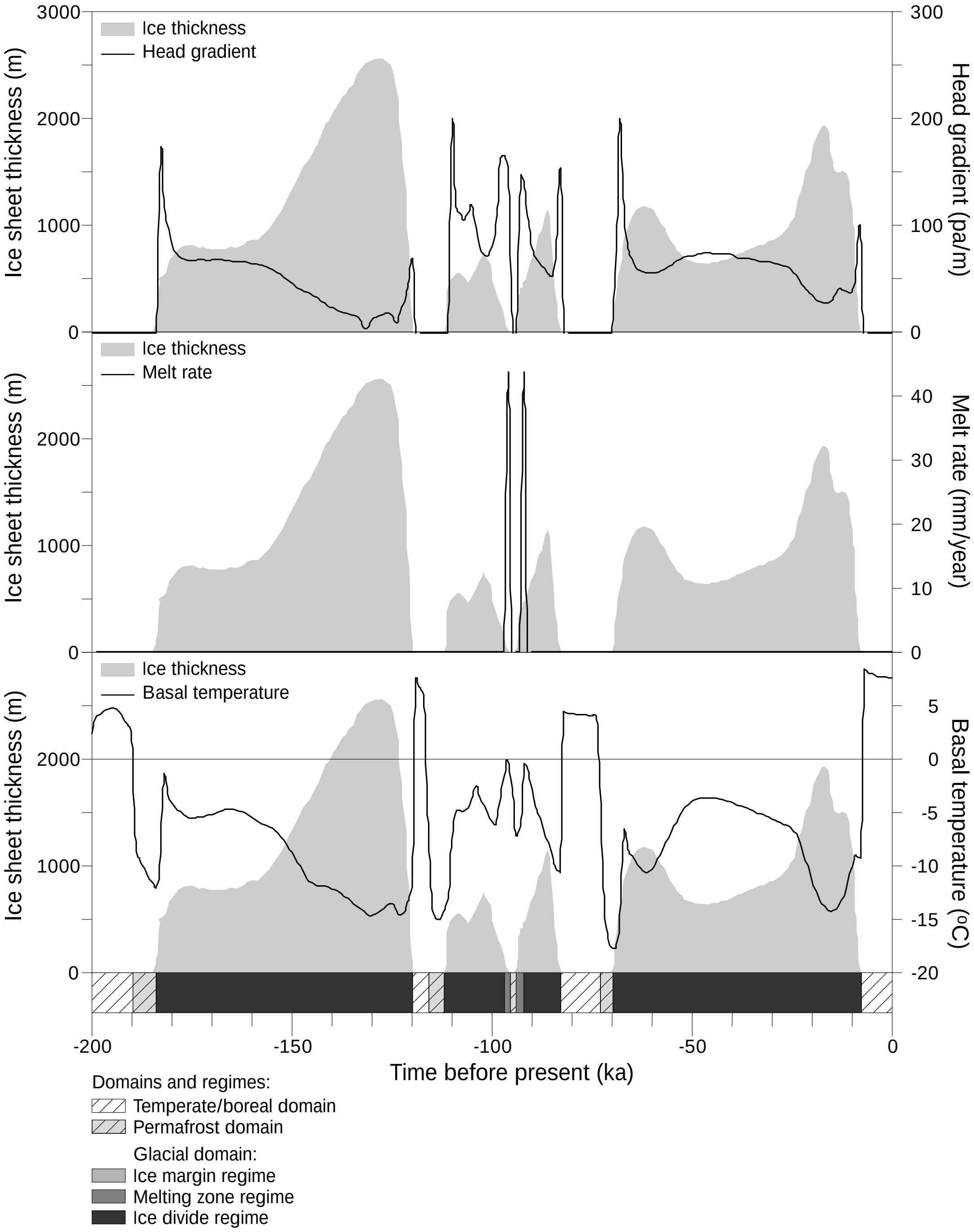

Figure C2-10. *Site specific climate driven boundary conditions at the inland site for the last 200,000 years.*

2.2.1 Palaeoenvironmental time sequences in specific settings

The continuous and discontinuous patterns of variation shown in Figure C2-10 and Figure C2-11 are based on spatially integrated, time dependent models. In this way, by the analysis of spatially varying but integrated properties, such as the thermodynamics of ice sheets, the sequence of extrinsic properties in time at a specific location can be simulated, and expressed as discontinuous or continuously varying properties. An analysis of time/space variation is necessary to achieve this.

