Technical Report

TR-01-19

Climate and shoreline in Sweden during Weichsel and the next 150,000 years

Lena Morén, Svensk Kärnbränslehantering AB
Tore Påsse, Sveriges Geologiska Undersökning

August 2001

Svensk Kärnbränslehantering AB
Swedish Nuclear Fuel
and Waste Management Co
Box 5864
SE-102 40 Stockholm Sweden
Tel 08-459 84 00
+46 8 459 84 00
Fax 08-661 57 19
+46 8 661 57 19

Climate and shoreline in Sweden during Weichsel and the next 150,000 years

Lena Morén, Svensk Kärnbränslehantering AB
Tore Påsse, Sveriges Geologiska Undersökning

August 2001

Summary

In this report scenarios of the climate, ice sheet and shoreline in Sweden during the Weichselian and the next 150 000 years are presented. The scenarios are intended to be used in performance and safety analysis of a deep geological repository, as a framework for the analysis of the impact of climate induced changes. First scenarios of the past and future climate are lined out. Based on these and observations of past ice sheets, scenarios of the evolution of the Scandinavian ice sheet are described. Finally the evolution of the shoreline is calculated using an empirical model based on observations from the Late Weichselian and the Holocene.

Climate and ice sheet

The earth climate system comprises the atmosphere, the biosphere, the oceans, the ice sheets, and the surface of the lithosphere. The components of the climate system interact via physical, chemical and biological processes. Solar radiation is the primary energy source driving the climate system. The fact that the radiation energy reaching the earth varies with latitude and that different surfaces, mainly land and water, absorb 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. The climate system is very complex. It includes processes acting on different scale both in time and space, various feedback processes and also a chaotic element.

There can be several causes of climate change. External causes are variation of solar radiation and dissipation of internal earth energy producing volcanism or shifts in earth physiography. Change in the 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 the dynamics of ocean currents and ice sheets, albedo and biological processes.

Our knowledge of past climate is based on measurements of climate variables such as temperature and precipitation. Observations are only available for the last 100-150 years or so and are mainly from northern hemisphere land areas. For information of the climate for other parts of the world, and for longer time periods we rely on proxy records. During the last century various climate data have been collected and compiled. Two very important sources of information about long term climate changes are deep-sea sediments and ice cores. The availability to data has together with the development of weather forecasting models and computers made it possible to construct computer codes that can be used to simulate the behaviour of the climate system. Data and computer codes has contributed to increase our knowledge of the complex climate system, but still it is not possible to make any long term predictions of either past or future climate evolution.

Changes of the earth orbit around the sun cause variations in the seasonal distribution and amount of solar radiation reaching the earth. Records of past climate show that there is a correlation between these variations and long-term climate changes. The theory that climate changes are triggered by variations in the earth orbital parameters is refereed to as the astronomical climate theory or the Milankowich theory. In spite of some ambiguities this theory is generally accepted. The orbital parameters, and from them the variation of insolation, can be calculated from the equations of planetary mechanics. Changes in insolation can then be linked to climate changes. In this report results from three different models based on the astronomical climate theory are utilised. Simulations are compared to observations of past climate and ice sheets. The climate and ice sheet scenario for the Weichselian is based on deep-sea sediment data and a reconstruction of the Scandinavian ice sheet. The future scenario is based on an arbitrary compilation of the results from the three models and the reconstruction of the Weichselian ice sheet.

Shoreline

Shore level displacement is due to two interactive vertical movements, glacio-isostatic depression/uplift and global eustatic sea level lowering/rise. The outermost parts of the earth are in this context referred to as the lithosphere and the asthenosphere. The lithosphere behaves like a brittle solid, and the asthenosphere exhibits plastic behaviour. The isostatic movements are primarily controlled by viscous flow of mantle material in the asthenosphere while the lithosphere often can be regarded as an elastic layer. During glacial phases water from the oceans is stored in ice sheets all over the world. The transport of water from the oceans unloads the oceanic bed and gives rise to a flow of mantle material from the continents towards the oceans causing uplift of the oceanic bed and lowering of coastal areas. Simultaneously a flow of mantle material out from glacialted areas takes place, causing depression of the underlying lithosphere. Stresses in the lithosphere caused by the ice load can produce a depression far beyond the limits of the ice sheet. The outflow of mantle material can also cause an elastic upward bending of the crust – a forebulge – further out from the ice edge. This makes the crustal movements in peripheral parts of the ice sheet – especially in coastal areas – very complex.

As seawater is stored in ice sheets during cold phases, changes of the global sea level – the euststatic component of shore level displacement – can be correlated to extensions of ice sheets. Crustal movements however make it complicated to derive past global sea levels from geological and other data.

The course of glacio-isostatic uplift since the Late Weichselian can be described by means of mathematical expressions adapted to empirical data. Isostatic crustal movements start slowly, reach a maximum rate and thereafter follow a retarded course. *Arctan* functions have proved to be suitable tools for describing the long-term evolution of glacio-isostatic uplift in time. The functions include three factors, which have been determined through an iterative comparison to empirical data. The factors are; the inertial factor (B), the download factor (A_T) and the time for maximal subsidence/uplift rate (T). The internal factor is related to the crustal thickness and determines the changes of the subsidence/uplift rate over time. The download factor is related to the ice load and represents the subsidence/uplift at the time for the maximal subsidence/uplift rate (A_T). The ice thickness is of great importance for the amount of isostatic depression, due to the inertia and slow development of the viscous flow of mantle material the duration of the load is also of importance.

To describe the shoreline displacement during the Weichselian and the next 150 000 years download factors (A_T) and times for maximum subsidence uplift (T) have been determined given the postulated ice sheet scenarios. The inertial factor has been determined from Late Weichselian and Holocene observations. The past eustatic component consist of a modification of published curves which has been revised to fit the postulated ice sheet scenario. The future eustatic curve is derived from the ice sheet scenario.

Uncertainties

The presented scenarios involve great uncertainty. The climate and ice sheet scenario for the Weichselian is based on observations, and the involved uncertainties are related to lack of data and uncertain dating. Lack of knowledge and capabilities to simulate the climate system are the main sources of uncertainty for the future climate and ice sheet scenario. The shoreline displacement scenario is based on a purely empirical model and the assumption that the evolution of the isostatic component always follows the same pattern as observations implies it did since the Late Weichselian. The model takes the crustal thickness into account, but it does not include any physical explanation or expression for the subsidence/uplift. In reality isostatic movement is a complex process affected by the properties of the crust and the mantle material and by the size and duration of the ice load. In spite of the shortcomings of the used models, the scenarios are considered to give an overview of the evolution during the studied periods. Such general descriptions of the evolution, including several complex processes, would have been hard to derive from more complex/realistic models.

Sammanfattning

I denna rapport presenteras scenarier för klimat, istäcke och strandlinje under Weichsel och 150 000 år framåt i tiden. Scenarierna är avsedda att användas i säkerhetsanalyser av djupförvar för använt kärnbränsle, som bakgrund till analysen av hur förändringar orsakade av ändrade klimatförhållanden påverkar förvaret och dess säkerhet. Först skisseras scenarier av gånget och framtida klimat. Baserat på dem och observationer av gångna tiders istäcken beskrivs sedan scenarier för isutbredningen. Slutligen beräknas kustlinjens utveckling med hjälp av en empirisk modell baserad på observationer från Sen-Weichsel och Holocene.

Klimat och istäcke

Jordens klimatsystem består av atmosfär, biosfär, hav, istäcken och ytan av litosfären. Klimatsystemets olika delar samverkar via fysiska, kemiska och bilogiska processer. Solinstrålning är klimatsystemets främsta energikälla. På grund av att strålningsenergin som når jorden varierar med latituden och pga att olika ytor, främst land och vatten, absorberar strålning i olika gard uppstår en ojämn uppvärmning. Klimatet är ett resultat av denna ojämna uppvärmning och de överförings processer den ger upphov till. Klimatsystemet är mycket komplicerat, med processer som verkar i olika skala både i tid och rum, återkopplingar av olika slag och även ett kaotiskt inslag.

Det finns flera orsaker till klimatförändringar. Orsaker som är externa i relation till klimatsystemet är variationer i solinstrålningen och frigörelse av energi från jordens inre i form av vulkanutbrott eller ändringar av topografi, landutbredning mm. Ändringar i klimatsystemets inre dynamik är en annan orsak till klimatförändringar. Koncentrationen av växthusgaser i atmosfären påverkar både jordens energibalans och meteorologiska processer och därmed klimatet. Havens och istäckenas dynamik, förändringar av albedo och biologiska processer kan också orsaka klimatförändringar.

Vår kunskap om gångna tiders klimat baserar sig på mätningar av olika klimatvariabler som temperatur och nederbörd. Observationer finns endast för de senaste ca 100–150 åren, och då huvudsakligen från norra halvklotets landområden. För att få information om klimatet i andra delar av världen eller längre tidsperioder är man hänvisad till olika spår klimatet lämnat i växlighet, jordlager, sediment, inlandsisar mm. Under det senaste århundradet har diverse klimatdata samlats in och sammanställts. Två mycket viktiga källor till information om långtida klimatförändringar är djuphavssediment och iskärnor. Tillgången till data har tillsammans med utvecklingen av väderprognosmodeller och datorer gjort det möjligt att göra datorprogram som kan användas för att simulera klimatsystemet. Tillgången till data och datormodellerna har bidragit till att öka vår förståelse för det komplexa klimatsystemet, men det är fortfarande inte möjligt att göra några prognoser, vare sig över gånga tiders eller framtida klimatutveckling.

Förändringar av jordens bana runt solen medför att fördelningen av solinstrålning över jorden och den totala mängden solinstrålning som når jorden varierar med tiden. Spår av gångna tiders klimat visar att det finns ett samband mellan dessa variationer och långtida klimatförändringar. Teorin att klimatförändringar orsakas av variationer av jordens omloppsparametrar kallas den astronomiska klimat teorin eller Milankowich teori. Trots att det finns vissa tvetydigheter är teorin allmänt accepterad. Jordens omloppsparametrar och

från dem solinstrålningen kan beräknas med god noggrannhet. Förändringarna i solinstrålning kan sedan kopplas till klimatförändringar. I denna rapport har resultat från tre olika modeller som baserar sig på den astronomiska klimat teorin använts. Modellsimuleringar har jämförts med observationer över gångna tiders nedisningar och klimat. Scenariot över Weichsel periodens klimat och istäcke baserar sig på data från djuphavssediment och en rekonstruktion av det skandinaviska istäcket. Framtidsscenariot baserar sig på en godtycklig jämförelse av resultat från de tre modellerna och isutbredningen under Weichsel.

Strandlinje

Strandlinje förskjutning är ett resultat av två samverkande vertikala rörelser, glacio-isostatisk nedtryckning/upphöjning av jordskorpan och global eustatisk förändring av havsytans nivå. De yttre delarna av jordskorpan benämns i detta sammanhang litosfären och astenosfären. Litosfären beter sig som ett sprött solitt material medan astenosfären visar plastiska egenskaper. De isostatiska rörelserna orsakas framförallt av ett visköst flöde av mantelmaterial i astenosfären, medan litosfären kan betraktas som ett elastiskt skikt. Under glaciala perioder samlas vatten från haven i inlandsisar världen över. Transporten av vatten från haven till inlandsisarna medför att havsbottnarna avlastas. Det ger upphov till ett inflöde av mantelmaterial ut från kontinenterna till området under havet, varpå havsbotten höjs och kustområdena sänks. Samtidigt sker ett utflöde av mantelmaterial från områdena under inlandsisarna som får jordskorpan under isarna att tryckas ned. Spänningar som byggs upp i litosfären pga av islasten kan leda till nedtryckning av jordskorpan utanför iskanten. Utflödet av mantelmaterial till området utanför isen kan orsaka en uppböjning av jordskorpan längre ut från iskanten. Detta medför att de isostatiska rörelserna i de yttre delarna av nedisade områden blir svåra att beskriva, speciellt i kustområden.

Eftersom havsvatten lagras i inlandsisar under kalla perioder kan förändringar i havsnivån – den eustatiska komponenten av strandlinjeförskjutningen – relateras till istäckenas utbredning. Jordskorpans rörelser medför dock att det är svårt att bestämma gångna tiders havsnivåer från geologiska och andra data.

De glacio-isostatiska rörelserna sedan Sen-Weichsel kan beskrivas genom matematiska utryck anpassade till empiriska data. Isostatiska rörelser inleds långsamt, når en maximal hastighet och avtar sedan. *Arctan* funktioner har visat sig vara lämpliga för att beskriva den långsikta utvecklingen av de isostatiska rörelserna i tiden. Funktionerna innehåller tre faktorer, som alla bestämts genom ett iterativt jämförande mot empiriska data. Faktorena är; tröghetsfaktorn (B), nedtryckningsfaktorn (A_T) och tiden för den maximala nedtrycknings-/höjningshastigheten (T). Tröghetsfaktorn är relaterad till jordskorpans tjocklek och avgör landsänknings/-höjnings förloppets utveckling i tiden. Nedtryckningsfaktorn är relaterad till islasten och representerar nedtryckningen/höjningen vid tiden för dess maximala hastighet (T). Isens tjocklek har stor betydelse för hur stor nedtryckningen blir, pga av tröghet och den långsamma utvecklingen av flödet av mantelmaterial påverkar också lastens varaktighet nedtryckningen.

För att beskriva strandlinjeförskjutningen under Weichsel och de kommande 150 000 åren har nedtryckningsfaktorer (A_T) och tider för maximal nedtryckning/upphöjning (T) bestämts baserat på scenarierna för klimat och isutbredning. Tröghetsfaktorn har bestämts utifrån observationer från Sen-Weichsel och Holocen. Gångna tiders eustatiska komponent utgörs av en modifierad version av publicerade kurvor över havsnivåförändringar som anpassats till den postulerade isutbredningen. Den framtida eustatiska utvecklingen har härletts från scenariot över den framtida isutbredningen.

Osäkerheter

De presenterade scenarierna rymmer stora osäkerheter. Scenariot över klimat och istäcke under Weichsel är baserat på observationer, och osäkerheterna är relaterade till brist på data och osäker datering. Kunskapsbrist och bristande möjligheter att simulera klimatsystemet är de huvudsakliga källorna till osäkerheter runt framtidsscenariot för klimat och isutbredning. Scenariot över strandlinjeförskjutningen baserar sig på en rent empirisk modell och antagandet att den isostatiska utvecklingen alltid följer samma mönster som observationer visar att den gjort sedan Sen-Weichsel. Modellen tar hänsyn till jordskorpans tjocklek, men den innehåller inga fysikaliska förklaringar eller uttryck för nedtryckningen/upphöjningen. I verkligheten är isostatiska rörelser en komplicerad process som påverkas av jordskorpans och mantelmaterialets egenskaper samt islastens storlek och varaktighet. Trots bristerna i de använda modellerna bedöms scenarierna ge en överblick över utvecklingen i det studerade tidsintervallet. En sådan överblick, omfattande flera komplexa skeenden, skulle vara svår att härleda ur mera komplexa/realistiska modeller.

Contents

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 1-1*.

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.

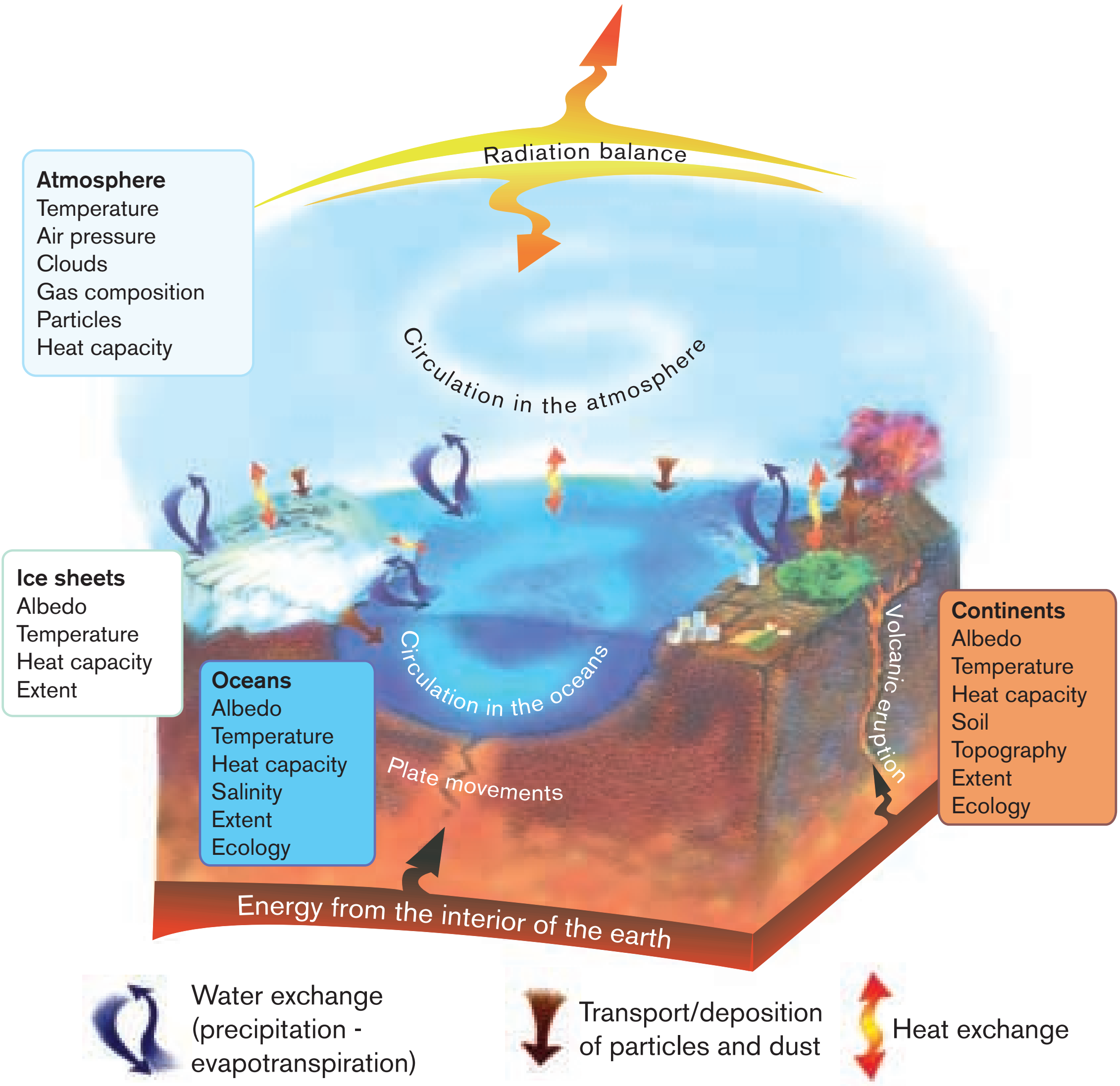


Figure 1-1. *The components of the climate system and some of their properties. The climate is determined by the properties of the components and how they are changed by internal processes and interaction between the parts. Externally the climate system is affected by insolation and energy from the interior of the earth.*

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. The parameters and their impact on the solar irradiation reaching the earth are accounted for in section 2.1 The earth orbit around the sun. 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.

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 and Karlén, 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 slightly 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 1-2*.

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.

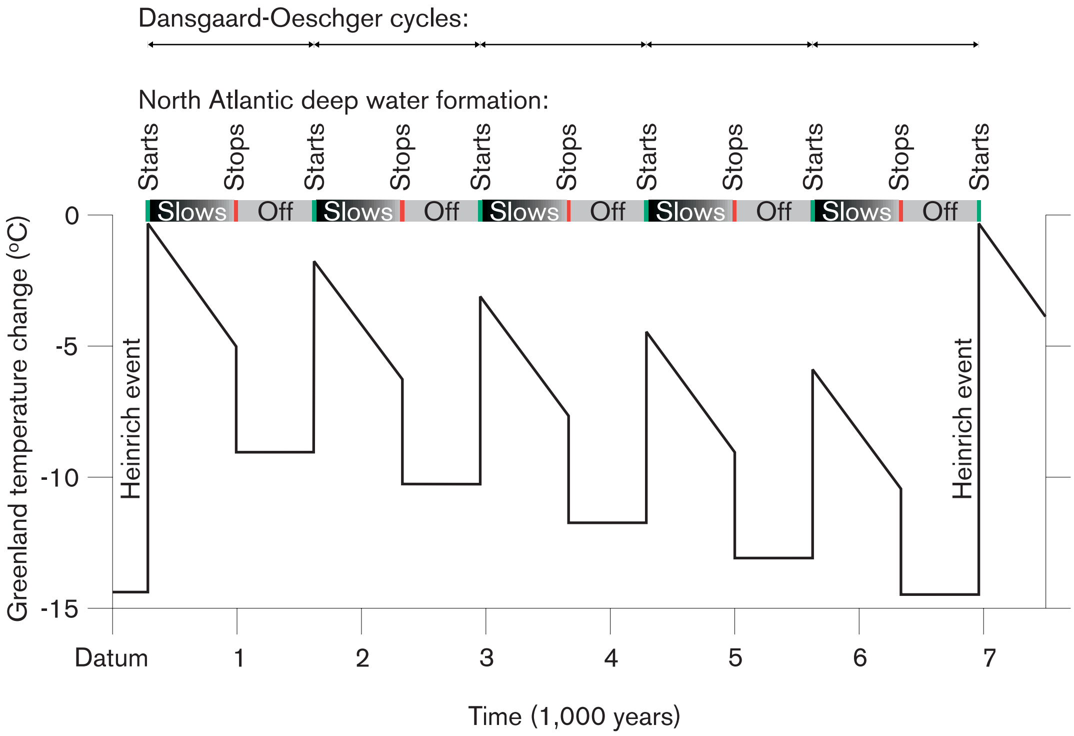


Figure 1-2. *Several successively cooler Dansgaard-Oescheger cycles are bundled together. In the beginning and ending of these bundled cycles Heinrich events occur. Dansgaard-Oescheger cycles start with a rapid warming believed to be associated with the start of North Atlantic deep-water formation. A shut off of the North Atlantic deep-water formation is believed to cause the abrupt cooling at the end of the cycles /revised from Holmgren and Karlén, 1998/.*

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 tenths 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 and Karlén, 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:

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

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 /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 and Karlén, 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, 1995a, 1995b 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 and Karlén, 1998/. Recently data from a core from Antarctica covering the past 420,000 years was presented /Petit et al, 1999/. Variations of $\delta^{18}O$ 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, 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 Holland et al, 1999/ 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.

