CLIMATE OF THE NORTHERN LATITUDES: PAST, PRESENT AND FUTURE

An International Conference held in Tromsø, 2–4 April 1990

Edited by

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CLIMATE OF THE NORTHERN LATITUDES:
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Introduction

The conference 'Climate of the Northern Latitudes: Past, Present and Future' was held in Tromsø, Norway, on 2–4 April 1990. It succeeded two previous Nordic meetings: 'Nordic Symposia on Climatic Change and Related Problems' held in Copenhagen in 1978 and in Stockholm in 1983. With the northern latitudes representing key areas for climatic research, what could have been more appropriate than a conference at the world's northernmost university, the University of Tromsø? Approximately 150 scientists from 13 countries attended.

The main topics of the conference were: climatic change over geologic time; climatic variations within historical time; ozone, climatic gases and the greenhouse effect, and climatic modelling – all topics at the head of the scientific as well as the political agenda.

The present volume of *Norsk Geologisk Tidsskrift* documents the results of the conference, in the form of 26 extended abstracts authored by 52 scientists from nine different countries. All the main topics from the conference are covered in this issue of the journal.

On reading through the titles it is striking to see the number of different types of studies that address climatic questions. Many papers deal with climatic changes over geologic time and discuss both long-term climatic evolution and short-term changes such as the Younger Dryas cooling. It is shown that a major climatic cooling during the last 40 million years of earth history, especially in the high latitudes (cf. Ruddiman), is linked to crustal uplift and the formation of high plateaux in America and Asia. Detailed palaeoclimatic reconstructions of the last interglacial–glacial cycle as well as from the last deglaciation are presented. We have learned a great deal, but still we do not fully understand why the earth responded so dramatically to Milankovitch forcing factors.

Climatic variations within historical time focus on social and economical consequences of climatic change during the last 1000 years. Contrasts in the climate of the last 1000 years are exemplified with the 'Little Ice Age' (AD 1550–1700) and the 'Little Climatic Optimum' (AD 1000–1150). The temperature difference between these periods may have been as much as 1.5°C and had an impact on agriculture and settlement.

The role of the ocean and atmosphere in climatic change is discussed in a separate section dealing with such different topics as carbon flux in the Barents Sea and the ozone layer in northern Europe. Variation in the atmospheric ozone layer during previous decades was much debated during the meeting. Different opinions exist about whether there has been ozone depletion or not.

The various topics of the conference clearly demonstrated that the study of climatic change is a multi-disciplinary activity. Thus, we were also faced with the 'classical' problem of interdisciplinary communication. The conference in Tromsø was clearly a step in the right direction with respect to both communication between disciplines and in-depth discussion within disciplines.

A general comment from many of the reviewers of the papers was that they would have liked more documentation of the data on which some of the results and conclusions were based. We understand and respect their views. However, the editors decided to publish extended abstracts rather than full papers, as this increases the number of contributions and reduces publication time. We hope you agree that we have made the correct choice.

Acknowledgements


Financial support was given by the University of Tromsø, Norway, the Norwegian Research Council for Science and the Humanities (NANF), *Norsk Geologisk Tidsskrift*, Nordisk Ministerråd, Office of US Naval Research (European Office, London) and the US Air Force Office of Aerospace Research and Development (Research Office, London).

We thank the secretary at the Geological Department of the University of Tromsø, Mrs Annbjørg Johansen, for typing and filing and for her patience.

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Stable Isotopic Evidence for Fluid-Rock Interactions in the Ivrea Zone, Italy. A.J. Baker

Reaction Between Ultramafic Rock and Fractionating Basaltic Magma I. Phase Relations, the Origin of Calc-alkaline Magma Series, and the Formation of Discordant Dunite. P.B. Kelemen

Metamorphic history of the Archean Pikwitonei Granulite Domain and the Cross Lake Sub-province, Superior Province, Manitoba, Canada. K. Mezger, S.R. Bohlen and G.N. Hanson

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Interpretation of laminated sediments from glacier-fed lakes, northwest Spitsbergen

MARIANNE CROMACK


Lithostratigraphy and magnetic susceptibility have been used to correlate cores within a series of glacier-fed lakes in northwest Spitsbergen. Variations in downcore laminated couplet thickness within these sediments are presented, compared with meteorological records and dated using 210Pb methods. If the couplets are assumed to be annual, then a close relationship between thickness and temperature variation should be visible. However, results of the dating experiments indicate that two to three laminated couplets may be formed annually, as a result of turbidity currents triggered independently of temperature and glacier size fluctuations. Sedimentation rate declines with distance from the glacier, and declines slightly at a site with glacier retreat. The formation of intermediate sediment traps within the linked lake system appears to have had no effect on the thickness and appearance of laminated couplets.

Marianne Cromack, Scott Polar Research Institute, University of Cambridge, Lensfield Road, Cambridge, CB2 1ER, UK.

Various attempts made to reconstruct high latitude Holocene climatic changes and associated glacial and geomorphic processes have relied largely on the interpretation of glacial and other terrestrial deposits (Innes 1982; Matthews 1985), but also more recently on the lacustrine record (Ashley 1975; Leonard 1986). However, the complexity of the glacial sedimentary system is such that terrestrial and lacustrine sedimentary records may not always record all of the same events. These two types of deposit should therefore be treated as complementary records of Holocene glacier activity.

Meteorological records for the last century from Spitsbergen are compared with variations in the sedimentary record contained in lacustrine deposits and moraine sequences. 210Pb-dating experiments allow deposition dates to be assigned particular horizons within the sediment. These, in turn, may be used to determine sedimentation rates for the catchment and to test the hypothesis that the laminations observed in the cores are annual varves.

Fieldwork for this study was undertaken in Signedalen, northwest Spitsbergen (Fig. 1). The Signedalen catchment is currently 15% glacierized, containing five linked glacier-fed lakes that drain into Signehamna, a small inlet off Lilliehöökfjorden.

During the course of a summer field season (1988) 40 piston cores, numerous grab samples, water samples and lichenometry data were collected. This note is concerned with the preliminary analysis of, and results obtained from the sediment cores.

Each core was measured for whole-core magnetic susceptibility at 1 cm intervals, split longitudinally, photographed and described. Several of the cores were also X-radiographed. Subsamples were taken from each core at 5 cm intervals for moisture content and loss-on-ignition determination.

In general, the cores consist of finely laminated silts and clays, with up to 14 couplets per cm, except in the most ice-proximal lake (Lake 6), where the sediments are sandy. Individual lamination thicknesses were measured in order to compare

![Fig. 1. Map of Signedalen, northwest Spitsbergen. Glaciers are shaded and dotted lines delimit catchment areas for Hajeren and Kleia. Inset. Location of the study area within Svalbard.](image-url)
variation in thickness with temperature fluctuations, and to test the hypothesis that climatic variability would be reflected in sedimentation rates. Subsamples from a number of horizons were dated using $^{210}$Pb activity (Flynn 1968; Appleby & Oldfield 1978) and compared with the number of laminations, in order provide absolute dating control.

Visual correlation of the sediment structures shows a clear similarity between cores from an individual lake (Fig. 2A). In the case of Cores 5/04 and 5/05, there is remarkable similarity between the appearance and thickness of individual laminations throughout the two cores. Results of loss-on-ignition analyses illustrate similar characteristics for all cores from Kløsa. These relationships suggest that the processes responsible for the formation of these laminations are acting on a basin-wide basis. Confirmation of this hypothesis is obtained from the comparison of whole-core magnetic susceptibility traces for each core taken from the lake (Fig. 2B). Similar peaks are illustrated from each core, suggesting a basin-wide deposition influence. The traces also illustrate a clear relationship between distance from source and accumulation rate. The more proximal cores (Cores 5/01 and 5/03) have a greater deposition rate than those more distal cores (Cores 5/04 and 5/05).

Where temperature has been relatively depressed for a sustained period (>5 years), Karlbreen appears to have advanced sufficiently to produce a distinct morainic ridge within its complex forefield. These ridges have been dated lichenometrically, using the growth curve presented by Werner (1988). Moraine stabilization shows excellent agreement with the retreat of Karlbreen after colder phases. Assuming, for the time being, that the couplets are varves, then a crude relationship appears between the couplet thickness variations and the temperature changes and moraine formation phases. When Karlbreen is more advanced, and temperatures reduced, there are thicker laminations deposited. Under circumstances where no other methods of dating were applied to the sediments, this could be taken as reasonable confirmation that the deposits were varves. Small deviations from a good general agreement between couplet thickness and annual temperature can be explained by numerous factors (Perkins & Sims 1983), for example miscounting, storm-induced deposition events within a varve cycle, and a changing relationship between summer and winter temperatures from which an annual temperature is derived.

Results of $^{210}$Pb dating suggest that the laminated couplets are not annual deposits. A total of 529 couplets were counted over the 64 cm of sediment recovered in Core 5/01, and represent around 240 or 250 years worth of sedimentation according to the lead dating (if accumulation rates for this century are extrapolated). For the period covered by the lead record (20th century), Core 5/01 sediment accumulation rates have been calculated as 3–9 mm per year (Table 1). The date of formation of Lake 6, within the Karlbreen Little Ice Age moraine, can be constrained by the lichenometrically dated moraine sequence of Karlbreen glacier and documentary evidence (Norsk Polarinstittut photograph S36 0637) to within the last 75 years at most. Surprisingly, there is no apparent change in sedimentation characteristics recorded by Core 5/01 that would indicate the formation of a sediment trap between the glacier sediment source and the lake Kløsa. At the LIA maximum Karlbreen terminated in Kløsa, so sedimentation would be expected to be similar to that found in Lake 6 for the period that the glacier was in contact with the lake. A slight reduction in the sedimentation rate suggested by lead-dating over this century may be indicative of the steady retreat of Karlbreen and increasing distance of the core site from the sediment source.

Radiocarbon-dating of samples from the base of cores from more distal lakes indicates a substantially lower sedimentation rate, suggesting that the intermediate lakes are effective sediment traps. Sediments from Lake 6 are also laminated, but are coarse in comparison to those of the Kløsa cores, suggesting that Lake 6 represents a reasonably effective sediment trap.

### Conclusions

With the discrepancy between $^{210}$Pb dated and the "varve" chronology for these cores it appears likely that the mechanism controlling the formation of these laminated couplets is turbidity current deposition. The trigger for such deposition is, as yet, unknown for this site, but may be a result of mass movement from the slopes of the LIA Karlbreen moraine as the ice core melts and the moraine stabilizes, or storm-induced flushing of sediment from the glacier and the surrounding catchment, for example.

The processes leading to the formation of laminations within the lakes of Signedalen affect the whole of each lake basin. Sedimentation

### Table 1. $^{210}$Pb dates for Core 5/01 sediments calculated using the CIC model (P. G. Appleby, pers. comm.). Samples from depths below 26 cm contained supported $^{210}$Pb only.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Total $^{210}$Pb activity (Bq kg$^{-1}$)</th>
<th>Age (yrs)</th>
<th>Accumulation rate (mm yr$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.25</td>
<td>110.9 ± 3.5</td>
<td>2</td>
<td>6.25</td>
</tr>
<tr>
<td>5.25</td>
<td>59.1 ± 1.4</td>
<td>43</td>
<td>5.41</td>
</tr>
<tr>
<td>10.25</td>
<td>68.6 ± 1.9</td>
<td>31</td>
<td>3.1</td>
</tr>
<tr>
<td>14.75</td>
<td>58.6 ± 1.4</td>
<td>45</td>
<td>3.2</td>
</tr>
<tr>
<td>20.25</td>
<td>53.2 ± 2.9</td>
<td>55</td>
<td>5.5</td>
</tr>
<tr>
<td>25.75</td>
<td>50.6 ± 2.8</td>
<td>61</td>
<td>9.2</td>
</tr>
</tbody>
</table>
Fig. 2. A. Composite photograph of core sediments from lake Kløsa, Signedalen, with tie-lines connecting layers of contemporaneous sedimentation within the lake. B. Composite plot of whole-core magnetic susceptibility traces for the same cores. Tie-lines connect similar peaks of susceptibility, confirming correlation obtained from core sediment structure.
rates are highest in the most ice-proximal parts of the lakes, and of the catchment as a whole. Dating experiments indicate low sedimentation rates and preclude the occurrence of varves within the sediments. More dating evidence is required to refine sedimentation rate calculations and to model the variations in rates throughout the Holocene. Longer cores and sub-bottom seismic profiling would greatly enhance interpretation of the sedimentary record.

References


The Early Weichselian climate in Finland

LARS FORSTRÖM


The presence of Early Weichselian interstadial organic deposits in northern Finland suggests that the region may have been completely free of ice during these interstadials. The vegetation represented by the deposits reflects the proximity of tundra, with an average temperature for the warmest month (July) of perhaps 10-11°C, some 4-5°C cooler than at present. No comparable interstadial deposits have been found in southern Finland, but it can be assumed that the region was occupied by coniferous forests at that time. It is possible, however, that these deposits, interpreted as interstadial, may represent the later stages of the last interglacial, the Eemian, in which case it is not certain whether evidence of any interstadials is to be found. Organic deposits found at Oulainen and Vimpeli in western Finland probably consist of redeposited material from the end of the Eemian. The fact that they have been preserved suggests that only one glacial advance occurred in the area after that interglacial. If that was the case, the Fennoscandian ice sheet was presumably so extensive throughout the Weichselian glaciation that only southern Finland was free of ice during the Early Weichselian interstadials, and had a climate corresponding to that of Greenland today.

Based on the evidence of a small number of organic deposits found beneath till in northern Finland, Korpela (1969) proposed that the Weichselian glaciation in that region was biphasic, comprising two stadials separated by the cool Peräpohjola interstadial. This conformed to the interpretation put forward for material from northern Sweden by Lundqvist (1967), in which he spoke of the Jämtland interstadial. Arguing from a radiocarbon date of 45,400 ± 2,000 BP (CrN-4491) obtained for the Kostonniska site in Taivalkoski (Fig. 1), Korpela assumed that the Peräpohjola was a Middle Weichselian interstadial. More recently, however, this date has been regarded as too young, and the Peräpohjola has been correlated with the Early Weichselian Brørup interstadial (Hirvas & Nenonen 1987), which was the warmest Weichselian interstadial in Central Europe (Andersen 1961; Behre & Lade 1986). On the other hand, it has since been suggested that the interstadial deposits identified in northern Fennoscandia do not all represent the same period, but are divided between the two main Early Weichselian interstadials recognized in Europe, the Brørup and the Odderade (Forsström 1982, 1988; Lagerbäck & Robertsson 1988; Olsen 1988; Robertsson 1988; García Ambrosiani 1990).

In the case of Finland, the discussion has mainly concerned two sites at Oulainen and Vimpeli in Ostro-
bothnia, in the western part of the country (Fig. 1), the deposits at which have variously been described as interglacial or interstadial (Forsström 1982, 1988, 1989; Aalto et al. 1983, 1989; Donner 1983). They represent a coniferous forest vegetation which does not conform to that reflected in the Peräpohjola deposits, even allowing for their more southerly location (Forsström 1988). The vegetation zones of Finland during the Peräpohjola interstadial, based on the deposits described by Korpela (1969), are shown in Fig. 1, while the corresponding zones in Sweden are based on observations made at Jämtland interstadial sites, of which perhaps the most important is that of Pilgrimstad (Lundqvist 1967). The southern limit of tundra, depicted in Fig. 1 based on Lundqvist (1978) and material presented by Korpela (1969), lies approximately 500 km further south than at present in Finland, implying average temperatures for the warmest month (July) about 4–5°C lower than those experienced nowadays at the same latitudes, i.e. 10–11°C. No deposits corresponding to the Peräpohjola have been identified in southern Finland, but if the coniferous forest zone in Sweden had the form presented by Lundqvist (1978), it can be assumed that such forests grew in the south of Finland as well.

A different explanation is also possible for these organic deposits and for the glacial history of northern Fennoscandia during the Weichselian. The interpretations given of the interstadial deposits in northern Finland and northern Sweden are grounded largely in the fact that they represent a cool or cold climate, with additional support adduced from their stratigraphic position (Lundqvist 1986; Hirvas & Nenonen 1987). It would seem, however, that the stratigraphic models do not take account of changes in the directions of ice flow during deglaciation. If these are noted, many organic deposits can be interpreted as material redeposited in the course of deglaciation which is of no stratigraphic significance (see also Punkari 1984; Forsström & Eronen 1991). The interstadial nature of the deposits can in turn be attributed to the fact that they represent the cold final phase of the last thermometer, regardless of whether this thermometer was an interstadial or an interglacial in northern Fennoscandia.

The Oulainen and Vimpeli deposits in western Finland (Fig. 1) resemble each other sufficiently for them to be generally considered to represent the same thermometer (Aalto et al. 1983; Donner 1983; Forsström 1984, 1985; Hyvärinen 1985), although the coniferous forest vegetation represents too warm a climate to be associated with the Jämtland–Peräpohjola interstadial (Fig. 1), or indeed any Weichselian interstadiial recognized in Central Europe (cf. Menke & Tynni 1984; Behre & Lade 1986; Behre 1989). The Vimpeli site, at least, is very clearly interglacial in character, because it possesses an organic horizon which can be identified as such on the grounds of both plant macrofossils and pollen (Aalto et al. 1989).

The Oulainen and Vimpeli sediments cannot represent the first half of the last interglacial, since they were deposited in a freshwater environment and the area was covered by the sea at that time (Forsström et al. 1988), but they could well correspond to the latter half, especially since they contain horizons indicative of a cold climatic phase. The main problem is that these horizons occur at the base of the organic units (Aalto 1982; Aalto et al. 1983), indicating that where a unit has been transported by glacial action it has been disturbed in the process. Transport of this kind is also evident at Oulainen from the fact that the whole organic unit is located on top of an esker chain associated with the last deglaciation. At Vimpeli, erratics of various sizes were found in the organic unit (Figs. 4 and 16 in Aalto et al. 1989). Thus, these organic deposits were probably transported to their present positions as frozen slabs in the base of the glacier, having been detached from the surface of interglacial deposits, and most likely represent the latter part of the Eemian interglacial, the horizons appearing in reverse order.

The discovery of deposits representing the very end of an interglacial in a redeposited position would seem to presuppose two facts: (1) that the advance of the ice into the area following the interglacial began in the form of an extension of the permanent snow cover, causing interglacial deposits to be buried beneath the ice in a virtually completely preserved form, and (2) that only one glacial advance could have taken place in the area after the interglacial, namely this phase of continuous ice cover. With these points in mind it is quite understandable that once the base of the glacier began to move, the ice would have been able to detach layers from the surface of the material representing the end of the interglacial and carry them as frozen slabs in its base to the sites at which they now occur. If the ice had advanced into the area on a number of occasions after the interglacial, the interglacial deposits would have been covered by till and other sediments at the time of the first advance, so that the ice of subsequent advances would have had to uncover them from beneath these sediments before transporting them to their present positions. In view of the considerable extent of the interglacial deposits, as observed at Oulainen (Forsström 1982) and the well-preserved nature of the material representing the end of the interglacial, the latter alternative may be considered highly improbable.

The ice sheet could be conceived of as having spread to Ostrobothnia only after the Early Weichselian interstadials, which would certainly have meant that only one Weichselian glacial advance occurred there, and that the interpretations put forward for the interstadial deposits of northern Scandinavia could be taken as correct. It is prob-
able, however, that the interglacial deposits of Ostrobothnia were in fact buried beneath sediments during the Early Weichselian even though the ice did not extend to the area, and it should also be borne in mind that the world’s glaciers were of considerable extent even by the first Weichselian stadial (Shackleton 1987), so that it is unlikely that the Fennoscandian ice sheet would have been particularly small. In fact, it is not at all improbable that the ice sheet may have reached the Baltic Sea in the south and the White Sea in the east at this period (Forström 1989; Forström & Eronen 1991).

As the author sees it, the Fennoscandian ice sheet covered the region of Ostrobothnia already during the first Early Weichselian stadial and melted only during the final deglaciation, i.e. in the Holocene. It was presumably at its minimum extent during the Early Weichselian interstadials, as indicated by both the sea-level record and the deep-sea core data (Labeyrie et al. 1987; Shackleton 1987), possessing approximately the limits depicted in Fig. 2. The model presupposes that the deposits interpreted as interstadial in northern Fennoscandia in fact represent the last interglacial, the Eemian. It is impossible to prove this to be so as regards individual sites, but it does appear to be the case that none of the deposits concerned has definitely been proved to represent an interstadial, so that one is forced to rely on indirect evidence in this matter.

Fig. 2. Extent of the Fennoscandian ice sheet during the same period as in Fig. 1 (the present author’s interpretation).

The organic deposit unearthed at Brumunddal in Norway (Fig. 2) represents a decisive site as far as determination of the minimum extent of the ice sheet is concerned, as its pollen includes fairly large quantities of larch, leading Helle et al. (1981) to correlate it with the Brørup interstadial, a period during which this species was growing in Denmark at least (Andersen 1961). The occurrence of larch does not prevent Brumunddal from representing the last interglacial, however, if we consider that the species could have spread to the area from the north, via Finnish Lapland (Forström 1990). It should also be noted that the peat horizon at that site must have been moved there by glacial action, as the till beneath it contains peat fragments (fig. 3 in Helle et al. 1981). In that case the organic deposit there would probably date from the last thermomer, on the same grounds as the Oulainen and Vimpeli sediments. This last thermomer could not have been the Brørup interstadial, however, as this was the older of the two more or less equally warm Early Weichselian interstadials, the Brørup and the Odderade (Behre & Lade 1986).

There are a number of reasons why the Fennoscandian ice sheet may have remained fairly extensive during the Early Weichselian interstadials. Firstly, these were relatively cool climatic periods by comparison with the Eemian or Holocene (Behre & Lade 1986), so that the melting of the glaciers must have been correspondingly more gradual. The second reason is connected with the glacial history. Prior to the Eemian interglacial and the Holocene, the ice sheets were so extensive and caused such a depression in the earth’s crust that they were surrounded for the most part by sea or vast ice lakes as they melted (Eronen 1983; Forström et al. 1988). This meant that most of them melted by calving, which would explain the disappearance of vast areas of glaciers within a comparatively short time (Denton et al. 1986). The ice sheets of the rela-
tively short-lived Early Weichselian stadials, on the other hand, will have been smaller and the depressions in the earth's crust shallower, so that the melting ice will have emerged in a supra-aquatic position more quickly, to become anchored on dry land. In the presence of a cool climate, the melting of the supra-aquatic ice would have been slow, so that the ice will not have retreated very much on the land during the interstadials. One indirect indication of this can be said to be the well-developed system of fjords on the Norwegian coast, which in the opinion of Porter (1989) is attributable to the fact that the ice margin hovered close to the coast for long periods under average glacial conditions in the course of the glaciations. In other words, the ice sheet was inclined to fluctuate around a mean extent, with both its maximum periods and its minima, the interglacials, remaining of short duration. Thus the Early Weichselian interstadials, like the stadials, may be said to represent largely internal fluctuations in mean conditions rather than exceptional conditions of the kind associated with interglacials, which caused the ice to melt almost entirely in Fennoscandia. The Greenland ice sheet is the closest present-day situation one can find to the supposed Weichselian minimum as shown in Fig. 2. By comparison, therefore, the climate in what is now southern Finland would have been an arctic one, similar to that of present-day Greenland.

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Early decay of the Barents Shelf Ice Sheet – spread of stable isotope signals across the eastern Norwegian Sea

MARA S. WEINELT, MICHAEL SARNTHEIN, ELKE VOGELSANG & HELMUT ERLENKEUSER


Based on planktonic δ¹⁸O records from 16 sediment cores obtained from the Norwegian Sea, we traced a meltwater signal marking the onset of Termination I. From its discharge point south of Bear Island, it spread alongshore the Norwegian margin up to 63°N, thereby suggesting a clockwise-directed drift of icebergs and surface water in the Norwegian Sea, the opposite of the present circulation pattern.

Based on seven sediment cores with high average sedimentation rates (8-100 cm/1000 years) (Table 1) from the Bear Island sediment fan and nine cores with medium high to low sedimentation rates from the central Norwegian Sea and the Fram Strait (1.5-15 cm/1000 years) (Fig. 1), we were able to map the oxygen and carbon isotope signals of the planktonic foraminifer *N. pachyderma* l.c., signals linked to this catastrophic event, and to identify the source and the probable regional distribution pattern of the meltwater spike, as defined by its δ¹⁸O amplitude (Fig. 2).

Oxygen isotope measurements from sediment records with medium to high resolution (e.g. cores 23258 to 23262) suggest that the meltwater event (δ¹⁸O shift) of the Barents Ice Sheet immediately succeeded the

Jones & Keigwin (1988) and Lehman et al. (1991) presented the first AMS-¹⁴C dated oxygen-isotope evidence from the Fram Strait suggesting that the marine-based Barents Shelf Ice Sheet had disintegrated rapidly about 14,500-13,500 years ago. This melting process occurred very early in the chain of deglacial events during Termination IA.

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Fig. 2. Glacial meltwater spike as defined by the δ¹⁸O amplitude between the end of the Last Glacial Maximum and the δ¹⁸O minimum peak (arrows) immediately subsequent, shown for a number of core transects from the Norwegian Sea. For locations see Table 1.
The age of Termination I was defined as lasting from 14,800 to 9,000 radiocarbon years BP and corrected for U/lb ages to 18,300-9,800 calendric years BP (after Bard et al. 1990).

Table 1. Locations, \( \delta^{13}C \) values (N. pachyderma l.c.) for the meltwater spike, and average sedimentation rates for Termination I for cores from the Norwegian Sea. The age of Termination I was defined as lasting from 14,800 to 9,000 radiocarbon years BP and corrected for U/lb ages to 18,300-9,800 calendric years BP (after Bard et al. 1990).