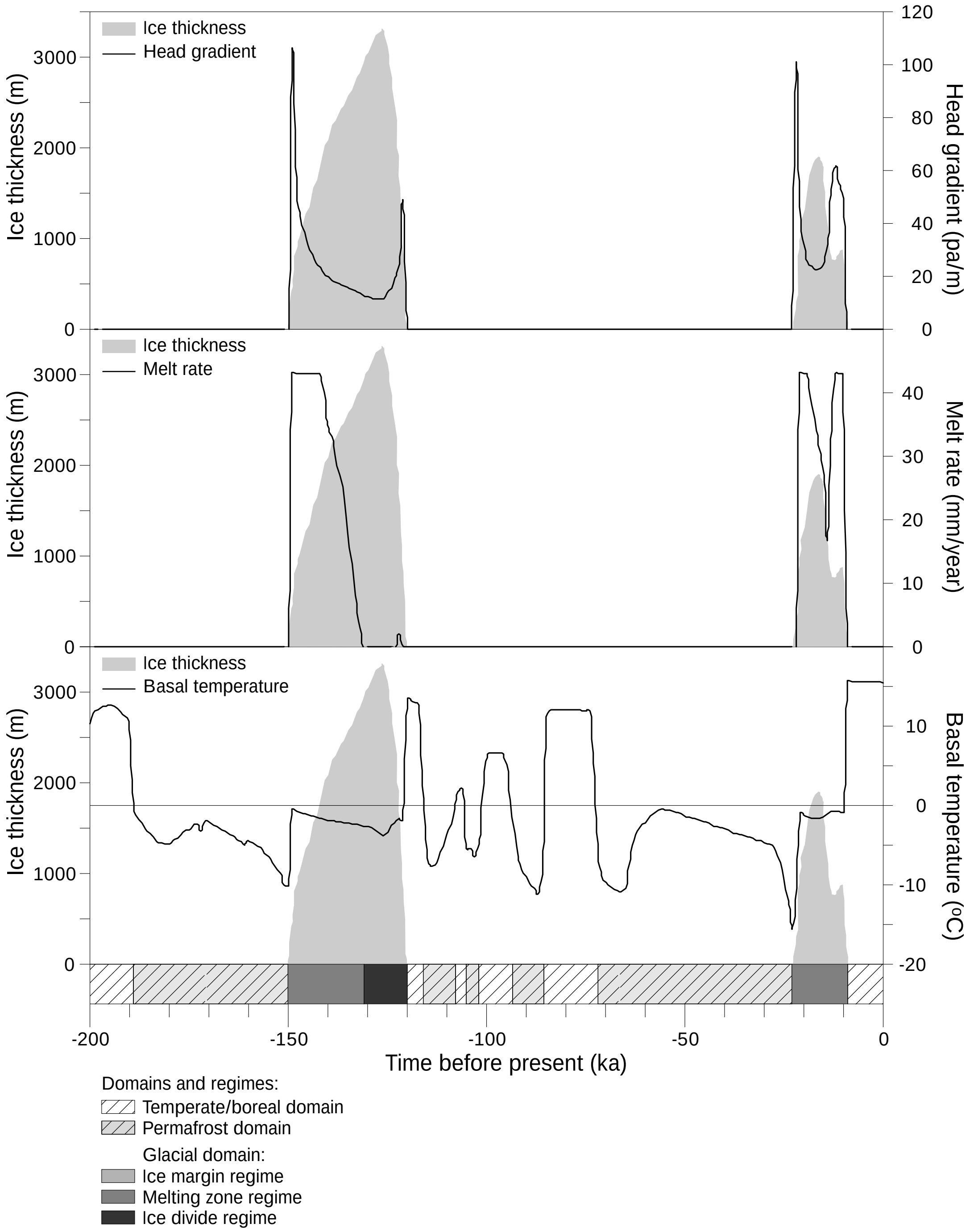

***Figure C2-11.** Site specific climate driven boundary conditions at the (southern) Baltic coast site for the last 200,000 years.*