For the assessment of the performance and safety of a deep geological repository for spent nuclear fuel we are mainly concerned with long-term climate changes causing transitions between different climate domains such as temperate/boreal, permafrost and glacial. These changes are believed to be 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 1980th, for instance /Imbrie and Imbrie, 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 1990th 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, 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.

2 The astronomical climate theory

The variations of the orbital elements affect the amount and distribution of solar irradiation reaching the earth. According to the astronomical climate theory the orbital variations are the main forcing factor of long-term climate changes.

2.1 The earth orbit around the sun

The shape of the Earth's orbit around the sun is fluctuating. The tilt of the equator with respect to the plain of the Earth's orbit and the time of the year when the earth makes its closest approach to the sun also vary. Variations in the orbital elements affect the seasonal distribution and amount of solar irradiation reaching the earth. Each of the orbital elements is a quasi-periodic function of time. The orbital elements are (also see *Figure 2-1*):

- Eccentricity (e = *focal distance/length of major axis*),
  the shape of the earth orbit around the sun changes from circular to slightly elliptical ($0 < e < 0.06$). The variation spectra include several components, the most important term has a period of 413,000 years. Eight of the next twelve terms range from 95,000 to 136,000 years, these terms contribute to a peak often referred to as the 100,000-year eccentricity cycle /Imbrie and Imbrie, 1980/. The eccentricity affects the total amount of solar energy reaching the earth. The effect is small, in the order of e^2, or about 0.1% at the top of the atmosphere. The present value of e is 0,017 /Imbrie and Imbrie, 1980/.
- Obliquity (ε),
  the tilt of the earth axis to the plane of its orbit around the sun varies between 21.8° and 24.4°. The basic period is about 41,000 years /Holmgren and Karlén, 1998/. Changes in obliquity affect the distribution of solar radiation reaching the earth. Low tilt reduces the amount of irradiation reaching the poles. The obliquity is at present 23.5° /Imbrie and Imbrie, 1980/.
- Precession,
  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 (passes perihelion). At perihelion the insolation reaches a maximum. Variations caused by precession can be quantified in different ways. The longitude of the perihelion (ω) measured as the angular distance of perihelion from the vernal or autumnal equinox is one possibility. To include the dependence between perihelion and eccentricity the variation can be expressed as $e \sin(\omega)$ or $\Delta e \sin(\omega)$, where Δ denotes the difference from the value of $e \sin(\omega)$ at a specific time. Another often-used measure is the June 21 distance to the sun[1]. The precession has an irregular periodicity of 13,000 to 25,000 years with periods of about 23,000 years and 19,000 years being the most common /Kukla et al, 1981/. At present the earth is closest to the sun in early January /Bogren at al, 1998/ ($e \sin\omega = 0.016$ /Berger et al, 1996/).

[1] When the value of $e \sin \omega$ for June 1950 is subtracted, the resulting precession index $\Delta e \sin \omega$ is approximately equal to the deviation from the 1950 value of the earth-sun distance in June, expressed as a fraction of the semi-major axis of the earth's orbit /7, 8/.

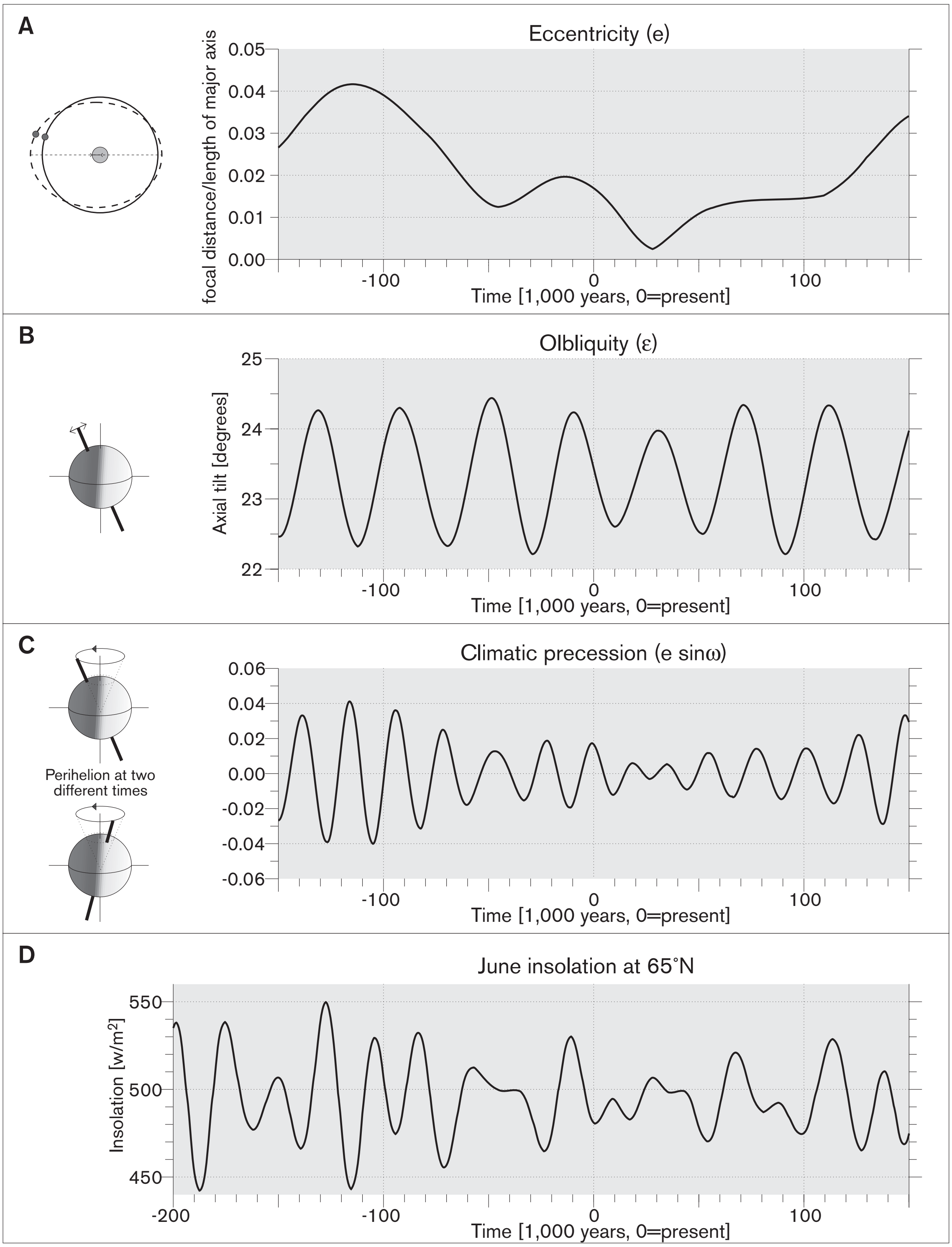


Figure 2-1. *The orbital parameters affect the seasonal distribution (obliquity and precession) and amount of solar irradiation reaching the earth (eccentricity). A) Eccentricity. B) Obliquity. C) Precession expressed as e sin(w). D) June insolation at 65°N. (Astronomical elements, insolation from /Berger, 1978a,Berger, 1978b/, graphs revised from /Berger and Loutre, 1997/.)*

Presuming a perfectly transparent atmosphere and a constant solar output the energy available at any given latitude of the Earth surface is a single-valued function of the orbital elements /Berger and Loutre, 1991/. The impact on the insolation of the orbital parameters varies as a function of latitude, season and time. Integrated over all latitudes and over an entire year, the energy influx depends only on the eccentricity (*e*) /Imbrie and Imbrie, 1980/. June insolation at 65°N is often used as a guideline for analysis of climate change, and it is shown in *Figure 2-1* to illustrate changes in insolation.

The Northern Hemisphere variations of insolation during the last glacial-interglacial are summarised in /Berger and Loutre, 1997/. Today, the Northern Hemisphere receives ~10.8 MJ/m^2 annually. During the last glacial-interglacial cycle this value has varied by 0.1% due to eccentricity. Variations of this energy during the astronomical summer – that is between the summer solstice and the autumnal equinox – and winter – between the winter solstice and vernal equinox – are due to obliquity. During the last glacial cycle the variation during the astronomical summer has been 1.8% (3.41 – 3.35 MJ/m^2), and during the astronomical winter the variation has been 3% (1.99 – 2.05 MJ/m^2). The seasonal contrast has varied by about 10% (total variation 1.36 MJ/m^2) with a periodicity governed by obliquity. The daily insolation values at the summer and winter solstices have varied by about 20% mainly with precession (440 – 520 W/m^2 at summer solstice and 190 – 230 W/m^2 at winter solstice).

2.2 Development of the astronomical climate theory

Evidences of the existence of major glaciations were put forward in the late 1800s and early 1900s. Contemporarily long-term changes in climate were predicted, and a connection between climate change and variations in insolation due to changes in the earth orbit around the sun was suggested. The first physical evidence of long term cyclic variation in climate became available about 50 years ago with improved deep-sea coring technique and the use of high precision mass spectrometry /Holmgren and Karlén, 1998, Imbrie et al, 1984/.

In the 1940s the astronomer Milutin Milankovitch put forward a full astronomical theory of the Pleistocene ice ages, often referred to as the Milankovitch theory. Milankovitch computed the orbital elements and the subsequent changes in the insolation and related them to climate change /Berger and Loutre, 1991/. Following the suggestions of contemporary researchers he concluded that the intensity of the radiation received at northern latitudes during summer was critical to the growth and decay of ice sheets /Imbrie and Imbrie, 1980/. Today the astronomical climate theory is basically accepted. However all mechanisms are not fully understood and there is a need for further investigations of the coupling between changes in insolation and climate change.

Research in the astronomical theory of palaeoclimates involves four main steps /Berger and Loutre, 1991/

1. The theoretical computation of the long-term variations of the Earth's orbital parameters and related geometrical insolations.
2. The design of climatic models to transfer the insolation into climate.
3. The collection of geological data and their interpretation in terms of climate.
4. The comparison of these proxy data to the simulated climatic variables.

The orbital parameters can be computed for any past or future time from the equations of planetary mechanics. Several researchers have improved the solution of these equations since Milancovitch presented his theory. The solutions presented by Berger in 1978 (BER78) /Berger, 1978a, Berger, 1978b/ and by Berger and Loutre in 1991 (BER90) / Berger and Loutre, 1991/ were used in the studies referred to in this report.

From the orbital parameters the insolation at different latitudes and times of the year can be calculated. With a model of how changes in insolation are linked to changes in climate the latter can be calculated. In the most simple models the search for solar-terrestrial links in the climate system are bypassed and the orbital perturbations are directly correlated to climate /Berger et al, 1980/. The most complex models try to include all the physical processes affecting the climate system and simulate its response to changes in the incoming irradiation. Since the orbitally forced changes in insolation have periodicities in the order of 10,000 to 100,000 years the models focus on climatic changes on the same time scale, like the waxing and waning of continental ice sheets.

The computed climate changes can be compared to geological evidences. A difficulty involved in this is the interpretation of the geological data. A key issue is to obtain an independent dating of the geological data and a chronology of climate events independent of the theory it self.

In the late 1960's improved techniques for dating and interpreting geological data in terms of palaeoclimate became available. The computations of the orbital parameters were also reviewed and improved. Long palaeoclimate records and accurately computed orbital parameters made it possible to test the astronomical theory in the frequency band. Hays et al /1976/ showed that the frequencies of climate change as depicted in palaeoclimatic variables in deep-sea sediments matched the frequencies of the orbital perturbations. Imbrie et al /1984/ calibrated a Quarternary time scale based on palaeoclimate data from deep-sea sediments and computations of the orbital elements.

Today the astronomical climate theory is basically accepted. But uncertainties still remain, one of the most important is related to the dominance of the 100,000-year cycle. The dominance of the cycle is evident the past 400,000 years or so, and it can be seen for the last 900,000 years. 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/. This observation and the small impact of eccentricity on insolation have made the role of the variation in eccentricity as the forcing factor of the glacial cycles a debated issue /Holmgren and Karlén, 1998/.

3 Sceanrios of past and future climate

Climatic driven changes such as the waxing and waning of ice sheets, extension of permafrost areas and shoreline displacements will essentially change surface environments, and most likely also affect subsurface conditions. In order to evaluate the evolution and safety of a deep geological repository for spent nuclear fuel we want to construct a scenario of future climate driven changes. *Climate change* refers to the long-term climate evolution related to variation in insolation due to orbital changes. The word *climate* is used in the sense "*gross climate state*". "*Cold climate*" refers to climate states with expanding or large global ice sheets, and "*warm climate*" refers to climate states with melting or small global ice sheets. 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. The mechanisms causing the shorter-term changes are not fully understood and they are not included in the scenario.

The radioactivity and thereby the radio-toxicity of the spent nuclear fuel decreases with time. After about 100,000 years the radioactivity of the spent nuclear fuel is comparable to the activity in the uranium ore once used to produce the fuel. We are therefor mainly concerned with the evolution the next 100,000–150,000 years, but the evolution in even longer time perspectives may also be of interest.

Geological evidence shows that the changes we focus on have the same periodicity as the orbital parameters. Even though uncertainties still exist we base our scenario on the astronomical climate theory. To construct scenarios we rely on models. Results from three different models have been regarded. The models and their predictions of the climatic changes for the past 200,000 years and future 150,000 years are presented below.

3.1 Models and model predictions

The referred models are all based on the astronomical climate theory. The modelled climatic changes are related to the predictable variations of the orbital elements. To illustrate the connection between climate and the orbital parameters the climate as depicted in $\delta^{18}O$ variation in deep-sea sediments (the SPECMAP record) are shown together with the variations in June insolation at 65°N, *Figure 3-1*. *Figure 3-2* shows the June insolation at 65°N for the next 150,000 years. The general pattern of changes in insolation is the same for all months and latitudes. The June insolation at 65°N was chosen for the illustration since it is often used when comparing orbital variations to climate change, especially when regarding the growth and decay of ice sheets /Berger et al, 1996/.

Regarding the recent geological past, isotopic stage 7 about 244,000–190,000 years before present, had a maximum amplitude in insolation changes due to large values in eccentricity and maximum variation of obliquity. During the period from 140,000–70,000 years before present, including isotopic stage 5, the amplitudes were almost as large. Isotopic stage 3 show small changes in insolation, the June values for 65°N are slowly decreasing towards a minimum followed by an increase of quite large amplitude during stage 2. The June insolation at 65°N reached a maximum in the beginning of stage 1 and is currently decreasing.

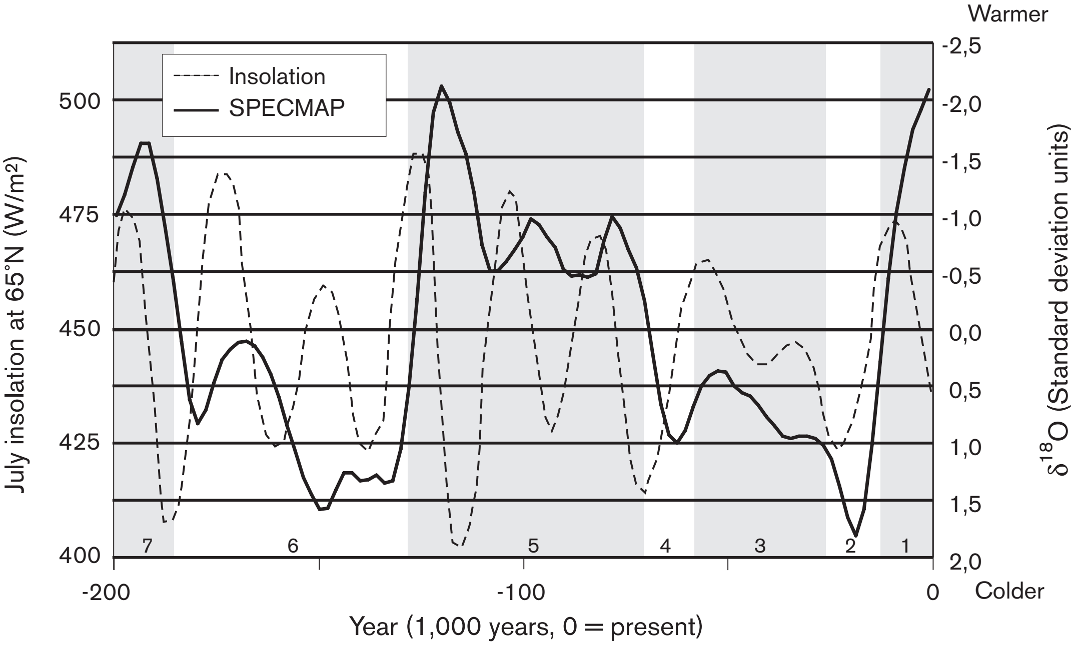


Figure 3-1. *The SPECMAP proxy record of global ice volume and the June insolation at 65ºN for the past 200,000 years. (SPECMAP revised from /Imbrie et al, 1984/, insolation based on /Berger and Loutre, 1991/. The numbered grey and white fields are the isotopic stages according to /Emeliani, 1955/. The ages for isotope stage boundaries are from /Martinson et al, 1987/.*

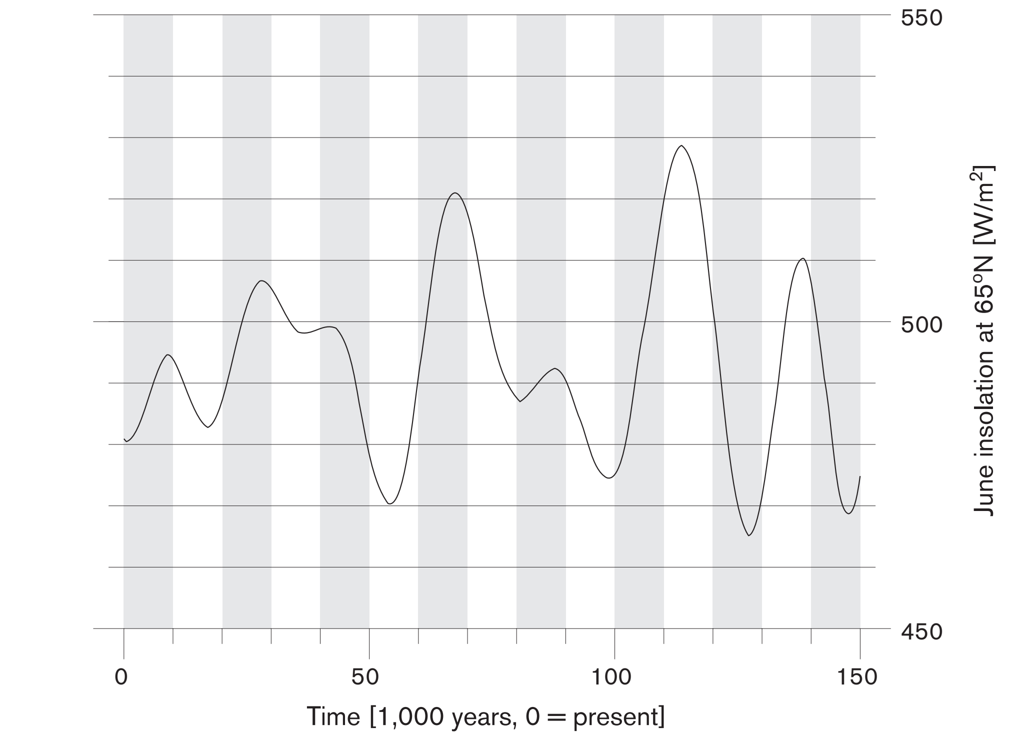


Figure 3-2. *June insolation at 65ºN for the next 150,000 years. (Based on /Berger, 1978a, Berger, 1978b/ graph revised from /Berger and Loutre, 1997/).*

For the future, the small insolation changes that characterise the next 50,000 years are calculated to be exceptional over the last 3,000,000 years /Berger et al, 1996/. The changes in June insolation at 65ºN during the period 50,000–110,000 years into the future are quite similar to the variations seen from the beginning of stage 4 to the end of stage 2.

Regarding the climate changes as depicted in the SPECMAP record it seems like changes in insolation act as a pacemaker of the ice ages as suggested by Hays et al /Hays et al, 1976/. Increased June insolation at 65ºN is followed by increased δ^{18}O-values and decreased insolation is followed by decreased isotopic values. The periodicity of change is similar while the amplitudes differ.

The SPECMAP record and the June insolation at 65ºN are used as a basis for an arbitrary comparison and judgement of the model results. Note that all the models have to some extent been calibrated against the geological evidence during the last glacial-interglacial cycle (the Weichselian). Also note that the SPECMAP time scale is tuned to match the orbital parameters in the frequency band. The variables used to depict the climatic changes are different in the different models. If all variations had been expressed in standard deviation units the comparison between models would have been facilitated, but such data processing has not been accomplished.

3.1.1 An astronomical climate index

1981 Kukla and others presented an astronomical climate index (ACLIN) /Kukla et al, 1981/. ACLIN is designed to predict the major climate changes in the late and middle Pleistocene and in the near future.The index is based on the observation that a specific orbital configuration – high obliquity combined with June perihelion – marked the beginning of the past three interglacials.