<table>
<thead>
<tr>
<th>Core</th>
<th>Geogr. position</th>
<th>Average sedimentation rate over Termination I (cm/1000 yr)</th>
<th>( % \delta^{13}C ) at the ( \delta^{18}O ) meltwater spike (maximum)</th>
<th>Source of ( \delta^{18}O ) data</th>
</tr>
</thead>
<tbody>
<tr>
<td>MG 123</td>
<td>79°16' N 00°48'E</td>
<td>6.1</td>
<td>?</td>
<td>Morris (1988)</td>
</tr>
<tr>
<td>PS 21295</td>
<td>77°59' N 02°25'E</td>
<td>1.8</td>
<td>?</td>
<td>Jones &amp; Keigwin (1988)</td>
</tr>
<tr>
<td>M 17724</td>
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<td>2.6</td>
<td>0.09</td>
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</tr>
<tr>
<td>M 23288</td>
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</tr>
<tr>
<td>M 17732</td>
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<td>0.33</td>
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</tr>
<tr>
<td>M 22329</td>
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<td>0.13</td>
<td>Own data</td>
</tr>
<tr>
<td>M 23260</td>
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<td>8.2</td>
<td>-0.19</td>
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</tr>
<tr>
<td>M 23261</td>
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<tr>
<td>M 23262</td>
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<td>ca. 100</td>
<td>-0.18</td>
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<tr>
<td>M 23068</td>
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</tr>
<tr>
<td>M 23071</td>
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<tr>
<td>M 23074</td>
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<td>14.4</td>
<td>0.01</td>
<td>Own data</td>
</tr>
<tr>
<td>PS 23205</td>
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<td>11</td>
<td>-0.16</td>
<td>Own data</td>
</tr>
<tr>
<td>HM 31-36</td>
<td>64°15' N 00°31'E</td>
<td>3.5</td>
<td>0.24</td>
<td>Zahn in:</td>
</tr>
<tr>
<td>HM 31-33</td>
<td>63°38' N 01°46'E</td>
<td>9.4</td>
<td>0.14</td>
<td>Ramm (1989) Jansen &amp; Erlenkeuser (1985)</td>
</tr>
<tr>
<td>HM 25-09</td>
<td>63°03' N 04°47'E</td>
<td>27.5</td>
<td>-0.24</td>
<td>Ramm (1989) Jansen &amp; Erlenkeuser (1985)</td>
</tr>
</tbody>
</table>

Last Glacial Maximum (\( \delta^{18}O \) values 4.7-4.9‰ PDB). The event was marked by a minor short-term precursor \( \delta^{18}O \) spike, which cannot be identified at more distal locations and lower sedimentation rates. The event ended as abruptly as it started with higher \( \delta^{18}O \) values of 3.8-4.0‰, prior to a second negative \( \delta^{18}O \) event, tentatively ascribed to the Allerod. The first chronostratigraphic correlations were based on cores PS 21295 and V27-60, where the \( \delta^{18}O \) events were AMS-dated by Jones & Keigwin (1988) and Lehman et al. (1991). Further \( \delta^{18}O \) records were recently dated by Duplessy, Gif sur Yvette (Sarnthein et al., in press).

The maximum \( \delta^{18}O \) spike (marked in Fig. 2) lay further offshore, SW of Bear Island, at 72°N, 9-11.5°E. The amplitude of the spike reached up to 3.2‰. This equates to a local salinity reduction of more than 3‰ (Vogelsang 1990). Contrary to our expectations, the amplitude of the spike decreased by 1‰ toward the nearby ancient shoreline and ice margin, despite the high data resolution of the records from cores 23261 and 23262. Moreover, the meltwater signal, which can be traced north toward the Arctic along the Spitsbergen margin, was very narrow and insignificant (amplitudes of less than 2.0 and 1.5‰). However, the meltwater signal spreading south expanded over a wide E–W range and reached up to 2.4‰ at the Voring Plateau and 1.8‰ off Trondheim. This dispersion of the meltwater signal, which was probably related to icebergs melting further offshore, suggests a dominantly clockwise surface-water circulation in the Norwegian–Greenland Sea during early deglacial times, exactly the opposite of the present counter-clockwise currents. Probably this circulation pattern was driven by prevailing northeasterly winds (as proposed by the model of Joussaume 1989). Modern summer winds are west–east. The deglacial NE winds and currents may have caused coastal upwelling of more saline subsurface water along the Bear Island and Norwegian margin, leading to the unexpected onshore gradient of the \( \delta^{18}O \) meltwater spike immediately offshore at 72°N and 63°N (Fig. 2). The upwelling-induced productivity resulted in extremely negative nearshore \( \delta^{13}C \) values along the continental margin downstream up to 63°N of –0.2 to –0.8‰ for N. pachyderma, equal to 0.6-0.0‰ \( \delta^{13}C \) of the ambient sea water (Labeyrie et al. 1985) (Table 1). Both the widespread meltwater and the clockwise surface circulation in the Norwegian Sea imply that the deep-water formation strongly declined during that time, which resulted in a regime of estuarine deep-water circulation and a breakdown of Atlantic deep-water ventilation during early Termination IA (Boyle & Keigwin 1986; Fairbanks 1989; Sarnthein & Tiedemann 1990).

Acknowledgements. – We thank G. Jones and L. Labeyrie for careful reviews. This study was supported by the Sonderforschungsbereich 313 at the University of Kiel.

References
Boyle, & Keigwin, L. D. 1986: Comparison of


Late Quaternary paleoceanography in the southern Barents Sea

MORTEN HALD, LAURENT D. LABEYRIE, DAVID A. R. POOLE, PERIVAR STEINSUND & TORE O. VORREN


The paleoclimatic evolution of the Barents Sea and continental slope occurred in three main steps: Prior to 13-12 kyr BP the area suffered from cold temperatures and the salinity of the upper water masses was reduced by glacial meltwater. After 12 kyr BP the salinity reduction was smaller and at 10 kyr BP oceanic paleoclimatic conditions similar to those of the present were established and have been fairly stable since.

The southwestern Barents Sea (Fig. 1) is an important area for reconstructing Quaternary paleoclimates for many reasons. This high latitude epicontinental sea is at present characterized by sharp climatic gradients and thus may prove an important modern analogue for shifts between past warm and cold environments on the Norwegian continental shelf during the Quaternary. The area is also close to the sites of formation of Norwegian Sea Bottom Water and the present shallow Barents Sea contributes to this water mass (Midttun 1985). In addition, the study area is located close to the northern extension of warm Atlantic water responsible for the poleward heat transport.

Sediments from the continental slope, shelf and shallow seas may provide records having higher resolution than in the deep sea due to high deposition rates. The continental slope off the western Barents Sea has been an important depocenter throughout the Quaternary (Vorren et al. 1989) and glacial troughs on the shelf/shallow Barents Sea represent local depocenters (Vorren et al. 1984). However, compared with those from the deep sea, records from the shelf and upper slope may be more problematic to interpret due to the pres-

Fig. 1. Bathymetric map of the Barents Sea showing the locations of the investigated sites and main trends in the oceanic surface circulation pattern. Solid arrows show transport of warm Atlantic water. Broken arrow shows transport of cold Arctic water. 1 = the Troms shelf record, 2 = the Barents slope record, and 3 = the Barents shelf record.
ence of unconformities, reworking and possible overprinting by local climatic and environmental signals.

The water masses of the Barents Sea are dominated by Atlantic water (Norwegian current) and coastal water (Norwegian Coastal Current) entering the study area from the south and Arctic water from the north (Mosby 1968). The average modern drift-ice limit is about 74°N (Vinje 1977). In the southwestern parts of the Barents Sea, fairly warm (5-7°C) Atlantic water dominates, whereas towards the east and north this water cools and mixes with Arctic water. Atlantic water dominates down to 600-800 m on the slope. Homohaline deep water is present below this depth and is characterized by a fairly constant temperature at ca. -1°C below 1200 m (Mosby 1959). North of the oceanic polar front, Arctic water dominates (Fig. 1) with temperatures frequently below 0°C.

Location study sites

Site 1, the southernmost of the three sites (Fig. 1), is located on the continental shelf (300-500 m depth) off Troms county in northern Norway. This location is closely connected to the mainland and the Norwegian Sea and provides a good link for correlating the continental and deep-sea records. Site 2 is located in the southwestern Barents Sea, which is more offset both to land and the Norwegian Sea. Hald et al. (1989) demonstrated that this site is a sensitive record of the last deglaciation and the onset of the North Cape current, an easterly branch of the Norwegian Current at present transporting warm Atlantic water into the southern Barents Sea. The hydrography of both of the shallow sites is at present dominated by Atlantic water. Site 3 is located on the continental slope off the western Barents Sea with the homohaline deep water at 1495 m depth. The present bottom water temperatures here are ca. -1°C, thus the isotope record from this site should reflect mainly isotopic composition of the sea water, while the temperature effect is negligible.

Material and methods

Oxygen isotopes on benthic foraminifera were investigated from the three sites in the Barents Sea area (Fig. 1). The sediments were sampled using a gravity sediment corer. Stable isotope measurements were performed at the Norwegian GMS laboratory at the University of Bergen and at the Centre des Faiblés Radioactivites, CNRS, Gif sur Yvette, France, using Finnigan MAT 251 mass spectrometers. The samples were prepared following the procedures described by Shackleton & Opdyke (1973), Shackleton et al. (1983) and Duplessy (1978).

As no single benthic species is abundant enough for analysis throughout the stratigraphies, we have analyzed six different benthic species. These are: Cassidulina reniforme (Nørvang), Cassidulina teretis (Tappan), Cibicides lobatulus (Walker & Jacob), Cibicides wuellerstorfi (Schwager), Nonion barleeanum (Williamson) and Nonion labradoricum (Dawson). Their δ18O disequilibrium fractionation values relative to the ambient sea water (given in Table 1) have been corrected for in all the plots (Fig. 2).

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**Table 1.** Delta 0-18 disequilibrium fractionation values relative to the ambient water for some benthic foraminiferal species. For value for C. wuellerstorfi cf. Labeyrie et al. (1987); for C. reniforme see Hald & Vorren (1987b) and for C. teretis cf. Jansen et al. (1988). The others are from Hald et al. (in prep.).

<table>
<thead>
<tr>
<th>Benthic foraminiferal species</th>
<th>Delta 0-18 (% vs. PDB) disequilibrium value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cassidulina reniforme</td>
<td>0</td>
</tr>
<tr>
<td>Cassidulina teretis</td>
<td>0</td>
</tr>
<tr>
<td>Cibicides lobatulus</td>
<td>0.6</td>
</tr>
<tr>
<td>Cibicides wuellerstorfi</td>
<td>0.64</td>
</tr>
<tr>
<td>Nonion barleeanum</td>
<td>0.4</td>
</tr>
<tr>
<td>Nonion labradoricum</td>
<td>-0.1</td>
</tr>
</tbody>
</table>

---

**Fig. 2.** Oxygen isotope (benthic foraminifera) versus age from site 1 (Troms shelf), site 2 (Barents shelf) and site 3 (Barents slope). For location see Fig. 1. Site 1 curves are based on measurements of the following species: 0-11.7 kyr BP, C. lobatulus; 11.7-13.9 kyr BP. Site 2 curve is based on: 0-11 kyr BP, C. lobatulus; 11-12 kyr BP, N. barleeanum; 12-12.7 kyr BP, C. lobatulus and N. labradoricum. Site 3 curve is based on *N. pachyderma* (sinistral). All the values are corrected for disequilibrium effects (see Table 1).
The last deglaciation

Detailed palaeoenvironmental reconstructions of the last deglaciation and the present interglacial (Holocene) have previously been discussed based on the following stratigraphical investigations: chronostratigraphy (conventional radiocarbon), lithostratigraphy (Vorren et al. 1983a, b, 1984), foraminiferal stratigraphy (Hald & Vorren 1984, 1987a; Hald et al. 1989), macro fossil stratigraphy (Thomsen & Vorren 1986) and stable oxygen and carbon isotope stratigraphy (Hald & Vorren 1987b).

The early phase of the deglaciation was characterized by deposition of glaciomarine sediments with very sparse fossil assemblages. On the Troms shelf area (site 1) there are laminated fine sediments with semi-barren foraminiferal assemblages and a total lack of macrofossils, indicating that the sea surface was sea-ice covered, at least seasonally. Following this, there was a gradual immigration/diversification of both macrofauna and foraminifera. Immediately prior to 14 kyr BP there are indications of open but fairly cold oceanic shelf waters. During this period the ice margin receded from a position close to the shelf edge in the south with at least two halts or readvances, named the Flesen and D events. The latter event is reflected in the stratigraphy by a high dropstone content and fossil assemblages indicative of a near-glacial environment. During this time, further north in the Barents Sea, Vorren et al. (1988) tentatively suggest a deglaciation of both Bjørnøyrenna and the banks and troughs in the southern Barents Sea. Between 14 and 13 kyr BP near-glacial conditions occurred during the D event in the Troms shelf record (site 1). Benthic foraminiferal transfer functions indicate that the bottom water temperatures were cold during this period, around -1°C (Steinsund et al. this volume). During the period 13-10 kyr BP the ice margin receded to a mid-inner fjord position along coastal northern Norway and western Svalbard, while the eastern Svalbard archipelago was still glaciated. On the Troms shelf a decline in iceberg rafting in the troughs, incipient winnowing on the banks and shifts in the fossil macro- and microfauna which occurred around 13 kyr BP, are linked to an incipient intrusion of Atlantic Water. Possibly a branch of this water reached western Svalbard concurrently (Forman et al., 1987; Mangerud et al. 1987), while the data from the Barents shelf (site 2) indicate that a corresponding change was delayed by ca. a thousand years in the southwestern Barents Sea. Around 12 kyr BP there is a marked increase in the bottom water temperatures (Steinsund et al. this volume) reaching modern temperatures (4-6°C). Between 12 and 10 kyr BP, large temperature fluctuations are seen which did not stabilize until the early Holocene, when values close to modern ones were reached.

The Pleistocene/Holocene boundary marks a drastic and rather sudden environmental shift from glaciomarine to open marine conditions. In the southwestern parts of the investigated area, Atlantic Water in the Norwegian current totally replaced the glacially influenced water masses. South of 72°N most of the foraminiferal and macrofauna were replaced by other species (Vorren et al. 1978; Hald & Vorren 1984, 1987a; Thomsen & Vorren 1986), whilst in the northwestern Barents Sea a somewhat larger number of the glaciomarine species persisted into the Holocene (Elverhøi & Bomstad 1980; Østby & Nagy 1981; Mackensen & Hald 1988), probably as a response to lower temperatures in the more northerly areas of the Barents Sea.

Isotope stratigraphies

The chronology of the Barents slope record is based on five AMS radiocarbon datings (Hald et al. in prep.) The Troms and Barents shelf records (Fig. 2) are composite records, dated by radiocarbon and discussed in detail by Hald & Vorren (1987b), Hald et al. (1989) and Hald et al. (in prep.).

The light isotope event between 13 and 14 kyr on the Troms shelf and the light values at ca. 13 kyr on the Barents shelf occurred at a time of cold (negative) bottom water temperatures at these sites (Steinsund et al., this volume). Thus these events probably reflect meltwater pulses correlating to the D event off northern Norway.

Between 12 and 10 kyr BP the isotope difference between the Barents shelf and Barents slope records can mainly be explained in terms of the temperature gradient established during this period (Steinsund et al. this volume). But for the Troms shelf, parts of this curve, compared with the other two curves, exceed the temperature effect, indicating a salinity reduction.

Termination I B, centered around 10-8 kyr BP, is marked in all the records and correlates to the marked faunal and sedimentological shifts (cf. the section above). We interpret the large light isotopic excursion seen in the Troms record and the slope record as the final meltwater discharge of the last deglaciation. After this event, the records are characterized by fairly stable isotope values indicating that fairly constant ocean temperatures prevailed throughout most of the Holocene. The isotope differences between the curves can be explained by the present-day temperature difference between the sites.

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References

Spatial and temporal distribution of Laurentide and Fennoscandian meltwater during the last deglaciation

GLENN A. JONES

The oxygen isotopic composition of planktonic foraminifera in eleven AMS 14C-dated cores from throughout the North Atlantic and Norwegian-Greenland Sea has been used to examine the temporal and geographic distribution of salinity anomalies in these ocean basins during the last deglaciation. Four regional patterns emerge, with the earliest anomaly found throughout the Norwegian-Greenland Sea at ~15000 BP and interpreted as the decay of the Barents/Fennoscandian Ice Sheet. Three additional anomalies occur in the Gulf of Mexico and the subtropical-to-subpolar North Atlantic at ~13500, 12000, and 9500 BP. The approximate 2000 years periodicity of these four salinity events is consistent with that predicted by the salt oscillator hypothesis of Broecker et al. (1990).

Despite long-term scientific interest in understanding the temporal and geographic distribution of meltwater events during the last deglaciation (~15000-18000 BP), it has only been in recent years that appropriate data (e.g. Fairbanks 1989) have been available to allow accurate and detailed marine geological reconstructions to be made. Emiliani et al. (1975) and Kennett & Shackleton (1975) documented, using the stable isotopic composition of planktonic foraminifera, that major low salinity events occurred in the Gulf of Mexico during the last deglaciation. These events had their origin in the discharge, through the Mississippi River, of isotopically light meltwaters derived from the decay of the southern margin of the Laurentide Ice Sheet. Berger et al. (1977) and Berger (1978) speculated further that the observed meltwater events were recorded globally. However, it was argued that the deleterious effects of low sedimentation rates and bioturbation smooth these events in ocean sediments to the point where they could only be observed by deconvolution processing. Jones & Ruddiman (1982) showed that incorrect assumptions about bioturbation mixing intensities were used in these deconvolution studies and that the most likely distribution of meltwater events were not global but rather restricted largely to the North Atlantic. All of the above-mentioned studies were limited by inadequate age control, and little more could be said about the timing of meltwater events other than that they occurred during deglaciation.

By the mid-1980s radiocarbon dating by Accelerator Mass Spectrometry had become available and approximately one high-quality AMS-dated record from the North Atlantic was being published per year (e.g. Duplessy et al. 1986; Bard et al. 1987; Jones & Keigwin 1988;
Keigwin & Jones 1989; Broecker et al. 1989). Combining these well-dated, high-quality stable isotopic records with the first good measure of the temporal evolution of eustatic sea-level rise for the past 17,000 years and hence the global ice-volume curve (Fairbanks 1989), it is now possible to remove from each stable isotopic record the ice-volume effect which is common to all cores. Furthermore, by making some first-order assumptions about the temporal change in sea-surface temperature from the glacial to the Holocene, the effects of temperature upon these isotope records can be removed as well, resulting in a temporal record of the local salinity anomalies at each core location. These observed anomalies are directly related to the rapid transfer of isotopically light glacial meltwaters from the ice sheets to the isotopically heavy ocean.

I report on eleven stable isotope records directly dated by 146 AMS radiocarbon measurements (Fig. 1).

My approach is as follows: the oxygen isotope record from each core is plotted to time based upon the AMS radiocarbon measurements. For this study there is an average of 13 (5 minimum and 24 maximum) AMS radiocarbon age control points per core. This record is then compared with the global ice-volume record derived in Fairbanks (1989). For most of the cores the glacial-Holocene amplitude exceeds that predicted from the contribution by global ice volume (Fig. 2c) and in

Fig. 2. End-member examples of how salinity anomalies are calculated. a, e. Core SU81-18 (A) exhibits a large glacial-Holocene \( \delta^{18}O \) amplitude resulting from large-temperature changes. Heavy solid line indicates data reported in Bard et al. 1987. The light solid line indicates the global ice-volume curve derived in Fairbanks (1989). Core PS21295-4 (E) exhibits no temperature anomaly (i.e. measured glacial-Holocene \( \delta^{18}O \) range between is equal to global ice-volume signal). Data from Jones & Keigwin 1988. b, f. Glacial-Holocene temperature change required to 'expand' ice-volume curve to equal the amplitude of the measured values shown in a and e. For internal consistency the temperature increase was assumed to have commenced at all core sites at 14,000 BP and ended at 8,000 BP. c, g. The global ice-volume curve 'expanded' by the temperature effect in b and f is shown as a dashed line and now matches the amplitude of the measured values (solid line). d, h. Measured \( \delta^{18}O \) (heavy line in c and g) minus the temperature adjusted global ice-volume curve (dashed line in c and g). For SU81-18 two anomalies have amplitudes greater than 0.5 per mille and are interpreted as low salinity having influenced this site at approximately 12,000 and 9,500 BP. For PS21295-4 one anomaly is larger than 0.5 per mille and is interpreted as low salinity having influenced this site at approximately 15,000 BP.
other records the measured amplitude agrees with that of global ice volume (Fig. 2g). This increased amplitude is a result of temperature increases from the last glacial maximum to the Holocene, as each 1°C increase adds 0.23 per mille to the oxygen isotopic amplitude (Epstein et al. 1953). The temperature increases calculated from the excess amplitudes of the isotopic records are consistent with those temperature increases calculated funnally (CLIMAP 1981).

The global ice volume curve (Fig. 2a, e) can be ‘expanded’ to match the measured glacial-Holocene isotopic amplitude in each core by calculating the difference between the measured glacial-Holocene amplitude and the global ice-volume amplitude. This difference is assumed to be solely due to temperature effects. I then make the simplified assumption that the glacial-Holocene temperature increase at each site began at 14,000 BP, ended by 8,000 BP, and occurred uniformly over this time interval (Fig. 2b, f).

Although one could argue for a different temperature history from each and every core, this approach is designed to be robust and every core is treated in the same manner. The ‘expanded’ global ice volume and measured isotopic records are compared (Fig. 2c, g) and any deviation is assumed to be a result of local temperature effects in the vicinity of the core. Repeating this procedure for each of the eleven cores shown in Fig. 1, we obtain a record of the temporal distribution of local salinity anomalies at each core site. The quality of each record will vary depending on several factors, such as how well the local temperature record has been incorporated, the sedimentation rate and bioturbation intensity, and the density of radiocarbon age-control points. All salinity anomalies exhibiting an amplitude greater than 0.5 per mille are shown (Fig. 3). The duration of each anomaly is determined by the width (i.e. time duration) at the half height of the anomaly. This study does not address the amplitude distributions of the salinity anomalies at each core, as this will be treated more thoroughly elsewhere.

There are two initial results that can be drawn from this study. The first is that there are four geographic patterns to the salinity anomalies. The first region is the Norwegian-Greenland Sea and it exhibits the earliest evidence of decay of the Northern Hemisphere ice sheets, beginning at approximately 15,000 BP. The five records obtained from throughout this basin demonstrate that the entire Norwegian-Greenland Sea isotopically records the decay of the Barents Shelf Ice Sheet (Jones & Keigwin 1988). Although the timing of the initiation of this event can be determined, the duration is artificially extended due to a combination of relatively low sedimentation rates, bioturbation mixing, and limited 14C age control points. This is especially true for cores V27-86 and V23-74.

There is no evidence of any younger discrete events observed in these records. The second pattern characterizes the Gulf of Mexico and the western Atlantic in the vicinity of the Blake-Bahama Outer Ridge and the Bermuda Rise. These records exhibit three anomalies centered around 13,500, 12,000, and 9,500 BP. The origin of these low salinity waters is inferred to be meltwater discharge from the Mississippi River. The third pattern characterizes the eastern North Atlantic with only one or two events observed at 12,000 and 9,500 BP, respectively. The origin of these events is inferred to be the Mississippi River, and these correspond in timing with the maximum meltwater discharge predicted in Fairbanks (1989). The 13,500 BP event does not appear to have been large
enough to manifest itself in the sea-level record of Fairbanks (1989) or in those regions distal from the inferred meltwater discharge down the Mississippi River. The final pattern characterizes those records south of 20°N in the subtropical Atlantic. These records (one shown; two others not shown) exhibit no anomalies and are used to infer that the influence of deglacial meltwater discharge was restricted to the North Atlantic and Norwegian–Greenland Sea surface waters as predicted by Jones & Ruddiman (1982).

The second result of this study is that the timing of the four observed salinity anomalies (−15,000, 13,500, 12,000 and 9,500) has an apparent periodicity of near 2,000 years when using the radiocarbon time-scale. This is a result consistent with the predicted time-scale of ice-sheet-ocean interactions as outlined in the salt oscillator hypothesis (Broecker et al. 1990). The data in this paper offer some of the first empirical constraints of this hypothesis by presenting both the temporal and geographic distributions of meltwater discharge into the North Atlantic during the last deglaciation. This study demonstrates that with high-quality AMS radiocarbon-dated isotope records and a high-quality estimate of eustatic sea level the regional and temporal distribution of glacial meltwater can be mapped. However, the details of the climatological implications of a study of this kind are beyond the scope of this presentation and must be covered elsewhere.

References


Ice-sheet/ocean interaction at the mouth of Hudson Strait, Canada, as a trigger for Younger Dryas cooling

GIFFORD H. MILLER & DARRELL S. KAUFMAN


Two major Late Wisconsin expansions of the Labradorian sector of the Laurentide Ice Sheet are recognized on SE Baffin Island. They occurred immediately prior to and immediately after the Younger Dryas cold interval of NW Europe; during most of the Younger Dryas chron, the Labradorian sector receded. We postulate that the change to more meridional atmospheric and oceanic flow during hemispheric deglaciation, beginning about 13 ka ago, produced greater snowfall over Labrador/Ungava, resulting in a vigorous NE flow from the Labradorian dispersal center by 11 ka BP. The flux of icebergs produced at this time (between 300 and 2400 km$^3$ a$^{-1}$) was about the same as the freshwater influx to the North Atlantic from the diversion of Lake Agassiz drainage to the St. Lawrence (Teller 1987). A sustained flux of icebergs to the North Atlantic may have significantly affected regional oceanographic conditions by cooling seawater temperature, increasing sea-surface albedo, and reducing wind mixing of surface waters, thereby reducing the volume of water required to cap the North Atlantic. We suggest that a combination of a massive iceberg flux and increased freshwater discharge via the St. Lawrence (cf. Broecker et al. 1989) may have been required to initiate Younger Dryas cooling.

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Although the role of Hudson Strait, the major trough connecting the former Laurentide Ice Sheet with the open ocean, has been the object of considerable interest, its control on ice-sheet dynamics has, until recently, remained unconstrained by field observation. Based on geologic field mapping of ice-directional features (Fig. 1), coupled with till composition, geochronology, and evidence for large-scale oceanic/atmospheric changes that occurred during the last deglacial cycle, we have developed a model of self-oscillating glacier dynamics that explains the glacial activity at the mouth of Hudson Strait, and provides a trigger for Younger Dryas cooling. The field evidence requires a dynamic ice sheet, operating independently of major topographic constraints and with shorter response times than previously envisioned. Ice-directional features and erratic lithologies of the last glacial demonstrate that ice flowed NNE across outer Hudson Strait, rather than down the strait as hypothesized in previous ice-sheet reconstructions. Ice advanced more than 600 km from a Labradorian source, overtopped southeastern Baffin Island, crossed Frobisher Bay, overtopped Loks Land (400 m a.s.l.) and advanced well onto Hall Peninsula (Fig. 2). A more complete presentation of the field evidence and its implications is given by Miller & Kaufman (1990).

Temporal control is provided by radiocarbon dates on marine shells in ice-contact marine deposits, individual shell fragments from till, and foraminifera in marine cores recovered from the adjacent shelf, with supplemental age control from amino acid geochronology. The timing of the oldest (maximum) Late Wisconsin advance is poorly con-

Fig. 1. Southeastern Baffin Island showing orientation of ice flow based on ice-erosional features on bedrock in regions free from topographic channeling of the ice (low-relief summits, rolling uplands). Flow pattern across Meta Incognita Peninsula is from Miller et al. (1988).
strained on land, although the duration of the Late Wisconsin advance(s) can be shown to have been very brief. Marine sediments of the Loks Land interstadial (Miller 1985), that underlie the Late Wisconsin drift of a Labradorean advance on eastern Loks Land, contain an abundant in situ fauna with non-finite radiocarbon ages. Temperature reconstructions, based on the limited difference in the racemization of amino acids in shells from the interstadial and adjacent dated deglacial sites, limit the duration of the most extensive advance(s) to a brief interval of no more than 2 ka.