Model simulations permit us to extract a continuous or discontinuous series of boundary conditions at any given site along the transect. The boundary conditions can be used in performance and safety assessments. Time sequences are shown for two sites. One situated 460 km from the assumed ice divide, representing an inland site, and the other situated 800 km from the ice divide representing a (southern) Baltic coast site. The boundary conditions include groundwater head and head gradient and are derived from simulation N.

The Baltic coast site is intended to simulate a pattern of variation at a site in the general area of Äspo. The transect however does not run through the Äspo area, but some 100 km to the west of the current Baltic coast. The general pattern of glacial behaviour at Äspo will however be much as shown for the area to the west. In order to simulate the pattern of glacially related lake or sea level change, we have used the simulated isostatic behaviour along the transect and corrected for the difference in land elevation. We believe that the discontinuous sequences of past regimes provide a good template for scenarios of future change.

3 Scenarios of future climate change

The orbital parameters governing variations of solar insolation received by the Earth can be correlated with long-term records of climate change. If it is assumed that the processes by which changes in insolation have forced past climate change will also force future change, we can use computations of future insolation to extrapolate future climate change. We have used calculations of the orbital parameters and insolation for the next 200,000 years to produce the variation of future sea level air temperatures (SLAT). To

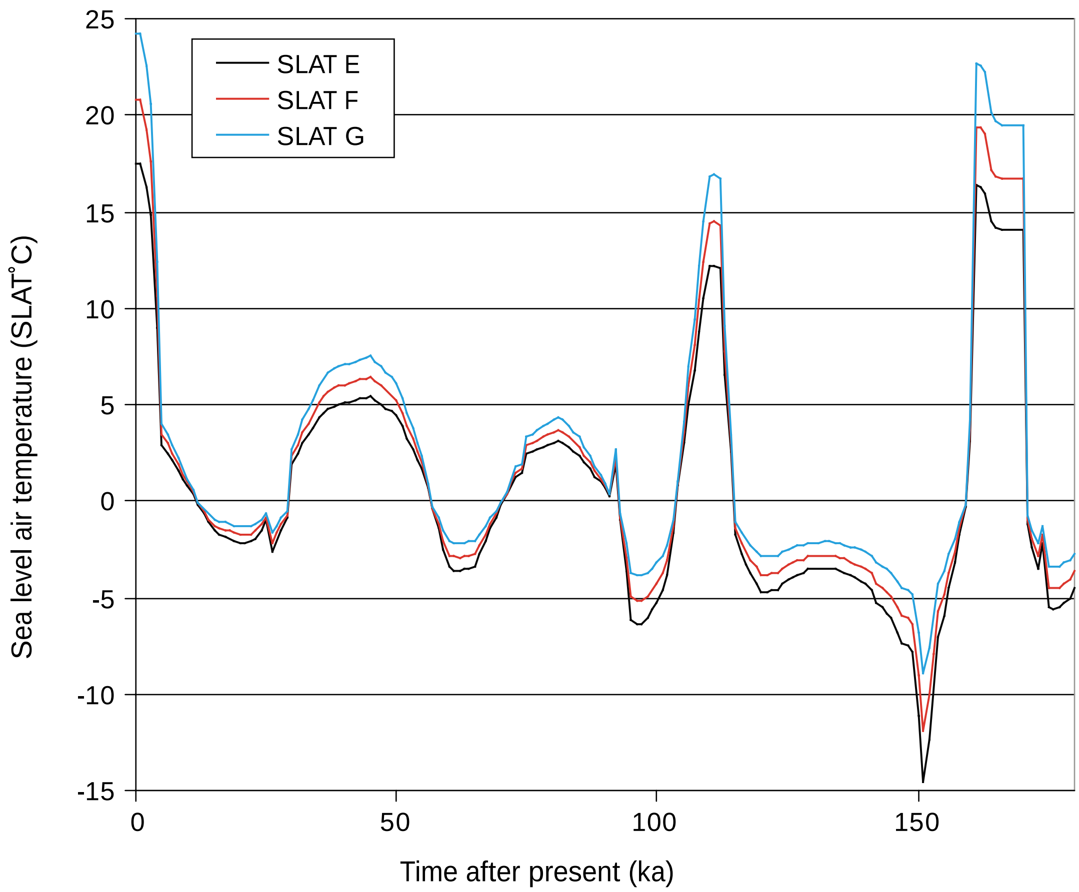

Figure C3-1. *Scenarios of sea level air temperature in southern Sweden for the next 180,000 years.*

compute future equilibrium line altitudes (ELA) the same approach and range of parameters have been used as when generating the hindcasts. How this is achieved is accounted for in Appendix A: Climate and climate change, and the resulting SLAT and ELA scenarios are shown in Figure C3-1 and Figure C3-2 respectively. The scenarios show cold troughs at 20,000, 62,000, 95,000, 120,000, 150,000 and 175,000 years after present, with intervening warmer periods, and the most severe cold period at 150,000 years after present. The warmest peaks, equivalent to interglacial peaks, are at 110,000 and 165,000 years after present. The temporal constancy of the peaks and troughs reflects the best fit in the statistical method. When a longer time series of European climate change (Maar Lake, see Appendix A) becomes available an estimate of the goodness of fit will permit us to explore the variability of peak and trough timing. The variability of amplitude shown in Figure C3-1 reflects the variety of ways in which orbital parameters and variation in insolation can be linearly fitted to the used proxy of global climate change (SPECMAP, see Appendix A).

The climate change scenarios have been used to drive environmental change along the transect (Figure C1-1). Results from three simulations are presented, their designations and the used climate data are presented in Table C3-1.

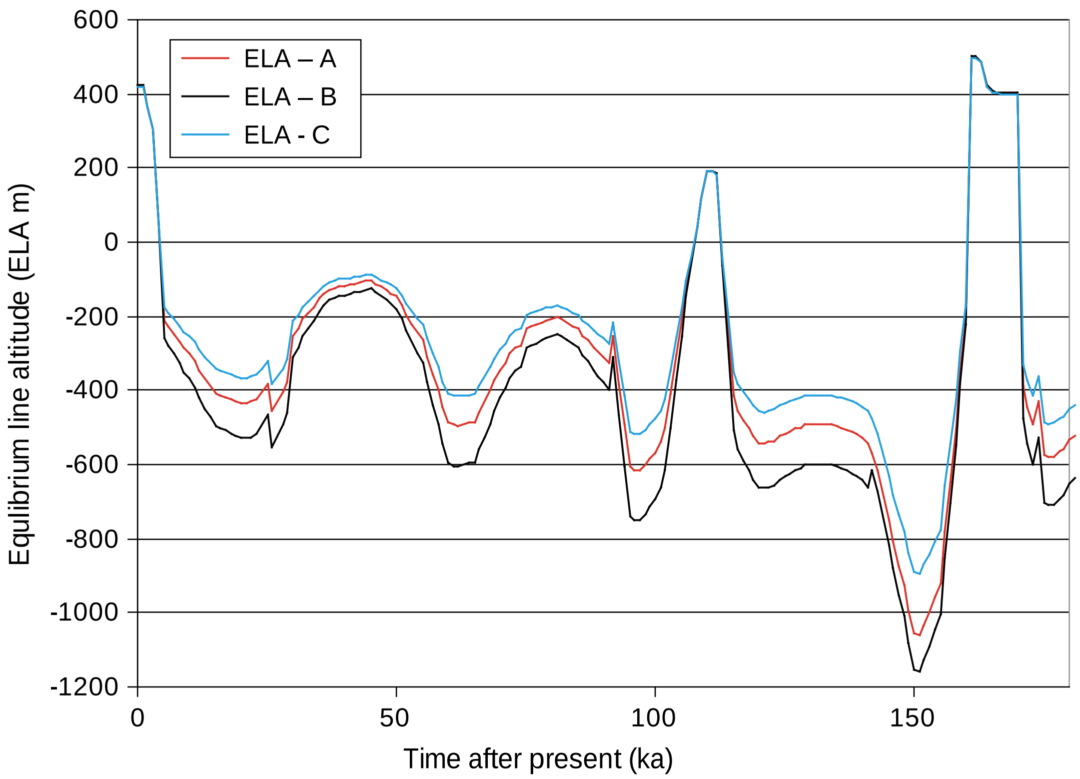

Figure C3-2. *Scenarios of equilibrium line altitude for the next 180,000 years.*