Radiometrically dated evidence of climatic changes from three sets of proxy records – pollen, sea level and δ^{18}O – were used to create a climate severity index for the past 130,000 years. Within each group of records data from several sites were included. The data were transferred to a scale were 100 units correspond to the last glacial maximum (15,000–20,000 years ago) and 200 units correspond to the peak present interglacial (about 6,000 years ago). The index was averaged arithmetically, first within each group and second among groups.

Variations with a wavelength of at least 10,000 years were analysed. The timing and climate indices for their peaks and troughs were settled. A starting time (entry) for each episode was defined as half way between neighbouring highs and lows. The climate indices of the peaks and troughs were plotted as a function of contemporary values of the orbital parameters and as a function of the orbital signatures of their entries. For the current and last interglacials (peaks at 6,000 years and 123,000 years before present) specific orbital characteristics were identified. They start at a time of June perihelion, reach a climax at the time of September perihelion and the obliquity is high throughout the first half of the interval.

Based on the above observation an index predicting the gross climate state as a function of ε, ω and e were formulated. This astronomical climate index (ACLIN) should be high when ε and e is high and the Earth passes perihelion in September (ω is close to 360º). The approach lead to the following, purely empirical formula:

$$\alpha_1 = \left|\frac{\omega_1 - 180}{90}\right| + \varepsilon_2 - 22 + 500e_1^2 \qquad (3\text{-}1)$$

where α_1 is the climate index at time t_1, ω_1 the longitude of perihelion at time t_1 (measured as the angular distance from the autumnal equinox), e_1 the eccentricity at time t_1 and ε_2 is the obliquity at time t_2. The time t_2 is defined by $\omega_2 = \omega_1 - 90$ and $|t_1 - t_2| = \min$, which corresponds to a time lag of ~5,000-6,000 years.

The value of the index varies between 0 and 6; the corresponding climate states are accounted for in Table 3-1. The simple ACLIN formula reproduces the principal features of the palaeoclimate record very well. This is interpreted as an evidence for the dominant role of orbital variations in shaping the climate. *Figure 3-3* shows the ACLIN index and the SPECMAP record for the past 200,000 years, and *Figure 3-4* shows the index and the June insolation at 65°N for the next 150,000 years.

Table 3-1: ACLIN index and climate states

Climate state	ACLIN index
Interglacial	≥ 4.3
Temperate interstadial	≥ 3.5 – < 4.3
Interstadial	≥ 2.5 – < 3.5
Stadial and protostadial	< 2.5

The pattern of change of the ACLIN index is very similar to the changes in insolation. This is not surprising since they are both functions of e, ε and ω. For the geological past the ACLIN index reproduces the intensity and timing of mild events well while cold events are not reproduced as well. Comparing the ACLIN index to the SPECMAP record for the Weichselian shows that the cold ACLIN peaks in the beginning of the cycle have greater amplitude than the corresponding $\delta^{18}O$ peaks.

Regarding the ACLN prediction of the future climate the next period with interglacial conditions is expected to occur in about 120,000 years. Two periods with "temperate interstadial" conditions are predicted during the intervening glacial period, the first around 30,000 years into the future and the second around 72,000 years into the future. Looking at the past the Weichselian also include two ACLIN periods of "temperate interstadial" conditions, namely stage 5c and 5a, both correspond well to the contemporary $\delta^{18}O$ record. If periods of cold conditions are regarded, 5 periods with "stadial and protostadial" conditions are predicted during the next glacial cycle. The first peaks at about 5,000 years into the future, the second at about 23,000 years, the third at about 38,000 years and the fourth at about 60,000 years into the future. The fifth period of "stadial and protostadial" conditions include two peaks, one at about 83,000 years and the second at about 108,000 years into the future. The ACLIN prediction of the Weichselian includes three periods with "stadial and protostadial" conditions, they occur during stage 5d, 4 and 2. They all correspond to cold peaks in the SPECMAP record. In relation to the SPECMAP record the ACLIN index indicates a colder climate.

In spite of the week insolation forcing the ACLIN model predicts colder conditions during the next glacial cycle than during the Weichselian. An explanation to this could be that the ACLIN index does not include the relation between eccentricity and precession. The reduced amplitude of insolation changes caused by low values of e is not reproduced in the climate index. Considering this and the colder condition ACLIN predicts in relation to SPECMAP for the early Weichselian it is presumed that the next 4 cold peaks are not so severe as indicated by the index value. The warm periods are though presumed to occur as predicted.

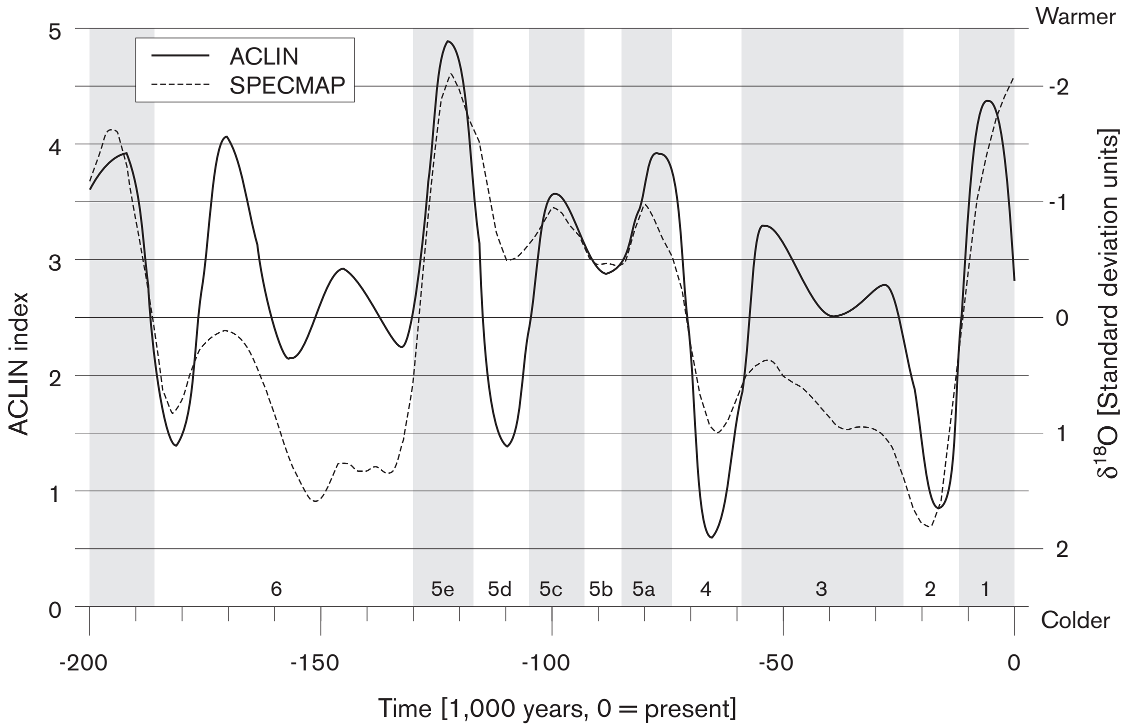


Figure 3-3. *The ACLIN index and the SPECMAP record for the past 200,000 years. (Revised from /Kukla et al, 1981, Imbrie et al, 1984/). Numbered grey and white fields ares isotopic stages (see fig. 3-1).*

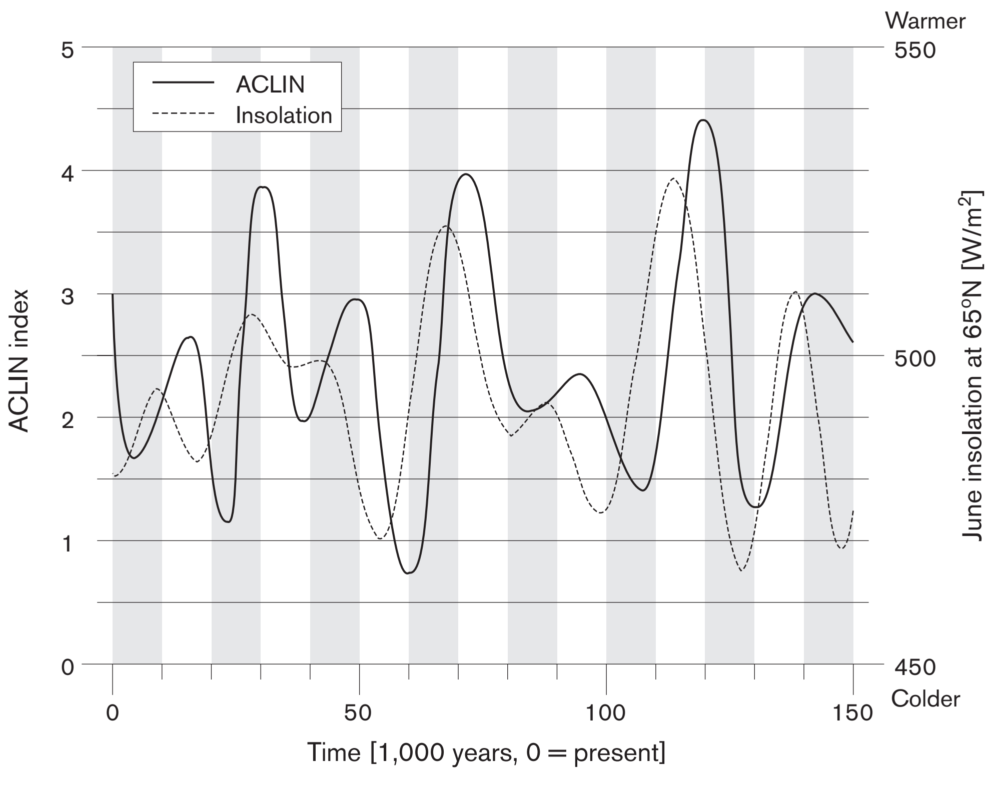


Figure 3-4. *The ACLIN index and the June insolation at 65°N for the next 150,000 years. (Revised from /Kukla et al, 1981, Berger and Loutre, 1997/.)*

3.1.2 The Imbrie and Imbrie model

Imbrie and Imbrie /8/ have developed a differential model describing the relation between orbital variations and global ice volume. By using models the physical mechanisms by which the climate system responds to orbital forcing can be investigated. One approach to achieve this is to include available knowledge of climate physics and develop a model that is as realistic as possible. An alternative strategy – applied by Imbrie and Imbrie – is to attempt to capture the essential features of more complex models in a class of simple models.

Imbrie and Imbrie concluded that a differential model of the form $dy/dt = f(x,y)$ (where x is the orbital condition, y is the climate state and f a system function relating y to x) is required in order to include the time-dependent behaviour of the system, including the observed lags between orbital forcing and climate response. In order to be able to tune the model against geological evidence a minimum of complexity was desired. Model complexity was defined to be the number of adjustable parameters included in the model. The parameters can occur both in the model input and in the system function.

Linear combinations of irradiation curves at various latitudes and seasons are a natural choice of input to a simple model of climate change. As any insolation curve is a linear combination of the three orbital elements, three parameters are required to scan all possible inputs. If only the shape of the input curve is important and the scale is arbitrary the number of parameters can be reduced to two. This fact was applied and the model input was defined as a single function of time:

$$x = \varepsilon + \alpha e \sin(\omega - \Phi) \tag{3-2}$$

where ε, e and ω are the orbital elements and α and Φ the adjustable parameters. Φ controls the phase of the precession effect and is linked to the season of the insolation curve that x would represent. α controls the ratio of precession and obliquity effects and is related to the latitude of insolation curve that x would represent.

When designing the system function (f) physical principles and the climatic record were kept in mind, and as for the input a minimum of complexity was sought. In a first approximation the system was assumed to be linear. The resulting simulation of the real isotopic curve was poor, for instance non of the periods associated with eccentricity were included. To extract eccentricity frequencies some form of nonlinearity is required. One observed source of non-linear behaviour is the tendency for ice sheets to shrink faster than they grow. This observation lead to a simple non-linear model including different time constants (T) depending on whether the climate is warming (T_w) or cooling (T_c). The system equation was written as:

$$\frac{dy}{dt} = \begin{cases} \dfrac{1+b}{T_m}(x-y) & if\ x \geq y \\ \dfrac{1-b}{T_m}(x-y) & if\ x \leq y \end{cases} \tag{3-3}$$

where T_m is the mean time constant and b is a nonlinearity coefficient ($0 = b = 1$). Large values of b correspond to large values of T_c / T_w.

The model was run with different values on the four adjustable parameters (α, Φ, b and T_m) and compared with isotopic data of the past 500,000 years. The isotopic data was derived from deep-sea cores. The model was fine-tuned against certain features of the geological record of the past 150,000 years. The tuning indicated an optimum model in the neighbourhood of the following values:

T_m = 17,000

b = 0.6 (corresponding to T_c = 42,500 and T_w = 10,600 years)

α = −2 (corresponds to 65°N)

Φ = 2000 years (corresponds to July, 2π corresponds to 22,000 years)

The model simulation of the past 150,000 years is quite good, between 150,000 and 250,000 years most features are reasonably well simulated, back to 350,000 there are significant correlations but beyond that the correlation with the geologic record diminishes significantly. Of the past four sharp changes in the orbital record proceeding the interglacial periods the last three are well simulated. The model's ability to simulate the relative magnitude of main peaks and valleys decreases systematically with age. Comparing the spectrum of the model output to the spectrum of the geological record shows that the model has too much power at the 413,000 year eccentricity period and too little at the 100,000 period. The failures are related to a shortcoming in the generation of the 100,000-year periodicity. The model output compared to the SPECMAP record of the past 200,000 years is shown in *Figure 3-5*. The model prediction of the next 100,000 years is shown in *Figure 3-6* together with the June insolation at 65°N.

The model has been calibrated against the last 130,000 years, and the reconstruction of the Weichselian correlates very well to the SPECMAP record. Note that both model results and the SPECMAP record are expressed in standard deviation units. Regarding the Weichselian, the large amplitude variations in insolation during stage 5 correspond to larger variations in the model results than in the SPECMAP record. The maximum peak in insolation in the beginning of stage 1 does not make the modelled ice sheets melt away to the same extent as the $\delta^{18}O$ record suggests they did.

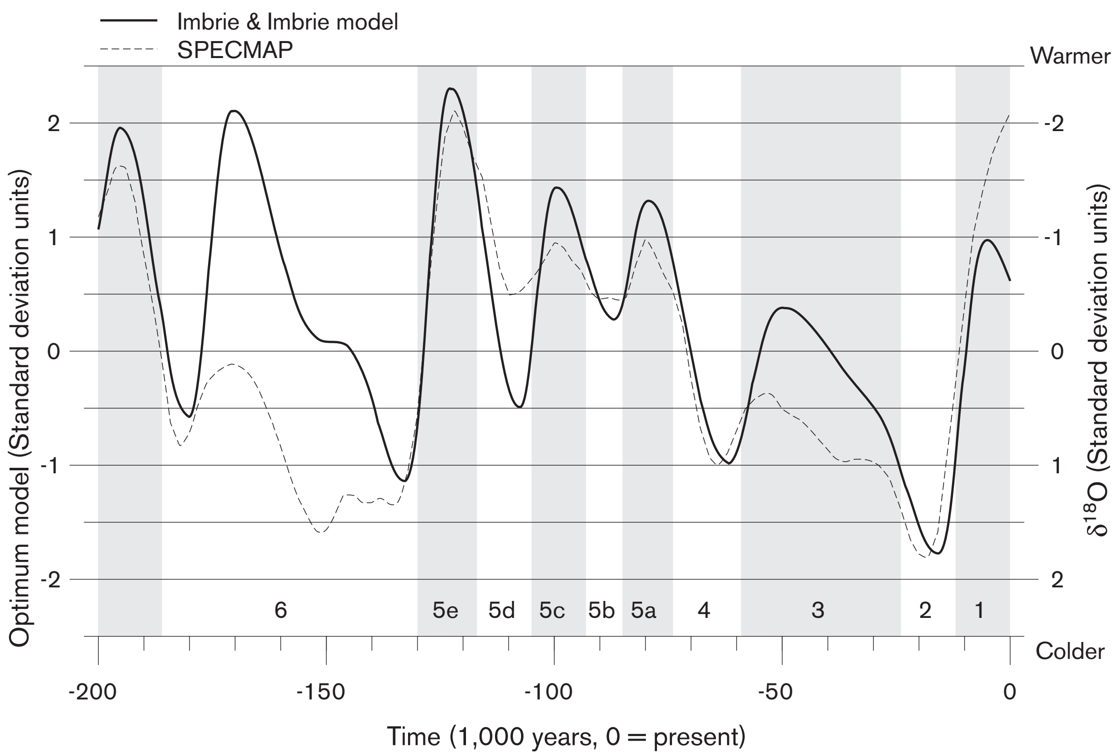


Figure 3-5. *Results from the optimum model compared to the SPECMAP record of the past 200,000 years. (Revised from /Imbrie et al, 1984, Imbrie and Imbrie, 1980/). Numbered grey and white fields are isotopic stages (see figure 3-1).*

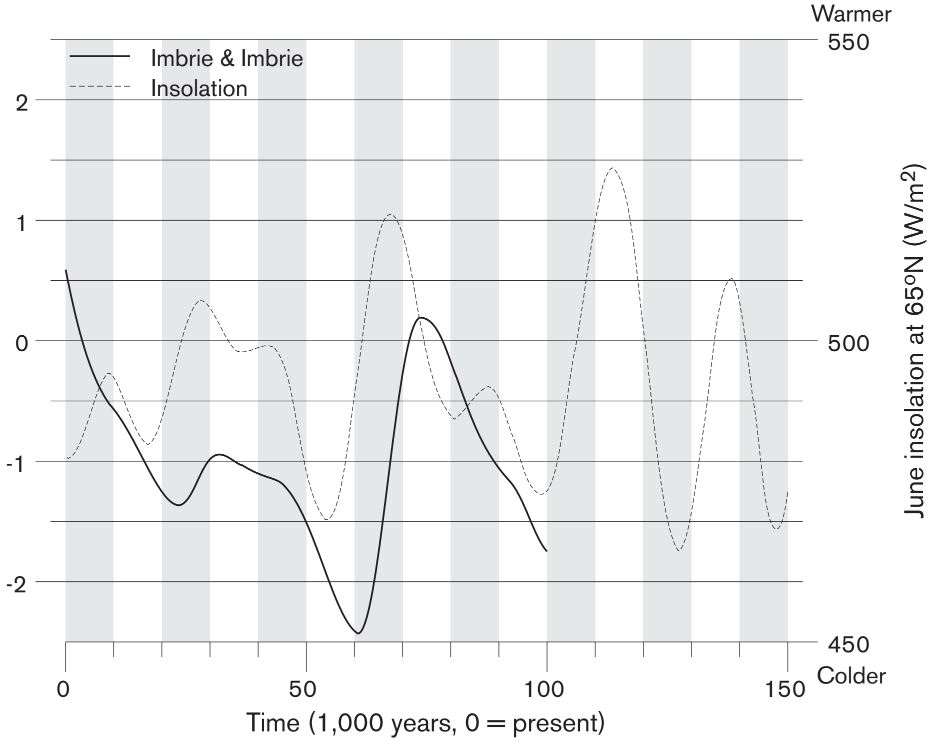


Figure 3-6. *Results from the optimum model for the next 100,000 years, and the June insolation at 65°N the next 150,000 years. (Revised from /Imbrie et al, 1984, Berger and Loutre, 1997/.)*

Regarding the future, neither the present low magnitude minimum in insolation or the following maxima can be seen as peaks in the model prediction. For this period the model predicts gradually colder conditions, ending with a cold maximum at about 23,000 years into the future. The cold peak is followed by an about 10,000 years long period of slightly warmer conditions. After this follows a transition to more severe conditions ending with a cold peak at about 60,000 years into the future. The magnitude of this peak is greater than the magnitudes of both the Weichselian and Saalian maxima. After this cold event the climate gets warmer and reaches a warm maximum at about 73,000 years, then the climate gets colder again.

As the ACLIN model the Imbrie & Imbrie model predicts colder conditions during the next glacial cycle than during the Weichselian in spite of the week insolation forcing. This may be a consequence of the model's incapacity to reproduce the current interglacial.

3.1.3 A coupled sectorially averaged climate-ice sheet model

A sectorially averaged seasonal model for simulating the response of the climate system to changes in insolation has been developed at the University of Louvain-la-Neuve (the LLN model) /Berger et al, 1996, Gallée et al, 1991, Gallée et al, 1992, Boulton and Curle, 1997/. The model simulates the Northern Hemisphere climate. To simulate the waxing and waning of ice sheets as a response to the climate changes a model of the Northern Hemisphere ice sheets and their underlying bedrock has been asynchronously coupled to the climate model.

The LLN model is a two-dimensional climate model. It is latitude, altitude and time dependent and simulates the seasonal cycle of Northern Hemisphere climate. The model domain has a resolution of 5° in latitude. Within each latitude band the surface is divided into at most five surface types, these are ice-free ocean, sea ice cover, snow covered or snow free land and ice sheet. Each surface type interacts separately with the sub-surface and atmosphere. A zonally averaged quasi-geostrophic model represents the atmospheric dynamics. A variable depth and temperature mixed-layer model represents the upper-ocean. Sub-grid scale mechanisms eliminated by the zonal average are taken into account by parameterisations. At the top of the model solar radiation and emitted infrared fluxes contribute to the net energy flux available to the system.

