Meddelelser om Grønland

Late Quaternary stratigraphy and glaciology in the Thule area, Northwest Greenland

Edited by Svend Funder

Geoscience
22 · 1990

Meddelelser om Grønland

The series *Meddelelser om Grønland* was started in 1879 and has since then published results from all fields of research in Greenland. In 1979 it was split into three separate series:

Bioscience
Geoscience
Man & Society

The series should be registered as *Meddelelser om Grønland, Bioscience (Geoscience, Man & Society)* followed by the number of the paper. Example: *Meddr Grønland, Biosci.* 1, 1979.

The new series are issued by Kommissionen for Videnskabelige Undersøgelser i Grønland (The Commission for Scientific Research in Greenland).

Correspondence

All correspondence and manuscripts should be sent to:

The Secretary
Kommissionen for videnskabelige Undersøgelser i Grønland
Danish Polar Center
Hausergade 3
DK-1128 Copenhagen K.

Questions concerning subscription to all three series should be directed to the agent.

Agent

Nyt Nordisk Forlag – Arnold Busck A/S, Købmagergade 49, DK-1150 Copenhagen K. Tlf. +45 33 11 11 03.

ISSN 0106-1054
ISBN 87-17-05971-2

Meddelelser om Grønland, Geoscience

Meddelelser om Grønland, Geoscience invites papers that contribute significantly to studies in Greenland within any of the fields of geoscience (physical geography, oceanography, glaciology, general geology, sedimentology, mineralogy, petrology, palaeontology, stratigraphy, tectonics, geophysics, geochemistry). Papers primarily concerned with other areas in the Arctic or Atlantic region may be accepted, if the work actually covers Greenland or is of direct importance to continued research in Greenland. Papers dealing with environmental problems and other borderline studies may be referred to any of the series *Geoscience, Bioscience,* or *Man & Society according to emphasis and editorial policy.*

Scientific editor – Geology

T. C. R. Pulvertaft, Institute of General Geology, Øster Voldgade 10, DK-1350 Copenhagen K. Telephone +45 33 11 22 32.

Instructions to authors.
See page 3 of cover.

Late Quaternary stratigraphy and glaciology in the Thule area, Northwest Greenland

edited by

Svend Funder

MEDDELELSER OM GRØNLAND, GEOSCIENCE 22 · 1990

Contents

Late Quaternary stratigraphy and glaciology in the Thule area, Northwest Greenland

EDITED BY SVEND FUNDER

Funder, S. (ed.) 1990. Late Quaternary stratigraphy and glaciology of the Thule area, Northwest Greenland. Meddr Grønland, Geosci. 22, 63 pp. Copenhagen 1990-06-27.

This report contains the geological and glaciological results of the NORDQUA 86 expedition to Thule, Northwest Greenland.

Coastal sections along Wolstenholme Fjord provide a detailed record of glacial and marine events during isotope stage 5 (74–134 ka), on the northern perimeter of Baffin Bay.

The record has been dated by a combination of thermoluminescence and C-14 dating. Amino acid analyses of marine mollusc shells afford local and regional correlations, while periods with penetration of warm subarctic water have been identified by their foraminifer and mollusc faunas. There were two marine episodes with influx of subarctic water. Between the two marine episodes (at 114 ± 10 ka) maximum, although restricted, Weichselian ice coverage was attained. After this, and until the Late Weichselian, ice coverage was similar to or smaller than at present. In Late Weichselian times there was a readvance of glaciers, also associated with influx of subarctic water.

The record provides for the first time a link between events in arctic Canada and Greenland, and shows that in the northern Baffin Bay region there is a causal relationship between hydrography and glaciation.

Samples of glacier ice from two cross sections of the ice margin all show a Holocene stable isotope signal. Although there are several possible explanations for this, this finding is in agreement with the small extent of ice cover during the Weichselian. It also seems likely that the local "Tuto ice dome" did not exist during the Holocene climatic optimum.

Contributors:

H. B. Clausen, Geophysical Institute, University of Copenhagen, Haraldsgade 6, DK-2200 Copenhagen N, Denmark.

R. W. Feyling-Hanssen, Geological Institute, University of Århus, DK-8000 Århus C, Denmark.

P. Frich, The Danish Meteorological Institute, Lyngbyvej 100, DK-2100, Copenhagen Ø, Denmark.

S. Funder, Geological Museum, Østervoldgade 5–7, DK-1350 Copenhagen K, Denmark.

M. Houmark-Nielsen, Institute of General Geology, University of Copenhagen, DK-1350 Copenhagen K, Denmark.

M. Kelly, Dept. of Environmental Science, University of Lancaster, 4LA 1YQ Lancaster, England.

C. Kronborg, Geological Institute, University of Århus, DK-8000 Århus C, Denmark.

J. Y. Landvik, Institute of Soil Sciences, Agricultural University of Norway, P.O. Box 28, N-1432 Ås-NLH, Norway.

V. Mejdahl, The Nordic Laboratory for Thermoluminescence Dating, Risø National Laboratory, DK-4000 Roskilde, Denmark.

N.-A. Mörner, Institute of Geology, Geological Institute, University of Stockholm, S-10691 Stockholm, Sweden.

N. Reeh, Alfred Wegener Institute for Polar and Marine Research, Postfach 120161, D-2850 Bremerhaven, Federal republic of Germany.

H.-P. Sejrup, Dept. of Geology, Sect. B, University of Bergen, N-5007 Bergen, Norway.

L. Sorby, Dept. of Quaternary Geology University of Lund, Sölvegatan 13, S-22362 Lund, Sweden.

H. Højmark Thomsen, The Geological Survey of Greenland, Østervoldgade 10, DK-1350 Copenhagen K, Denmark.

Introduction

SVEND FUNDER

The Baffin Bay region is a climate-sensitive area, and climatic and hydrographic changes in this region have been held responsible for the initial formation and growth of the Laurentide Ice Sheet over North America during the last ice age.

The present report describes results of a study in the Thule area, northwest Greenland, which yielded new evidence on this issue, and for the first time provides a link between events in North America and Greenland at the beginning of the last ice age. The evidence supports the theory that in this region glaciation is associated with influx of warm subarctic water along the coast.

The results are a product of the "NORDQUA 86 Project", based on field work carried out in August 1986 by 33 Quaternary scientists from the five Nordic countries and Britain, under the auspices of NORDQUA – the association of Quaternary scientists from the Nordic countries. The organisers were S. Funder, then President, and Kaj Strand Petersen, General Secretary of NORDQUA.

Preliminary results have been presented in a series of papers at the 18. Nordic Geologic Winter Meeting in Copenhagen (Feyling-Hanssen 1988; Funder 1988; Houmark-Nielsen 1988; Mörner 1988; Sejrup 1988; Sorby 1988; Reeh & Thomsen 1988), and NORDQUA 86 Participants (1989).

Although responsibilities for interpretation and conclusions lie with the authors of each section, it must be noted that the results are due to all participants (see below).

Participants in the NORDQUA 86 expedition

Björn Andersen, Dept. of Geology, University of Oslo, Norway
Roar Austad, Geological Institute, University of Bergen, Norway
Cato Erichsen, Dept. of Geology, University of Bergen, Norway
Karin Ericson, Dept. of Quaternary Research, University of Stockholm, Sweden.
Poul Frich, The Danish Meteorological Institute, Copenhagen, Denmark
Svend Funder, Geological Museum, University of Copenhagen, Denmark
Morten Funder, Hillerød, Denmark
Michael Houmark-Nielsen, Institute of General Geology, University of Copenhagen, Denmark
Olafur Ingolfsson, Dept. of Quaternary Geology, University of Lund, Sweden
Marit Karlsen, Geological Institute, University of Oslo, Norway
Michael Kelly, Dept. of environmental Science, University of Lancaster, U.K.
Ove Klakegg, Geological Survey of Norway, Trondheim, Norway.
Lars Einar Kringstad, Dept. of Geology, University of Bergen, Norway
Christian Kronborg, Geological Institute, University of Århus, Denmark.
Lars-König Königsson, Dept. of Quaternary Geology, University of Uppsala, Sweden
Robert Lagerbäck, Geological Survey of Sweden, Upssala, Sweden.
Jon Y. Landvik, Agricultural University of Norway, Ås, Norway
Jan Lundqvist, Dept. of Quaternary Research, University of Stockholm, Sweden
Jan Mangerud, Dept. of Geology, University of Bergen, Norway
Arve Misund, Dept. of Geology, University of Bergen, Norway
Nils-Axel Mörner, Geological Institute, University of Stockholm, Sweden
Mats Olvmo, Institute of Physical Geography, University of Göteborg, Sweden
Kaj Strand Petersen, Geological Survey of Denmark, Copenhagen, Denmark.
Mikko Punkari, Dept. of geology, University of Helsinki, Finland
Niels Reeh, Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Federal Republic of Germany.
Arne Reite, Geological Survey of Norway, Trondheim, Norway.
Lars Rodhe, Geological Survey of Sweden, Upssala, Sweden.
Lars Eirik Skylvik, Dept. of Geology, University of Bergen, Norway
Oddmund Soldal, Dept. of Geology, University of Bergen, Norway
Lennart Sorby, Dept. of Quaternary Geology, University of Lund, Sweden
Joar Sættem, Continental Shelf and Petroleum Technology Research Institute a/s, Trondheim, Norway
Henrik Højmark Thomsen, Geological Survey of Greenland, Copenhagen, Denmark.
Morten K. Thoresen, Geological Survey of Norway, Trondheim, Norway.

Previous studies of Quaternary geology

Observations on geomorphology, and its relevance to earlier glaciation in the Thule area were discussed by Chamberlin (1895), Koch (1928) and Schytt (1956). A comprehensive study and mapping of the Quaternary geology in the region here dealt with was carried out by Krinsley (1963). Non-finite C-14 ages, the first in Greenland, from a coastal cliff on Saunders Ø played an important part in the stratigraphical outline, showing that although the entire region had been overridden by ice, the ice cover began to melt before 32 000 yrs BP and "marine till" with *in situ* molluscs was deposited in front of the receding Wolstenholme Fjord glacier. This was followed by sea-level rise which reached a culmination before 8500 yrs BP, and after this the glacier receded up to 24 kilometres behind its present boundary. Goldthwait (1960) and Crane & Griffin (1954) obtained C-14 dates from Nunatarssuaq on the north side of Wolstenholme Fjord supporting the early recession, and showing that the present extent of glaciation was attained before 9000 yrs BP. This was supported by later C-14 dates from Thule and Narssârssuk (Weidick 1977).

Blake (1975, 1977, 1987) revisited and reinterpreted the sections on Saunders Ø, and made the important observations that below an upper till bed there are marine sediments with a thermophilous mollusc fauna which again overlie an older bed of till. The marine sediments had non-finite C-14 ages, and were correlated with a similar occurrence on Coburg Island to the east of Thule, and referred to the Sangamonian interglaciation (Blake 1973: 56).

At Qarmat in the interior Wolstenholme Fjord, Kelly (1980a, 1986) found a somewhat similar sequence of marine sediments with thermophilous molluscs, non-finite C-14 age, and amino acid ratios comparable to those of occurrences in Melville Bugt to the south, and correlated them with the "intra last glaciation" Kogalu aminozone of Baffin Island (Miller *et al.* 1977).

From these studies it emerged that the Thule region, and especially the coastal cliffs on Saunders Ø, had a longer and more complex record of Late Quaternary glacial and marine events than known elsewhere in West Greenland, and in general stratigraphic reviews the Saunders Ø occurrence attained a key role for the Eemian-Weichselian stratigraphy of West Greenland (e.g. Funder 1984, 1989; Kelly 1985, 1986), a role that probably had not been anticipated by the field workers.

With this background one objective of the NORDQUA 86 expedition was to make detailed sedimentological investigations supplemented by modern dating techniques at the sites on Saunders Ø and Qarmat. During the field work a third site for pre-Holocene sediment, Narssârssuk, was discovered.

Methods and organisation of this report

The field work comprised two projects: Quaternary geological work on the ice free land, and glaciological work on the adjacent ice sheet. The work on land was concentrated on three sites where coastal sections in raised alluvial cones give a detailed record of isotope stage 5 (74–134 ka) marine and glacial events (Fig. 1). The results of lithologic logging at these sites are described in the first section, and summarised in Figs 4, 8 and 9. Periods of cold and warm water influx are identified by micro- and macropalaeontological analyses, as described in the second section, and summarised in Fig. 11 and Tables 4, 5 and 6. Absolute dates are provided by thermoluminescence and C-14 ages, described in the third section and appendix, and summarised in Table 9. The third section also gives a description of amino acid analyses of marine bivalve shells, essential for correlating between sites, and listed in Table 10. Finally, in the fourth and fifth sections this evidence is combined into an event-stratigraphy (Fig. 24), and compared to that from adjacent areas in Baffin Bay (Fig. 27).

Throughout this part of the work samples are referred to by their field numbers and the lithologic units to which they belong. The location of samples and lithologic units appear from Figs 4, 8 and 9.

The sixth section deals with glaciological work aimed at unravelling the dynamic and climatic history of the Northwest Greenland ice cover by studying stable isotopes in surface samples of glacier ice. Two sites were selected for this work: one on the Inland Ice margin north of Wolstenholme Fjord, and the other on the margin of the "Tuto ice dome" south of the fjord.

Chronostratigraphic nomenclature follows the recommendation of Mangerud *et al.* (1979), and isotopic substage 5e (123–130 ka) is regarded as equivalent to the Eemian interglaciation, while the Weichselian glaciation covers isotopic stages 2–5d (12–123 ka). Ages cited for isotopic stages are from Martinson *et al.* (1987).

Acknowledgements

There is no precedence for bringing a large group of scientists to work on a single project in this part of Greenland, and had it not been for the extraordinary goodwill and helpfulness offered by many persons and institutions, the project would have foundered long before the departure from Copenhagen.

During the planning phase the Commission for Scientific Research in Greenland, and Sven Adsersen, of the then Ministry for Greenland cleared many obstacles for us. The Geological Survey of Denmark kindly undertook the printing of an 80 page excursion guide "Istiden i Thule – Nordqua 1986, Grønland".

The field work took place in the period 7-21.8. 1986, and in Thule smooth cooperation with the air base authorities was handled efficiently by the Danish liaison officer, Commander Erik Thomsen and his staff, whose advice and help was essential for the success of the field work in spite of bad weather and ice-filled waters. Also the help from Hans J. Østergaard, commissary, was important for our general well-being. A major problem was the transportation within the area of such a large group of people. Here we had the unexpected luck to be able to use a U.S. Army landing vessel under the command of sgt. W. Turner, as well as the Danish naval vessel Tuluaq.

Financial and logistic support for the field work was provided by The Geological Survey of Greenland, The Geological Survey of Denmark, The Joint Committee of the Nordic Natural Science Research Councils, The Danish Natural Science Research Council, and The Commission For Scientific Research in Greenland.

During the important work-up phase the participants' institutions provided facilities and technical assistance. In addition, C-14 dates were provided as part of Nils-Axel Mörners research project from Laboratoriet för isotopgeologi at Naturhistoriska Riksmuseet, Stockholm, while others were supplied by the C-14 Laboratory of The Geological Survey of Denmark and the National Museum, Copenhagen – by courtesy of The Geological Survey of Denmark, and the Physical Institute, Århus University. The Polar Continental Shelf Project, Ottawa, Canada, and its former director George Hobson, made Niels Reeh's participation and the glaciological programme possible. Substantial funds for TL dating were obtained from the Danish Natural Science Research Council.

On 4th–7th May 1987 the contributors to this report discussed NORDQUA 86 results during a meeting at Mønsted Field Station, Denmark. Funds for this important meeting were provided by the Commission For Scientific Research in Greenland, and the Joint Committee of The Nordic Natural Science Research Councils.

Finally, we wish to thank J. T. Andrews, D. Fisher, S. Foreman, R. LeB. Hooke, Karen Luise Knudsen, J. Lundqvist, G. H. Miller, and an anonymous referee for valuable and constructive comments on the manuscript.