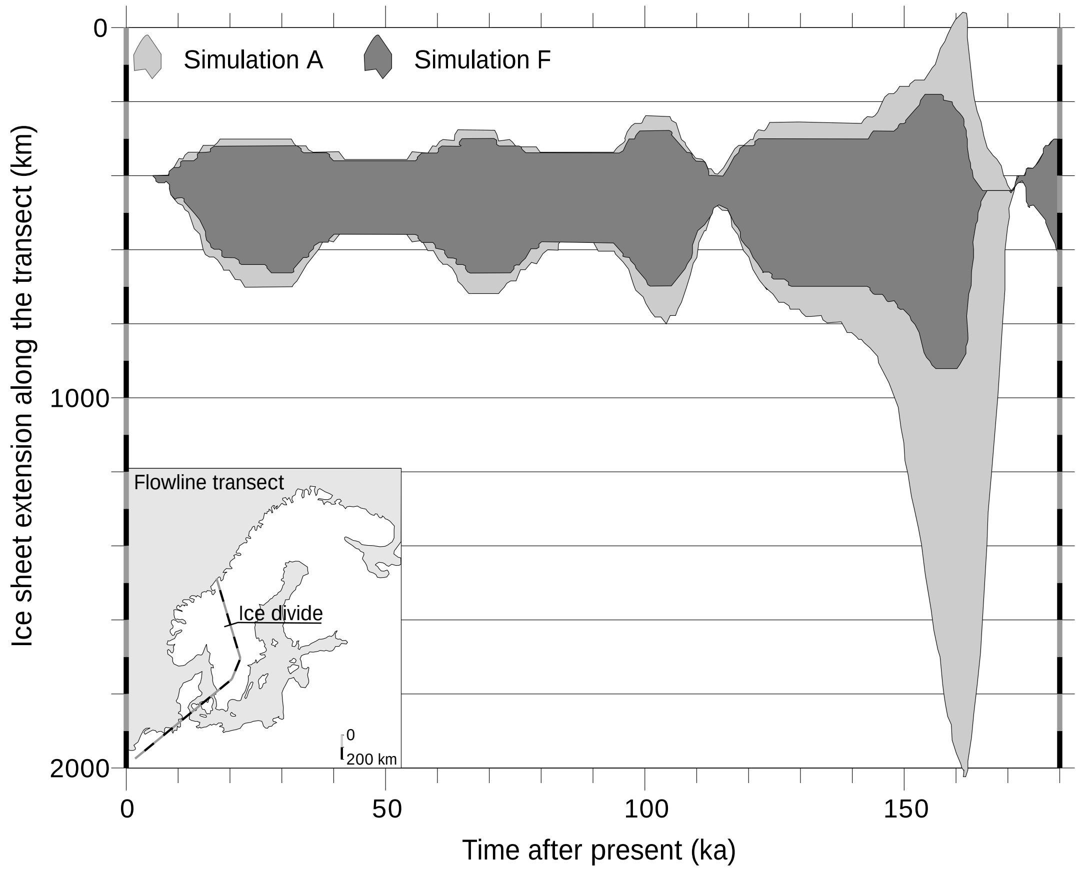

Figure C3-3. *Ice sheet variations along the transect for simulations A and F.*

The fluctuation of the ice sheet margin in simulations A and F are shown in Figure C3-3. Simulation C represents a scenario that resembles simulation S of past change. Figure C3-4 and show site-specific scenarios for the inland site (460 km from the ice divide) and the (southern) Baltic coast site (800 km from the ice divide) based on simulation C.

