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Climatic changes and uplift patterns – past, present and future

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CLIMATIC CHANGES AND UPLIFT PATTERNS - PAST, PRESENT AND FUTURE

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Abstract

Our knowledge about the Pleistocene (= the last 2.5 million years) climatic changes and their global environmental effects on the Earth system, e.g. the glacial- interglacial cycles, the sea level changes, and the significant crustal movements in glaciated regions, has increased greatly during the last decades. This report outlines the historical background and the present state-of-the-arts on these matters. Because the driving mechanisms and feed-back effects behind these changes have been more and more discussed in earth-science literature, analysed, and probably also better and better understood, it has become possible to present theoretical models for future climates (not including man's influence on the earth system). The report presents and discusses one such climate model (short of predicting man's future behaviour and its consequent effect on climate) and its likely implications on future climatic and glacial conditions, and bedrock movements, with focus on the Stockholm region. Possibilities for Quaternary geologists to establish and map postglacial fault zones, related to irregular bedrock movements, are also briefly outlined in the report.

Sammanfattning

Vår kunskap om kvartärperiodens glaciala/interglaciala klimatvariationer liksom därav beroende havsytefluktuationer och jordskorperörelser har ökat markant under de senaste årtiondena. I denna rapport beskrivs den historiska bakgrunden liksom det nuvarande kunskapsläget inom detta område. Då klimatets drivkrafter och återkopplingseffekter nu nått viss förståelse har teoretiska modeller för jordens framtida klimat kunnat upprättas. Rapporten presenterar och diskuterar en sådan modell (utan att hänsyn tas till mänskligt introducerade klimateffekter). Modellen som baseras på observerade periodiska ändringar i vissa jordbaneelement (Milankovitch teorin) förutsäger att jorden är på väg mot en ny istid. Med hjälp av denna klimatmodell uppskattas klimat, nedisning och jordskorperörelser, särskilt i Stockholmsområdet. Rapporten diskuterar även möjligheter att dokumentera och kartlägga postglaciala förkastningszoner och oregelbundna berggrundsrörelser.

Acknowledgement

For the palaeogeographic reconstructions for Fig. 5-3, 5-4, 5-5, SKB has kindly allowed use of digital elevation data. Mr. Magnus Odin at SKB is thanked for his help with the computing and drawing of these reconstructions.

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1 INTRODUCTION

The aim of this report is to give a brief presentation of the present state-of-the-arts concerning our knowledge about Pleistocene (= last 2.5 million years) climatic changes, glaciations, possibilities to calculate future glaciations and sea level changes, the Swedish uplift history since deglaciation, possibilities to map any former and future fault zones, possibilities to calculate the effect of future glacial stages on the Swedish bedrock, and finally also to present some means and methods within Quaternary geology which could be used to map differential horizontal and vertical movements in the Swedish bedrock. It would of course require a massive work achievement to be able to give an exhaustive presentation of the above listed subjects. This report can be regarded as an introduction to a very complex, partly enigmatic research field, where certain achievements have to be carried out in order to answer the many questions we still have.

Svensson is responsible for chapter 5 and for the editing. Björck is responsible for the remaining chapters.

Earth today (right, top) and during the last ice age (right, below). 20,000 years ago, great ice sheets covered parts of North America, Europe, and Asia, surface waters of the Arctic and parts of the North Atlantic Oceans were frozen, and sea level was more than 100 m lower than it is today. Many parts of the continental shelf became dry land. Drawing by Anastasis Sotiropoulos, based on information compiled by George Denton and other members of the CLIMAP project. From Imbrie and Imbrie (1979). Reprinted with permission from Enslow Publishers. MacMillan Press Ltd. Ice Ages – solving the mystery, 224 pp. By J Imbrie and K P Imbrie 1979.

2 PLEISTOCENE CLIMATIC CHANGES

2.1 SHORT HISTORICAL BACKGROUND

When Louis Agassiz (1840) presented the first scientific evidence (Figure 2-1) that Europe had once been covered by much more extensive glaciers than today, Quaternary Geology gradually emerged as a separate geological subject in many countries where glacial deposits are abundant. This also created interests in glacial environments (e.g. Svalbard, Iceland, Greenland, Antarctica, Arctic Canada, and alpine glacial environments), glacial processes, and glacial deposits and their genesis. Research in these areas led the geologists to realize that not only the Quaternary was characterized by larger glaciations, but that the Earth has undergone glacial periods during the major part of its existence. It has, however, become obvious that some of the answers to the underlying causes and mechanisms behind these climatic changes are most easy to obtain from the geological records of the most recent glacial period, the Quaternary.

When it was realized that these glacial-interglacial cycles were not regional phenomena but of global character, geologists working in areas far outside the actually glaciated regions also became interested in climate studies of the last 2–3 millions of years. In fact, the geologists working within glaciated regions had to work with restricted records, both in space and time. On the other hand they could give detailed accounts of the last glacial and deglacial episodes (and scattered finds of older interglacials, interstadials, glacials, and stadials) as well as of the last late glacial and Holocene (last 10,000 years) history. In fact, the abundance of lake basins in glaciated regions, due to the repeatedly glacially eroded landscape, make these regions most suitable to study the terrestrial environmental and climatic changes since the last deglaciation. This owes to the fact that lake sediments are not only good archives of local/regional environmental history, but usually are also datable and have a good time resolution. Furthermore these glaciated regions have been characterised by a unique property that very early attracted attention by earth scientists: 10,000–15,000 years of more or less rapid and continuous land uplift. However, the increased interest in the period covering the last few millions of years even outside the glaciated regions resulted almost in a scientific revolution. Four types of long and continuous records showed very promising results, but from slightly different palaeoclimatic and palaeoenvironmental aspects: long lake sediment sequences (e.g. Hammen & Gonzalez, 1964), loess deposits (e.g. Kukla, 1970), ice cores (Dansgaard et al. 1971), and marine sediments (Emiliani, 1955). The two latter are generally regarded as the best types of records with respect to global climatic aspects as well as to the possibilities of quantifying these geological records to past temperatures, precipitation and evaporation values,

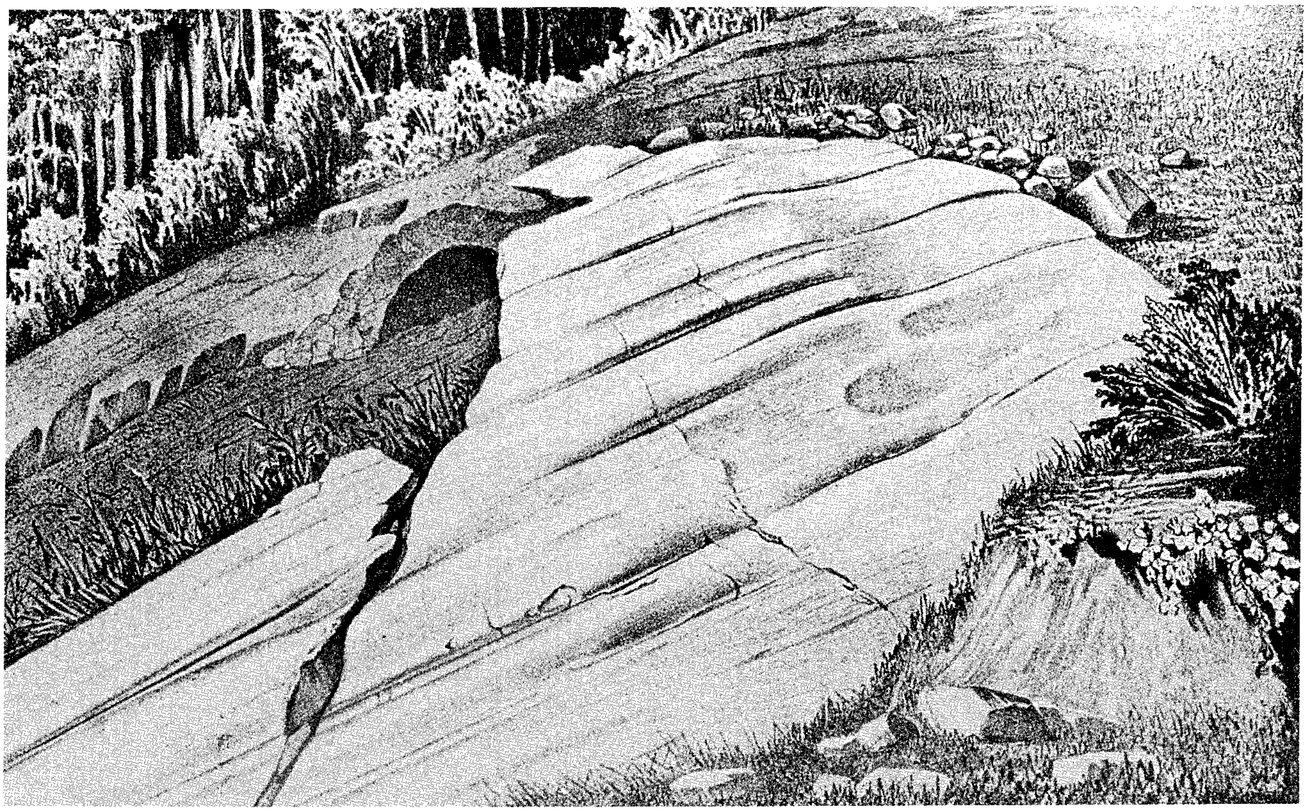

Figure 2-1 An illustration of polished bedrock near Neuchatel, Switzerland, published by Louis Agassiz in 1840. Agassiz argued that polished and grooved surfaces, occurring many miles from existing glaciers, were clear evidence of a former ice age. From Imbrie and Imbrie (1979) who published it from Carozzi (1967).

and glacier ice/water budgets. The records from ice cores and marine sediments are well-correlated and closely linked with each other via the global hydrologic cycle. There are, however, abundant marine sediment cores in comparison to the very few ice cores. Therefore we will mainly focus on the relationship between marine records, past climates, and astronomical theories.

2.2 MARINE RECORDS — PAST CLIMATE — ASTRONOMICAL THEORIES

In the middle of this century, with the advent of deep-sea piston coring during the Swedish Deep-Sea Expedition 1947–48, the first steps towards today's large ODP (Ocean Drilling Programme) programme were taken. It was not only the startpoint in confirming the "Continental Drift Theory", but it also resulted in the possibility of obtaining long continuous sediment records from the Quaternary.

By analysing the ratio of the two oxygen isotopes ^{16}O and ^{18}O in foraminiferas found in these marine cores Emiliani (1955) could show that this ratio had changed through time (downcore) in a cyclic manner (Figure 2-2). Emiliani argued that these changes were caused by the ambient water tem-

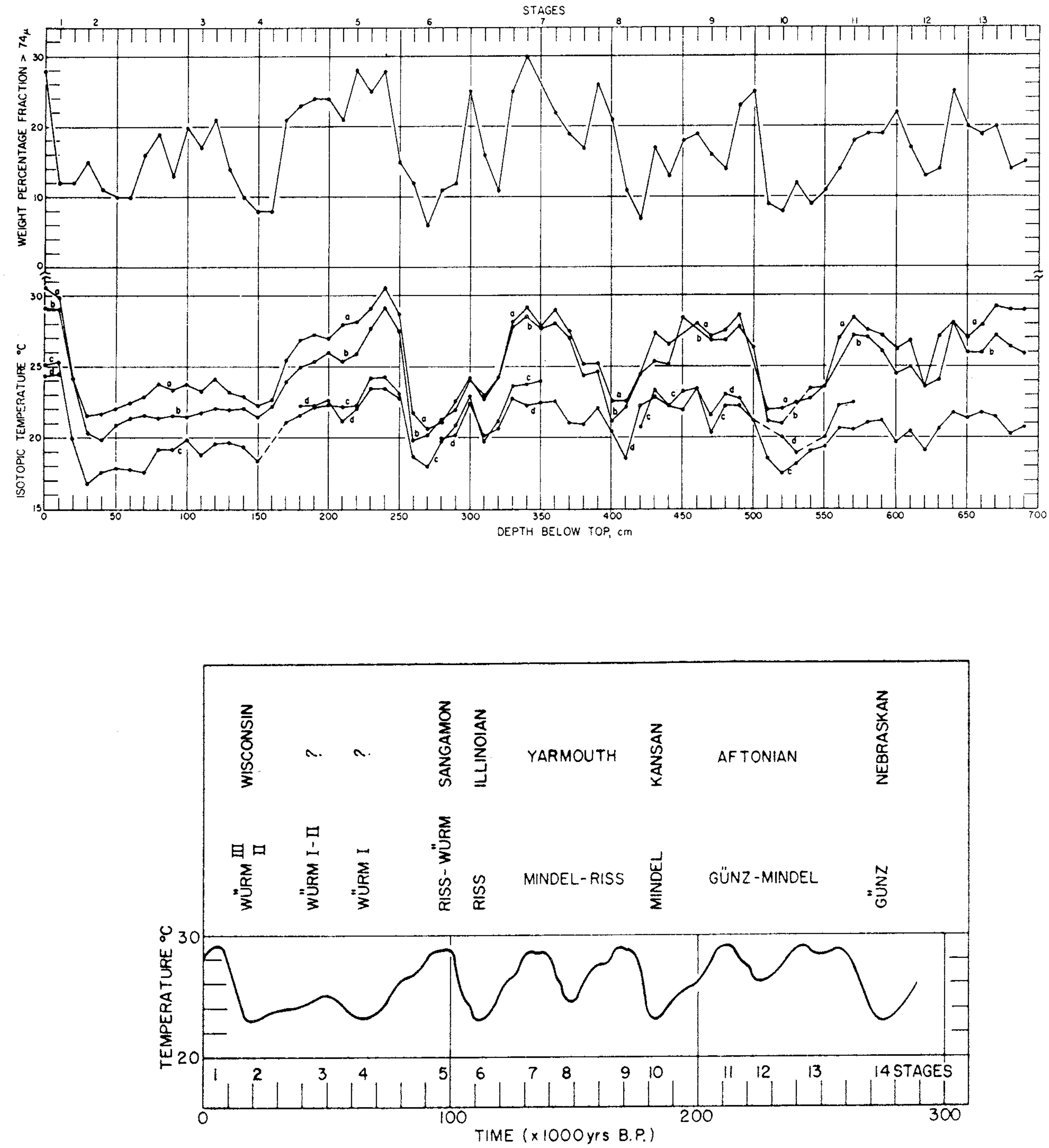

Figure 2-2 Above: core A179-4; percentages of the fraction larger than 74 μm and isotopic temperatures obtained from *Globigerinoides rubra* (a), Globigerinoides sacculifera (b), *Globigerina dubia* (c), and *Globorotalia menardii* (d). From Emiliani (1955). Below: generalized temperature variation, based on the temperature graphs of the cores and on the astronomical time scale. From Emiliani (1955). Reprinted from The Journal of Geology 63, pp. 538-578. Plestocene temperatures, by C. Emiliani, 1955. Publisher The University of Chicago. Copyright 1955 The University of Chicago.

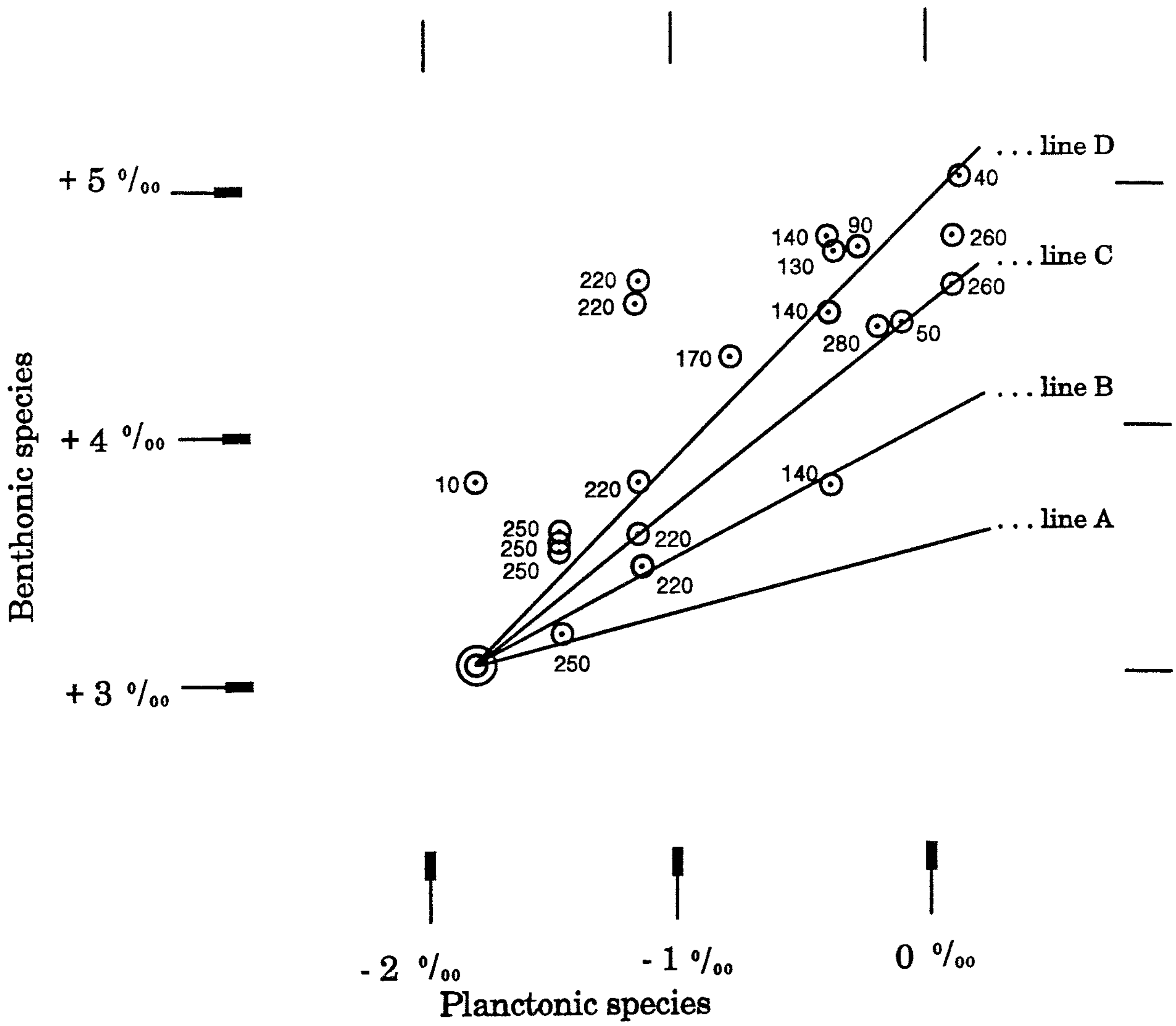

Figure 2-3 Oxygen isotope composition of benthonic foraminifera (numbers denote the depth of each sample in cm) from Lamont core A 179-4 plotted against oxygen isotopic composition of the planktonic species *Globigerinoides sacculifer* (Brady) as determined by Emiliani (1955). If there were no systematic change in temperature with changing isotopic composition, the points should lie on line D. The fact that the points lie closest to line D shows that changing ocean isotopic composition is certainly the primary cause of the isotopic changes observed in planktonic foraminifera. From Shackleton (1967). Reprinted with permission from NATURE 215, pp. 15-17. Copyright 1967 Macmillan Magazines Ltd. Oxygen isotope analyses and Pleistocene temperatures re-assessed, by N J Shackleton 1967.

perature changes reflecting the shifts between glacial and interglacial periods. It was later found (e.g. Olausson, 1965; Shackleton, 1967), and generally accepted that these downcore changes in the oceanic oxygen isotope composition were mainly caused by the growing and shrinking inland ice sheets: the more water that is trapped by glacier ice the higher ^{18}O values in the oceans (Figure 2-3). This also means that sea level changes are more or less directly linked to ^{18}O changes (1 m of sea level change corresponds to a c. 0.01 per mil ^{18}O change according to Fairbanks and Matthews, 1978).

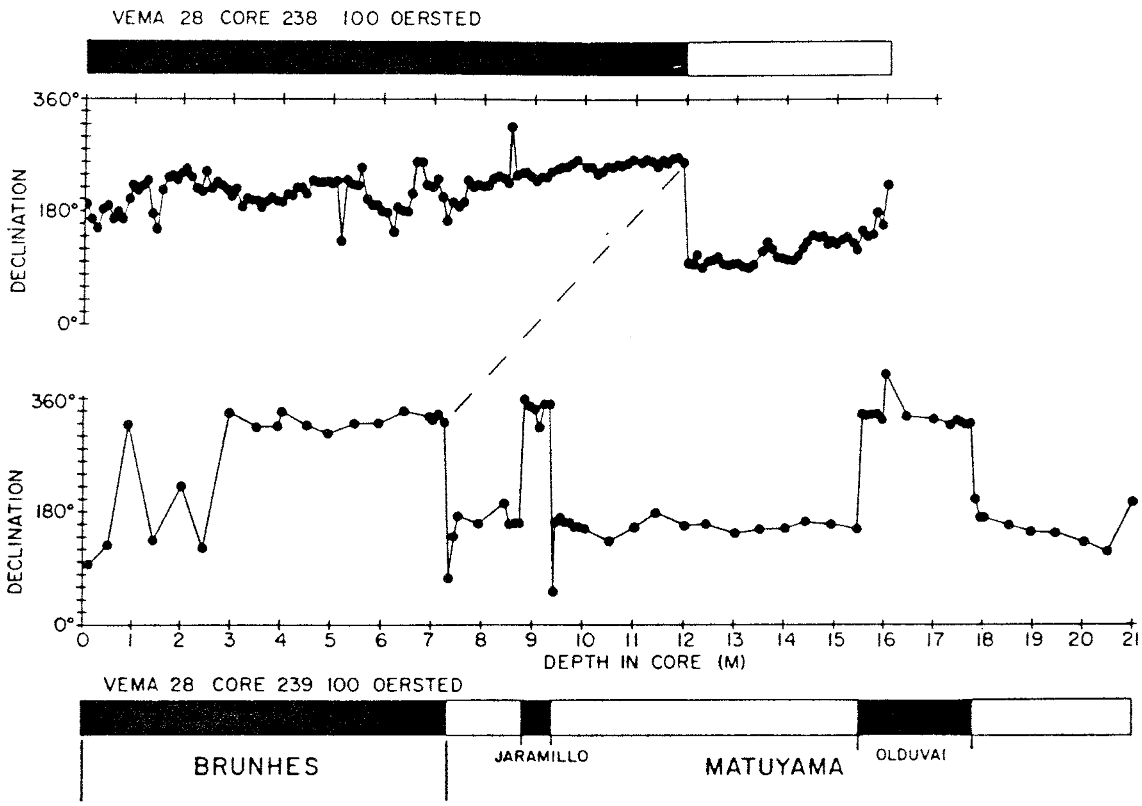

Figure 2-4 Plot of the change in declination for cores V28-238 and V28-239 plotted against depth in core and its correlation with the standard magnetic time scale. The cores are not oriented with respect to true north. From Shackleton and Opdyke (1973). Reprinted with permission from Academic Press. Quaternary Reasearch 3, pp. 39-55. Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacipif core V28-238; oxygen isotope temperatures and ice volumes on a 105 and 106 year scale, by N J Shackleton and N D Opdyke 1973.

It has, however, been found (e.g. Duplessy et al., 1981; Dodge et al. 1983) that certain ^{18}O curves do not only reflect the global ice volume but are also influenced by other factors such as e.g. the ambient water temperature (cf. Emiliani, 1955), local evaporation-precipitation ratios, vital and ecological effects of individual foraminifera species, differential dissolution, and bioturbation. During the 1980's certain sampling strategies were designed to avoid these problems in order to enhance the ice volume effect (e.g. Imbrie et al. 1984).