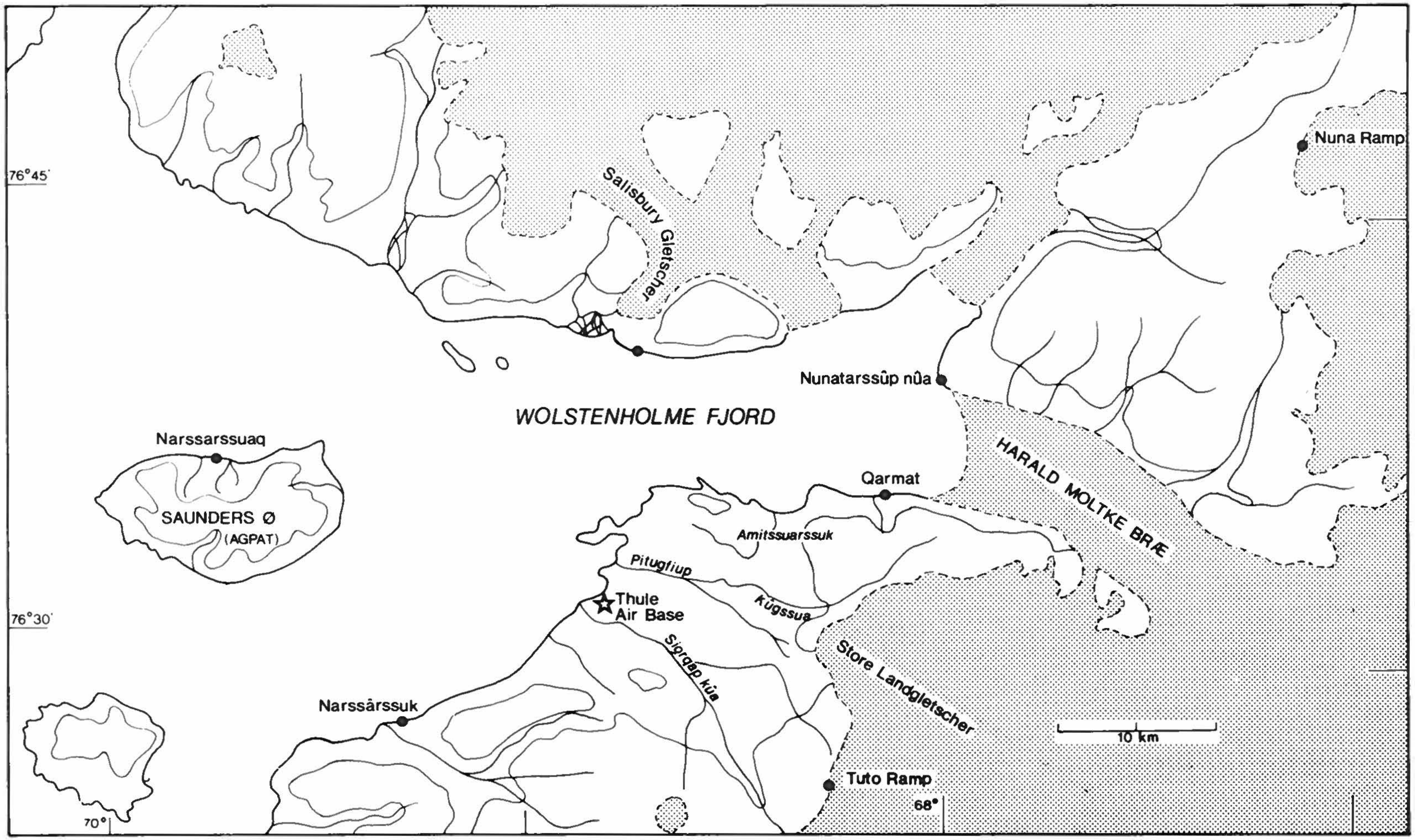


Fig. 1. The Thule area with localities described in this report. Black circles are principal studied sites. Present ice cover is shaded, thin line shows 300 m contour.

The pre-Holocene Quaternary: Lithostratigraphic and geomorphic evidence

MICHAEL HOUMARK-NIELSEN, MICHAEL KELLY, JON Y. LANDVIK and LENNART SORBY

Detailed lithostratigraphic evidence has been obtained from sections excavated in pre-Holocene deposits in three areas: Saunders Ø, Narssârssuk and Qarmat. The geology and geomorphology of the areas surrounding these sites has provided additional evidence.

Saunders Ø

(Michael Houmark-Nielsen and Lennart Sorby)

General site description

The island of Saunders Ø is situated at the mouth of Wolstenholme Fjord, about 20 km west of Thule Air Base (Fig. 1). The island forms a plateau at 300 m above sea-level with steep coastal cliffs. It is built up of Pre-cambrian sedimentary rocks (Davies *et al.* 1963, Dawes *et al.* 1982). A narrow beach is locally present. A thin till cover drapes the plateau with frequent boulders of crystalline rocks, showing that the island has been over-ridden by a glacier from the mainland. An embayment along the north coast at Narssarssuaq has been filled with Quaternary sediments belonging to a system of prograding alluvial fans (Fig. 2). In its western part, Narssarssuaq consists of a plateau at about 30 m above sea-level with raised beach ridges, the surfaces of which dip gently eastwards. An alluvial plain ending in a re-cent beach ridge and originating from two canyons, cuts off the plateau with beach ridges in the eastern area. A 20–30 m high coastal cliff exists along the 800 m long western coast of the plateau (Fig. 3), and reveal moder-ately eastward dipping strata composed of till, glacio-

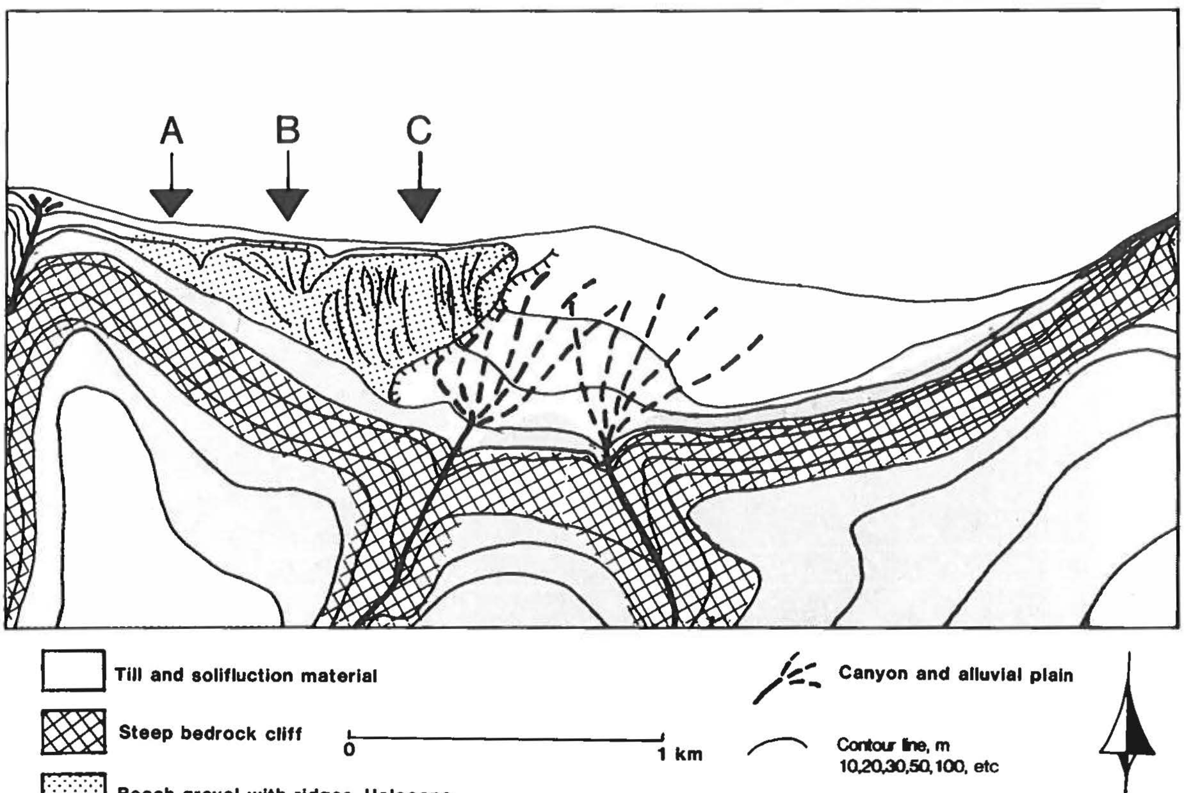


Fig. 2. Quaternary geology of Narssarssuaq, northern Saunders Ø. Letters mark described sections illustrated in Fig. 4. (Topography drawn by computer-assisted air photo stereometry by O. Winding, Geological Survey of Greenland).

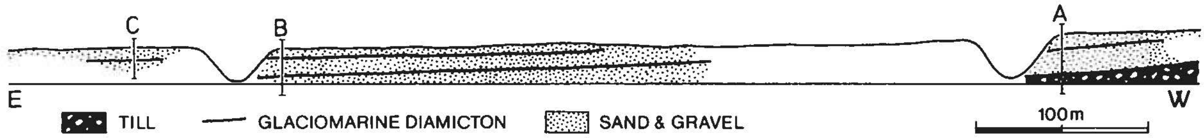


Fig. 3. Narssarssuaq, Saunders Ø, general stratigraphy of the coastal cliff. The section shows three successive transgression-regression cycles, each starting with glaciomarine diamicton or till. Letters mark described sections illustrated in Fig. 4.

marine diamictons and marine sand and gravels, as described in detail below.

Krinsley (1963), who previously investigated the cliff site, concluded that "marine till" is overlain by marine sand and beach deposits with a combined thickness of about 30 m. The upper half of the sequence was ascribed to the Holocene, whereas the lower half gave non-finite C-14 age. Observations by Blake (1975) indicated the presence of two diamict units interbedded and overlain by marine sand and gravel. A lower shell-bearing bouldery till is overlain by a sandy to gravelly coarsening upwards sequence. A thin till bed that overlies these strata was ascribed to the last glaciation. This till bed is covered by another muddy, sandy to gravelly coarsening-upwards sequence deposited during the Holocene. Blake (1975) reported non-finite C-14 ages from shells beneath the upper till bed. The need to reconcile the opposite viewpoints of Krinsley (1963) and Blake (1975) about the genesis of the diamict beds and subsequent glaciation history was one of the major tasks undertaken by the NORDQUA reinvestigation of the cliff.

Field observations

Detailed investigations of the sediments in three sections in the coastal cliff (Figs 2 and 4), and in two complementary sections in erosion valleys perpendicular to the coast, were undertaken. Sections A and C are those investigated by Blake (1975). Seven major sedimentary units, each composed of individual facies or facies associations, have been recognised. Facies associations largely comprise till, glaciomarine ice-rafted diamictons and marine mud, sand and gravels. The coastal cliff was partly snow-covered, preventing full visual correlation between diamict units as pictured in Fig. 3. However, field evidence supported by a number of laboratory analyses suggests the stratigraphic relationship shown on Fig. 4, and described below.

Lithostratigraphy

Unit S1 – This is a diamicton found in the lower part of section A. It could not be investigated in detail because it was covered by scree. The diamicton consists of a muddy matrix packed with stones and boulders primarily of crystalline basement rocks, and is the oldest visible sediment in the sequence (Fig. 3). A glacigenic origin, which is in accordance with Blake's interpretation, is ascribed to this unit because of its massive character with dominating exotic boulders and the lack of *in situ* shells. It was deposited either by melt-out or gravity flow processes from an active or stagnant glacier descending from the mainland towards the east.

Unit S2. – This unit overlies unit S1 and consists of a complex facies association. As a whole this unit is built up of muddy or sandy diamictons interbedded with marine shell-bearing mud and lenticular bodies of sand. A transitional facies at the boundary with unit S1 is thin matrix-supported gravel, grading into gravel with subrounded clasts. The matrix is sand grading to sandy silt with shell fragments.

Above the gravel, the lower part of unit S2 consists of a brown sandy-silty diamicton rich in subangular clasts with low sphericity. The sediment contains only few shell fragments. It is overlain by grey clayey and sandy mud with numerous shell fragments and occasional mollusc shells and a few clasts. It is interbedded with lenses of clay and muddy diamicton also containing shell fragments. Dropstones featuring both bottom and top contact structures (Thomas & Connell 1985), and clasts draped with clay laminae have been found in the sediment.

The upper part of the unit consists of massive beds of silt with clasts laterally grading into silty diamictons, and interlayered with silty diamicton and sand beds. Shell fragments occur mainly in the sand beds.

The clasts in this sediment are a mixture of local bedrock and far transported exotic basement rock types. The facies of unit S2 were deposited by gravity flow processes, rain-out from suspension of marine mud and drop and dump of ice-rafted debris. The overall muddy appearance and the presence of *in situ* shells in combination with ice drop structures, suggests deposition at some depth in a glaciomarine environment with influx of icebergs. This interpretation is contradictory to that of Blake (1975), who apparently combined our unit S2 with the underlying till bed (our unit S1), and described them both as shell-bearing till.

Unit S3. – A thin bed of normal-graded sandy diamicton forms the transition upwards into unit S3. It disconformably overlies unit S2 and comprises a thick sequence of stratified material with inclined alternating beds of clast-supported gravel and stratified sands, dipping about 30° NNE and containing shell fragments. Clasts of local origin totally dominate the material. Towards the top, unit S3 is made up of horizontally bedded sand and coarse gravel. Unit S3 is interpreted as foreset beds in a coarse clastic submarine delta fan. The clasts were probably derived locally through canyons to the west, and mixed with redeposited marine components during delta progradation. As the beds of the growing fan reached wave base, reworking of topset beds into littoral gravels shaped the uppermost part of this unit.

A gully perpendicular to the coast exhibits a complex and muddy diamicton in the otherwise coarse clastic unit S3. It consists of a lower bed of clayey silty diamicton with far-transported clasts and shell fragments. It is conformably overlain by a thin sand bed rich in shells, which in turn is covered by a massive silt, rich in shell fragments and with bladed clasts of local character. Laterally, towards the coast, the diamicton wedges out and is replaced by sand. This subunit is probably a glaciomarine deposit, in its lower part deposited by rain-out of mud from suspension, gravity flows and ice-rafted debris. It

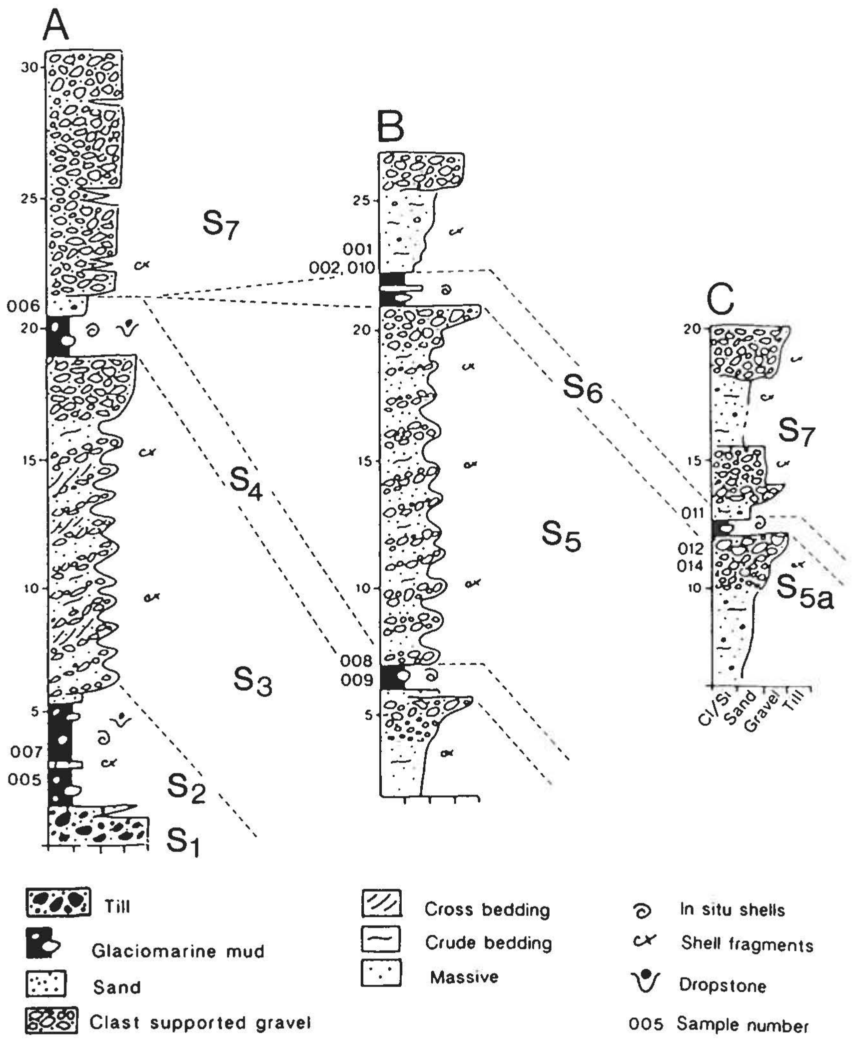


Fig. 4. Lithological sections and correlation at Narssarssuaq, Saunders Ø. Small numbers are m above sea level. Sample numbers refer to samples described in later chapters.

passes upwards into a gravity flow deposit. Unfortunately, the stratigraphic relationship of these sediments to units exposed in the coastal cliff is not fully understood.

Unit S4 – This unit conformably overlies unit S3. It is a horizon of crudely bedded diamicton and shell bearing mud and sand. The diamicton facies is silty with clasts of mainly local origin, although some are far transported. Dropstone structures sometimes occur in connection with clasts. Shell fragments occur in all facies. The unit forms the upper diamicton in section A, and the lower one in section B. It was deposited from gravity flows, rain-out from suspension of marine mud and ice-rafted debris. Sand was deposited by traction currents.

Unit S5. – This unit disconformably overlies unit S4. It consists of a thick coarsening upward sequence of alternating beds of clast-supported gravel and stratified sand. The beds contain shell fragments and dip 15° N. Unit S5 is found in the sections B and C. The interpretation is as for unit S3.

A matrix-poor gravel facies of littoral origin, which is gradually better sorted towards the top, occurs only in section C (subunit S5a), and is characterised by its abundant fragments of *Mytilus edulis* shells which gave a non-finite C-14 age (Blake 1975, 1987). TL-dating indicates that this is the youngest pre-Holocene sediment in the Saunders Ø sequence (see later sections by Funder and by Kronborg, Mejdahl & Sejrup).