Table C3-1. Designations of simulations and climate input

Simulation	SLAT	ELA	Mass balance pattern
A	E	B	Most maritime
C	F	A	Intermediate
F	G	C	Most continental

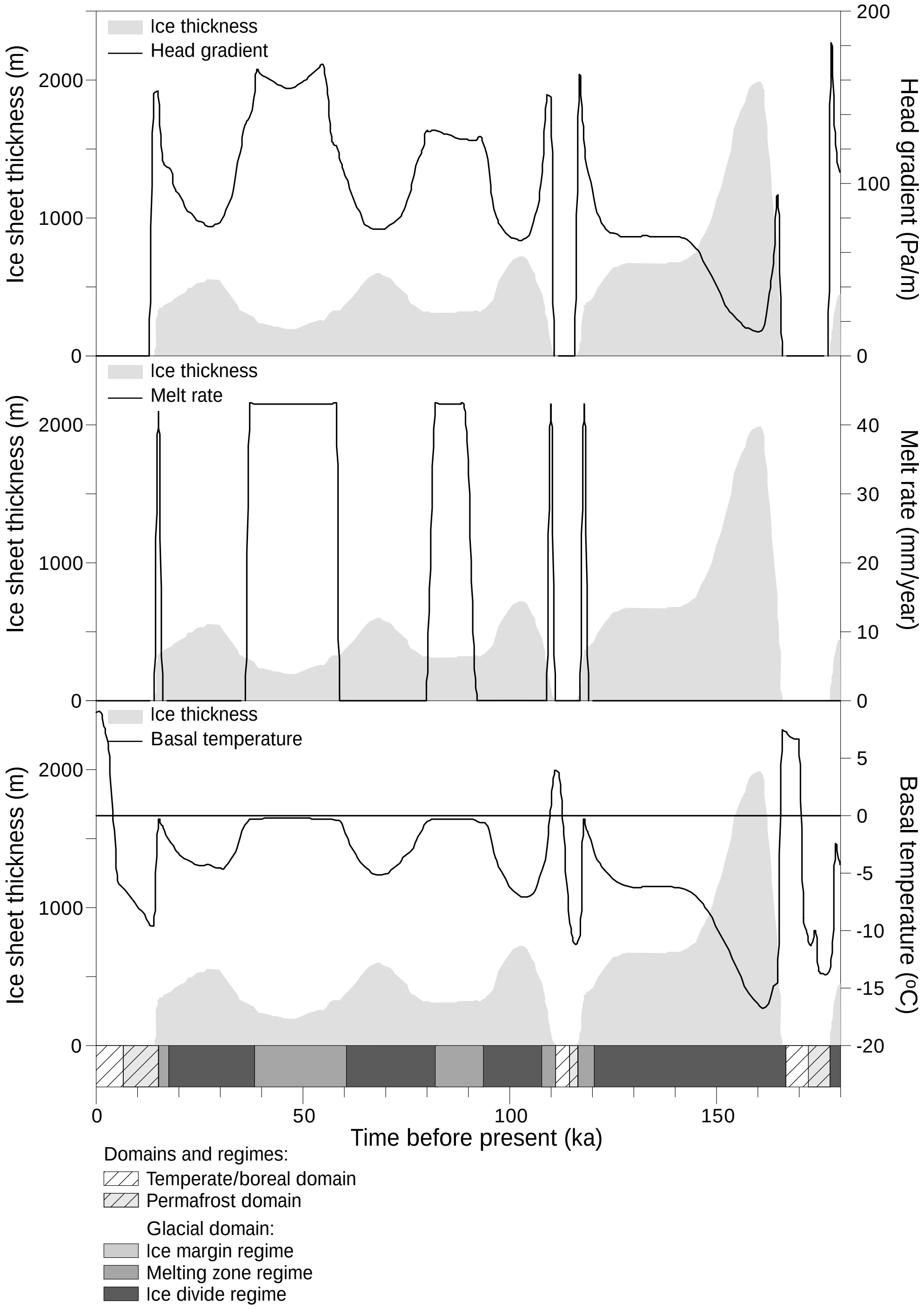

Figure C3-4. *Site specific climate driven boundary conditions at the inland site for the next 180,000 years.*