Two separate Late Wisconsin advances are inferred from morphostratigraphic relations on Hall Peninsula. Radiocarbon dates on individual shells collected from till of the younger event, the Gold Cove readvance, and on in situ shells from deglacial marine sediments constrain the age of this readvance. Eleven reservoir-corrected dates on single or paired valves of Mya truncata collected from Gold Cove till and marine deposits overridden during the readvance cluster between 10.0 and 10.5 ka BP. They demonstrate that the Gold Cove readvance postdates an interval of open water of at least this duration. The readvance was brief; by 9.4 ka BP ice had receded to the southern shore of Hudson Strait (Lauriol & Gray 1987). The land-based chronology is consistent with that of a well-dated piston core raised from the mouth of Frobisher Bay (HU 84-008; Evans 1990; Andrews et al. 1990). The 8 m core records ice-distal marine sedimentation from ca. 14 ka until 11.2 ka BP, followed by a brief advance of ice over the site between 11.2 and 10.8 ka BP. Open marine conditions after 10.8 ka BP were followed by a brief interval of extreme ice-proximal sedimentation about 10 ka BP, correlative with the Gold Cove readvance. The terrestrial evidence also documents a later, less extensive readvance about 8.6 ka BP during the Cockburn Substage (Fig. 2, Miller et al. 1988), but the ice margin did not reach the core site.

We postulate that the shift to more meridional atmospheric and oceanic flow during hemispheric deglaciation beginning about 13 ka ago increased the moisture flux to the Labrador sector of the Laurentide Ice Sheet, leading to an advance of its NE margin that culminated about 11 ka ago. The configuration of the northeastward advancing ice margin can be estimated using available terrestrial field evidence, the bathymetry and seismic stratigraphy of the adjacent shelf, and marine-core data. The flux of ice necessary to maintain such a margin positioned in deep water (at least 600 m) facing the Labrador Sea must have been substantial. Based on a minimum cross-sectional area and reasonable estimates of ice velocity, the iceberg flux to the North Atlantic at the maximum extent of the Labradorean advance, coincident in time with the onset of the Younger Dryas, was between 300 and 2400 km$^3$ a$^{-1}$, about the same as the freshwater influx to the North Atlantic from the diversion of Lake Agassiz drainage to the St. Lawrence (Teller 1987). The discharge of a significant volume of ice into the North Atlantic may have substantially altered regional oceanographic conditions, which in turn could have imparted an important feedback to the glacial system. A sustained iceberg flux into the North Atlantic would have cooled sea-surface and air temperatures, increased sea-surface albedo, and diminished the effectiveness of wind mixing of surface waters, thereby reducing the volume of water required to cap the North Atlantic. The return to a 'glacial' ocean circulation mode during the Younger Dryas resulted in reduced precipitation over Labrador/Ungava so that the extension of Labradorean ice onto the shelf could not be maintained. By 10.5 ka ago the terminus had withdrawn into Hudson Strait. The Gold Cove readvance reflects a dynamic response of the ice sheet to increased precipitation immediately following the Younger Dryas, although the ice sheet was unable to sustain the advance against rising terrestrial temperatures and sea level. By 9.5 ka BP Labradorean ice had again receded within Hudson Strait, so that when the St. Lawrence outlet reopened the iceberg flux from Hudson Strait was minimal, and the

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**Fig. 2.** Time-distance diagram for the NE margin of the Labrador sector of the Laurentide Ice Sheet; cross-section (from Ungava Bay northeastward across Loks Land) is located in Fig. 1. Radiocarbon dates are keyed as follows: (1) Solid symbols are AMS $^{14}$C dates on single shells in till or conventional dates on paired valves subsequently overridden by ice; open circles and triangles represent dates on shells collected from postglacial sediments. (2) Triangles represent dates on shells associated with the Gold Cove readvance; circles represent shells associated with the Cockburn readvance; squares represent AMS dates on foraminifera from core HU 84-008 (see Andrews et al. 1990, for paleoenvironmental interpretations). Abbreviations are as follows: C = Cockburn readvance; GC = Gold Cove readvance; Max = maximum Late Wisconsin advance; Ungv = Ungava Bay; MIP = Meta Incognita Peninsula; FB = Frobisher Bay; LL = Loks Land; Shelf = Baffin Shelf north of Loks Land.
meltwater alone was insufficient to cap the North Atlantic. We suggest that a combination of a massive iceberg flux and increased St. Lawrence discharge (cf. Broecker et al. 1989) may have been required to initiate the Younger Dryas.

Ruddiman & McIntyre (1981) looked for evidence of a major increase in ice-rafted detritus (IRD; defined as quartz/feldspar grains >125 µm) into the North Atlantic at the onset of the Younger Dryas, but found no discernible increase in IRD flux. However, a sample of ice-proximal sediment from eastern Leks Land contains 40% carbonate; of the remaining non-carbonate grains, less than 5% are larger than silt. Similarly, both ice-proximal and ice-distal glacial-marine facies in cores at the mouth of Hudson Strait contain close to 50% detrital carbonate and are dominated by grains <125 µm; 5 to 15% of the grains in ice-proximal facies are >125 µm, whereas ice-distal facies contain less than 2% grains >125 µm (Evans 1990). Consequently, the iceberg flux from the Labrador sector cannot be monitored by the classically defined IRD fraction in deep-sea cores, and thus its absence cannot be used to reject the hypothesis of an iceberg flood from the Labradorian sector at the onset of the Younger Dryas.

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References

The climate of the northern latitudes as a function of ocean/atmosphere circulation and the Earth's rate of rotation

NILS-AXEL MÖRNER


Major short-term climatic changes on the decadal to century basis are controlled by the delicate balance between northward penetration of warm Atlantic water – the Gulf Stream variability – and southward penetration of cold Arctic water. These ocean current changes are driven by the interchange of angular momentum – in a feed-back mechanism between the 'solid' Earth and the hydrosphere. This mechanism has been identified for (1) the El Niño events, (2) the last 300 years' instrumental records, (3) the Holocene short-term fluctuations (50–150 years), (4) the Milankovitch variables in glacial as well as pre-glacial times, and (5) certain long-term changes. This also holds for the major late glacial changes within the 13–10 ka interval.

This theory of redistribution of energy and mass via the ocean current systems due to a feed-back interchange of angular momentum seems also to apply to the high-amplitude

The climate of the northern latitudes is primarily controlled by the changes in the atmospheric circulation (cf. the present cold poles versus the Early Cenozoic warm poles) and in the ocean circulation (cf. the present contrast between the eastern and western sides of the Greenland–Norwegian Seas due to the effects of the Gulf Stream).

The main conclusion of the Second Nordic Symposium on Climatic Changes and Related Problems (Mörner & Karlén 1984) was that the short-term climatic changes (i.e. in the time range of decades up to 100–150 years) do not represent global increases and decreases in incoming solar radiation, but rather the redistribution of heat within the terrestrial system. This redistribution of heat can only take place via ocean circulation changes which, in their turn, must be driven by the interchange of angular momentum between the 'solid' Earth and the hydrosphere (Mörner 1984a, b). In subsequent papers (Mörner, e.g. 1987, 1988, 1989), I have expanded on this theory and demonstrated that it seems well documented not only for major Holocene and Late Glacial climatic shifts and changes, but also instrumental records of recent centuries and ENSO events of recent years. Even the Milankovitch variables must have strong effects on the global ocean current systems (via differential rotation); by that also affecting the global paleoclimate.

Redistribution of heat via the ocean current systems implies short-term changes of compensational type instead of globally uniform type. This means that most short-term changes are of local to regional validity when we look at the similarity in characteristics. A strong signal, however, may well have a global significance only, in this case, the changes must be of compensational nature with a total conservation of angular momentum, energy and mass; i.e. the time of change is similar but the direction and amplitude differ.

This theory of redistribution of energy and mass via the ocean current system due to a feed-back interchange of angular momentum seems also to apply to the high-amplitude

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**Fig. 1.** The high-amplitude climatic changes recorded in NW Europe between 13,000 and 10,000 BP are here proposed to represent the redistribution of heat over the globe via the ocean current system due to a feed-back interchange of angular momentum between the 'solid' Earth and the hydrosphere. The sea-level records from NW Europe are directly opposite to those recorded off West Africa, indicating the displacement of ocean water masses between the two sides of the Atlantic (when water masses hit the American coasts, the Gulf Stream is intensified, and vice versa). The Gulf Stream is counterbalanced by the cold Labrador Current. It is therefore consistent with the proposed model that the new glaciation curve of Hudson Strait is directly the opposite to the classical curve of NW Europe.
changes between 13,000 and 10,000 BP as illustrated in Fig. 1. The classical Scandinavian paleoclimatic record gives a sudden warming at about 13,000 BP, a sudden cooling at the onset of the Younger Dryas Stadial at about 11,000 BP, and a drastic warming at the end of the Younger Dryas Stadial at about 10,300–10,000 BP. This scheme has been widely used (even in a global context) though it should be limited to NW Europe (Mörner 1984b). The corresponding sea-level changes are a rising, a falling and a rising trend, respectively. From West Africa, Tastet (1989) has presented a sea-level record which gives exactly the opposite trends, however. This seems to indicate that we are primarily dealing with ocean circulation changes (due to the interchange of angular momentum between the ‘solid’ Earth and the hydrosphere) causing the redistribution of mass (sea level) and heat (climate) in line with the theory of Mörner (1984a, b, 1988). The cold Labrador Current has a counter-balancing relation to the warm Gulf Stream. In further support of the ocean circulation model for the high-amplitude changes 13–10 Ka, it therefore seems highly significant that Miller & Kaufman (1990) reported glacial fluctuations in the Hudson Strait region that are directly opposite to those recorded in Fennoscandia. All this is illustrated in Fig. 1.

Consequently, the Younger Dryas event seems to represent a very strong change in the ocean circulation system that led to a significant cooling in Europe and southward penetration of Arctic water in the North Atlantic (Mörner 1990). This change seems to have affected many other parts of the globe. In this respect it may have global dimensions. At the same time, it does not seem to represent a global loss and gain in solar radiation, but a redistribution of energy within the terrestrial system (Mörner 1988).

References
In the last few decades significant progress has been made in the modelling of land ice masses and in understanding the role they play in the climate system. Although many questions remain unanswered, a number of processes have been identified that establish a clear and important link between ice-sheet evolution and changes in the climate system. Several of these will be discussed in this contribution: direct implications of ice-sheet growth/decay for the annual energy budget, the albedo feedback, the elevation-mass balance feedback, and the effect on ocean circulation and precipitation rates.

Whether a glacier or ice sheet influences the climate depends very much on the scale: the effect of a mountain glacier will be restricted to the valley in which it flows. Many glaciers and small ice caps in a single mountain area may modify climate on a regional scale, but only for large ice sheets does one expect world-wide effects. The interesting aspect is that an effect on the local climate can still make an ice mass grow larger and larger, thereby gradually increasing its radius of influence.

At present, land ice covers an area of 15,861,766 km² (Haeberli et al. 1988), which is slightly more than 3% of the total area of the earth’s surface and about 11% of the continents. At the peak of the last glaciation, the latter figure was in the order of 25%. By far the largest amount of present ice is stored on the Antarctic continent, about 90% of the total. Of the remainder, most is found on Greenland. Mountain glaciers and small ice caps contribute very little to the total ice volume, in fact. This total volume is estimated to be in the order of 32 million km³, equivalent to approximately 70 m of sea-level rise when melted and spread uniformly over the world ocean. For an up-to-date summary of the distribution of glaciers and ice sheets over the globe, the reader may consult the reference quoted above.

The purpose of this paper is to discuss how ice sheets interact with the climate system. No attempt is made to review the paleoclimatic evidence of the Pleistocene glacial cycles, nor to discuss the wide variety of ice-age theories that have been proposed in the last few decades. A rather qualitative description of physical mechanisms will be given, showing that ice sheets form an active component of the climate system.

Ice and the energy budget of the climate system

Irrespective of the details of ice-sheet growth or decay, it is interesting to compare the amounts of energy involved with the basic energy flows in the climate system. The global balance between incoming solar radiation and outgoing terrestrial (infrared) radiation can be written as

\[(1 - A)S = \sigma T^4 = \tau \sigma T^4, \quad (1)\]

where \(S\) (341 W/m²) is the solar constant divided by 4 (because the area of the surface of the earth is four times the cross-section intercepting the sun’s rays), \(A\) the mean planetary albedo (currently 0.3), \(\sigma\) the Stefan-Boltzmann constant \((5.67 \times 10^{-8} J/(m^2 K^4 s))\), and \(T\) the effective radiative temperature of the planet. The outgoing terrestrial radiation is frequently expressed in the surface temperature \(T_s\), which makes it necessary to introduce an effective transmissivity of the atmosphere for radiation emitted at the surface (denoted by \(\tau\)). By inserting the values for \(A\), \(S\) and \(\sigma\) it follows that \(T \approx 255 K\), i.e. the radiative temperature is much lower than the mean surface temperature and the radiating level must be at several kilometres’ height. This is typical for a greenhouse planet, where the atmosphere is not transparent for infrared radiation. To arrive at the observed mean surface temperature (about 288 K), it appears that \(\tau = 0.61\).

The annual amounts of solar and terrestrial radiation passing the ‘top of the atmosphere’ are about
3.8 \times 10^{24} \text{ J/yr. Table 1 shows how this compares to typical amounts of energy needed to heat the ocean or melt an ice sheet. Accumulating the solar radiation absorbed at the surface during 3.5 years would be sufficient to melt all present land ice. In fact, this would also be required to heat the ocean by about 1.7 K. So the amounts of energy involved in ice-sheet growth and decay are relatively small (see also Oerlemans & Van der Veen 1984). This is further illustrated by considering a rapid deglaciation: melting of all ice-age ice within 5000 years requires 3.8 \times 10^{24} \text{ J per year, which is only 0.084\% of the absorbed solar radiation in the present climate. So it is hard to imagine that the solar radiation absorbed at the surface during 3.5 years would be sufficient to heat the ocean or melt this requires:

<table>
<thead>
<tr>
<th>Mass of land ice ( x \times 10^{19} \text{ kg} ) for melting</th>
<th>( 1.6 \times 10^{25} \text{ J} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mass of ice in full glacial ( 7.3 \times 10^{19} \text{ kg} ) return to interglacial</td>
<td>( 9.3 \times 10^{24} \text{ J} )</td>
</tr>
</tbody>
</table>

Table 1.

is implicitly taken into account. So, instead of equation (1), we now use

\[ (1 - \alpha)S = a + bT^3, \]

The coefficients derived from Fig. 1 are: \( a = 206 \text{ W/m}^2 \text{K} \) and \( b = 2.2 \text{ W/(m}^2\text{ K)} \).

The sensitivity of surface temperature to a change in the absorbed amount of solar radiation is determined by the coefficient \( b \). If this amount were to change by 10 W/m\(^2\), a temperature change of 4.5 K would be needed to compensate for this in the outgoing infrared flux. So the smaller the \( b \), the larger the 'climate sensitivity'. It is interesting to make a comparison with a straightforward linearization of the right-hand side of equation (1). Without a greenhouse effect (\( r = 1 \)), it is found that \( b = 5.42 \text{ W/(m}^2\text{ K)} \), and the temperature change would only be about 1.8 K. With a constant greenhouse effect, \( b = 3.36 \text{ W/(m}^2\text{ K)} \), and the temperature change would be 3 K. This is for a linearization around the present climatic conditions, with a mean surface temperature of 288 K. Although this analysis is extremely simple, it clearly demonstrates the importance of the water vapour feedback.

We now turn to the albedo feedback. The change of mean planetary albedo for a small change in surface temperature is denoted by \( \beta = \delta A/\delta T \). The linearized form of the radiation balance is then easily derived and the solution for the temperature perturbation \( \delta T \) reads:

\[ (1 - A_o)\delta S - \beta S \delta T = b \delta T \]

\[ \delta T = (1 - A_o)\delta S / (b + \beta S) \]

In this expression \( A_o \) is the present-day mean planetary albedo of 0.3, and \( \delta S \) represents a change in absorbed amount of solar radiation received by the earth. Equation (3) shows that climate sensitivity is enhanced by the albedo feedback (\( \beta < 0 \)). Theoretically, when the albedo feedback is so strong that \( b + \beta S < 0 \), a runaway situation occurs in which the entire earth would soon be covered by ice.
There are various ways to estimate the value of $\beta$. One is to compare the climate of an ice age as simulated by General Circulation Models (GCMs) of the atmosphere to the present climate, and compare the radiation budgets. Several studies of the ice-age climate have been undertaken, with different GCMs, and all based on the surface conditions as provided by CLIMAP (1976). A particular comprehensive analysis has been performed by Manabe & Broccoli (1985), who have compared the total radiation budget of the climate system for present and ice-age surface conditions with the same orbital forcing. During full glacial conditions the amount of solar radiation absorbed in the climate system is decreased by 3 W/m$^2$, which corresponds to a change in mean planetary albedo of slightly less than 0.009. With a decrease in mean surface temperature of about 3 K this implies that $\beta = 0.003$ K$^{-1}$. A similar value was derived in Oerlemans & Van den Dool (1978) from experiments with an energy-balance climate model with a fairly detailed parameterization of the albedo. The value of $\beta S$ is about $-1$ W/(m$^2$ K), so the denominator in equation (3) becomes significantly smaller and climate sensitivity is enhanced.

In Table 2 some results are summarized in terms of the change in global mean surface temperature for a 1% increase in isolation. The significance of the albedo feedback is obvious, but it should be added that this is the combined effect of snow cover, sea ice and land ice.

<table>
<thead>
<tr>
<th>Mass balance of the present ice sheets</th>
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| As noted in the Introduction, the surface conditions at the ice/snow surface ultimately determine what will happen to an ice sheet. The surface mass balance, frequently expressed in m water equivalent per year, forms the link between ice sheet and meteorological conditions and thus deserves careful study. A natural start is to look at the two large ice sheets we can observe today. In Fig. 2, two mass balance ‘profiles’ for the ice sheets of Greenland and Antarctica are shown. The mass balance of the Antarctic ice sheet is positive everywhere, at least when average values over elevation intervals are considered (there are places where the mass balance is locally negative because of very dry conditions and/or snowdrift). The largest values are found in the coastal regions, at elevations between 0 and 1000 m. The accumulation in the interior is very low: the polar desert. This is due to very low absolute humidity associated with the extremely low air temperature. The total annual accumulation on the ice sheet is estimated to be 1800 km$^3$. Virtually all loss of ice is by iceberg calving, but it is not known whether the ice sheet is close to equilibrium. A totally different picture applies to the Greenland ice sheet. There is a large accumulation zone with a positive balance and a smaller but important ablation zone. The ablation rates on the lowest parts are very high, and in fact comparable to what is observed on the tongues of mountain glaciers in, for instance, southern Norway and the Alps. The low values of surface albedo due to accumulated atmospheric dust are the common factor here. A recent estimate of the total accumulation on the ice sheet is 535 km$^3$ (Ohmura & Reeh 1990). It is generally assumed that half of the loss is by iceberg calving, the other half by melting and runoff. As for the Antarctic ice sheet, it is not known to which degree gain and loss match. The most important quantity governing the mass balance at a particular location is the annual mean surface air temperature $T_{sa}$. The dependence of accumulation and ablation on $T_{sa}$ is shown in Fig. 3. The graph is schematic and it is supposed that cloudiness, seasonal temperature cycle, etc., are constant under a change of $T_{sa}$. Depending on local conditions, snowfall reaches a maximum for annual temperatures somewhere between $-5$ and $+5$°C. Ablation becomes significant when summer temperature exceeds the melting point, so the curve shown here applies to a location with an annual
temperature range of 30 K. This is a large range, but not unusual in high latitudes. The resulting mass balance curve shows a peculiar behaviour: for $T_w < -10^\circ C$ the mass balance increases with temperature (range A in the figure), for $T_w < -12^\circ C$ it decreases with temperature (range B). The value of $-12^\circ C$ should not be taken as a universal number, of course. Depending on local conditions it may vary in a range running from -20 to $-5^\circ C$. The important point to make here is that no monotonic, let alone linear, relation between glacier mass balance and annual air temperature exists.

It is interesting to consider the Antarctic and Greenland ice sheets in the light of Fig. 3, as this may give a first indication about changes in ice-mass accumulation in a warmer world. Annual surface temperature on the Antarctic continent ranges from roughly $-10^\circ C$ in the coastal areas to $-55^\circ C$ on the interior plateau. So virtually all of the Antarctic ice sheet is in range A, and the mean surface balance would certainly increase when climate would get warmer. On the other hand, a significant part of the Greenland ice sheet is in range B. In fact, in several studies it has been shown that the mean surface balance of this ice sheet would become significantly smaller in a warmer climate (Ambach & Kuhn 1989; Oerlemans et al. 1990). Here the increase in snow accumulation is not large enough to compensate for the stronger melting.

Mass-balance conditions on the ice-age ice sheets of the northern hemisphere must have been similar to those of the Greenland ice sheet today, with significant mass turnover and large melting rates in the margin, and lower accumulation rates towards the north. In most ice-age climate simulations with GCMs the mass balance of the ice sheets has not been studied explicitly. However, Manabe & Broccoli (1985) give the geographical distribution of the 'net ice accretion rate' and this confirms the basic similarity with Greenland conditions: accumulation of the order of 0.5 m w.c./yr, ablation up to several m w.c./yr in the lowest parts of the southern margin.

A further remark concerns the fact that the accumulation rate is bounded while ablation can become very large for sufficiently high temperatures. This implies that ice sheets can decay much faster than they grow, which is in accordance with the palaeoclimatic evidence. As will be discussed later, other mechanisms exist that make this asymmetry even more pronounced.

Feedback between elevation and mass balance

As surface temperature is an important factor with regard to the mass balance, the feedback between increasing surface elevation and ice accumulation can be powerful. The full implications of this for the Pleistocene glacial cycles were first realized by Weertman (1961), who gave an elegant mathematical analysis of the problem.

The feedback is illustrated in Fig. 4 for typical northern hemisphere conditions. The continent is bounded in the north by the Arctic ocean, and climatic conditions are represented by an equilibrium line that slopes upward to the south. The intersection of the equilibrium line with sea level is termed the climate point $P$. Climatic change can now be considered as shifting the equilibrium line up and down, or moving the point $P$ back and forth. In the event of climatic cooling, $P$ shifts southward, and as soon as $P$ is on the continent an ice sheet will form. Whether it will grow or not depends on the location of the equilibrium point $E$, the intersection of the equilibrium line and ice-sheet surface. The surface elevation of a growing ice sheet increases, which automatically shifts $E$ southwards, even without further cooling. In fact, we can be sure that the growth of the large ice-age ice sheets was only possible because of this powerful positive feedback mechanism.

In a (very) first approximation, the profile of a continental ice sheet only depends on its size $L$, and its mean elevation is given by $0.67 \times \sqrt{(\Delta L)}$, see e.g. Oerlemans & Van der Veen (1984). Here $\Delta L$ is a parameter determined by the stiffness (yield stress, actually) of the ice. The equilibrium size of an ice sheet can then be obtained from the condition that the total net balance is zero. For a sloping equilibrium line as shown in Fig. 4, this can be worked out to find $L$ for any prescribed position of the equilibrium line, i.e. for any value of $P$. Fig. 5 shows a typical solution diagram. The heavy line represents stable steady states, the dashed line unstable steady states. Two critical

![Fig. 4. An illustration of the feedback of surface elevation on mass balance. $P$ is the 'climate point' (intersection of equilibrium line with sea level) and $E$ the equilibrium point (intersection of equilibrium line with the ice-sheet surface). $E$ may shift southward due to ice-sheet growth, even if the climate does not change ($P$ stays in position).](image-url)
Mechanisms that can destabilize ice sheets

The discussion in this section will be qualitative. Although several mechanisms described here have been studied by modelling, a discussion of this would be too technical in the present context. Also, many of those modelling studies are really sensitivity studies, of which the output still depends in a critical way on the boundary conditions and input parameters. They have shown, however, that the mechanisms depicted in Fig. 6 are potentially important. This figure summarizes, in a sketchy way, the possible history of a glacial cycle, without referring to a particular period or time-scale. The various factors are discussed below.

It is easy to imagine that isostatic sinking of the bed on which an ice sheet grows must affect the mass balance. The associated lowering of the surface implies higher air temperature and, when averaged over the ice sheet, smaller mass balance (most of a growing northern hemisphere ice sheet will be in region B of Fig. 3). The outflow of material in the asthenosphere takes time, so the isostatic adjustment lags the increasing load. Without detailed calculations we may conclude that bedrock sinking will slow down the growth of an ice sheet, or even stop it when the climatic cooling that initiated the ice sheet is not large enough or is of short duration only.

The same lagged response of the bedrock accelerates ice-sheet decay once it has been started. First, during the process of retreat of the ice edge, the bed will be much lower than the corresponding isostatic equilibrium state. The ice retreats in a region where ice thickness was large 'shortly' before, and the bed may still be below sea level. As the bed will slope downwards towards the ice sheet, large proglacial lakes and seas can form. Calving ice fronts, in combination with high sliding velocities of the ice lobes, can then lead to very fast retreat in certain places.

Time-dependent models of the

points appear, reflecting the occurrence of hysteresis. There is a range of values of the climate point \( P \) for which two stable states are possible: no ice sheet or a large one, which is maintained by the elevation-mass balance feedback. When \( P \) is located south of the critical point \( C_2 \), there must be an ice sheet, when \( P \) is located north of the critical point \( C_1 \), a stable ice sheet cannot exist.

A few important conclusions can be drawn from Fig. 5. For convenience, we denote the characteristic response time of the northern hemisphere ice sheets by \( T_{ICE} \), the characteristic time-scale of a forcing mechanism of the climate system by \( T_{CLIM} \). If \( T_{ICE} \gg T_{CLIM} \), then the ice sheet will hardly react to the forcing. Its equilibrium size will correspond to the mean forcing. In the case of \( T_{ICE} \ll T_{CLIM} \), the ice sheet will always be in equilibrium, i.e. on the steady state curve of Fig. 5, thereby following the arrows. The most complicated case occurs when \( T_{ICE} \approx T_{CLIM} \), because of transient effects. For the ice sheet to jump from one stable branch of the solution diagram to another it is necessary that \( P \) changes, and that this change lasts long enough.