Unit S6. – This unit is found above unit S5 in sections B and C. It is a muddy diamicton with mainly local clast material, containing shell fragments and sometimes crudely bedded mud and sand lenses. The diamicton is overlain by homogenous silt and laminated silt and sand. West of section B the unit thins out and is eroded away. This diamicton was described as "till-like sediment" by Blake (1975), and we suggest a glaciomarine origin because of the similarity with the older ice-rafted diamictons (units S2 and S4) in associations and sedimentary features such as dropstone structures, suspension fall-out, gravity flows and *in situ* shells. C-14 dating indicates an Early Holocene age, although TL-dating suggests an older age (Kronborg & Mejdahl, and appendix, this volume).

Unit S7. – This unit constitutes the uppermost facies in all sections and disconformably overlies the units below. It consists of a thick sequence of clast-supported sand and gravel. On the surface of the alluvial plain it forms several beach-ridges. In the poorly exposed section C there seems to be a gradual transition from unit S6 to S7. Shell fragments are frequent, and unit S7 is interpreted as littoral gravel, including both supra- and sublittoral facies.

Interpretation

Unit S1 was deposited by an ice stream originating to the east, filling up Wolstenholme Fjord. The till is provisionally correlated with the heavily soliflucted till that covers the plateau of the island, and signifies the last occasion when glaciers reached Saunders Ø, the Agpat stade (Houmark-Nielsen, this volume). During the melting of this ice sheet, the transitional facies between units S1 and S2 was deposited by melt-out from stagnant or active ice and redeposited by gravity flow processes in a near-shore submarine environment.

Unit S2 was deposited under glaciomarine conditions after melting of the ice on Saunders Ø. Glaciers had retreated eastwards and faunal analyses (Feyling-Hanssen & Funder, this volume) indicate marine conditions as favourable as those of today. Eventually a prograding delta (unit S3) was built out from the island, interrupted however by one or several phases of more distal marine sedimentation during halts in delta progradation. Delta sedimentation ceased and topset beds were reworked into littoral gravels.

Unit S4 represents a renewed transgression phase with sedimentation of glaciomarine mud and ice-rafted debris. Foraminifer assemblages indicate transition from high arctic to less extreme conditions (Feyling-Hanssen, this volume), suggesting proximity of a glacier when the base of this unit was deposited. Sea-level probably stood more than 20 m above present.

This transgression was followed by regression and deposition of coarse gravel by local streams and reworking of marine sediment, which accumulated as submarine fans ending up with littoral gravels, unit S5.

The marine clayey sediments in unit S6, reflect a third transgression followed by regression represented by sandy gravel with beach-ridges, unit S7. TL-dates, amino acid ratios, radiocarbon ages suggest that the sequence in units S6-S7, is of Late Weichselian to Early Holocene age (Kronborg, Mejdahl & Sejrup, and appendix, this volume). The underlying units, S2 to S5, are defined on similar grounds within the Thule aminozone (Sejrup, this volume), TL-dated to the time span 136–69 ka. Consequently the Agpat glaciation (unit S1) was older than this.

Narssârssuk

(Michael Kelly and Jon Y. Landvik)

A complex of Quaternary deposits occupies the low ground at the mouth of the Narssârssuk valley, and extends for some distance along the foot of the bedrock cliffs which bound the southern side of Wolstenholme Fjord (Fig. 5). Good exposures occur in these deposits in low cliffs along the coast. Older deposits and landforms cover the surface of the surrounding slopes and hill tops.

The Narssârssuk upland areas

The plateau areas, which extend up to 300–400 m on both sides of the Narssârssuk valley, are extensively covered with till, and bedrock exposures are rare. Where examined on the north side, the till is predominantly rich in coarse clasts with a sandy silt matrix. Erratic metamorphic rocks are frequent, as well as the local mudstones, sandstones and dolerites. The tills are highly cryoturbated and soliflucted. Weathering of the deposits, however, is only moderate, with granular surface roughening of metamorphic, dolerite and sandstone clasts, some disintegration of coarse gneisses, iron mobilisation and staining of basic rock clasts, and a general shallow oxidation of the deposit.

Relict ice margin features occur around the plateau in the form of benches up to 30 m high and low ridges and channels, all of which are much degraded by solifluction. Along the western edge of the plateau are low boulder ridges and channels (Figs 5 and 6) related to a +400 m thick ice stream in Wolstenholme Fjord.

Lowland areas

North of the Narssârssuk river is a massive body of deposits, extending up to 113 m a.s.l., which partially blocks the mouth of the U-shaped valley. Basically, the deposit consists of the remnants of a terrace 40–50 m above the river, and a major moraine ridge to the north (moraine I, Figs 5 and 7). The terrace surface is covered by sandy tills, and has a gradual slope in the up-valley direction. It exhibits shallow depressions and low ridges interpreted as meltwater channels and thermokarst collapse features. The morphology of this whole body suggests it is an ice-contact complex, formed as a moraine ridge and a kame terrace by a glacier in Wolstenholme Fjord, at a time when the valley was ice-free. On the southwest side of the river, breaks of slope may represent the continuation of this ice marginal deposit (Fig. 5). A second feature that can be related to the presence of ice in the fjord is a discontinuous moraine ridge up to 70 m a.s.l. which parallels the coast (moraine II, Figs 5, 7, 10). We propose to name these features the Narssârssuk moraines.

The deposits forming the main body of the ice contact feature consist of a complex of tills and sorted sediments >50 m thick. Their stratigraphy has not been worked out in detail, but on the steep slope adjacent to the river it appears to consist of five units (a-e, section K, Fig. 8), including an upper and lower series of diamictons alternating with sorted sediments with sands and gravels overlying silts. The lowest unit (d) of the upper diamicton, which outcrops at 65 m above sea-level, is notable in being rich in fragmented bivalve shells and having a matrix of grey clayey silt, in which the silt itself appears to be present as clasts. The unit is interpreted as a till formed from reworked marine silts.

Silts with fragmented shells also floor the flat-lying

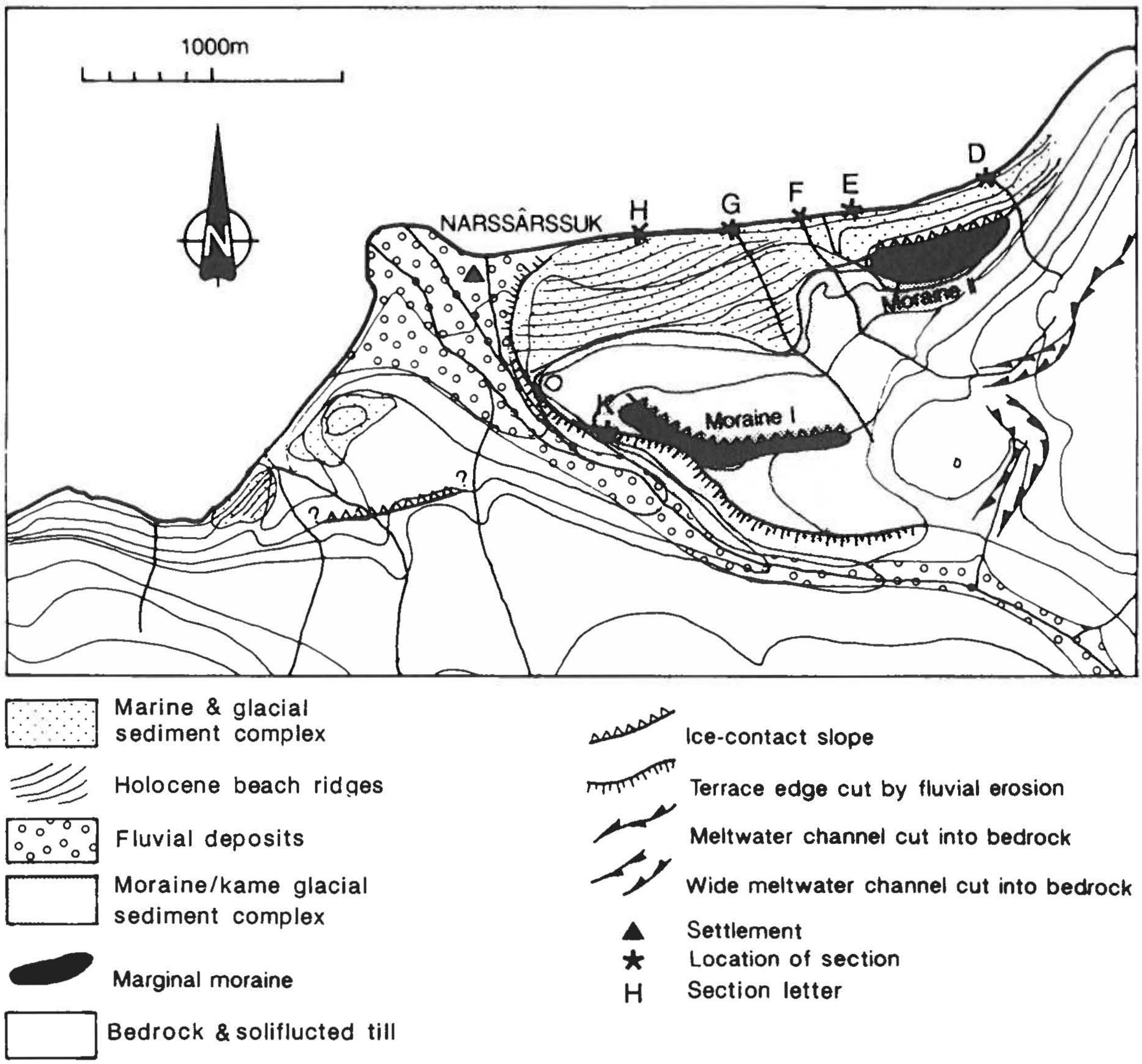


Fig. 5. Map of the Narssârssuk area showing major Quaternary features. Letters refer to sections described in text. Moraines are shown on Fig. 7. (Topography drawn by computer-assisted air photo stereometry by O. Winding, Geological Survey of Greenland).

area at *c.* 50 m a.s.l., below the steep northern side of moraine I (Fig. 5).

The clasts in these tills and gravels are the same suite of rocks as in the plateau tills, and the surface of the deposit exhibits the same degree of weathering. South of the river, exposures at the coast reveal 9 m of shell-free laminated sands with graded bedding, on 4 m of till. These may be proximal glacio-lacustrine turbidites.

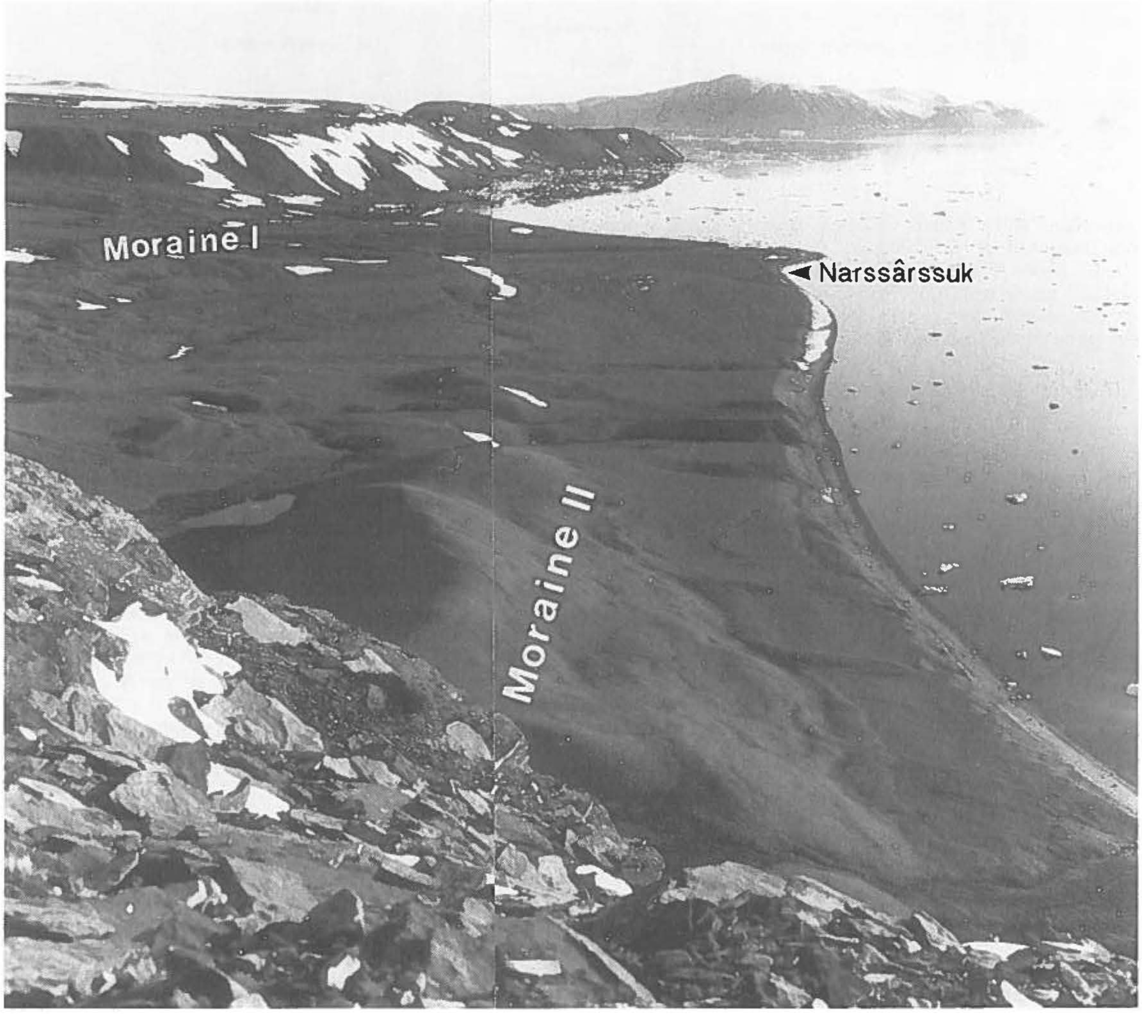
Lithostratigraphy

Exposures along the 2 km of low coastal cliffs show a complex stratigraphy with considerable lateral variation. In the excavated and measured profiles six units (N1 – N6) can be identified (Fig. 8). Most of these main units can be correlated laterally by following their boundaries across poorly exposed ground as breaks in slope and sediment colour changes.

Unit N1. – The lowest unit is exposed at the base of section E; it is a compact, over-consolidated diamicton with many clasts in a predominantly silty matrix. The diamicton is rich in shell fragments and, as with the unit in the ice contact complex, it is interpreted as a till largely made of reconstituted marine sediment. The presence of whole *Hiatella arctica* shells in this unit in a nearby, incompletely investigated profile may indicate a larger scale slice of derived material, or even the presence of less disturbed, *in situ* marine muds.

Fig. 6. Large spillway c. 200 m above sea-level, formed by drainage from Wolstenholme Fjord into Narssârssuk valley (Photo J. Landvik).

Fig. 7. The Narssârssuk area looking towards the southwest. Moraines I and II (Narssârssuk stade) and localities of sections described in text are marked. (Photo K. S. Petersen).



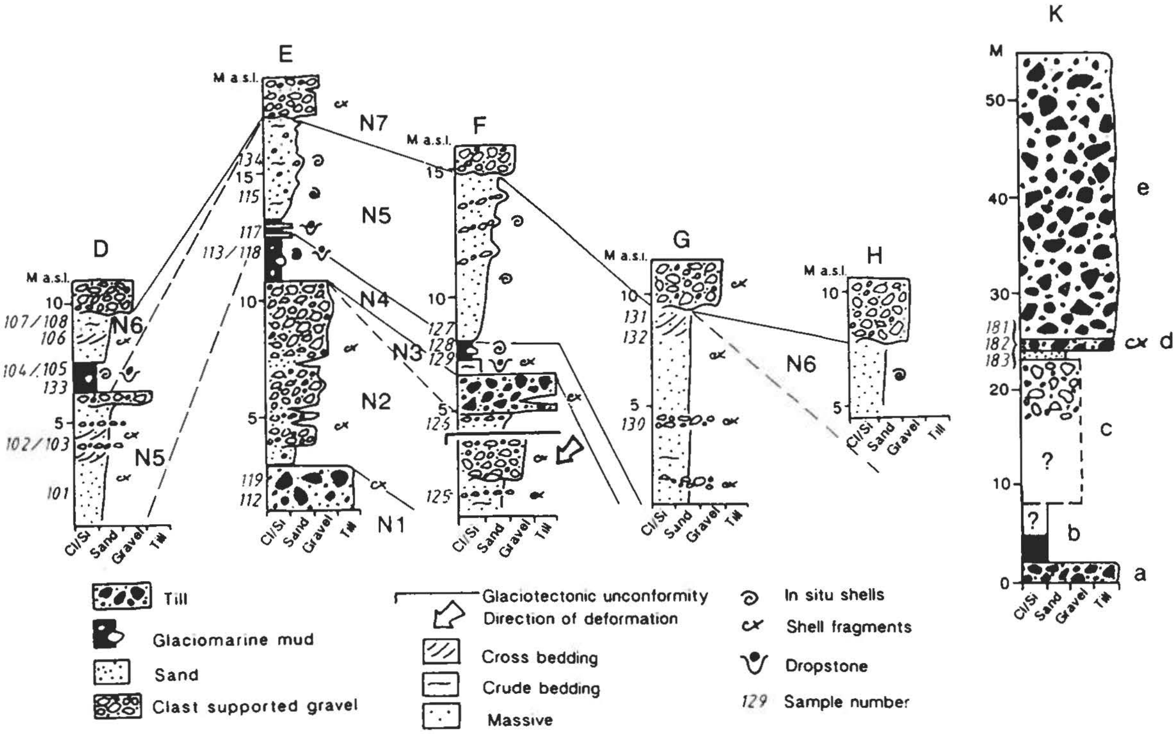


Fig. 8. Lithological sections from Narssârssuk. Solid lines are correlations traced in the field. Small numbers are m above sea level. Notice difference in vertical scale for section K. Sample numbers refer to samples described in later chapters.