Surface and subsurface processes that are included in the LLN model are; precipitation, evaporation, vertical heat fluxes, surface albedo, oceanic heat transport and oceanic mixed-layer dynamics. Potentially important processes that are not taken into account are variations in water vapour transport, cloudiness, atmospheric dust content and deep-water circulation. Variations in insolation force the model. As the model does not include a carbon cycle, the carbon dioxide (CO_2) concentration in the atmosphere is also considered as an external forcing. The LLN model has been used to simulate the present climate of the Northern Hemisphere and it reproduces the main climatic characteristics well /Gallée et al, 1991/.

In the ice sheet model the evolution of the ice sheet within each latitude band is determined by a vertically integrated equation of mass (or volume) conservation. The change in ice sheet thickness is a function of instantaneous mass balance, ice mass discharge and ice velocity. The model resolution is 0.5° latitudes. The vertically integrated ice velocity is deduced from the generalised flow law and from the equations of momentum balance. The ice mass discharge is parameterised and taken to be proportional to the ice sheet altitude at the centre of the ice sheet and in inverse to the square of the longitudinal extent. Assuming that the ice behaves in a perfect elastic manner the east-west profile of the ice sheet will be parabolic. Its longitudinal extent within the latitude band can be calculated from the altitude at the centre of the ice sheet and the yield stress. The yield stress is unique for each ice sheet and calibrated over the ice sheet size at the last glacial maximum. The deflection of the underlying bedrock is computed using a time-dependent diffusive equation of the asthenosphere along the latitude /Gallée et al, 1992, Berger and Loutre, 1997/.

Three ice sheets, the Greenland, the North American and the Eurasian are assumed to occur in the Northern Hemisphere. The extension of each ice sheet is calculated separately. The ice sheets longitudinal (north – south) profile is prescribed, it is symmetrical from the ice sheet crest. The ice sheets are not allowed to expand further than the underlying continents. If an ice sheet extends into the ocean lateral calving is assumed to occur. The position of the ice sheet crest is artificially displaced in such a way that one of its longitudinal edges is always located at the coast.

The climate and ice sheet models are asynchronously coupled. The climate model is run to reach quasi-equilibrium given the prevailing insolation and CO_2 value. The yearly net snow mass balance is computed for each latitude band. The computed snow mass balances[2] serve as an input to the ice sheet model, which is run separately for each ice sheet over a time interval of 1000 years. The new ice sheet extents and altitudes change the surface conditions in the climate model. A new net snow balance is calculated for the

[2] The net mass balance is the difference between the local snow precipitation and ablation. The ablation is computed from the balance of heat fluxes at the snow or ice surfaces.

time t + 1000, using the updated surface conditions and possibly new insolation and CO_2 values. The climate and ice sheet models have different latitude resolution, 5° and 0.5° respectively. The ice sheet extents and altitudes used in the climate model are the average from the ice sheet model, and the snow mass balances computed in the climate model are interpolated on the ice sheet model grid /Gallée et al, 1992/.

Comprehensive descriptions of the climate, ice sheet and coupled models and the results of various sensitivity tests are presented in /Gallée et al, 1991 and Gallée et al, 1992/. Some of the conclusions from these studies are that:

- Thanks to parameterisations and corrections of its zonal character the climate model reproduces the prevailing climate conditions well.
- The net mass balance variations are dominated by ablation variation and highly correlated with June insolation variations.
- Ablation variations are affected by snow ageing and parameterisations of surface processes like albedo and temperature variations.
- Forced by changes in insolation alone the coupled model reproduces the geological record of ice sheet volume quite well.
- The reproduction of the geological record is improved when CO_2 variations are added.

Model results for a simulation with a constant CO_2 concentration of 210 ppmv and a case with varying CO_2 values derived from the Vostok core /Jouzel et al, 1993 in Berger et al, 1996, Berger et al, 1996 and Berger and Loutre, 1997/ are shown in *Figure 3*-7. A CO_2 content of 210 ppmv corresponds to cold glacial conditions, simulations using a higher CO_2 content show poor correlation with geological evidence /Berger et al, 1996, Berger and Loutre, 1997/. *Figure 3*-7 also shows the SPECMAP record.

The coupled climate ice sheet model has been used to predict future conditions. The results from two simulations, one with constant CO_2 and the other using forecasted CO_2 variations /Boulton and Curle, 1997/, are shown in *Figure 3-8*. *Figure 3-8* also shows the June insolation at 65°N.

When comparing the model results to the SPECMAP record it is important to keep in mind that the LLN-ice sheet model simulates the conditions in the Northern Hemisphere, while the SPECMAP record is a proxy for the global ice volume. The modelled magnitude of change depends on the CO_2 concentration. Sensitivity studies comparing the model response to variations in insolation and reconstructed CO_2 variations, show that variations in insolation is the dominant forcing factor /Gallée et al, 1992/. If the model is driven by variations in insolation only and the CO_2 concentration is kept constant, a low CO_2 concentration is required in order to reproduce past conditions /Berger et al, 1996/. High CO_2 concentrations generate too small ice volume. The sensitivity of the simulated ice volume to the atmospheric CO_2 concentration is not constant in time. Sensitivity analysis imply that the climate system is more sensitive to the CO_2 concentration when the magnitude of insolation minima is small /Berger et al, 1996/.

The model reproduces the Weichselian cycle well. In the simulation using the CO_2 record from Vostok the quite moderate response of the $\delta^{18}O$ record to the large amplitude variations in insolation during stage 5 are well captured.

For the future, regarding the simulation using forecasted CO_2 concentrations the low magnitude variations in insolation the next 10,000 years are not reproduced as peaks, the climate is only getting slightly colder. A low magnitude cold peak occurs at about 23,000 years, but the Northern Hemisphere ice sheets will not start to grow significantly

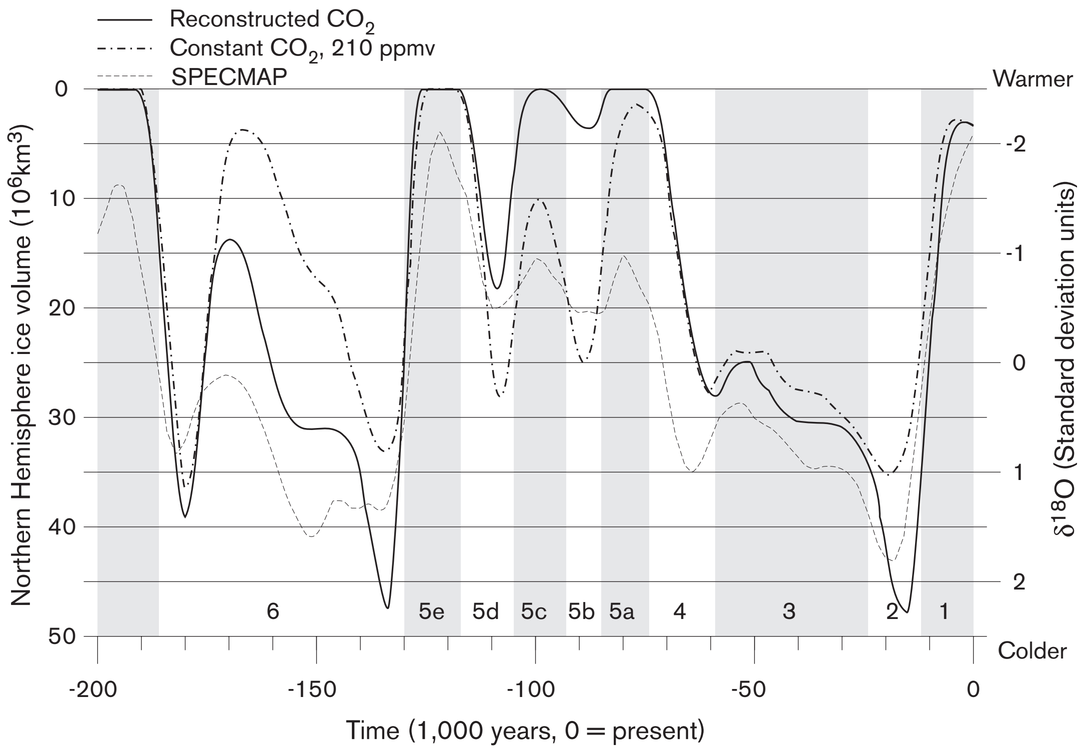


Figure 3-7. *Results from the LLN model asynchronously coupled to an ice sheet model, and the SPECMAP record. The simulated ice volume is shown for a case with constant CO_2 content and a case where the CO_2 content is varying according to a reconstruction from the Vostok core /Jouzel et al, 1993/. (Revised from /Imbrie et al, 1984, Berger and Loutre, 1997/.) Numbered grey and white fields are isotopic stages (see figure 3-1).*

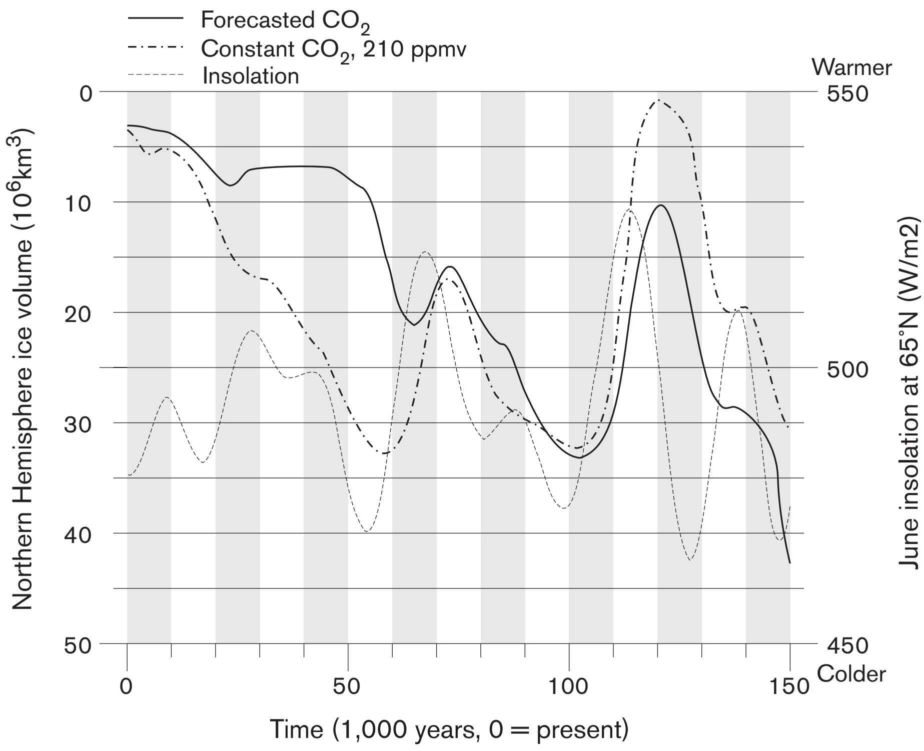


Figure 3-8. *The LLN-ice sheet model forecast for the next 150,000 years, and the June insolation at 65°N.*

until about 50,000 years into the future. The ice volume reaches maxima at about 65,000 years and 102,000 years. During the intervening period a moderate decrease in ice volume is followed by ice sheet growth. A warm peak occurs at about 120,000 years, but the modelled ice volume does not correspond to interglacial conditions.

In the simulation using a constant CO_2 concentration, the ice sheet growth during the next 50,000 years is larger. A maximum occurs at about 5,000 years into the future. There is no maximum at 23,000 years instead the ice sheet grows continuously towards a maximum. The maximum is reached at about 58,000 years, that is earlier than in the case with varying CO_2 concentration. The subsequent warming is larger, but the following cold maximum is comparable to the case with varying CO_2 concentration. A warm maximum with interglacial conditions is reached at about 120,000 years.

3.2 Climate during the Weichselian

For the reconstruction of the climate conditions during the Weichselian the above models are not utilised. Instead the scenario of past climate changes is based on the SPECMAP record and a reconstruction of the Scandinavian ice sheet made by Mangerud /1991/. Mangerud's reconstruction is based on a compilation of various geological evidences of past glaciations in Scandinavia. It is shown in *Figure 3-9* together with the SPECMAP record.

The climate scenario for the Weichselian is summarised in Table 3-2 and illustrated in *Figure 3-10*.

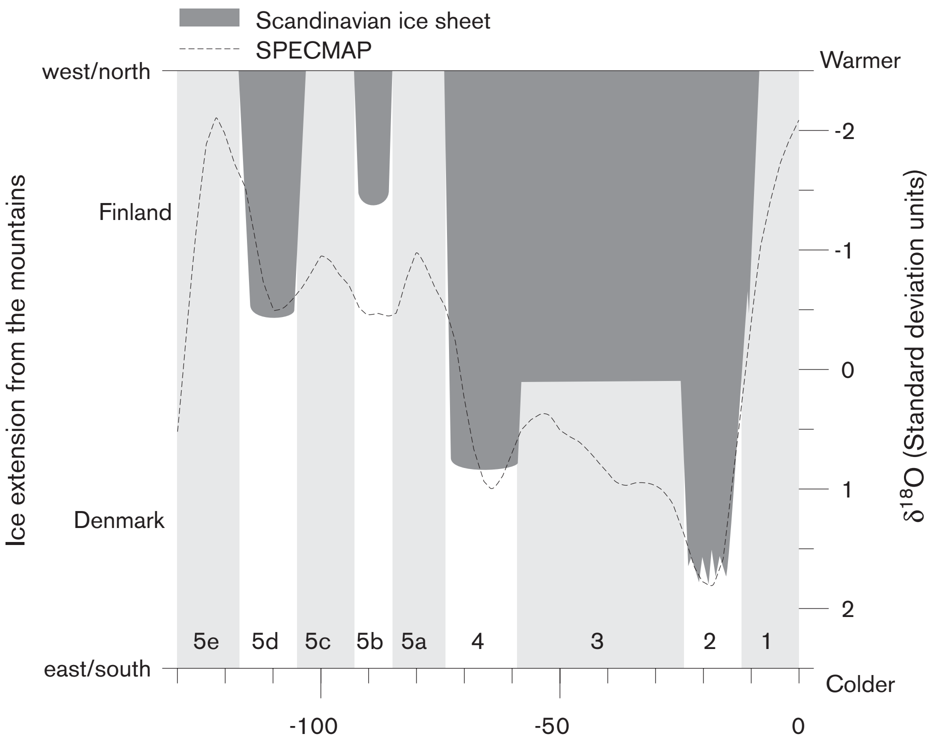


Figure 3-9. *Schematic graph for the Scandinavian ice sheet during the Weichselian according to Mangerud, and the SPECMAP record. The schematic graph shows the ice sheet extension in an arbitrary scale, from the mountains towards the east (north Sweden and Finland) and south (south Sweden and Denmark). (Revised from /Imbrie et al, 1984, Mangerud, 1991/.)*

Table 3-2. Climate and ice sheet scenario for the Weichselian

Period [1,000 years ago]	Climate and ice sheet
130 – 120	Intergalcial.
120 – 100	Gradually colder, at about –115,000 years an ice starts to grow in the mountains, it reaches its maximum extent at about –108,000 years and covers all the country.
100 – 90	Interstadial, the ice sheet melts except for ice caps in the mountain area.
90 – 80	The climate gets colder again, and the ice sheet expands, at its maximum as far as to northern Skåne in the south and western Finland in the east.
80 – 70	A warm period, the ice sheet melts except for ice caps in the mountain area.
70 – 30	A long cold period with variable climate, the ice sheet covers large parts of the country during the whole period, in the south the extension varies from the current lake Mälaren to northern Skåne, and in the east from the current Swedish Baltic coast to central Finland.
30 – 20	The coldest period of the glacial cycle, the ice sheet reaches its maximum extension, it extends to northern Germany in the south and far in to Russia in the east.
20 – 10	The climate gets warmer, in spite of cold drawbacks the ice sheet gradually melts away.
10 – 0	Interglacial.

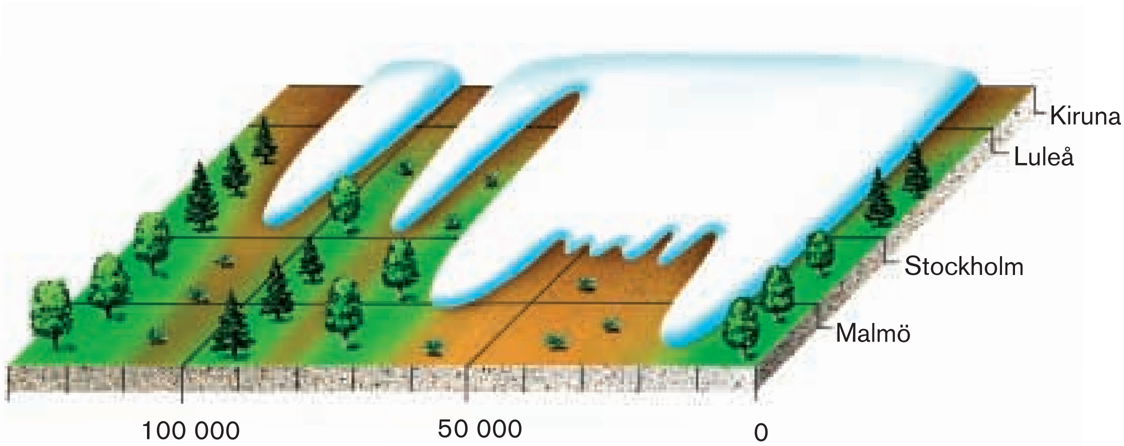


Figure 3-10. *Sketch showing the assumed ice sheet extension in Sweden during the Weichselian.*

3.3 A climate scenario for the next 150 000 years

A scenario for next 150 000 years describing the evolution of the climate and related environmental changes has been constructed. The time period 150 000 years has been chosen because it is the approximate time period required encompassing an interglacial-glacial cycle.

The scenario is based on geological reconstructions of past conditions and available model predictions. The intention has been to outline how climate conditions in Sweden may vary during the next glacial cycle. It must be emphasised that the scenario is not a prediction of the future but merely a description of an evolution similar to one that may occur. The scenario can also be seen as a general description of the evolution during any of the glacial cycles of the Quaternary. The scope of climate related changes are such that the impact on repository performance is not underestimated.

3.3.1 The gross climate states

Results from the models described in chapter 5 have been used when constructing a climate scenario for the next 150,000 years. ACLIN is a very simple empirical model that bypasses the search for solar-terrestrial links in the climate system and directly correlate orbital perturbations to climate as depicted in various geological records. The Imbrie & Imbrie model includes a simple model of the non-linear response of the climate system to orbital forcing. In the coupled LLN-ice sheet model the climate system, its response to orbital forcing and changes in CO_2 concentration is simulated in a relatively simple two-dimensional model and linked to the growth of Northern hemisphere ice sheets. The model results have been used to construct a scenario of the gross climate states during different periods.

Common for all the models is that they predict cold maxima between 20,000–30,000 years, 60,000–70,000 years and 100,000–110,000 years. (The Imbrie & Imbrie model only covers the next 100,000 years.) The LLN-ice sheet model simulation with constant CO_2 concentration does not have a cold peak between 20,000 and 30,000 years, but the case with higher, varying CO_2 concentration is judged to be more credible for this period. The ACLIN model predicts more cold maxima than the three mentioned, these are assumed to be a consequence of the simple model structure. When the inertia of the climate response is taken into account they are assumed to be insignificant. Consequently three cold peaks occurring between 20,000–30,000 years, 60,000–70,000 years and 100,000–110,000 years are predicted for the next 150,000 years.

Of the three cold peaks the one occurring between 100,000 and 110,000 is assumed to be the most severe. This assumption is based on the following:

- The variation in insolation between 70,000–100,000 years into the future is similar to the variation proceeding the Weichselian glacial maxima.
- Geological evidence suggests that the cold maxima in the end of glacial cycles give rise to larger ice sheets than the ones occurring at the beginning of the cycles.
- Results from the LLN-ice sheet model (the case with varying CO_2 concentration).

Of the two other maxima the one occurring between 20,000–30,000 years is assumed to be the mildest, based on the following:

- The insolation forcing is weak.
- Results from the LLN-ice sheet model (the case with varying CO_2 concentration).

Warm periods are predicted between 30,000–40,000 years, 70,000–80,000 years and around 120,000 years into the future. The LLN-ice sheet model case with varying CO_2 concentration predicts a long period of quite warm conditions between about 30,000–45,000 years, during which the northern hemisphere ice volume is almost constant. In the case with constant CO_2 concentration the peak in insolation is only depicted as a temporary stop in a continuos ice sheet growth. As for the cold conditions ACLIN predicts more peaks, but the three mentioned above are the warmest ones. ACLIN predicts interglacial conditions for the peak around 120,000 years.

Of the three warm periods the one peaking around 120,000 years is assumed to be the warmest, this assumption is based on the following:

- The ACLIN index and the fact that it predicts warm periods well.
- The corresponding peak in insolation is similar to the one before the present interglacial.

The other two warm periods are assumed to be quite similar. The ACLIN index suggests they were similar. According to the Imbrie & Imbrie model the peak between 70,000 and 80,000 years is warmer with less ice than the 30,000–40,000 years peak. The LLN-ice sheet model on the contrary predicts a larger Northern Hemisphere ice sheet between 70,000–80,000 years than during the period between 30,000–40,000 years.

3.3.2 The Scandinavian ice sheet

Mangeruds reconstruction of the last interglacial-glacial cycle /Mangerud, 1991/, *Figure 3-9*, is used as a basis for the judgement of the extension of the Scandinavian ice sheet during the different periods accounted for above. It is assumed that the 150,000-year period encompasses a full interglacial – glacial cycle and that it will include a glacial maximum comparable to the Weichselian maximum.