When Shackleton and Opdyke (1973) managed to apply palaeomagnetic chronostratigraphy on the ocean cores (Figure 2-4) it meant that the supposed climatic cycles could at least be dated roughly and cores (and isotope curves) could be correlated in spite of large distances between sampling points. This also meant that the global validity, in terms of waning and waxing of the great ice sheets, could be determined. Hence the global nature of the $^{18}O/^{16}O$ curves was confirmed. However, as long as the marine cores are mainly dated with the palaeomagnetic reversals the chronology will be based on rather few dating points with interpolated ages between these. This procedure is obviously a bit uncertain, and so in order to overcome this pro-

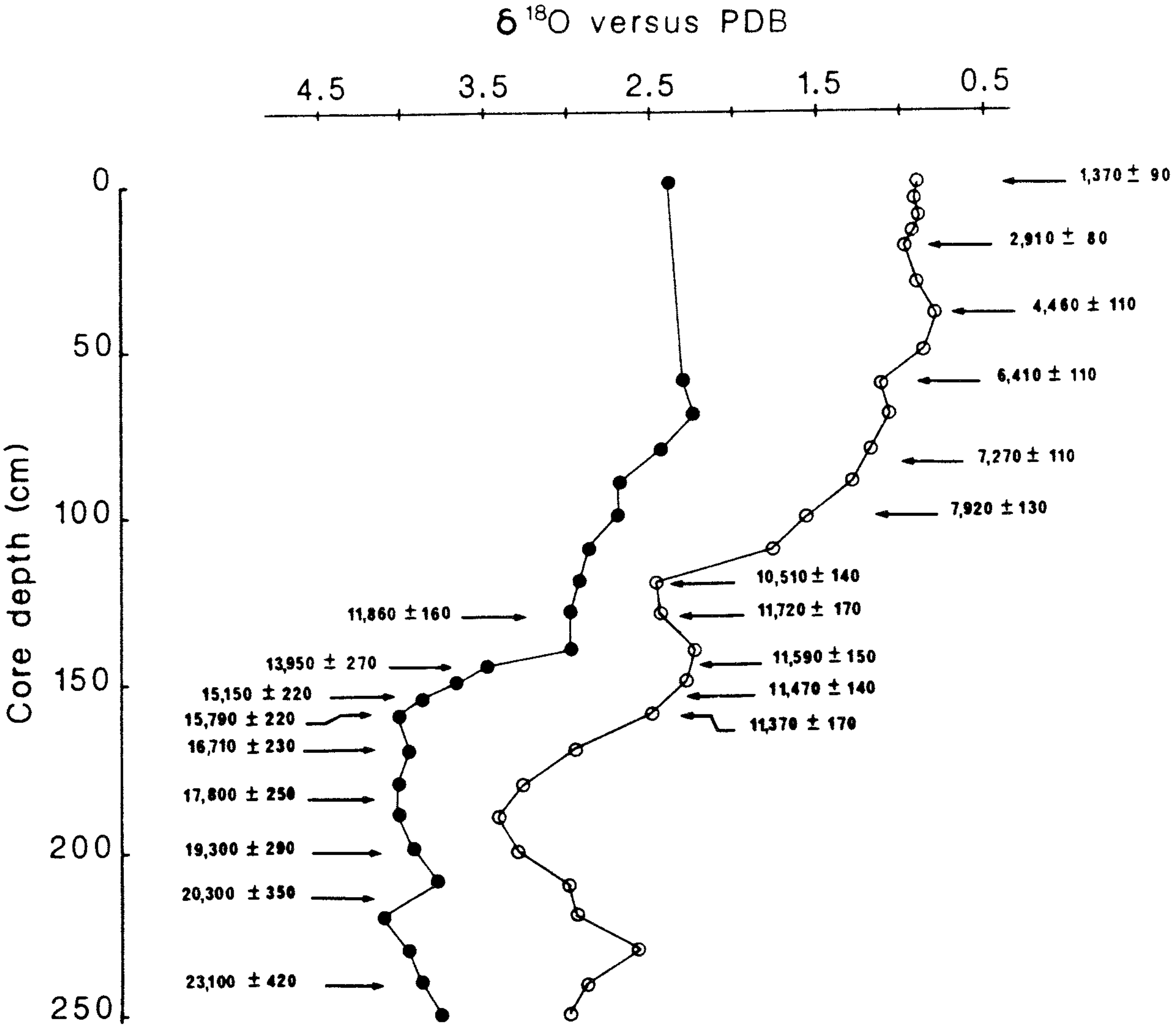

Figure 2-5 Oxygen-isotope records and AMS ^{14}C ages of the two foraminiferal species *G. bulloides* and *N. pachyderma* in the core CH73-139C. From Duplessy et al. (1986). Reprinted with permission from NATURE 320, pp. 350-352. Copyright 1986 Macmillan Magazines Ltd. Direct dating of the oxygen-isotope record of the last deglaciation by ^{14}C accelerator mass spectronometry, by J-C Duplessy, M Arnold, P Maurice, E Bard, J Duprat and J Moyes 1986.

blem the isotope curves have been tuned against different so-called target curves, which will be discussed below. This tuning is of course not necessary for the youngest parts of the curves, where the sediments can be dated by radiocarbon analysis. Bulk samples on sediments poor in organic material are, however, often problematic to date (Björck & Håkansson, 1982) so the use of the rather recently deployed AMS-technique (accelerator mass spectrometry) on pure foraminifera shells (Duplessy et al., 1986; Bard et al., 1987) will certainly create a very detailed chronology (Figure 2-5) from at least the last deglacial phase up to present time.

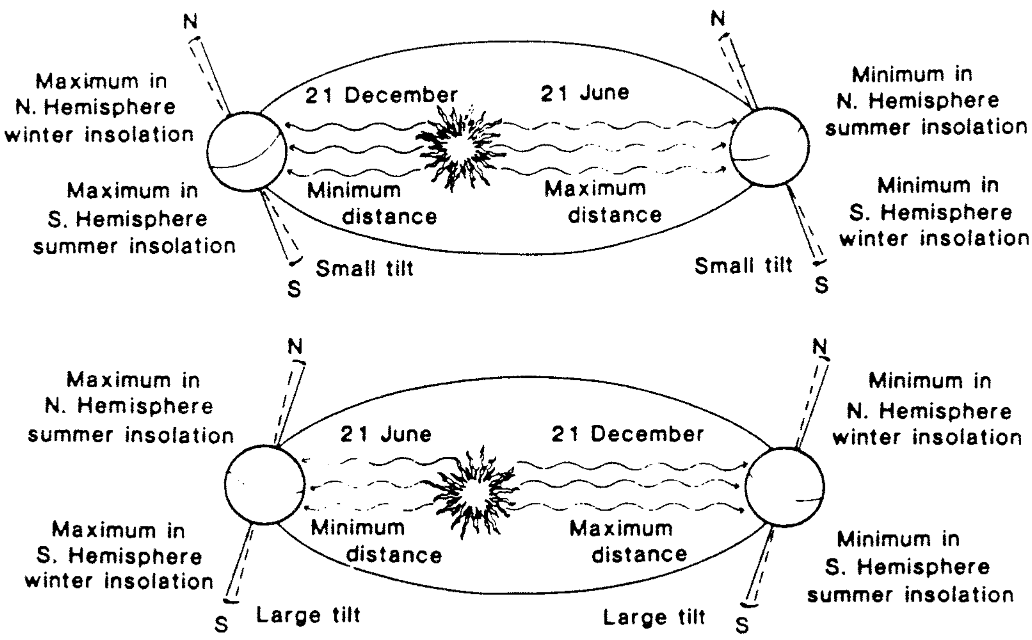

Figure 2-6 Schematic representation of the earth's orbital elements (eccentricity, tilt, and precession). From Ruddiman and Wright (1987). Reprinted by permission from The Geological Society of America. The material was originally published by GSA. The Geology of North America Vol. K-3, pp. 1-12. Introduction. In W F Ruddiman and H E Jr Wright (eds). North America and adjacent oceans during the last deglaciation, by W F Ruddiman and H E Jr Wright 1987.

Already in 1941 the astronomer Milankovitch postulated that the glacial-interglacial cycles were controlled by seasonal, especially summer insolation variations around the latitude of 65° N. These insolation anomalies were caused by cyclic changes in the three Milankovitch orbital parameters (Figure 2-6): eccentricity with cycles of almost 100,000 years (94,000 and 125,000 years) and 413,000 years, the obliquity with a 41,000 year cycle, and the precession with cycles of 19,000 and 23,000 years. The seasonal insolation received at the top of the atmosphere is a function of latitude. At middle and low latitudes the precessional parameters are strongest, while the 41,000 year tilt period is strongest at high latitudes. The eccentricity variations cause only minor shifts in the average annual amount of insolation received (Berger, 1984); however, eccentricity is the primary control on the amplitude of the precessional signal, which in turn controls the seasonal distribution of insolation received (Ruddiman & Wright, 1987). From astronomical calculations we know the form of these orbital changes (Figure 2-7). The calculations of the obliquity variations (Berger, 1984) are accurate to within 5000 years even as far back as 2,500,000 years at the onset of major glaciations (i.e. the beginning of the Quaternary). The ages of the other two parameters' variations are at least as well-known for the last 1,000,000 years, but prior to that the uncertainty is significant.

Hays et al. (1976) were the first to demonstrate detailed evidence of the fact that these orbital cycles dominate the ^{18}O and other paleoclimatic sig-

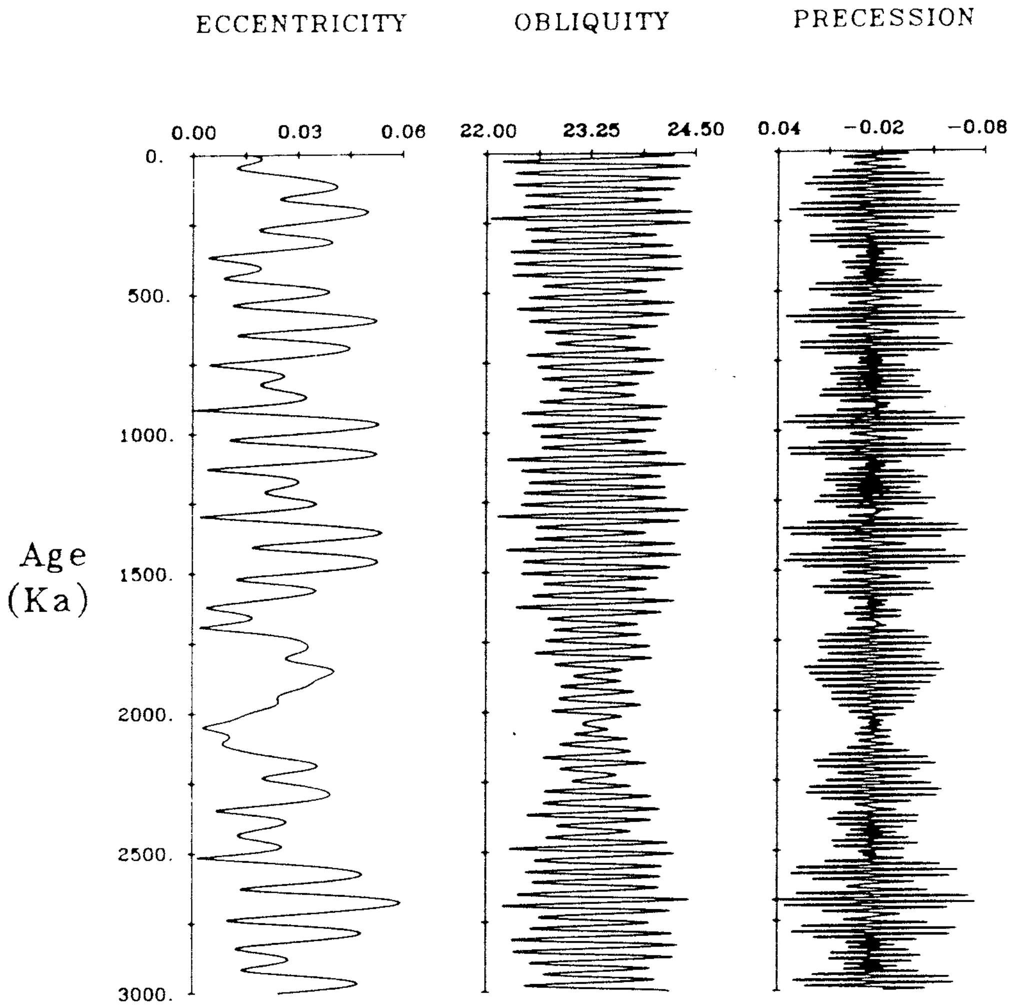

Figure 2-7 Variations over the last 3,000,000 years of the three major components of the earth's orbit around the sun; eccentricity, obliquity, and precession. Precession is shown as the precessional index (e sin w), which incorporates the modulating effect of eccentricity. From Ruddiman and Wright (1987). Reprinted by permission from The Geological Society of America. The material was originally published by GSA. The Geology of North America Vol. K-3, pp. 1-12. Introduction. In W F Ruddiman and H E Jr Wright (eds). North America and adjacent oceans during the last deglaciation, by W F Ruddiman and H E Jr Wright 1987

nals. Later Imbrie et al. (1984) refined the methodology and analysed the last 700,000 years. So the ultimate explanation for the ice-sheet responses at periods of 100,000, 41,000 and 23,000/19,000 years appears almost certainly to be the insolation variations on earth caused by changes in the earth's orbit.

It is generally accepted that the Pleistocene, or the period with larger glaciations, began rather precisely at 2,500,000 B.P. (Figure 2-8), when the initiation of moderate-sized ice-sheets occurred in the Northern Hemisphere (Backman, 1979; Shackleton et al., 1984). However, already 3 million years

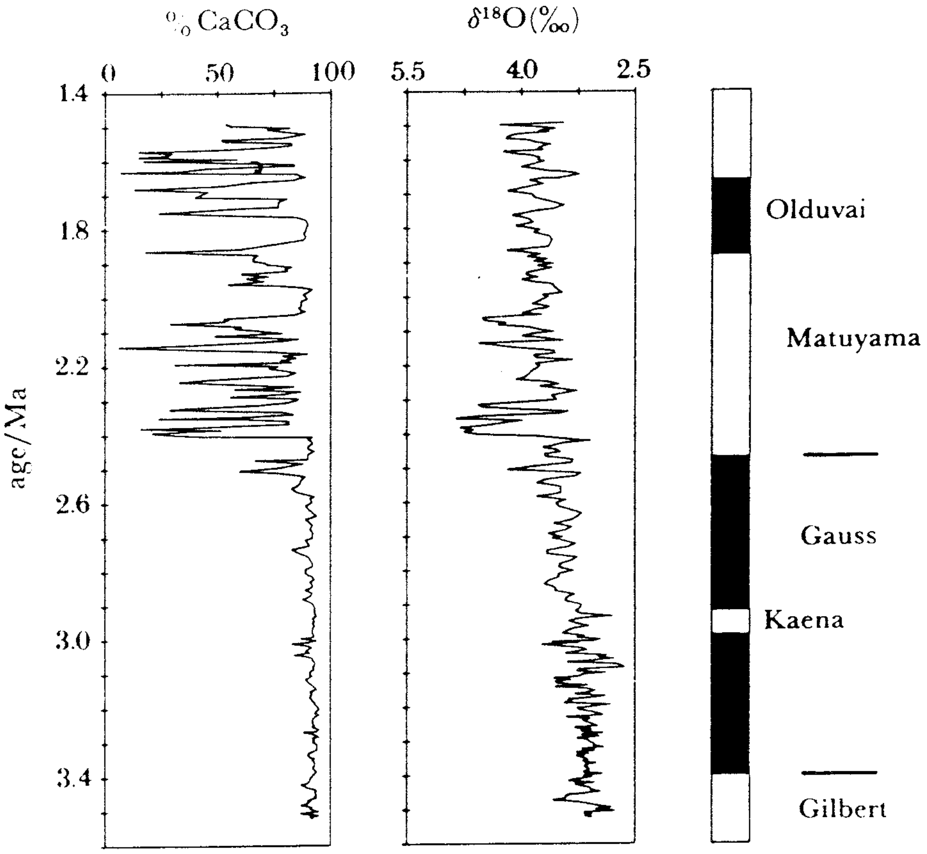

Figure 2-8 Late Pliocene and early Pleistocene % $CaCO_3$ and stable-isotope records at North Atlantic DSDP site 552 (Shackleton et al. 1984). The abrupt decrease of % $CaCO_3$ values at 2,400,000 years B.P. marks initiation of deposition of ice-rafted sand. A correlative late-Pliocene increase in ^{18}O values is superimposed on progressive long-term increase after 3,150,000 years B.P. From Ruddiman and Raymo (1988). Reprinted with permission from The Royal Society. Phil. Trans. R. Soc. Lond. B 318, pp 411-430. Northern Hemisphere climate regimes during the past 3 Ma: possible tectonic connections, by W F Ruddiman and M E Raymo 1988.

earlier large ice sheets with a calving ice-shelf already covered large parts of the Antarctic continent (Jansen, 1990). Between 2,500,000 B.P. and 735,000 B.P. all the measured paleoclimatic parameters were dominated by the 41,000 year cycle (Figure 2-9), which strongly suggests that the ice-sheets fluctuated at the same period (Ruddiman et al. 1986). Sometime between 900,000 and 700,000 B.P. the amplitude of the paleoclimatic parameters increased significantly (cf. Shackleton and Opdyke, 1973) indicating that the ice sheets grew to volumes twice as large as before (Ruddiman and Wright, 1987). The Scandinavian Ice Sheet seems to have reached the Continental Shelf for the first time at c. 1,000,000 B.P. (Jansen, 1990). At the same time the three main paleoclimatic indicators (^{18}O changes, $CaCO_3$ variations, and estimated sea-surface temperatures (SST)) begin to show increasing variance at or near the cycles of precession (23,000/19,000 years) and eccentricity (100,000 years). Gradually the 100,000 year rythm has become totally dominant (Figure 2-10) in the global ^{18}O records (Imbrie et al., 1984) and in the $CaCO_3$ and SST signals in the North Atlantic. Spectral analysis of the global ^{18}O record (Imbrie, 1985) does, however, show that also the tilt (41,000 years) and precession cycles (23,000/19,000 years) over the last 800,000 years have influenced the ^{18}O signal (Figure 2-10).

The dominance of the 41,000 rhythm during the time of less extensive glaciations (the Matuyama chron) is understandable because this orbital parameter is strongest at high latitudes where most of the ice sheets at that time probably occurred. Because summer insolation varies with this rhythm at high latitude it is also natural that this parameter has influenced the timing of ice-sheets during the whole of the Pleistocene since summer ablation of ice and snow is generally regarded as perhaps the most critical parameter for the growth and decay of glaciers and ice-sheets. The importance of the 23,000/19,000 year cycle, with its strong influence at middle and low latitudes, during late-Pleistocene was a natural consequence of the much more extensive and southerly spread ice-sheets during the last c. 1 million years. The great enigma is the dominance of the 100,000 year cycle during late-Pleistocene since changes in the amount of direct insolation available at the eccentricity cycle is almost negligible (Hays et al., 1976; Berger, 1984). One explanation might be that nonlinear responses in the climate system amplify the effects of direct insolation forcing (Hays et al., 1976), e.g. ice-sheets decay much faster than they grow. Imbrie and Imbrie (1980) for example assumed that ice-sheets have a 17,000 year average time constant of physical response to insolation forcing and that the ice decay is four times faster than the ice growth. These ideas did already exist in glaciological theory (Weertman, 1964). With these assumptions it was found that the summer insolation signal at 65° N produced a good approximation of the global ^{18}O record. The perhaps most prominent features of the late-Pleistocene ^{18}O record are the sudden deglaciations that occur approximately every 100,000 years, showing the effect of the nonlinear processes. Generally speaking the major deglaciations coincide with summer insolation maxima in the N hemisphere (Figure 2-11), but it is often hard to explain the abruptness of the deglaciations only in terms of insolation maxima. The last deglaciation is a good example of such a mismatch (Ruddiman and Wright, 1987). This suggests that internal feedback processes (e.g. bedrock rebound, iceberg calving, CO_2, and moisture feedback) are involved in the climate processes (Ruddiman and Wright, 1987).

Oerlemans (1980) and Birchfield et al. (1981) proposed that a suggested delay in bedrock rebound at the onset of deglaciation could explain the accelerated ice-sheet disintegration during the glacial terminations. Peltier and Hyde (1984) and Peltier (1987) refined and modelled these mechanisms. The actual mechanism depends on accentuated isostatic sinking beneath maximum ice-sheet loads. If positive summer insolation anomaly occurs in such a situation the southern ice margin after a while will be situated in the depression that is maintained by the delayed rebound and where ablation is high. This will result in a steep ice gradient which will increase the ice flow and transfer northern ice to the south where it melts rapidly. By this mechanism and when forced by realistic summertime seasonal insolation anomalies

(65° N) Peltier's (1987) ice-sheet model produces sawtoothed 100,000 year cycles of the varying ice volume for the last 800,000 years. These cycles are rather similar to the SPECMAP curve (Imbrie et al., 1984) calculated from ^{18}O records.

Broecker and Denton (1989) even suggests that the glacial cycles were driven by mode changes that are part of a self-sustained internal oscillation that would operate even in the absence of changes in the Earth's orbital parameters. This means that the orbital cycles can merely modulate and pace a self-oscillating climate system (cf. Saltzman et al., 1984). The main point is that the glacial-to-interglacial transitions (terminations) involve major reorganizations of the ocean-atmosphere system (Figure 2-12) and that these terminations constitute jumps between stable modes of operation which cause changes in the greenhouse gas content and albedo of the atmosphere. The main evidence supporting the significance of non-orbital causes of climatic change is the combination of synchronous climate changes of similar severity in both polar hemispheres especially the rapid glacial terminations in both hemispheres. According to Broecker and Denton (1989) the most likely connection between insolation and climate is through impacts of fresh water transport on the ocean's salinity distribution.

2.2.1 The Younger Dryas — an enigma or future analogue?

The importance of the 100,000 year cycle is not the only enigma and challenge for the paleoclimatologists and climate modellers. Another problem is the Younger Dryas event (or stadial), which has been known for a long time in NW European terrestrial records (Hartz and Milthers, 1901; Nilsson, 1935; Jessen, 1935; Iversen, 1942). It comprises a 500 to 1000 year long period (11,000–10,000 B.P.) characterized by a very sudden transition

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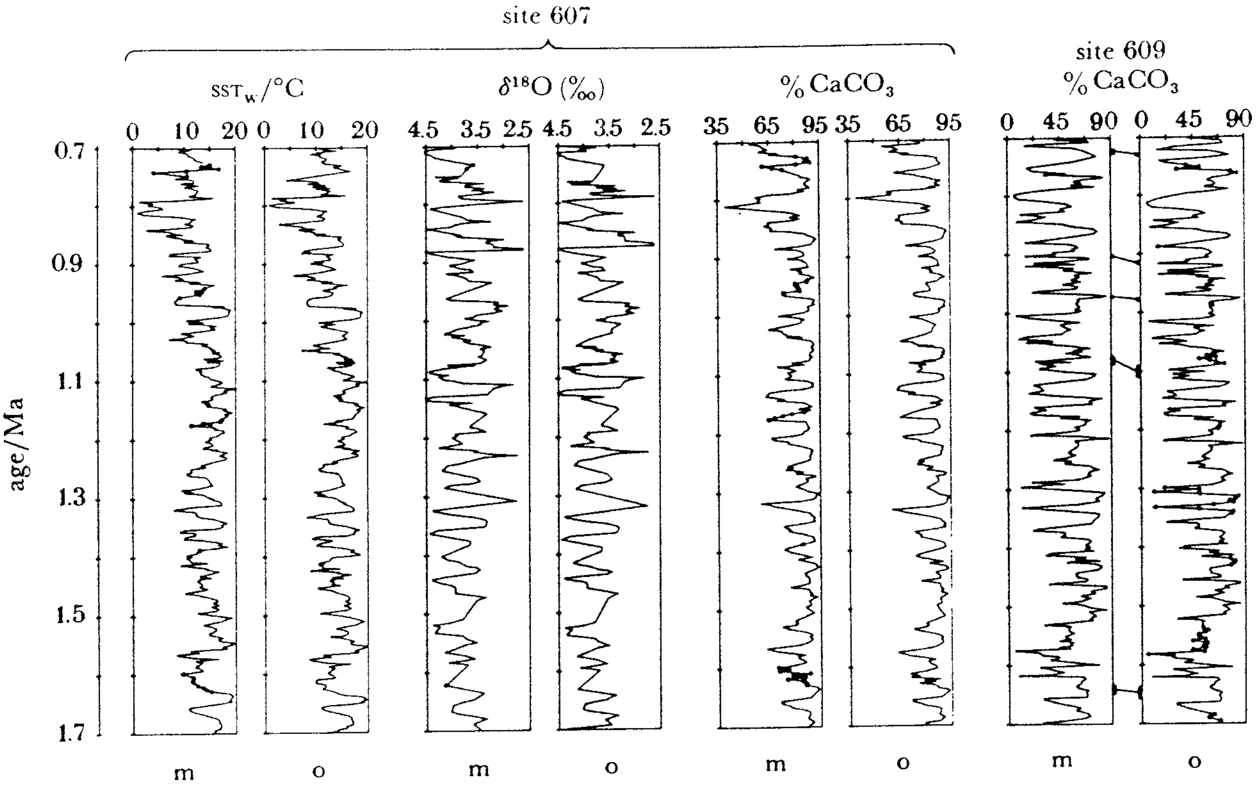
Figure 2-9 (right) Above: early Pleistocene climatic trends at North Atlantic DSDP sites 607 (Ruddiman et al. 1986). SSTw=winter sea-surface temperature. Magnetic timescale based on linear interpolation between magnetic datums; obliquity timescale derived by tuning records to obliquity. Large dots are values spliced into main sequence from offset holes. From Ruddiman and Raymo (1988). Reprinted with permission from The Royal Society. Phil. Trans. R. Soc. Lond. B 318, pp 411-430. Northern Hemisphere climate regimes during the past 3 Ma: possible tectonic connections, by W F Ruddiman and M E Raymo 1988.
Below: spectral analysis of the early Pleistocene records shown above. Spectra from both magnetic (broken lines) and obliquity timescales (solid lines) show prevalent 41,000 year periodicity during the late Matuyama. (a) Site 607, SSTw; (b) site 607, ^{18}O; (c) site 607, % $CaCO_3$; (d) site 609, % $CaCO_3$; age range for all sites, 735,000–1.640,000 years B.P. From Ruddiman and Raymo (1988). Reprinted with permission from The Royal Society. Phil. Trans. R. Soc. Lond. B 318, pp 411-430. Northern Hemisphere climate regimes during the past 3 Ma: possible tectonic connections, by W F Ruddiman and M E Raymo 1988.