Unit N2. – This is a thick sequence of coarse clastic sediments, predominantly crudely bedded gravels and boulder gravels. Exotic rocks form a significant portion of these, indicating glacially related transport from the inland areas. The sediments are locally glaciotectonically deformed (Fig. 8, section F), and the trend of the fold axis shows ice push from the fjord.

Unit N3. – This unit is a gravelly sandy diamicton containing large boulders and shell fragments which is interpreted as a basal till, exposed only in section F. This till unconformably overlies the gravels of unit N2, and wedges out to the northwest against the thicker development of the gravels. The till and the top of the gravels (unit N2) can be correlated by the high boulder content at this level of the cliff, and the boundary with the overlying clayey silts (unit N4, Fig. 8).

We consider that units N1, N2 and N3 are genetically related glacial and proximal glacial facies. The latter were probably submarine, judging from the evidence of the overlying marine unit (see below). North of sections E and F the gravels (unit N2) pass laterally first into a poorly investigated, thick sequence of bedded sands, variously dipping to the northeast and southwest, and then back into bedded gravels.

Unit N4. – This unit consists of 3 m of clayey silts with clasts which become less frequent upwards (N4). Whole *Hiatella arctica*, partly in life position, occur in several horizons, the lowest *c.* 30 cm above the base of the unit. These are associated with thin sand laminae with rust coloration indicating that decrease in mud supply and more aerobic conditions favoured colonisation by a benthic fauna.

Unit N5. – Conformably overlying unit N4 is a thick sequence of sands beginning with laminar bedded silty sands with faunas in life position of *Mya truncata* and *Serripes groenlandicus* together with burrow traces. The dominant sand grain size becomes coarser upwards in Sections E and F and gravel layers occur, probably representing lag gravels from periods of wave erosion. Internal structures include planar and cross lamination marked by heavy mineral laminae, colour mottling suggestive of bioturbation and sparse trace fossils in the form of vertical burrows with escape structures, all of which suggest a shallow-water marine depositional environment.

In section G the sediments are also interbedded coarse and fine sands and thin gravels. At the top, 1.4 m of delta-bedded sands with laminae of organic detritus include mosses. Individual beds are also rich in *Mytilus edulis* fragments.

Occasional clasts, presumably dropstones, occur throughout the sequence.

Unit N6. – This unit is exposed at each end of the series of sediments. It is composed in its lower part of a bed of glaciomarine mud with shell fragments and large clasts, grading upwards into laminated sandy silt with *Hiatella arctica* in life position. Amino acid ratios indicate a Late Weichselian/Early Holocene age for this unit (see Sejrup, this volume).

Unit N7. – The uppermost unit in most of the coastal exposures is a sandy gravel, the surface of which forms a gravel pavement. In the central sector this pavement has low coast-parallel beach ridges which extend up to a minimum altitude of 36 m a.s.l., where they are overridden by solifluction. These are considered to be Holocene in age, formed by the reworking of the older sediments during the Holocene transgression/regression. In section H, sands with abundant mollusc shells have given Holocene radiocarbon dates (see appendix), and may be referred to this unit, which together with unit N6 seems to represent a Late Weichselian-Early Holocene transgression-regression cycle.

Interpretation

Based on the stratigraphy in the sections at Narssârssuk and the geomorphic evidence from the area, the following geological history is suggested:

1. The till deposits and features on the plateau clearly belong to a phase of retreat of an ice sheet which must have essentially covered the whole area. Krinsley (1963) mapped these plateau tills out to the south west coast at Kap Atholl and this ice sheet phase is correlated with the Agpat glaciation (see Houmark-Nielsen, this volume). Retreat of the ice sheet resulted in the deglaciation of the land and area, whilst the 200 m deep Wolstenholme Fjord trough was still filled with ice. The large spillways running from the fjord side of the mountain into the Narssârssuk valley (Figs 5 and 6) were probably formed during this retreat. Identification of the Agpat glaciation in the lowland and coastal sections is necessarily speculative. However, the stratigraphically lowest glacial unit found, i.e. the till at the base of section K, may belong to this ice sheet phase.

2. The coarsening-upwards sequence, revealed both in section K and in the coastal sections (units N1-N3), suggests a readvance of ice in the fjord in the Narssârssuk stade. The lateral margin of the glacier lay across the mouth of the valley (moraine I), where it first overrode glacial lake sediments and marginal outwash, to finally lay down tills, which included marine sediment carried up from the fjord floor. The lithologically similar reworked marine sediment at the base of the coast sections (unit N1) may correlate with this event.

The gradual retreat of the ice margin from the coastal region produced the complex of landforms (Fig. 5), and the ice-proximal deposits found in the coastal exposures (Fig. 8); the till (units N1 and N3), coarse gravels (unit N2) and turbidites. We assume that deposition of the till (unit N3) occurred as a minor readvance during this phase, which is also in accordance with the direction of deformation seen in the underlying gravels, which is from the fjord.

3. The succeeding coarsening-upwards sequence of marine sediments (units N4 and N5), laid down after the last retreat of the ice margin, was a consequence of the change in sediment supply, from ice front to river and coast erosion, superimposed on a lowering of sea-level. Sea-level regression can be attributed to the glacioisostatic rebound following the Narssârssuk stade ice advance. The position of the coastline during this interval is not known, but it could be expected to be higher than the Holocene marine limit from the greater degree of ice cover, and it may be represented by the benches developed around 50 m a.s.l.

Non-finite radiocarbon dates (see appendix), amino acid values (Sejrup, this volume) and biostratigraphic data suggest that this period of marine sedimentation dates from an Early Weichselian warm period (Feyling-Hanssen & Funder, this volume).

4. The succeeding long interval up to the Late Weichselian and Holocene is not represented in the stratigraphy of the area. Sea-level was probably below that at present and processes on land were predominantly erosional. In the Late Weichselian/Early Holocene a marine transgression brought sea-level to at least 36 m a.s.l., followed by a gradual regression. The initial transgression recorded by the diamicton (glaciomarine mud) in unit N6 may be synchronous with the Late Weichselian/Early Holocene glacier advance of the Wolstenholme Fjord stade, known from the interior fjord region.

Qarmat and the adjacent area

(Michael Kelly and Jon Y. Landvik)

Pitugfiup kûgssua and adjacent valleys

In addition to widespread till deposits the area around the valleys of Pitugfiup kûgssua and Siorqap kûa (Fig. 10) are notable for the development of meltwater deposits and landforms. A system of meltwater channels, formed in tills and outwash gravels, cut across the upper Siorqap kûa with an east – west alignment with the uppermost end forming the watershed between this valley and the southerly flowing Narssârrsuk drainage system, down which meltwater must have flowed. In part the channels also drain into the system of deep gorges formed in the easily eroded Proterozoic rocks, where subglacial or marginal glacial meltwater cut through the plateau margin to the lower base level of the fjord area. Outwash gravels form the floor of the valley of Pitugfiup kûgssua, where they have been worked extensively for construction purposes.

All of these deposits have the same degree of weathering and degradation as the deposits around Narssârssuk, and they appear to be part of the same sequence. An ice margin on the Narssârssuk-Siorqap kûa watershed would correspond approximately with an ice front in the fjord near Narssârssuk. It is therefore likely that the deposits in these valleys date from the retreat stages of the Narssârssuk glacial phase.

Along the northern edge of the area are a series of ice marginal deposits and landforms which contrast with those to the south in their degree of weathering, preservation and continuity. This is the Wolstenholme moraine system, mapped by Krinsley (1963) and named informally by Kelly (1985). On the steep sides of the innermost part of Wolstenholme Fjord they comprise a series of lateral moraine ridges and benches, the uppermost at *c.* 250 m a.s.l. being in places 5 m high. At Amitsuarssuk valley (Fig. 10) the outer part of the

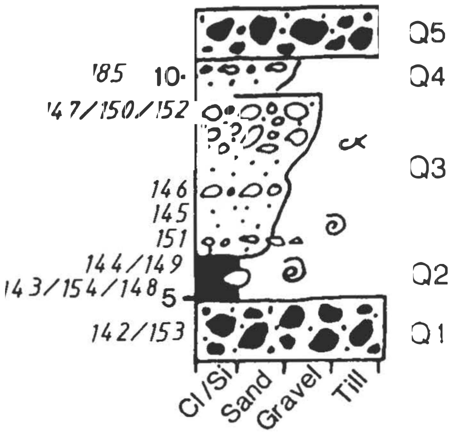


Fig. 9. Lithological log from section at Qarmat. Small numbers are m above sea level. Sample numbers refer to samples described in later chapters. Symbols as in Fig. 8

system swings south to occupy the watershed between this valley and Pitugfiup Kûgssua. Glacio-lacustrine deposits are common in the former valley, in part overlain by till and coarse outwash sediments, i.e. dating from a readvance stage. The moraine system does not extend beyond the narrow section of the fjord and the ice terminus was presumably fixed by the widening of the fjord, implying a floating terminus.

The state of weathering of these deposits is relatively slight, with surface clasts showing less granular disintergration and oxidation than in the older deposits to the south, whilst the depth of oxidation also appears to be less, around a few cm compared with *c.* 10 cm. Although a detailed study of the degree of weathering of the two sets of deposits has not been carried out, the contrast in their overall appearance and the evidence of a readvance are taken to indicate that a substantial period intervened between them.

Deposits comparable in age to the Wolstenholme moraine system do not occur in front of the ice caps along the eastern margin. In fact their present margins appear to be lying on deposits and landforms formed during retreat of the ice margin of the Narssârssuk stade; thus, during this century the Store Landgletscher advanced across a meltwater channel, cut in rock, from the earlier glaciation. (See also Reeh et al., below).

Qarmat, lithostratigraphy

The pre-Holocene deposits mentioned by Kelly (1980a) cover a small area on the south side of the inner part of Wolstenholme Fjord, close to the mouth of a stream. It lies about 1.5 km in front of the terminus of the lateral moraine from the historical readvance of Harald Moltke Bræ. The steep slopes above are covered with till and moraines of the Wolstenholme moraine system. The succession is shown in Fig. 9.

Unit Q1. – This unit is a red tinged diamicton with a sandy silt matrix, shell fragments, fine gravel and coarser clasts. It is interpreted as a till, although it might be a proximal glaciomarine sediment. Above this is a coarsening-upwards sequence of marine sediments.

Unit Q2. – This is a grey green marine diamicton of sandy silt with dropstones. At the base is a horizon with *Hiatella arctica* in life position, and near the top another with *Mya truncata* similarly in life position

Unit Q3. – This unit consists of 2 m fine to coarse grained sands, with thin gravel layers as well as several horizons with shells in life position – *Mya truncata* and *Serripes groenlandicus*.

Unit Q4. – This unit uncontormably overlies Unit Q3, and consists of 1.7 m of fine gravel with a sand matrix, indistinctly bedded in units of a few centimetres thickness, some of which show trough cross stratification. The bedding in the gravel dips offshore in the direction of the present stream, and the sediment is interpreted as fluvial. Its TL-age is 58 ka (Kronborg & Mejdahl, this volume). This sediment provides a glimpse from a long-lasting hiatus in the record, and shows that at this time the glacier in the fjord was not further advanced than now.

Unit Q5. – The topmost unit is a 1.4 m cryoturbated diamicton, clast rich with a silty sand matrix, interpreted as a till.

An adjacent exposure showed more than 8 m of subrounded to rounded clast-supported cross-bedded gravels with foreset beds dipping towards the fjord. These are overlain by 1 m of clast supported diamicton dominated by flat-lying subrounded boulders, some of them > 1 m in diameter, in a sand and gravel matrix. This unit is capped by 2 m of Holocene beach gravels. The relationship of these beds to those in the neighbouring sequence is uncertain but the two lower units may be proximal glacial facies related to unit Q5.

Interpretation

The northern half of the Thule mainland was occupied be the retreating ice margin during the Narssârssuk stade (unit Q1). The importance of meltwater phenomena from this interval suggests an ameliorating climate and rapid retreat.

The succeeding marine sediments (units Q2 Q3) therefore correlate with the marine beds (units N4-N5) at Narssârsuk. Like them, they show a changing marine environment, with loss of ice margin influence, changing sources of sediment and decreasing water depth. Amino acid analyses (Sejrup, this volume) agree with this correlation.

The TL-age of 58 ka for the fluvial sediment of unit Q4 indicates that at this time the fjord glacier in Wolstenholme Fjord was not further advanced than at present.

The subsequent ice advance of the Wolstenholme Fjord stade deposited the Wolstenholme moraine system at its maximal and early recessional stages, and unit Q5 is assigned to this phase. The relative freshness of the deposits of this advance suggests correlation with the Late Weichselian glaciation which affected West Greenland (Kelly 1985). In this case, the Holocene marine transgression and the raised marine sediments

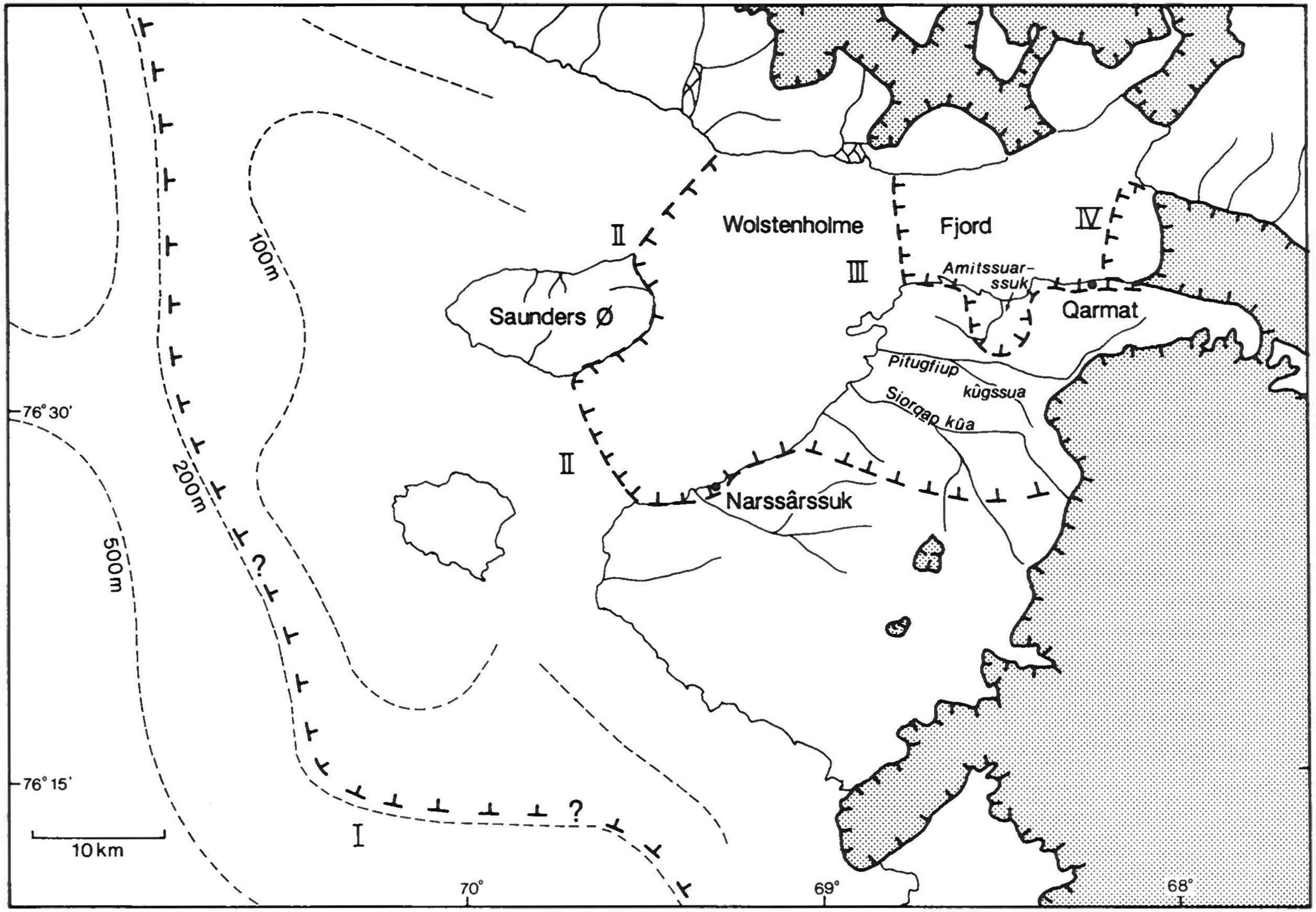


Fig. 10. Postulated ice margin positions in the Wolstenholme Fjord area during the Agpat glaciation (I), the Narssârssuk and Wolstenholme Fjord stades (II, III), and in historical times (IV).

preserved in the area can be related to the glacioisostatic movements of the Wolstenholme Fjord stade.