When considering climatic forcing by changes in the earth's orbit (the Milankovitch theory, which is currently the most popular theory of the Pleistocene glacial cycles — for a review, see Berger 1988), we are in the situation that \( T_{ICE} \approx T_{CLIM} \). The precession of the equinoxes has an apparent period of 22 ka (1 ka = 1000 years). A corresponding time-scale for the forcing then is 5 to 10 ka, which is somewhat less than the time ice sheets need in order to grow to a size where they start to affect the global climate. For insolation variations associated with changes in the obliquity (major period 44 ka), \( T_{ICE} \) and \( T_{CLIM} \) are quite comparable. The effect of changes in the eccentricity of the earth's orbit are on a somewhat larger time-scale (periods of \( \approx 100 \) ka and \( \approx 400 \) ka).

In view of these facts, the response of the climate system to Milankovitch insolation variations cannot be linear, at least not when the build-up and decay of large ice sheets are involved.

Weertman (1976) was the first to take the consequence of this observation. He forced a simple time-dependent model for a single northern hemisphere ice sheet with the Milankovitch insolation variations, by assuming that the equilibrium line moves up and down in proportion to changes in summer insolation at high latitudes. His attempt was partly successful. The model was able to reproduce realistic variations in ice-sheet size, but did not generate as much power at the longer time-scales (100 ka, in particular) as is normally seen in the deep-sea oxygen-isotope records. The slope of the equilibrium line and the accumulation rate are the most important parameters concerning the behaviour of a model ice sheet. The smaller the slope, the larger the ice sheet, whereas the time-scale for growth is mainly determined by the accumulation rate.

In several studies with ice-sheet models (Oerlemans 1980; Pollard 1982) it has been shown that a full-grown northern hemisphere ice sheet, generated with realistic model parameters, will not disappear because of high summer insolation alone. Rapid ice-sheet decay must have involved one or more destabilizing mechanisms.

Fig. 5. Equilibrium size of a northern hemisphere ice sheet, in dependence of climatic conditions (position of \( P; P > 0 \) implies 'climate point' on the continent). Note that hysteresis occurs. The open circles indicate critical points and the dashed part of the curve represents unstable equilibria. From Oerlemans & Van der Veen (1984), original analysis by Weertman (1961).

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response of the solid earth to ice load range from extremely simple local isostatic adjustment (Oerlemans 1980), thin-channel models (Pollard 1982; Oerlemans 1982) to spherical visco-elastic representation of the mantle (Birchfield & Grumbine 1985; Hyde & Peltier 1985). The response of combined ice-sheet/geodynamic models to Milankovitch forcing is discussed in these papers, and significant differences have been found. Still, the results demonstrate that the interaction of geodynamics and ice-sheet growth/decay is capable of generating glacial cycles with a timescale significantly larger than that of the forcing.

The thermal regime of an ice sheet can also contribute to destabilization. The temperature distribution is determined by advection (vertical and horizontal), internal heat generation by shearing, and geothermal heat input through the base of the ice sheet. The higher ice temperatures are always found at the base. During the process of growth, the basal ice layers get warmer and at some stage reach the (pressure) melting point. Subglacial meltwater is then produced. When the bed consists of rock, the meltwater may be discharged in sheet or channel flow (Weertman 1972), unless the entire warm zone is surrounded by ice, frozen to the bed. In that case subglacial lakes may occur (they have been found in Antarctica). The subglacial water may also saturate a till layer to generate a deformable bed, which may favour rapid ice discharge and lead to thinner ice sheets (e.g. Boulton & Jones 1979; Fisher et al. 1985). Recent investigations of the ice streams of West Antarctica have made clear that such till layers are present there, and seem to be the reason for the high ice velocities (e.g. Alley et al. 1987). To what extent similar situations occurred on the Laurentide and Fennoscandian ice sheets is not clear. ‘Downdraw’ of ice from the major ice domes through deep outlet glaciers has also been suggested as a significant destabilizing factor (Denton & Hughes 1981). In any case, there is theoretical and observational evidence for a number of processes that may accelerate the decay of an ice sheet when the base is sufficiently warm, having in common that they all effectively reduce the friction of ice flow at the bed.

When a glacial period approaches its maximum, sea ice cover in the Atlantic ocean extends much further southward, and the moisture supply to the more northerly parts of the Fennoscandian and Laurentide ice sheets will be restricted. Oerlemans & Vernekar (1981) conducted many experiments with a zonal climate model, including atmospheric dynamics and a full hydrological cycle, to see how the precipitation regime would change due to extensive glaciation. It was found that precipitation decreases by up to 50% north of about 60°N, and increases by up to 30% in the 40 to 60°N latitude belt, particularly in winter. The reason for this is the enhanced storm activity along the southern ice and sea-ice margin. A similar conclusion was reached by Manabe & Broccoli (1985), in an experiment with a general circulation model of the atmosphere, run for ice-age boundary conditions. The net effect appears to be a gradual drying out of northern hemisphere ice sheets when conditions get colder, and in particular when sea-ice cover becomes extensive.

The interaction between ice-sheet evolution and ocean circulation is complicated, and may also involve components that destabilize ice sheets. For example, when deglaciation has been initiated, a low-salinity meltwater layer may cover part of the North Atlantic Ocean and lead to more extensive sea-ice
cover (for a more extensive discussion, see for example Ruddiman & McIntyre 1981). A likely consequence is the reduction in snowfall on the ice sheets, thereby contributing to the deglaciation process. On the other hand, the more stable stratification associated with input of meltwater affects the deep ocean circulation. There is evidence from proxy palaeoclimatic data and from modelling studies that the production of North Atlantic Deep Water (NDAW) is extremely sensitive to small changes in fresh-water input (Broecker et al. 1985; Manabe & Stouffer 1988; Maier-Reimer & Mikolajewicz 1989). In fact, it has been suggested that the Younger Dryas climatic event was forced by a substantial reduction in NADW production and the consequent decrease in poleward heat transport by the meridional circulation in the Atlantic Ocean.

Apart from processes directly linked to the shape and state of the northern hemisphere ice sheets, other feedback loops in the climate system had a major influence on the evolution of the Pleistocene climate. The atmospheric concentration of a number of greenhouse gases, in particular, appears to vary significantly between glacial and interglacial conditions (e.g. Barnola et al. 1987). Attempts to delineate the effects of albedo feedback, atmospheric CO₂ concentration and orbital insolation variations suggest that, in terms of radiative forcing of the climate system, these are all of the same order of magnitude (Broccoli & Manabe, 1987; Lorius, pers. comm. 1989).

Concluding remarks

There is general consensus that the processes outlined above are of direct importance to the growth and decay of northern hemisphere ice sheets, but the details remain unclear. More work with coupled models, in which atmospheres, ocean, land ice and solid earth are treated in a time-dependent (but not necessarily synchronous) way, is needed for a further diagnosis of the glacial cycle. Such models should also comprise a component describing the geochemical cycles that have a direct effect on the energy balance of the climate system. It will be crucial to test models against field data. This must involve: (i) The use of large modern climatological data sets to test in detail the mathematical representations of physical processes included in the models. (ii) Further construction of palaeoclimatic data sets, mainly in the form of snapshots of the Pleistocene climate, for verification of climatic states generated by coupled models. This will imply large interdisciplinary efforts along the lines set by the COHMAP Project (1988).

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Late Cenozoic benthonic foraminiferal stratigraphy from outer Bjørnøyrenna, Barents Sea: palaeoclimatic implications

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A benthonic foraminiferal stratigraphy based on a 140 m long shallow drilling from outer Bjørnøyrenna is presented. The foraminiferal fauna mainly reflects a polar shelf environment. However, a zone with an abundant and diverse microfauna was found. Based on the fauna, the sediments and amino acid measurements, we conclude that part of this zone was deposited during the Eemian interglacial (isotope substage Se).

In this study we present a foraminiferal stratigraphy of a 140 m long shallow drilling (7317/10-U-01) located close to the shelf break southwest of Bjørnøya, at a water depth of 465 m (Figs. 1 and 2). Our aim is to reconstruct palaeoenvironments and ages of the recovered stratigraphy. The study consists of a detailed analysis of 52 sediment samples (starting at 7.55 m under the sea bed), where the following investigations have been carried out: (1) Grain-size distribution, (2) foraminiferal analyses, and (3) amino acid measurements.

Benthonic foraminiferal stratigraphy

The analysed interval is divided into five assemblage zones (Fig. 2), which are defined by the quantitative distribution of benthonic foraminiferal species and major faunal parameters such as number of foraminifera per gram sediment, species diversity and planktonic foraminiferal content (Fig. 2).

Zone A

Of the few specimens found in this zone *Elphidium excavatum* f. cla-

Most of the published Quaternary stratigraphic work published from the Barents Sea has been based on seismic data and/or short gravity cores (for example, Elverhøi & Solheim 1983; Solheim & Kristoffersen 1984; Vorren et al. 1988, Vorren et al. 1989). The distribution and timing of marine-based ice sheets and the palaeoceanographic development of the Barents Sea shelf are still debated themes (Hughes et al. 1977; Boulton 1979; Matishov 1980; Elverhøi & Solheim 1983; Vorren & Kristoffersen 1986; Hald et al. 1989; Sættem et al. this issue). Continuous cores through the upper stratiform sequence (Sættem & Hamborg 1987; Vorren et al. 1988) in the Barents Sea would be of great importance in order to improve our understanding of long-term (10^4-10^5) climatic shifts in high northern latitudes.

The analysed interval is divided into five assemblage zones (Fig. 2), which are defined by the quantitative distribution of benthonic foraminiferal species and major faunal parameters such as number of foraminifera per gram sediment, species diversity and planktonic foraminiferal content (Fig. 2).

Zone A

Of the few specimens found in this zone *Elphidium excavatum* f. cla-
vata (Terquem) and *Cassidulina reniforme* (Nørvang) dominate. These species have their main distribution in glacially influenced shelf waters and/or other stressed waters characterized by rapid changes in salinity and temperature (Cushman 1948; Osterman 1984; Hald & Vorren 1987). The *E. excavatum-C. reniforme* fauna occurs very rarely in modern sediments, but has been reported from glacial marine deposits around the North Atlantic (e.g. Barents Sea, Hald et al. 1990; Canada, Vilks & Rashid 1976). Accessory species are inter alia *Cassidulina teretis* (Tappan) and *Islandiella norcrossi* (Cushman). This species composition can indicate a glacial marine environment with bottom water temperatures close to zero degrees Celsius. However, the presence of accessory taxa such as *Epistominella nipponica* (Kuwano) and *Bulimina marginata* (d'Orbigny) that are linked to interglacial (Boreal) environments may indicate that part of the sediment has been reworked.

Zone B

This zone is characterized by a marked influx of the Boreal species *B. marginata*. The zone shows an increase in foraminifera per gram sediment and in planktonic foraminifera. *B. marginata* is found today on the southern Norwegian continental shelf (van Weering & Qvale 1983) and is not common in the Barents Sea (Østbye & Nagy 1981; Hald & Steinsund, in print). Due to the presence of high arctic foraminifera, we suggest that the *B. marginata* have been reworked from an earlier interglacial. However, it cannot be totally excluded that this species is in situ.

Zone C

This zone is marked by an increase in foraminifera per gram sediment, planktonic foraminifera and a Boreal component of benthonic foraminifera dominated by *E. nipponica*, *Nonion barleeanum*
Fig. 1. Location of the shallow drilling and bathymetry of the investigated area.

Fig. 2. Foraminiferal stratigraphy of shallow drilling 7317/10-U-01 based on selected foraminiferal taxa and faunal parameters.
(Williamson) (Fig. 2), Pullenia bulloides (d’Orbigny), Trifarina angulosa (Williamson) among others (not shown in Fig. 2). This overwhelming dominance of Boreal foraminifera indicates quite clearly that this zone is of interglacial origin. The low number of E. excavatum f. clavata suggests that there is little or no reworking of the sediments. The foraminifera in this zone are well preserved and show no signs of corrosion, which can sometimes be seen in the two underlying zones. Both faunal and sedimentological evidence (Sættem et al., this volume) confirm this zone to be of in situ interglacial origin. Amino acid measurements, the ratio of D-alloisoleucine-L-isoleucine measured on E. excavatum f. clavata, B. marginata, and unidentified molluscs indicate part of Zone C to be of Eemian age (120 ka BP, isotope stage 5e). The fauna found in this zone was compared with the present fauna in the southwestern Barents Sea. From running transfer functions (Steinsund et al., this volume) it appeared that the water mass during the Eemian in the study area was possibly warmer than it is today.

Zone D

This zone is marked by a drastic decrease in the number of foraminifera per gram sediment and high content of Cretaceous species. However, high Arctic species such as E. excavatum f. clavata and C. reniforme are present. This mixed assemblage might indicate erosion of Cretaceous rocks, but other processes such as ice-rafting or reworking by winnowing/turbidite processes are also feasible.

Zone E

This zone is dominated by E. excavatum f. clavata and C. reniforme and can be interpreted as being of glacial marine origin. Zone E is very similar to Zone A, but contains a larger component of Boreal species and the cosmopolitan C. teretis.

Conclusions

(a) Zones A, B, D and E are of possible glacial marine origin and the assemblages seem to have been affected by reworking at different levels.

(b) Part of Zone C, which possibly represents the Eemian interglacial (isotope substage 5e), was deposited under conditions somewhat similar to, or maybe warmer than, the present southwestern Barents Sea. There is little evidence of reworking in Zone C and the unit represents the first direct indication of interglacial sediments, other than the Holocene, in the Barents Sea.

(c) The Boreal faunal component in Zone B is supposed to have been reworked from an earlier interglacial/interstadial. The species composition shows that the interglacial faunas might differ from one interglacial to another through the Quaternary in the southwestern Barents Sea.

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Notes on long-term forcing of Arctic climate

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Explanations for Cenozoic cooling of the Arctic tend to focus on proximal causes such as slow northward drift of circum-Arctic land masses or changes in meridional fluxes of heat and moisture through narrow gateways to the Arctic Ocean. Recent studies suggest that an important part of the forcing may lie in lower middle latitudes far from the Arctic. Uplift of massive plateaus in southern Asia and the American west appears to have fundamentally rearranged planetary-scale atmospheric circulation, including jet stream meanders and subtropical monsoons, with resulting climatic effects on all circum-Arctic lands. The onset and progressive intensification of strong monsoons in southeast Asia and elsewhere may have increased chemical weathering of freshly exposed rocks including silicates, causing a drawdown of atmospheric CO₂ and thus a global cooling.

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climate: (1) altered mean positions and increased amplitude of the planetary-scale (Rossby) waves; and (2) intensification of previously weak monsoonal circulations. Many long-term Cenozoic climatic trends deduced from a wide array of paleobotanical and geological evidence in specific regions agree with the model-simulated effects of uplift (Ruddiman & Kutzbach 1989). The uplift experiments particularly simulated the kind of regional differentiation of wet and dry climates observed in late-Cenozoic geologic data.

Although these experiments suggested fairly widespread cooling of high northern latitudes as a result of uplift, the amplitude of cooling away from the uplifted regions was modest, especially in the critical summer ablation season. The direct physical effects of uplift thus do not provide a convincing hemispheric-scale explanation for the pervasive long-term Cenozoic cooling. As noted earlier by Barron (1985), a decrease in CO2 still appears to be a necessary part of such an explanation.

Uplift may provide a key mechanism for long-term CO2 decline. Raymo et al. (1988) noted that tectonic deformation associated with broad-scale uplift exposes fresh bedrock on steepened plateau and mountain slopes and also intensifies strong seasonal monsoonal circulations that drench these slopes with rain. One result is to increase the rate of chemical weathering in these regions, as well as the runoff that flushes erosion products away. As a result, rivers issuing from high regions in Asia and elsewhere have the highest chemical dissolved loads in the world. Raymo et al. (1988) hypothesized that these uplift-related processes cause increased chemical weathering not just locally, but also in terms of the global mean. They noted that an increase in global mean weathering of silicate rock would cause a drawdown of atmospheric CO2. Berner et al. (1983) also included chemical weathering in their CO2 model, but did not include the strong dependence on elevation that is now justified by recent geochemical studies (Edmond 1987).

If these results prove to be correct, the primary cause of Arctic (and global) cooling in the late Cenozoic may be uplift of massive mid-latitude plateaus. These uplifting plateaus rearrange basic atmospheric circulation and produce numerous regional climatic effects, one of which includes stronger subtropical monsoons, increased chemical weathering, and decreased CO2 values in the atmosphere.

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Modern benthic foraminiferal distribution in the southwestern Barents Sea: paleo-oceanographic applications

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The distribution of total (living + dead) benthic foraminifera has been studied in 66 surface sediment samples from the southwestern Barents Sea. They respond to bottom-water conditions by correlating well with the summer bottom-water temperatures and the sediment texture, carbonate content and organic carbon. Based on factor analysis, four assemblage zones have been identified. The two easternmost assemblages are influenced by carbonate dissolution that may be a response to formation of corrosive, CO₂-rich bottom water. Based on multiple regression, equations relating the fauna to the bottom-water temperature within a temperature interval from 2 to 6°C have been calculated.

Factor analysis
We have divided the faunas into four assemblages zones, using Q-mode factor analysis (Fig. 1). These four factors explain 91.8% of the variance, and show communalities larger than 0.8 for all except five samples. Further information is given in Hald & Steinsund (in press).

Nonion assemblage zone
The Nonion assemblage zone, where Nonion barleeanum is the dominant species, occurs in pelitic sediments with a bottom-water temperature between 3 and 4.5°C. The assemblage zone is partly influenced by calcium carbonate dissolution. N. barleeanum may not be affected by this as much as other foraminifera, possibly because it is an infaunal species, and thus is protected from corrosive bottom water.

Trifarina/Cibicides assemblage zone
The Trifarina/Cibicides assemblage zone is dominated by the relatively large species Trifarina angulosa and Cibicides lobatulus. The predominant sediment is carbonate-rich gravelly sand. The CaCO₃ is mainly of biogenic origin, and suggests a low terrestrial sediment supply. The enrichment of the sediment in larger foraminifera is ascribed to the winnowing caused by the influx of Atlantic water. The bottom water has a temperature of between 5 and 5.5°C.

Reophax assemblage zone
The Reophax assemblage zone is dominated by agglutinating foraminifera, of which Reophax guttifera and Reophax atlantica are the most common species. The almost total lack of calcareous tests is due to calcium carbonate dissolution. The sediment is an organic-rich pelite. The assemblage zone is developed in the coldest part of the study area where the temperature range is between 2 and 4.5°C.

Epistominella assemblage zone
The Epistominella assemblage zone is found on the flanks of the banks and is dominated by the small Epistominella nipponica. This assemblage zone is enriched by small foraminifera that supposedly have been transported from the surrounding banks.

Carbonate dissolution
Our investigation has shown that carbonate dissolution is a characteristic phenomenon in the modern southwestern Barents Sea. The carbonate dissolution pattern seems to follow the distribution of dense bottom water. Formation of dense bottom water in the Barents Sea is connected to formation of sea ice and influx of Atlantic water. In the study area, increasing dominance of dense bottom water is found towards the east and north where there is sea ice (Fig. 1), as well as in the troughs (Midttun 1985; Anderson et al. 1988).

According to Anderson et al. (1988) the dense bottom water is
Estimating temperature

Using multiple regression analysis we have calculated an equation relating bottom-water summer temperature to the 13 most important benthic foraminiferal species, organic carbon, carbonate and pelite content. Planktonic foraminiferal temperature equations based on multivariate analysis (Imbrie & Kipp 1971) have been used with success in recent decades. An important difference between the planktonic and benthic foraminifera is that the benthic species are influenced by the substrate in addition to the ambient bottom water. We have therefore added the percentage values of organic carbon, carbonate and pelite to the regression analysis. An analysis without these three sedimentological parameters is much less statistically significant. With the inclusion of these sedimentological parameters the analysis shows a high significance ($R = 0.93$), the largest difference between the estimated temperature (based on benthic foraminifera) and measured temperature (based on temperature data from the Norwegian Oceanographic Data Centre) is 0.66°C, with a mean of 0.25°C (Fig. 2).

By adding assemblage zones from Svalbard, which were sampled in an area with temperatures down to -1.8°C, we have been able to expand the temperature range where the first equation can be used. Using a detailed foraminiferal stratigraphy from the southwestern Barents Sea by Hald et al. (1989), we calculated palaeotemperature through the last deglaciation and the Holocene. We modified our temperature equation for this purpose. The agglutinated foraminifera were excluded because of their low fossil

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**Fig. 1.** Distribution of recent benthic foraminiferal assemblage zones (= factors) in the southwestern Barents Sea. Each assemblage is plotted as a function of varimax factor matrix values equal or greater than 0.3. Dots show the location of the box-core samples of surface sediment. The index map shows surface waters (modified from Mosby 1968). (1) Arctic water, (2) mixing water, (3) Atlantic water, (4) coastal, Baltic, and North Sea water, (5) surface currents, (6) sea-ice border, April.

**Fig. 2.** Estimated versus observed summer bottom-water temperatures in the southwestern Barents Sea.
preservation potential. Thus the equation in this case is based on 10 calcareous benthic foraminiferal species which occur in both the stratigraphical and the modern data sets.

The calculated summer bottom temperatures (Fig. 3) show that there are large fluctuations in temperature before 10,000 years BP, and temperatures below 0°C before 11,800 years BP. The temperatures in the early Holocene are probably estimated too low due to reworking of glacial foraminifera.

Conclusions

Calcium carbonate dissolution in the Barents Sea seems to follow the distribution of dense bottom water. The dissolution can be used as a paleo-indicator for sea-ice formation and thus for influx of high salinity water.

The good correlation between temperature and benthic foraminifera together with sedimentological parameters shows that benthic foraminifera are suitable for temperature estimations.

References


Glacial geology of outer Bjørnøyrenna, western Barents Sea: preliminary results

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We present a local integrated borehole and seismic stratigraphic framework for the glacialic sediments in outer Bjørnøyrenna, southwestern Barents Sea, and discuss the depositional environment of the cored sediments. The lower part of the investigated succession may be Late Pliocene in age, and is partly dominated by shallow continental shelf sands with a suggested glaciofluvial sediment supply. The upper part of the outer shelf succession mainly comprises Middle and Upper Pleistocene muddy glacialic diamictons which we interpret to be mostly tills. Our stratigraphy suggests younger age and hence higher depositional rates for the upper part of the clastic wedge in front of Bjørnøyrenna than thought previously, and supports earlier suggestions of extensive Plio–Pleistocene glacialic erosion in the Barents Sea. A buried acoustically stratified seismic unit is interpreted from amino-, bio- and lithostratigraphy to include Eemian sediments. This unit has been overrun by a late Middle–Upper Weichselian grounded ice extending to the outermost shelf.

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Seismostratigraphic evidence suggests several episodes of glaciation of the Barents Sea, with grounded ice eroding its substratum all the way to the outer shelf (e.g. Elverhøi & Solheim 1983; Solheim & Kristoffersen 1984; Vorren et al. 1988), and a large clastic wedge in front of Bjørnøyrenna (e.g. Nansen 1904; Eldholm & Ewing 1971; Sundvor 1974; Spencer et al. 1984; Eidvin & Riis 1989; Vorren et al. 1990a, c) has recently been interpreted by Eidvin & Riis (1989) to comprise largely Upper Pliocene–Pleistocene glacialic sediments. The number of Quaternary glaciations and the depositional environments during the ice-free periods have hitherto largely remained unknown due to lack of control on age and lithology. Different views have existed regarding the existence as well as the extent of the Upper Weichselian ice sheet in the southwestern Barents Sea (Hoppe 1970; Schytt et al. 1967; Rokoengen et al. 1979; Boulton 1979a; Elverhøi & Solheim 1983; Solheim & Kristoffersen 1984; Vorren & Kristoffersen 1986; Vorren et al. 1989; Sættem 1990). A complex succession of late Middle–Upper Weichselian glacialic sediments with several internal erosional boundaries in the southern Barents Sea suggests, however, extensive glaciation in the area during the late Middle–Late Weichselian (Hald et al. 1990; Vorren et al. 1990a).

Our study utilizes IKU borehole core material from the glacialic sequence and shallow seismic data collected by various institutions since 1970, in an area straddling the northeastern boundary of the clastic wedge in front of Bjørnøyrenna (Fig. 1, Table 1). Here we give some preliminary results from the investigation of this material. Our aim is to provide new input to the reconstruction of glaciations and palaeoenvironments in the area and the development of the prograding clastic wedge in front of Bjørnøyrenna. The foraminiferan biostratigraphy of the cores utilized in this study is presented by Poole et al. (this volume) and Hald et al. (1990).

Stratigraphy

Seismic units

A brief overview of the architecture of the glacialic sequence in outer Bjørnøyrenna can be achieved from the seismostratigraphy given in Figs. 2 and 4, and the main relationships to earlier regional studies are given in Fig. 3.

Palaeocontinental shelf breaks are seen at the upper and lower boundaries of unit B in Fig. 2, and the westward dip of the unit boundary unconformities seen farther east is probably caused by compaction and subsidence of the clastic wedge. The ‘sawtooth’ reflection pattern within unit A0 (Fig. 2) is interpreted as large-scale sand structures (Sættem 1991), possibly formed on a shallow continental shelf with fluctuating sea level, grading eastwards into a coastal–non-marine environment. Large amounts of sediments were probably supplied to the shelf system through channels of possible glaciofluvial origin which are cut into this unit. These channels are partly re-excavated by glacial/glaciofluvial erosion in unit A time.

The seismic section in Fig. 4 illustrates different acoustic characters within the upper limits. The coherent internal stratification of unit E is not likely to develop under subglacial conditions, and may reflect deposition in a marine or glaciomarine environment (King & Fader 1986). The chaotic, hummocky part of unit G is interpreted by Sættem (in prep.) as buried glacioclastic cupola hills, possibly partly with a gradational transition into more homogeneous unit G till.

Lithology and depositional environments

Four main lithologies are distinguished:
1. Gravelly sand grading upwards into gravelly till (originating from unit A0 channel sands?). This sediment was cored in the lower part of the glacialic sequence in borehole 7316/06-U-01, located a few hundred metres beyond the seis-
mically mapped pinch out of a major unit A₀ channel. The lower gravelly sand is poorly sorted but has less than 5% silt and clay. The till seems partly to be derived from the underlying sand, but is intermixed with more clay- and silt-rich intervals. The sand may possibly correlate with the infill of the nearby unit A₀ channel. A primary glaciofluvial (subglacial?) origin of both the channel and the coarse sediment is suggested, but the till may originate from a later phase of reworking, related to formation of channels/depressions at the units A₀/A₀ boundary.