site 607
SST_w/°C
$\delta^{18}O$ (‰)
% $CaCO_3$
site 609
% $CaCO_3$
0 10 20 0 10 20
4.5 3.5 2.5 4.5 3.5 2.5
35 65 95 35 65 95
0 45 90 0 45 90
age/Ma
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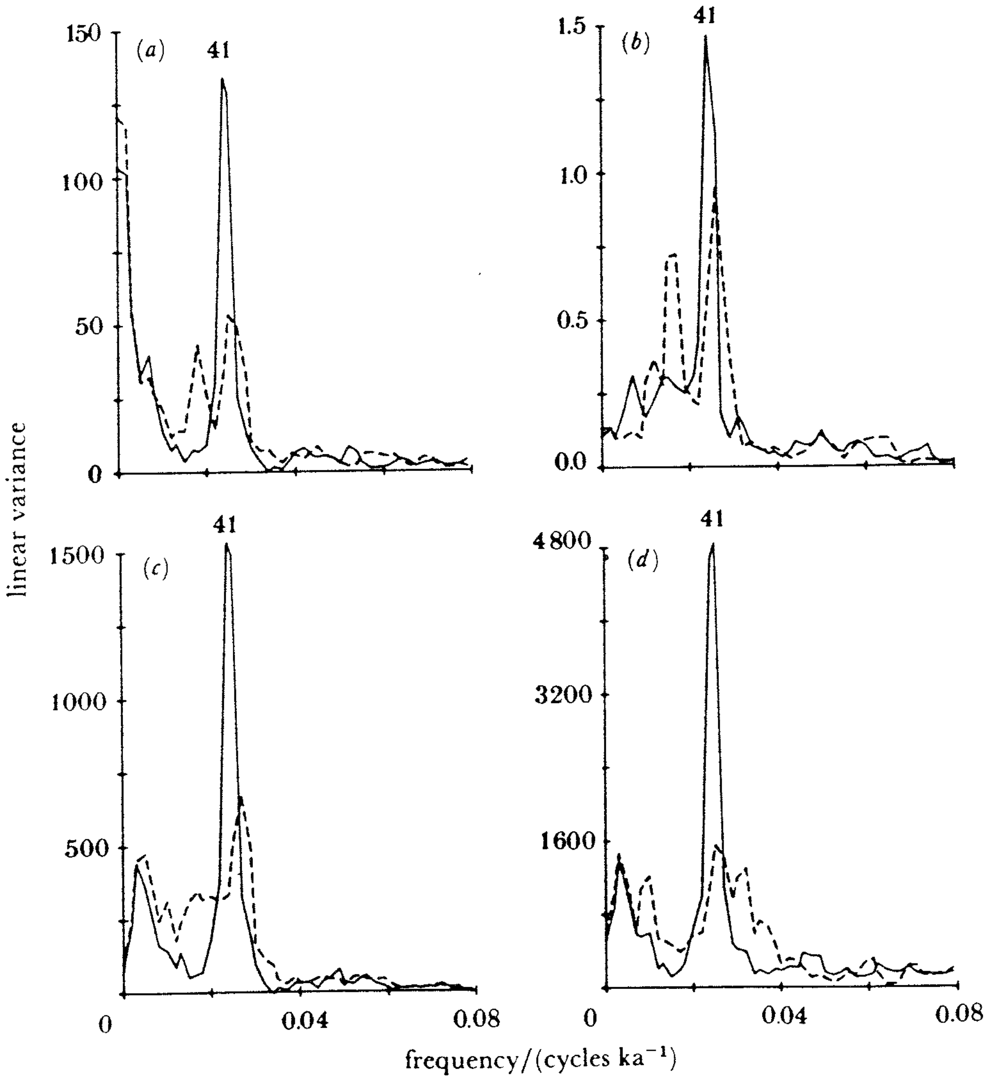
(a)
(b)
(c)
(d)
41
linear variance
150
100
50
0
1.5
1.0
0.5
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1500
1000
500
4800
3200
1600
0
0.04
0.08
frequency/(cycles ka^{-1})

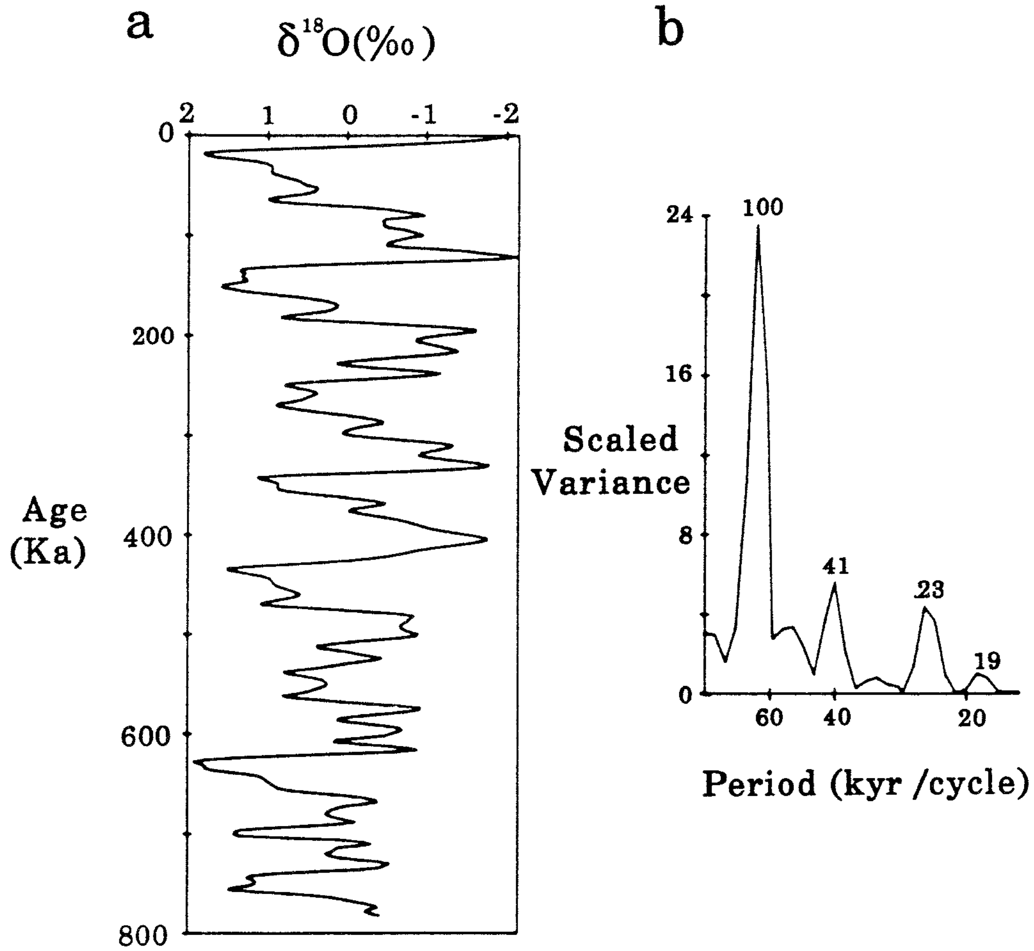
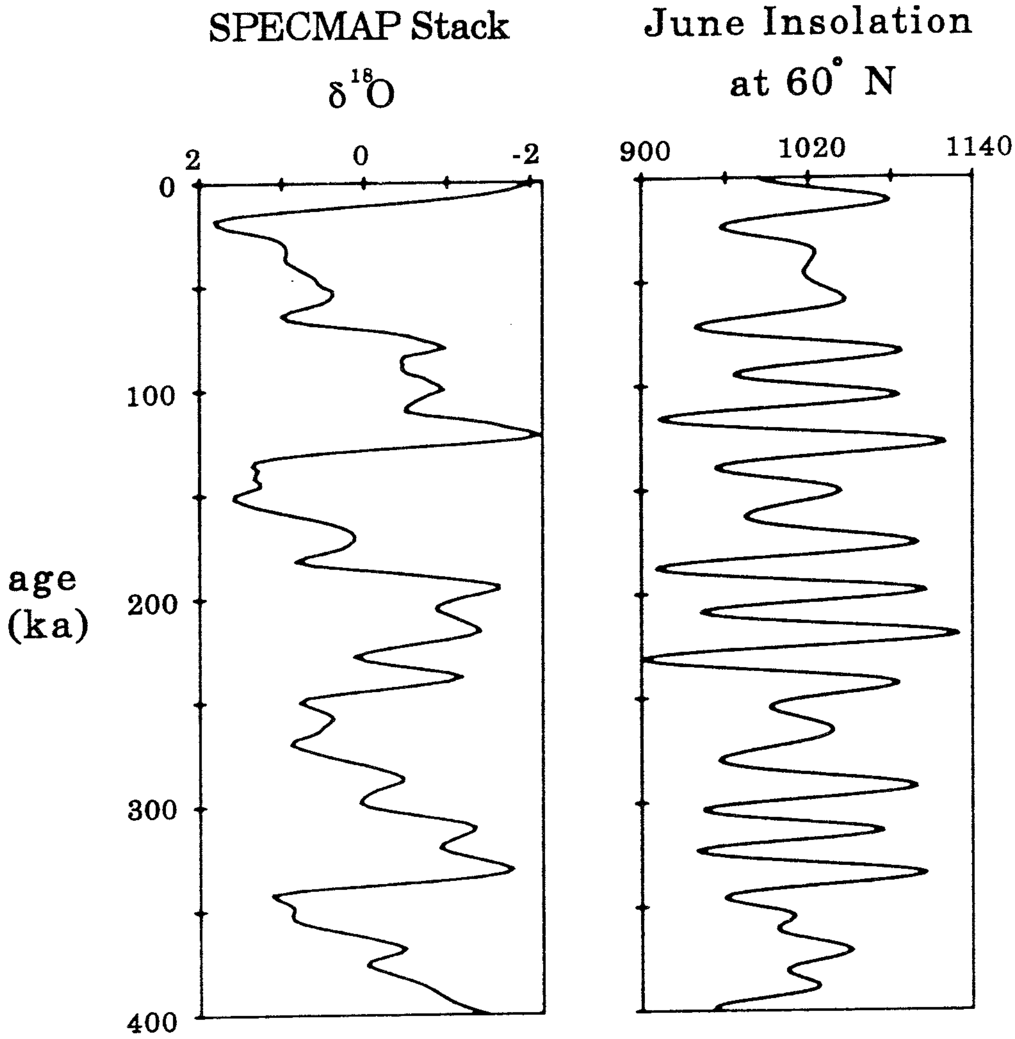


Figure 2-10 (a) Composite global ^{18}O record from Imbrie et al. (1984). Units are standard deviations from the mean. (b) Spectral analysis of ^{18}O record (after Imbrie 1985) in (a). From Ruddiman and Wright (1987). Reprinted by permission from The Geological Society of America. The material was originally published by GSA. The Geology of North America Vol. K-3, pp. 1-12. Introduction. In W F Ruddiman and H E Jr Wright (eds). North America and adjacent oceans during the last deglaciation, by W F Ruddiman and H E Jr Wright 1987

Figure 2-11 (right) Above: changes in ^{18}O over the last 400,000 years (after Imbrie et al. 1984), and changes in June insolation at 60° N over the last 400,000 years. From Ruddiman and Wright (1987). Below: changes in tilt, precession (precessional index), and summer half-year (calendar) insolation at 60° N over the last 25,000 years. From Ruddiman and Wright (1987). Reprinted by permission from The Geological Society of America. The material was originally published by GSA. The Geology of North America Vol. K-3, pp. 1-12. Introduction. In W F Ruddiman and H E Jr Wright (eds). North America and adjacent oceans during the last deglaciation, by W F Ruddiman and H E Jr Wright 1987


SPECMAP Stack
δ18O
2
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June Insolation
at 60° N
900
1020
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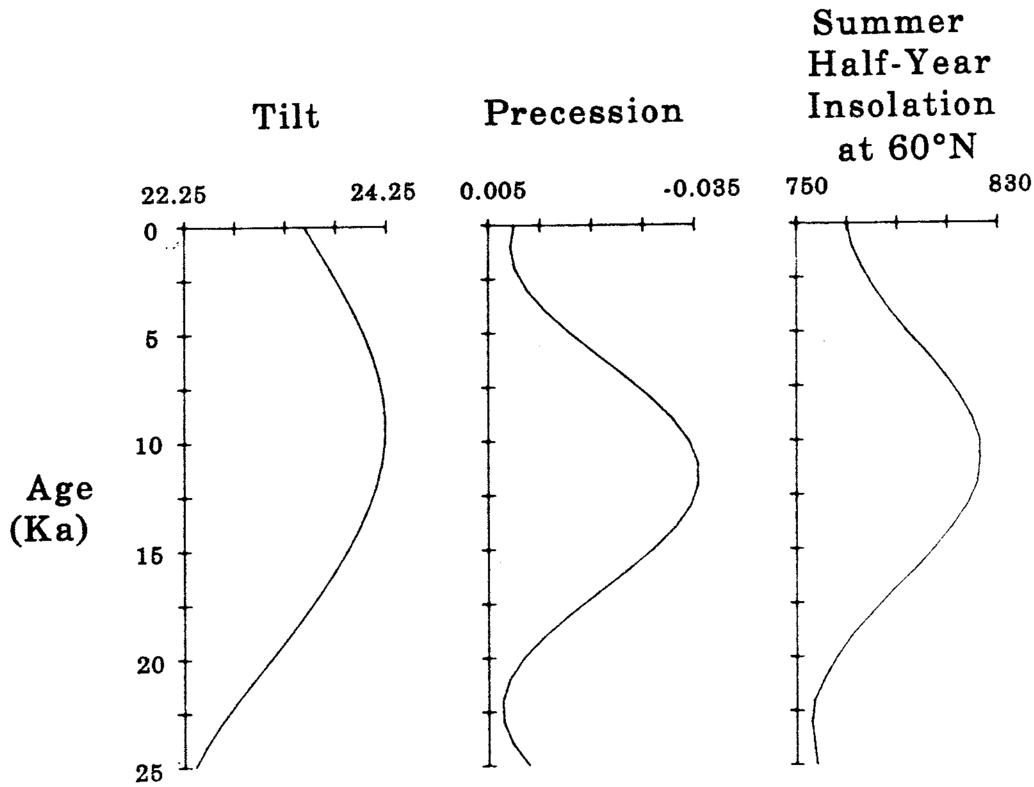
Tilt
Precession
Summer
Half-Year
Insolation
at 60°N
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830
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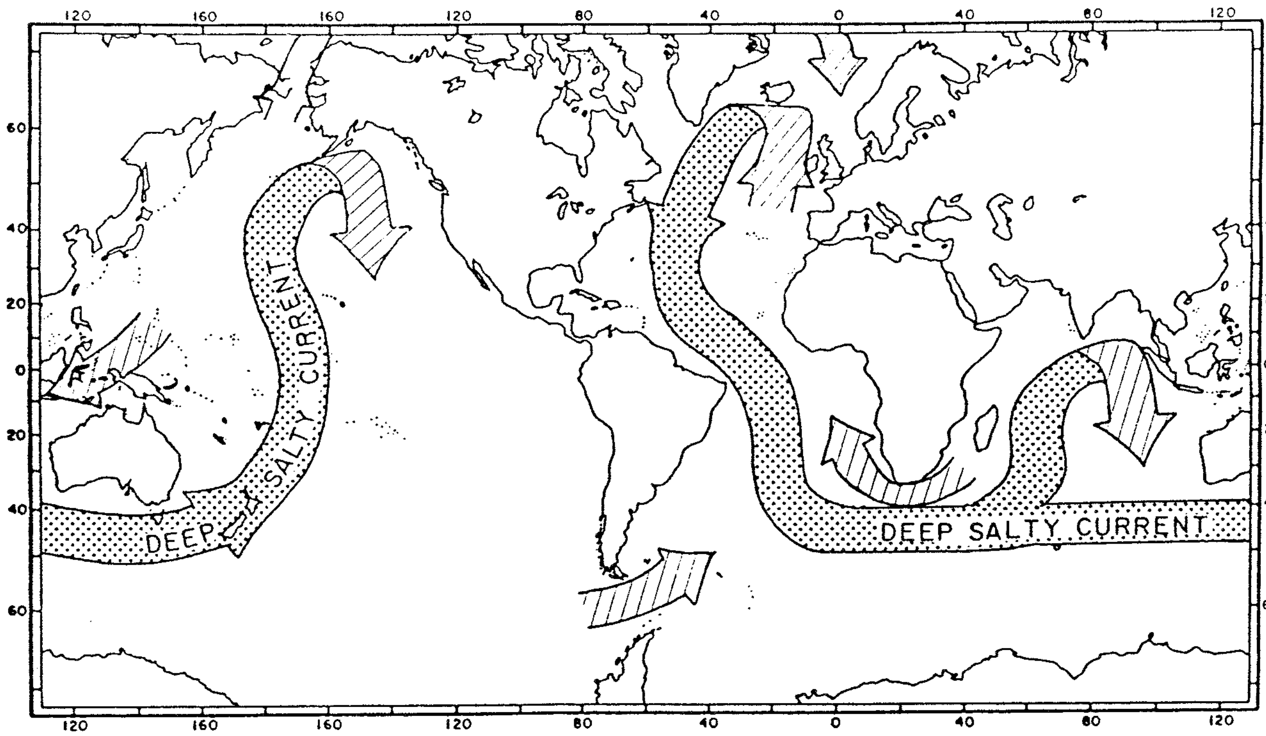

Figure 2-12 A large scale salt transport system operating intoday's ocean compensates for the transport of water (as vapor) through the atmosphere from the Atlantic to the Pacific Ocean. Salt-laden deep water formed in the northern Atlantic flows down the length of the Atlantic around Africa through the southern Indian Ocean and finally northward in the deep Pacific Ocean. Some of this water upwells in the northern Pacific, bringing with it the salt left behind in the Atlantic due to vapor transport. Records from ocean sediments tell us, however, that the so-called Atlantic conveyor was disrupted during glacial time and was replaced by an alternate mode of operation. From Broecker and Denton (1989). Reprinted with premission from Geochimica et Cosmochimica Acta, Vol. 53, Broecker/Denton, The role of Ocean-Atmosphere reorganizations in Glacial Cycles, 1988, Pergamon Press PLC.

(Björck et al., 1987; Dansgaard et al., 1989) into more or less full glacial conditions during 500–800 years followed by the postglacial warming. During the last decades this event has frequently also been found in marine records from the North Atlantic (especially in the northeastern sector) illustrating how the deglacial polar front advanced far south (Figure 2-13) during the Younger Dryas (Ruddiman and McIntyre, 1981). Over the years scientists have claimed its existence in many other parts of the world (Asia, Australia, South America), but most of these finds have been criticised and reevaluated. During the last decade, however, many studies indicate that at least the east coast (especially Nova Scotia) of North America (Mott et al., 1986; Stea and Mott, 1989) was influenced by the Younger Dryas event and possibly also areas much further to the west on the North American continent (Björck, 1985; Shane, 1987). Still unpublished studies from e.g. Glacier Bay in Alaska indicate that it might also be found in that part of North America (H. E. Wright Jr., pers. comm.). The difficulty in understanding

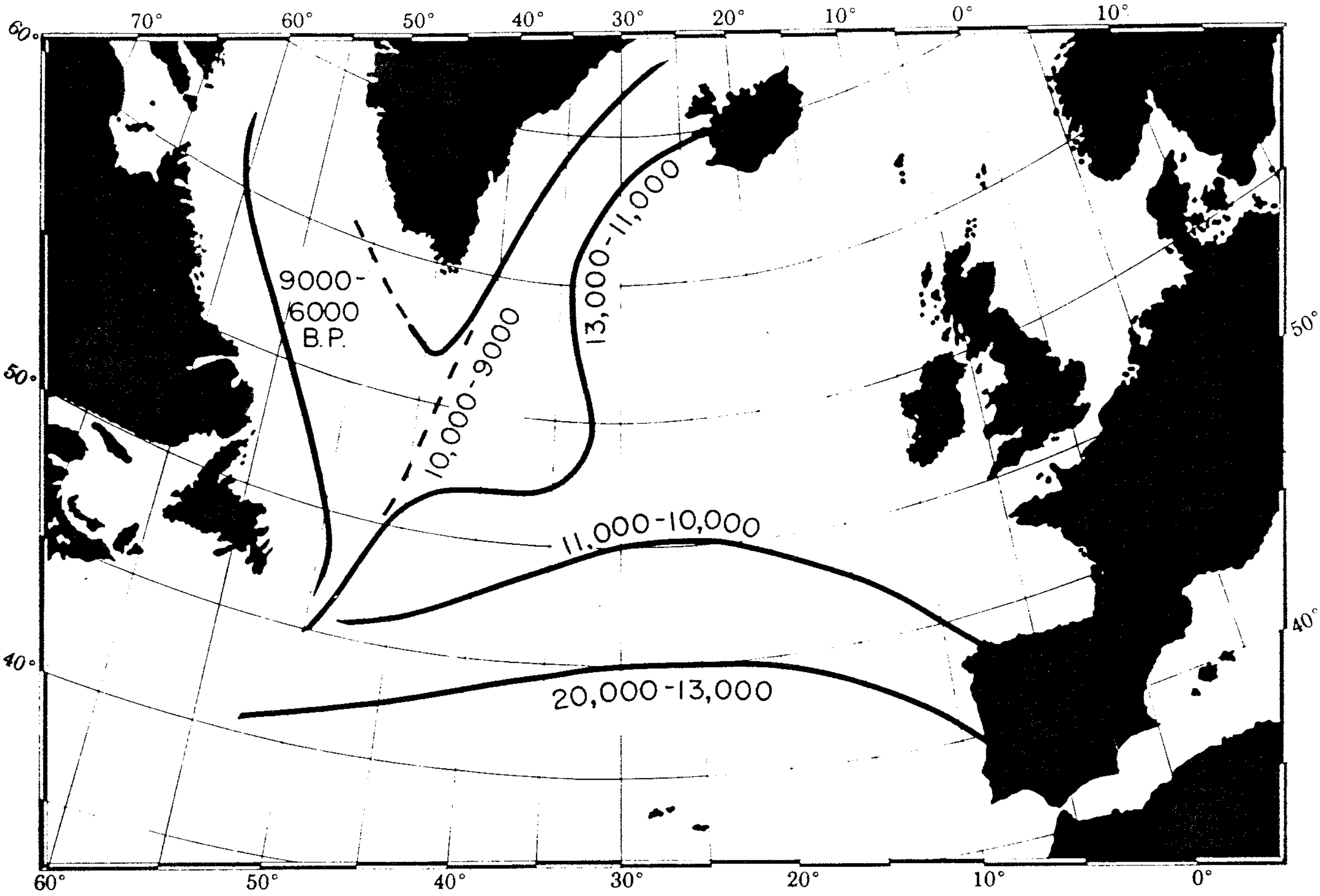

Figure 2-13 Summary of the polar front positions and changes during the last deglaciation. From Ruddiman and McIntyre (1981). Reprinted with permission from Academic Press. Quaternary Reasearch 16, pp. 125-134. The Mode and Mechanism of the Last Deglaciation: Oceanic Evidence, by W F Ruddiman and A McIntyre 1981.