Summary of regional glaciation history

(M. Kelly and J.Y. Landvik)

The lithostratigraphic and geomorphic evidence from Saunders Ø and the Thule mainland records the occurrence of at least three glacial – deglacial cycles; maximum ice margin positions during each are shown in Fig. 10.

1. The Agpat glaciation, in which an ice sheet ostensibly covered the whole land area, and extended onto the shelf. The subsequent deglaciation, which may have been associated with interglacial environmental conditions (Feyling-Hanssen & Funder, this volume), brought the ice margin north of Narssârssuk, at least.

2. The Narssârssuk stade, during which ice advanced in Wolstenholme Fjord as far south as Narssârssuk. In the northern fjord arm it did not reach Narssârssuaq on Saunders Ø, but may have been held at the narrowest point between Saunders Ø and the north coast of Wolstenholme Fjord, where morainic features were observed by Poul Frich (pers. comm. 1987). Deglaciation resulted in an ice cover hardly more, or perhaps even less extensive than at present, which is compatible with the interglacial environmental conditions suggested by Feyling-Hanssen & Funder (this volume).

3. The Wolstenholme Fjord stade resulted in a limited ice advance to the mouth of the inner part of Wolstenholme Fjord.

4. A minor readvance has taken place during historical times and probably reached a maximum in 1920 (Koch 1928). This is recorded by fresh unvegetated moraine some kilometres in front of the present glacier terminus.

Fauna and flora

ROLF W. FEYLING-HANSSEN and SVEND FUNDER

Foraminiferal assemblages

(R.W. Feyling-Hanssen)

A total of 51 species of benthonic foraminifera occur in the 16 investigated samples from the Thule area. Approximately 20 of these are common in at least a few of the samples. Of these 20 species only 10 are very frequent in one or more of the samples (Fig. 11). The most important are *Islandiella helenae, Astrononion gallowayi, Cassidulina reniforme, Islandiella norcrossi, Elphidium excavatum, Elphidium subarcticum, Elphidium albiumbilicatum,* and *Nonion orbiculare*. Planktonic foraminifera, represented by two specimens of *Globigerina pachyderma*, occur only in one sample.

In order to obtain a clearer picture of the palaeoenvironments of the Quaternary in the Thule district, it would be of interest to compare these Quaternary assemblages with arctic foraminiferal faunas of the present day. Such assemblages have been studied by Aksu (1985) from Baffin Bay between Baffin Island and the west coast of Greenland. However, Aksu's results are not comparable with the present Thule faunas due to the very deep water (minimum 900 m) in his study area.

Phleger (1952) has described recent foraminiferal assemblages from six bottom samples in the Thule district. These were all taken at 31 m depth, and the benthonic foraminiferal populations contained a high frequency of agglutinated specimens. This contrasts with the fossil assemblages dealt with in the present investigation, in which no agglutinated foraminifera occur.

Other investigations of recent foraminifera of arctic shelf waters show that many arenaceous species and specimens occur in the assemblages (e.g. Cushman 1948, Loeblich & Tappan 1953). Vilks (1964, 1969) found that the recent foraminiferal populations of the Canadian Arctic are dominated by arenaceous forms, and Schafer & Cole (1985) published similar observations from the eastern Baffin Island shelf.

Quaternary arctic assemblages, on the other hand, contain few if any arenaceous foraminifera (cf. Gudina 1966, 1969; Feyling-Hanssen 1976, 1980, 1985).

Environmental conditions could hardly be responsible for these differences. Rather, the underrepresentation of agglutinated foraminifera in fossil assemblages has probably resulted from *post-mortem* destruction of their delicate tests in the sediment. Nagy (in Elverhøi *et al.* 1980), for example, has argued that the decreasing frequency of the arenaceous *Spiroplectammina biformis* downwards in Spitsbergen cores is a function of the solution of wall cement combined with increasing pressure in the sediment. In particular, it is probable that serious decimation of the arenaceous tests takes place in permanently frozen sediment bodies or during freezing and thawing in outcrops. If, as an experiment, the arenaceous specimens of the richest of Phleger's recent Thule assemblage; (sample OP15) are ignored, the resulting faunal composition would be as follows, (names changed to present taxonomic usage):

Islandiella helenae	60%
Astrononion gallowayi	15%
Elphidium excavatum	7%
Buccella frigida	4%
Nonion labradoricum	4%
Elphidium subarcticum	4%
Lagenidae	1%
Stainforthia fusiformis	1%
Cibicides lobatulus	1%
Angulogerina fluens	1%
Cassidulina reniforme	1%
Nonion barleeanum	< 1%

This faunal composition compares well with the fossil *Islandiella helenae* assemblages in the youngest samples from the area. The resulting Sanders (1960) similarity index for the above assemblage and that of sample no. 010 (unit S7) is high (78%).

The foraminiferal assemblages of the Quaternary samples from Thule are inner shelf faunas indicating arctic marine climatic conditions. A few subarctic species do however occur in the assemblages. These species are known in the arctic today, but they are not confined to the arctic, and they may not have occurred there during glacial periods of the past.

Different views exist on which foraminiferal taxa should be considered as indicating subarctic, or warmer, as opposed to the arctic conditions (cf. Gudina & Evserov 1973, Feyling-Hanssen 1980). In the present study a very limited number of species have been grouped under the heading subarctic (Fig. 11) viz. *Buccella calida, Rosalina vilardeboana, Epistominella vitrea, Gavelinopsis praegeri, Cibicides lobatulus, Nonionella auricula, Nonion barleeanum*, and *Elphidium incertum*. The presence of one or more of these taxa in significant numbers in a fossil assemblage is probably a definite indication of ameliorated conditions, and they are used here as an index of warming.

Another palaeoenvironmental parameter is the faunal diversity index of Walton (1964), which is defined as "the number of ranked species in a counted assemblage whose cumulative percentage accounts for 95% of the

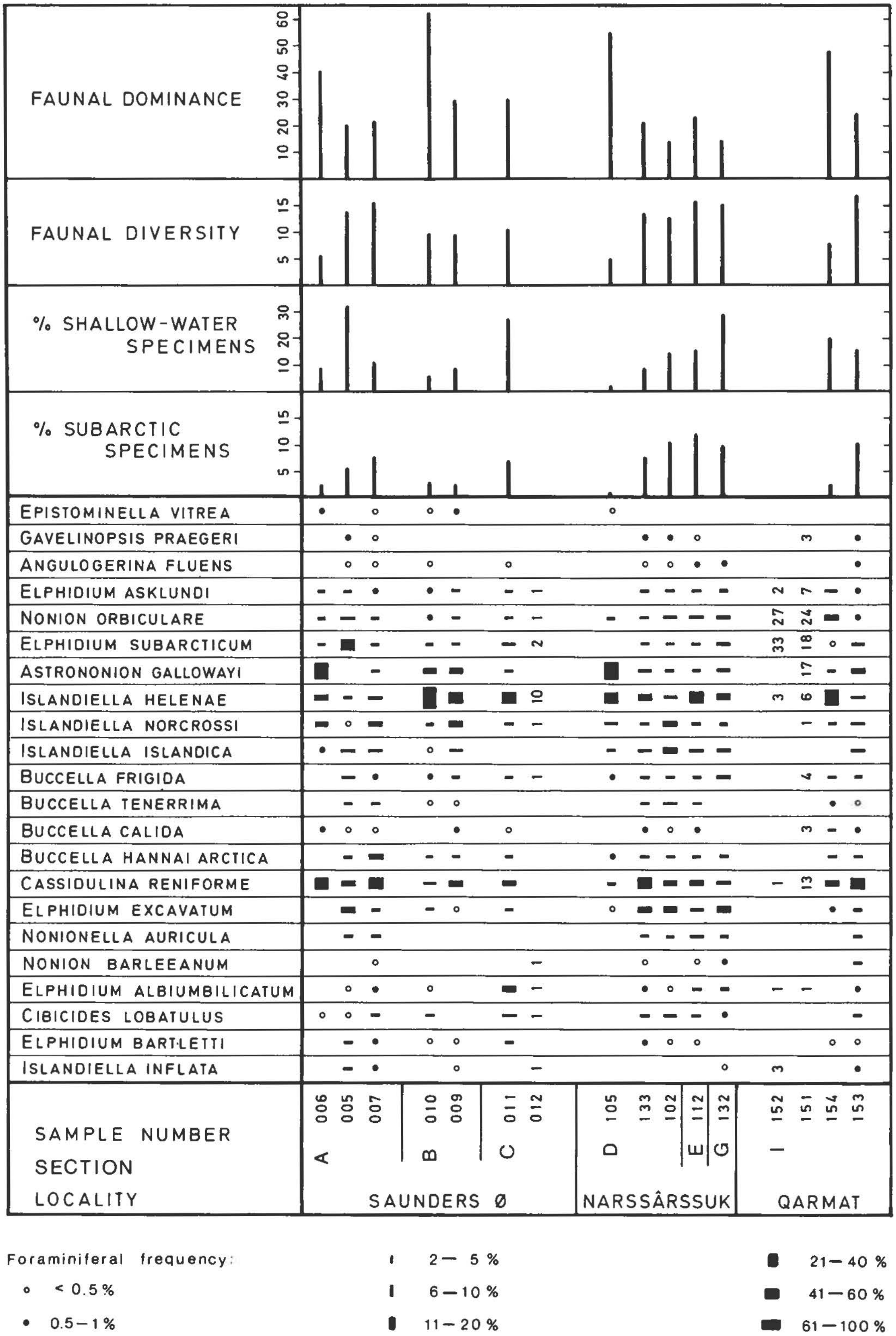
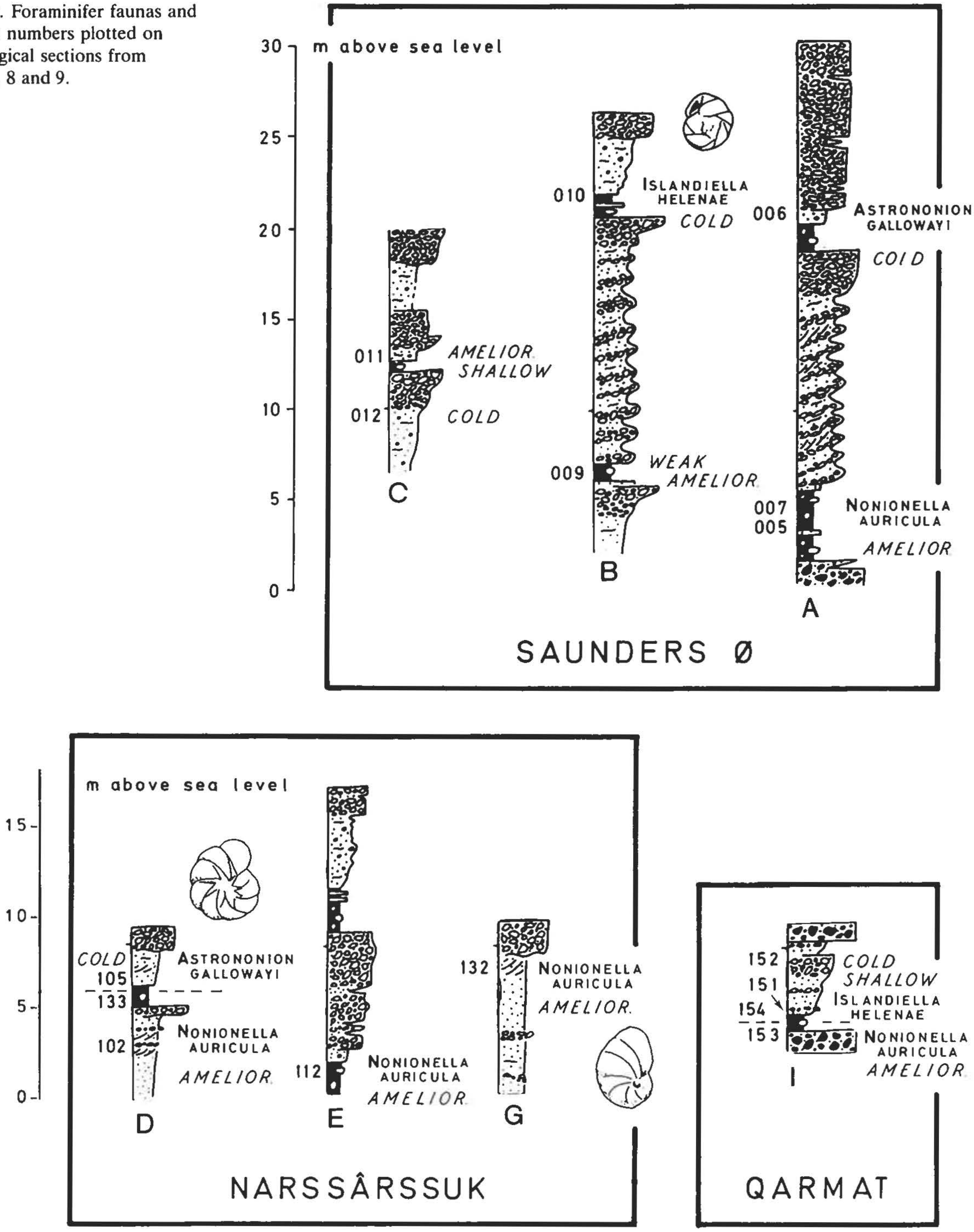


Fig. 11. Foraminifer range-chart for the sections on Saunders Ø and at Narssârssuk and Qarmat. Sample numbers and foraminiferal assemblages are indicated for each section.

Fig. 12. Foraminifer faunas and sample numbers plotted on lithological sections from Figs 4, 8 and 9.



total assemblage". High diversity would, in general, indicate stable and favourable environmental conditions, whereas low diversity would reflect unstable and severe conditions. A third indicator is the faunal dominance index, which is smply the percentage of the most frequent species in a counted assemblage. High faunal dominance would in general indicate extreme environmental conditions, and would normally occur together with low faunal diversity (Fig. 11).

The *Nonionella auricula* assemblage

Fossil assemblages reflecting arctic but ameliorated conditions occur in the lower part of the Saunders Ø sequence (samples 007 and 005, unit S2), in the lower part of the Narssârssuk sequence (sample 112, unit N1), as well as in its upper parts (samples 102, 132, unit N5), possibly reworked in unit N6 (sample 133), and reworked in till at Qarmat (sample 153, unit Q1).

In addition to their ameliorated character (high diversity, low dominance), these assemblages are characterised by the presence of *Nonionella auricula*, which is absent in high-arctic fossil assemblages from the Thule area. The samples in question have been grouped as the *Nonionella auricula* assemblage. The faunal composition of sample 007 from Saunders Ø is typical for this assemblage (Table 1). Another fauna of the same assemblage but from shallower water (sample 132, at Narssârssuk) contains the same percentage of *Nonionella auricula* but fewer *Cassidulina reniforme* (10%) and more *Elphidium excavatum* (13%). Shallow-water species are more frequent, e.g. *Nonion orbiculare* (9%), *Elphidium subarcticum* (9%), *E. albiumbilicatum* (5%), and *E. asklundi* (4%). Shallow-water specimens account for 28% of this assemblage.

Nonionella auricula was originally described from off Plymouth, England (Heron-Allen & Earland 1930), but it also occurs in the present day Arctic, e. g. off Point Barrow, Alaska, at depths from 21 m to 223 m (Loeblich & Tappan 1953). Gudina & Evserov (1973) considered it to be an arctic-boreal species. It has been recorded from interstadial Weichselian layers in southwestern Norway (Feyling-Hanssen 1971, Mangerud *et al.* 1981, Sejrup 1987), and has also been found in early Eemian or pre-Eemian deposits in Norway (Mangerud *et al.* 1981). Other records include the Upper Pliocene of eastern Baffin Island (Feyling-Hanssen 1980).

Most of the samples from the Thule area containing the *Nonionella auricula* assemblage originate from units of shell-bearing diamicton deposited in deep water; only samples 102 and 132 are from sandy deposits indicating shallower water.

Table 1. Foraminifer-frequencies in sample 007, unit S2, weight 100 g.

Species	Percentage
Cassidulina reniforme Nørvang	22
Islandiella norcrossi (Cushman)	18
Buccella hannai arctica Voloshinova	12
Islandiella helenae Feyling-Hanssen & Buzas	9
Islandiella islandica (Nørvang)	8
Elphidium subarcticum Cushman	5
Nonionella auricula Heron-Allen & Earland	5
Nonion orbuculare (Brady)	4
Elphidium excavatum (Terquem)	4
Buccella tenerrima (Bandy)	2
Cibicides lobatulus (Walker & Jacob)	2
Astrononion gallowayi Loeblich & Tappan	2
Bucella frigida (Cushman)	1
Elphidium bartletti Cushman	1
Elphidium albiumbilicatum (Weiss)	1
Islandiella inflata (Gudina)	1
Nonion labradoricum (Dawson)	1
Elphidium asklundi Brotzen	1

Thirteen other species, each accounting for less than 1% of the counted assemblage.
Counted: 494 specimens = 4/15 of the sample.
Six additional species in the uncounted part of the sample.
Number of species: 37.
Number of specimens/100 g sediment: 1850.
Faunal diversity: 16. Faunal dominance: 22.
Subarctic specimens: 8%.
Shallow-water specimens: 11%.