2. Muddy, greyish diamictons (seismic units B, D, F and G). Most of the cored sediments consist of silty sandy clay with scattered fine gravel (muddy diamictons), which typically is non-stratified. This sediment type contains dominantly Arctic foraminifera (Hald et al. 1990; Poole et al. this volume). Also a few, scat-
tered shell fragments were found. The minerogenic clast composition is strongly dominated by fragments of sedimentary rocks from the continental shelf. Overconsolidation of sediments was found in the northernmost boreholes, but also sediments apparently 'under-consolidated' compared to present burial depth are found. The stiff diamictons have a well-developed, often lenticular fissility, probably due to imposed shear from grounded ice. The lateral transitional boundary of the glaciotectonic structures within unit G (Fig. 4) inferred by Sættem (in prep.) may represent a boundary between two modes of subglacial deformation, and it is possible that most of the muddy, greyish diamictons on the shelf are finally deposited in a soft sediment subglacial traction layer environment as described by for example Boulton (1979b) and Alley et al. (1986, 1987). The fossils may represent earlier or intermittent periods of glaciomarine conditions (Hald et al. 1990). A component of final deposition by glaciomarine sedimentation, which is emphasized by Vorren et al. (1989, 1990a), is not excluded.

3. Bioturbated, partly stratified olivish sediments (mainly seismic unit E). The acoustically stratified seismic unit E (Fig. 4) was cored in borehole 7317/10-U-01. The lithology of this unit consists of soft strongly bioturbated or laminated silty, olivish sandy clay with only scattered gravel (lithozone L4 and foraminiferal assemblage zones C and D of Poole et al. this issue)). In one of the samples from assemblage zone C the gravel was dominated by pyritized burrows, and the lithology and fauna suggest a marine interglacial environment with only occasional ice-rafting. A sheared and folded, partly stratified, bioturbated sediment recovered from deeper in this borehole (lithozone L2 of Poole et al. this issue) may represent another marine or glaciomarine event possibly correlating with the seismic units D1/D2 boundary.

Both these intervals are 'under-consolidated' compared to present burial depth. The low consolidation probably indicates rapid deposition of overlying sediments, possibly under confined drainage conditions.

4. Allochthonous bedrock rafts. In borehole 7316/06-U-01 cores were obtained of rafts of Cretaceous sedimentary rocks embedded in glaciogenic sediments. The rafts cover at least 15 m of the borehole and are interpreted by Sættem (in prep.) as part of buried glaciotectonic cupola hills.

Age

Preliminary correlations of borehole aminosteratigraphy and seismic units

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**Table 1. Overview of shallow drilling boreholes and sample recovery from the glacigenic sediment sequence in the study area. Most of the boreholes penetrate into underlying preglacial sediments, but only the glacigenic interval is included in the table.**

<table>
<thead>
<tr>
<th>Site</th>
<th>7316/03-U-01</th>
<th>7316/06-U-01</th>
<th>7316/06-U-02</th>
<th>7317/10-U-01</th>
<th>7317/02-U-01</th>
<th>7317/02-U-02</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude (N)</td>
<td>73°52'55.8&quot;</td>
<td>73°33'16.5&quot;</td>
<td>73°34'7.7&quot;</td>
<td>73°8'58.5&quot;</td>
<td>73°49'10.92&quot;</td>
<td>73°47'08.06&quot;</td>
</tr>
<tr>
<td>Longitude (E)</td>
<td>16°54'28.9&quot;</td>
<td>16°49'59.6&quot;</td>
<td>16°50'0.9&quot;</td>
<td>17°16'13.0&quot;</td>
<td>17°22'04.53&quot;</td>
<td>17°23'35.51&quot;</td>
</tr>
<tr>
<td>Water depth (m)</td>
<td>297</td>
<td>428</td>
<td>421</td>
<td>465</td>
<td>297</td>
<td>324</td>
</tr>
<tr>
<td>Glacigenic sediment thickness (m)</td>
<td>31.9</td>
<td>65.0</td>
<td>68.5</td>
<td>143.2</td>
<td>45.69</td>
<td>-50</td>
</tr>
<tr>
<td>Cores (no.)</td>
<td>3</td>
<td>11</td>
<td>11</td>
<td>35</td>
<td>7</td>
<td>1</td>
</tr>
<tr>
<td>Total core length (m)</td>
<td>0.62</td>
<td>13.67</td>
<td>2.08</td>
<td>12.98</td>
<td>2.35</td>
<td>0.47</td>
</tr>
<tr>
<td>Core recovery (%)</td>
<td>1.9</td>
<td>21</td>
<td>3.0</td>
<td>9.1</td>
<td>5.1</td>
<td>1</td>
</tr>
</tbody>
</table>

---

**Fig. 2. East-west profile along 73°30′ in the northern flank of Bjarøyrenna showing the prograding top of the clastic wedge in front of Bjarøyrenna (partly modified from Vorren et al. 1990). The different seismic units are informally labeled with capital letters. The ages of units A0 and A are based on correlation with the stratigraphy of Eidvin & Riis (1989). Our dating of units B and younger is based on aminosteratigraphy, mainly correlated from borehole 7317/10-U-01. Magnetopolarity confirms that all sediments in this borehole were deposited during the Brunhes normal polarity chron.**
This work, Eidvin & Riis (1989) (Figs. 2, 3) suggest that units B and younger, which correspond to Vorren et al.'s (1990b, c) unit TeE, may be less than about 0.44 m.y. old. The ratios measured on foraminifera and mollusc fragments obtained from the marine sediments correlating with the lower part of the acoustically stratified unit E are comparable with Eemian values (e.g. Sejrup et al. 1989; Miller & Mangerud 1985). This, combined with the biostratigraphy (Poole et al. this issue) and lithology, suggests that the Eemian interglacial (117–130 ka, Mangerud 1989) may be represented within the lower part of this unit. The amino acid ratios further indicate a late Middle–Late Weichselian age for the lower part of the overlying units F and G, including the glaciotectonic cupola hills (Fig. 4). This would imply that grounded ice has extended to the shelf break in the deep outer Bjørnøyrenna during that time.

The base unit A₀ unconformity
corresponds to the base of the Plio-Pleistocene wedge as outlined by Eidvin & Riis (1989, fig. 2) and the ages assigned to units A₀ and A are based on Eidvin and Riis's (1989) datings.

Conclusions

Preliminary data on lithology, aminostratigraphy and seismostratigraphy from the glaciogenic succession in outer Bjoernoyrena, southwestern Barents Sea, provide new information about the glacial history of the area and about the development of the clastic wedge deposited in front of Bjoernoyrena. A buried acoustically stratified seismic unit is interpreted from amino-, bio- and lithostratigraphy to include Eemian marine sediments. This unit has been overrun by a late Middle–Upper Weichselian grounded ice extending to the outermost shelf. An older interval containing possible teconitzed marine sediments is separated from the Eemian sediments by a diamicton which we interpret as a till. With the exception of these two intervals we suggest that most of the sediments deposited on the palaeoshelves of our seismic units B and younger were deposited as tills by Middle and Upper Pleistocene ice sheets extending to the shelf break.

Our seismic unit A₀ is partly dominated by shallow continental shelf sands with a suggested glacialfluvial sediment supply. We correlate the unconformity at the base of this unit with the base of the Pliocene–Pleistocene clastic wedge described by Eidvin & Riis (1989), and our data indicate a dominating glaciogenic environment in the Barents Sea since unit A₀ time. Our results further suggest that the part of the prograding clastic wedge inferred by Vorren et al. (1989c) to originate from about 150 m of glacial bedrock erosion in the southwestern Barents Sea (Vorren et al. 1990c, table 4) has been deposited since about 0.44 Ma. This may imply that a more elevated bedrock surface possibly forming extensive, partly ice-covered land areas may have existed in the Barents Sea region during the Pliocene and possibly part of the Pleistocene.

Acknowledgements. – The present study is a cooperative research effort between IKU and the Universities of Bergen, Trondheim and Tromsø. The borehole cores and part of the shallow seismic data were collected within two IKU Barents Sea Mapping Projects in 1985 and 1989, financed by 13 oil companies and the Norwegian Petroleum Directorate (NPD), and funding from Mobil Exploration Norway inc., Norsk Hydro, Statoil and the owner of the drillship 'M/S Bucentaur' allowed improved sample coverage from the glaciogenic sequence during the IKU drilling campaign in 1989. The remaining shallow seismic data were collected by the Seismological Observatory, University of Bergen, the Norwegian Polar Research Institute, the Norwegian Petroleum Directorate, and the University of Tromsø. The English text was corrected by Stephen Lippard and the figures were drawn by Anne Irene Johansen, Berit Fossum and Jan Helge Johansen. To all the above institutions, companies and people we extended our sincere thanks.

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Analyses of climatological series in relation to the detection of the CO₂ effect

ARNOLDUS JOHANNES COOPS


A method has been developed for analysing climatological series based on the assumption that climate undergoes abrupt changes by natural means. A gradual climatic change due to the increasing CO₂ concentration in the atmosphere is superimposed upon the natural abrupt changes. The method has been applied to both European summer and winter temperatures and to the precipitation and air pressure series for De Bilt. The method has been combined with another one that reduces noise to a minimum. These methods are applied to spatial combinations of temperature as well as to combinations of temperature with precipitation and air pressure. The results give a survey of climatic changes during the last 120 years; although the presence of a CO₂ effect has not yet been demonstrated convincingly, the methods of analysis may prove useful for CO₂ detection in the future.

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Methodology

**Analysis of climatological series**

Climatic changes may occur rather abruptly by natural means. This idea originated from an analysis of the series of cumulative deviations from the long-term average value of the summer temperatures from the year 1706 (De Bilt, The Netherlands). This series has a zig-zag shape, characterized by relative extremes that are connected by straight lines rather than, for example, by parts of a sine wave. Since the cumulative series can be considered to be a primitive function of the original series, the relative extremes can be conceived as change points of the climate. Hence, a climatological series can be partitioned into subperiods with on average low and high values alternately. In order to establish the significance of each jump in climate, the values in two successive subperiods can be compared, for instance by means of the two-sample Wilcoxon test statistic. This is a non-parametric method, that is to say, it is distribution-free. The problem here is that the test statistic is calculated with the same data upon which the positions of the relative extremes of the cumulative series were based. This selection effect has been taken into account by Pettitt (1979) for the case where no more than one jump may occur in a series. The position of a change-point is then fixed by the value of \( k \), which defines the maximum \( W_2 \) of the test statistic of Wilcoxon and the significance level \( P \) is given by an (over)estimate:

\[
W_2 = \max_{2 \leq k \leq N-2} \left\{ \sum_{j=1}^{k} \sum_{i=k+1}^{N} \text{sign}(X_i - X_j) \right\}
\]

\[
P \leq \exp\left\{ -\frac{6W_2^2}{N^3(N+1)} \right\}
\]

\((N \) is the length of the series \((X_i)\) and \(W_2\) the statistic used). A generalization of this method as developed by the present author involves optimizing the position of each point successively, according to Pettitt, and calculate the probability \( P \) that the value of the test statistic appears by chance, that is to say, occurs under the null hypothesis that climate did not actually change.

1. Choose a number of \( M \) equidistant points; optimize the position of each point successively, according to Pettitt, and calculate the probability \( P \) that the value of the test statistic appears by chance, that is to say, occurs under the null hypothesis that climate did not actually change.

2. Compare the maximum \( P_{max} \) of the \( M \) values of \( P \) with the significance level \( 1/M \) (this level was chosen in such a way that ultimately one jump may occur by chance, on the assumption that the change-points are independent of each other, which is not exactly the case): if \( P_{max} < 1/M \), the process is finished; otherwise the jump with \( P_{max} \) is cancelled, together with the adjacent one with the highest \( P \) value; thus the block structure is retained.

3. Repeat the procedure described above, replacing \( 1/M \) by \( 1/(M-2) \) and optimize the positions of the remaining points. In this manner an iteration process is started, which always converges to a solution with at least one jump. (Details of this method are given in the Appendix.)

**Detection of the CO₂ effect**

As a result of the increasing CO₂ concentration in the atmosphere, a gradual climatic change is to be expected. This change is superimposed upon the jump-like behaviour that is present by natural means. Jumps in the direction of a climate alteration owing to the rising CO₂ content should be stronger than those in the opposite direction (level behaviour) and than preceding ones in the same direction in the case of a non-linear trend (signal behaviour). A third criterion is the behaviour of the signal-to-noise ratio, which is better if noise is allowed to change due to the CO₂ increase; here noise is defined as the average of the interannual variability (which is a measure for the derivative)

\[
l_j = \left| X_{j+1} - X_j \right|
\]
If variations in the length of the sub-periods are also involved the significance behaviour is even more relevant.

The various criteria were not fulfilled by the summer temperatures of De Bilt. (This series was chosen in the first instance because of the expected favourable CO₂ signal-to-noise ratio at moderate latitudes in summer, according to Wigley & Jones (1981)). Possible causes are: (1) De Bilt is not representative and/or the series is unreliable. (2) The effects of volcanism, which might be greatest in summer, have been neglected. (3) Temperature is not the best indicator of CO₂-induced climatic changes. Other variables, such as precipitation and air pressure, may be more sensitive to CO₂ changes. (4) The noise of the series is as yet too large with respect to the present-day CO₂ signal.

Solutions to these problems were sought by the following procedures. 1. Analysing summer temperature series at other places in Europe.

2. Analysing winter temperature series in Europe. This may be interesting with respect to a presumed temperature decrease in Central Europe in winter owing to a possible CO₂-induced rise in the frequency of blocking, according to Palutikof et al. (1984).

3. Analysing the precipitation and air pressure series of De Bilt, for summer and winter. In summer, precipitation is negatively correlated with temperature, in winter, positively; the opposite holds for air pressure.

4. Reducing noise by means of a linear combination of variables that are sensitive to CO₂ changes (in connection with a possible northward shift in summer of the Azores high): air pressure and precipitation which exhibit analogous jumps. The optimum combination is a weighted average whose coefficients are chosen in such a way that noise is minimized by applying the theorem on the multipliers of Lagrange (analogous to the maximization of the signal-to-noise ratio according to Bell (1982)). This method can also be applied to spatial combinations of temperature. However, seasonal temperatures are highly correlated regionally, which leads to a small reduction in variance; opposite CO₂ effects are possible globally (which is not favourable for CO₂ detection either, in combination with an inhomogeneous station distribution). When the new method was applied to an a priori combination (based on the mutual correlations only), the significance levels in the stop criterion of the iteration process were raised by a factor 2, on the assumption that the probabilities might next be halved, if differences in the signal-to-noise ratios of the separate variables in the combination are also taken into account. In the cases where temperature was combined with precipitation and air pressure this assumption was not fulfilled. In this situation the original level was maintained, which nevertheless led to a halving of the original probabilities.

Results
As far as summers are concerned, the most remarkable and coherent picture emerging from Fig. 1A is the warm period 1930–50, which
appears almost everywhere. It seems that in the course of a few years the mean position of the boundary between the region with 'fine' weather (on average) and that with 'bad' weather shifted from the south to the north and then in a few years it came from the north to the south. Besides, this period appears at the coastal stations a few years later than at the continental stations, presumably because of the presence of the ocean. The average number of jumps increases from \(3\frac{1}{2}\) in the separate variables to 7 in the combinations.

Winter temperatures are characterized by rises at the end of the last century and at the beginning of this century, and around 1950. Furthermore, it is striking that in Fig. 1B the 1970s started with some six or seven mild winters, but by the end of the 1970s winters had become colder again, starting from the north. When temperatures are compared at the beginnings and ends of the various series, a rising tendency end of the 1970s winters had become colder again, starting from the north. When temperatures are compared at the beginnings and ends of the various series, a rising tendency

The agreement between temperature \((T)\), precipitation \((Pr)\) and air pressure \((Ap)\) in De Bilt in this century is not so clear in Fig. 1C, because the correlations \((\tau)\) between the variables are lower (in an absolute sense) than the spatial correlations in temperature in summer \((S)\) and winter \((W)\): \(\tau_s(T, Pr) = -0.17; \tau_s(T, Ap) = 0.12; \tau_s(Pr, Ap) = -0.47; \tau_s(T, Pr) = 0.29; \tau_s(T, Ap) = -0.02; \tau_s(Pr, Ap) = -0.34\). Precipitation and air pressure do not satisfy all criteria respectively (except the periods in the series of summer temperatures for southern Europe), but not very significant.

Conclusions

The method presented here makes it possible to distinguish between jumps and trends in a time series. Application of this method in combination with the one developed by Bell leads to the following conclusions.

The behaviour of the European summer and winter climate can be described mainly in terms of jumps. The two seasons behave differently. The strongest rises in temperature did not happen in the recent past (in winter around 1910, in summer about 1930). With regard to the criteria used, it should be emphasized that the assumption of \(CO_2\) warming is generally less fulfilled the stricter the criterion. In this century the summer climate in southern Europe satisfies the assumption best. This statement links up with the expectation formulated by Wigley & Jones (1981). A \(CO_2\)-induced drop in winter temperature as suggested by Palutikof et al. (1984) might be inferred from the downward jump at the end of the 1970s, after a period of (very) mild winters since the beginning of that decennium.

Hardly any trends could be detected within the subperiods and spurious trends could be attributed to the fact that jumps had not been detected. Differences in the variability of successive subperiods were mostly not significant. These statements justify the application of the method developed here. The results do not demonstrate convincingly that there has been a \(CO_2\) effect upon the European climate so far, nor do they demonstrate that the increasing \(CO_2\) content in the atmosphere has not had an influence upon the European climate.
Reliability, sensitivity, accuracy

In the case of white noise (no jumps) the method yields two jumps, with probabilities of more than 50% for both. However, they appear in the middle of the series instead of at the ends, and, hence, they are not realistic. For up to six jumps, inclusive, the number of jumps detected ($N_2$) exceeds the real number ($N_1$) by a value varying from 0 to 2. If $N_1$ is larger than 0, only spurious jumps are found at the ends of the series. When more than six jumps are induced, the resolution of the method diminishes ($N_2 < N_1$), since the lengths of the subseries decrease and the probabilities rise accordingly (despite a theoretically constant signal-to-noise ratio), while the stop criterion becomes stronger with increasing $N_2$.

When induced jumps are detected, their positions are accurate in almost half the cases. The (absolute) timing error varies from 0 to 6 and averages out at less than 1. A positive relation between the accuracy and the signal-to-noise ratio of the jumps is found.

Autocorrelations

In our search for an explanation of the jump-like behaviour of climate we also study the influence of persistence. Up till now, the temperature series have been assumed to consist of independent variables. However, in the winter and summer temperatures of De Bilt for instance, the lag-1 autocorrelations $r_1$ amount to +0.11 and +0.12, respectively (during the period 1869–1988). The original random series has a negative persistence ($r_1 = -0.22$), which is removed by the transformation

$$Y_i = X_i & Y_i = \frac{X_i - r_1X_{i-1}}{\sqrt{1 - r_1^2}} \quad (i = 2, 3, \ldots, 120). \quad (5)$$

In that case a third jump is found, but not at an end. Next, various positive autocorrelations $\rho$ are introduced:

$$Z_i = Y_i & Z_i = \rho Z_{i-1} + \sqrt{(1 - \rho^2)} Y_i, \quad (6)$$

with $Y_i$ standard normally distributed. The number of jumps detected ($N_2$) increases gradually with increasing $r_1$ until $r_1 \approx 0.5$; thereafter $N_2$ decreases slightly, despite a continuing rise in the significances. These results are summarized as follows:

$$\begin{align*}
N_1 & \quad r_1 \quad 0.22 \quad 0.0 \quad 0.1 \quad 0.2 \quad 0.3 \quad 0.4 \\
N_2 & \quad 2 \quad 3 \quad 4 \quad 5 \quad 6 \quad 7 \quad or \quad 8 \\
& \quad 0.5 \quad 0.6 \quad 0.7 \\
& \quad 10 \quad 7 \quad or \quad 8 \quad 6 
\end{align*}$$

Comparing these results with the number of jumps in the series of De Bilt (4 in winter, 9 in summer), we conclude that the jump-like behaviour in winter might be associated with the weak persistence, according to the Markov process considered above. For summer, however, autocorrelation can be regarded as no more than a partial explanation for the jump-like behaviour.

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Peculiarities of polar ozone annual course: analysis of satellite and ozonesonde data and model results

ALEXANDER N. GRUZDEV, IGOR L. KAROL, ALEXANDER P. KUDRYAVTSEV, IGOR I. MOKHOV & SERGEY A. SITNOV

Analysis of satellite data and model results

There have been many climatological studies of global ozone vertical distribution (e.g. Wilcox et al. 1977; Dutsch 1978; Bojkov 1987), some of which have been based on a harmonic analysis of ozone annual variations. The following analysis of satellite ozone data (Keating & Young 1985) and the results of two-dimensional photochemical modelling (Gruzdev et al. 1988) are based on a method of amplitude-phase characteristics (Mokhov 1985; Gruzdev & Mokhov 1988). Similarly to Mokhov (1985) and Gruzdev & Mokhov (1988), height-altitude annual dynamics of ozone are determined: (1) using phase characteristics – by means of reaching the local annual mean values with positive (0-phase, or phase of increase) and negative (π-phase, or phase of decrease) time derivatives and also by means of reaching the local annual maxima and minima; (2) using amplitude characteristics – by moving the boundaries (isochrones) of the areas with ozone changes by a certain value relative to non-uniform distribution of certain months.

Fig. 1 shows the isochrone boundaries (integers correspond to the mid-month) of areas with ozone-mixing ratio increase (full curves) and decrease (broken curves) by 1 ppmv against March (a) and September (b) regimes, during a successive six months' satellite data set. The amplitude dynamics exhibits the opposite tendencies of ozone annual variations in the upper (40–50 km) and middle (25–40 km) stratosphere with the division boundary (DB) at the 3–4 mb level (40 km) in different seasons. This also includes changes relative to December and June (not shown). In the southern hemisphere (SH), opposite tendencies are observed during all seasons, but in the northern hemisphere (NH) they are only observed relative to the winter and summer regimes. At polar latitudes, the evolution of ozone amplitude characteristics displays mainly dynamics of divergent type: the boundaries of areas with ozone increase or decrease extend, as a rule, from the polar to moderate latitudes. However, there is a difference between the evolution of increased and decreased isochrones: some areas of decrease in Fig. 1 have divergent type dynamics of expansion from stratospheric domains at moderate latitudes.

Fig. 2 shows the latitude-altitude dynamics of 0- and π-phases and phases of ozone maximum and minimum. In areas with significant semiannual harmonics there are two families of phase boundaries shown by full (1st family) and broken (2nd family) curves. According to Fig. 2A, the earliest reaching of the 0-phase (for 2nd family curves) occurs in the Arctic stratosphere, at the 15 mb level, where 0-phase isochrones have divergent latitude-altitude dynamics, and near 70°N in the neighbourhood of the stratosphere from which the isochrones descend to the DB. By March, from the tropical and Arctic phase boundaries, the common 0-phase isochrone is formed, exhibiting peculiarities of the lower part of the NH middle stratosphere from the equator to the middle latitudes, with maximum delay in reaching the 0-phase in NH spring. The latitude-altitude dynamics of the 1st family of the 0-phase isochrone reveals essential differences between NH and SH upper stratospheric ozone evolution. The effect of phase rotation of reverberator type (spiral structure) occurs near the 0.7 mb level at 30°N, which also influences the 0-phase isochrone evolution at the Arctic latitudes. A spiral-type feature is also noted in the vicinity of the DB in the SH.

For π-phase dynamics (Fig. 2B) the feature of divergent spreading from the polar latitudes is displayed in the NH middle stratosphere (the spreading is slower compared with that of the 0-phase in the same domain). A substantial difference can be noted between the 0- and π-phase dynamics in the upper stratosphere. Particularly in the SH, the
0-phase boundary spreads horizontally from the high latitudes while the $\pi$-phase boundary spreads vertically from the upper to middle stratosphere.

From February to May in the NH middle stratosphere, the slow spreading of successive maximum-phase isochrones from the polar to middle latitudes takes place (Fig. 2C). In the NH upper stratosphere, isochrones spread from the tropical stratopause to the polar DB (in December–March). During October–November, successive isochrones move in the opposite direction from the DB and from the polar latitudes. Outside the tropical latitudes of the SH upper stratosphere, the boundary of ozone maximum spreads quickly in June from the sub-Antarctic stratopause, and turns to the equator near the DB. Fully formed by the end of July, the successive isochrones, similar to those descending over the tropics, penetrate into the tropical middle stratosphere in August and spread to the Antarctic latitudes. During October–November, an abrupt descent of isochrones (of injection type) takes place from the DB to the lower part of the middle stratosphere, at the 60th SH latitudes, with horizontal components in the Antarctic direction. This charac-

Fig. 1. Successive isochrone boundaries of areas with ozone mixing ratio increase (full isochrones) and decrease (broken isochrones) by 1 ppmv against March (A) and September (B) regimes according to satellite data. Integers correspond to the mid-month.

Fig. 2. Isochrone boundaries of 0-phase (A), $\pi$-phase (B), maximum phase (C) and minimum phase (D) of stratospheric ozone mixing ratio as derived from satellite data.
terizes the process of filling up the Antarctic ozone 'hole'.

In phase dynamics of minimum ozone values (Fig. 2D) the spreading of appropriate isochrones from the tropical to polar latitudes is displayed from April to July in the NH upper stratosphere, while during November–December in the upper and middle stratosphere the spreading occurs from polar to middle latitudes. In the SH upper stratosphere, the boundary of minimum values spreads during October–December from the tropical to Antarctic latitudes, while in the SH middle stratosphere, isochrones spread slowly from the polar to sub-Antarctic latitudes. The essential differences between the dynamics of maximum and minimum ozone values are displayed in the SH.

Fig. 3A, B shows the latitude-altitude dynamics of 0- and π-phases of ozone concentration annual course based on the results of two-dimensional photochemical modelling. The descent of 0- and π-phase isochrones is displayed in the middle and lower polar stratosphere. In the SH lower stratosphere these isochrones spread simultaneously from the moderate to Antarctic latitudes. By October, the boundary of the π-phase reaches the SH tropopause. The boundary of the 0-phase reaches the 22 km altitude at the 60th SH latitudes by October, and reaches the tropopause level by March–April. Unlike the SH stratosphere, the phase dynamics of NH polar ozone does not, on the whole, display the same latitudinal delay. In the Arctic lower stratosphere the 0-phase isochrones spread above 20 km from the polar latitudes but a tendency to spread to the North Pole is displayed below 16 km. The 0- and π-phase isochrones reach the tropopause level, respectively, by October and May.