(and modelling) the Younger Dryas event is that it took place during the culmination of the "termination" and at a time when the summer insolation reached a maximum at middle and high latitudes of the Northern Hemisphere. Many different explanations have been presented for this event. The latest but certainly not last contributions on this controversial topic were recently presented by Fairbanks (1989) and Jansen and Veum (1990) from two completely different parts of the North Atlantic region. They strongly support the "French two-step" deglaciation model, based on 16 cores from coral reefs on the coast of Barbados and isotope records from benthic and planktonic foraminifera in the northeastern Atlantic, respectively. Fairbank's (1989) study shows two very distinct melt-water pulses (12,000 B.P. and 9500 B.P., respectively) that are separated by very low melt-water discharges between 11,000 and 10,500 B.P. (Figure 2-14). Furthermore, sea level rise was slow 11,000–10,000 B.P., preceded and followed by rapidly rising sea levels. So in contrast to what Broecker et al. (1985) and Broecker and Denton (1989) postulate, the Younger Dryas seems to have been characterized by low melt-water fluxes. The main argument behind Broecker's

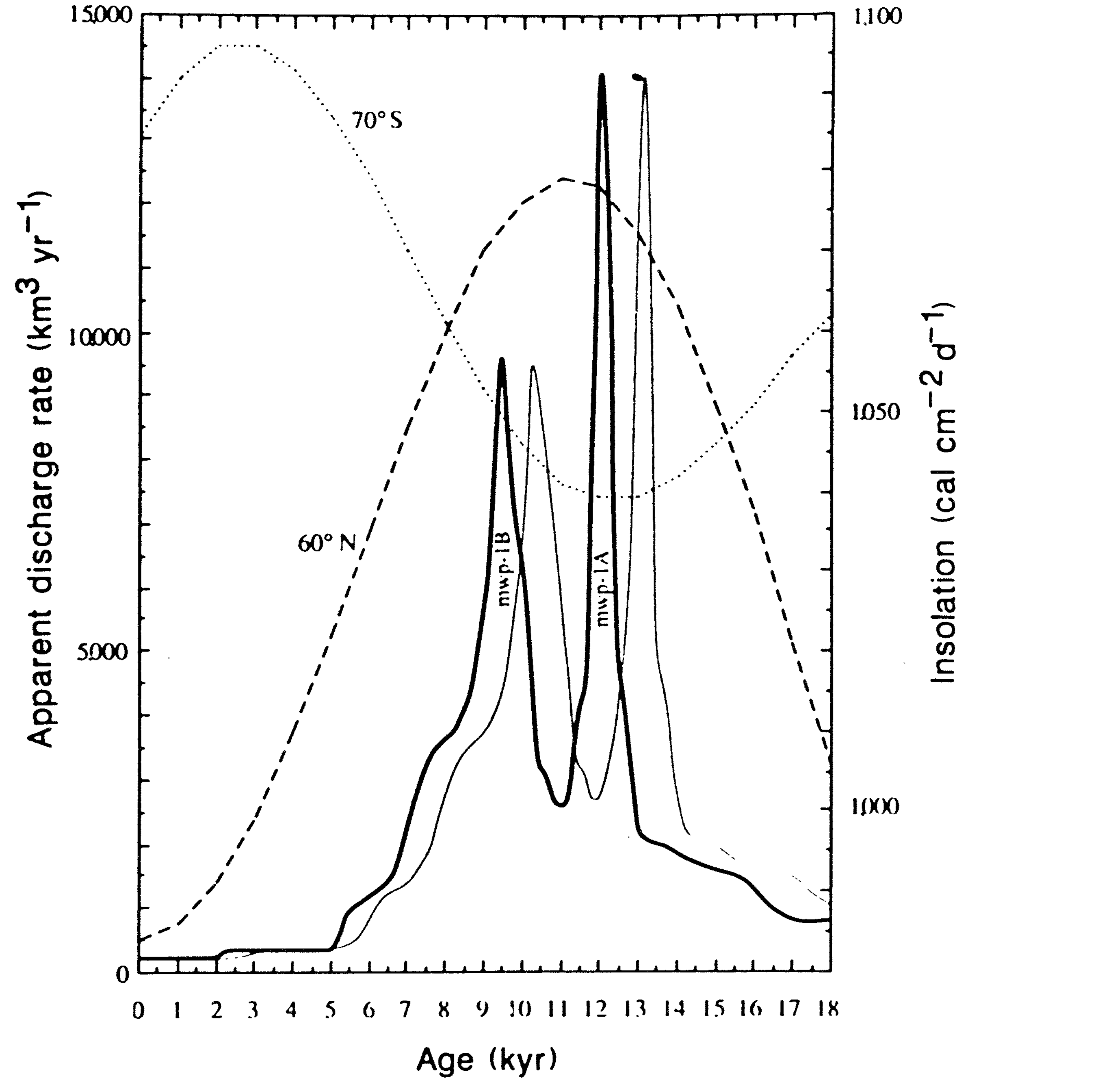

Figure 2-14 The rate of glacial melt-water discharge calculatedfrom the Barbados sea level curve compared to summer insolation. The discharge curve (thick line) is plotted according to radiocarbon years uncorrected for secular changes in atmospheric ^{14}C. Converting radiocarbon years B.P. to calendar years B.P. using the calibration in Stuiver et al. (1986) shifts the ages of mwp-IB mwp-IA as shown by the thin line. Preliminary $^{230}Th/^{234}U$ dates indicate that this correction is too small for mwp-IA (Bard, Hamelin and Fairbanks in prep.). By comparison, the insolation at 60° N (dashed line) and 70° S latitude (dotted line) are plotted against absolute time (Berger 1978). From Fairbanks (1989).Reprinted with permission from NATURE 342, pp. 637-642. Copyright 1989 Macmillan Magazines Ltd. A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation, by R G Fairbanks 1989.

(and colleagues') model (that huge melt-water influx to the North Atlantic during the Younger Dryas upset the density stratification by lowering the surface salinity, which prevented formation of North Atlantic deep water) is thus contradicted by the very important studies of Fairbanks (1989) and Jansen and Veum (1990). The Younger Dryas is thus still a large, enigmatic anomaly during the last deglaciation.

It is thus quite clear from numerous studies that the Younger Dryas was characterized by an arctic climate in the ice-free areas of NW Europe. One of the many questions regarding future glacial stages has concerned the relationship between the extent of these stages and the formation of permafrost. In this context it is therefore worthwhile mentioning that at least the western part of South Sweden was characterized by permafrost conditions from areas just south of the ice-margin (e.g. Johnsson, 1986) down to at least the Laholm Plain in southern Halland (Svensson, 1974, 1982) during the Younger Dryas. Whereas other permafrost indicators found elsewhere in South Sweden might either have been formed during previous stadial conditions or during the Younger Dryas. In any case this tells us that the climatic conditions during the very temporary Younger Dryas were severe enough to produce permafrost in large areas outside the actual ice-margin. We do, however, not know to what depths the permafrost reached during this quite restricted time period. Two questions arise from the above: Is it possible to compare (from a climatic viewpoint) the future glaciations with the Younger Dryas? and, At which future glaciation(s), and at which stage(s) of it (them), can we expect permafrost conditions?

The existence of the Younger Dryas certainly also raises some generally important and thought-provoking questions. Can the regional/global climate change from an interglacial to a glacial state in a matter of decades? How finely balanced is the global climate and what can upset this balance?

2.3 CONCLUSIONS

In spite of different opinions regarding the relative importance of the many different climatic processes that may be involved in triggering the glacial and interglacial cycles there exist certain, very important, agreements concerning these cycles: A) During the last c. 2.5 million years the Earth's climate system has been characterised by continuous interglacial-glacial cycles. B) The interglacial-glacial cycles have, during the last 700,000–900,000 years, been dominated by the 100,000 year rhythm, but the 41,000 and 23,000/19,000 year rhythms do also appear. This is undoubtedly good evidence for the importance of the Milankovitch orbital parameters. C) The interglacial-glacial cycles are characterized by a rather gradual transition from interglacial to glacial conditions. The "pure" interglacial part of the cycle seems to be short (10,000–20,000 years). The glacial part is long and culminates in an extensive glaciation followed by a very rapid termination.

D) The link between insolation anomalies, oceans, atmosphere, and ice-sheets. This is particularly well evidenced by the good correlation between the orbital cycles, the marine records, and the ice core records.

3 CLIMATE MODELS

Today many scientists are developing techniques to create general-circulation models (GCMs) that, at least partly, can explain the relationships between different climate parameters and how they have influenced the past global climate (e.g. COHMAP, 1988). A global array of well-dated palaeoclimatic data is assembled and then the model tries to identify and evaluate causes and mechanisms of climate change. The model is usually run for well-defined time-slices. Comparisons of palaeoclimatic data with the model simulations are essential for the testing and calibration of the GCMs. Hence our confidence in GCM results is dependent on the quality of the palaeoclimatic data, which is far better for the most recent times (i.e. the Holocene) than for older times. which means that the possibility of testing these GCMs. So the GCMs have become particularly useful in explaining the climate changes of the last c. 20,000 years. To a certain degree they have also been used to predict possible climatic changes in the near future. They can, however, still not be used to simulate the long glacial-interglacial cycles. Below we will thus concentrate on models (not GCMs) that mainly relate to the orbital parameters and the waning and waxing of the great ice-sheets (the changing amount of water that was tied to the continental ice-sheets). The latter can be calculated from the globally significant ^{18}O curves.

If we accept that the Milankovitch orbital parameters are the driving mechanisms behind the Pleistocene climatic changes this would obviously imply a challenge for climate modellers to try to predict future climates, since the future behaviour of these orbital parameters is possible to calculate. We do know the past relationship between orbital changes and the changes in global ice volume (^{18}O curves). This is useful in two ways: it makes it possible to understand and analyse the links between changing solar radiation and changing ice volumes and the model outputs can be tested against the ^{18}O curves that are assumed to be the best measures for the global ice volume changes.

The first models were so-called equilibrum models of the form y=f(x). Some of these models (e.g. Shaw and Donn, 1968; Budyko, 1969; Berger, 1977; Schneider and Thompson, 1979; Suarez and Held 1976) showed some promising results. Successes, however, were mixed with failures that are typical of equilibrum models. Equilibrum models presume an instantaneous response of climate to a change in boundary conditions, whereas in reality the true climate system, and the geologic record of it, can lag significantly behind such a change in boundary conditions.

A differential model of the form dy/dt=f(x,y) is required in order to understand the time-dependent behavior (including the observed lags between orbital forcing and the climate response) of the system (Imbrie and Imbrie, 1980). The need for a differential model has been recognized since the time of Milankovitch (1941). A simple differential model over the past 800,000 years was set up by Calder (1974) in which he achieved good results for the last 150,000 years. In 1976 Weertman made a pioneering study in modelling the fluctuations of a continental ice-sheet as a function of changing solar radiation. He not only got significant responses at the forcing cycles (41,000 and 23,000/19,000 years), but also in the range 100,000 to 400,000 years. Pollard (1978) coupled Weertman's ice-sheet model to an energy-balance model of the atmosphere, which became a considerable advance in modeling capacity. This model gave a reasonably accurate simulation of the climate record of the last 150,000 years. Although the above cited models represented considerable progress in applying all available knowledge of climate physics to create the most realistic model on the behaviour of the great ice-sheets their results fell short of an adequate simulation. In order to overcome this problem Imbrie and Imbrie (1980) followed an alternative strategy by trying to capture the important features of more complex models in a class of simple models. This makes it possible to disclose the full information power of a given class of models by adjusting the model parameters until an optimum model is identified. The big advantage of developing simple models is that it can be tested and developed over and over again and that it is easier to understand what features are responsible for failures and what are the basis of its success. Six radiometrically dated events were used in tuning the model and the model ages came out very well. The model did also respond to the five main orbital (and ^{18}O) cycles (Figure 3-1), but as usual the 100,000 cycle was less prominent than in the orbital data and ^{18}O records. As will be shown in the next section Imbrie and Imbrie's (1980) model output simulates the ^{18}O curves over the last 350,000 years in a rather impressive way. It should, however, be pointed out that there are still no models (stochastic) that can provide a good basis for estimating the degree of unpredictability in climate. A significant event like the Younger Dryas, which obviously occurred and dramtically influenced large parts of the North Atlantic region, is hardly understandable and has been very difficult to model. Attempts have been made (e.g. Rind et al., 1986) but the output models have often failed to simulate the climate indicated by the geologic records. As such it is an important and instructive experience for climate modellers and palaeoclimatologists.

3.1 A MODEL OF FUTURE CLIMATE

The model output from Imbrie and Imbrie's (1980) response model is shown in Figures 3-1 and 3-2. The agreement between the output, the orbital input (with the time-lag), and the two assumed globally valid ^{18}O curves (from the southern Indian Ocean and from the Pacific Ocean, respectively) is striking. According to the output curve the present day interglacial conditions are gradually changing to glacial conditions. This transition began c. 6000 B.P. and will culminate in 23,000 years in what could be called a stadial. The model output values of that stadial indicates that the amount of ice tied to continental ice-sheets will be similar to the situation we know from isotope stage 4, when parts of Scandinavia (e.g. Mangerud, 1981; Lundqvist, 1986) and North America most likely were glaciated. In marine sediments from Denmark (Jutland) Lykke-Andersen (1987) shows that her zone V, which is correlated to isotope stage 4, marks the first high arctic period (based on foraminifera) during the Weichselian in the Jutland-Kattegatt area. There are no certain evidence how extensive this glaciation was. It might have reached down to the Göteborg area (Hillefors, 1969) but not further westwards or southwestwards (Lykke-Andersen, 1987). According to some Polish TL-dates it might have filled the Baltic basin down to the Polish coast in the form of a Baltic ice lobe (Lagerlund, 1987). According to Berglund and Lagerlund (1981) there are no traces of glaciation in Skåne before the Late Weichselian. This glaciation was thus much more restricted than the Late Weichslian one. Although the actual extent of it is unknown the ^{18}O curves suggest that almost full-glacial conditions prevailed and that this period was the second coldest (at least in the amount of ice that was tied up in glaciers) during the Weichselian. So according to Imbrie and Imbrie's (1980) model parts of Sweden will start to become glaciated at least within 20,000 years with a culmination at c. 23,000 years from now.

The above outlined stadial conditions will, according to the model, be followed by an interstadial climate, similar to the Middle Weichselian climate that preceded the Late Weichselian glaciation maximum. In most parts of Sweden, with a possible exception for the Swedish mountain areas, the ice-sheet will have disintegrated at c. 30,000 years from now. That situation will probably prevail for c. 20,000 years. If this interstadial can be comparable with Lagerbäck's (1988) Tärendö Interstadial, which he tentatively correlates with isotope stage 5a, but in our opinion also may be correlated

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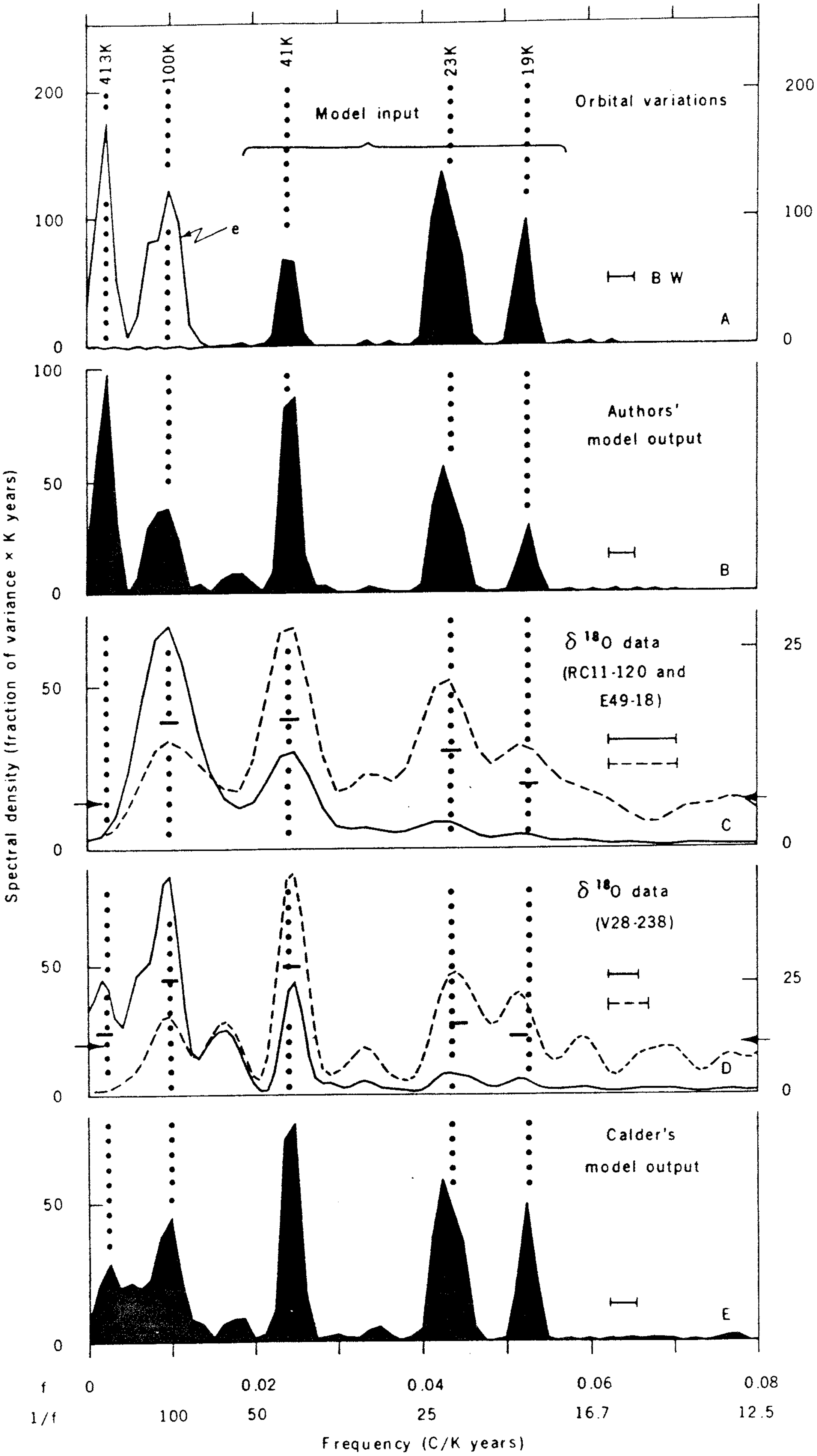
Figure 3-1 (right) Spectral density as a function of frequency f (cycles per 100 years). Climatic spectra are compared with input spectra of two climate-response models. (A) Solid peaks are the spectrum the input to our optimum model. Open peaks are the spectrum variations in eccentricity calculated from data of Berger (1978). The dominant periods of these orbital variations, calculated by Berger (1977), are indicated in thousands of years and by dotted vertical lines. (B) Spectrum of the output of our optimum model. Note the low-frequency peaks not found in the model input. (C) Solid line shows spectrum of variations in ^{18}O from a 468,000-year record in the southern Indian Ocean. Dashed line shows spectrum of same data after prewhitening with a first-difference filter to emphasize details in the higher frequencies. Data are from Hays et al. (1976), with the spectrum calculated on their TUNE-UP time scale. (D) Solid line shows spectrum of variations in ^{18}O from a 730,000-year core record in the Pacific Ocean. Dashed line shows spectrum of the same data after prewhitening with a first-difference filter to emphasize details in the higher frequencies. Isotopic data are from Shackleton and Opdyke (1973). (E) Spectrum of the output of a response model due to Calder (1974). From Imbrie and Imbrie (1980). Reprinted with permission from American Association for the Advancement of Science, SCIENCE 207, pp. 943-953. Modelling the climatic response to orbital variations, by J Imbrie and J Z Imbrie 1980. Copyright 1980 by the AAAS.


413K
100K
41K
23K
19K
Model input
Orbital variations
e
B W
A
Authors'
model output
B
δ ^{18}O data
(RC11-120 and
E49-18)
C
δ ^{18}O data
(V28-238)
D
Calder's
model output
E
Spectral density (fraction of variance × K years)
200
100
0
50
25
f
1/f
0.02
0.04
0.06
0.08
12.5
16.7
Frequency (C/K years)

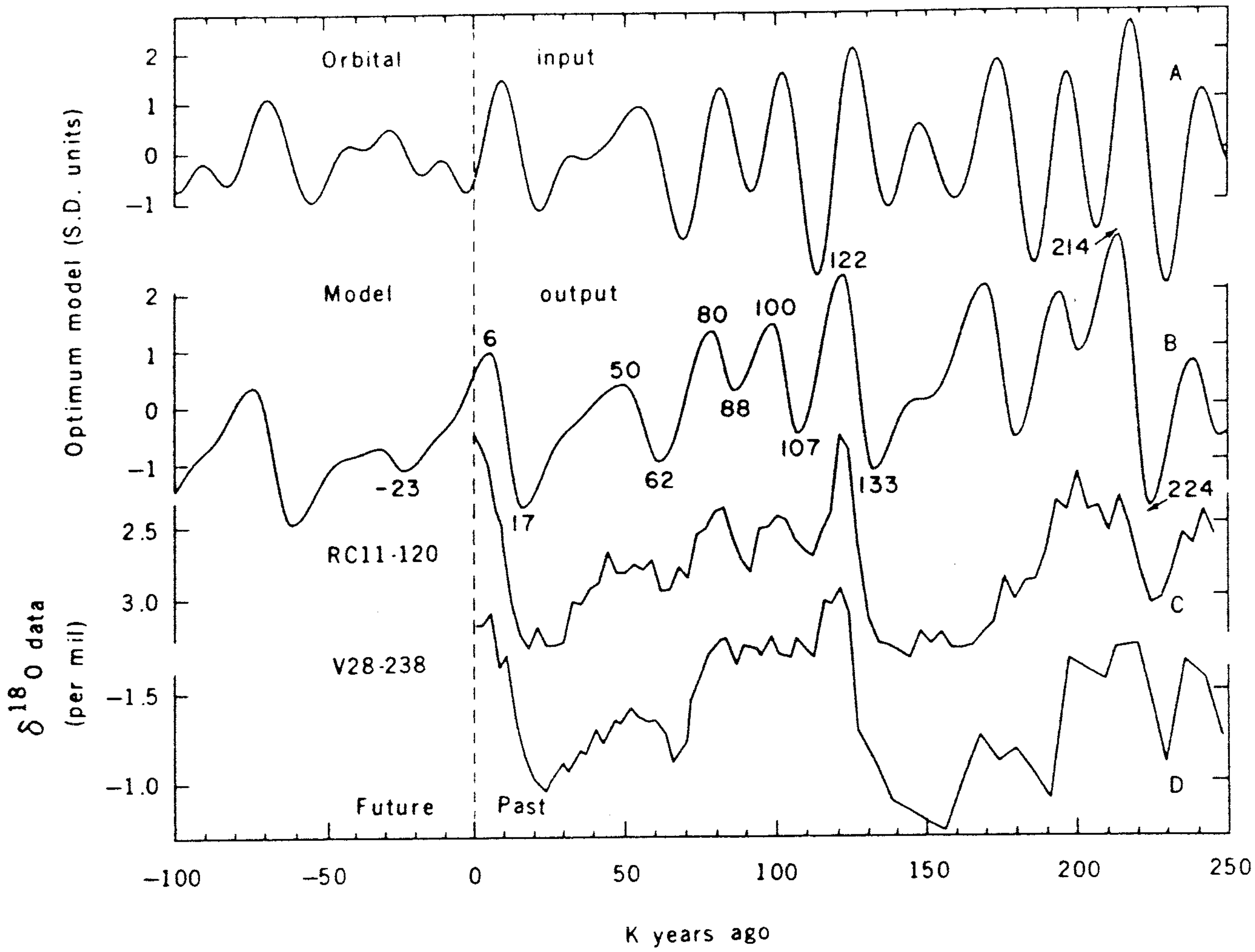


Figure 3-2 Input and output of Imbrie and Imbries (1980) response model compared with isotopic data on climate of the past 250,000 years. (A) Orbital input corresponding to an irradiation curve for July at 65° N. (B) Output of a system function with a mean time constant of 17,000 years and a ratio of 4:1 between the time constants of glacial growth and melting. Ages of selected maxima and minima are given in thousands of years. According to this model, the influence of orbital variations over the next 23,000 years will be to enlarge continental ice-sheets. (C) Oxygen isotope curve for deep-sea core RC11-120 from the southern Indian Ocean (Hays et al. 1976). (D) Oxygen isotope curve for deep-sea core V28-238 (Shackleton and Opdyke 1973) from the Pacific Ocean. Curves C and D are plotted against the TUNE-UP time scale of Hays et al. (1976).

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with e.g. stage 3, it means that at least the end of it might be characterized by an extremely harsh climate in Norrland, comparable to Greenland or Antarctic conditions, with extensive periglacial activity.

At c. 50,000 years from now the main glaciation of the coming glacial will start to build up. According to Imbrie and Imbrie (1980) the maximum glaciation will be reached at c. 63,000 years from now. According to the model output the glaciation will be as extensive as the Weichselian glaciation maximum at 18,000–20,000 B.P., i.e. isotope stage 2, which would mean that the whole of Scandinavia, Finland, the Baltic republics, western Russia, large parts of Poland and the German republics become glaciated. The effect on the Swedish bedrock would in that case be similar to what happened during the Late Weichselian.

The termination of the assumed future glaciation maximum seems to become as rapid as it was at the end of the last glaciation maximum between 15,000 and 10,000 B.P. If this will be the case the uplift features of the Swedish bedrock will probably also be rather similar to what we know about the Late Weichselian-Holocene uplift history. What we, however, know much less of, actually almost nothing, are the Early and Middle Weichselian isostatic changes, i.e. how the bedrock behaved during the upbuilding of a continental ice-sheet.

According to Imbrie and Imbrie's (1980) model the next interglacial optimum will be reached at c. 75,000 years from now with a gradual change into more or less full-glacial conditions (cf. isotope stage 4) at c. 100,000 years from now.

If we accept that the above outlined scenario is likely to occur it implies that large parts of Sweden will be glaciated three times in the coming 100,000 years. This also implies that the Swedish bedrock will experience three major down-warping phases and two uplift phases during that time period.