Table 2. Foraminifer frequency in sample 006, unit S4, weight 100 g.

Species	Percentage
Astrononion gallowayi Loeblich & Tappan	41
Cassidulina reniforme Nørvang	21
Islandiella norcrossi (Cushman)	13
Islandiella helenae Feyling-Hanssen & Buzas	12
Elphidium subarcticum Cushman	5
Nonion orbiculare (Brady)	3
Buccella calida (Cushman & Cole)	1
Islandiella islandica (Nørvang)	1
Epistominella vitrea Parker	1

(One specimen of the planktonic *Globigerina pachyderma* Ehrenberg has been observed).
Three other species, each accounting for less than 1% of the counted assemblage.
Counted: 378 specimens = 7/90 of the sample.
Number of species: 12.
Number of specimens/100 g sediment: 4850.
Faunal diversity: 6. Faunal dominance: 41.
Subarctic specimens: 3%.
Shallow-water specimens: 9%.

Table 3. Foraminifer frequency in sample 010, unit S7, (Holocene), weight 100 g.

Species	Percentage
Islandiella helenae Feyling-Hanssen & Buzas	63
Astrononion gallowayi Loeblich & Tappan	11
Cassidulina reniforme Nørvang	7
Buccella hannai arctica Voloshinova	4
Islandiella norcrossi Cushman	2
Cibicides lobatulus (Walker & Jacob)	2
Elphidium excavatum (Terquem)	2
Elphidium subarcticum Cushman	2
Elphidium asklundi Brotzen	1
Buccella frigida (Cushman)	1
Nonion orbiculare (Brady)	1

Nine other species, each accounting for less than 1% of the counted assemblage.
Counted: 366 specimens = 4/15 of the sample.
One additional species in the uncounted part of the sample.
Number of species: 21.
Number of specimens/100 g sediment: 1370.
Faunal diversity: 10. Faunal dominance: 63.
Subarctic specimens: 3%.
Shallow-water specimens: 6%.

Sample 132 and also sample 005 contain considerable numbers of shallow-water foraminifera.

The distinctly ameliorated *Nonionella auricula* assemblage is similar, to assemblages in sections of the Qivituq Peninsula on the northeast coast of Baffin Island (Feyling-Hanssen 1980), where they overlie layers with strongly ameliorated foraminiferal assemblages, probably representing oxygen isotope stage 5e, the Eemian. *Nonionella auricula*-bearing deposits have also been found in North Sea boring 2501, directly overlying an Eemian sand unit (Feyling-Hanssen 1981).

The *Astrononion gallowayi* assemblage

An assemblage dominated by *Astrononion gallowayi* occurs in unit S4 on Saunders Ø (sample 006. Table 2), while unit N6 from Narssârssuk carries an assemblage with a similar faunal composition (sample 105). Furthermore this foraminifer is encountered in fewer numbers in nearly all the other investigated samples. *Astrononion gallowayi* is an arctic species which is known to occur in some abundance in the vicinity of calving glaciers (Nagy 1965). It would appear sometimes to occur in increasing amounts, associated with indications of somewhat ameliorated environmetal conditions, of which calving or melting ice may be a result (Feyling-Hanssen 1980, Feyling-Hanssen & Ulleberg 1984).

The *Astrononion gallowayi* assemblage of the Thule area may, nevertheless, reflect high-arctic conditions (faunal dominance high, diversity low, few subarctic specimens), and the low content of shallow-water specimens indicates a sublittoral environment. The lower parts of units S4 and N6 are therefore interpreted as ice proximal and deposited during the Narssârssuk and Wolsteholme Fjord stades respectively, an interpretation that is compatible with lithological and stratigraphical observations.

The *Islandiella helenae* assemblage

A third type of assemblage had *Islandiella helenae* as its dominant species. This *Islandiella helenae* assemblage occurs in the Early Holocene of Saunders Ø (sample 010, unit S7, Table 3), and in unit Q2 of Qarmat (sample 154, Figs 11 and 12). At both localities it reflects high-arctic conditions and at Qarmat also a littoral environment (less than 20 m depth). The assemblage in sample 010 in particular resembles recent faunas in East Greenland fjords, and, as illustrated on p.19, the recent fauna of the Thule area when the arenaceous component is ignored.

Other assemblages

The assemblage in sample 009, unit S4, is also dominated by *Islandiella helenae*, but not to the same extent as the two above-mentioned assemblages. *Astrononion gallowayi* and *Islandiella norcrossi* are also important, and *Islandiella islandica* is common. This assemblage reflects arctic conditions but probably not high-arctic. Its content of subarctic specimens is low, only 3%, but the faunal diversity index is 10 and the dominance only 30%.

Slightly ameliorated conditions were also observed in an assemblage from the Early Holocene unit S7 on Saunders Ø. This is a shallow-water sample with 7% subarctic specimens, a faunal diversity index of 11, and a dominance of 30.

Three of the samples from the area contain less than 100 foraminifers in 100 g sediment, whereas most have between 200 and 5000, and a single sample, 009 from unit S4, contain 12 000. The absence of planktonic foraminifera may be explained by the fact that deposition occurred in a nearshore environment that was effected by the influx of meltwater, which must have caused somewhat reduced salinities.

Conclusions

The investigation has led to the distinction of four different assemblage types in the Quaternary sediments of the Thule area: the subarctic *Nonionella auricula* assemblage, the high-arctic *Astrononion gallowayi* assemblage, the high-arctic *Islandiella helenae* assemblage, and an ameliorated assemblage with *I. helenae* as the dominant species, which has only been registered in a single sample. As discussed in the following section the environmental implications are in close agreement with those obtained from other faunal and floral evidence. Especially important is the recognition of the cold water *Astrononion gallowayi* assemblage, correlated with the two periods of glacier advance known from glacial deposits in the interior parts of the region.

Molluscs, barnacles and other marine and terrestrial fauna and flora

(S. Funder)

Blake (1973, 1975, 1977) demonstrated that pre-Holocene sediments on Saunders Ø and elsewhere in northern Baffin Bay contain the subarctic molluscs *Mytilus edulis* and *Chlamys islandica* outside their present limits, and used this to correlate the sediments with the Sangamonian (Blake 1973: 56, although revised by Blake 1987: 26).

One purpose of the NORDQUA 86 work was to obtain a more detailed knowledge of the faunas, and relate them to regional oceanographic changes. The new results show that besides the two species mentioned above, the subarctic barnacles *Balanus balanoides* and *B. crenatus* also occurred, as well as remains of a few thermophilous terrestrial organisms, and that subarctic organisms lived in the area during at least two phases of the last interglacial/glacial cycle.

This section gives a brief survey of the present-day oceanography and fauna in Baffin Bay and a description of the subfossil marine faunas, with emphasis on the subarctic species and their environmental implications.

Present fauna and oceanography

Early observations on marine faunas in the Thule district have been summarised by Posselt & Jensen (1898), Stephensen (1913) and Thorson (1951). Detailed studies have been made by Vibe (1939, 1950) and Theisen (1973), and in August 1968 extensive dredging was carried out by J. Just and C. Vibe, especially in the area between Saunders Ø and Narssârssuk (e.g. Just 1980). The molluscs from this work have been identified by K. Ockelmann, and have been made available to me at The Zoological Museum, Copenhagen.

The Baffin Bay distribution of the four species mentioned above is given in Fig. 13, and the results show that while *Balanus balanoides* is absent, *Mytilus edulis*, *Chlamys islandica* and *Balanus crenatus* have their Baffin Bay northern limits in the Thule area. However they occur sparsely, and the area is apparently critical for them. Thus only two populations of *Mytilus* are known, at Thule and at Siorapaluk, 150 km to the north (Vibe 1950). Theisen (1973) found that both populations consisted only of old individuals that for some time had not been able to reproduce. *Balanus crenatus* is known from Etah 200 km to the north of Thule, and from Cape Herschel on the Canadian side of Smith Sound. Like *Mytilus*, this species is restricted to rock crevices in a narrow zone at low water mark, where the animals, protected from abrasion by ice, may take advantage of high temperatures in the surface water, thus underlining the critical situation for them (Vibe 1950).

Chlamys islandica lives at greater depth, and in the 1968 dredging at Thule a few adult individuals of *Chlamys* were found at 30 and 50 m at two localities, between Saunders Ø and Narssârssuk, and in Murchison Sund 150 km to the north. Empty shells have been dredged at Siorapaluk (Vibe 1950), and at Narssârssuk, and were found on the beach of Wolstenholme Fjord (1968 collections). The observations show that also this species is sparse in the area.

For all the three species mentioned the populations in the Thule district appear to be disjunct from their main areas which have their northern limit at Upernavik, 400 km to the south. This area also marks the northern limit for *Balanus balanoides*, but Petersen (1962, 1966) noted that its populations here may be sustained by larvae brought up from the south. He also showed that its distribution has apparently varied considerably during this century, in harmony with the known changes in extent of the West Greenland Current along the coast.

On the Canadian side of Baffin Bay *Mytilus* has a disjunct occurrence at Pond Inlet, northern Baffin island, while its main area of distribution begins 800 km to the south. The occurrence at Pond Inlet may be conditioned by insolation and local warming in this sheltered area (e.g. Ellis 1960). The population has lived here for at least a millenium (Blake 1987: 23), and is considered a relict from a continuous distribution along the Baffin Island coasts that was severed 3000 years ago (Andrews *et al.* 1981).

The summer surface circulation in Baffin Bay is shown in Fig. 14. The exchange with adjacent oceans is afforded by the West Greenland Current moving northwards along the Greenland coast, and the Labrador Current carrying cold polar water from the Arctic Ocean south along the Canadian coasts.

The West Greenland Current contains a mixture of cold polar and underlying warm Atlantic water. However, the relative proportion of the two components varies, and in periods the warm Atlantic water is known to invade even shallow depths, to the great benefit of the Greenland fishing industry (Jensen 1939).

The greater part of the water in the West Greenland Current is deflected into central and eastern Baffin Bay, and some occurs as surface water in Cumberland Sound and Frobisher Bay on the Canadian side (Kramp 1963, Ellis 1955). However, small amounts have been traced in coastal shallow water as far north as the Thule district and even eastern Ellesmere Island (Aarkrog *et al.* 1987). Since the water spends some time in transit one effect is that it prolongs summer warmth at high northern latitudes (e.g. Smidt 1979).

From Figs 13 and 14 there appears to be a causal relationship between the occurrence of subarctic water masses in Baffin Bay and the distribution of the four species mentioned. This has been pointed out already by Madsen (1936, 1940), while Dunbar (1968) defined his important marine subarctic zone as the area where polar and Atlantic water masses mix, with its northern limit marked by the northernmost penetration of warm Atlantic water.

In summary, there is a large amount of evidence to show that the distribution of the four species in Baffin Bay is associated with the occurrence of warm subarctic Atlantic water, and consequently especially *Mytilus* and *Chlamys* have often been used as palaeooceanographic indicators for this type of water in Baffin Bay (e.g. Andrews *et al.* 1981, Funder & Símonarson 1984, Funder 1984, 1989, Kelly 1985, 1986). Marine oceanographic and biological studies show that the Thule area today marks the northernmost extension for the conveyance of Atlantic water in the West Greenland Current, and the fossil occurrence of subarctic species in this area is therefore interpreted as reflecting a palaeooceanographic regime similar to the present one, or with deeper penetration of The West Greenland Current.

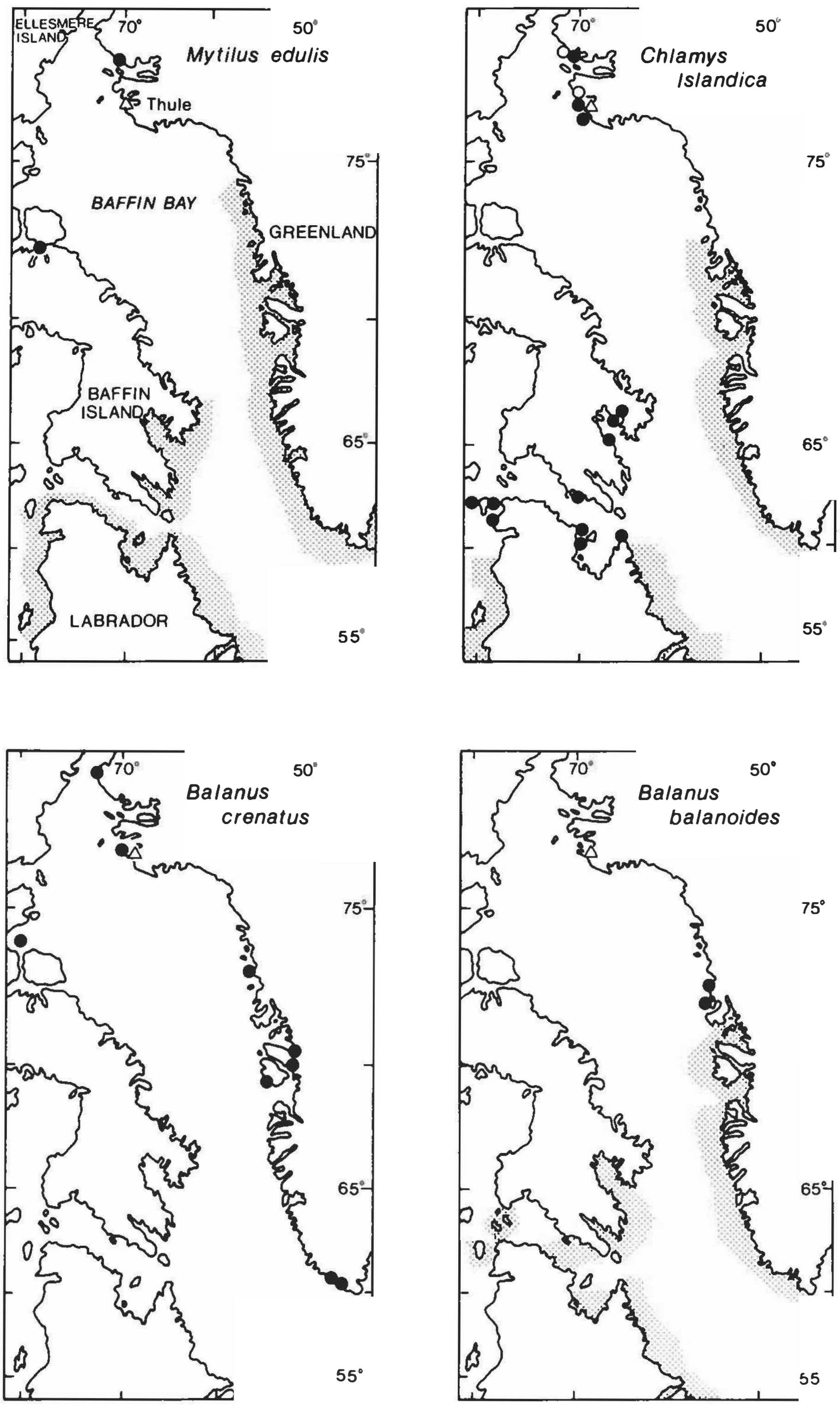


Fig. 13. Present distribution of some marine subarctic species in Baffin Bay. Shading shows continuous distribution, black dots are isolated northern occurrences, open dots are empty shells. (sources: Posselt & Jensen 1898, Stephensen 1936, Vibe 1950, Barnes 1957, Petersen 1962, Theisen 1973, Lubinsky 1980, Dale 1985).

Table 4. Pre-Holocene marine flora and fauna from the Thule aminozone. (Location of samples in sections appear from Figs 4, 8, 9).