The ice sheet extension during the 20,000–30,000 years stadial is assumed to be similar to the one Mangerud reconstructed for stage 5b or 5d. During the period 60,000–70,000 years the ice is assumed to extend as far as Mangerud suggests it did during stage 3 or 4. The maximum occurring between 100,000 and 110,000 years is assumed to give rise to an ice sheet extending as far as the Weichselian maximum. During the interstadial between 30,000 and 40,000 years the ice sheet is assumed to melt away except for ice caps in the mountains. The interstadial between 70,000 and 80,000 years is assumed to be almost as warm, but as the ice sheet is much larger at the beginning of this interstadial, more ice is assumed to remain. The warm period peaking at about 120,000 years is assumed to be an interglacial. This gives the scenario for the next 150,000 years accounted for in Table 3-3 and *Figure 3-11*.

Table 3-3. Climate and ice sheet scenario for the next 150,000 years

Period [1,000 years]	Climate and ice sheet
0 – 20	Gradually colder, an ice sheet is starting to grow in the mountains at about 5,000 years.
20 – 30	Stadial, the ice extends to the current Swedish Baltic coast in the east and to the lake Siljan in the south.
30 – 40	Interstadial, the ice sheet melts except for ice caps in the mountain area.
40 – 60	The climate is getting colder again and the ice sheet expands to the south-east.
60 – 70	Stadial, the ice sheet covers Finland and extends over the lake Vättern in the south.
70 – 80	Interstadial, the ice sheet melts away rather quickly to an extension that is a little bit less than the 20,000–30,000 years maximum.
80 – 100	A new stadial is initiated, the climate is getting colder and the ice sheet expands.
100 – 110	Stadial, the glacial maximum is reached, the ice sheet extends into Russia in the east and into northern Poland and Germany in the south.
110 – 130	Interglacial, the ice melts away and the climate gets similar to the present, the warm maximum is reached at about 120,000 years.
130 – 150	A new glacial period is initiated, the ice sheet grows and reaches almost as far as during the 60,000–70,000 years maximum at the end of the period.

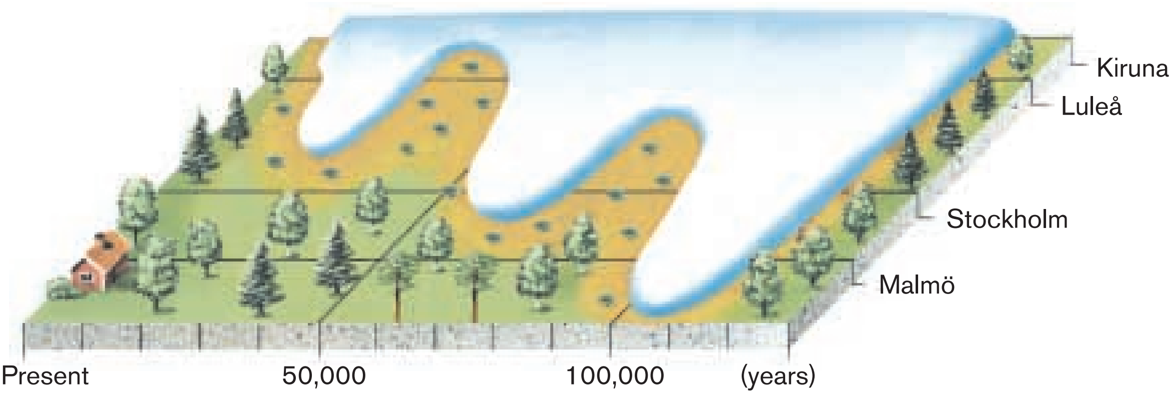


Figure 3-11. Sketch showing the assumed ice sheet extension in Sweden during the next glacial cycle.

4 Scenarios of shoreline development in Scandinavia

4.1 Background and objective

4.1.1 Objective

Cold climate and the presence of ice sheets will significantly alter the surface environment. Another climate induced environmental change with significant impact on surface conditions is shore level displacement. Whether an area is covered by the sea or not impact groundwater flow and groundwater composition and is of major importance for the biosphere. The occurrence of watercourses, lakes and bays also affects the possible development of permafrost.

Shore level displacement in a glaciated area is due to two interactive vertical movements; glacio-isostatic depression/uplift and global eustatic sea level lowering/rise. Shore level displacement is estimated by:

$$S = U - E \qquad (4\text{-}1)$$

where S = shore level displacement

U = glacio-isostatic depression/uplift

E = eustatic sea level lowering/rise

A prerequisite for making a credible modelling of past and future shore level displacement is thus a reasonable knowledge of glacio-isostatic and eustatic changes. Påsse (1996a, 1997) presented an empirical model of shore-level displacement in Fennoscandia during Late Weichselian and Holocene. The objective of the present project is to extend this model back in time; to include the whole Weichselian glacial and the Eemian interglacial, and into the future; to make a prediction of shore-level displacement during the next full glacial cycle.

4.1.2 Background

Empirical data from which the course of glacio-isostatic uplift can be estimated have recently been reported from investigations of lake tilting (Påsse 1990b, 1996b, 1998a). By magnifying the function, which describes the lake tilting, it was possible to start an iteration process, which gave mathematical expression for factors involved both with the isostatic and the eustatic component of shore level displacement. The model was tested by a comparison between shore-level curves based on empirical data and those predicted by the model. The results of the modelling are presented in two SKB Technical Reports (Påsse 1996a, 1997).

4.2 Isostatic component

The earth is from the center to the surface divided into inner and outer core, the mantle and the crust. Each part is composed of material having different properties. The crust is the outer most and the part of the earth, which we can observe. The crust is subdivided in continental crust and oceanic crust, which have quite different qualities. The thickness of the continental crust varies between 25 and 90 km, while the oceanic crust is much thinner, between 6 and 11 km. Two additional terms are used within the vertical division of the earth, the lithosphere and the asthenosphere. The lithosphere includes the crust and upper mantle and is considered to be cool and rigid. It can be defined as the layer of the earth cool enough to behave like a brittle solid. The asthenosphere is situated within the mantle and is located between the depths of 100 to 350 km and may extend down to 700 km. This part of the earth is capable of gradual flow. The material within the asthenosphere behaves like a solid under certain conditions and like a fluid under other conditions. Material of this type is generally described as exhibiting plastic behaviour. This layer is not able to store elastic energy like a brittle solid and is thus incapable of generating earthquakes. The asthenosphere may be very thin or absent below the older parts of the continental shield.

The isostatic movements of the earth crust corresponding to glacial loading and unloading, is primarily controlled by viscous behaviour of the asthenosphere while the lithosphere often can be regarded as an elastic layer. Despite that the main features of crustal movements are known the viscous flow within the asthenosphere in response to the glacial loading and unloading is not fully understood.

4.2.1 Viscous flow

Ekman & Mäkinen (1996) have by repeated high-precision relative gravity measurements proved that viscous flow of mantle material is a necessary part of isostatic movements. During glacial phases there is a horizontal viscous outflow of mantle material below the crust in ice covered areas. During interstadial phases this outflow is restored by an inflow. The reverse situation occurs below the oceanic crust where unloading of water during glacial phases provide areas of compensatory uplift (Bowlby). During interglacial and interstadial phases viscous flow is directed towards the continents to accommodate the subsiding oceanic crust due to the reloading of water masses. Mantle material is thus alternately transported from the areas beneath the continents to the areas benaeth the oceans and vice versa during the glacial-interglacial cycles, also see Figure 4-1.

4.2.2 Global isostatic development

During glacial phases water is stored in inland ices and the sea levels can be down to about 100 m below the present levels. The transport of water from the oceans gives rise to crustal uplift of the oceanic bed. This uplift can be in the order of 30 m. Coastal areas are simultaneously lowered as mantle material is transported from beneath the continents towards the area beneath the oceans when the oceanic bed rises. At the same time glaciated continents are depressed by the ice load. The glacial-isostatic depression will cause crustal rise in the surrounding areas, as mantle material is transported from the areas beneath the ice sheets. The course of crustal changes in the peripheral parts of an ice sheet are thus assumed to be different depending on whether the peripheral parts lie in continental, or oceanic coast environments. Continental environments exist to the east and to the south of a Scandinavian ice sheet. Oceanic coast conditions prevail to the west and to the north. These environmental differences make the modelling of shore level displacement in Scandinavia a complex issue.

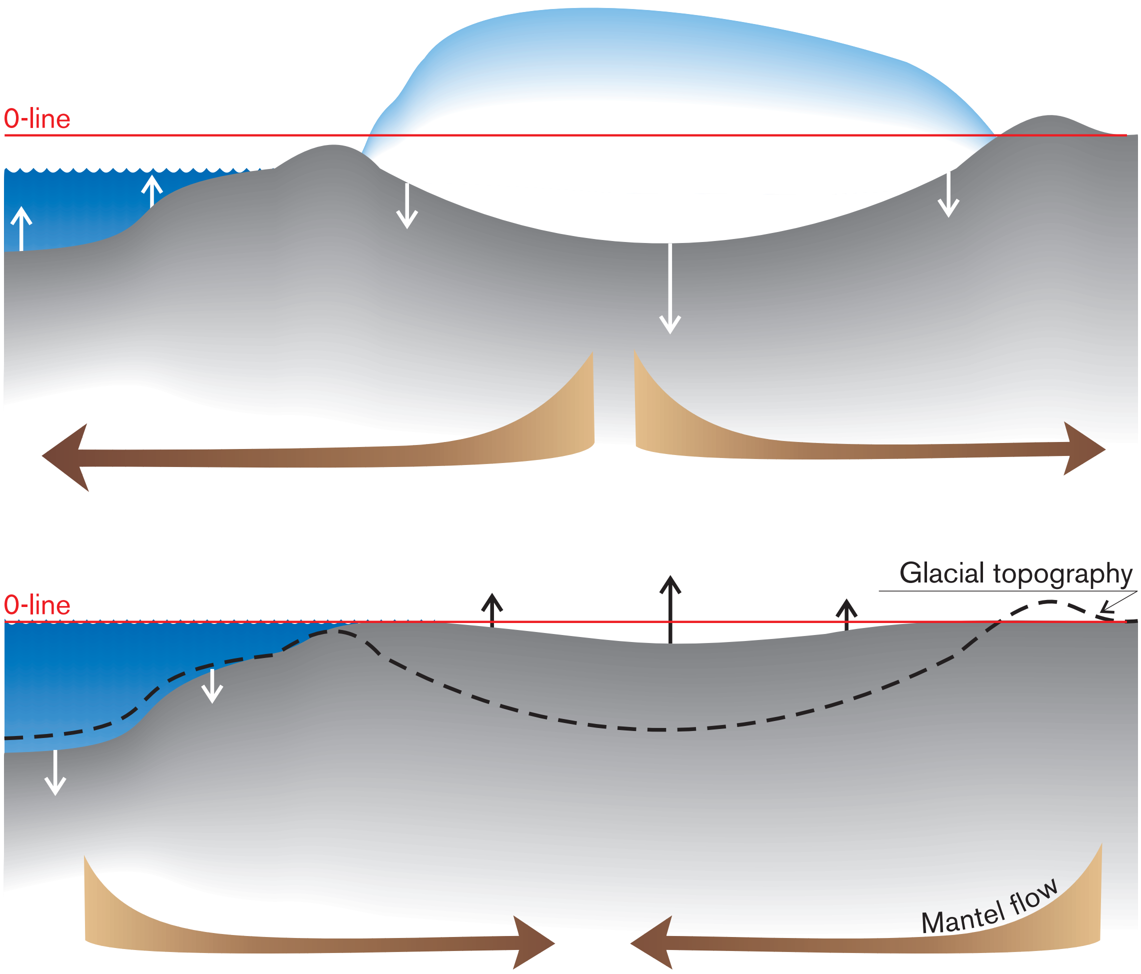


__Figure 4-1.__ Schematic illustration of crust movements and the viscous flow of mantle material beneath the crust.

4.2.3 The shape of the depression of the crust produced by the ice load

There are several models describing the isostatic mechanism. According to Fjeldskaar & Chatles (1991) these can conceptually be dived into two categories: the bulge models and the punching models. In the bulge models the loading produces large peripheral accumulations of mantle material squeezed from the loaded areas through a viscous channel. In the punching models, or deep flow models, the loading or unloading drags the peripheral regions with them in a sympathetic motion. With exception of the distal area the equilibrium depression produced by the load of a large ice sheet will not differ much between the two hypotheses. Close to the ice edge large divergences, depending on the strength of the lithosphere, are possible. McGinnis (1968) has suggested that the stresses in the lithosphere caused by the load will produce a proglacial depression far beyond the limits of the load but also an elastic upward bending of the crust, a forebulge, further out from the ice edge.

Contradictory models of the isostatic mechanism may very well predict similar behaviour of the central part of the depressed area. However, the differences are of vital importance for the predicted crustal movements in peripheral parts, especially for the area outside the ice sheet.

Jutland in northern Denmark in the peripheral parts area of the maximal Weichselian ice sheet was relatively largely depressed. This indicates the existence of a proglacial depression extending outside the ice margin. In the model presented in this report proglacial depression is assumed to exist relatively far from the ice margin.

The development of forebulges seems probable but is neglected in the presented model. This is justified by the following; Forebulges exist in the most peripheral areas, thus mainly outside the modelled area. The development of forebulges would depend on whether the peripheral parts are continental or oceanic coast. The crustal movements related to the development of forebulges are assumed to be relatively small. Walcott (1970) has presented theoretical calculations of the size and position of forebulges peripheral to an ice load. These calculations are mainly based on the effective flexural rigidity of the elastic lithosphere and the ice thickness. According to Walcott (1970) a situation similar to the last glacial maximum gave rise to a forebulge of 20 m situated c. 80 km outside the ice front.

4.3 The modelling

4.3.1 Eustatic component

As a response to climatic variations water is stored in the global ice sheets during cool phases and in the oceans during warmer phases. The changes of the global sea level, eustatic changes, could thus be correlated to the extensions of the ice sheets.

Geological and other data concerning eustatic development during the Weichselian glacial is very uncertain. Lundberg & Ford (1994) have presented an intermittent curve based on coral datings. This curve includes hydro-isostatic changes and can not be used to describe shoreline displacement solely caused by eustatic lowering/rise. Shackelton (1987) has from oxygen isotope variations calculated an eustatic curve which is in wide use.

The eustatic curve used in this report is based on modifications of the curves presented by Lundberg & Ford /1994/ and Shackelton /1987/ and revised to fit the used glacial chronology. The eustatic curve is shown in Figure 4-2.

4.3.2 Isostatic component

The shore-level model presented by Påsse (1996a, 1997) indicates that there are two components involved in glacio-isostatic uplift, one slow and one fast. The main uplift, still in progress, acts slowly. The slow mechanism can be linked to viscous flow and to the response of the crust to long-term evolution of crustal stresses. During the last deglaciation and also during the ice advance and retreat during the Younger Dryas a fast component of subsidence/rebound can be observed. The fast component give rise to crustal movements during a very restricted period, lasting a few thousand years. The underlying physical processes causing this movement are not known. For the modelling of the gross glacio-isostatic changes during a full glacial period the fast component is judged to be of limited importance, and it is neglected in the present modelling. More detailed information regarding the late Weichselian and Holocene shore level displacement, where the fast component is included, is presented in Påsse (1996a, 1997).

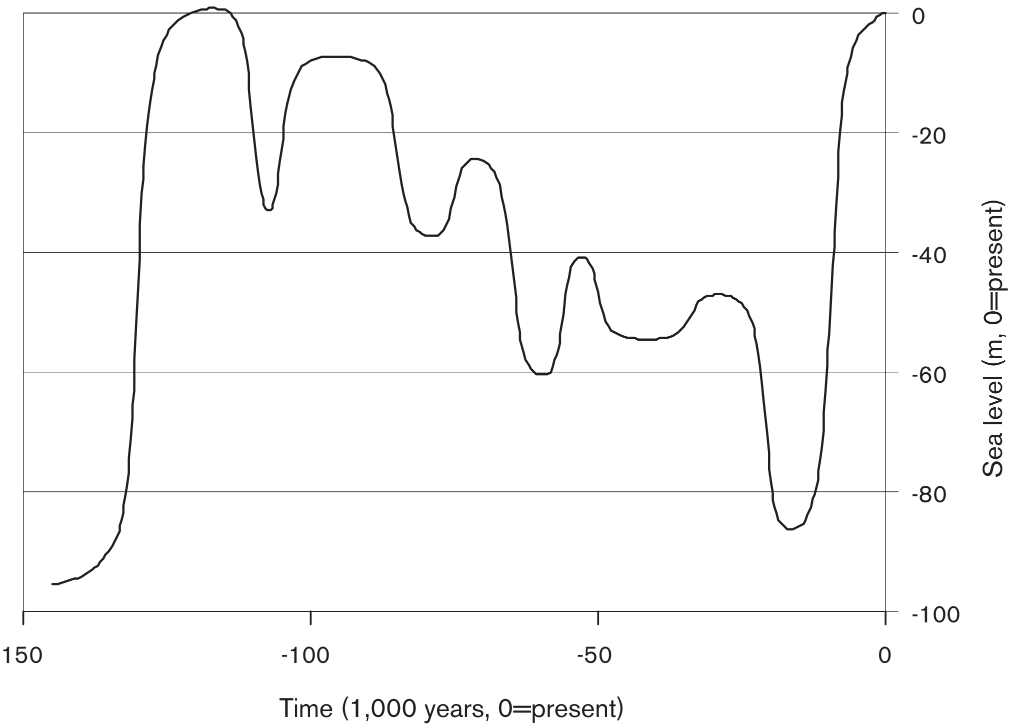


Figure 4-2. *The eustatic curve used in this report. The curve is based on modifications of curves presented by Lundberg & Ford (1994) and Shackleton (1987) and revised to fit the used glacial chronology.*

Isostatic crustal movements start slowly, reach a maximum rate and thereafter follow a retarded course. *Arctan* functions have proved to be suitable tools for describing the evolution of slow glacio-isostatic uplift in time. The basic formula for the evolution of land subsidence/uplift derived in Påsse (1996a, 1997) can be expressed as

$$U = \frac{2}{\pi} A_T \left[\arctan\left(\frac{T}{B}\right) - \arctan\left(\frac{T-t}{B}\right) \right] \tag{4-2}$$

Where T = the time for the maximal subsidence/uplift rate (years)

A_T = download factor (m), subsidence/uplift at the time T

B = inertial factor (years)

t = variable time (years)

There is a relationship between the inertial factor (B) and crustal thickness. In areas with greater crustal thickness the inertial factor is higher than in areas with thinner crust. The inertial factor was originally named the declining factor. The regional pattern of the inertial factor was estimated in Påsse (1997). Figure 4-3 shows a map of the isobases of the *B* factor.

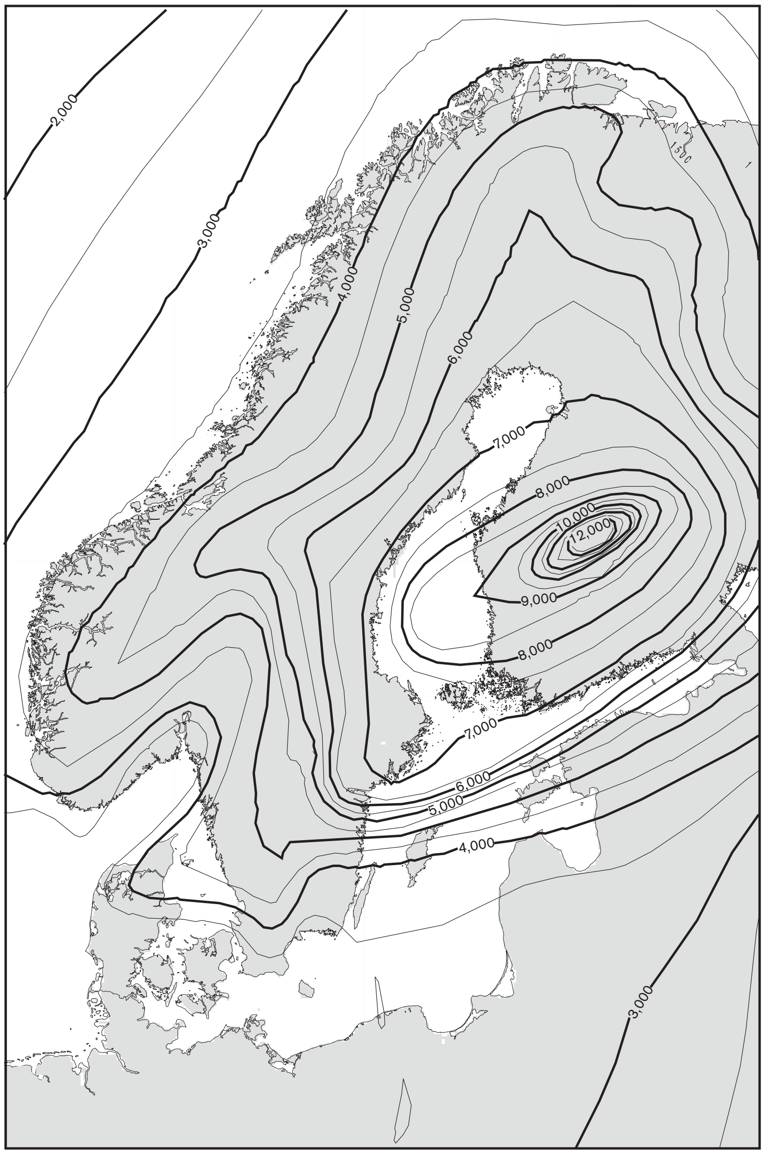


Figure 4-3. *Isobases for the inertial factor (B) (from Påsse, 1997).*

Crustal uplift is markedly delayed in relation to deglaciation. This is due to the slow viscous flow mechanism and flow velocity is concluded to govern the development of crustal changes in time. Response to glacial loading is assumed to follow a similar course of events as unloading. The ice thickness is of great importance for the amount of isostatic depression. However, the viscous flow mechanism also implies that the duration of the glacial load is very important for the depression. A thick ice existing during a short period may produce a small depression, while a thinner ice existing during a long period may produce a similar or even bigger depression. When the ice melts the uplift starts at the contemporary level of depression. If an ice sheet exists during a short period isostatic balance is not reached as the download process is interrupted when the ice melts. This also implies that during ice free periods, uplift usually ends at a level ”far” from isostatic equilibrium. The growth and decay of ice sheets is a fast process in relation to mantel flow, thus subsidence or uplift postulated from the asymptotic parts of formula 4-2 will most probably never occur.