4 PLEISTOCENE - HOLOCENE SEA LEVEL CHANGES

One of the main features of the Pleistocene are the repeated glacial/interglacial cycles. A main effect of these cycles has been significant global sea level oscillations. These oscillations are best demonstrated by the marine ^{18}O records, whose signals are dominated by the amount of water tied up in glaciers and continental ice-sheets. Other good indications of the past sea level history are e.g. coral reefs, coastal wet land stratigraphies in tropical regions, older shorelines, and isolations and transgressions of lake basins. According to the ^{18}O curves and estimations of maximum ice volumes the sea level changes have been estimated to the order of 80–150 m between the interglacials and the main glacial phases. For many years attempts were made to construct a globally uniform sea level curve (=eustatic curve) for the last 18–20,000 years (=since the last glacial maximum). This has, however, shown to be impossible since the eustatic changes are composed of different non-globally uniform components (except for changes in the volume of the hydrosphere). The importance of geoidal changes for understanding past sea level changes led Mörner (1976a) to define eustasy as vertical sea level changes regardless of causation. Three main types of eustasy can be distinguished: A) Glacial-eustasy, which is controlled by ocean water volume changes that are caused by the waning and waxing of glaciers and continental ice-sheets. B) Geoidal eustasy, which is controlled by changes (in time and space) in the Earth's gravitational field. C) Tectono-eustasy, which is controlled by changes in the ocean basin volume. The most important factors are sedimentation and volume variations of ocean ridge systems.

The sea level history of an area is mainly a function of the eustatic, glacio-isostatic, hydro-isostatic, and tectonic changes it has undergone. The eustatic components have been described above. Glacio-isostasy has influenced glaciated regions, as well as their peripheral areas, to a greater or lesser extent depending on the temporal and areal extent and the thickness of the regional/local ice-sheets/glaciers. The hydro-isostatic changes are caused by sea level changes that deflect the ocean floor to attain isostatic equilibrum. E.g. a eustatic sea level rise of 3 m would depress the ocean floor by 1 m (1/3 of the sea level change), which in turn would result in a mean uplift of the continents by c. 2 m (the oceanic area is approximately twice the continental area). It is evident that sea level changes, wherever they occur, are a complex of many interrelated and complicated factors, which are often also difficult to quantify.

The latest contribution to our knowledge about sea level changes since the last glacial maximum is a sea level curve based on drillings in *Acropora palmata* coral reefs on the south coast of Barbados (Fairbanks, 1989) covering the last 18,000 radiocarbon years. The advantage by this study is that a) this coral species is restricted to the upper five metres of water, b) the tec-

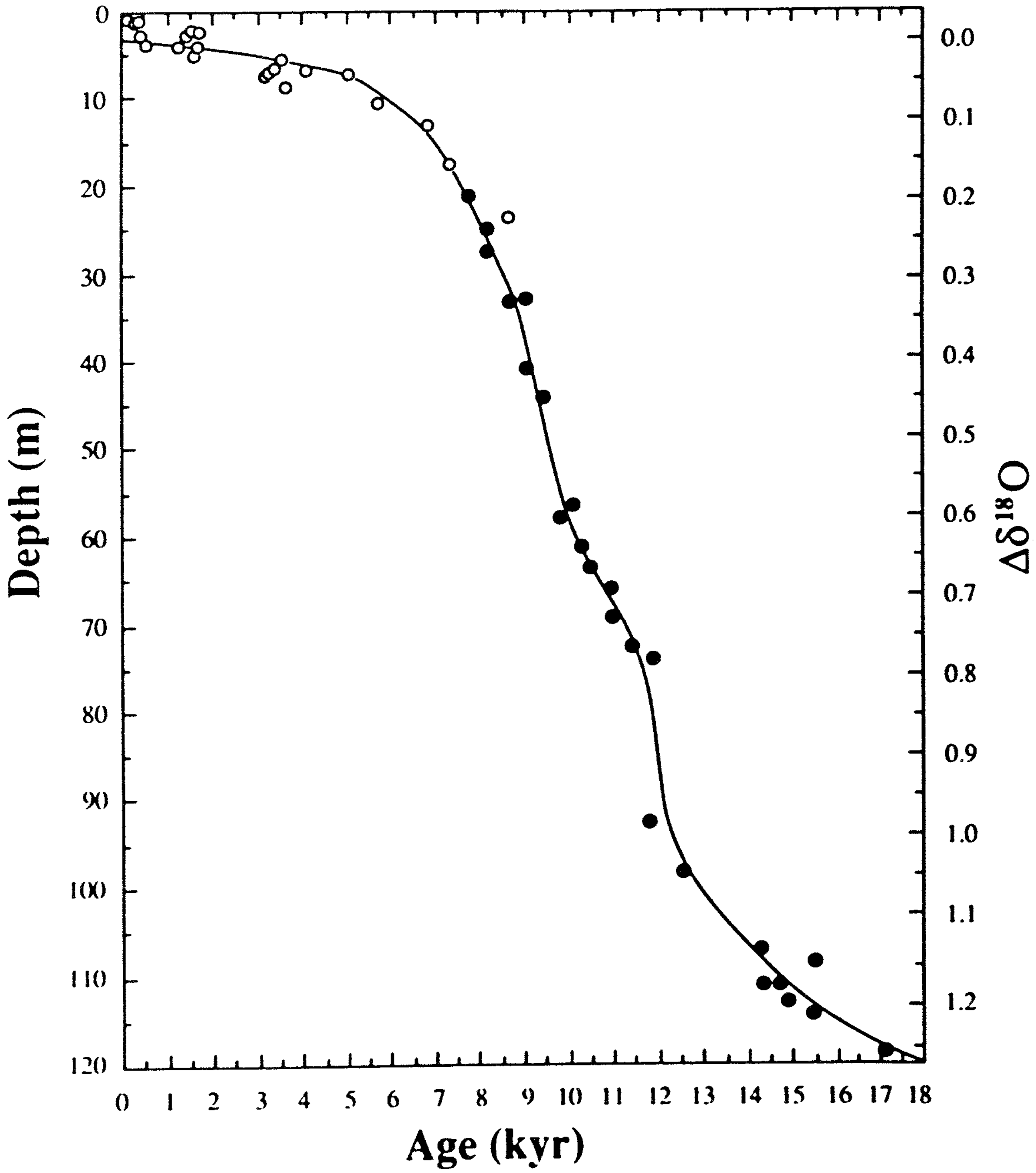


Figure 4-1 Barbados sea level curve based on radiocarbon-dated *A. palmata* (filled circles) compared with *A. palmata* age-depth data (open circles) for four other Caribbean island locations. All radiocarbon ages in this figure are corrected for local seawater ^{14}C by substracting 400 yr from the measured radiocarbon ages but they are not corrected for secular changes in atmospheric ^{14}C levels (Stuiver et al. 1986). The Barbados data are corrected for the estimated mean uplift of 34 cm/1000 yr. The right-hand axis of the Barbados sea level curve is scaled to the estimated ^{18}O change of mean ocean water using the calibration in Fairbanks and Matthews (1978). From Fairbanks (1989). Reprinted with permission from NATURE 342, pp. 637-642. Copyright 1989 Macmillan Magazines Ltd. A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation, by R G Fairbanks 1989.

tonic uplift rate is calculated to be very small (c. 34 cm/1000 yr), c) the offshore morphology is gently sloping and extremely well-mapped with three parallel offshore ridges containing thick sequences of *A. palmata*, and d) Barbados is situated in an area where sea level was hardly influenced by changes in the equipotential surface that were due to the changing gravitational attraction of the melting Laurentide Ice Sheet. According to the curve obtained (Figure 4-1) sea level was situated at c. -120 m at 18,000 B.P. The hydro-isostatic effect is, however, not estimated and an island moving with the sea floor will record the full sea level change (Fjeldskaar, 1989) and not the net change. If Barbados has sunk in pace with the sinking sea floor, which is not likely, the net sea level rise will be in the order of 80-90 m, which is also what e.g. Mörner (1976b) has calculated from NW Europe.

Ever since the last glacial maximum sea level has been rising, but with slightly different rates (Figure 4-1). The most significant rates occurred 13,000–11,500 B.P. and 10,000–9000 B.P. as a probable result of high meltwater fluxes from the Fennoscandian and North American ice-sheets. As a result of this unloading (incl. the hydro-isostatic component) of former glaciated regions, these regions have experienced a more or less extensive uplift during late-, and postglacial time. As such Fennoscandia is not only a suitable "sea level laboratory" (Mörner, 1980a), but also a region where the geodynamic features and processes of the mantle can be studied and understood. The research history on the Fennoscandian relative sea level and uplift history has been treated exhaustively by Mörner (1979) and will thus only be shortly dealt with here.

5 LATE WEICHSELIAN–HOLOCENE UPLIFT IN SWEDEN

5.1 SHORT HISTORICAL BACKGROUND TO STUDIES ON LAND-UPLIFT

The study of land-uplift in Sweden has a long history, beginning in the late 17th century when Urban Hiärne (Hiärne 1706) began a survey of the withdrawal of the sea. During the 18th and 19th centuries many Swedish scientists studied the land-uplift, such as Emanuel Swedenborg, Anders Celsius, Carl von Linné, P Kalm, Efraim O Runeberg, Sven Nilsson, Axel Erdmann, Gerhard De Geer, Henrik Munthe, Leonard Holmström, Nils Olof Holst and many others. The work during these two centuries led to the understanding that the observed uplift had an isostatic origin caused by the glacial deloading at the end of the last glaciation (De Geer 1888). The results obtained during late 19th and early 20th centuries concerning the elevation of major shoreline stages of different ages (e.g. the highest shoreline, the Ancylus shoreline and the postglacial transgression shoreline (Litorina and Tapes), (De Geer 1910, Munthe 1902, 1910ab, Lundqvist 1928), as well as the "processes" behind the complicated Baltic development are generally still valid.

During the 20th century two other important properties of the land uplift were established:

- The principles of the eustatic interaction on the uplift were described by Nansen (1922) and Ramsay (1924);
- The finding that the geoid of the earth is not constant through time (Mörner 1977, 1979, Fjeldskaar & Kanestrøm 1980, Fjeldskaar 1989) has invalidated the use of global sea level rise curves.

The last century has been a period of intense collection of shore level data, from both morphologically identified shorelines and stratigraphic sequences. Especially, over the three last decades the use of ^{14}C dating has yielded detailed knowledge about the shore displacement from many areas in Sweden.

5.2 SEA-LEVEL MEASUREMENTS AS SOURCES OF INFORMATION ON LAND UPLIFT

Registrations of land uplift by measuring sea-level relative to watermarks chiseled into bedrock is a method practised since the early 18th century, and the use of tide gauges and mareographs had begun already in 1774 in Stockholm (Sjöberg 1984, Ekman & Sjöberg 1984, Ekman 1986).

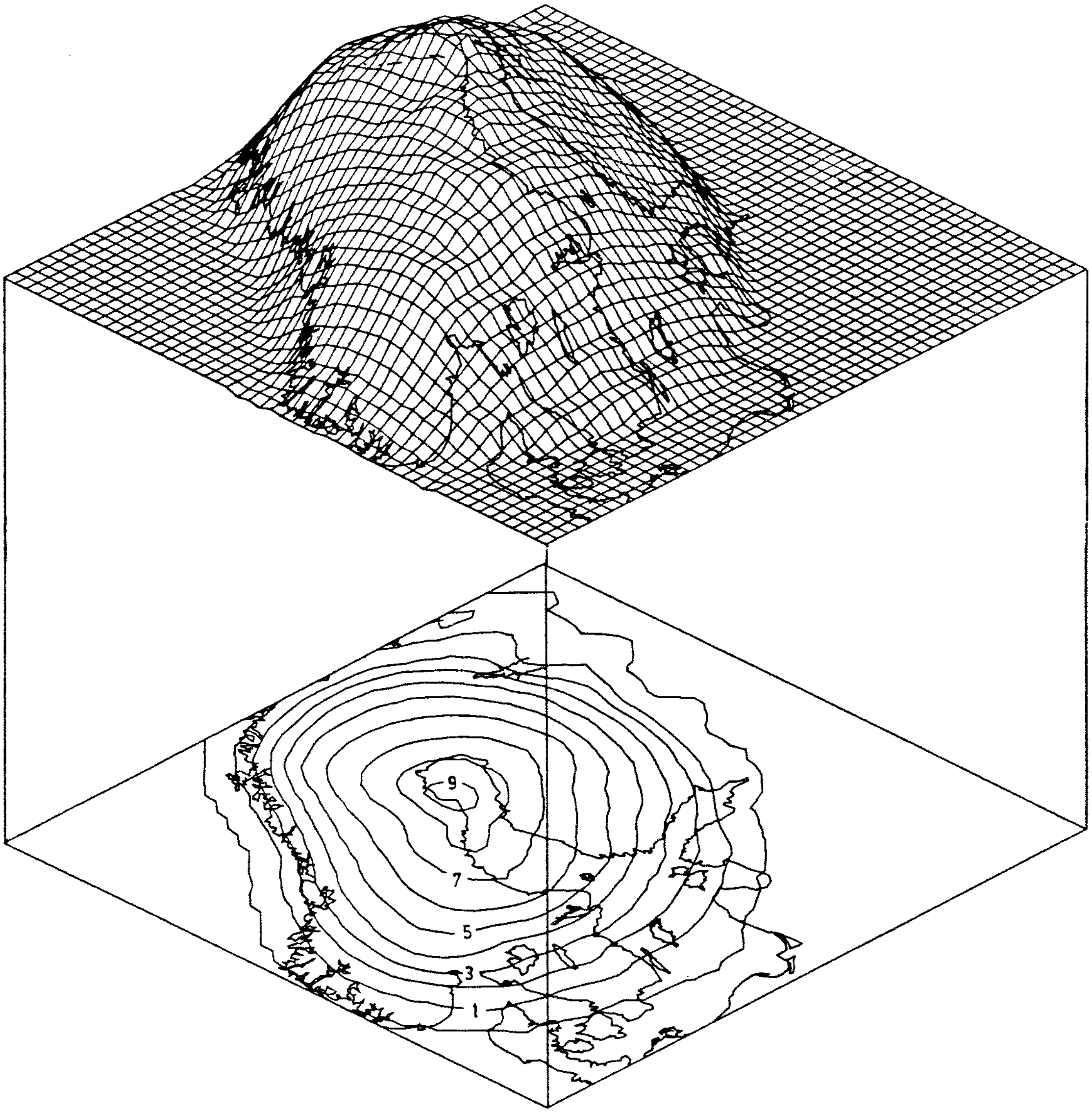


Figure 5-1 Observed land uplift in the Fennoscandian area; in mm/yr. Based on data from Ekman (1987).

5.3 PRECISION LEVELLINGS

Repeated high-precision levellings, related to mareograph registrations of sea-level, give rather exact values for the apparent uplift for the period between the levellings. Such a study, based on the Swedish high-precision levellings from 1886–1903 and 1951–1967, has been presented by Ussisoo (1977), and compilations of data from the whole Fennoscandian uplift area have been done by e.g. Ekman (1987). Around 1997 the third high-precision levelling should be finished resulting in a much more detailed picture of the present land uplift.

This method only gives results on the land-uplift relative to sea-level. Estimates of the rate of change in sea-level show rather large variations. A value of c. +1mm/yr (Lisitzin 1974) is often used, which, however, may be too high (Pirazzoli 1989). Ekman (1986) calculated variations in apparent land-uplift based on observations since 1774. He found that the uplift rate relative to sea-level was 1 mm/yr higher between 1774 and 1884 than it was from 1884 to 1984. He explained this as a higher eustatic rise during the latter period as a result of climatic amelioration since the "Little Ice Age".

5.4 SHORELINE STUDIES

Studies of ancient shorelines of the sea (and lakes) is the most common method to document the land uplift for the period following the deglaciation until recent time. Former shorelines can be identified either as a morphological feature, such as a beach ridge, or from stratigraphical evidence (e.g. from the sediment sequence formed during the isolation of a lake from the sea, or in some cases inundation by a rising sea). The shorelines are usually dated by the radiocarbon method, either by dating organic deposits (shells, peat or wood) within or below beach deposits, or by dating organic lake sediments deposited in connection with the isolation of the lake.

By dating the isolation of several lakes at different elevation situated within a small area, very accurate knowledge of the shore displacement through time can be obtained. By mapping the elevation of dated shorelines in larger areas it is possible to display the spatial pattern of uplift, which can be especially detailed if distinct morphological shorelines can be used.

One problem is that the datings (except in a few cases when annual laminations have been used for dating) are performed by the ^{14}C method. The ages are not true years, but depend instead on the ^{14}C production in the atmosphere, and show a deviation of up to c. 900 yrs from the sidereal timescale. For the period when the deviation is known (0– c. 8000 yrs ago) correction is possible, but for the time before, the deviation is largely unknown.

5.4.1 Morphologically identified shorelines

The study of morphological shorelines, mainly beach ridges, delta surfaces, boulder fields, boulder rims, and eroded beach notches, has a long tradition. The major shorelines were identified, both on their morphology and stratigraphy, (e.g. finds of shell, and biostratigraphic studies on peat and gyttja beneath transgressive shorelines) and were geographically mapped already by the late 19th and early 20th centuries (e.g. De Geer 1888, Munthe 1902, 1910ab). During the period c. 1920–1950 intense studies of shorelines took place by measuring and correlating not only major but also minor shorelines. Unfortunately many correlations from this period were based on neither stratigraphical evidence nor continuously mapped shorelines, but seemingly only by correlations in shoreline diagrams constructed in various ways. These graphical correlations, often performed throughout very large areas (e.g. Nilsson 1953, 1968) are today regarded to be of little value in regional studies of the land uplift. However, in intensely studied, limited geographic areas these older studies could be of great value, especially when applying ages derived from modern shore displacement studies.

The general opinion today regarding morphologically identified shorelines is rather restrictive. Shorelines correlated regionally need to be morphologically pronounced, either by size or shape or by being the first or last in a sequence. Shorelines of the pronounced type are formed during the culmination of transgressions, or during rather long periods of stable (or nearly stable) sea-level. Such shorelines were formed;

1. By the transgressive Ancylus Lake in the Baltic,
2. During the (metachronous) culmination of the Postglacial transgressions (the Litorina transgression(s) in the Baltic).

The shorelines distinguished by their appearance in a sequence are:

3. The highest shoreline, which is the uppermost shoreline of the sea or the dammed Baltic.
4. The last Baltic Ice Lake shoreline, formed in the Baltic just before the final drainage.
5. The first Yoldia shoreline, formed in the Baltic just after the drainage of the Baltic down to sea-level.

Further, to be certain of the age of a shoreline, datings are needed, preferably both from the central and marginal areas of a shoreline's occurrence. These could be obtained by ^{14}C dating and biostratigraphically investigated organic material, either within or below a transgressive beach deposit, or in lake-sediment stratigraphy directly related to the shore displacement.

In the following section these five distinct shorelines will be further discussed and their appearance and elevation are summarized.

5.4.1.2 The highest shoreline

The highest shorelines were formed by the sea or the dammed Baltic, normally at the time of deglaciation, but at places possibly during a later transgression. This shoreline is thus metachronously formed in different parts of Sweden. Its age is c. 13,000 B.P. in southern Sweden and c. 8000–8500 B.P. in northern Sweden.

The advantage of this shoreline is that it is rather easy to observe, and is thus generally the most frequently studied shoreline. The feature identified as the highest shoreline is important when regional syntheses are made, e.g. whether the feature is the highest limit of wave-wash, a boulder rim, or an erosion terrace could produce significant differences in elevation. In many locations there also exist higher shorelines, formed in local dammed glacial lakes, which should not be mistaken as having been formed by the sea or the dammed Baltict. Due to its metachronous character, the gradient of the highest coastline displays the combined effect of the water-level changes of the sea or the Baltic, and the isostatic land uplift. It could also be emphasized that due to its metachronous character it is **not** formed as a single shoreline, but rather as the upper end of a sequence of shorelines where each one has a greater gradient than the actual highest shoreline (Ramsay 1924) (Figure 5-2).

As discussed later, the highest shoreline gives possibilities for identifying areas with differential uplift (e.g. tectonic zones) which has been active after deglaciation. The reason is firstly that the shoreline is old, commonly formed at deglaciation, which means that all later tectonic events

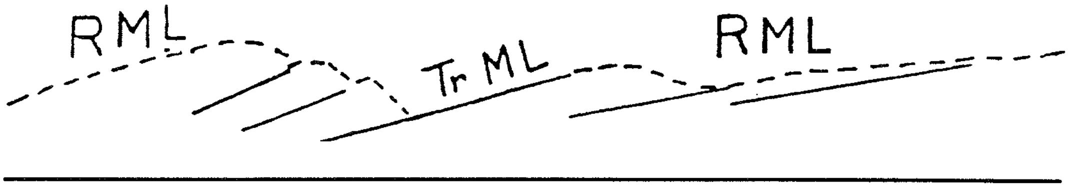


Figure 5-2 Illustration on how the highest shoreline is formed during deglaciation as a sequence of metachronous shorelines normally with a gradient less than the highest shoreline. From Ramsay (1924).

should show up, and secondly that this shoreline can usually be determined for many sites in an area. However, using this shoreline for such a purpose should be done with care, considering the effect of regional sea-level changes, as well as more local glacio-eustatic influence.

Regional synthesis: Based on a compilation of published data, isolines of the highest shoreline were constructed (Fig. 5-3). By combining these shorelinedata with a digital elevation model (with 500 m spacing between datapoints compiled by the National Land Survey of Sweden) land areas above the highest shoreline could be illustrated as a map (Fig. 5-3).

From the Swedish west coast information on the highest shoreline is given in the geological maps and their descriptions (from Sveriges Geologiska Undersökning, SGU ser. Aa och Ae). Some recent studies includes Gillberg (1952), Wedel (1967, 1971), Hillefors (1969), Mörner (1969), Digerfeldt (1979), Lind (1983), Sandgren (1983), Robison (1983). Recently new results from Skåne and southern Halland (Lagerlund 1983, 1987, Adrielsson 1984, Malmberg-Persson 1988) indicate that the deglaciation sea-level was low and later followed by a transgression, reaching higher levels (in Skåne even much higher) than assumed in previously published studies on the highest shoreline. However, the question of deglaciation and highest shoreline in this region is debated and Ringberg (1989) presents results favouring the more traditional view of the Late Weichselian glacial stages, and thus a lower elevation of the highest shoreline in Skåne.

The elevation of the highest shoreline decreases southwards from c. 215 m a.s.l. at Höljes far north in the Klarälven river valley (Gillberg 1952), c. 160 ma.s.l. at Dals Ed (Gillberg (1952), 134 ma.s.l. at Hunneberg (Digerfeldt 1979), to c. 90 ma.s.l. at Sandsjöbacka 15 km south of Göteborg (Påsse 1987). Further south, in the debated area, the highest shoreline is found at either c. 45 m a.s.l. south of Lund according to Ringberg (1989), or somewhat above 75 m a.s.l. in central Skåne according to Malmberg-Persson (1988).

In the southern part of the Swedish east coast from Scania up to the Middle Swedish end-moraine zone the highest shoreline was formed by the Baltic Ice Lake which at most times was dammed above sea-level. A summary of published as well as new information from this area were done by Agrell (1976).

The ice-marginal position at the Baltic Ice Lake drainage in the northern part of the Middle Swedish end-moraine zone shows a drop of the highest shoreline of c. 25 m (Bergsten 1943, Björck & Digerfeldt 1989) due to the final drainage of the Baltic Ice Lake. During the following c. 600–800 years (Svensson 1989, 1991a) the Baltic was in level with the sea. During the next c. 500 yr followed the dammed Ancylus stage, but afterwards again the highest shoreline was formed at the true sea-level.

The elevation of the highest shoreline continues to increases towards the north on the east coast and is c. 160 m at Kilsbergen in Närke (Halden 1934), and c. 200 m a.s.l. 40 km east of Gävle. Close to Örnsköldsvik in Ångermanland the highest value of the highest shoreline, c. 285 m a.s.l. is recorded. Further to the north as well to the west the highest shoreline is found at successively lower elevations, e.g. c. 170 m a.s.l. at Pajala, northern Norrbotten (G. Lundqvist 1961).