Fig. 3C shows the evolution of areas of ozone increase (on the shaded side of the isochrones) and decrease (on the non-shaded side of isochrones) against December (C) and June (D) as derived from two-dimensional photochemical simulation. Connected points indicate the tropopause.
decrease (on the non-shaded side of the isochrones) with respect to the December regime. The opposite tendencies toward ozone changes in the middle and lower stratosphere are displayed during winter–spring and summer–autumn (relative to June) seasons in the SH, and only in the spring season (Fig. 3C) in the NH. The isochrones spread downwards at subpolar regions, along with the simultaneous displacement of isochrones toward the South Pole. In the NH middle stratosphere (above 20 km), the boundary of ozone increase spreads during January–February from the polar to moderate latitudes. During summer–autumn, in the NH only, an ozone increase is noted in the lower stratosphere, with divergent-type isochrone dynamics at moderate latitudes (Fig. 3D). A qualitative correspondence can be observed between ozone annual course peculiarities through both the satellite data and model calculations, although essential quantitative differences exist.

Analysis of ozonesonde data

The preceding analysis of the ozone vertical distribution field can be supplemented by an analysis of ozone sounding data over two polar stations: Resolute (74.43°N, 94.59°W) for 12 years (1975–86) and Amundsen–Scott (89.59°S, 24.48°W) for 7 years (1967–71, 1987–87). Figs. 4 and 5 show the time–altitude cross-sections for ozone concentration month-to-month changes above Resolute and Amundsen–Scott (A). Cross-sections are also shown for deviations of the ozone monthly means from the long-term annual means (B), calculated on the basis of long-term monthly means.

The annual ozone evolution above Resolute reveals certain features in the boundary layer, in the troposphere above the boundary layer, and in the stratosphere. Various phases of the annual course (extreme values, local annual mean values) are reached first in the stratosphere and in the surface layer; in the mid-troposphere the respective regimes occur 3 to 6 months later (see zero isolines). The spring stratospheric ozone decrease (from March to July) is largely concentrated at 7–22 km, originating in the mid-stratosphere and gradually penetrating into the upper troposphere (Fig. 4). The autumn ozone increase (November–December) also occurs in a downward direction, but originates in higher layers of the stratosphere (30 km). In addition, it can be observed clearly that below the ozone maximum during October–November (Fig. 4A), there is an additional height and time-localized centre of ozone increase. Between December and January, a sharp decrease in ozone content can be observed above Resolute above 22 km. This decrease is accompanied by a drastic rise in temperature and a reversal of zonal wind in this layer of the atmosphere, most likely due to sudden stratospheric warmings (Gruzdev & Sitnov 1989). The line of advance of ozone increase and decrease from the stratosphere to the troposphere above Resolute (Fig. 4A) indicates stratospheric–tropospheric exchange in the northern polar region. The content and time regimes of ozone in the surface and boundary layers of the Arctic atmosphere are largely controlled by photochemical and synoptical processes (Kelley 1973; Barrie et al. 1988). The delay in reaching the x-phase regime in the middle troposphere (as much as 6 months) is quite long, compared to that in the surface layer (Fig. 4B). This is likely to be associated with the spring ozone destruction induced by the Arctic surface temperature inversion, which occurs as the sun rises (Barrie et al. 1988).

Ozone annual variation above the South Pole is characterized by isolated evolution in some atmospheric layers, including the maximum ozone layer and the near surface layer (Fig. 5A, B). So, from March to June there is an ozone increase in the wide layer of the middle stratosphere, while outside this layer the ozone decreases (Fig. 5A). The most rapid increase takes place in the neighbourhood of the maximum ozone layer in April–May. From July to October there is a reduction.

Fig. 4. Monthly ozone concentration changes (A) and ozone concentration deviations from local annual mean value (B) by ozonesonde data above Resolute (units: $10^{12}$ mol cm$^{-3}$). Upper dotted-broken curves indicate the maximum ozone level and lower ones indicate the tropopause.
in the low and middle stratosphere ozone concentration most pronounced in the maximum ozone layer again, which is known as the Antarctic ozone 'hole' (Solomon 1988). The spring ozone increase following the spring decrease occurs throughout the entire sounded layer of the stratosphere. The rise in the altitude of ozone maximum (see the upper (dotted) broken curves in Fig. 5A, B) indicates that ozone has been advected from the lower latitudes. It should be noted that although the spring ozone deficit above Antarctica (the ozone 'hole') manifested itself in the 1980s (Solomon 1988), a notable decrease in stratospheric ozone above the South Pole also occurred in 1971 (Fig. 5C). The essential differences seen in Fig. 5A, C are: first, the ozone decrease in 1971 was displayed approximately one-and-a-half months earlier, and, second, the greatest decrease in 1971 was noted considerably above the maximum ozone layer.

Tropospheric ozone above the South Pole is characterized in particular by the contrary annual variations of the near surface and upper troposphere ozone (Fig. 5A, B).

### Conclusion

This analysis reveals certain features in the polar ozone annual course at different atmospheric levels. It is highly significant that these features are not the same for the different seasons, a conclusion that cannot be reached through carrying out standard harmonic analysis. In some respects, the ozone evolutions in the Arctic and Antarctic atmospheres are similar, but they can also manifest fundamentally different behaviour. In particular, one can conclude that the ozone 'hole'-type effect is more peculiar to the Antarctic than to the Arctic. As shown by the direction of latitudinal displacement of different isochrones in Figs. 1–3, similar ozone regimes are reached very late in Antarctica, relative to neighbouring latitudes, due to the relative dynamical isolation of the Antarctic atmosphere.

### References


Stable ozone layer at Tromsø

KJELL HENRIKSEN, TROND SVENØE & SØREN H. H. LARSEN


Long-term measurements of the thickness of the ozone layer, carried out in Tromsø since 1935, show that the annual mean value can vary by as much as 20%. However, no decreasing trend and no correlation with the increasing amounts of chlorine and halogen compounds in the atmosphere occur in the Scandinavian sector of the Arctic.

At the Auroral Observatory, Tromsø, measurements of the thickness of the ozone layer were carried out from 1935 to 1969, and these have been normalized by Bojkov (1988). Unfortunately, there was a break in the recordings from 1969 to 1984.

A review of the measurements is given in Fig. 1, where the mean value of each year is given. For some years, however, the data are incomplete, lacking measurements for two or three months. Therefore no mean values are given for these years. Using years with complete data sets, the overall mean value of the measurements is calculated. This mean value is 341 Dobson Units, corresponding to a mean thickness of 3.41 mm.

During the period 1935–69 the emission of Freon gases was minor, and the ozone layer was not affected by these anthropogenic ozone-destroying gases. The measurements of this period are used to calculate the mean annual, undisturbed variation in the thickness of the ozone layer (see Fig. 2) and this variation is indicated by the thick continuous curve. The amount of data is sufficient to calculate the uncertainty of the annual variation, which is plotted by the dashed curve on both sides of the thick curve.

Also for the last five years, 1984–89, the mean annual variation is calculated and is indicated in Fig. 2 by the thin continuous curve. Comparing the annual variations of these two periods, it can be concluded that the thickness of the ozone layer of the latter period has not decreased in average. For the summer months the ozone layer is significantly thicker because the measurements of the last five years stay above the mean value plus the uncertainty of the period 1935–69. It is only in the summer months that the ozone layer is most important as protection against dangerous solar ultraviolet radiation.

The Auroral Observatory has obtained the second longest series of ozone measurements in the world. During recent years several ozone measuring stations have been established in the neighbouring parts of the Arctic, in Sodankylä, Murmansk, Longyearbyen, Ny-Alesund, and Franz Josef’s Land. With some tolerance of measurements and systematic deviations the same ozone thickness and variations are obtained at all these stations. These results were documented by

Fig. 1. Averaged annual ozone values connected by straight lines for the period 1935–89 in Tromsø. The mean ozone thickness is 341 DU or 3.41 mm.
Annual Variation of Ozone in Tromsø

Fig. 2. Annual variations in the thickness of the ozone layer in Tromsø. The thick continuous curve represents the period 1935-69, and the thin curve the period 1984-89.

...the Finnish scientists Taalas & Kyrö (1990).

The world's longest series of ozone measurements derives from Arosa, Switzerland. This series shows that the mean thickness of the ozone layer has increased by 2% from 1926 to the beginning of the 1950s. Since then the thickness has decreased by 4% on average (Dütsch & Staehlin 1989).

The world's longest series of ozone measurements therefore indicates that the long-term ozone variations are different in different geographic regions. Today it is too early to claim that anthropogenic ozone-destroying gases have caused a global ozone decrease. This view is mostly accepted, but it was criticized at the conference because the existing data do not show a clear trend. A global ozone decrease, however, is supported by numerous prognoses and model calculations claiming a considerable ozone reduction due to emission of Freon gases. The measurements from Tromsø, on the other hand, indicate that the thickness of the ozone layer is not affected by the increasing amounts of Freon gases in the atmosphere.

This paper is an extended abstract of a paper presented at the climate meeting in Tromsø, 2-4 April 1990, and was completed on 1 May 1990. The full paper has been submitted to the Journal of Atmospheric and Terrestrial Physics (Henriksen et al. 1990).

References


Global, spectral ultraviolet and visible measurements in the Arctic

KJELL HENRIKSEN, STEFAN CLAES, TROND SVENØE & KNUT STAMNES


Spectral measurements in the region from 2900 to 4500 Å were carried out during the midnight-sun period. Meteorological conditions have a stronger effect on the irradiances than the ozone layer. The maximum absolute value of ultraviolet B, however, is about 20% of the value at the Equator. Data on this irradiance are scarce even though the threat of a decreasing ozone layer and the subsequent increasing UV irradiance have been of major public concern for several years. These campaigns are continuations of solar spectral measurements started recently at the University of Tromsø (Henriksen et al. 1989a, b).

There is no reliable support for the public concern about a decreasing ozone layer. Treating the long-term ozone measurements from Tromsø, an increase in the thickness of the ozone layer can be documented during the summer months only for the last five years (Henriksen et al. 1990a). For the remainder of each year there is no significant change in the ozone layer at Tromsø when measurements for the last five years, 1984–89, are compared with those for the period 1935–69.

However, spectral irradiances have a crucial effect on the biosphere and life on Earth (see collection of papers published in Ottar (1986: 159). Therefore our spectral measurements were extended beyond the UV region.

A spectrum obtained in the Barents Sea (shown in Fig. 1) is characteristic of observations described in this study. There is an abrupt decrease in irradiance around 3300 Å caused by the ozone layer. A pronounced decrease towards longer wavelengths starts at 4500 Å, caused by off-axis absorption of the teflon sheet used as a diffuser in front of the entrance slit. Unfortunately, this defect in the teflon diffuser was not discovered until the end of the investigations, and therefore the data are not usable for quantitative analysis beyond 4500 Å.

Integrated irradiances for the period 7 July to 30 July are shown in Fig. 2. The measurements were carried out during the midnight-sun period and the solar elevation varies between 2° and 30°, causing clear diurnal irradiance variations. During the day the UVB irradiance reaches 600 mW/m² Å, whereas during the night it is two orders of magnitude less. The maxima on clear days are about five times higher than during completely cloudy days, and this variation is much stronger than expected for variations occurring in the ozone.
Fig. 2. Integrated UVB irradiances from Longyearbyen. The largest decrease in the maxima occurring at mid-day is due to meteorological factors such as clouds and rain.

Fig. 3. Spectral ratios between UVB and visible blue, 400-450 nm. The variations in the maximum heights occurring at mid-day are mainly caused by the varying thickness of the ozone layer.
layer. The maximum UVB irradiance at this latitude is about five times less than the amount at the Equator, and therefore the Arctic regions are preferable to people who are too sensitive to UVB irradiance (Henriksen et al. 1989b).

For the first time large variations in the UVB/visible blue spectral irradiances are documented (Fig. 3). During the day this ratio is about 0.015, whereas during the night it is 0.0005. This variation is mainly due to the effect of the ozone layer. During the day the optical path of the UVB radiation is much less than during the night and therefore the attenuation is less, giving a subsequent higher UVB irradiance. The spectral ratio is much less affected by meteorological conditions than the thickness of the ozone layer. The pronounced diurnal variation of the spectral ratios may have a crucial effect on biological clocks, indicating the time of the day also during the midnight-sun period.

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References


On the oceanic control of the global surface air temperature

ASMUNN MOENE


It follows from the first energy law of thermodynamics that the large time-scale fluctuations of the mean global surface air temperature have to be completely determined by the fluctuations of the total solar heat energy accumulated in the world oceans. The negligible term through the variable mean annual cloudiness within 40°S and N constitutes, in a statistical sense, a continuous white noise thermal forcing of the oceans. A spectral component with a period of about 2500 years, determined by palaeoclimatic records of the global mean temperature, can then be ascribed to stochastically forced inertial oceanic oscillations in a world ocean with a thermal-mixing time constant, i.e. a turnover time of about 500 years.

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On the interannual change of the global surface temperature

As no external work is performed at the top of the atmosphere and at the bottom of the oceans the first thermodynamical energy equation of the total solar heat-absorbing mass (M) of the earth is given as

\[ cM \frac{dT}{dt} = R' \]  

(1)

where \( c \) is the specific heat and \( T \) the spatial mean annual temperature of the mass units. The net annual radiation balance at the top of the atmosphere is given as \( R = Q(1 - \alpha) - I \), where \( Q \) is the total mean annual incoming solar radiation, \( \alpha \) is the albedo and \(-I\) the infrared heat loss to space. \( R \) is the quantity that induces the ensuing global climate change, i.e. the change in the mean global surface temperature. \( R \) can be written as the long-term average plus the deviation, \( R = \bar{R} + R' \). Due to the dependence of \( I \) upon the temperature, we have a long-term radiation balance \( \bar{R} = 0 \).

If \( \tau \) is the response time needed to obtain a temperature change \( \Delta T \) with a given impulse \( R' \), we get by time integration of (1)

\[ \tau = cM \Delta T (R')^{-1} \]  

(2)

i.e. the thermal response time \( \tau \) is proportional to the heat capacity \( C = cM \).

The internal energy is additive and we get from (1)

\[ c_l m_l \frac{dT_l}{dt} + c_a m_a \frac{dT_a}{dt} + c_o m_o \frac{dT_o}{dt} = R' \]  

(3)

where \( l, a \) and \( o \) represent the crust of land areas, the atmosphere and oceans respectively.

Equation (3) is the basic energy equation for global climate changes.

Since \( m_l \ll m_a \) and \( m_o = 1.3 \cdot 10^6 m_o \), the basic equation is reduced to

\[ C_o \frac{dT_o}{dt} = R' \]  

(4)

The main conclusions drawn from (4) is that the derivable large time-scale fluctuations of the annual global surface temperature are determined solely by the fluctuations in the total heat content of the world oceans through the belonging sea-surface temperature fluctuations.

Because of the small heat capacity with corresponding small thermal response time of the atmosphere, the surface air must, in a statistical sense, be in thermal equilibrium with the surface of the oceans at all times.

The global surface temperature is thus completely determined by the huge amount of heat accumulated in the world oceans. The negligible accumulation of heat energy absorbed by an earth without oceans (and consequently without water vapour and latent heat) would cause the surface to warm up until it radiated to space as much energy as it absorbed. For the radiation actually absorbed by the earth, such an equilibrium would be achieved when the temperature at the Equator reached 270°K, the temperature at the South Pole 150°K and at the North Pole 170°K.

Below the surface the water can usually be divided into three zones, in terms of its temperature structure. The three zones are not thermally independent. For time-scales up to a few years only the upper mixing layer has to be included.

For greater time-scales the complete turnover caused by the large-scale thermohaline circulation has to be included. The term age, the time since the water mass was last at the surface and in contact with the heat source and sinks, is estimated to be about 500 years for the world’s deep oceans (Broecker 1979). The thermal response time \( \tau_0 \) of the total mass of the oceans is thus determined by the turnover time to be in the order of 500 years.
On the creation of very low frequency oscillations of the annual global surface air temperature

The perturbation-forcing term $R'$ on the right-hand side of (1) appears through the variable albedo and infrared radiation terms. Clouds are effective reflectors of short-wave solar radiation. A quantitative estimate of the degree to which clouds affect the albedo can be obtained from satellite measurements. The minimum albedo is presumably close to the absence of clouds, and comparison of the minimum albedo with the average albedo shows the reflectory effect of clouds. Most of the oceans within 40° of the Equator have a minimum mean annual albedo below 0.1, but the average albedo is about 0.25. It is clear that the variable cloudiness is a very important factor, as the main accumulation of solar heat radiation takes place in the oceanic given area. The relative importance of factors other than the variable cloudiness is given by the relations that a 3% and 100% increase in water vapour and CO$_2$ respectively can be compensated for by only a 1% increase in cloudiness (Møller 1963).

The atmospheric perturbation inputs $R'$, dominated by the variable cloudiness, must in a statistical sense be a continuous sequence of perturbations $R'(t)$ of the global heat balance, i.e. the white noise thermal input to the oceans confirmed by the observed interannual change in the global sea-surface temperature. This thermal input induces only small perturbations $\Delta T_0$ of the stable time-mean $T_0$ determined by the long-term global radiation balance. For small perturbation inputs the restoring force is a negative linear feedback.

The behaviour of such a system can be described by linear integro-differential equations with constant coefficients.

As a first approximation, the system can be regarded as a first-order autoregressive linear process given

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**Fig. 1.** $\omega > \tau_0^{-1}$: inertial interval. $\omega = \tau_0^{-1}$: with $\tau_0 = 500$ years, period $2\pi \tau_0 = 3000$ years. $\omega < \tau_0^{-1}$: the oceanic mean temperature $T_0$ oscillates with the same low frequency as that of the input forcing.

**Fig. 2.** A review of the mean global surface air temperature changes during the past 10,000 years. The five periodic functions, with periods from 100 to 100,000 years, from which the mean temperature was constructed were derived from a variety of palaeoclimatic records. The maximum rate of change is given for the different fluctuation periods at the left end of each curve (NAS 1975; after Kellogg 1978).
from (4) as
\[ \tau_0 \frac{dT_0}{dt} = R'(t) - T_0, \quad (5) \]
where \( \tau_0 \) is the thermal response time, or time constant of the oceans. The power spectrum is
\[ P = \frac{K}{1 + (\omega \tau_0)^2}, \quad (6) \]
where \( K \) is the constant spectrum of the white noise input and \( \omega \) the radian frequency (Jenkins & Watts 1968, p. 227).

In the frequency interval where \( \omega > (\tau_0)^{-1} \), the term \( (\omega \tau_0)^2 \) is the dominating one and in this inertial interval we have with high accuracy
\[ P = \text{const.} \omega^{-2}. \quad (7) \]

The model given by (7) corresponds to the stochastic model of climate variability (Hasselmann 1976).

As shown in Fig. 1, the white noise thermal input in the oceanic heat accumulation area thus generates in the frequency interval \( \omega > \tau_0^{-1} \) low-frequency inertial oscillations of the mean temperature \( T_0 \) of the oceans. In the frequency interval \( \omega < \tau_0^{-1} \) the temperature \( T_0 \) will oscillate with the same frequency as excited by a very low frequency input.

Introducing \( \tau_0 = 500 \) years in \( \omega = \tau_0^{-1} \) we get the maximum inertial generated spectral power at a frequency with a period of about 3000 years. A second-order autoregressive process includes a spectral peak at the frequency \( \omega = \tau_0^{-1} \), i.e. at the resonant response frequency.

The history of past climates has been reconstructed well enough for us to apply statistical tests to the indication of a spectral peak near a period of 3000 years. The panel on Climatic Variations of the US Committee of GARP, NAS (1975), has summarized such statistics, based on a number of palaeoclimatic records, and has described the history of the mean surface temperature in terms of a superposition of five periodic functions with different amplitudes, given in Fig. 2 (after Kellogg 1978). It appears that the amplitudes are increasing in the interval up to periods of about 2500 years, which is well in accordance with our estimate of the oceanic inertial interval. The fluctuation components with the much larger periods of 20,000 and 100,000 years then have to be ascribed to forced oscillations caused by a very low-frequency external thermal input to the oceans, probably astronomically controlled by the Milankovitch mechanism.

The maximum rate of change for a given perturbation is, according to (5), determined by the thermal response time, i.e. by the great heat capacity of the oceans. The estimated maximum for different fluctuation periods is noted at the left end of each curve in Fig. 2. It appears that the upper limit of this major criterion is not greater than 0.15°C per decade.

References
Exchange of CO₂ between the atmosphere and the ocean

REIDAR NYDAL


This paper focuses on CO₂ uptake by the ocean at higher latitudes. ¹⁴C data from previous nuclear tests have been used as a tracer for the study. Some preliminary ¹⁴C depth profiles in the Norwegian and Greenland Seas provide valuable information about circulation and deep-water formation. In addition to plans for direct CO₂ measurements in the ocean, further ¹⁴C tracer studies will be continued in order to predict the CO₂ development in the atmosphere.


CO₂ in the atmosphere is one of the most important climate gases and more knowledge about CO₂ exchange in nature is necessary at a time when there are increasing levels of CO₂ from combustion of fossil fuel. In order to predict the further development of the CO₂ concentrations in the atmosphere, it is important to know the rate of CO₂ uptake by the ocean and the land biosphere.

An attack on the CO₂ exchange problem in nature was devised in our laboratory in 1962, and based on radioactive CO₂ (¹⁴C) data from nuclear tests in the atmosphere (Nydal 1968). Over a period of more than 25 years it has been possible to trace the excess of ¹⁴C through the atmosphere and down to the ocean surface (Fig. 1). Measurements in the atmosphere have been carried out on samples from several land stations between Svalbard and Madagascar. Samples from the ocean surface have been collected by the Wilhelmsen and Fred Olsen shipping companies in the Atlantic, Pacific, and Indian Oceans (Nydal & Løvsen 1983; Nydal et al. 1984).

The ¹⁴C data have provided an important tool in CO₂ modelling work (Bergh 1989) and give an independent check on global models which are tested against atmospheric CO₂ data from Mauna Loa (CDIAC 1988).

It is realized that the ocean at higher latitudes acts as a sink for CO₂, and it is expected that much effort will be provided in future to study CO₂ uptake and circulation in the ocean. ¹⁴C information from previous nuclear tests also represents one of our most efficient tracers for circulation studies. This was realized with the earlier GEOSECS programme for the ocean in 1972–74, where some depth profiles were also performed on higher latitudes (Østlund et al. 1976).

A further attempt was made in summer 1989 to study ¹⁴C profiles in the Norwegian and Greenland Seas. The observed curves in Fig. 2 indicate a change with time towards the right, according to the penetration of ¹⁴C into the deep ocean. The horizontal scale of the graphs (Δ¹⁴C) is given in per mille excess above the normal level (existing level at 1950). Fig. 2 indicates that the bomb carbon was observed down to a depth of ca. 1000 m in 1972, and the ¹⁴C level had increased by a few percent at the bottom of the ocean in 1989. The immediate impression from these observations is that the renewal of the deep water is very slow, with a maximum age of more than 100 years. If, however, we take

![Fig. 1. ¹⁴C measurements in the troposphere and ocean surface. Δ¹⁴C indicates per mille excess above normal level (¹⁴C level before 1950).](image-url)
Fig. 2. $^{14}$C-profiles in the Norwegian and Greenland Sea. The steep vertical part of the GEOSECS profile indicates the $^{14}$C value before disturbance by the atomic bomb. $\Delta^{14}$C = $-50\%$ is the previous steady state value for the ocean in this area.

... into account an exchange with the deep water in the polar basin ($k_3 = 0$), the apparent age ($\tau_3$) of the deep water changes in magnitude. The behaviour of $\tau_3$ is demonstrated in a 3-box model (Fig. 3), which is applied to the profiles in the Norwegian and Greenland Seas. A preliminary estimate on a few data indicates that the deep water has only reached an average value of ca. 20% of the $^{14}$C concentration in the
ocean surface during a period of 25 years. This calculation demonstrates in a simple way the dependence of the apparent age on the circulation in the deep ocean. More knowledge about this circulation is necessary before $r_2$ can be satisfactorily determined.

A more extended programme in the Nordic seas is planned in July/August 1990 with the Norwegian research vessels ‘M/S Lance’ and ‘G.O. Sars’. Ten to 14 samples will be collected here in each of 10 depth profiles between 66°N and 80°N. In addition to $^{14}$C measurements, it will be possible to carry out measurements on dissolved inorganic carbon (DIC), $\delta^{13}$C, pH, and partly $O_2$. Most of the $^{14}$C data will be based on accelerator measurements (AMS) on small CO$_2$ samples (5 ml).

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References


Correlation between observed anomalies of sea surface temperature and the sea-level pressure above Scandinavia and the north-east Atlantic

KAARE PEDERSEN & INGOLF KANESTRØM


The geographical distribution of persistent sea surface temperature (SST) anomalies and the persistence characteristics of SST anomalies over the North Atlantic Ocean are studied. SST anomalies with a duration beyond about 20-25 days are found to have a constant half-life time of 16 days. For the large-scale, persistent anomaly episodes in the region of strong westerlies, there is a negative correlation, significant at the 2% level, between the SST anomaly and the local zonal index.

J. Bjerknes published a famous paper about air-sea interaction in the North Atlantic (Bjerknes 1962). Lately, this topic about sea-air interaction has received considerable attention. Bjerknes discussed annual anomalies of sea surface temperature (SST). Later studies have mainly been concerned with seasonal or monthly anomalies and possible feedback effects on the atmosphere, with the greatest attention given to the El Niño phenomenon in the Equatorial Pacific. In the Atlantic Ocean, SST anomalies and their possible feedback in the atmosphere have been studied, for instance, by Ratcliffe & Murray (1970), Rowntree (1976), Palmer & Zhaobo (1985) and Karaca & Müller (1989).

Bjerknes found for the annual SST anomalies a negative correlation between the SST anomalies east of New Foundland and the local zonal index (the pressure difference between latitudes 35° and 55°N). Ratcliffe & Murray found that a colder than usual ocean surface in an area south of Newfoundland in September could be associated with positive sea-level pressure (SLP) anomalies the following month over Scandinavia, while an unusually warm ocean in the same area favours the incidence of negative SLP anomalies in the subsequent month over Scandinavia.

Palmer & Zhaobo (1985) suggested that the SST anomalies are initially forced by perturbations in the atmospheric circulation. Bjerknes (1962) discussed possible atmospheric processes that could explain the anomalies; processes such as turbulent flux of latent and sensible heat from the ocean to the atmosphere, horizontal advection in the ocean and vertical advection, or upwelling. In discussing these questions it may be helpful to know more about the character of the SST anomalies, both where they mainly occur and their characteristic time and space scales.

Data

The sea surface data used were obtained from daily 1200 UGT analyses provided by the National Centre for Atmospheric Research (NCAR). The data used cover the period 1962-83. We intended to analyse the SST as far north as possible. However, in connection with the Norwegian Polar Low Project, we studied the cold air outbreaks over the Norwegian and Barents Seas (Hansen 1987). It was found that prior to 1974, the NCAR data were rather unreliable in this region. Therefore, the area north of 60°N is excluded in this discussion.