Observed and modelled land uplift since the last glacial maximum indicates that the rate of crustal uplift was low during the initial part of crustal recovery, i.e. when the peripheral parts of the ice sheet were deglaciated. Shortly later the uplift rate reached a maximum. This is interpreted as variations in the rate of viscous inflow of mantle material. When the inflow to the centre of the depressed area had reached a maximum, the flow of mantle material was maximal at every spot simultaneously. This could also be expressed as crustal recovery, or subsidence, occurs in sympathetic movements. In the model presented in this report T (the time for maximum subsidence/uplift) has specific constant values for each glacial event.

The essence of the presented model can simplified be expressed as estimating the variations in A_T (download factor) and T (the time for maximum subsidence/uplift) through two full glacial periods. These estimations are based on the glacial chronology and the ice sheet extension given as postulates. The estimated values of A_T and T are used for calculating a crustal change curve (CCC), defined as a curve describing the variations in level of the crust through time in relation to the present level.

As crustal changes occur in sympathetic movements a proportional relation exists between the movement for each site in an area. This implies that glacio-isostatic movements only have to be calculated in detail for a few sites and that the regional course of crustal movements can be extrapolated from these.

4.3.3 Crustal change curves

The Weichselian glacial cycle is subdivided into phases with growing ice sheets and phases of vanishing or no ice sheets. For the development of crustal change these phases represent periods of subsidence and land uplift respectively. The duration of the phases and the values of A_T and T, used in the calculations, are presented in Table 4-1. Subsidence during phases of growing ice sheets are given negative values of A_T, while land uplift during phases with vanishing or no ice sheets are given positive numbers. Crustal changes are calculated according to formula 4-2. The subdivision of the Weichselian is given as a postulate and mainly follows the chronology presented by Mangerud /Mangerud, 1991/.

Table 4-1. Values used for calculating a reference curve of crustal changes

Download factor A_T (m)	Start of phase (y before present)	Time for maximum subsidence/uplift T (y before present)	End of phase (y before present)
–550	190,000	148,000	145,000
450	145,000	132,000	115,000
–300	115,000	110,000	107,000
225	107,000	105,000	90,000
–300	90,000	85,000	80,000
200	80,000	75,000	70,000
–325	70,000	65,000	60,000
175	60,000	56,000	52,000
–137	52,000	39,000	27,000
–200	27,000	24,000	20,000
500	20,000	12,500	0

The modelling is an iterative process. First a crustal change curve for one hypothetical site is determined. This site is situated in the central part of Fennoscandia where A_{12500} (the value of A at 12 500 BP) is set to be 500 m /Påsse 1996a, 1997/. An isobase map for A_{12500} is shown in Figure 4-4. The crustal level during the end of the Eemian interglacial is assumed to be approximately the same as the present level. It is presumed that the evolution of subsidence/rebound in time can be expressed by arctan functions; subsidence/rebound starts slowly, reaches a maximum rate and eventually declines. The inertial factor B represents the evolution of subsidence/uplift in time. B is initially given a constant value of 4000. The extent of ice loading is expressed in the value of A_T, the download factor. T is the time when the maximum uplift rate occurs, at this time the subsidence/uplift is assumed to be equal to A_T. The values of A_T determining the glacio-isostatic subsidence during the glacial phases, are estimated based on assumptions of the ice extensions during each specific phase. The values of A_T, determining the following recovery, are dependent on crustal level reached during the previous phase and these values are estimated based on the iteration results. The resulting crustal change curve is shown in Figure 4-5.

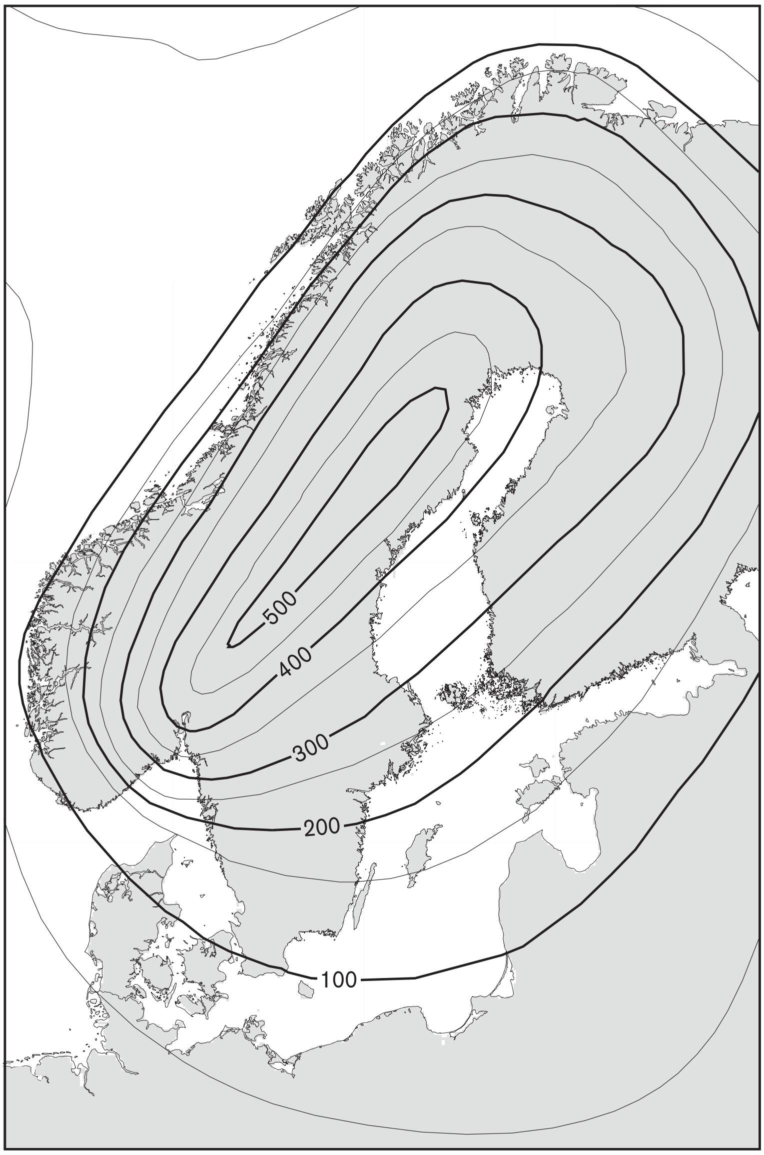


Figure 4-4. *Isobases for the uplift at 12 500 years before present (A_{12500}).*

The estimations of A_T, which in turn are based on assumptions of ice extensions, involve great uncertainty. The Eemian isostatic level is assumed to be similar to the present, and the crustal change curve is assumed to be close to zero during the end of the Eemian interglacial. Most of the Weichselian larger or smaller parts of Fennoscandia were covered by ice sheets. Seen over the total glacial cycle, the crust must have been gradually depressed, from the Eemian level down to the crustal level recorded during the Late Weichselian. During the glacial cycle the isostatic level is presumed to never exceed the Eemian level and never to be below the minimal level occurring at the Late Weichselian. To demonstrate the plausibility of the model it is compared to a linear depression from the Eemian level to the maximal depression occurring at the Late Weichselian, see Figure 4-5.

The next step in the modelling was to construct three more crustal change curves applying the same approach. These curves were calculated with different values of the inertial factor *B:* 3500, 6500 and 9000. The values of A_T for cold phases, determining the subsidence and derived from assumptions of the ice extension, were not altered. The values of A_T for warm phases, determining the recovery, were changed by a few tenths of metres to fit the level to which the subsidence had reached. The calculated curves are shown in Figure 4-6.

The differences between the four crustal change curves are related to the inertial factor. A linear relation between the curves using different values of B has been estimated, formula 4-3. The curve calculated by B = 4000 is chosen as a reference curve. This crustal change curve is designated CCC (500/4000) and is constructed based on the

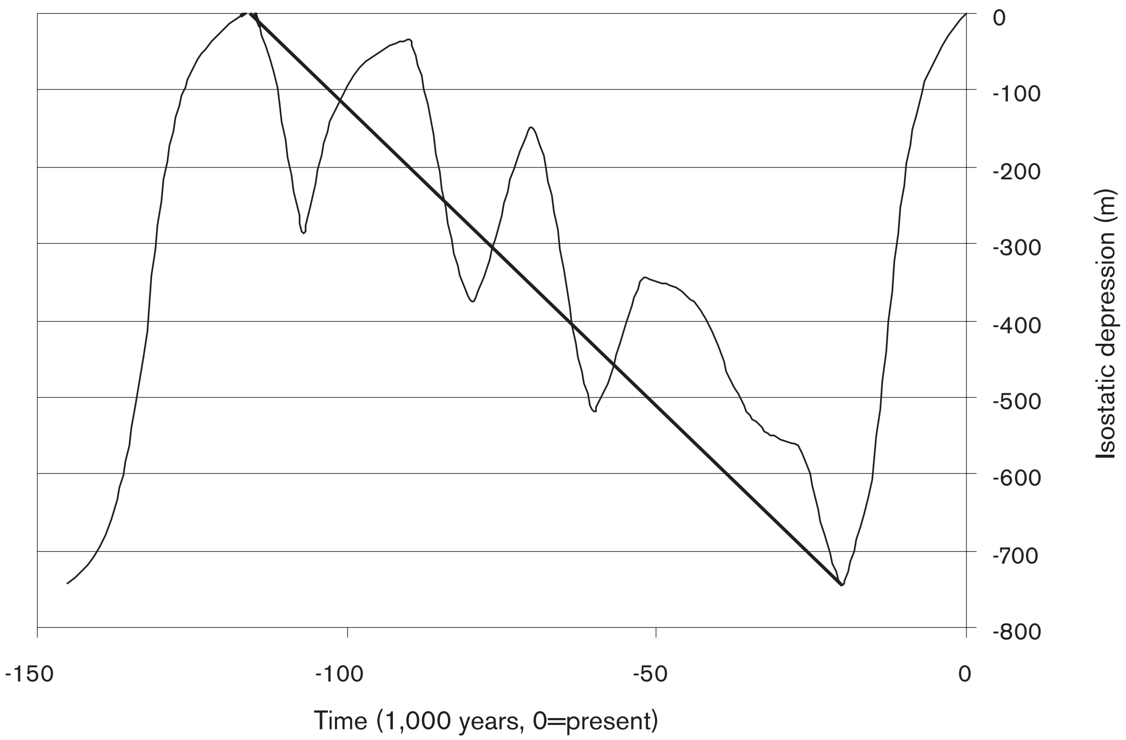


Figure 4-5. *The course of modelled crustal change during the Weichselian compared to a linear depression from the assumed Eemian level to the maximal depression occurring at the Late Weichselian. Seen over the total glacial cycle the crust have been gradually depressed, from the Eemian interglacial level down to the lowest level recorded during the late Weichselian.*

assumption that A_{12500} = 500 m and B = 4000. Crustal change curves (CCC) for sites with other values of B can now be calculated according to:

$$CCC(500/B) = (-0.75\,10^{-4}B + 1.32) * CCC(500/4000) \tag{4-3}$$

Examples of crustal change curves estimated according to formula 4-3 are shown in Figure 4-6. A comparison between Figure 4-6 and Figure 4-7 shows that the differences between the curves are very small, which implies that crustal change curves can be calculated for any B values using the relation 4-3.

Crustal change curves (CCC) for the uplift since the Late Weichselian can be estimated for other sites by proportionality to the reference curve according to:

$$CCC(A_{12500}/B) \equiv (A_{12500}/500)\,(-0.75\,10^{-4}\,B + 1.32)\,CCC(500/4000) \tag{4-4}$$

where A_{12500} and B are site specific values. Presuming that the pattern of crustal changes during the Late Weichselian is valid for other periods, crustal changes through time can now be estimated for any site in Fennoscandia. The crustal change curves are derived from three sources; the reference curve, the assumed A_T value in relation to the A_{12500} value received from Figure 4-4 and the value of B received from Figure 4-3.

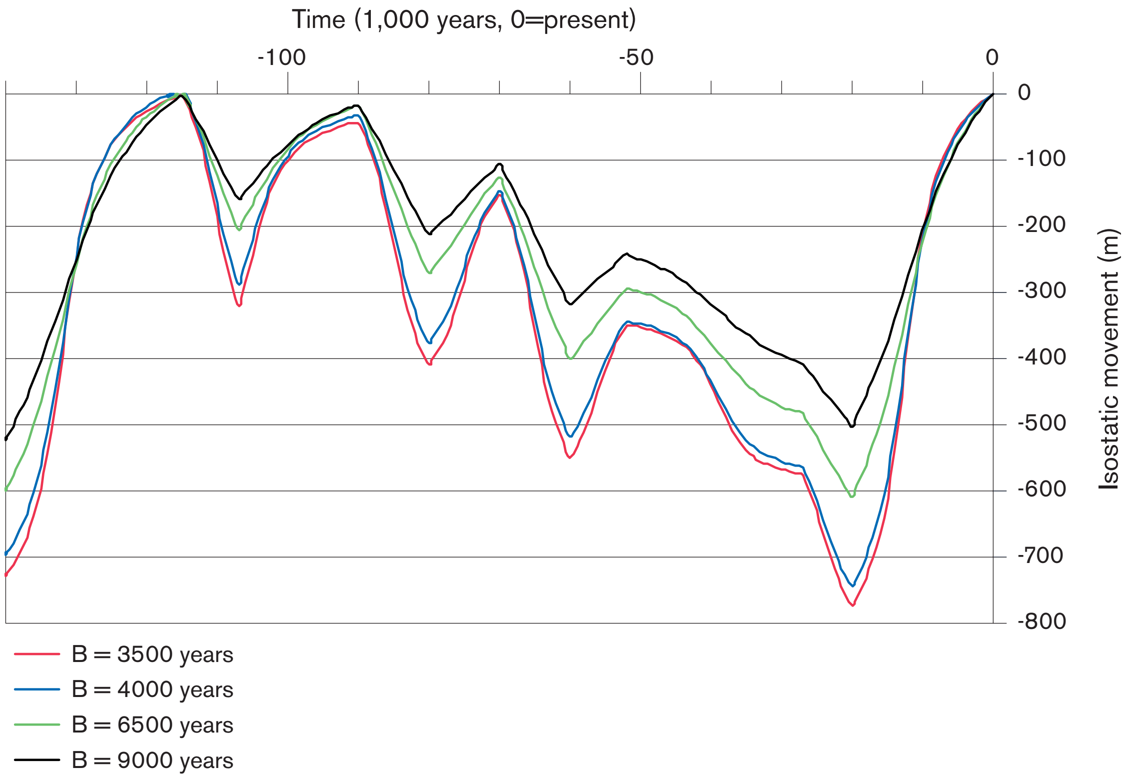


Figure 4-6. *Crustal change curves (CCC) calculated for a hypothetical site in central Fennoscandia using different values of the internal factor B.*

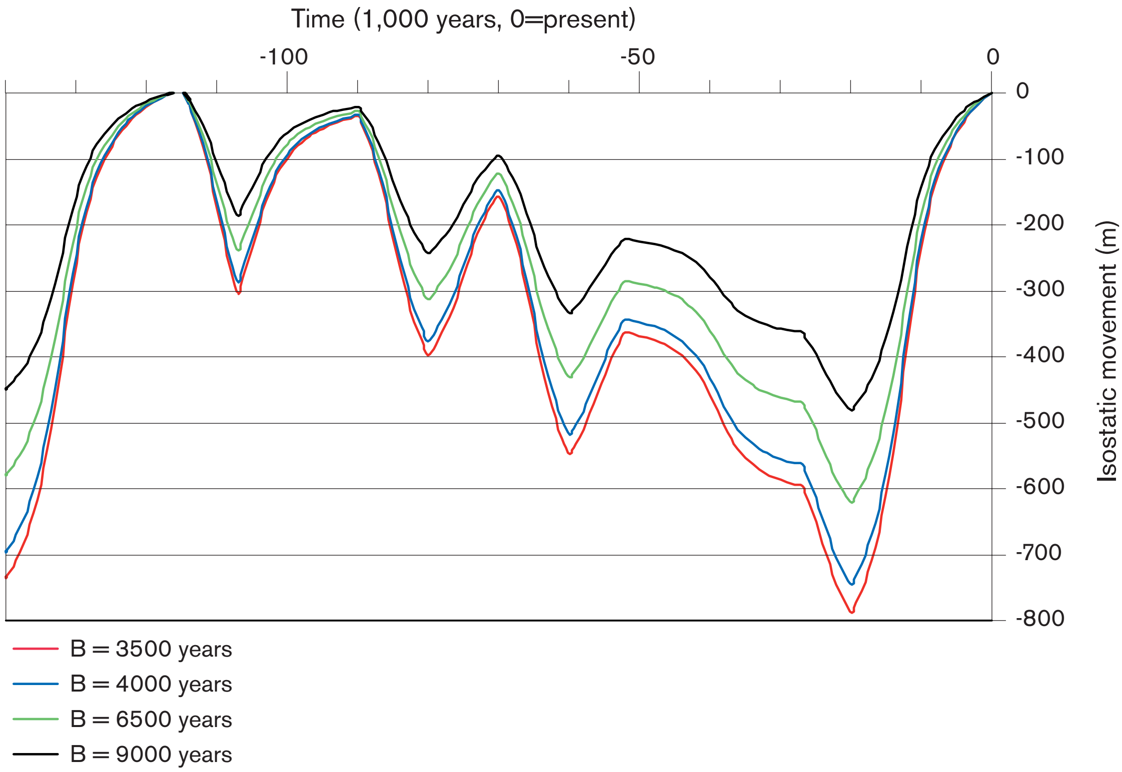


Figure 4-7. *Crustal change curves estimated by applying formula 4-3.*

4.4 Past shore level displacement

By adding the eustatic curve to the isostatic balance curve shore level curves can be calculated for any given site. Shore level displacement during the last 140 000 years at a few sites are showed in Figure 4-8 and Figure 4-9. The locations of the sites are shown in Figure 4-10. The shore level curve representing the Öresund strait indicates that the Baltic was connected to the sea during all "ice free" phases during the Weichselian glacial. However, regarding the uncertainties both in the crustal levels and the eustatic levels this conclusion is not unambiguous as the threshold level at the Öresund strait is situated at a level only tenths of metres from the calculated shore level.

Modelled shore level data have also been supplied to a GIS (Geographic Information System) application. Through the GIS-application the modelled geographical evolution of Scandinavia can be depicted as topographic maps. Maps for the assumed warm and cold peaks are shown in Figure 4-11 together with the assumed ice sheet extension for each phase.

4.5 Discussion of the model and the model results

Empirical data to test the model are very scarce. Only one shore level curve can be used for validation of the results. This curve represents the shore level displacement during Late Saalian, Eemian and Early Weichselian in the Varberg-Falkenberg area (Påsse 1998b). The data is based on empirical information but the chronology is relative. A comparison between these data and a shore level curve obtained from the modelling is shown in Figure 4-12. The relative chronology used for constructing the empirical curve is somewhat revised in order to match the chronology used in this report.