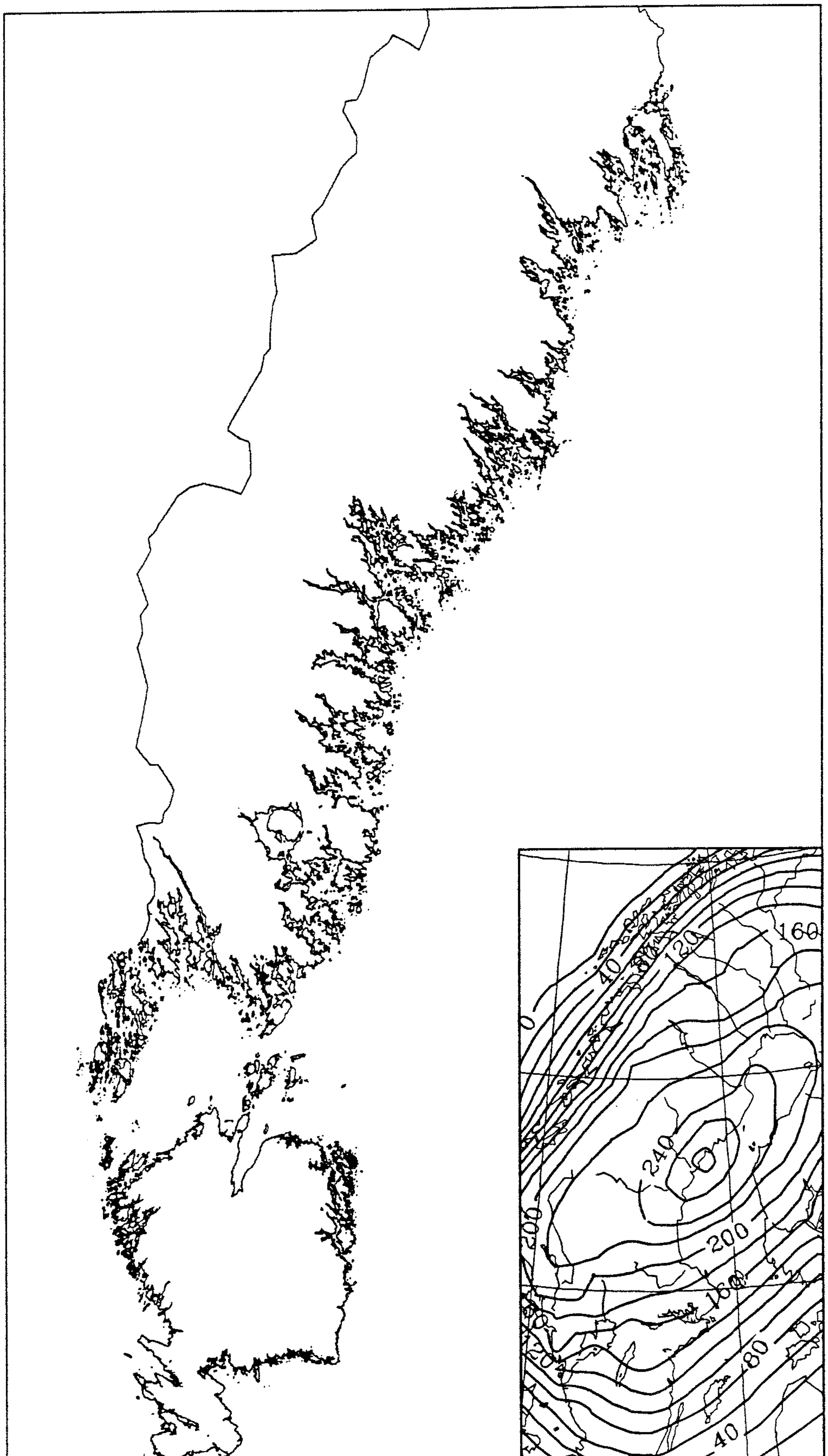
Information on the highest shoreline on the Swedish eastcoast is given in the descriptions to the geological maps (SGU ser. Ae, Ca) and for example in Asklund (1935), Halden (1934), G.Lundqvist (1961), and Hörnsten (1964), Agrell (1976), Ringberg (1971), Cato & Lidén (1973), Rudmark (1975).

5.4.1.2 The last Baltic Ice Lake shoreline, 10.300 BP

This shoreline was formed by the dammed Baltic Ice Lake during late Younger Dryas, around 10,300 (Björck 1979, 1981, Svensson 1989, 1991a), just before the final drainage (Fig. 5-4). This drainage took place when the ice margin retreated north of Mt. Billingen in south central Sweden, when the low-lying area north of Billingen was deglaciated. The drainage took place during the later part of Younger Dryas when the ice margin still was within the Middle Swedish end-moraine zone.

The final drainage of the Baltic Ice Lake has engaged many researchers. This drainage of the Baltic Ice Lake down to sea-level was first proposed by Munthe (1902) in the description to the geological map sheet Kalmar. Published work dealing with this drainage includes Munthe (1910b), Lundqvist (1921, 1931), S. Johansson (1926), Bergsten (1943), Berglund (1966), Strömberg (1974, 1977, 1984, and 1986), Björck (1979, 1981), Donner (1982), Björck & Digerfeldt (1984, 1986, and 1989), and Svensson (1989, 1991a).

Figure 5-3 The highest shoreline in Sweden. The map is based on shoreline data from the 'Swedish shoreline database' and a digital elevation model from the National Land Survey of Sweden (LMV) with 500 m data spacing. The inserted map shows isolines of highest shoreline observations, irregularities in southeastern Sweden are mainly an effect of highest shoreline formation during a dammed Baltic stage. In the province of Skåne, isolines as well as the map, shows the deglaciation shoreline, not the later transgressional stage (cf Lagerlund 1983, 1987).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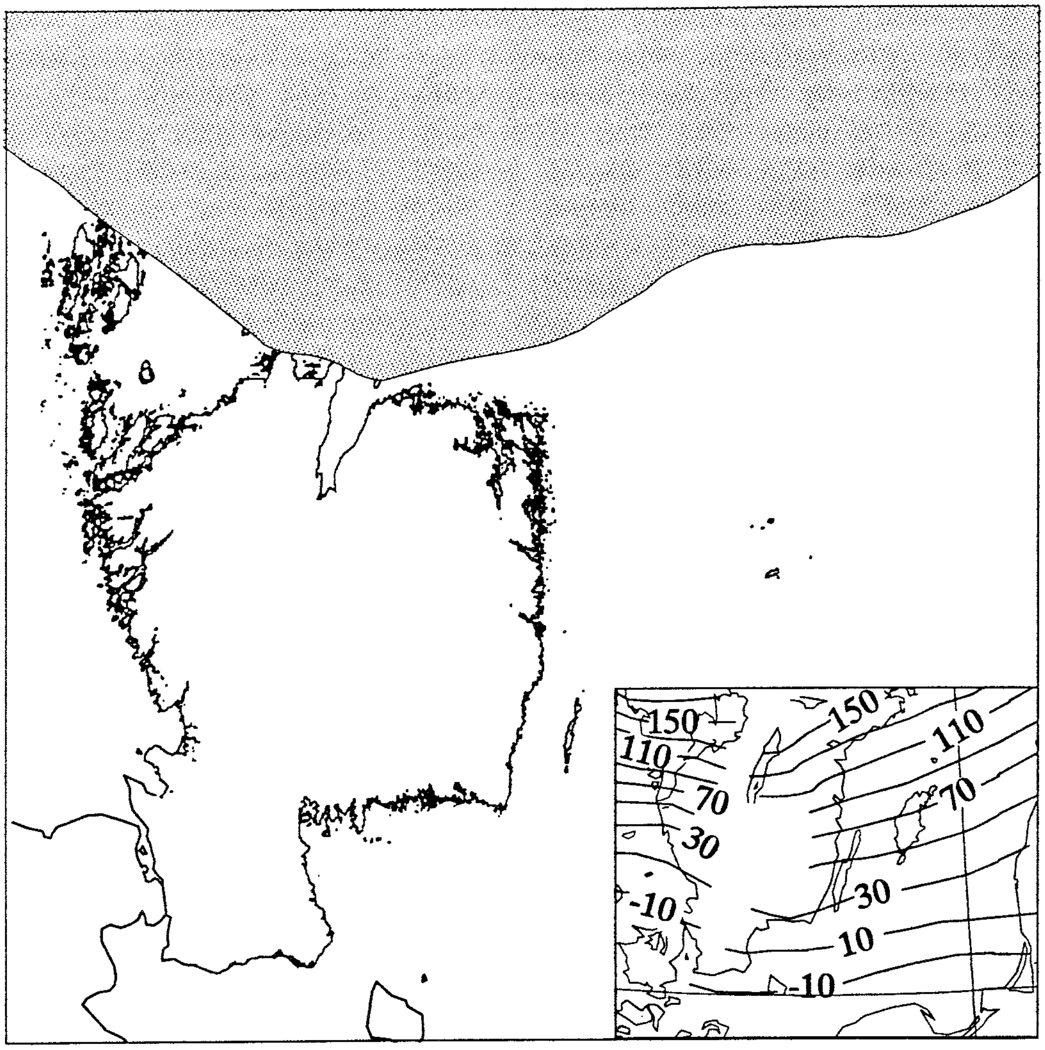


Figure 5-4 Palaeogeographic reconstruction of Southern Sweden 10.300 BP. At this time the Baltic were dammed c. 25 m above sea-level. The map is based on shoreline data from the 'Swedish shoreline database 'and a digital elevation model from LMV with 500 m data spacing.

The drainage amount is estimated to be 26 m in south-central Sweden (Bergsten 1943), 26–28 m in Finland (Donner 1969, 1978, 1982), at least 26 m in Blekinge (Björck 1979, 1981), and c. 25 m on Gotland and in the Oskarshamn area (Svensson 1989, 1991a).

This event was dated by Nilsson (1968) to 8213 varve years B.C. Björck (1979) dated the drainage to 10,200–10,300 B.P. (B.P. denotes ^{14}C years before present) in Blekinge, but Björck & Digerfeldt (1984) concluded that it was somewhat older, 10,400–10,500 B.P. Svensson (1989, 1991a) dated the drainage to c. 10,300 B.P. in the Oskarshamn area. Recently, Strömberg (1986) used distinct changes in the varved clay sequences to correlate this drainage to the revised Swedish varve chronology (Cato 1987), and thus dated the event to 10,690 varve years before present.

Based on assumed similar shore-displacement curves for the Baltic and the sea in the west, Svedhage (1985) proposed that the Baltic Ice Lake did not exist as a dammed stage, but instead was in level with the open sea. As a consequence no drainage would have occurred, and the fast regression at this time is explained by rapid tectonic uplift. Svensson (1989) concludes on the basis of investigations on Gotland and in the Oskarshamns area, that the drainage actually took place. He describes four different sets of observations speaking in favour of a drainage of c. 25 m in the later part of the Younger Dryas:

1. A very rapid regression, dated to c. 10,300 B.P., in an interval of c. 25 m.
2. The absence of beach-ridges in the interval, in contrast to their presence above and below.
3. A unit of coarse grained sediments associated with this regression indicates a sudden redeposition of sediments.
4. Similar gradients on the shorelines formed before and after the drainage indicates a fast regression.

The last shoreline formed by the Baltic Ice Lake is in the ice-marginal zone established by surfaces of ice marginal deltas, where a higher surface shows the pre-drainage shoreline and a younger c. 26 m lower surface shows the sea-level after drainage (Bergsten 1943). South of the ice marginal zone this shoreline is formed either as a distinct eroded shore notch, as a boulder field or as the lowest ridge in a sequence of beach-ridges.

This shoreline should be synchronously formed around the whole Baltic Ice Lake c. 10,300 B.P., and in general it is easy to identify both due to the absence of shorelines in an interval below, and the fact that the conditions for shoreline formation were especially favourable at this time (Svensson 1989). As a synchronously formed shoreline, with such a wide geographical extent, this shoreline is of significant interest for evaluation of land-uplift and various related parameters.

Regional synthesis: By the help of two new shore-displacement curves from Gotland and the Oskarshamn area (Svensson 1989), shore displacement curves from Blekinge (Björck 1979), Östergötland (Fromm 1976), and data from the Billingen area (Strömberg 1974, 1977, Björck & Digerfeldt 1984, 1989) the elevation of the 10,300 B.P. shoreline is rather well known. With this as a basis Svensson (1989) was able to correlate new shoreline levellings in the county of Kalmar and from Gotland with older studies from south Sweden to draw an isobase map for the last Baltic Ice Lake shoreline.

Based on a compilation of published data and new studies by Svensson, isolines of shorelines formed c. 10.300 BP were constructed (Fig. 5-4). At this time the Baltic was dammed c. 25 m above the ocean sea-level which formed shorelines on the Swedish West Coast. By combining these shorelinedata with a digital elevation model the palaeogeographic reconstruction of Fig. 5-4 were drawn.

It seems most probable that with further investigations the last Baltic Ice Lake shoreline could be mapped in more detail. Together with confirmation of its age in some more areas, as well as establishing some more shore displacement curves, such detailed knowledge of a synchronous shoreline would be of major importance for understanding the past and present land-uplift, as well as for documenting local disturbances in uplift and possible neotectonic activity.

5.4.1.3 The first Yoldia shoreline

This Baltic shoreline is defined as formed directly after the last Baltic Ice Lake drainage when the Baltic had reached sea-level. Assuming an instantaneous drainage it should have the same age as the previously described last Baltic Ice Lake shoreline, but situated at c. 25–26 m lower elevation. This shoreline could be more suitable to study in some areas, although it is often less well developed. As it is believed to be directly related to the last Baltic Ice Lake shoreline the discussion above is applicable. However, as it was formed by the sea, tides and meterological caused water-level changes might have had some influence.

Regional synthesis: This shore line is not as well documented as the last Baltic Ice Lake shoreline, partly because the fact that it was destroyed during the Ancylus transgression south of the 25–30 m a.s.l. Ancylus isobase (running approximately from south central Gotland to Mönsterås).

This shoreline has been documented on Gotland (Philip 1989, Svensson 1989), and in Östergötland and Västergötland (e.g. Nilsson 1937, Bergsten 1943, Fromm 1976).

5.4.1.4 The Ancylus shoreline

Shells of the fresh-water snail *Ancylus fluviatilis* in beach deposits on Gotland lead to the recognition of an early fresh-water stage of the Baltic (Munthe 1887). A damming above sea level seems probable during the first part of the Ancylus stage (e.g. Munthe 1902, Björck 1987); during its later parts, however, the Baltic presumably was in contact with the sea for a period prior to the transition to saline water.

On Öland and on Gotland this shoreline, mainly developed as a beach-ridge, is characterized not only by its extraordinary size but also its content of fresh-water snail shells, especially *Ancylus fluviatilis* and *Limnæa peregra* (Munthe 1910a; Königsson 1964, 1967a). The Ancylus beach-ridge shows the transgressive nature of the stage as the ridge often overlies terrestrial or limnic organic deposits (Munthe 1910b, 1940, Lundqvist 1928, 1965; Königsson 1968b; Persson 1978). Well preserved *Pinus* logs are often found buried beneath the beach-ridge or corresponding minerogenic deposits on southern Gotland (Lundqvist 1965; Persson 1978).

The first part of the Ancylus stage is characterized by a transgression that seems to take place without any halts and is so rapid that its rate is hard to determine in detail. In the Oskarshamn area the transgression amounts to c. 11 m and the rate could be almost 5 m/100 years (Svensson 1989).

The absolute dating of the Ancylus transgression has been an important subject in the study of the Baltic shore displacement. The age determinations of the Ancylus stage varied considerably. Some studies have concluded that the transgression had begun c. 9500–9700 B.P. and culminated well before 9000 B.P. (e.g. Berglund 1964; Glückert & Ristaniemi 1982; Svensson 1985, 1989, 1991a; and Ristaniemi & Glückert 1987); others concluded that the whole sequence was much younger, not beginning until after 9000 B.P. and culminating 8300–8500 B.P., (e.g. Lundqvist 1965; Königsson 1968a; Kessel & Punning 1974; Kessel & Raukas 1979).

Svensson (1989) discusses these differences in age and relates them partly to the type of sediment investigated and dated. Investigations yielding older ages for the transgression were often based on lake sediment ^{14}C dates (e.g. Berglund 1964; Glückert & Ristaniemi 1982; Svensson 1985, 1989; and Ristaniemi & Glückert 1987), while investigations yielding young ages often involved organic deposits (normally of terrestrial origin) buried beneath minerogenic shore deposits, (e.g. Königsson 1968a; Lundqvist 1965). Some of the difference in age could thus be ascribed to the reservoir effect which could be expected to give dates a few hundred years too old when ^{14}C dating lake sediments. As most of the dated peats are found below only a few m of gravel and sand, contamination from downgrowth of roots seems highly possible in many cases. By comparing dates from timber (mostly very well preserved) buried beneath Ancylus gravel on Gotland and dates from peat in similar situations, Svensson (1989) found a difference of more than 500 yrs — probably caused by contamination of the peat by younger material.

Another reason for differing age of the transgression between areas is the different rates of land-uplift in the areas compared. Observations from around the Baltic, however, do not confirm that this is the main reason for the age differences discussed, as no clear relationship between uplift rate and the age of the transgression maximum is found except possibly between eastern and western Finland (Ristaniemi & Glückert 1987). The influence of different rates of land uplift on the culmination age of the transgression must also be limited due to the fast transgression.

It can also be argued that variations arise because two different transgressive stages have been studied. This, however, seems less probable, because no single site with two clearly distinguished transgressions of this age appears to have been documented in modern investigations.

This shoreline is of great value for studies of land uplift as it is possible to study regionally, and was formed almost synchronously. The shoreline is well known and easily mapped on the islands of Öland and Gotland. In the county of Kalmar and Östergötland some rather easily performed studies of lake stratigraphies will give a better knowledge of this shoreline's elevation. A direct mapping of the shoreline will then be possible, which will give us a detailed knowledge of its elevation in the region. Further north, possibly even in southern Norrland, it could be established through shore displacement curves based on lake isolations. The reduced rate of relative uplift during the Ancylus culmination has probably resulted in a shoreline detectable by the help of the shore displacement curves.

Regional synthesis: This shoreline formed at the maximum of the Ancylus transgression is south of the 10–15 m isobase destroyed by the later Litorina transgression. The shoreline is generally very distinct on the islands of Öland and Gotland, where it was mapped in detail by Munthe (1910a) on Gotland, and by Munthe (1902), Munthe & Hedström (1904), and Lundqvist (1928) on Öland. In the county of Kalmar (except Öland) observations are from scattered localities, e.g. Munthe (1902, 1940), Munthe & Hedström (1904), Thomasson (1926), Nilsson (1953, 1968), Königsson (1967b, 1968ab), Mikaelsson (1978), Rudmark (1988), and Svensson (1989).

Assarsson (1927) presented information, based on lake sediment stratigraphies and shorelines, about the highest limits of the Ancylus transgression (c. 70 m asl) in eastern Östergötland. The Linköping area seems to display the northernmost evidence of a transgressive Ancylus Lake at c. 75 m a.s.l. (Munthe 1940, Fromm 1976). In northern Östergötland the Ancylus transgression could not exceed the land uplift, and Persson (1979) shows that this stage is here represented by a stillstand or a decreased rate of regression. Correspondingly, no transgressive shoreline is found here and further northwards. However, the Ancylus shoreline (at least up to the c. 110 m isobase) still often forms a significant beach-ridge in southernmost Svealand (Munthe 1940).

5.4.1.5 The Postglacial transgression maximum (Litorina) shoreline

Because the eustatic rise in the middle of the Holocene was more rapid than the uplift rate in large parts of Sweden this transgression continued around the Swedish coasts, culminating in the Postglacial transgression maximum on the west coast and the Litorina transgression on the Swedish east coast. The culmination of this transgression was complex, with a series of several minor culminations in which the highest (in absolute terms) was not the first. Thus, depending on the rate of land uplift, it varies which of these culminations caused the transgression maximum between different regions in Sweden. In areas of high uplift rate the first culmination reached the highest, whereas in areas with low land uplift a later culmination reached highest. This shoreline is thus metachronous; in Blekinge it is dated to c. 5000-5500 B.P. (Berglund 1964, Liljegren 1982), in western Skåne c. 4800 B.P. (Digerfeldt 1972, 1975), in the Göteborg area somewhere between 7700 and 7200 B.P. (G. Persson 1973, Mörner 1976b, Påsse 1983), on Gotland c. 6700 B.P. (Persson 1978), in the Nyköping area c. 6200 B.P. (G. Persson 1973), and in the Stockholm area c. 7000 B.P. (Miller & Robertsson 1981, Sandgren & Risberg 1990).

As this shoreline can be identified both on the west and east coasts of Sweden, and mostly is of significant size, it is of considerable interest for reconstructing the former land uplift. However, its metachronous character has to be considered in the interpretations and detailed biostratigraphical investigations and datings are needed in more areas to regionally establish the age of the shoreline.

Regional synthesis: Based on a compilation of published data, isolines of shoreline observations dated to around 6500 BP were constructed (Fig. 5-5). At this time the postglacial transgression (Tapes/Litorina) reach one of its earlier culminations. By combining these shorelinedata with a digital ele-

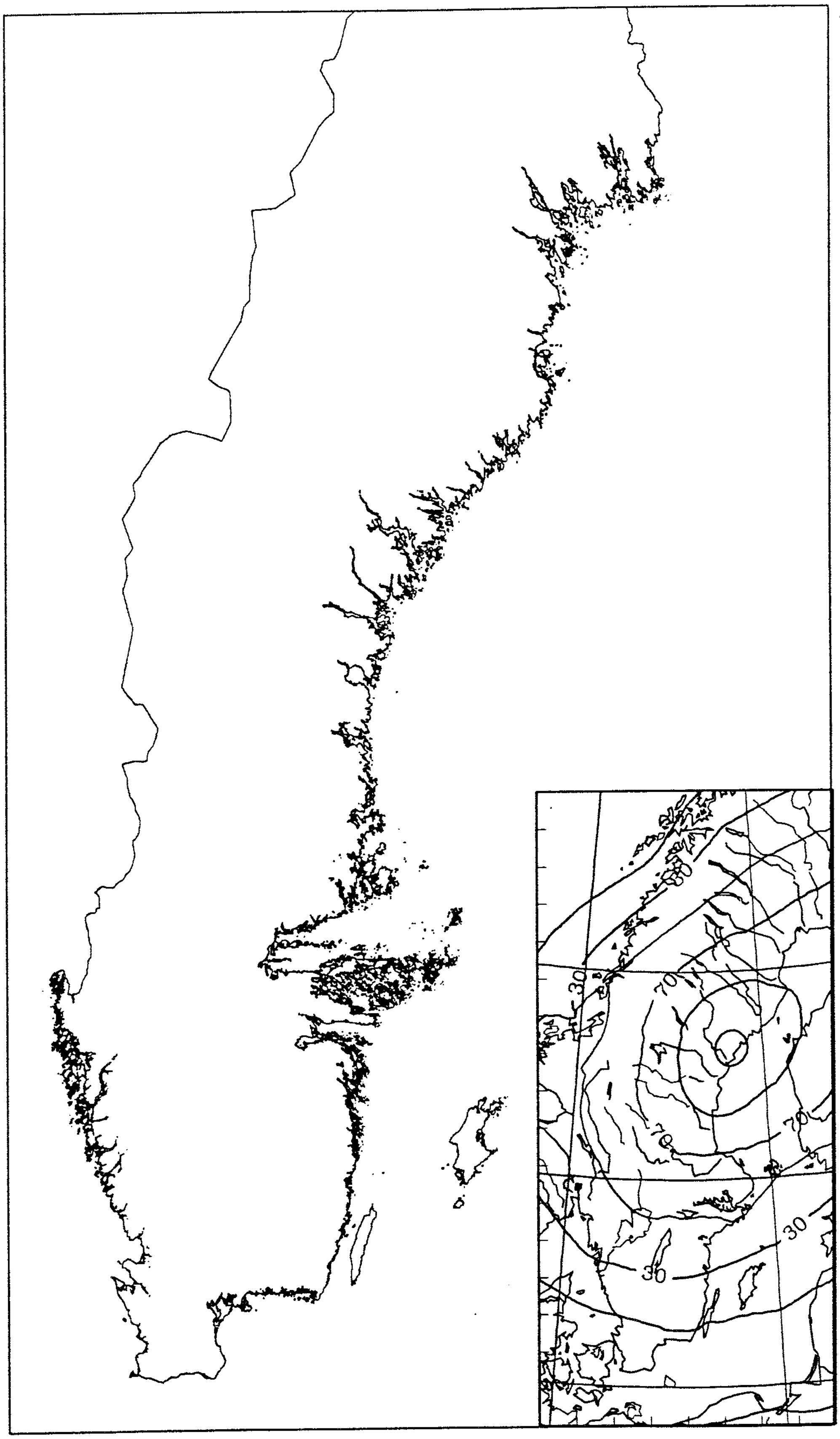


Figure 5-5 Palaeogeographic reconstruction of Southern Sweden c. 6500 BP. At this time the first postglacial transgression (Tapes/Litorina) culminates. The map is based on shoreline data from the 'Swedish shoreline database 'and a digital elevation model from LMV with 500 m data spacing.

vation model the palaeogeographic reconstruction of Fig. 5-5 were drawn. The shoreline formed by the highest postglacial transgression has been frequently studied and is thus well represented in the literature, especially in the descriptions to the geological maps, and below only a few authors are cited.

The shoreline of the postglacial transgression maximum along the whole Swedish west coast is mostly very beautifully developed as a beach ridge or as a shore cliff or shore cut (Mörner 1969).