Gridpoint SST anomaly events

We calculated the area averaged SST anomaly, where the area is a circle centred at the gridpoint with a diameter of 15° latitude. If the numerical value of the anomaly at that particular gridpoint remained equal to or greater than 0.55 K for at least 10 days, we defined the occurrence of a persistent SST anomaly event. The geographical distribution of the number of positive events for the winter season (November through March) are given in Fig. 1a. We find three primary regions with high numbers of events:

- The belt of westerlies.
- The trade wind region.
- The intertropical convergence zone.

In the westerlies the distribution shows two maxima of frequency of occurrence; one located in the Gulf Stream region near Cape Hatteras and the other in the Labrador current region near Newfoundland. Also in the trade wind belt two maxima are found; in the upwelling region near Africa and in mid-ocean. The high number of anomaly events is located in zones where the wind speed is strong.

The distribution of negative events is given in Fig. 1b. We find this distribution to be very similar to that of the positive events. The half-life time of the anomaly events (the time required for half of the anomaly events to be terminated) with a duration beyond 20-25 days is found to be about 16 days.
Fig. 1. The geographical distribution of the number of gridpoint SST anomaly events; (a) positive events for the winter season, (b) negative events for the winter season.

Fig. 2. The composite sea surface temperature anomaly map (contour interval 0.5 K) for the anomaly episodes in the westerlies for the winter season; (a) warm episodes, (b) cold episodes.
Anomaly episodes

We are interested in investigating stationary, persistent SST anomaly episodes which have a relatively large magnitude and horizontal scale. An anomaly episode is defined as a group of 15 or more connected gridpoints each of which simultaneously fulfills the magnitude criterion for an event. The duration period is now 20 days, and the time-averaged SST anomaly must have a numerical value equal to or greater than 1.5 K at least in one gridpoint.

Using these criteria for the SST anomaly episodes, we find for the three geographical regions 89 episodes in all (42 for the winter season). Most episodes start with several small-scale anomalies, growing in amplitude and extending to a coherent area. Especially in the trade wind zone, the anomaly extends across the width of the ocean basin.

The anomaly episodes in the westerlies

The anomaly episodes located in the westerlies are assumed to have greatest impact on the climate at high latitudes. Therefore these will be investigated in more detail. We find that 11 warm and 7 cold episodes have extreme values in the region 40°–52°N and 40°–55°W. A composite SST anomaly map of the 11 warm episodes is shown in Fig. 2a. The anomaly amplitude for each episode is weighted according to its duration. The extreme value of this composite picture, 2.8 K, is situated at 47°N and 47°W. The sign of the SST anomalies in the trade wind region is also seen to be positive, but their values are less than 0.4 K.

A composite picture of the 7 cold winter episodes is given in Fig. 2b. The minimum value of −2.1 K is located at 45°N and 45°W.

Bjerknes (1962) found that the SST anomalies (based on annual SST values) and the local zonal index were correlated in such a way that warm SST anomalies were associated with low index and cold SST anomalies with high index situations. To investigate whether this is also true for the anomalies based on daily SST values, we have constructed the composite sea-level pressure (SLP) anomaly map for the warm and cold winter episodes (see Fig. 3). For the cold episodes we find a negative pressure centre (−4.3 hPa) at 62°N, 40°W, a positive pressure centre (4.9 hPa) at 58°N, 55°E and a negative pressure centre (−2.6 hPa) at 58°N, 55°E. We also find a positive pressure centre (2.9 hPa) at about 35°N, 45°W. For the warm episodes we find four corresponding pressure centres with the values 3.7 hPa, −1.0 hPa, 2.6 hPa and −2.0 hPa, respectively. As seen, the composite SLP anomalies for the warm and cold episodes have the opposite sign. This change in the main pattern of the SLP is especially marked at 45°W. From Fig. 3 we see that there is a negative correlation, significant at the 2% level, between the SST anomaly and the local zonal index. It should therefore be evident that the SST anomalies have a great influence on the atmospheric circulation and the connected weather or climate. For a cold episode, an atmospheric high is built up over Scandinavia, and this high will block the air flow over north-west Europe. In the warmer episode, we find a more progressive type of air flow over the north-east Atlantic and north-west Europe.

The turbulent flux of latent and sensible heat from ocean to atmosphere is of great importance. With an increased zonal index the sea surface will be cooled and will create a stronger north–south temperature gradient, which in turn will increase the baroclinity in the atmosphere. Thus we get a feedback mechanism between the local zonal index and the SST anomaly.

References


P. TAALAS & E. KYRÖ

As a consequence of the detection of Antarctic ozone depletion (Farman et al. 1985), ozone research in the Arctic regions has been intensified. A considerable number of new ozone observation stations have been established and several ozone research campaigns have been commenced in the Arctic regions during recent years (Evans 1990; Turco et al. 1990; Hofmann et al. 1989). So far, no severe ozone depletions comparable with that in the Antarctic have been found in the northern polar region. The Trend Panel of NASA (Watson et al. 1988) and WMO (1990) have reported long-term decreasing ozone in the northern hemisphere with the largest deviations in January at high northern latitudes, whereas Larsen & Henriksen (1990) have reported that no negative trends of total ozone in Norway have been observed.

The total ozone observations of Tromsø (Northern Norway), Sodankylä (Northern Finland) and Murmansk (Northwestern Soviet Union) for 1987–89 have been studied, and comparisons of total ozone with stratospheric temperatures have been made. These values have also been compared with the long-term means of total ozone and stratospheric temperatures. No severe ozone depletions were observed. The exceptionally high total ozone values at these stations in February 1990 were connected with the abnormally high stratospheric temperatures. The 1989 monthly mean ozone sounding observations at Sodankylä, Bear Island and Ny Ålesund (Spitsbergen) have also been studied. Temperature observations made in Sodankylä over 24 years revealed the existence of a potential for polar stratospheric cloud formation in the lower stratosphere in winter and early spring. A trend analysis of 30–100 hPa temperatures revealed a negative trend of \(-3.85\) K in January and a positive trend of \(3.49\) K in April; the year-to-year trend was only \(-0.47\) K for this period.

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Fig. 1. The monthly means of total ozone at three sites in Northern Scandinavia and one site at the Antarctic Peninsula, and the monthly means of 50 hPa temperatures in Sodankylä. The 1935–69 and 1984–88 means of total ozone in Tromsø and the 1965–88 means of 50 hPa temperature in Sodankylä are plotted as references.
Island and Ny Ålesund (Spitsbergen) reveal any indications of pronounced ozone depletions. A comparative layered study of ozone, temperature and relative humidity in Sodankylä indicated that the springtime variability of ozone in the lower stratosphere was clearly connected with the meteorological variability, whereas the minima observed in summer was not linked to meteorological variability and may therefore be of chemical origin. The lower tropospheric ozone had two distinct maxima, one in spring with large-scale photochemical causes, the other in summer connected with the emissions of hydrocarbons and oxides of nitrogen in Europe. The causes of high surface ozone concentrations have been studied by Taalas (1988).

Temperature observations carried out in Sodankylä over 24 years (1965–88) reveal the existence of a potential for polar stratospheric cloud formation above northern Scandinavia in the lower stratosphere in winter and early spring. The percentages of temperatures below −78°C for each month are plotted in Fig. 3 (−78°C is the critical temperature for nitrous acid-trihydrate cloud formation; Arnold et al. 1989). A trend analysis of 30–100 hPa temperatures revealed a

Fig. 2. The monthly mean ozone profiles at three North-European sites in 1989. The profiles are calculated as 500 m means.
negative trend of $-3.85\ K$ in January and a positive trend of $3.49\ K$ in April; the trend of yearly averages was only $-0.47\ K$ for this period. The connections of the stratospheric temperatures with the quasi two-year variability of the tropical stratosphere and the 11-year cycle of the sun's activity have been studied. The connection between the stratospheric temperatures in Sodankylä and the phase of QBO in the tropics alone was not pronounced. On the contrary, while separating the January–February mean temperatures according to the phase of the QBO in the tropical stratosphere, connections between temperatures and sunspot numbers are found. This method has been described by Labitzke & van Loon (1988).

Fig. 3. The frequency of the stratospheric temperatures below $-78^\circ C$ in Sodankylä at $30-100\ hPa$ levels during 1965–88.

References


Ultraviolet radiation: the missing link in the ozone debate

ANN R. WEBB


The need for spectral measurements of solar ultraviolet radiation is discussed and illustrated with preliminary results from work now in progress to provide such spectral data.

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Recognition that ozone in the stratosphere is being depleted by man-made chlorofluorocarbons has focused world attention on these pollutants and the effects of their continued usage. The direct effect is a reduction in total column ozone; the resulting effects range in estimate from an increased incidence of skin cancer to disruption of the food chain and destruction of whole ecosystems (Russell-Jones 1987; Calkins 1982). The link between direct and resulting effects is ultraviolet (UV) radiation (200–400 nm).

Ozone is a strong absorber of the shortest wavelength UV radiation, preventing any UVC (200–280 nm) radiation from reaching the earth’s surface. However, absorptivity decreases through the UVB (280–320 nm) part of the spectrum so that as wavelength increases more and more radiation reaches the earth’s surface. The amount of ozone along the pathlength of radiation through the atmosphere thus determines both the short wavelength limit and intensity of the ground level UV spectrum.

At the surface, the UVB part of the solar spectrum contains less than 1% of the total solar energy, but photon for photon it is the most energetic radiation to which we are exposed and those packets of energy cause biological reactions, many of them detrimental to life.

To assess the biological significance of changing ozone, and hence UVB climate, it is necessary to know: (a) the distribution of effective UVB in the natural environment; (b) the exposure of a target body to the available UVB; and (c) the rate of photoreaction per unit of incident UVB. The overall result of irradiation is a function of (b) and (c). The latter (c) is effect specific, each photoreaction has its own action spectrum. The exposure of a target body (b) depends upon the available radiation (a) and any modification which may take place, e.g. wearing a sunscreen, moving into shelter. For many organisms modification on a short time-scale is not possible, though adaptation can occur over a longer period (e.g. the cuticular surface of plants can thicken). Exposure then depends on the natural distribution of UVB (a), which is a function of latitude, season and ozone amount (Bener 1969). All forms of life have evolved to maximize survival rates in their natural climates. If that climate is subtly changed by changing ozone and hence ambient UVB levels, then species susceptible to UV damage must either adapt or weaken and decline.

To understand the extent to which changing UVB radiation may affect life systems, knowledge of current ambient levels of UVB around the globe is needed; the range of intensities to which species are exposed and survive now. Next, if and how UVB at the surface is changing, stratospheric ozone depletion does not automatically mean increased UVB at the ground. Besides various feedback processes involving UV photons in ozone chemistry, man is also polluting the troposphere with ozone (Bruhl & Crutzen 1989). In the biosphere ozone is undesirable in that it damages plant life and adversely affects the respiratory system of mammals (Schneider et al. 1989), but it also absorbs UVB radiation and in this sense helps to offset any stratospheric ozone depletion (if the high tropospheric/low stratospheric concentrations occur in the same region). Further complication comes when cloud, aerosol and other pollutants have to be considered.

To gain this knowledge of UVB, climatology measurements are needed: measurements of natural seasonal and spatial variability; measurements over a long period to assess changes/trends; measurements upon which to base realistic biological research and to verify model simulations of radiative transfer: spectral measurements.

These applications require a detailed knowledge of the UVB spectrum rather than a broadband energy measurement (Caldwell et al. 1986). Consider the biological effects: many biological action spec-
tra decrease sharply with increasing wavelength across the UVB waveband, whereas the irradiating solar spectrum increases dramatically in the opposite sense (Fig. 1). Thus, a broadband measurement may indicate doubling of UVB, but if this energy increase is at the long wavelength end of the waveband it will have little biological effect. Conversely, a slight shift of the spectrum to shorter wavelengths may not be detectable in broadband energy terms but may have a profound biological effect. For the same reason model predictions of pollution scenarios must be able to accurately provide similar spectral data.

To make accurate spectral measurements requires an instrument with high sensitivity, good wavelength resolution and rigorous stray light rejection, and in this instance the ability to function outdoors. Until recently the majority of solar UVB measurements have been made with broadband instruments, e.g. the frequently used Roberton-Berger meters (Barton & Robertson 1975; Scotto et al. 1988) a lack of demand for spectral data combining with technological and cost restraints to preclude widespread solar spectral measurements. This situation is now changing and more spectral measurements are being made (Webb 1985; Henriksen et al. 1989; Roy et al. 1989).

At Reading, a commercially available Optronics 742 spectroradiometer is used for this task. The instrument is basically a double holographic grating monochromator with photomultiplier tube detector and cosine corrected teflon diffuser. Measurements have been made on clear days since August 1989 over the waveband 280–400 nm, sampling every 2 nm. The monochromatic irradiance at a wavelength of 300 nm and the total UVB irradiance (the integrated spectrum for wavelengths ≤ 320 nm) are shown in Fig. 2 for two entirely clear days: 23 August and 29 November 1989. This illustrates both the daily and seasonal changes in UVB radiation and emphasizes the greater relative changes at short wavelengths. Total solar radiation, both global and diffuse, is also recorded to assess how the UVB compares with these more widely available and routinely measured parameters. Ozone measurements are available from the UK Meteorological Office.

Other sites in Europe also have the capability to make spectral solar UVB measurements; in Australia and New Zealand there are plans to supplement the existing broadband UVB network with spectral measurements, and scientists in America are also addressing this question. Thus it is to be hoped that in a few years we will hear not only of worldwide atmospheric ozone

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Fig. 1. Generalized erythema (E), plant (P) and DNA (D) action spectra and the solar spectrum measured at noon on 23 August 1989 in Reading.

Fig. 2. Irradiance at wavelength = 300 nm (----) and integrated over the whole UVB spectrum, λ ≤ 320 nm, (-----) on two clear days in Reading: 23 August (A) and 29 November (N).
trends but also about the ground level consequences of those trends—the global patterns of UVB radiation to which we are all exposed.

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References


Peat humification and climate history

EILIF NILSSEN & KARL-DAG VORREN


Fourteen ombrogenic peat sequences from northern and central Norway have been submitted to colorimetric determination of humification degree at every vertical centimetre, and six additional sequences from the same area have been analysed by von Post's method. The peat horizons with transitions from comparatively high to comparatively low humification degree have been dated by means mostly of interpolations between radiocarbon-dated levels. The preliminary survey of the temporal distribution of the transitional humification horizons back to 4600 BP tit very well with those found in Danish bogs (Aaby 1978).

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The relation between climate and peat humification, especially in ombrotrophic bogs, has been discussed by various authors for at least a century (see Barber 1981). Well known are the studies of von Post & Granlund (1926) and Granlund (1932) in southern Sweden. Granlund named the shifts between dark and light peat in raised bogs 'recurrence surfaces' if they occurred over a large area of the bog. He maintained that precipitation was the most important factor for the development of recurrence surfaces. Later, G. Lundqvist (1962) questioned the synchronicity of these recurrence surfaces, based on $^{14}$C datings. However, several researchers still believe in an intimate connection between bog peat humification and climate.

Aaby (1978) explained this connection as follows: The main factors determining peat colour are temperature and precipitation. The balance between these two factors, i.e. basically the humidity, determines the humification degree and the degree of peat decomposition. Thus, if precipitation during the growing season is low and temperature high, peat becomes dark; if the precipitation is high and the temperature low, the peat becomes light.

The wet parts of a bog, the hollows, may be permanent features extending over several thousands of years (see Foster & King 1984; Foster & Fritz 1987, Foster et al. 1988). The hummock centres may also have persisted for a long period, constituting a 'wall' of hummock peat downwards into the peat layers, and surrounded by hollow peat (cf. Walker & Walker 1961). The history of north European hummocks may, however, be more complex (Vorren 1972).

Many of the bog sections in the present study are eroded in the upper part. Erosion is still occurring, but at a negligible scale. Surface erosion was probably most active during the Little Ice Age. Erosion may have removed peat layers down to the 2000 years BP horizon in the bogs. In several cases, depressions of the entire bog plain may be recorded. The erosion was probably most active in the subarctic and subcontinental climate areas.

Most sections have been sampled without any associated investigation of the overall stratigraphy of the bog, so the sequences largely represent isolated sampling points, mostly from the marginal part of the bog plain. When sampling, coring was normally in hollows. The area of the different bogs sampled varies from ca. 0.4 to ca. 10 ha.

The sections are from ombrotrophic bogs in four different districts of central and north Norway (Fig. 1). The three southernmost districts belong to the Köppen's C-
which in our opinion is quite reliable when employed on fresh material. Colorimetric measurements have been carried out on 14 sequences, and von Post's method on 6.

The $^{14}$C datings of each section, numbered from 2 to 6, are rarely connected directly with the humification shifts, as they mostly concern palynological/biostatigraphical events. Most of the peat sequences have been pollen-analysed, especially with regard to agricultural history. Thus the datings of the humification shifts are usually based on extra- and interpolations. This is, admittedly, a method which can include major sources of error, as the peat is unlikely to have a constant accumulation rate.

Most of the datings are concentrated in the upper part of the sequences. Sequences reaching back beyond ca. 5000 BP number only 7, whereas 10 reach 4600 BP. Because of these two sources of error we only discuss humification shifts younger than 4600 BP.

Fig. 3 shows the different main humification shifts of the individual sequences, from dark to light, plotted against a $^{14}$C age scale. The horizontal black columns in the middle of Fig. 3 indicate the chronological extent and number of humification shifts in subjectively distinguished periods. Not all periods are represented in all sequences because of their individual chronological extent, and because all 'humification shifts' are not equally clearly marked in one and the same sequence. When considering the relative importance of the different periods of humification shifts, the second column from the right gives a summary. Here the periods have been ranged according to their relative representation in the sequences. Thus the period 3850–3750 BP and the 2950–2750 BP period seem to be most important. Also the periods 750–500, 2200–2050, 400–200 and 4600–4400 BP should be noted.

The two most important periods may coincide with agricultural expansion periods, and the humification shifts may thus be at least partly due to human impact. This is climate region of advective maritime climates, the very southernmost district being a little thermally continental. Stations 17–19 of the northernmost district, number 4, belong to a subcontinental D-climate, whereas the northernmost station belongs to a cool maritime C-climate. In all districts, with the exception of much of district 4, agricultural activity is recorded in deposits younger than 2000–3000 years.

Two methods of recording humification have been employed; a modified version (cf. Berglund 1986) of the colorimetric approach of Overbeck (1947) (cf. Fig. 2), and the far less time- and labour-consuming approach of von Post (1924),
also the case with the other humification shifts in Fig. 3.

Comparison with Danish data

Aaby (1978) dated humification shifts in 15 peat sequences from Danish peat bogs and plotted the datings against a calibrated time scale. He suggested that there might be a tendency of a 260-year regular cyclicity between the shifts.

In the present material such a cyclicity/pseudo-cyclicity may be traced only locally and for short time intervals (cf. Fig. 3, site Elgsnes between 5760 and 4850: a 200–250 year cycle, site Rystad between 2800 and 1750 BP: 100–140 year cycle).

If we group Aaby’s (1978) original (in radiocarbon age) datings into natural clusters, derive an average 14C age from each cluster, and compare his average datings with our group average datings (in radiocarbon age), a rather surprising similarity emerges between the ages of his and our humification shifts from dark to light (Table 1).

The chronological similarity demonstrated here supports the hypothesis that changes in peat humification are linked primarily with synchronous macroclimatic changes, in this case within the entire C-climate region of northern Europe. Thus the frequent correlation between agricultural expansions (cf. Vorren et al. 1990: 2900–2750, 2650–2480, 1850–1650, 1250–1100, and 500–400 BP) and humification shifts is probably mostly climatically induced. However, forest removal in connection with land clearings may cause a rise in the groundwater table, which may influence smaller bogs. Such a groundwater rise may theoretically cause a humification shift. However, we propose that geographically widespread and chronologically similar shifts in peat humification (Table 1) most probably result from macroclimatic changes.

References


Genetic distances between geographically isolated *Pedicularis dasyantha* populations in Spitsbergen, Svalbard Archipelago, Norway. Evidence of glacial survival?

ANN MARIE ODASZ, KATRI KÄRKKÄINEN, OUTI MUONA & GEROLD WEIN

Geographically isolated populations of *Pedicularis dasyantha*, an arctic herbaceous plant on Svalbard, show strong phenotypic variation suggesting routes of migration from sites of glacial survival. Electrophoretic isozyme analyses of 6-phosphogluconate dehydrogenase (6-PGD) revealed significant variation in allele frequencies between the isolated populations. The pattern of gene-pool variation and the important evolutionary development of self-compatibility in an otherwise self-incompatible genus suggests survival of the species during an extended unfavourable period in Svalbard. A scenario is outlined where development of self-compatible reproduction is set to the period of climatic deterioration which followed the Eemian interglacial (126,000 years BP). A decrease in population genetic variation due to the Founder effect and 'bottle-necking' followed by Holocene expansion of the limited gene-pools is interpreted from the extant distribution of the broad overlapping 'ancient' gene-pools.

Mutations are important events for introducing and increasing genetic diversity in populations. Introduction of such variation is especially important in environments that are heterogeneous in space and/or time. Maintaining the genetic diversity thus generated can therefore be critical for survival in severe environments that can vary considerably from one year to the next.

Self-fertilization assures the persistence and stabilization of a phenotype that is well adapted to the immediate environment (Lloyd 1965). Thus, in severe and unpredictable environments a self-compatible population may possess a selective advantage over a self-incompatible population because individuals are more closely adapted to the immediate environment such as the extreme short-growing seasons encountered in Svalbard (Odasz 1988; Gauslaa & Odasz 1990).

The requirement of outcrossing in the self-incompatible populations may be considered an evolutionary 'trap' (J. L. Harper, pers. comm.), and would eliminate mutations in corolla shape and colour that are unattractive to the required pollinators and would also eliminate all other phenotype changes associated with or linked to such mutations. In contrast, self-compatible populations are 'freer' phenotypically to express corolla mutations because they are no longer required to maintain the required attractiveness to insect pollinators. Such mutations and the associated genetic information might result in phenotypic variation leading to increased reproductive success and higher survival rates in the severe Arctic environment.

The selected species for this study, *Pedicularis dasyantha*, has a number of attributes which facilitate population genetics investigations. It is a diploid species making interpretation of genetic composition straightforward. It reproduces by seed and not vegetatively, which makes identification of individual plants simple (Odasz 1988). It occurs in isolated disjunct populations, which simplifies identification of separate and non-interbreeding populations. And there are numerous individuals in the sites where it does occur providing large sample sizes for analyses.

*Pedicularis* species thus far examined for their pollinating mechanisms (in North America, including Canada and Alaska) are found to be self-incompatible (Kevan 1972; McInnes 1972; Macior 1975, 1982). The *Pedicularis* populations in Svalbard, however, have been shown to be self-compatible and plants in insect exclosures produced as many seeds as control plants in nature (Odasz 1988).

During the field seasons 1985–89 phenotypically heterogeneous populations were recognized in the field on Svalbard. These populations have been geographically isolated from each other for different periods of time. They show phenotypic evidence of genetic divergence over possible time-spans as short as a few thousand years to as long as many glacial periods (i.e. Pleistocene and/or Weichselian Glaciations).

Plant population genetic structure has attracted considerable attention (Smith & Schaal 1979; Learn & Schaal 1987; MacDonald et al. 1988; King & Schaal 1989). In spite of the recent interest, there are still too few data to allow generalizations about the relative extent of population genetic structure in different species. However, the study described in this paper is the initial investigation to date addressing this question on Arctic plants.

Methods

Thirty plants were collected from each of the 13 isolated populations on Svalbard during the summer of 1989. Plant material was placed on dry ice and maintained at −30°C. This quick freezing of plant material was required because numerous previous attempts to germinate seeds has not been successful.

Preliminary electrophoretic analy-
Fig. 1. Dendrogram based on Nei's (1972) genetic distance and UPGMC clustering (Sneath & Sokal 1973) summarizing genetic relationships among 13 populations of *Pedicularis dasyantha* on Svalbard. The Isfjorden populations are in the top part of the dendrogram and the west coast populations in the lower part.

Fig. 2. Distribution of allele 1 and allele 2 for the enzyme 6-phosphogluconate dehydrogenase (6-PGD) in 13 geographically isolated populations of *Pedicularis dasyantha*.

Results

Despite reports of only monomorphic loci in the genus (Waller et al. 1987), the species *Pedicularis dasyantha* from Svalbard shows clear polymorphism at the 6-PGD enzyme locus. Genetic distances shown in the dendrogram (Fig. 1) are greatest between the populations south of Isfjorden and the northwest coast. The distributions of alleles 1 and 2 of 6-PGD are mapped (Fig. 2). The gene pools overlap north of Isfjorden.

Discussion

Variation in allele frequencies between isolated populations suggest general areas of possible glacial survival. The genetic distances (Fig. 1) and the distribution of alleles 1 and 2 at the 6-PGD locus (Fig. 2) suggest a scenario with the development of self-compatibility (Odasz 1988) during the last period that was warmer than present (probably Eemian interglacial, 126,000 BP), followed by restriction of population genetic variation due to the Founder effect and 'bottle-necking' by extreme environmental constraints (Fig. 3). These constraints on gene-pool diversity resulted in disjunct, isolated populations with limited gene-pools. The subsequent Holocene expansion of the species is suggested by the extant distribution of overlapping 'ancient' gene-pools (Fig. 2).

The pattern of genetic distance between populations (Figs. 1 and 2) supports the hypothesis of plant survival during a colder period, but better resolution is required to identify gene flow within and between the isolated populations, and to identify sites and times of population isolation (Eemian 126,000 BP? or Holocene 10,000 BP?).

During the 'International Conference on Climate of the Northern Latitudes: Past, Present, and Future' in Tromsø, 2–4 April 1990, there was general agreement on the glacial maximum ice distribution on Svalbard which allowed for ice-free refugia south of Isfjord, Svalbard (Mangerud et al. 1990) and ice-free sites on NW Svalbard (Forman & Miller 1984). This pattern correlates well with the isozyme results as
shown in Figs. 1 and 2. However, warm trends in Fram Strait during the early Holocene (10,000–12,000 BP) (Hald et al. 1990; Lehman et al. 1990), which also seem to correspond with conditions towards the west on Ellesmere Island, Canada (Koerner 1989) and with the early holocene mollusc distributions on Svalbard (Funder 1990; Salvigsen 1990) create a dilemma. The question arises: ‘Have plants like Pedicularis dasyantha survived in ice-free sites on Svalbard since the Eemian or are they survivors from the warm period during early Holocene?’