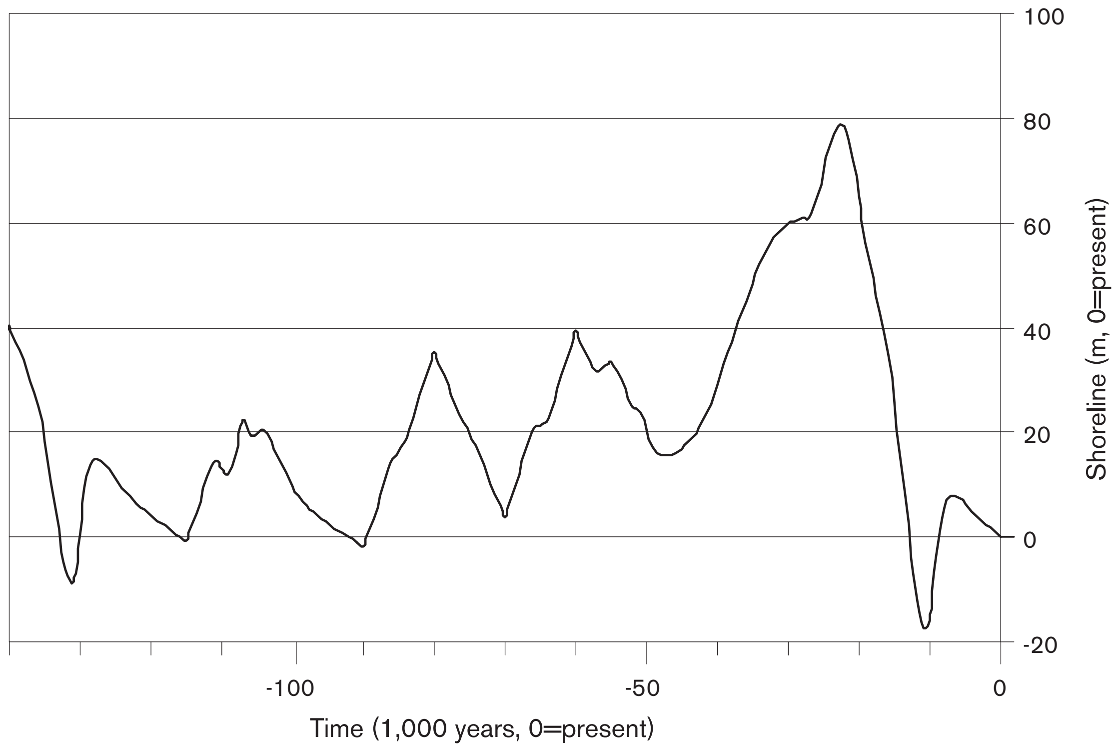


Figure 4-8. *Shore level displacement at Öresund during the last 140 000 years. The threshold at Öresund is situated 7 m below present sea level. Lake phases within the Baltic have only existed during the Holocene and possibly during a short phase of the Eemian interglacial.*

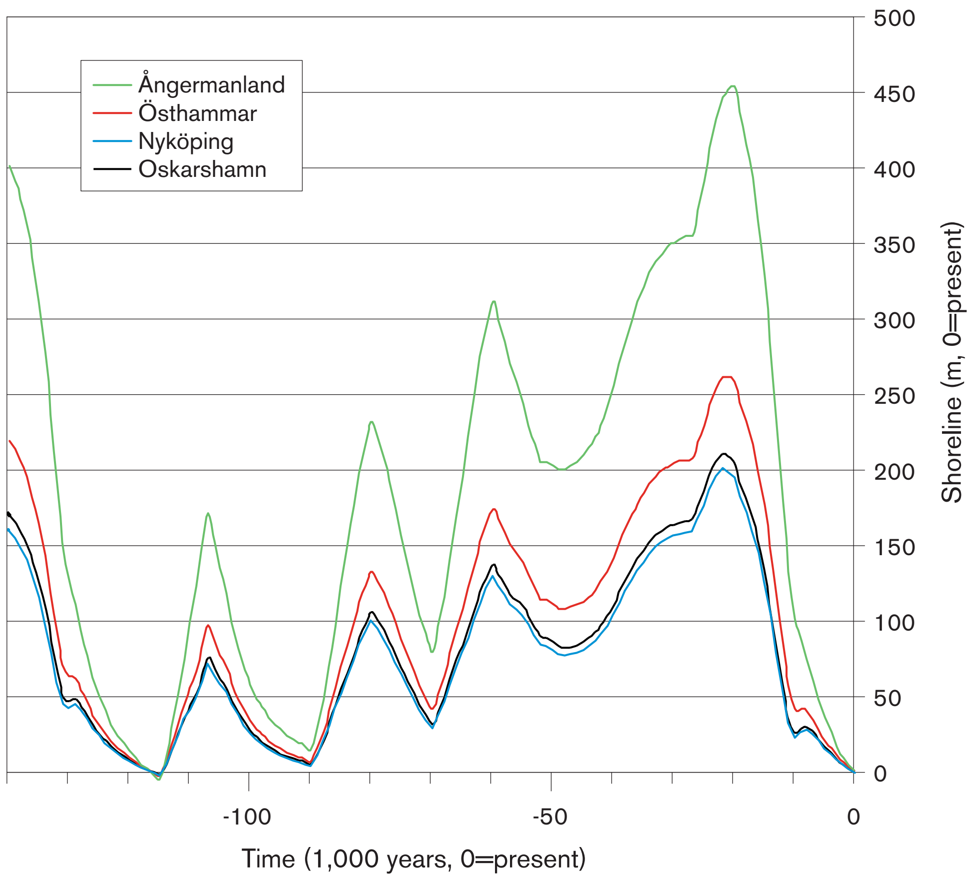


Figure 4-9. *Shore level displacement at Ångermanlad, Östhammar, Nyköping and Oskarshamn during the last 140 000 years.*

Figure 4-10. *Locations of the sites.*

4.5.1 Some remarks concerning the isostatic development

There are reasons to discuss some of the modelling output. A remarkable feature is how much the inertial factor, which depends on the crustal thickness, influences the crustal changes. This can be exemplified by a comparison between the shore level development at Oskarshamn and Nyköping, Figure 4-9. The crustal changes at these sites are quite similar despite the fact that the Nyköping area is situated closer to the uplift centre than the Oskarshamn area. The similarity in the calculated crustal change is due to the large difference in the internal factor. This can be expressed as the crust tends to subside and recover more easily in the Oskarshamn area than in the Nyköping area.

As mentioned before, the duration of the download is important for the amount of depression. As exemplified above the inertial factor also has a large influence on the depression. Large internal factors (large crustal thickness) and/or short duration of ice loading can act to restrain the depression. Including the duration of the download and the inertial factor in the modelling means that the maximal crustal depression can be more adequately estimated than before. The Ångermanland area is usually assumed to have had the maximal glacial depression. However, the modelling shows that the maximal depression in this area only was about –550 m below the present shore level during the

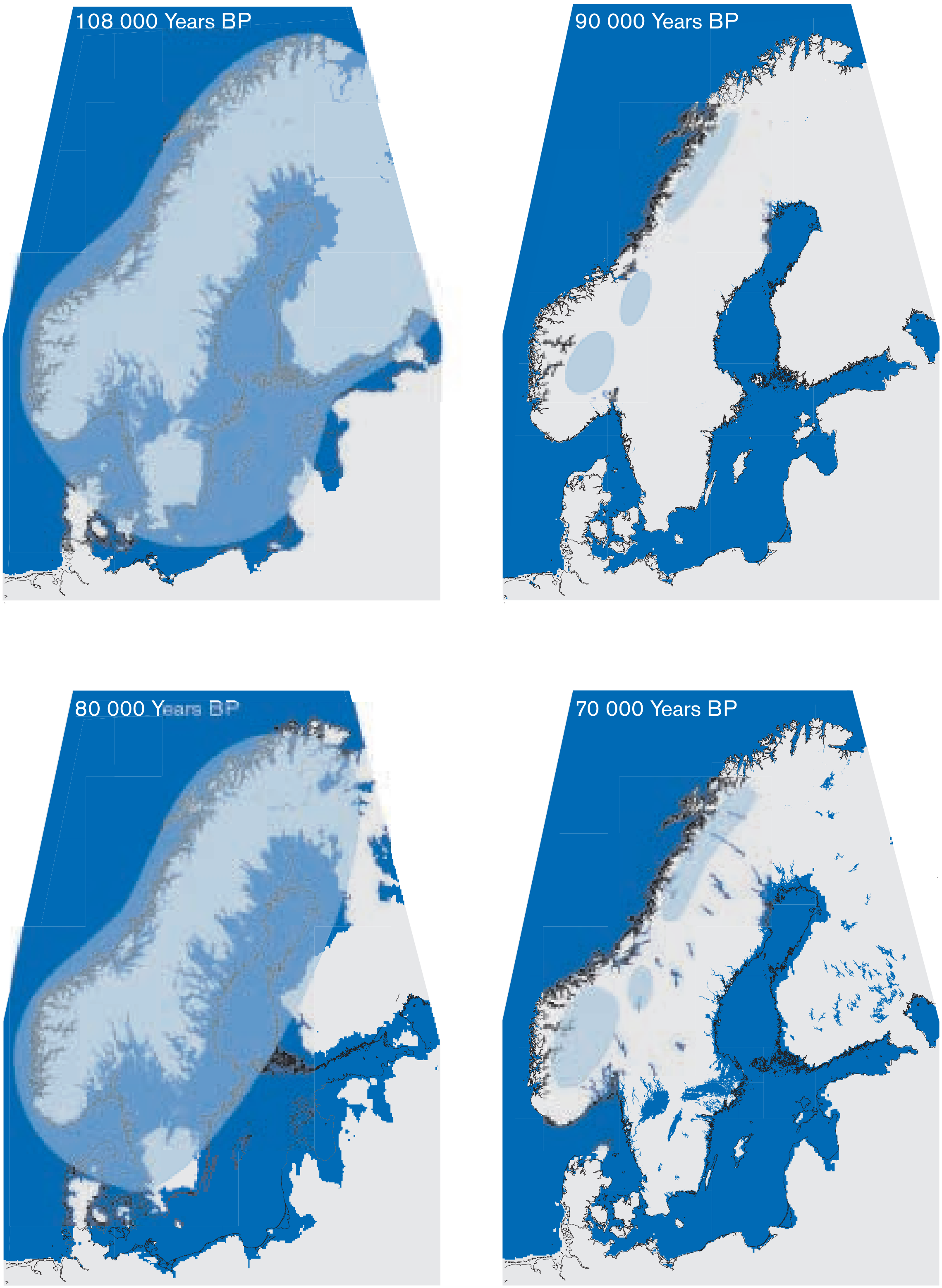
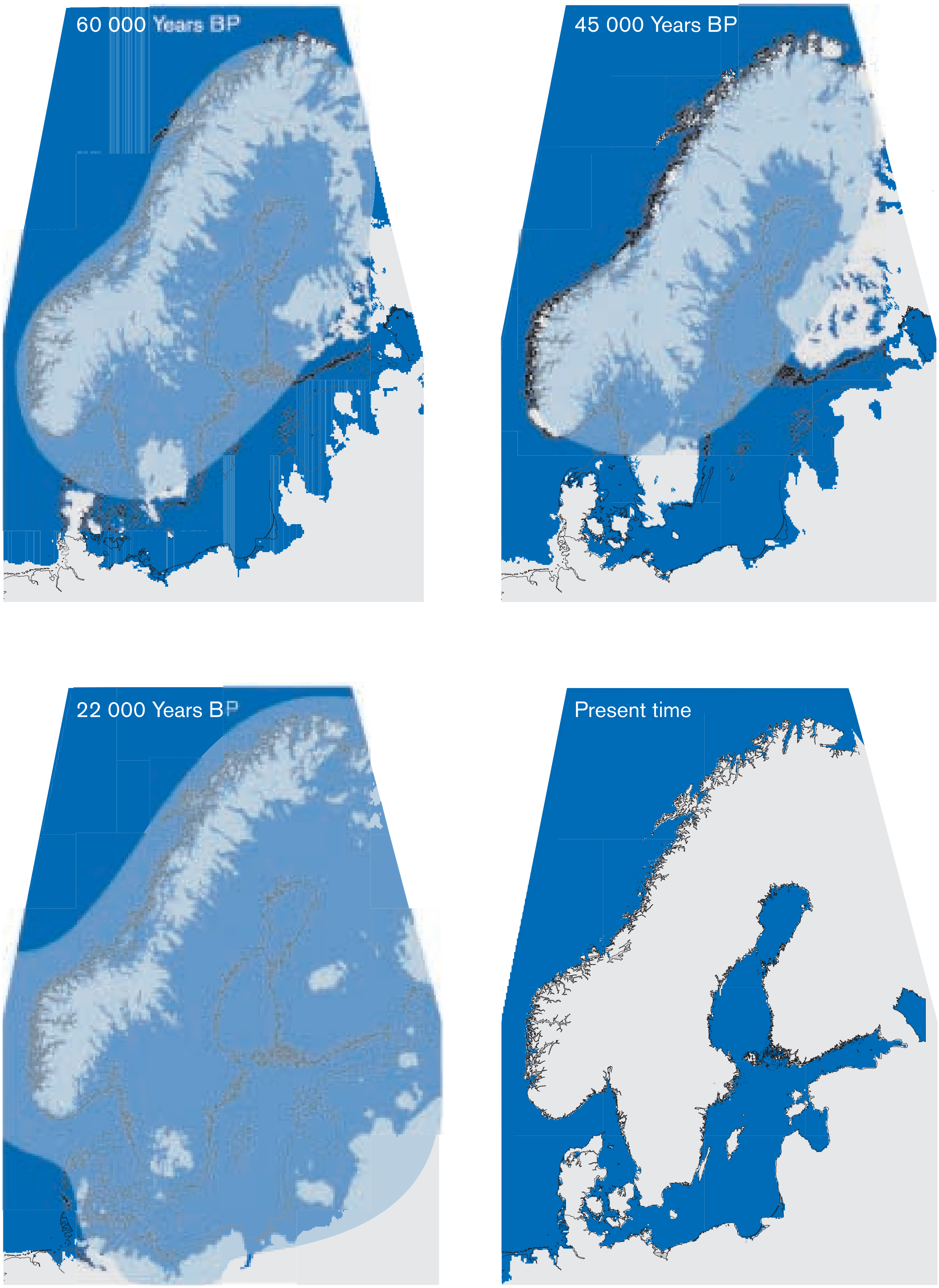


Figure 4-11. *Shore level and ice sheet extension during the cold and warm phases of the Weichselian glacial.*


60 000 Years BP
45 000 Years BP
22 000 Years BP
Present time

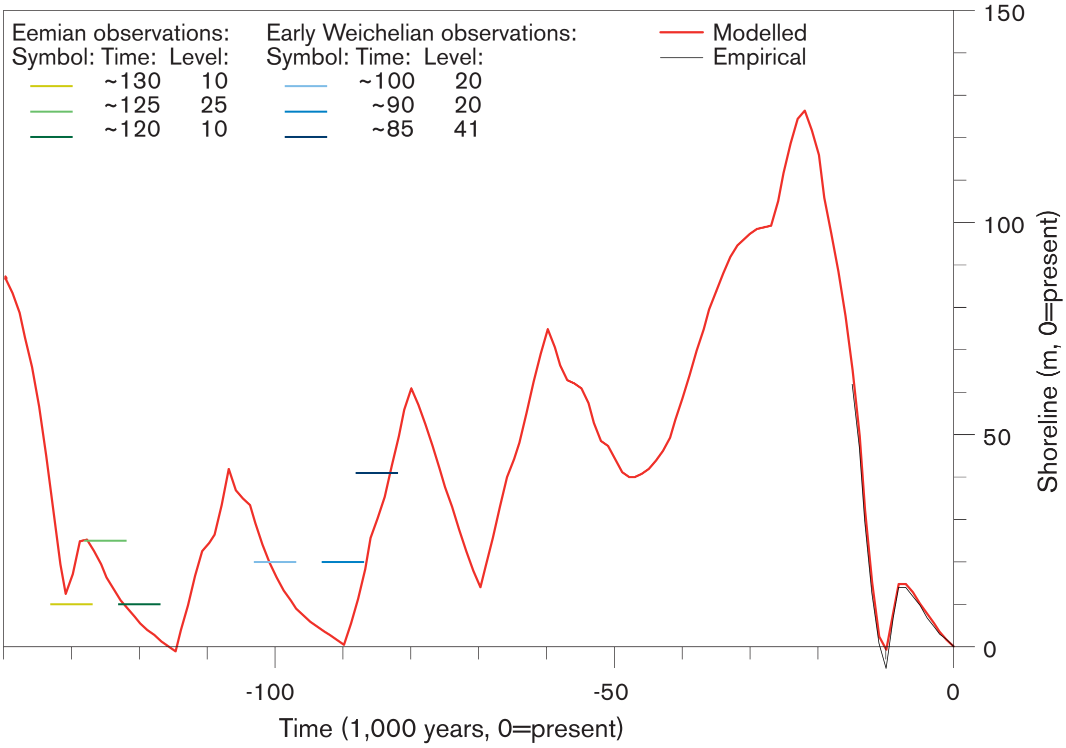


Figure 4-12. *A comparison between empirical data from Varberg-Falkenberg (Påsse, 1998b) and the shore level curve obtained from modelling.*

maximal extension of the ice sheet in the Late Weichselian, while the depression in the Oslo area was about –725 m below the present shore level. It should be added that the remaining land uplift in the Ångermanland area is about 90 m while the remaining uplift in the Oslo area only is 40 m according to Påsse (1997). It should be emphasised that the amount of depression not merely depends on ice thickness, but also is determined by the frequency and magnitude of ice loading, crustal properties and the slow process of mantle flow.

4.6 Future shore level displacement

The future shoreline was calculated applying the same method as when modelling the Weichselian shoreline. The assumed glacial development is presented in Table 3-3. The corresponding values of T and A_T used when constructing the reference curve for crustal change are reported in Table 4-2, and the resulting curve is shown in Figure 4-13. The assumed future eustatic development curve is constructed on the same basis as the eustatic curve for the Weichselian, it is shown in Figure 4-14.

Local future crustal change curves are calculated by formula 4-3 in the same way as the Weichselian site specific curves. Future crustal change through time can thus be predicted at any site in Fennoscandia from only three sources, the reference curve for the future crustal change shown in Figure 4-13, the value of A_{12500} received from Figure 4-4 and the value of B received from Figure 4-3.

Table 4-2. Values used for calculating the reference curve of crustal changes for the future.

Download factor A_T (m)	Start of phase (y after present)	Time for max. uplift T (y after present)	End of phase (y after present)
500	0	12,500 BP	5,000
–225	5,000	18,000	30,000
200	30,000	35,000	40,000
–300	40,000	55,000	70,000
400	70,000	75,000	80,000
–350	80,000	95,000	110,000
500	110,000	115,000	130,000
–250	130,000	140,000	150,000

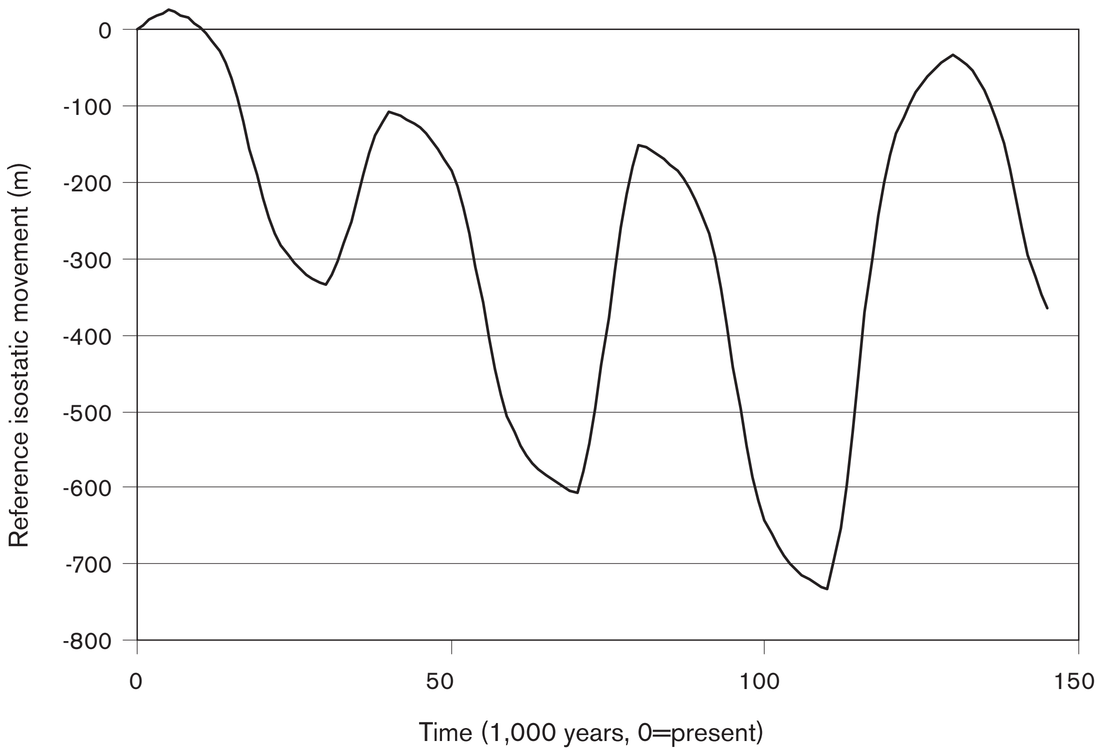


***Figure 4-13.** Reference crustal change curve used in the modelling of the future shoreline.*

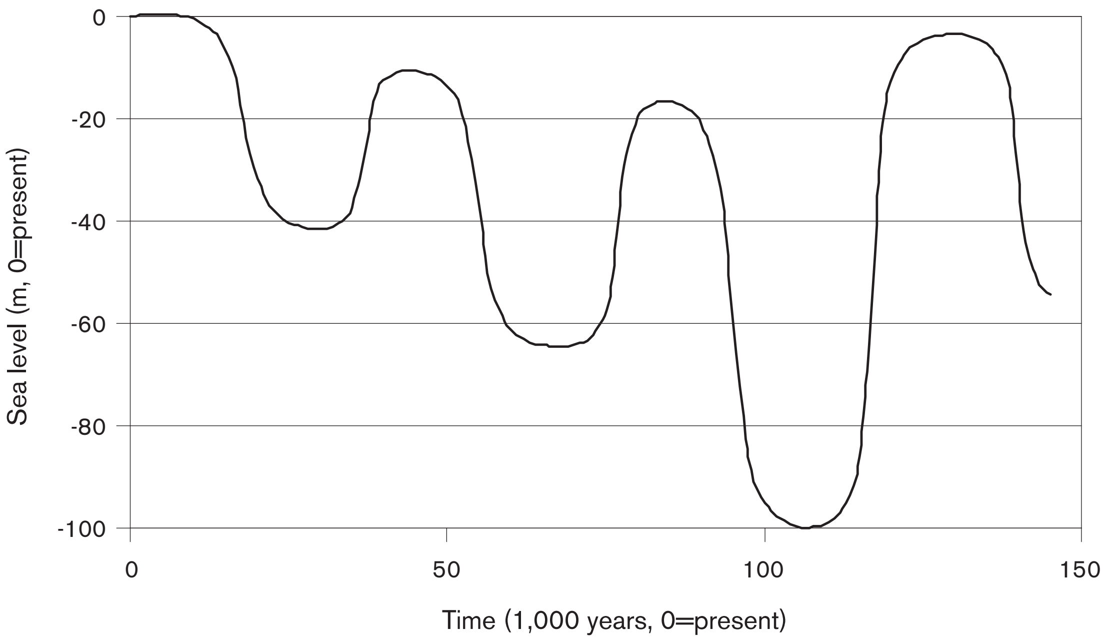


Figure 4-14. *The assumed future eustatic development.*

By adding the local crustal change curves to the curve of future eustatic development it is possible to predict future shore level displacement at any site in Fennoscandia. Examples of future shore level displacements are shown in Figure 4-15 and Figure 4-16. The predictions indicate that the threshold at the Öresund strait is below sea level during the next full glacial, which means that the Baltic will remain marine or brackish during ice free phases. Modelled shore level data have also been used in a GIS-application, and maps depicting the warm and cold peaks are shown in Figure 4-17.