In northern Bohuslän the Postglacial transgression could not exceed the land uplift — the northernmost signs of actual transgressions (reaching close to 45 m asl) are found in central Bohuslän somewhat south of Uddevalla (G. Persson 1973). Recently, Miller & Robertson (1988) briefly report some indications of the transgression c. 20 km further to the north. At Göteborg the transgression culminates at c. 23–25 m a.s.l. (G. Persson 1973, Påsse 1983). At Morup close to Falkenberg the transgression reached 13.5 m asl, at Båstad c. 8 m asl, and close to Höganäs c. 6 m a.s.l. (Mörner 1969). Further south at Barsebäck, c. 20 km north of Malmö, Digerfeldt (1975) found the maximum transgression to reach 4 m asl.

Along the south and east coasts of Götaland this shoreline is also often well developed, frequently of a significant size. In Blekinge it is situated at c. 7–8 m a.s.l. (Berglund 1964, Liljegren 1982). At Borgholm, on Öland, the transgression reached 17 m a.s.l. (Munthe & Hedström 1904), northernmost Gotland c. 27 m a.s.l. (Munthe 1910b), eastern Östergötland 37–40 m a.s.l. (Assarsson 1927), and Stockholm c. 50–58 m a.s.l. (Miller & Robertsson 1981, Sandgen & Risberg 1990). Further to the north the transgressive character of the shoreline disappears. In Norrland this shoreline is not easily recognized in the field and stratigraphical studies, mainly diatom analyzes are needed to identify it. In Northern Hälsingland it is found at c. 115 m a.s.l. (G.Lundqvist 1963), and close to Örnsköldsvik at c. 123 m a.s.l. (Miller et al 1979).

5.4.2 Stratigraphically determined shorelines

5.4.2.1 Methods

To establish the shore displacement for an area it is necessary to date deposits directly related to the shore displacement. This can be done in several ways:

- By dating organic deposits within or just below beach deposits.
- By dating peat formed by plants which have grown close to sea-level. To estimate the former sea-level the peat should rest on firm ground so that compaction below the peat can be avoided.
- By dating delta surfaces that were built up close to sea-level.
- By detailed stratigraphical studies of the sediments in a single basin. The sediment's composition and content of macro- and micro-fossils is analyzed and the former depth of the shallow sea is deduced from the ecology of the observed species.
- By dating lake sediment deposited either when the lake was separated from the sea (because of uplift) or when the sea inumdates into the lake

basin. The elevation of sea at the time of isolation/inundation is obtained from levelling of the lake threshold.

The dating methods used in the above methods are generally ^{14}C dating, either direct dating of the sediment or indirectly by pollen stratigraphical correlations to a thoroughly ^{14}C dated lake sediment stratigraphy. However, annually laminated sediments, e.g. lake sediments, varved clays, or varved delta sediments, have occasionally been used.

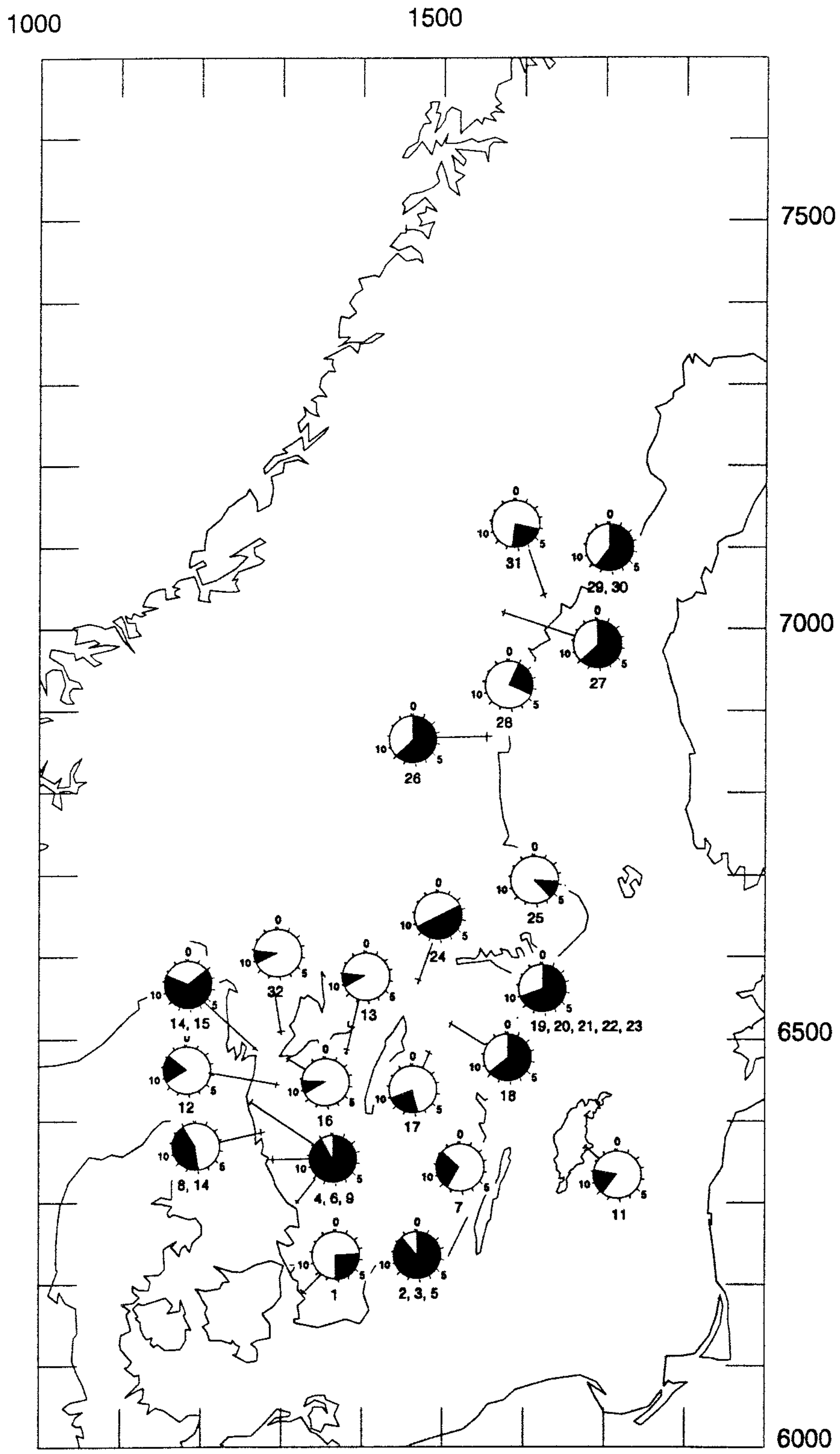
In one case (Lidén 1938) delta surfaces dated by varve chronology have been used to establish a shore displacement curve. The usual method to establish the shore displacement in Sweden is to date lake isolations and thus this method will be described more in detail.

The method requires the use of a number of lakes situated at different elevations so that the ages of lake isolations are spaced in time. The lakes need to be situated in a rather small area so that the differential land uplift between the lakes is not too large. This differential uplift is normally compensated for, but as the pattern of uplift through time is not known in detail, this compensation is more or less uncertain and must be kept rather small. The levelled threshold elevations of the lakes are used as a measure of the former sea-level. It is therefore important that the threshold has not been severely altered by later erosion or deposition. Thresholds on easily eroded soils are thus more uncertain than bedrock thresholds. The sediments from a suitable lake (or overgrown lake) are sampled by some type of corer, and the sediment description usually gives a rough indication of where the isolation horizon is to be found. The isolation (or inundation) horizon is established in detail by determining the sediment composition and studying the microfossil content — an isolation from the saline sea is indicated by the change from saline to fresh water diatom floras. The time of isolation can be obtained by a ^{14}C date of the sediment at the isolation level. However, this often yields erroneous ages, especially in Late Weichselian and early Holocene sediments. To check the obtained dates a good pollen stratigraphy is needed for the time around the isolation. In many situations (especially for Late Weichselian and early Holocene isolations) it is even better to carefully ^{14}C date the regional pollen stratigraphy in a lake with suitable sediments, and then apply ages to the individual lake isolations by detailed pollen stratigraphical correlations.

5.4.2.2 Swedish shore displacement curves

A number of modern shore displacement curves have been established in Sweden by the help of ^{14}C dates or dates by annually laminated sediments, and Figure 5-6 shows the location and time interval covered by these investigations. Figure 5-7 permits comparisons between the different shore displacement curves. To space the different curves in the figure, each one was started at the present land uplift of the site. This comparison emphasizes the regional differences, some of which can be mentioned:

- The dammed Baltic stages e.g. the Baltic Ice Lake and the Ancylus lake, clearly affects the shore displacement as seen in curve 2, 3, 5, 7, and 11.
- The rates of present land uplift are correlated to former rates of shore displacement.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2, 3, 5

- The minor fluctuations in shore displacement, mainly believed to be of eustatic origin, shows large discrepancies in amplitude between different investigations. The expected behaviour — that the amplitude should decrease in areas with higher uplift rates — is not clear from all investigations. In fact several curves show rather the opposite relation.

5.4.2.3 The course of shore displacement

During deglaciation of the southernmost part of Sweden, Skåne, at c. 13,000–14,000 B.P. relatively little is known about shore displacement and the configuration of sea and dammed glacial lakes. It seems clear that the west coast of Skåne have been under marine influence for at least part of this period as polar cod (*Boreogadus saida*, formerly *Gadus polaris*) were found in clays probably deposited at this time (de Geer 1884). In areas where the shore displacement is established for this period (Fig. 5.7) a fast regression takes place.

During Allerød (c. 11,000–12,000 B.P.) the regression in southern parts of the Swedish West coast continues to be rather fast, whereas in western south-central Sweden the regression slows down until close to 11,000 B.P. when it again becomes faster. On the east coast the Baltic Ice Lake prob-

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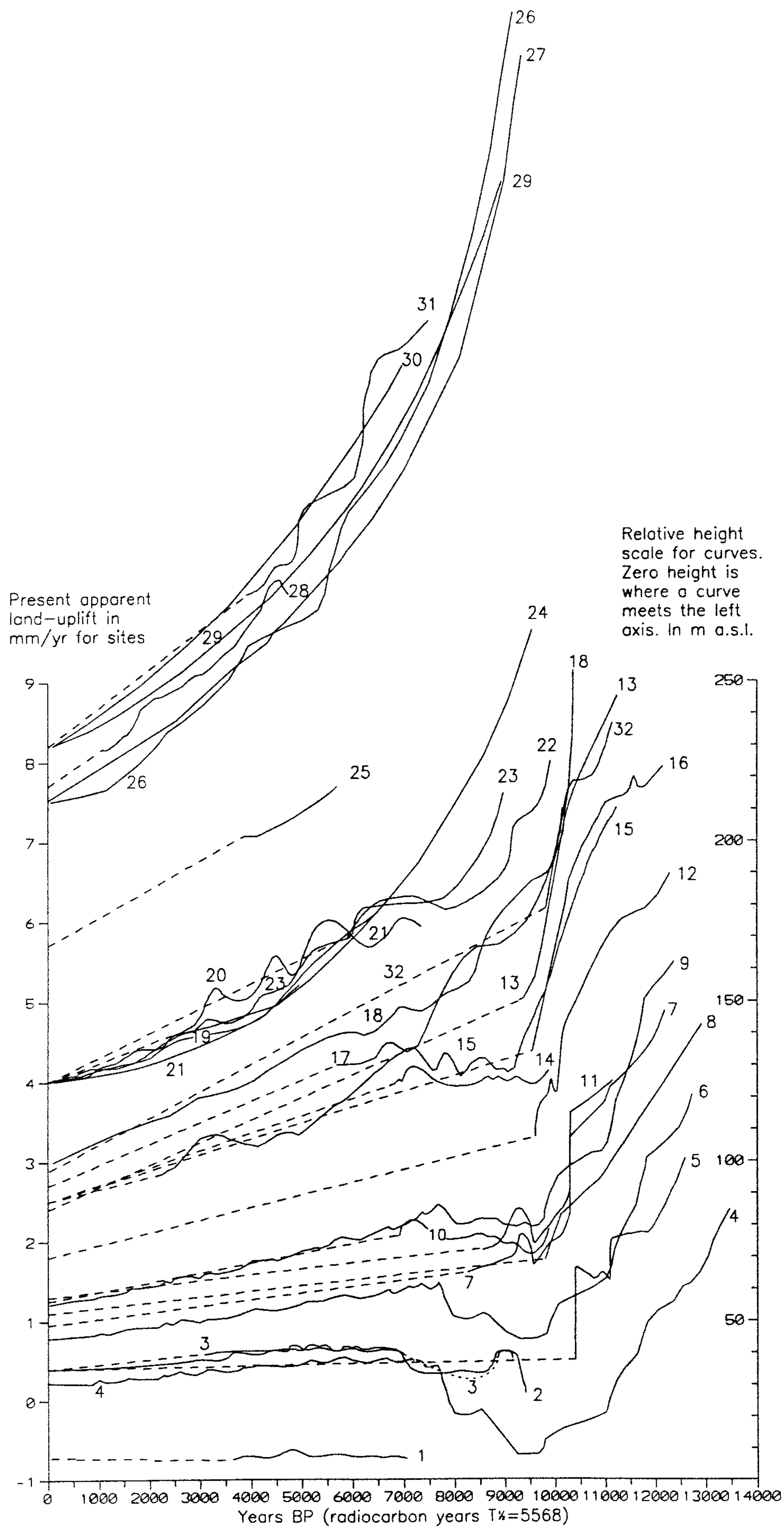
Figure 5-6 Areas with modern shore displacement curves. The watch at each site (or region) show the time interval covered by shore displacement curves, each 'hour' corresponds to 1000 years. Numbers in square brackets refers to the shore displacement curve(s) in Fig. 5.7. Data from the 'Swedish Shoreline Database'.

Central Bohuslän [14, 15]. Persson (1973), Miller & Robertsson (1988). *Northern Halland and Göteborg area* [8, 14], Persson (1973), Påsse (1983,1987). *Kroppefjäll* [32], Björck & Digerfeldt (1991). *Hunneberg* [16], Björck & Digerfeldt (1982). *Risveden* [12], Svedhage (1985). *Billingen* [13], Björck & Digerfeldt (1986). *Viskan valley* [6], Mörner (1969, 1980b), these investigations covers besides from the main area, Viskan Valley, the Swedish West Coast from Göteborg southwards to Skåne. *Barsebäck* [1], Digerfeldt (1975). *Central and eastern Blekinge* [2, 3, 5], Berglund (1964), Björck (1979, 1981), Björck & Möller (1987), Liljegren (1982). *Oskarshamn area* [7], Svensson (1989, 1991a). *Central Gotland* [11], Svensson (1989, 1991a). *Linköping* [17], Fromm (1976). *Rejmyra* [18], Persson (1979). *Stockholm region* [19, 20, 21, 22, 23], Möller & Stålhös (1964, 1969), Åse (1970), Digerfeldt (1981), Digerfeldt et al (1980), Miller & Robertsson (1981), Åse & Bergström (1982), Miller (1982), Sandgren & Risberg (1990). *Örebro* [24], Magnusson (1970). *Uppland* [25], Robertsson & Persson (1989). *Northern Hälsingland* [26], G. Lundqvist (1962, 1963). *Västernorrland* [28], J. Lundqvist (1987). *Sollefteå* [27], Lidén (1938). *Anundsjö* [31], Miller & Robertsson (1979). *Västerbotten* [29, 30], Broadbent (1979), Renberg & Segerström (1981).


Present apparent land-uplift in mm/yr for sites
Relative height scale for curves. Zero height is where a curve meets the left axis. In m a.s.l.
Years BP (radiocarbon years T½=5568)
0 1000 2000 3000 4000 5000 6000 7000 8000 9000 10000 11000 12000 13000 14000
-1 0 1 2 3 4 5 6 7 8 9
50 100 150 200 250
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25 26 27 28 29 30 31 32

ably becomes successively dammed to a higher level, thus decreasing the rate of the regression. At c. 11,200 B.P. a fast drop of the Baltic's level is recorded in a study from Blekinge (Björck 1979, 1981) interpreted as a drainage down to sea-level taking place when the ice margin withdrew north of Mt. Billingen in south central Sweden.

During the Younger Dryas (c. 10,000–11,000 B.P.), which was a very cold period, the ice advanced and closed the passage at Billingen (Björck & Digerfeldt 1984, 1986) and once again a damming of the Baltic took place. The advancing ice at this time also affected the shore displacement west of Billingen. Björck & Digerfeldt (1991) have observed distinctly reduced rates of land uplift at this time at Kroppefjäll, Dalsland, caused by the near-by ice-loading and possibly glacio-geoidal effects. During late Younger Dryas, at a time when the ice was retreating again, high rates of regression are observed at the west coast. The Baltic Ice Lake once again drained (by c. 25 m) down to sea level when the ice retreated north of Billingen (Munthe 1910b, Björck & Digerfeldt 1984, 1986) after which a fast regression

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Figure 5-7 Swedish shore-displacement curves. Ages in ^{14}C years, (T 1/2 5568) except for curve 27 dated in sideral years by varve chronology and curve 29 dated by varve counting. The left vertical axis shows the apparent land uplift rates for the areas of the shore displacement curves (uplift values from Ussisoo 1977). The right vertical scale shows the relative elevation scale in m a.s.l. for each curve, each curve has an individual zero point where it intersect with the left axis. The hatched continuation of the curves is for finding the zero m a.s.l. point and is not derived from the original author. Data from the 'Swedish Shoreline Database'.

1. Barsebäck, Skåne (Digerfeldt 1972), 2. Southeastern Blekinge (Berglund 1964). 3. Spjälkö, Blekinge (Liljegren 1982). 4. Southern Halland (-50 km) (Mörner 1980b). 5. Central Blekinge (Björck 1979, 1981, Björck & Möller 1987). 6. Viskan valley, Halland (0 km) (Mörner 1980b). 7. Oskarshamn, Småland (Svensson 1989, 1991a), 8. Sandsjöbacka, Halland (Påsse 1987), 9. Göteborg (+50 km) (Mörner (1980b), 10. Göteborg (Persson 1973), Göteborg, 11. Central Gotland (Svensson 1989,1991a), 12. Risveden, Västergötland (Svedhage (1985), 13, Billingen Västergötland (Björck & Digerfeldt (1986), 14. Kolbengtserödsjön, Bohuslän (Persson 1973), 15. Central Bohuslän (Miller & Robertsson 1988), 16. Hunneberg, Västergötland (Björck & Digerfeldt 1982), 17, Linköping (Fromm 1976), 18. Rejmyra, Östergötland (Persson 1979), 19. Stockholm (Åse 1970), 20. Stockholm (Miller & Robertsson 1981), 21. Stockholm (Möller & Stålhös 1964), 22. Ådran, Stockholm (Sandgren & Risberg 1990), 23. Stockholm (Åse & Bergström 1982), 24. Örebro (Magnusson 1970), 25. Northern Uppland (Robertsson & Persson 1989), 26. Northern Hälsingland (G. Lundqvist 1962, 1963), 27. Ångermanälven, Ångermanland (Lidén 1938), age in sidereal years before present, redated according to Cato (1987), 28. Västernorrland (J. Lundqvist 1987), 29. Southern Västerbotten (Renberg & Segerström 1981), age in sidereal years before present, 30. Västerbotteno (Broadbent 1979), 31. Anundsjö, Ångermanland (Miller & Robertsson 1979). 32. Kroppefjäll, Björck & Digerfeldt (1991)

began.

In Early Holocene (c. 8000–10,000 B.P.) the rates of glacio-isostatic land uplift where still higher than the eustatic rise of the sea during the first part of this period and around southernmost Sweden the sea-level fell far below present day sea-level. On the West Coast this regression ceased at c. 9500–9300 B.P. and the sea soon began to rise — the onset of the post glacial transgression(s). In the Baltic the regression ceased even somewhat earlier at c. 9700–9600 B.P. when the outlets straits of the Baltic in central Sweden began to close due to the on-going land-uplift (Munthe 1910b, Björck 1987). This caused the Ancylus transgression, which culminated somewhat before 9000 B.P. (Svensson 1989, 1991a) when the outlet was redirected to the Danish straits. After this change in outlet a rather fast regression took place in the Baltic, and the Baltic quite soon became in level with the sea (Björck 1987, Eronen et al. 1990).

Mid and late Holocene (c. 8000 B.P. – present time), this period is first characterised by the rising sea, the Postglacial transgression, or the Litorina transgression in the Baltic. The culmination of this transgression is metachronous, and youngest in the south due to the low rates of glacio isostatic land-uplift there. This transgression is divided into several oscillations to which some agreement possibly exist in time but which amplitude in different parts of the country seems hard to explain. After the culmination of this transgression a regression down to present day sea-level took place during which some oscillations is documented. In the Baltic the lowered sea-level reduced the area of the Danish straits. As a consequence the Baltic become less saline.

5.5 NEOTECTONICS

One of the most clear evidence of movement in the earth's crust is well developed tectonic structures. However, such distinct structures of proved Quaternary age, thus possibly of a glacioisostatic origin, from Sweden is not commonly described in the literature. This interesting and debated field of study will only be briefly mentioned below, but is further treated in chapter 7 and 8.

The dating of these structures is important and their age is mainly derived from the deposit or the structure the dislocation cuts, some times the date is further improved by a non-dislocated deposit found above the fault. Possible evidence for a Late Quaternary age could be:

- Faulting of a glacially striated bedrock surface,
- Formation of tectonic structures which not would be able to survive a glaciation:
- Dislocation of linear Quaternary features such as shorelines, eskers, or dislocation of sediment deposits, possibly overlain by younger non-dislocated sediments.

The type of structure indicating tectonics could be:

- Faults, either as large dislocations along a fault-line, a good example of this is the Pärve Fault in Swedish Lapland (Lundqvist & Lagerbäck

1976, Lagerbäck 1979), or of a smaller extent such as described from central and southern Sweden by Lagerlund (1977), Björkman & Träd-gårdh (1982), Mörner (1978, 1985). Further can vertical displacement within Quaternary sediments indicate neotectonic activity, of such gives Flodén & Winterhalter (1981) an example of from the central Baltic Sea.

- Boulder caves formed by large accumulations of huge boulders which are broken up from the bedrock. The boulders are often glacially abraded and striated which indicates a post-glacial age for the formation of the cave. Such caves are described from central and northern Sweden by R. Sjöberg (1987), Agrell (1986), and Ekman (1988). These authors ascribes the formation of the caves to earthquakes occurring at the end of deglaciation. Ekman (1988) further discusses their formation and suggest that three very large boulder caves at the coast of Medelpad and northern Hälsingland were formed during major earthquakes produced by the effect of the Gaussian curvature. The effect of the Gaussian curvature was according to Ekman (1988) concentrated to the area of maximum uplift rate close to deglaciation — the coast of Medelpad and northern Hälsingland, and was most intence in early Post glacial time. However, the theory does not explain the occurrence of the smaller boulder caves in other regions of Sweden.

Regional studies of shore lines can reveal important information on larger irregularites in the uplift pattern. An example of such a disturbance with relatively higher land-uplift, observed in an area extending from southeast Finland to western Estonia, northern Lithuania and across the Baltic to Gotland and Öland, is given by Svensson (1991b). The explanation of this feature is not clear, possibly could it be related to large scale tectonics, deep crustal structures or higher ice load during the last glacial period.

6 GEOPHYSICAL MODELLING — ISOSTASY/EUSTASY — FUTURE PROGNOSES

The long tradition of research in Fennoscandian relative sea level history has led to an almost unique amount of data/unit area. In spite of the relatively good (but obviously not sufficient) information different scientists reach different conclusions about the geophysical characteristics and thus also about the behaviour of the lithosphere and the upper and lower mantle since the last deglaciation. This can partly be explained by the fact that different approaches and prerequisites have been used by scientists from different disciplines. It is hardly surprising that modellers (e.g. Clark, 1980; Peltier 1985, 1987), geophysicists (e.g. Anderson, 1984; Fjeldskaar, 1989) and geologists (e.g. Mörner 1981) come to smaller (or larger) differences concerning the properties of the lithosphere and mantle. Being a geologist without geophysical skills it is almost impossible to evaluate and review the different theories. It is, however, obvious that foreign (outer Scandinavian) scientists often do not have access to or are not aware of the up-to-date and best geological data from the Fennoscandian region. Therefore we will mainly focus on the results from Scandinavian researchers.

According to Mörner's (1979) calculations, based mainly on late-, and postglacial uplift data, the absolute Fennoscandian uplift in the centre of the uplift cone is slightly more than 800 m (Figure 6-1). The model of Fjeldskaar (unfortunately still being a secret report which prohibits citing actual values or figures) gives slightly lower maximum uplift values, based mainly on a thick ice model and best-fitting earth rheology parameters. In e.g. the Stockholm region Mörner's (1979) uplift model gives values of c. 500 m absolute uplift while Fjeldskaar's computer based model shows significantly larger values of uplift. The latter also holds true for the whole of Sweden, except for the uplift cone region.