	Saunders Ø											
Sample No. (sectionwise-upwards)	005	007	006	009	008	012	101	102	112	119	110	113
Section	A	A	A	B	B	C	D	D	E	E	E	E
Unit	S2	S2	S4	S4	S4	S5a/2?	N5	N5	N1	N1	N4	N4
Type (1)	sd	sd	sd	sd	sd	sd	sd	sd	sd	pk	pk	sd,pk
State (2)	is,x	is,x	is,x	is	is,x	x	x	x	is,x	x	is,x	is
BIVALVIA												
Portlandia sp.												
Nuculana pernula Costigera	2											
Modiolaria nigra (Gray)									3			
Mytilus edulis Linne						f						
Chlamys islandica (Møller)	r					2			r			
Palliolum greenlandicum (Sowerby)									2			
Astarte sp.			r									
Axinopsida orbiculata Sars												
Clinocardium ciliatum (Fabricius)												
Serripes groenlandicus (Chemnitz)											r	
Macoma calcarea (Chemnitz)	2		r	c				3				c
Hiatella arctica (Linne)	r		f	c	r	c		c	c		c	f
Mya truncata Linne	f	c	f	c	r	c		c	f	c	c	f
GASTROPODA (3)												
Buccinum cf *hydrophanum* Hancock												
Buccinum sp. (?*belcheri* Reeve)												
CIRRIPEDIA												
Balanus balanoides Linne												
Balanus cf. *balanoides*												
Balanus balanus (Linne)		r	c			r			c		2	2
Balanus crenatus Bruguiere								r				
Balanus sp.	c	c	c					c				
ECHINODERMATA												
Strongylocentrotus droebachensis (Møller)			+						+		+	+
POLYCHAETAE												
Sprirorbis sp.												
BRYOZOA (4)												
Myriapora coarctata (M. Sars)									r			
Myriapora subgracilis (d'Orbigny)												
Celleporina surcularis (Packard)									r			
Celleporina ventricosa (Lorenz)												
ALGAE (5)												
Sphacelaria arctica Harvey												
FORAMINIFER ASSEMBLAGES (from Fig. 10)												
Monionella auricula ass.	+	+					+	+	+			
Islandiella helenae ass.				+								
Astrononion gallowayi ass.			+									
TL age, ka (from Table 9)	136	119	114	113	89	69						
Total alle:Ile, x' (from Table 10)		.036	.021	.049		.041		.027	.030			

Frequencies: f: frequent, c: common, r: rare, 1–3: very rare, number of fragments, +: frequency unspecified, ?: identification doubtful.
(1): sd: bulk sediment, pk: shells picked from section
(2): is: *in situ*, x: fragmented shells
Identification other than the author: (3): L. A. Símonarson, University of Reykjavik; (4): Karen Bille Hansen, Zoological Museum, Copenhagen; (5): P. M. Pedersen, Botanical Institute, Copenhagen.

Narssarsûk														Qarmat						
117	115	134	125	126	129	128	127	130	132	131	181	182	183	142	143	144	151	145	146	147
E	E	E	F	F	F	F	F	G	G	G	K	K	K	I	I	I	I	I	I	I
N5	N5	N5	N2	N2	N4	N4	N4	N5	N5	N5				Q1	Q2	Q2	Q3	Q3	Q3	Q3
sd	pk	sd	pk	pk	sd	sd	pk	pk	sd	sd	pk	pk	pk	pk	sd,pk	sd	sd	sd	pk	pk
is	is	is	x	x	x	x	x	is	is	is	x	x	x	x	is	is	x	is,x	x	is
								+	f								?			c
		r			2	r											?			
		2																		
		2												?						
									r											
											+				r					
	c							+	c									2	2	f
r	3								c					3			c	c	c	f
f	f				c	c		+	c			c		r	c	c	c	r	c	f
	f	c		c	c	c	c	f	c		+	c	r	c	c	f	c	c	c	f
										r										
										a										c
					c	3											r			3
						1											+			r
		c			c									c						
+	+				+	+			+					+	+	+	+			+
									+	+										
									+					+						
															+					
															86		80			80
				.032	.026			.035					.058	.037	.025					.025

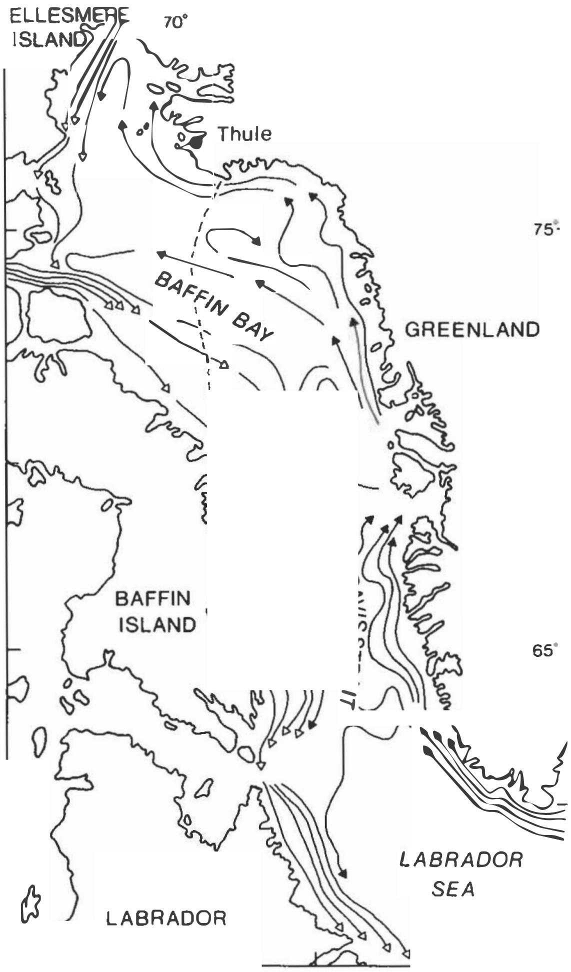


Fig. 14. Summer surface circulation in Baffin Bay, simplified from Kramp (1963) and Jacobs *et al.* (1985). Stippled line is northern boundary for marine subarctic zone from Dunbar (1968). Black and white arrows are warm and cold water.

Material

Faunal analysis has been carried out on 70 samples from 5 localities. The samples generally consist of 3 kg bulk sediment, but some are smaller and comprise shells picked out from the sections. 54 samples are considered pre-Holocene in age, and out of these 29 are from the youngest pre-Holocene marine episode, the Qarmat interstade, while only 2 are from the older Saunders Ø interstade. However, rather well preserved faunas occur in some samples of till and glacially reworked sediment from the Narssârssuk stade, and are – on the basis of their amino acid ratios – interpreted to be derived from the underlying Saunders Ø interstade (units N1-N3 and Q1, Table 4). Hence different stratigraphic levels are variously represented, and this must kept in mind in the environmental interpretation.

Besides shells from bivalves and barnacles, needles and plate fragments from sea urchins occur frequently. Less frequent are fragments of gastropods, bryozoans and marine algae which have been identified by L.A. Símonarson, Karen Bille Johansen and Poul Møller Pedersen. A few samples contain washed-out remains of terrestrial flora and fauna. Moss remains in these samples have been identified by G.S. Mogensen, angiosperm leaves and seeds by O. Bennike, and a single well preserved beetle fragment by J. Böcher (1989).

Lists of the identified pre-Holocene and Holocene marine species are given in Tables 4 and 5, and remains of terrestrial flora and fauna are listed in Table 6.

The shell material varies from highly fragmented in coarse grained sediments to well preserved with articulated shells in silt sediments. From field observations the latter have generally been interpreted to be *in situ*, while the fragmented faunas, living in a high energy environment, have been displaced after death. The faunas have a heavy bias towards species with robust shells, and are dominated by forms that are very widespread in northern waters (e.g. *Mya truncata*, *Hiatella arctica*, *Macoma calcarea*). Apart from the subarctic species mentioned above there are few ecological indicators. Especially noteworthy is the rarity of arctic forms in the pre-Holocene faunas. Thus, while *Portlandia arctica* is rather common in the Holocene faunas, it is absent from the older ones. The only arctic representative in the pre-Holocene samples is *Palliolum greenlandicum* which, however, is known as a bathyal form in southern areas (Thorson 1951). The environmental interpretation of the faunas therefore relies mainly on the presence or absence of subarctic fauna elements, and some notes on their occurrence in the material, and in the Baffin Bay region are given below, and discussed in a later section. Event names are shown on Fig. 24.

Mytilus edulis. – This species has been found in 11 pre-Holocene samples from all three pre-Holocene localities, and in three Holocene samples. The pre-Holocene material is highly fragmented, and only unit N5 (sample 132) contains fairly intact shells. In some samples the fragments are extremely abundant; this is the case for unit S5a where they are a distinct component of the gravel sediment, as noted also by Blake (1975: 436). On the other hand, in the samples from unit Q3, they are both tiny and rare.

Mytilus occurs at all three localities only in the littoral sediments of the younger pre-Holocene units, best preserved and most frequent in unit N5. However, amino acid ratios from unit S5a indicate that the worn shell fragments here may be eroded from older sediments equivalent to unit S2 from the Saunders Ø interstade, and thus would be 50000 years older than the gravel in which they are embedded.

Mytilus edulis therefore occurred in the area during the Qarmat interstade, whereas its presence during the earlier Saunders Ø interstade rests on circumstantial evidence. Its apparent rarity in the latter, however, should be viewed in the light of the very sparse sample material, and lack of littoral sediments from these units.

In the Baffin Bay region pre-Holocene *Mytilus edulis* has been observed on Coburg Island, where it was referred to the Sangamonian (Blake 1973, 1977). In western Greenland it occurs in the Early Quaternary Pátorfik Formation and in the Laksebugt aminozone referred to the stage 6/5 boundary (Funder & Símonarson 1984). It has also been observed at Svartenhuk, and referred to the Kaffehavn interglaciation correlated with isotopic stage 5e (Kelly 1986). On Baffin Island *Mytilus* is known from the early Quaternary Cape Christian Member, but is conspicuously absent from younger pre-Holocene sediments, including the Kogalu aminozone (Andrews *et al.* 1981, Miller 1985).

Although *Mytilus* is rare in the area at present, it occurs frequently in the raised Holocene deposits, and apparently immigrated in the Early Holocene. Thus the C-14 date of 9200 yrs BP for a sample from Wolstenholme Fjord containing *Mytilus* (sample 021, Table 5) shows that the West Greenland Current was functioning as at present at this early stage. This is in agreement with a date of 8500 yrs BP obtained earlier on whale bones found in association with *Mytilus* (Krinsley 1963). Along the coast of Baffin Island it was more frequent at 8400 yrs BP than it is now (Andrews *et al.* 1981).

Table 5. Holocene marine faunas and ages. (Location in sections appear from Figs 4, 8, 9).

Sample No.	010	011	013	133	104	135	138	021(1)	161	172
Section	B	C	surf.	D	D	H	H		J	J
Unit	S6	S6		N6	N6	N6	N7			
BIVALVIA										
Nucula tenuis Reeve								r		
Nuculana pernula Costigera								r		
Nuculana minuta (O. F. Møller)						2	r	r		
Portlandia arctica (Gray)	+									
Mytilus edulis Linne								r		
Chlamys islandica (Møller)				3				r		
Palliolum greenlandicum (Sowerby)				1						
Astarte sp.				r						
Axinopsida orbiculata Sars								c		
Clinocardium ciliatum (Fabricius)					3			c	s	
Serripes groenlandicus (Chemnitz)								r		
Macoma calcarea (Chemnitz)					3	r	f	c		
Hiatella arctica (Linne)	c		+	r	f	f	f	f		s
Mya truncata Linne	c	c	+	c	f	f	f	f	a	s
GASTROPODA										
Acmaea testudinalis (O. F. Møller)								1		
?*Colus togatus* (Mörch)								1		
?*Buccinum undatum* Linne										
CIRRIPEDIA										
Balanus balanus (Linne)	c				r	f	f			
Balanus crenatus Bruguiere	+	r					3	f		
ECHINODERMATA										
Strongylocentrotus droebachensis (Møller)	+	+			+		+	+		
BRYOZOA (2)										
Celleporina surcularis (Packard)				r						
Nonionella auricula ass.										
Islandiella helenae ass.	+	+								
Astrononion gallowayi ass.										
C-14 age, ka (See appendix)	8.2	8.0 (3)	8.8			8.7		9.2	9.0	7.0
TL age, ka (from Table 9)	14	36								
Total alle:Ile, x' (from Table 10)	.014		.013&.023		.016					

Frequencies: f: frequent, c: common, r: rare, 1–3: very rare, number of fragments, +: frequency not specified.
(1): 1 m above sea-level in coast cliff on north shore of Wolstenholme Fjord, near Salisbury Gletscher.
(2): Identified by Karen Bille Hansen, Zoological Museum, Copenhagen.
(3): Sample GSC-2079, Blake (1975).

Chlamys islandica. – This species occurs in 16 pre-Holocene samples and in one Holocene sample. Like *Mytilus* it occurs at all three localities where pre-Holocene sediments are known, and only as fragments. However, in agreement with its sublittoral habitat, it occurs mainly in fine-grained deep-water sediment. Generally the samples contain only 2–5 fragments, and only in units N4 and N5 (samples 128 and 102) are they rather frequent. A few fragments are found in one of the two samples from the Saunders Ø interstade (unit S2), and may correlate with those found in glacially reworked sediments of the Narssârssuk stade (units N1-N3).

Chlamys occurs together with *Mytilus* on Carey Øer and Coburg Island (Bake 1977), and – probably reworked – in Olrik Fjord where the Olrik Fjord aminozone is tentatively correlated with isotopic stage 5e (Weidick 1978, Kelly 1986). In West Greenland it is known from the Early Quaternary Pátorfik Formation and also from Mudderbugt and Eqalugsugssuit in northern Melville Bugt. These deposits were grouped in the Meteorbugt marine event, considered to be older than isotopic stage 5 (Kelly 1986).

On Baffin Island *Chlamys* is known from all marine episodes since the Cape Christian Member, and it occurs in deposits on Clyde Foreland and Broughton Island which are referred to the Kogalu aminozone (isotopic stage 5), and most abundantly in its upper part (Andrews *et al.* 1981).

In the Holocene, *Chlamys* apparently occurred in the Thule region as early as 9200 yrs BP (sample 021, Table 3). Around southern Baffin Island where the species is now absent, it occurred already at 9700 yrs BP (Andrews *et al.* 1981).

Balanus balanoides. – This barnacle has been identified in three pre Holocene samples, one from Narssârssuk (unit N5) and two from Qarmat (unit Q3). The species, which is very rare in the fossil record (e.g. Feyling-Hanssen 1953), has been identified from its compartment plates, using the characters noted by Broch (1924) and comparison with type material from Greenland. Certain identification of the delicate plates can only be made on well preserved material and plates lacking some diagnostics are referred to *B.* cf. *balanoides*. The richest material is from Qarmat where 10–15 well preserved plates occur in samples 150 and 152.

B. balanoides is thus known only from the youngest pre-Holocene units, i.e. the Qarmat interstade; this is not surprising since it is an intertidal species, and shallow water deposits are known only from this marine episode.

B. balanoides is not known from other pre-Holocene depos-

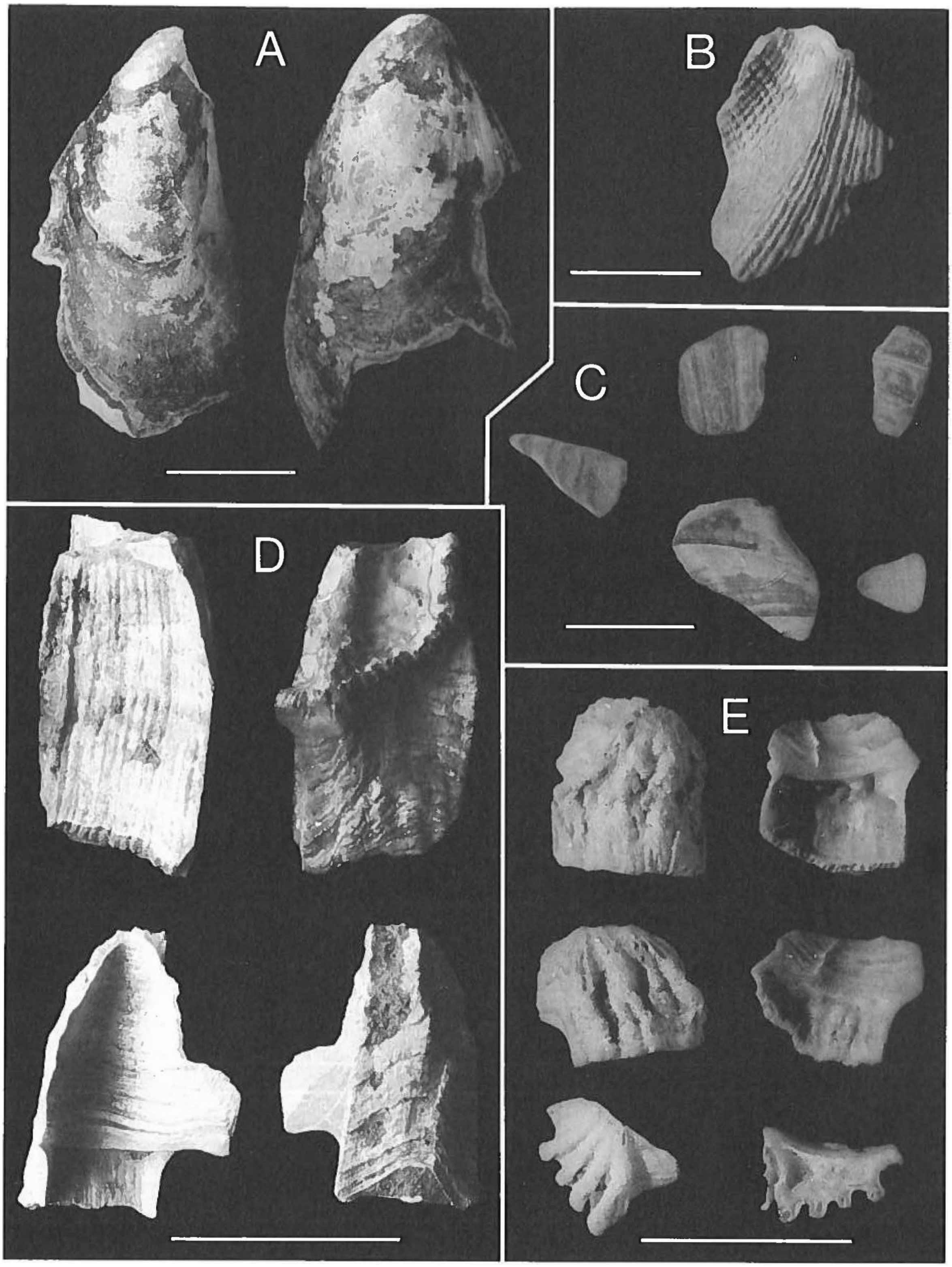


Fig. 15. Pre-Holocene subarctic species from Thule. A: *Mytilus edulis*, B: *Chlamys islandica*, C: *Mytilus edulis*, worn fragments from unit S5a. D: *Balanus crenatus*, inner (left) and outer (right) side of compartment plates, E: *Balanus balanoides*, outer (left) and inner (right) side of compartment plates. White bar is 1 cm.