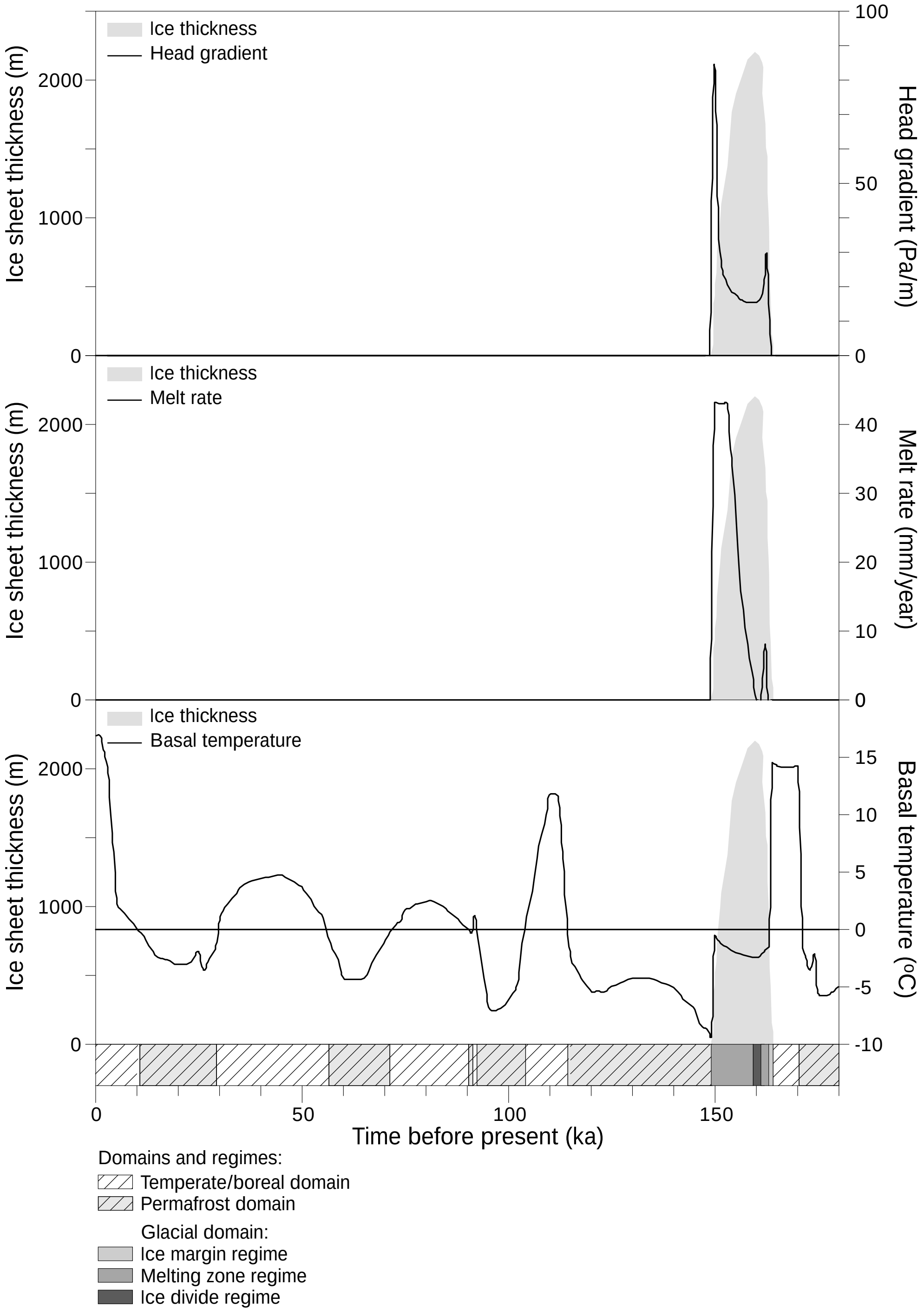

Figure C3-5. *Site specific climate driven boundary conditions at the (southern) Baltic coast site for the next 180,000 years.*