Estimations of remaining land uplift seem more uncertain. This is mainly due to uncertainties about the values on remaining rise of the geoid and to what extent these values reflect postglacial mass deficiency and/or mass deficiency of other origin. With Ekman's (1977) formula and Bjerhammar's (1980) and Kakkuri's (1985) value of c. 10 m for remaining geoid rise the remaining land uplift becomes c. 150 m. Most people working with these problems do, however, seem to agree that a correction for the crustal structure according to Anderson (1984) has to be made. This correction reduces the remaing geoid rise to c. 2 m which in turn results in a much lower remaining land uplift value of c. 30 m (Ekman 1989). Together with the previously mentioned uplift data this suggests that the absolute uplift in the uplift centre since the last glaciation is in the order of 800–850 m. This gives an intimation of the expected down-warping in an uplift centre during future full glacial conditions.

If we focus our predictions to the Stockholm region with three different glacial scenarios we might expect the events described below. The predictions are complicated since it is unlikely that the area was characterized by isostatic equilibrum under any of the phases described below. Most other parameters are also uncertain so the following section should only be regarded as a rough attempt to illustrate how we can use our knowledge to describe possible future situations.

- **A glaciation restricted to Northern Sweden and the mountain area**: As the glaciation increases the area will possibly be uplifted to a small extent, both absolutely and relatively, as an effect of a southwards moving small forebulge and falling sea levels. These two uplifting effects will thus be greater than the ice-sheet gravity effect (causing a higher gravimetric geoid close to the ice-margin) since the postulated ice-margin is situated too far from the area to have a significant effect (e.g. Fjeldskaar and Kanestrøm, 1980). During the deglaciation of such a minor glaciation the area will submerge but probably without a marine transgression.
- **A glaciation restricted to Northern and Central Sweden down to approximately Stockholm's latitude**: During the build-up of such a glaciation the area will first experience the uplift effects of the southwards migrating forebulge and the falling sea level (in order to keep in pace with the falling sea level the Baltic waters will have to erode and deepen the Danish straits at a critical point). As the ice-margin gets closer the down-warping will begin, local sea level will rise as a consequence of the ice-sheet gravity effect, and the global sea level will keep falling. The added effect will probably not result in any transgression of the Stockholm region. With a final ice-margin at Stockholm's latitude and an ice-sheet profile according to Hughes (1981) we could expect an ice thickness at Stockholm of c. 800–1000 m. Let us regard this as a temporary advance of the ice-sheet and use the same prerequisites as e.g. Quinlan and Beaumont (1981) used for their calculations: a 1 km thick advance into a 5° x 5° (550 km x 400 km) area during 3000 years. The down-warping effect would under such circumstances be c. 50 m (Figure 6-2) and becoming greater the longer the advance lasts (a e.g. rather unlikely 10,000 year long advance will result in c. 90 m down-warping). If we use Imbrie and Imbrie's (1980) model output curve, sea level during such a glaciation would possibly fall 70–100 m. Although we might expect a rapid deglaciation (according to the marine records the most significant "terminations" seem to be surprisingly rapid) the area will probably, under the above outlined circumstances, not be flooded by the Baltic. This will be followed by an isostatic uplift with an uplift maximum quite close to deglaciation. Thus it is not likely that the area will be transgressed by Baltic waters during the described scenario above.
- **A full glaciation, roughly corresponding to isotope stage 2**: During the early build-up stage the area will be uplifted (see above), but this will soon be followed by a gradual down-warping as the ice-margin gets closer and the area is finally covered by the growing ice-sheet. The down-warping will continue until slightly after the maximum ice thick-

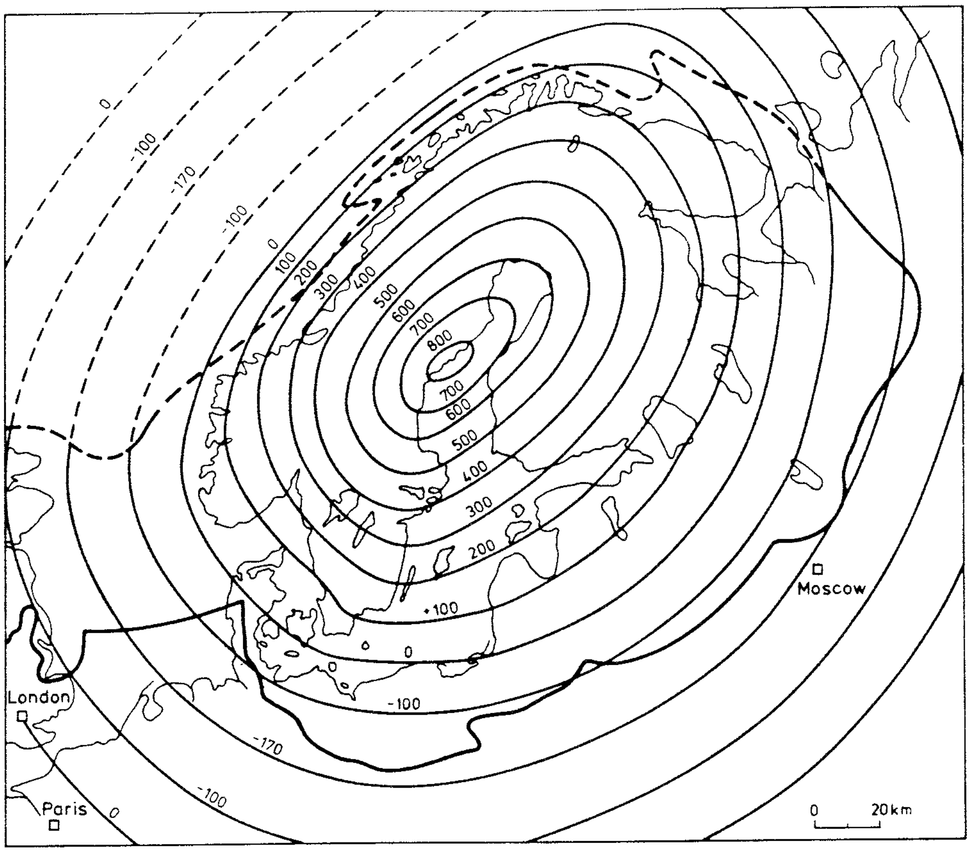
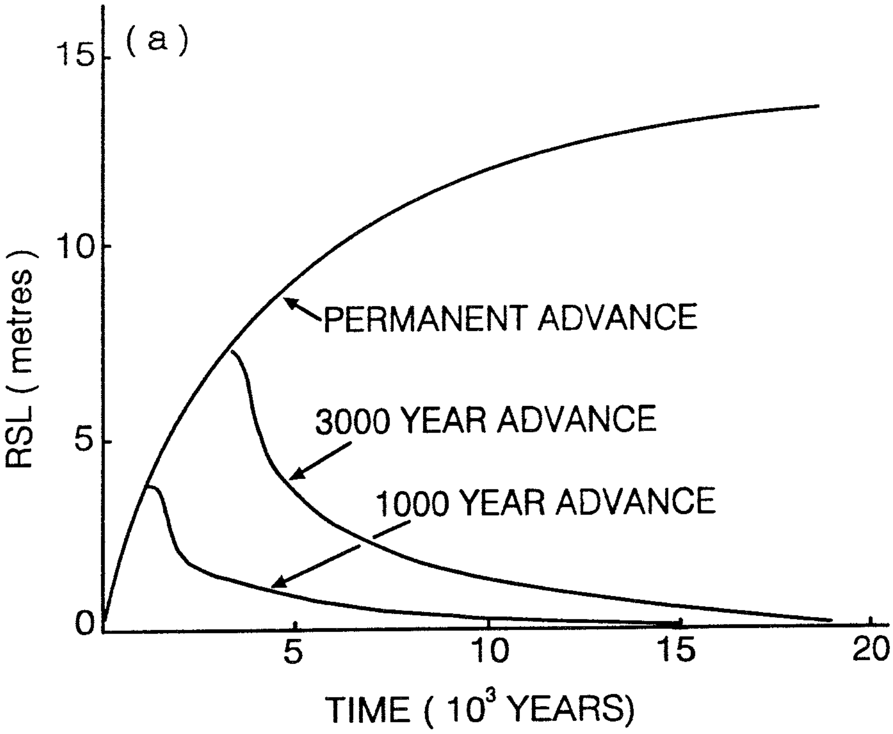


Figure 6-1 Contours of the total absolute uplift and subsidence in relation to the last glaciation of the Fennoscandian Shield. Heavy line gives the extension of the Late Weichselian ice cap. The North Sea forebulge trough is assumed to circumscribe the uplift cone uniformly. From Mörner (1979). Reprinted with permission from N-A Mörner from 'The Fennoscandian uplift and Late Cenozoic geodynamics: geological evidence', by N-A Mörner 1979. GeoJournal 3.3, pp. 287-318.

ness has been attained as an effect of the time-lag between ice load and isostatic response. The area will be covered by ice, with a maximum thickness of 2000–3000 m (Wu & Peltier, 1983; Denton & Hughes, 1981, respectively) during a time period of 10–20,000 years. The maximum down-warping of the area will possibly be in the order of 500–700 m (see above). As the final deglaciation sets in the ice-sheet will soon start to become thinner, and, slightly delayed, the uplift will commence. It is not completely unrealistic to postulate that the deglaciation pattern and relative sea level changes will roughly follow the same development known from our studies on the Late Weichselian-Holocene. Many Finnish studies show that the Eemian development was in many ways, broadly speaking, similar to the last lateglacial-postglacial development (Eronen, 1989). With this postulation the Stockholm region will undergo quite significant changes. In spite of high uplift rates at the time of deglaciation (in the order of 10 m/100 years) the area will be inundated by glacial lake waters at and some time after deglaciation. It is, however, not likely that the Baltic Sea will experience the same post-glacial development as it did during the Holocene, since this complex development seems to have been very dependent on the altitudes and uplift rates of the shifting thresholds of the Baltic basin (Björck, 1989). Since we know that glaciations are capable of altering the landscape significantly many of the recent preconditions will most likely have changed at the termination of the next glacial stage and at the beginning of the following interglacial. It does, on the other hand, seem likely that the Stockholm region, during a future interglacial, will be inundated by fresh and marine (brackish) water. In this context it is, however, important to stress that the Finnish-Russian-Estonian studies on Eemian sea level history (e.g. Forsström et al., 1987, 1988; Liivrand, 1987) suggest striking differences in the configuration (e.g. shoreline tilts) of Eemian and Holocene shorelines. Whether these differences have been caused by local/regional anomalies in the uplift pattern (e.g. tectonic movements) or by differences in the "threshold history" of the Baltic Sea is not clear.



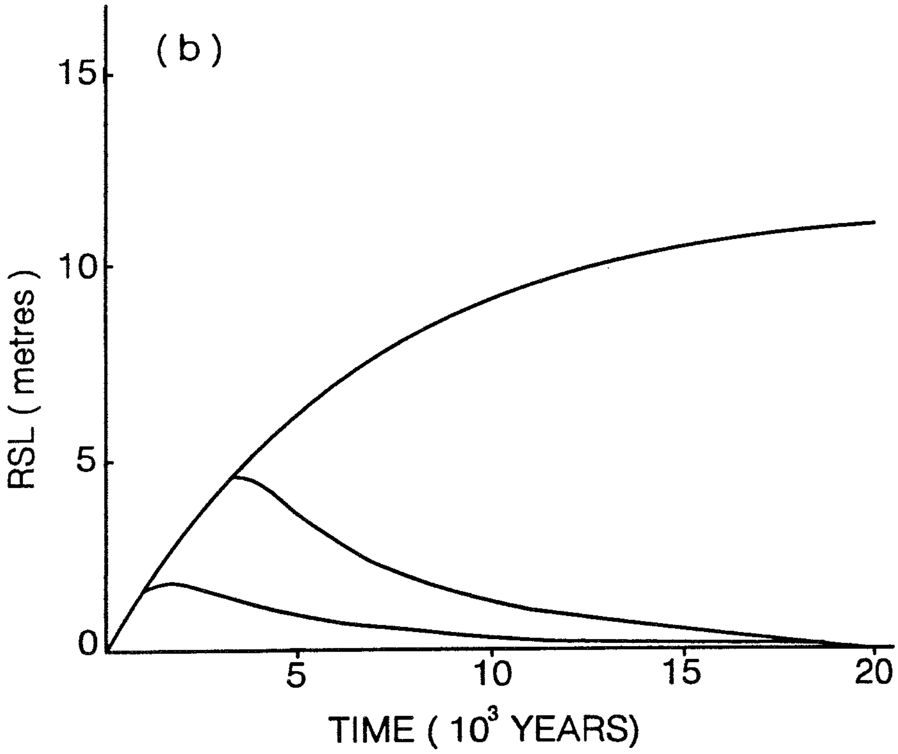


Figure 6-2 The relative sea level history resulting from a 1 km thick ice advance into (a) a 1° x 1° area and (b) a 5° x 5° area, both centred on 45° N latitude. The ice remains permanently, 1000 years, or 3000 years. the calculation is for a point at the edge of the ice advance. From Quinlan and Beaumont (1981). Reprinted with permission from National Research Council of Canada. Canadian Journal of Earth Sciences 18, pp. 1146-1163. A comparision of observed and theoretical postglacial relative sea level in Atlantic Canada, by G Quinlan and C Beaumont 1981.

7 FAULT ZONES

Numerous reports on non-uniform Late Weichselian–Holocene uplift patterns have been presented from glaciated regions of NW Europe (Fennoscandia and the British Isles). Many of these have been attributed to neotectonic activity, especially in connection with the high uplift rates during the deglaciation of the different areas. We think it is very likely that sudden tectonic movements were one part of the uplift pattern during (and even after) the deglaciation of the Fennoscandian Ice Sheet. We also think that the actual extent of such movements is far from known. This is owing partly to the fact that the small amplitude vertical movements (in the order of a few metres) are often hidden in the large scale pattern of high uplift rates that characterized the termination of the last glaciation. It is also partly due to the absence of any detailed summary of the Late Weichselian-Holocene uplift history of Scandinavia despite the vast amount of data potentially available. One of the main reasons for creating the data base described above is to try to map anomalies in the uplift pattern in order to find bedrock zones that have been subject to neotectonic activity. However, it should be noted that small amplitude movements (1–2 m) will be almost impossible to find without using some of the field work methods mentioned below. In order to find tectonic movements that are less than c. 5 m with the help of the data base one would need very detailed studies in carefully selected areas with exceptionally favourable conditions. Although a few such studies already exist, there is certainly a need for more research, e.g. at important bedrock boundaries or fracture zones, and around the Protogine shear zone or when possible fault zones have been indicated from the data base analyses.

We see no point in trying to pick out or map fault zones that might have been activated during late-, or postglacial time before a proper and detailed analysis of a large amount of high-quality data has been carried out (cf. the shoreline database). Only then will it be possible to try to draw some conclusions about future fault zones, if this is ever possible!

8 FIELD METHODS FOR ESTABLISHING POSTGLACIAL IRREGULAR HORIZONTAL/VERTICAL BEDROCK MOVEMENTS - SOME EXAMPLES

In the following section only Quaternary Geological methods will be discussed. Apart from researchers like e.g. like N-A Mörner (e.g. Mörner 1978; Mörner et al. 1989 and other studies) and R Lagerbäck (e.g. Lundqvist and Lagerbäck 1976, Lagerbäck 1979 and other studies) very few Swedish Quaternary researchers have seriously worked with the subject of neotectonics. Those two have together covered most of the methods described below. In the following section only field methods will be dealt with since the only Quaternary "laboratory" method, i.e. the shoreline data base, has been mentioned in chapter 5 and 7. It should also be pointed out that paleoenvironmental reconstructions are often necessary in order to ascribe many of the neotectonic indicators mentioned below to a real neotectonic event instead of relating them to other processes, such as e.g. dead-ice melting (faults and collapse structures) or water level changes (slumpings, slidings, and hiatuses).

8.1 STRATIGRAPHIC METHODS

Perhaps the most powerfull method of establishing and dating vertical tectonic movements is to find evidence for displacement of a stratigraphic sequence. If other possibilities (e.g. glaciotectonics or other sediment disturbing processes) can be ruled out this type of evidence can give both an age and an amplitude of the neotectonic activity. This type of neotectonic evidence is obviously best shown in open sections, but can also be demonstrated by e.g. corings and seismic studies. The "school-book example" is a sequence of displaced layers (units) covered by layers not affected by the displacement. One would also expect to find slumpings and slidings of sediments in connection with the fault zone. If it is possible to date the youngest displaced layer the maximum age of the displacing event will be obtained. Additionally, if it is also possible to date the oldest, not displaced layer, the event will possibly be dated accurately providing a good dating tool is available. In cases like this it is also often possible to determine the amplitude of the vertical displacement.

Another way of finding these types of vertical movements is through corings, seismic profiles or even better; a combination of both. These studies can be performed in the sea, in lakes, and on land, but are often expensive. They do, however, have the undisputable advantage of being applicable in most areas, i.e. investigations can be made in areas where possible fault zones occur. The lake based operations, on ice, or with boats, are undoubtedly the cheapest kind of operations, while the other types might be very costly. Many of the likely fault zones are also possibly situated along lake systems. The basic precondition to success lies in finding and being able to follow synchronous units with a combination of seismic profiles and corings. When a possible fault has been detected by e.g. displacement of synchronous units draped by a covering unit detailed corings might be carried out on both sides of the probable fault. In such cases one would also expect to find different sediment disturbances in the cores as e.g. slumping and sliding of layers in the submerged parts and hiatuses in the uplifted parts of the stratigraphy. The amplitude of such vertical movements can usually be estimated.

Horizontal movements are usually more difficult to detect in stratigraphic sequences, since the layering usually is horizontal itself and the effects of such movements often can be ascribed to other processes. The amplitude may also often be difficult to estimate. However, when horizontal movements can be confirmed in stratigraphic sequences, the dating possibilities are the same as for the vertical movements.

8.2 MORPHOLOGIC METHODS

Displaced morphological units such as e.g. displaced raised beaches or glaciofluvial eskers, do probably provide us with some of the best examples of neotectonic activity. This owes to the fact that they are usually clearly distinguished features and that any distinct horizontal or vertical displacement of such formations are easy to distinguish in the field. The amplitude of the displacement can also be estimated. If the age of e.g. the local deglaciation (eskers) or a certain shore line height (raised beaches) is known the maximum age of the displacement will be obtained. A targeted inventory on these kind of features combined with mapping of possible fault zones would certainly provide us with new data concerning the frequency of postglacial neotectonic activity.

The abundant, often glacially eroded, bedrock surfaces in Sweden constitute a potentially vast source for studies of neotectonic activity. It is possible to investigate the effects of larger or smaller, horizontal or vertical, irregularities in bedrock movements by mapping and in detail studying fracture systems on top of such surfaces. Such types of neotectonic indicators are, however, often afflicted with interpretation problems. One of the main reasons is that they are often difficult to date. Without a way of obtaining at least maximum ages of a supposed bedrock displacement it is naturally difficult, at least for Quaternary Geologists, to classify the displacement as having a neotectonic origin. However, when weathering of bedrock surfaces has been restricted by e.g. covering soils or by the hardness of the bedrock type, glacial striations may be a very useful dating tool. Since the predominant part of the glacial striations (and other ice-movement indicators) found on bedrock surfaces were partly formed during the last glaciation (c. 30,000 to c. 18,000 B.P.) but mostly during the deglacial part of that phase (c. 18,000 to c. 9000 B.P.) they constitute a rough dating tool. Under favourable circumstances it is thus possible to find cut off and displaced (horizontally or vertically) striations at fracture zones. If e.g. the striations coincide with the last local ice-movement the maximum age of the displacement will be approximately the same as the local deglaciation age. This will usually also be the case when a significant vertical displacement is found facing the last local ice-movement without any traces of glacial erosion.

8.3 CONCLUSIONS

It is hopefully clear from the above that more targeted efforts have to be made to be able to establish, date, and map the extent of postglacial neotectonic activity. We propose that earth-scientists from different disciplines, with "open eyes", work in common projects to solve these often debated questions. The Quaternary Geologists' best contribution to such projects should probably be by combining shore line data, by e.g. analyses within the shore line data base, with the above mentioned field work methods. Through collaboration with other suitable earth-scientists this could probably be a fruitful starting-point to understand more about the irregularities that most probably have characterized parts of the Fennoscandian uplift history since the last deglaciation.

9 GENERAL CONCLUSIONS

9.1 THE GEOLOGIC DATA

This report has briefly presented some basic knowledge about the Pleistocene climatic changes and their effects on the Earth system as evidenced by different types of geologic data. From those data we can conclude that:

- There is a close linkage between marine records, past climate, and Milankovitch's orbital parameters, but there are also more or less enigmatic linkages that are not well understood.
- The glacial/interglacial cycles have caused sea level changes in the order of 100–150 m.
- The glaciated/deglaciated regions, such as Sweden, have experienced periods of major, sometimes complex, down-warping/uplift movements. that may also involve neotectonic activity.
- Based on a shoreline database it is possible to map the lateglacial/postglacial uplift pattern in detail to detect areas and periods with irregular bedrock movements.

9.2 IMPLICATIONS AND SPECULATIONS FOR THE FUTURE

The report has also outlined the probable driving mechanisms and feed-back effects behind the climatic/environmental changes (excluding man's influence on the atmospheric, terrestrial, and marine systems). Since the basic driving mechanisms most likely have an astronomical origin (the Milankovitch orbital parameters) it is possible to mathematically calculate these parameters into the future. The shoreline database combined with targeted fieldwork will also give us possibilities to map possible future fault zones. We can thus also conclude that:

1. By testing the model output to the past orbital and ^{18}O cycles it is theoretically possible to create a model that produces a likely ^{18}O curve into the future. Based on such a curve (Imbrie and Imbrie, 1980), by relating it to the ^{18}O and climatic changes of the past as revealed by our geologic data, and as long as man's influence on the system is negligible, some general conclusions can be drawn about major future climatic changes and their effects in a chosen region. In this case the conclusions below are mainly focused to the Stockholm region, and that during the coming 100,000 years we may expect approximately the following scenario:
   a) after a long period of increasing mountain glaciation and permafrost extension (both horizontally and vertically) the first glacial episode

will culminate in c. 23,000 years from now, but has from a theoretical viewpoint almost already begun. Although this stadial period seems to become the least extensive of the three, it will most likely mean that large parts of Sweden, probably including the Stockholm region, gradually will be covered by an ice-sheet. The glacial peak will last perhaps 5000 years, but

b) during the following rather cool and dry interstadial conditions larger glaciers will possibly survive in the Swedish mountains. Those conditions together with a probably quite harsh climate at the end of this interstadial suggest that

c) the next ice-sheet will grow quite rapidly. This glaciation will culminate in c. 63.000 years from now but most parts of Sweden will possibly be glaciated some thousands of years before that. It will be a major glaciation, comparable to the maximum of the last glaciation, and the Stockholm region will probably be covered by the ice-sheet for at least 10,000 years, and possibly much longer. A rapid deglaciation will

d) lead to and culminate in interglacial conditions, comparable with todays, in c. 75,000 years from now. These relatively warm conditions may last for 10,000–15,000 years, but will slowly grade into

e) a glacial climate that will culminate in c. 100,000 years from now. This glaciation will probably be rather extensive, but will possibly not develop as rapidly as the former one.

It is obviously impossible to give true quantitative values for any parameter during the five different hypothetic stages (a–e) outlined above. We will, however, try to date and define these stages with purely hypothetical estimations on their age, glacial and permafrost extent, ice thickness (in the Stockholm region), position of relative sea level (in relation to the present sea level in Stockholm), down-warping and uplift. The ages used (years from now) for the glaciations correspond to the periods when the Stockholm region will be covered by ice. The presented estimations can only be looked upon as rough measures and comparisons between the stages. It should also be noted that these parameters during transition periods between the stages are the most difficult to estimate. That was slightly discussed in the sections above, but is not dealt with here.

Stage a): 20,000–25,000 years, glaciation down to the Stockholm region, permafrost in South Sweden, 800 m ice thickness, relative sea level at -25 m, down-warping of 60 m. Stadial conditions.

Stage b): 25,000–45,000 years, mountain glaciation, permafrost in northern and central Sweden, no ice in Stockholm, relative sea level at -60 m, uplift of 50 m. Interstadial conditions.

Stage c): 55,000–70,000 years, a complete Fennoscandian glaciation including Finland, the Baltic republics, W Russia, northern Poland and Germany, and Denmark, permafrost in large areas of Europe, 2500 m ice thickness, relative sea level at +500 m (at glacial maximum when the Stockholm region is ice-covered) to +100 m (at the end), downwarping of 600 m. Full glacial conditions.

Stage d): 70,000–85,000 years, small mountain glaciers, permafrost in the very north, no ice in Stockholm, relative sea level at 0 m, uplift of 600 m. Interglacial conditions.
Stage e): 95,000–? years, glaciation of larger parts of Fennoscandia, permafrost in large areas outside the ice-margin, 1500 m ice thickness, relative sea level at +400 m to +100 m, down-warping of 500 m. Full glacial conditions.

2. It may be possible to detect and map zones which during lateglacial/postglacial time have been characterised by irregular bedrock movements and which could be reactivated during the future down-warping (glaciations) and uplift (deglaciations) phases.

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Thermodynamic modelling of bentonite-groundwater interaction and implications for near field chemistry in a repository for spent fuel

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