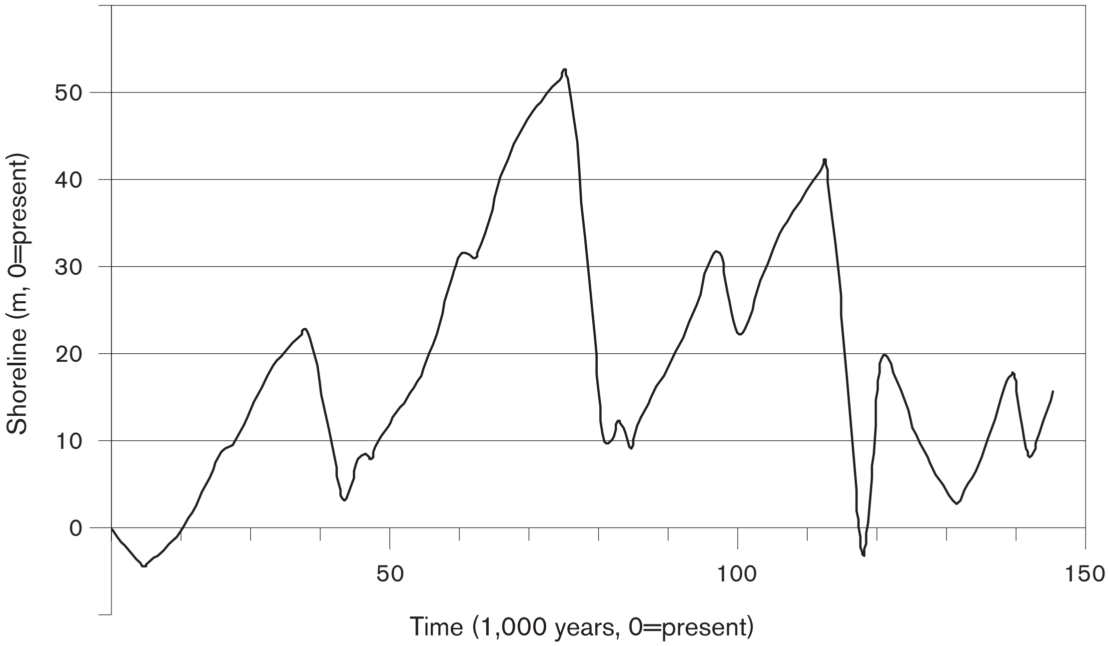


Figure 4-15. *Scenario of future shore level displacement at Öresund*

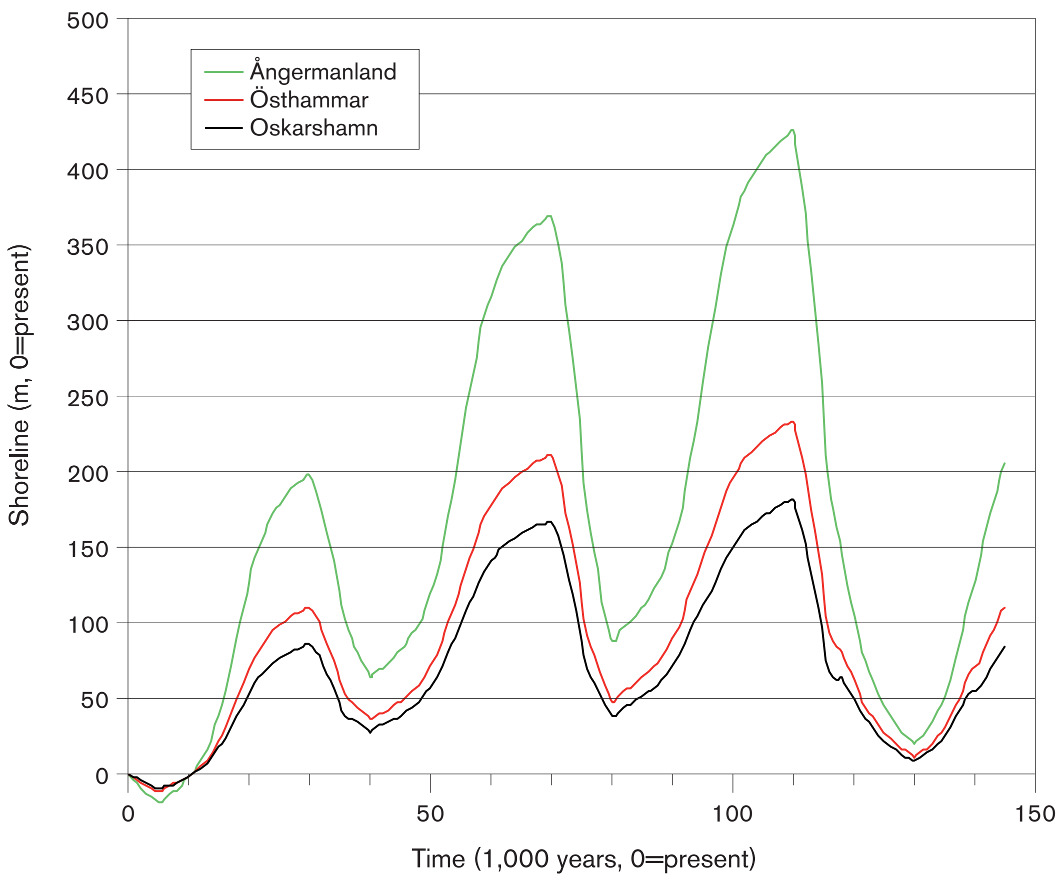


Figure 4-16. *Scenario of future shore level displacement at Ångermanland, Östhammar and Oskarshamn.*

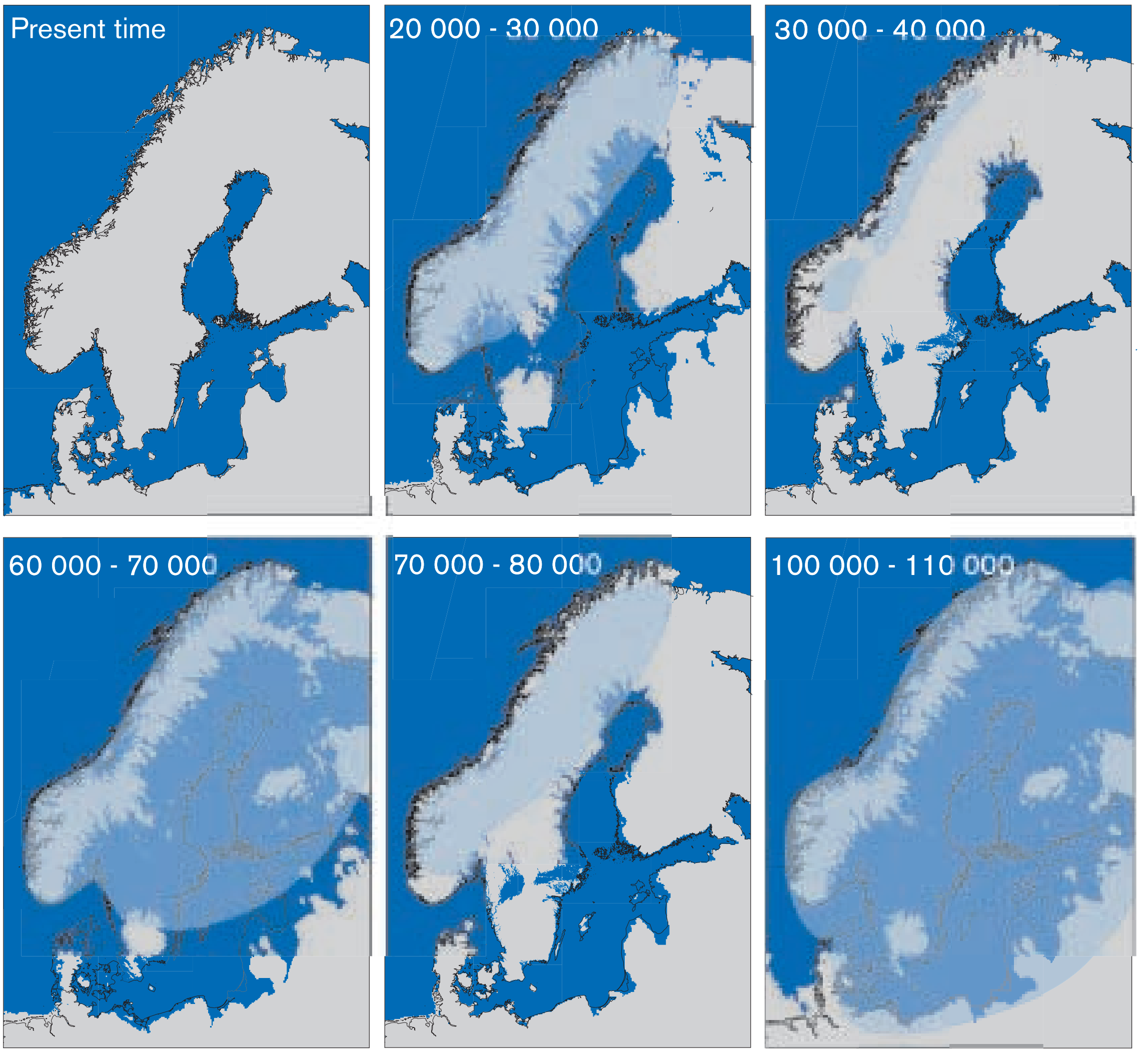


Figure 4-17. *Shore level and ice sheet extension during the warm and cold peaks of the scenario for the next 150 000 years.*

5 Involved uncertainties

5.1 Climate and ice sheet extension

The scenario for the Weichselian is based on empirical observations. Lack of data and uncertainties in dating are the main sources of uncertainties for this scenario.

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.

The scenario for the next 150,000 years is based on the results from simulations using three different models. All the models reconstruct the long-term evolution during the Weichselian quite well. The models are all based on the astronomical climate theory. We assume that the astronomical climate theory is valid. As the orbital parameters can be predicted with high fidelity this means that the predicted timing of peaks and turning points in the climate evolution should be quite appropriate. Due to lack of knowledge and capabilities to describe and simulate the climate system the ranges of climate change predicted by the models are though very uncertain. Only the LLN-model includes a description of the earth climate system. Thus the LLN-model is the only of the three models which can be used to investigate the importance of the different components and processes included in the climate system.

Results from simulations with the LLN-model indicate that human induced climate changes may perturb orbitally driven climate changes. It also predicts that when the orbital forcing is weak the climate system is sensitive to internal dynamics. The variations in insolation are extremely small the next 50,000 years. If the LLN-model captures the main features of the climate system this fact in combination with human induced greenhouse warming may make the current interglacial period extremely long – maybe 60,000 years /Berger and Loutre, 1996/. Due to our lack of knowledge of the climate system it is not possible to judge whether this scenario is credible or not.

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. The presented scenarios of past and future climate give a simplified and smoothed picture of the climate evolution.

5.2 Shoreline displacement

The model used to estimate the shoreline displacement is purely empirical. It is based on the assumption that the evolution of the isostatic component in time always follows the same pattern as observations implies it did since the Late Weichselian. Via the inertial factor the model takes the crustal thickness into account, but it does not include any physical explanation or expression for the subsidence/uplift. In reality subsidence/uplift is a complex process affected by the properties of the crust and the mantle material and by the size and duration of the ice load.

The model is based on a compilation of data from many sites all over Scandinavia. It describes the evolution since the Late Weichselian very well, and the extrapolation of the remaining uplift since the last glaciation ought to be adequate. The modelled shoreline displacements for all other periods are though very uncertain. This is due to uncertainties in the climate scenario and the estimated values of the download factor. It is also due to the simplicity of the model. The model is based on the assumption that the evolution of subsidence/uplift in time always follow the same pattern as it has done since the Late Weichselian, independently of the ice load. This is most probably not true.

In the model the subsided area is always the same independently of the ice sheet extent. For a smaller ice sheet the download factor has been proportionally changed all over the area. In reality the subsided area is most probably related to the extension of the ice sheet, a smaller ice sheet will thus generate a smaller subsided area and the modelled subsidence far from the ice margin for small ice sheets is overestimated in the model.

5.3 The future scenario

Both climate and shoreline displacement scenarios include great uncertainty. The scenarios can only be seen as rough pictures of an evolution that may occur. Lack of knowledge and the complexity of the involved processes currently make the usage of simple models the only available tool to create scenarios of past and future evolution. In spite of the limitations of the models the fact that they are capable to generate a picture of a continuous evolution is believed to result in a more comprehensive scenario than if more complex and possibly more physically correct models had been used. Complex models may be used to generate snapshots of past or future situations, but it is usually not possible to use complex models to simulate long-term evolution. They are often designed for detailed studies of different phenomena, an example are the GCM-models mentioned in section 1.3.2 Climate models. Complex models can be used when evaluating the impact of the scenario on repository performance and safety. More complex and physically correct models could, together with empirical data, also be used to quantify the uncertainties involved in a scenario, but no such studies have been performed in this report.

References

Berger A, 1978. A simple algorithm to compute long term variations of daily or monthly insolation. Contribution No. 18, Institute d'Astronomie et de Géophysique G. Lemaître, Université Catholique de Louvain, Louvain-la-Neuve.

Berger A, 1978. Long-term variations of daily insolations and Quaternary climatic changes. Journal of Atmospheric Sciences, 35(2), 2362-2367.

Berger A, Guiot J, Kukla G, Pestiaux P, 1980. Long-Term Variations of Monthly Insolation as Related to Climate Changes. International Alfred Wegener Symposium, Band 70, Heft 2, 748-758.

Berger A, Loutre M F, 1991. Insolation values for the climate of the last 10 million years. Quaternary Science Rewiews, Vol. 10, 297-317.

Berger A, Loutre M F, Gallee H, 1996. Sensitivity of the LLN 2-D climate model to the astronomical and CO_2 forcings (from 200 kyr BP to 130 kyr AP). Scientific Report 1996/1. Institut d'Astronomie et de Géophysique Georges Lemaître. Universite Catholique de Louvain.

Berger A, Loutre M-F, 1997. Palaeoclimate Sensitivity to CO_2 and Insolation. Ambio Vol. 26, No. 1, 32-37.

Bogren J, Gustavsson T, Loman G, 1998. Klimatförändringar – Naturliga och anropogena orsaker. Studentlitteratur, Art.nr 6452, ISBN 91-44-00320.

Boulton G S, Curle F (eds.), 1997. Simulation of the effects of long-term climatic change on groundwater flow and the safety of geological disposal sites. Report EUR 17793 EN. Luxembourg: Office for Official Publications of the European Communities. IBSN 92-828-0789-4

Bowlby J R, 1984. Glacio-isostatic deformation in Ontario. Proceedings of a workshop on transitional processes. May 1984. 308 p. p. 109-122, Ed. Heinrich W F. AECL–7822 (AECL7822).

Ekman M, Mäkinen J, 1996. Recent postglacial rebound, gravity change and mantle flow in Fennoscandia Geophysical Journal International 126, 229-234.

Emeliani C, 1955. Pleistocene temperatures. Journal of Geology 63 (1955), 538-578.

Fjeldskaar W, Cathles L, 1991. Rheology of mantle and lithosphere inferred from post-glacial uplift in Fennoscandia In: Sabdini *et al.* (eds): Glacial isostasy, sea-level and mantle rheology, Kluwer Academic Publishers, 1-19.

Gallée H, van Ypersele J P, Fichefet Th, Tricot Ch, Berger A, 1991. Simulation of the Last Glacial Cycle by a Coupled, Sectorially Averaged Climate-Ice Sheet Model 1. The Climate Model. Journal of Geophysical Research, vol. 96, NO D7, 13,139-13,161.

Gallée H, van Ypersele J P, Fichefet Th, Tricot Ch, Berger A, 1992. Simulation of the Last Glacial Cycle by a Coupled, Sectorially Averaged Climate-Ice Sheet Model 2. Response to Insolation and CO_2 Variations. Journal of Geophysical Research, vol. 97, NO D14, 15,713-15,740.

Hays J D, Imbrie John, Shackleton N J, 1976. Variations in the Earth's Orbit: Pacemaker of the Ice Ages. Science, vol. 194 (4270), 1121-1132

Henderson-Sellers A, 1996. Climate modelling, uncertainty, and response to predictions of change. Mitigation and Adaptation Strategies for global change 1: 1-21 Kluwer Academic Publishers.

Holland W R, Joussaume S, David F (editors), 1999. Les Houches Session LXVII – Modelling the earth's climate and its variability. Elsevier Science B. V., ISBN 0444 503382.

Holmgren K, Karlén W, 1998. Lete Quarternary changes in climate. SKB Technical Report TR-98-13.

Houghton J T, Jenkins G J, Ephraums J J eds., 1990. Climate Change. The IPCC Scientific Assessment. World Meteorological Organization/United Nations Environment Programme. Intergovernmental Panel on Climate Change. Cambridge University Press. IBSN 0 521 40360 X.

Imbrie J, Hays J D, Martinson D G, McIntyre A, Mix A C, Morley J J, Pisias N G, Prell W L, Shackleton N J, 1984. The orbital theory of Pleistocene climate: Support from a revised chronology of the marine $d^{18}O$ record. Berger A L et al. Eds., Milankovitch and Climate, Part 1, 269-305. Reidel Publishing Company.

Imbrie J, Imbrie J Z, 1980. Modelling the Climatic Response to Orbital Variations Science, Vol 207, 943-953.

Jouzel J, Barkov N I, Barnola J M, Bender M, Chapellaz J, Genthon C, Kotlyakov V M, Lipenkov V, Lorius C, Petit J R, Raynaud D, Raisbeck G, Ritz C, Sowers T, Stievenard M, Yiou F, Yiou P, 1993. Vostok ice cores: extending the climate records over the penultimate glacial period. Nature 364 (6436), 407-412.

Kukla G, Berger A, Lotti R, Brown J P, 1981. Orbital signatures of interglacials Nature, 290(5804), 295-300.

Lundberg J, Ford, D C, 1994. Late Pleistocene sea level change in the Bahamas from mass spectrometric U-series dating of submerged speleothem. Quaternary Science Reviews 13, 1-14.

Mangerud J, 1991. The Scandinavian ice sheet through the last interglacial/glacial cycle In Frenzel B ed. Klimatgeschichtliche Probleme der letzen 130,000 Jahre. Stuttgart, Fisher Verlag, 307-330.

Martinson D G, Pisias N G, Hays J D, Imbrie J, Moore TC, Shackleton N J, 1987. Age dating and orbital theory of the ice ages: Development of a high resolution 0 to 300 000 year chronostratigraphy. Quaternary Research 27, 1-29.

McGinnis L D, 1968. Glacial crustal bending. Geological Society of America Bulletin, 59, 697-710.

Naturvetenskapliga forskningsrådet, Åberg L (redaktör), 1996. Jordens klimat, Naturvetenskapliga forskningsrådets årsbok 1996. Swedish Science Press.

Petit J R, Jouzel J, Raynaud D, Barkov N I, Barnola J M, Basile I, Bender M, Chappellaz J, Davis M, Delaygue G, Delmotte M, Kotlyakov V M, Legrand M, Lipenkov V Y, Lorius C, Pépin L, Ritz C, Saltzman E, Stievenard M, 1999. Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. Nature 399, 429-436, 1999.

Påsse T, 1990. Empirical estimation of isostatic uplift using the lake-tilting method at Lake Fegen and at Lake Säven, southwestern Sweden. Mathematical Geology 22, No. 7, 803-824.

Påsse T, 1996a. A mathematical model of the shore level displacement in Fennoscandia

Påsse T, 1996b. Lake-tilting investigations in southern Sweden. SKB Technical Report 96-10. Swedish Nuclear Fuel and Waste Management Co. Stockholm. 1-34

Påsse T, 1997. A mathematical model of past, present and future shore level displacement in Fennoscandia. SKB Technical Report 97-28. Swedish Nuclear Fuel and Waste Management Co. Stockholm. 1-55.

Påsse T, 1998a. Lake-tilting, a method for estimation of isostatic uplift. Boreas 27. 69-80.

Påsse T, 1998b. Early Weichselian interstadial deposits within the drumlins at Skrea and Vinberg, south-western Sweden. Geologiska Föreningens i Stockholm Förhandlingar, 120, 349 -356.

Påsse T, Andersson L, 2000. A mathematical shore level model presented in GIS. Geologiska Vintermötet i Trondheim. Abstract.

Shackelton N J, 1987. Oxygen isotopes, ice volume and sea level. Quaternary Science Reviews 6, 183-190.

Shackleton N J, Crowhurst S, Hagelberg T, Pisias N G, Schneider D A, 1995a. A new late neogene time scale: Application to leg 138 sites. In Pisias N G, Mayer L A, Janecek T R, Palmer-Julson A, van Andel T H (eds), Proceedings of the Ocean drilling Programme, Scientific Results, Vol 138 (1995) pp 73-101.

Shackleton N J, Hall M A, Pate D, 1995b. Pliocene stable isotope stratigraphy of site 846. In Pisias N G, Mayer L A, Janecek T R, Palmer-Julson A, van Andel T H (eds), Proceedings of the Ocean drilling Programme, Scientific Results, Vol 138 (1995) pp 337-353.

SKB Technical Report 96-24. Swedish Nuclear Fuel and Waste Management Co. Stockholm. 1-92.

Walcott R I, 1970. Isostatic response to loading of the crust in Canada. Canadian Journal of Earth Sciences, 716-727.