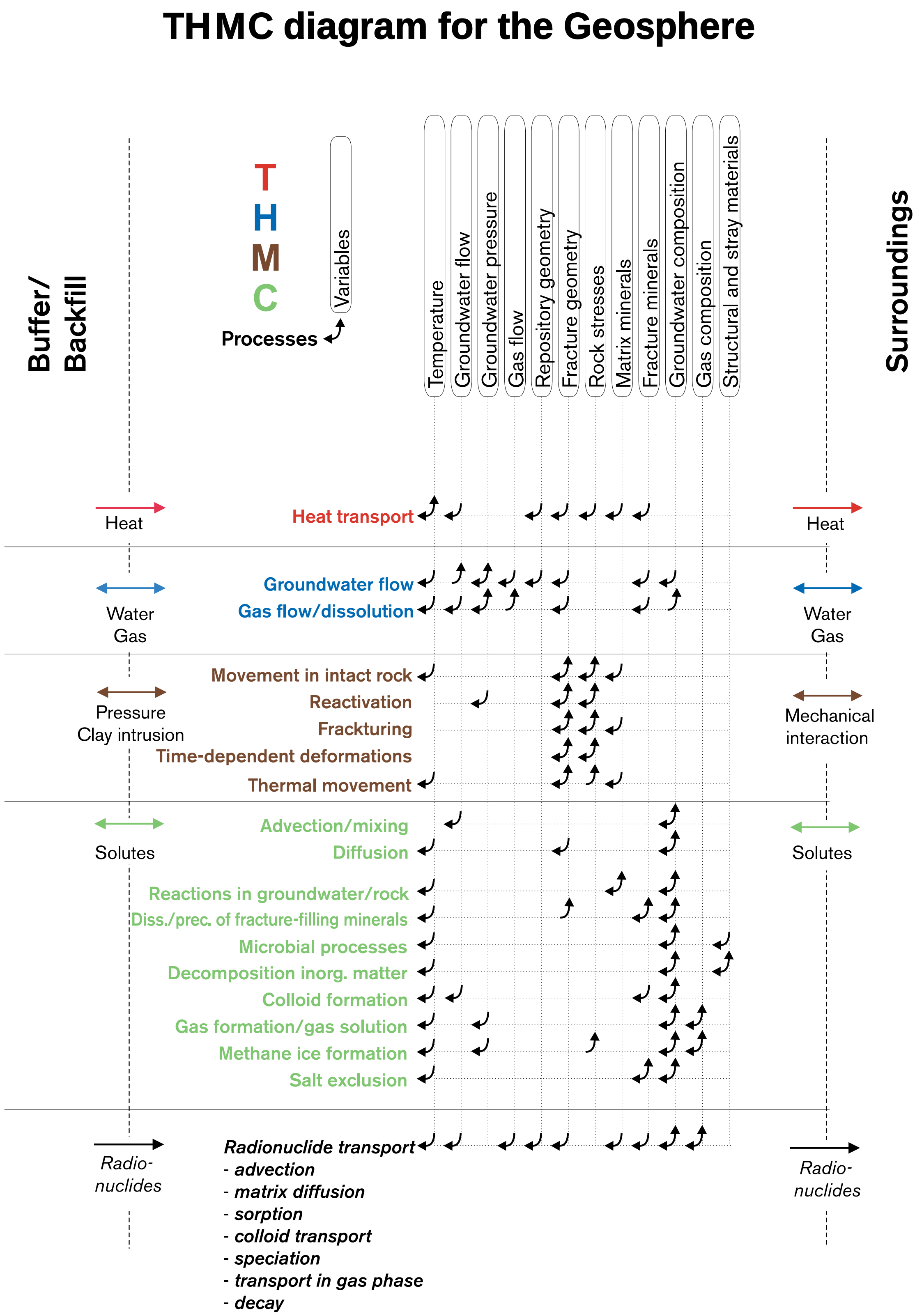

Figure D-1. *Process and interactions in italics only accour when the insolation of the cooper canister is broken.*