Plants persisting in ice-free sites during the arid and cold climate (Schweger et al. 1982; Kutzbach 1987) of the glacial maximum require eco-physiological adaptations to conditions that are more extreme than today. Pedicularis dasyantha and other tundra species from Svalbard exhibit adaptations to such cold and dry conditions (Odasz 1988, 1990; Gauslaa & Odasz 1990) and presently thrive in arid periglacial habitats.

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Fig. 3. Diagram illustrating the suggested scenario with decline in population genetic variation during a cold period in Svalbard and subsequent population expansion with ameliorating temperatures.
A preliminary history of the Little Ice Age in a mountain area in SW Norway

LOTTE SELSING, ODDVEIG FOLDØY, TROND LØKEN, EINAR SOLHEIM PEDERSEN & ERIK WISHMAN

Pollen analytical, archaeological, ethnological, historical and meteorological evidence are combined to establish a preliminary history of the Little Ice Age in a mountain area, Øystøl, in SW Norway. A climatic deterioration in the period AD 1300-1800 is suggested. Ripening of barley in this period was uncertain at the summer farm areas (550-700 m a.s.l.) which were deserted in the first part of this period, and presumably also in the farm area of Hamrabø (300 m a.s.l.).

This study was carried out in 1983 as part of a registration of prehistoric and historic monuments. The investigated area in Suldal, SW Norway, includes Hamrabø (ca. 300 m a.s.l.) with four farms and their summer farms (550–700 m a.s.l.), one of which is Øystøl (650 m a.s.l.) (Fig. 1).

The bedrock (Sigmond 1975, 1978) is characterized by Precambrian metamorphic rocks, the Ryfylke Schists and the Caledonian allochthonous gneisses. Nemoral deciduous trees are present in the lowlands. At higher altitudes, boreal pine forest takes over. This forest is replaced by subalpine birch forest which reaches to 750–800 m a.s.l. at higher levels. A reforestation in the summer farm area is in progress due to a reduction in grazing and felling of timber for firewood. Above the forest limit low alpine and alpine heaths and grasslands exist.

Results

The pollen investigated site, Øystøl, 650 m a.s.l.

A small peat deposit, 50 cm deep, located on the outskirts of a till deposit close to one of the house ruins (Fig. 2) was sampled. Samples were collected directly from the exposed wall of a trench (stratigraphy, Fig. 3). An open birch forest, heather and grassland pastures surround the site. Preparation of the palynological samples mainly fol-
Fig. 2. The pollen site at Øystøl (650 m a.s.l.) is a small peat deposit located on the outskirts of a till deposit close to one of the house ruins. Presumably the peat has earlier covered a greater area, some of which is seen in the foreground. Photo: Jo Selsing.

Fig. 3. The peat stratigraphy at the pollen site at Øystøl. The deposit is 50 cm thick. Peat stratigraphy from top to the bottom: Layer 1 – Dark brown peat with a light grey horizon at the base (marked with an open circle). Layer 2 – Peat with changing brown colours and grassroots. Layer 3 – Dark brown peat with a prominent light grey horizon at the base (marked with open circle). The transition between layers 2 and 3 is not sharp. Layer 4 – Black peat with light grey horizons. Layer 5 – Till with two great stones; the one to the right was used as a foundation for the house, now ruined. Star – Radiocarbon date.

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Fig. 3. The peat stratigraphy at the pollen site at Øystøl. The deposit is 50 cm thick. Peat stratigraphy from top to the bottom: Layer 1 – Dark brown peat with a light grey horizon at the base (marked with an open circle). Layer 2 – Peat with changing brown colours and grassroots. Layer 3 – Dark brown peat with a prominent light grey horizon at the base (marked with open circle). The transition between layers 2 and 3 is not sharp. Layer 4 – Black peat with light grey horizons. Layer 5 – Till with two great stones; the one to the right was used as a foundation for the house, now ruined. Star – Radiocarbon date.

lowed the acetolysis method (Fægri & Iversen 1975). The pollen dia­
gram (Fig. 4) is a percentage dia­
gram subdivided according to Hedberg (1976) and Mangerud et al. (1974). The zone boundaries are

based on marked changes in the pollen assemblage composition.

Pollen analysis: Øy1 local pollen zone reflects a subalpine open birch forest mixed with pine and alder. Indicators of light demanding vege-
tation and of pastures (Plantago lanceolata) are already present.

Through the Øy2 local pollen zone tree values are reduced, while dwarf shrubs and herbs increase. Small scattered groups of birch for-
est may still occur, while the vegetation is dominated by herbs and dwarf shrubs. The pollen site was moist. Deforestation and culturally influenced vegetation resulted from human land use. The charcoal peak at the top of the zone may have derived from usage of the house ruin nearby.

In the Øy3 local pollen zone there is an increase in moisture indicating taxa with domination of Cyperaceae. Tree pollen values are very low. There is no indication of changes in the local topography, e.g. the drainage system. The very low tree pollen percentages may be explained by an overrepresentation of locally produced Cyperaceae, while no reforestation occurred. The observed changes are inter-
preted as the effect of a reduction or ceasing of cultural activity in the area, maybe combined with a cli-
matic deterioration.

Øy4 local pollen zone resembles zone Øy2. There is an increase in tree pollen in the top sample while the vegetation was still dominated by herbs. Light demanding taxa and indicators of pasture are present, while moisture indicating taxa are reduced. A small increase in char-
coal is seen in the zone.

Radiocarbon dating

The level 13–16 cm was radiocar-
bon-dated to 990 ± 60 years BP (NaOH soluble fraction, T-5797A) (AD 980–1050 calib., Stuiver & Becker 1986). The age is a maximum date of the first occurrence of Picea pollen in the lower part of zone Øy3. The charcoal peak in Øy2 must therefore be from the Early Middle Ages or later. The small rise in charcoal in the lower part of local pollen zone Øy4 may be from the period AD 1800–50, when the summer farms expanded and many of the summer farm buildings were prob-
**Biostratigraphy**

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Total pollen diagram</th>
<th>calculation basis Σ pollen</th>
<th>calculation basis Σ pollen + x</th>
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**Spore plants**

<table>
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<th>Total pollen diagram</th>
<th>calculation basis Σ pollen</th>
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Analyses: Lotte Selstng 1983

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**Fig. 4.** Lithostratigraphy and pollen percentage diagram from Øystøl. The calculation basis is total pollen, or total pollen + spores or antrax (charcoal fragments).
ably built. Presumed uniform peat deposition and a decreasing peat compression towards the top of the deposit, the period of climatic deterioration is dated to AD 1300–1800.

Archaeological evidence from Hamrabø–Øystøl relevant for climatic interpretations

Clearance cairns, house ruins, cattle paths and charcoal pits point to Øystøl and the area to the east of Øystøl as the only one in the Hamrabø summer farm area with possible human occupation in the Late Iron Age/the Middle Ages (Lillehammer 1970; Bakka 1972; Løken 1989). Presumed permanent settlement from this period has been recorded in adjacent mountain areas (Hafsten 1965; Odner 1969; Høgestøl & Brøsch-Danielsen 1986; Brøsch-Danielsen 1990). We suggest that Øystøl was a permanent settlement in periods with pressure on agricultural resources.

Ethnological evidence from Hamrabø–Øystøl relevant for climatic interpretations

The summer farm area belonging to Hamrabø has been extensively exploited as a result of cattle breeding and fodder gathering in the period from the 17th century to 1960 (Hoffman 1948; Pedersen & Foldøy Solli 1979). The farmers exploited all available niches in a yearly summer farm cycle from sea level to 700–800 m a.s.l. (Table 2).

Hamrabø in historical sources

Hamrabø was probably deserted as a result of the Black Death between AD 1340 and 1350. It remained deserted for more than 200 years. Permanent settlement in Hamrabø is obvious from around AD 1600 (Tax records 1519, 1521, 1563 p. 37; County accounts 1601, 1983; Hoftun 1972; Rygh 1915). All four farms were in use for a few decades. For about 200 years (AD 1600–1800), between 30 and 40 people lived here with about 40 heads of cattle and horses and 100 sheep and goats (Land registers 1668, 1723; National censuses 1769, 1801, 1815; Censuses of agriculture 1802, 1808–1809). The summer farm area was only used to a small degree. In the course of a few decades the number of people rose to 100 (AD 1845) (Censuses 1815, 1825, 1835, 1845). The number of cattle and horses increased to ca. 100 and sheep and goats to ca. 400 (Censuses of agriculture 1835, 1845). Only 30 years later, in AD 1875, the numbers were reduced to 58 people, ca. 75 cattle and horses, and ca. 200 sheep and goats (Census and Censuses of agriculture 1875). The production of vegetable food (grain and from 1835 potatoes) follows the same pattern as the milk, wool and meat production. Full exploitation of Øystøl as a summer farm in the latter part of the 18th century lasted well into the 20th century with a peak around AD 1850 (Pedersen 1982). Permanent settlement has not existed in modern times at Øystøl.

Between AD 1660 and 1800 the population in Norway doubled, while in Hamrabø the number of people remained stable throughout this period. In all of Norway the population increased by about 50% from AD 1800 to 1850, while in the marginal area of Hamrabø it nearly tripled.

Cultivation of barley, and its relation to atmospheric circulation

In periods with little or no trade the minimum factor of making a living through agriculture has been the ability of the farm to produce sufficient quantities of grain. About 75% of the food, converted to calories, came from plants, mostly grain (Lunden 1978, 1988). If a farm could not produce enough grain, subsistence possibilities were small. The cereal generally grown in...
Table 3. Mean August temperatures in the lowlands (Suldal-Mo) and at the summer farm levels (Stråpa-Sandsa) during mean anticyclonic conditions (August 1976) and mean cyclonic conditions (August 1979) (Wishman 1985).

<table>
<thead>
<tr>
<th>Meteorological station</th>
<th>Mean temperatures (°C)</th>
<th>Aug. 1976</th>
<th>Aug. 1979</th>
<th>Temp. decrease</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>(anticycl.)</td>
<td>(cycl.)</td>
<td></td>
</tr>
<tr>
<td>Suldal-Mo</td>
<td>58</td>
<td>14.9</td>
<td>13.0</td>
<td>−1.9</td>
</tr>
<tr>
<td>Stråpa-Sandsa</td>
<td>633</td>
<td>13.2</td>
<td>10.2</td>
<td>−3.0</td>
</tr>
</tbody>
</table>

A change from anticyclonic to cyclonic circulation will impose a stronger fall in temperature at the summer farm levels than in the lowlands (Wishman 1985). This can be illustrated by comparing the mean temperature distribution between Suldal-Mo and Stråpa-Sandsa in August 1976, when the general circulation was mainly anticyclonic with August 1979, when it was generally cyclonic. This change in circulation gives a fall in temperature of 1.9°C at Suldal-Mo and 3°C at Stråpa-Sandsa (Table 3).

According to Lamb (1977) anticyclonic circulation was more frequent over Western Europe during the Middle Ages warm epoch than during the following colder Little Ice Age period, when cyclonic circulation occurred more often. This climatic condition would also apply to western Norway. Consequently, if the mean summer temperatures are lowered by 1°C in the lowlands during the Little Ice Age simulating mean Little Ice Age conditions, it would be reasonable to suggest that it would fall more, e.g. by 1.5°C at the summer farm levels. This would imply that the mean heat sum at Stråpa-Sandsa would be reduced to 1040 degree-days, making cultivation of barley impossible. On the other hand, during the climatically most favourable epoch of the Middle Ages, cultivation of barley was possible at the summer farm level, and a fundamental basis for a permanent settlement at the summer farm level was present. However, with the changing towards a colder climatic phase, barley cultivation became increasingly difficult and eventually impossible. Thereby a fundamental basis for a permanent settlement in the summer farm level disappeared.

Correspondingly, the heat sum at Hamrabø, around 300 m a.s.l., during the most unfavourable part of the Little Ice Age was about 1500 degree-days. Taking into account the general year-to-year temperature fluctuation, cultivation of barley at Hamrabø during the coldest phase of the Little Ice Age would...
be uncertain, but the grains would probably ripen most years.

Conclusion
A climatic deterioration in SW Norway is indicated by pollen analysis and supported by data from cultural history and meteorology. It occurred in the period AD 1300–1800 when ripening of barley was uncertain at the summer farm areas (550–700 m a.s.l.) and presumably also in the farm area Hamrabø (300 m a.s.l.) both of which were probably deserted. A similar climatic deterioration in the High Middle Ages was recorded in Norway by, e.g. Hoel & Werenskiold (1962), Hafsten & Solem (1976), Holmsen (1977) and Nesje & Dahl (1990).

It is reasonable that during the previous period, the 'Little Climatic Optimum', impact of permanent occupation occurred in the summer farm areas based on barley cultivation. An expansion was recorded in Late Iron Age/Middle Ages in the Tengesdal-Lindvang area (Høgestøl & Frøschi-Danielsen 1986; Frøschi-Danielsen 1990). Identification of cereal pollen was interpreted as a plex correlation between the climate and population. Hamrabø became repopulated in the 17th century when the most unfavourable period of the Little Ice Age started. This resulted in a stagnation of the population for about 200 years. About AD 1800 (i.e. after the Little Ice Age maximum) there was an increase in the population and an expansion of agricultural activity also in the summer farm areas. This was indicated by the great number of registered historical monuments including the summer farm buildings.

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Land Registers 1668, 1723. Riksarkivet, Oslo.
National censuses 1769, 1801, 1815, 1825, 1835, 1845, 1875, Statsarkivet, Stavanger and Riksarkivet, Oslo.
(også trykt i Stavanger Turistforenings årbok for 1979.)
Tree-rings of Scots pine (Pinus sylvestris L.) as indicators of past climate in Central Norway

TERJE THUN


A tree-ring chronology back to AD 574 has been developed on Pinus sylvestris from Trøndelag. The pattern of the tree-ring width of this chronology correlates with the curve based on instrumental data for the mean summer temperatures (May-August) during the period 1870-1988. This shows that tree-ring chronologies can unveil detailed palaeoclimatic information.

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Regional tree-ring chronologies

In recent years tree-ring chronologies based on Scots Pine (Pinus sylvestris) have been developed for various parts of Norway. The longest chronology so far goes back to AD 574 and is based partly on material from standing buildings and partly on logs from archaeological excavations in Trøndelag, primarily the medieval city excavations in Trondheim. Approximately 200 samples have been collected within a wide range of Trøndelag, right from the coastal regions in the west and eastwards to the inland areas along the Swedish border. The part of the chronology older than ca. 1300, however, consists mainly of archaeological material excavated in the city of Trondheim. From these excavations approximately 1000 samples have been treated, a fact that demonstrates how easy it has been to provide a sufficient quantity of samples for constructing the chronology. The only exception is the period from ca. 1300 to 1500, for which little building material is available. An obvious explanation is the cessation of all construction activity due to the Black Death epidemic which reached Norway in 1349 and probably lay waste as much as two-thirds of the farms in Trøndelag (Sandnes 1971).

A prolongation of this chronology will hopefully be practicable by our making use of the abundant wooden material present in bogs and lakes in Trøndelag and adjacent counties. Tree-ring chronologies covering the last millennium are nearing completion also for the western and eastern parts of South Norway, viz. Vestlandet and Østlandet, and a standard chronology is under construction for parts of northern Norway as well, viz. Troms.

Climate reconstruction

An interesting aspect of tree-ring chronologies is the possibility of obtaining precise climate data back in time. To illustrate this the tree-ring pattern for Trøndelag for the period from 1870 to 1988 is compared with the instrumental data for the mean summer temperatures (May-August) during the same period, both quoted in a logarithmic scale (Fig. 1). It has become common practice to present tree-ring curves in a logarithmic scale because the non-linearity of the logarithm amplifies the minima and suppresses the maxima of a curve. This is necessary because the annual growth of individual trees might differ due to locational factors and would make comparisons of curves in a linear scale difficult.

The air temperature data referred to above were measured partly in Trondheim, at an altitude of 50 m (1870-1944), and partly at Værnes airport in Stjørdal, 35 km east of Trondheim, at 12 m altitude (1944-88). The correlation between the tree-ring curve and the temperature curve shows a r-value as high as 9.7 (Baillie & Pilcher 1973) and the percentage of agreement 73.7 (Eckstein & Bauch 1969). This demonstrates that the tree-ring chronologies contain important climatic information and, consequently, that a recon-

![Graph](Fig. 1. Comparison between the temperature (May-August tetratherm) from 1870 to 1988 in Trøndelag (lower curve) and the corresponding tree-ring width sequences (upper curve), both shown in a logarithmic scale.)
struction of the past annual summer temperatures is practicable by means of tree-ring studies.

For studying long-term variations one may use mean tree-ring values for a number of years, for instance 20-year or 50-year periods. By employing average tree-ring width for several years one avoids the problem of auto-correlation, i.e. that the climate of one year affects the tree-ring growth during the next. During a summer with good growth conditions extra needles will be produced, which the following year will result in good growth even if the growth conditions that year should turn out to be less favourable.

In a Palaeo-climatic context the pattern or cycle of narrow and broad tree-rings is of interest. The standard chronology for Trøndelag shows a number of such sequences which unveil several shifts in climate during the last 1400 years.

As referred to above, the part of the Trøndelag chronology spanning from ca. 1300 to 1500 is not based on an entirely satisfactory number of logs. This part consists of rather narrow tree-rings, which might reflect less favourable climate conditions. However, since one cannot entirely rule out the possibility of local effects on the growth, the author hesitates for the time being to go further into the climatic implications of the tree-ring chronology. However, publications by other tree-ring researchers, e.g. Briffa et al. (1990) and Schweingruber et al. (1988), demonstrate the palaeo-climatic potential in tree-ring chronologies.

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Annual dynamics of carbon flux in the Barents Sea: preliminary results

PAUL WASSMANN & DAG SLAGSTAD


Mathematical modelling was used to explore the annual variability of primary production and sedimentation between 72° and 80°N in the central Barents Sea during 1981 and 1982 (extensive ice coverage) and 1983 (little sea ice). The interannual sea-ice dynamics of the Barents Sea reflect the variability of the transport of warm Atlantic Water into the Barents Sea. The annual estimates of primary production, and sedimentation decreased on average from 73 to 18 and 48 to 9 g C m⁻² year⁻¹ between the southern and the northern parts of the Barents Sea, respectively. The annual estimates of carbon flux were much higher in 1983 than in 1981–82, especially to the north of 76°N where up to six times higher rates were calculated for 1983 than in 1981–82. The dynamics of sea ice are therefore of utmost importance for the interannual productivity of the Barents Sea. This changing productivity has consequences for the pelagic-benthic coupling, the flux of CO₂ from the atmosphere to the ocean as well as the advection of particulate organic matter in the nepheloid layer into the Polar Ocean and the Norwegian Sea.

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The Barents Sea is a marginal shelf sea of the eastern North Atlantic. It supports one of the richest fisheries of the world's oceans (Fig. 1). The two main water masses, Arctic Water entering the Barents Sea from the northeast and Atlantic Water entering from the southwest, are separated by the Polar Front. The Arctic Water is periodically ice-covered, and the maximum extension of ice is close to the Polar Front in the western and central parts of the Barents Sea, while ice cover is extensive in the eastern parts (Fig. 2). Icemelt during spring and summer gives rise to a strongly stratified and nutrient-rich euphotic zone, with a distinct phytoplankton bloom in the marginal ice zone (Marshall 1957; Bobrov 1985; Sakshaug & Skjoldal 1989). In contrast, in the areas dominated by Atlantic Water, stratification develops slowly due to solar radiation during spring and summer and the resulting phytoplankton bloom is less distinct (Rey & Loeng 1985).

The ice-cover of the Barents Sea varies greatly from year to year (Vinje 1983) (Fig. 2). This variation reflects the interannual dynamics of inflowing Atlantic Water (Midttun & Loeng 1987) which is warm, nutrient rich and transports variable amounts of zooplankton into the southern and central Barents Sea (Skjoldal & Rey 1989). The dynamics of sea ice in the Barents Sea and its suppression of primary production (light limitation) during the short productive period at latitudes greater than 70°N are, thus, capable of influencing the total productivity and sedimentation of organic carbon in the area, both during the course of the year and interannually.

The amount of data gathered by the Norwegian Research Program of Marine Arctic Ecology (PRO MARE) and other investigations

Fig. 1. Map of the Barents Sea with the stations I to IV where simulations were run. Also shown is the area in the central Barents Sea for which estimates of seasonal ice-coverage and primary production were calculated.
during 1979–89 is too small to adequately address the annual or interannual dynamics of carbon flux of a large shelf area like the Barents Sea. However, understanding the large-scale dynamics of carbon flux is nonetheless crucial for understanding the function of this ecosystem, the flux of phytoplankton-derived carbon to the sediment as well as the advection of detritus from the shelf into the deep sea. At present, the only adequate method of addressing these questions is mathematical modelling based on available information on the physical, chemical and biological oceanography of the area. This contribution represents a first step in addressing the dynamics of primary production and sedimentation in the Barents Sea, until more extensive data sets and more model results are available.

The present investigation is based mainly on observations made at a southwest–northeast transect in the central Barents Sea during 1981 to 1983 (Fig. 1). A short overview of the sampling and analytical methods has been presented by Skjoldal et al. (1987). Data have been published elsewhere (Rey & Loeng 1985; Skjoldal & Rey 1989; Rey et al. 1987; Wassmann 1989). A one-dimensional phytoplankton model was used to simulate the dynamics of photoplankton and nutrients as functions of seasonal light intensity, vertical mixing, ice cover, zooplankton grazing (Slagstad 1981, 1982) and sedimentation (Wassmann & Slagstad, in prep.) at four stations along a gradient from 72° (I) to 80°N (IV) (Fig. 1).

The spring phytoplankton bloom is delayed from May/June in the ice-free Atlantic Water until August/September in the northern part of the Barents Sea. The decrease in solar radiation limits primary production already in early autumn.

The impact of sea-ice dynamics on the carbon flux in the Barents Sea is given in Table 1. Annual estimates of primary production and sedimentation decreased on average by 75 and 82%, respectively, between stations I and IV. The annual primary production and sedimentation rates decreased with increasing latitude, indicating limited primary production in the northern Barents Sea due to the sea-ice coverage. Export of organic carbon to the aphotic zone is thus of greater significance in the southern Barents Sea compared to its middle and northern parts. However, significant variation occurred between different years. The annual variation in ice-cover was quite substantial during 1981–83 (Fig. 2). 1981 and 1982 were rich in sea ice while only small amounts of sea ice covered the Barents Sea during 1983, especially during spring. The annual estimates of primary production and sedimentation at station IV were four to six times higher in 1983 than in 1981 and 1982. The annual variation of sea ice appears to have a stronger impact on the flux of carbon.

The integrated estimates of primary production and sedimentation from the transect (Fig. 1) varied between 44 and 65 to 27 and 37 g C m⁻² yr⁻¹, respectively (Table 2). The integrated annual estimates from the transect varied by less than 20%, with highest rates during years with a northerly distribution of sea ice.
ice during the spring, e.g. 1983. The integrated primary production of the central Barents Sea (shaded area in Fig. 1) for the years 1981 to 1983 ranged from 1.3 to 2.0 × 10^7 t C, respectively (average: 1.6 × 10^7 t C ± 23%). The average integrated primary production of the central Barents Sea area during 1981–83 was 50 g C m⁻² yr⁻¹ (Table 2). This rate is only slightly different from the estimate of the average integrated primary production of the transect, i.e. 56 g C m⁻² year⁻¹ (Table 2). This comparison suggests that the annual variability of primary production in the western part and the total area of the Barents Sea is not significantly different, despite the substantial east–west differences in sea-ice coverage during spring and early summer (Fig. 2). This is most probably caused by low annual primary production rates in the northern Barents Sea.

Dense bottom water which accumulates in the deeper parts of the Barents Sea can give rise to pulses of cold water which flow through the various troughs into the Arctic Ocean and the Norwegian Sea (Quadfasel et al. 1988). The formation of this water results from winter cooling of surface water and is thus a function of climatic variability (Midttun & Loeng 1987). The downflow of dense bottom water at the steep shelf break between the Barents Sea and the Norwegian Sea may trigger an avalanche of a large-scale bottom flow over a wider area on the shelf. This flow, in turn, may flush out newly deposited unconsolidated sediment from the shelf and troughs, causing a ‘winter outburst’ transport of particulate matter (Honjo 1990). These ‘outbursts’ are reflected by winter sedimentation rates of particulate carbon, and especially lithogenic matter in the eastern Norwegian Sea, which are far higher than for the central Norwegian Sea (Honjo et al. 1988; Honjo 1990). These results are supported by annual sedimentation rates and hydrographic measurements south and west of Bear Island (Honjo & Wassmann, unpubl.). Transport of particulate material along the shelf break of the Barents Sea and Spitsbergen is also reflected in increased amounts of suspended matter (Kullenberg & Zhenglang 1986). This seems to imply that the Barents Sea supplies the deeper parts the Norwegian Sea with material which gives rise to additional signals in the sediment record. This is probably also true for the Arctic Ocean.

Anomalous periods, with temperatures and salinities below the long-term mean, occurred in the East Icelandic Current in the late 1960s. Similar periods occurred in the Atlantic inflow to the Norwegian Sea during the 1970s (Blindheim 1989). A downstream delay of about three years from the Rockall Channel to the southwest and to the entrance to the Barents Sea was recorded (Blindheim 1989), suggesting that the event is advective. The Barents Sea is thus an integrated part of the Norwegian Sea with interannual, climatically induced variations of Atlantic Water flowing into its southern part. This influences the sea-ice distribution, primary production and sedimentation rates. Occasionally, some of the sedimented phytoplankton derived detritus and other particles in the nepheloid layer and the sediment surface are advected along with dense bottom water into the deeper parts of the Norwegian Sea. These advective fluxes are not necessarily reflected by sediment trap studies in the Norwegian and Greenland Seas (e.g. Bathmann et al. 1990; Bodungen 1989), but influence benthic processes, nonetheless (Graf 1989).

These climatic variations show the dynamic nature of the Norwegian Sea and adjacent Barents Sea. The climatically deteriorating anomalies give rise to biological repercussions which are reflected in the supply of organic carbon to the sediment surface. Climatic changes in the eastern North Atlantic thus play a significant role in the primary production and sedimentation of organic carbon in the Barents Sea and are probably also reflected in variations of advected material into the deeper parts of the Norwegian Sea.

These preliminary results are based on a mathematical model for a transect of the central Barents Sea only. However, a three dimensional model of the carbon flux in the Barents Sea is under development. Mathematical carbon flux models of whole ecosystems represent a valuable method of investigating how climatic short-term and long-term variations and changes in ecosystem structure influence the supply of plankton-derived detritus to the sediment surface and its fate in the benthic boundary layer before these signals are stored in the geological record.

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