Table 6. Pre-Holocene terrestrial flora and fauna. (Location of samples in sections appears from Figs 4, 8, 9).

Sample No.	101	113	114	132	131
Section	D	E	E	G	G
Unit	Nx	N4	N5	N5	N5
ANGIOSPERMAE 1)					
Dryas integrifolia M. Vahl					+
Menyanthes trifoliata L.					+
Carex spp.					+
Salix sp.					+
Potentilla sp.					+
Eriophorum sp.					?
Empetrum sp.					?
BRYOPHYTA 2)					
Scorpidium scorpioides (Hedw.) Limpr.	+	+	+		+
Pogonatum dentatum		+			
Pohlia cruda		+			
Pohlia nutans		+			
Ditrichum flexicaule (Schwaegr.) Hampe		+			
Ditrichum cf. *flexicaule*	+				
Bryum spp.	+	+	+		
Rhacomitrium sp.	+	+			
Drepanocladus cf. *exannulatus* Warnst.			+		
Drepanocladus sp.	+	+			
Distichium sp.	+				+
COLEOPTERA 3)					
Amara alpina Paycull				+	

Identification: 1) O. Bennike, Geological Museum, Copenhagen; 2) G. S. Mogensen, Botanical Museum, Copenhagen; 3) J. Böcher, Zoological Museum, Copenhagen.

Balanus crenatus. – This species occurs infrequently in the pre-Holocene samples from Saunders Ø (unit S2), Narssârssuk (units N4) and Qarmat (units Q1 and Q3). It occurs equally infrequently in the Holocene samples.

its in the Baffin Bay region, and also its Holocene history is virtually unknown.

The species thus appears both in the Saunders Ø and Qarmat interstades.

B. crenatus was observed in the Early Quaternary Pátorfik Formation (Símonarson 1981), and in the Kogalu aminozone deposits (isotopic stage 5) on Broughton Island (Andrews et *al.* 1981). Its occurrence in sample 021 (Table 5) indicates an Early Holocene immigration to the Thule area together with *Mytilus* and *Chlamys*.

Terrestrial flora and fauna

The shallow water sediments from unit N5 at Narssârssuk contain scattered washed out remains of the terrestrial flora and fauna, as shown in Table 6. The list is interesting because it contains the first unambiguous evidence of Greenland's terrestrial flora and fauna during the last ice age.

Although the identified taxa generally belong to species and genera that are widespread in the area today there are two exceptions: a single seed of *Menyanthes trifoliata*, identified by O. Bennike, and a well preserved pronotum of the beetle *Amara alpina*, identified by J. Böcher (1989). *Menyanthes* has its present northern limit in Greenland in the interior Disko Bugt region, 800 km to the south of Thule, and shows that summer temperatures during the Qarmat interstade were significantly higher than at present. *Amara alpina* is absent from Greenland, and apparently was not able to reach the country during the Holocene, although it occurs as far north as Devon Island in the Baffin Bay region (Böcher 1989). It is interesting to note that this species has also been identified in peat from nearby Ellesmere Island, although its C-14 age here, 31 100 yrs BP, would indicate a considerably younger age

	Mytilus	*Chlamys s and ca*	*Ba anus alano des*	*B.*	*Menyanthes r ol a a*	*Non onella au cu a* ass. (w	*Island ella helenae* ass. (in r)	*As ronon on gallowayı* (c ld)	Number of analysed sample
HOLOCENE	+	+	–	+	–	–	+	–	21
QARMAT INTERSTADE	+	+	+	+	+	+	+	+	29
NARSSÅRSSUK STADE	–	–	–	–	–	–	+	–	5
SAUNDERS Ø INTERSTADE	+*	+	–	+	–	+	–	–	2 (6*)

* Reworked into younger glacial sediment

Fig. 16. Distribution of subarctic and cold water fauna in Thule aminozone deposits.

(GSC-3364, Blake 1982), if this age is accepted without reservation.

Conclusions

From the description above and Fig. 16 it appears that subarctic species are more abundant in the sediments related to the Qarmat interstade than in older or younger sediments. This is especially conspicuous when the record is compared to that from the Holocene. The presence of *Balanus balanoides*, a sensitive indicator of warm Atlantic water, reflects a West Greenland Current that was stronger and had a larger warm water component than is known from the Holocene, while the occurrence of *Menyanthes trifoliata* shows that summers on land were warmer.

When comparing this record with that of the earlier Saunders Ø interstade it should be kept in mind that the material from the latter is often glacially reworked, while the few *in situ* samples belong to a deep water facies where the littoral *Mytilus* and *Balanus balanoides* would not occur. However, the presence of *Chlamys*, *Balanus crenatus*, and the ameliorated *Nonionella auricula* foraminifer assemblage (see discussion by Feyling-Hanssen above), indicates that warm Atlantic water reached Thule also during this phase, and that marine conditions were similar to the present or warmer. That *Mytilus* may have occurred also during the Saunders Ø interstade is indicated by the circumstantial evidence from unit S5a.

A conspicuous feature of the faunas is the lack of arctic species. The only indication of colder than present conditions comes from the *Astrononion gallowayi* foraminifer assemblage recorded by Feyling-Hanssen (above).

Thermoluminescence dating and amino acid analyses.

CHRISTIAN KRONBORG, VAGN MEJDAHL and HANS-PETTER SEJRUP

Thermoluminescence dating of marine sediments from Saunders Ø and Qarmat

(C. Kronborg and V. Mejdahl)

Samples and dating technique

A total of twelve samples from marine sediments were taken for TL-dating, eight from Saunders Ø and four from Qarmat. Notes on the samples, together with ages expected from geological evidence, are given in Table 7. Sample numbers mentioned in the text are laboratory numbers, while lithological units and field numbers are added in parentheses.

The datings were made on coarse-grained alkali feldspars (0.1–0.3 mm) extracted by means of the heavy liquid technique (Mejdahl 1985a). The liquids were made up by dissolving sodium metatungstate in water. The feldspars were treated with 10% HF for 40 minutes, in order to remove the outer layer and thereby eliminate the effect of external alpha radiation. A natural glow curve for potassium feldspar exhibiting two characteristic peaks at 240° and 325°C (heating rate 2° C/s), is shown in Fig. 17. The curve is typical for all samples except one (R-861012, unit Q4, sample 185), for which the high-temperature peak is visible only as a shoulder on the 240°C peak (Fig. 18). A TL growth curve for potassium feldspar is shown in Fig. 19. The curve can be approximated by a saturating exponential function:

$$Y = 24 + 308\,(1 - \exp(-0.000881X)) \quad (1))$$

where the dose X is measured in grays.

It is interesting to note that the reciprocal value of the exponential term, 1135 Gy, is very close to the value 1260 Gy (average value for ten feldspars) used in Mejdahl (1988b, eq. 2). A regenerated growth curve for a potassium feldspar from Kap København, North Greenland, gave the value 1217 Gy (Mejdahl 1985b). There is thus indication that alkali feldspars from a wide geographical region have common TL growth characteristics.

The dating technique used was regeneration combined with the plateau criteria (Mejdahl 1986, 1988a). The bleaching time was adjusted so that maximum plateau length was obtained in a

Table 7. Description of samples taken for TL dating from Saunders Ø and Qarmat. All samples were shallow water marine sediments, except sample R-861012, which is fluviatile. Sample location in sections appears from Figs 4 and 9.

Risø TL no.	Sample no.	Unit no.	Description
Saunders Ø			
R-861001	005	S2	Gravel with shells
R-861002	006	S4	Sand with shells
R-861003	007	S2	Sand with shells
R-861004	008	S4	Well-sorted sand
R-861005	009	S4	Gravel with shells
R-861006	010	S6	Clay
R-861007	011	S6	Clay
R-861008	012	S5a	Gravel
Qarmat			
R-861009	143	Q2	Clay with shells
R-861010	144	Q2	Beach sand w. shells
R-861011	147	Q3	Beach gravel w. shells
R-861012	185	Q4	Fluvial sand

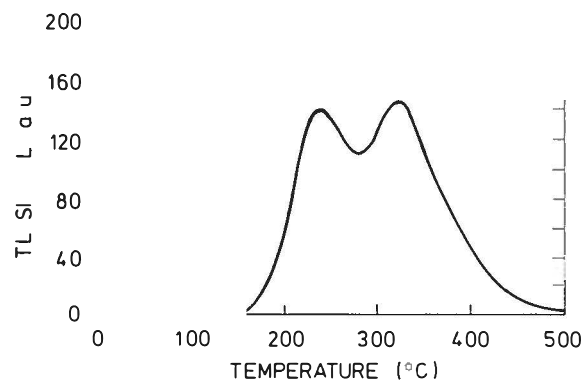


Fig. 17. Glow curve for potassium feldspar exhibiting characteristic peaks at 240 and 325°C. The curve is typical for all samples except R-861012. Heating rate 2°C/s.

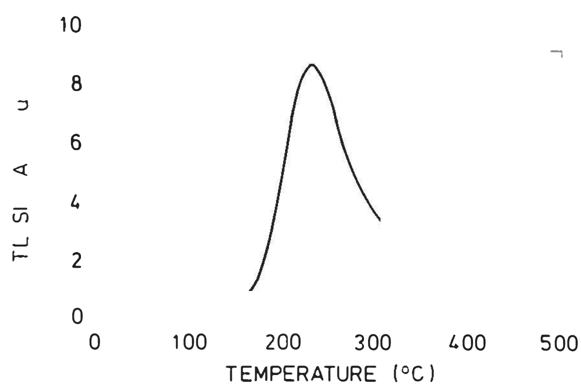


Fig. 18. Glow curve for R-861012, a potassium feldspar. The 325°C peak is visible only as a shoulder on the edge of the 240°C peak. Heating rate 2°C/s.

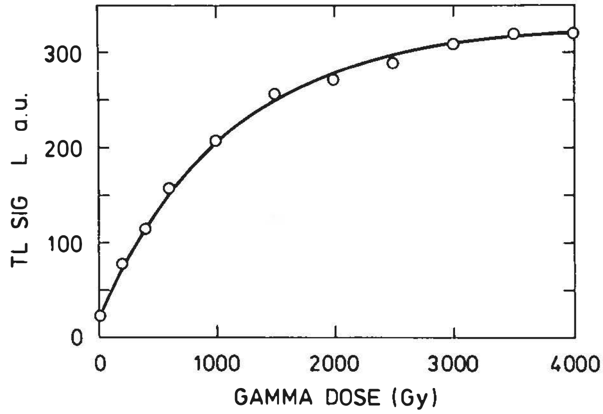


Fig. 19. TL growth curve for R-861006 obtained by adding doses to the natural dose. The TL signal was integrated over the temperature region 320–380°C (heating rate 8°C/s). The curve has been approximated by an exponential function:
$y = 24 + 308\,(1 - \exp(-0.000881X))$
where the dose X is measured in grays.

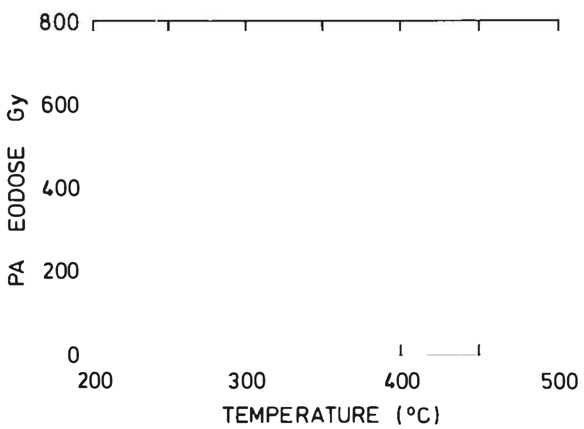


Fig. 20. Plot of palaeodose as a function of temperature for R-861005 (potassium feldspar). The plateau (standard deviation less than 5%) extends from 200 to 490°C.

plot of paleodose vs. temperature (Fig. 20). Our light sources were either the sun (in most cases), or a fluorescent lamp (Philips TL05/40 W) which was found to have the same bleaching effect as sunlight. The possibility of a sensitivity change as a result of bleaching was tested for two samples (R-861006 and R-861007, unit S6, samples 010 and 011), where the natural TL signal was in the linear region. The change was found to be negligible (less than 3%).

Dose rate determination

Dose rate components for alkali feldspar grains include: gamma + cosmic radiation, external beta radiation from the material surrounding the grains, internal beta radiation from potassium and rubidium in the crystal lattice, and internal alpha radiation from uranium and thorium embedded in the grains.

Gamma + cosmic radiation was measured on site, using a sodium iodide scintillation counter. The energy threshold was set at 0.4 MeV.

Non-finite-matrix beta radiation was determined by means of TL dosimeters (Mejdahl 1978) on samples that had their natural water content. This content was assumed to be representative for the total post-sedimentation period.

The uranium concentration in the samples was determined by the delayed neutron counting technique (Kunzendorf *et al.* 1980). For all samples, the contribution from uranium to the beta dose rate was less than 15%. Any effect of disequilibrium in the uranium series was, therefore, assumed to be negligible. Also the possible effect of excess Th-230 and Pa-231 (Wintle & Huntley 1980, Huntley & Wintle 1981) was neglected.

The potassium concentration in the grains was determined by beta counting, as described by Bøtter-Jensen & Mejdahl (1985). The beta dose rate contribution from rubidium was assumed to be 0.3 times that from potassium (Mejdahl 1987).

The thorium concentration was assumed to be 2.36 times that of uranium (Mejdahl 1987). Alpha dose rates were calculated using conversion factors from Nambi and Aitken (1986) and assuming an alpha efficiency factor of 0.2.

All dose rate data have been assembled in Table 8. The natural water content of the samples (wet weight over dry weight) is also shown. It is clear from Table 7 that some of the samples have very small water contents, which are hardly representative for the deposits in their marine environment. It is necessary, therefore, to estimate the water content that the deposits had while they were submerged.

A maximum value would be W = 1.59, as measured by Wintle and Huntley (1980) for a deep-sea core. The sediments from Greenland might, however, be more comparable to sediments in the North Sea. We have measured a water content of 18% (W = 1.18) for a drill core from the North Sea. The same value was found for a drill core from Storebælt. For comparison, the maximum value we have measured for water-saturated clay on land was W = 1.33.

Assuming a maximum water content of 30%, and assuming that the sediments have been above water for the last 70 ka, we can calculate corrections to the total dose rates in Table 8. The calculations show that the total dose rates for the first five samples would have to be reduced by about 5%. For the remaining seven, the correction would be negligible. If the maximum water content was 40%, the correction for the first five samples would be about 7%.

Palaeodoses and TL ages

As mentioned above, palaeodoses (doses accumulated since the last exposure to light) were obtained by the

Table 8. Dose rate data for feldspar grains, 0.1–0.3 mm. The external beta dose rate is the infinite-matrix beta dose rate (natural water content) multiplied by an attenuation factor 0.89. The internal beta dose rate originates from K and Rb in the grains. In the calculation of the internal alpha dose rate, an alpha efficiency factor of 0.2 was used. W is wet weight over dry weight. Field sample numbers appear from Table 7.

Risø TL No.	W	Gamma + cosmic (Gy/ka)	Beta Ext	(Gy/ka) Int	Alpha Int (Gy/ka)	Total (Gy/ka)
R-861001	1.12	0.93	1.42	0.79	0.08	3.22
R-861002	1.15	1.05	2.24	0.08	0.18	3.55
R-861003	1.10	0.95	2.05	0.70	0.15	3.85
R-861004	1.18	0.97	1.97	0.64	0.13	3.71
R-861005	1.14	1.00	1.98	0.78	0.15	3.91
R-861006	1.08	1.04	1.67	0.48	0.18	3.37
R-861007	1.09	0.86	1.42	0.71	0.12	3.11
R-861008	1.04	0.94	2.09	0.04	0.19	3.26
R-861009	1.24	1.34	2.33	0.40	0.15	4.22
R-861010	1.09	1.00	1.51	0.72	0.13	3.36
R-861011	1.05	1.03	1.78	0.48	0.23	3.52
R-861012	1.04	1.03	1.75	0.07	0.19	3.03

regeneration technique combined with the plateau criteria. The effect of short-term fading was eliminated by storing irradiated samples at 100°C for one week before measuring the TL signal.

Plateau regions, palaeodoses and TL ages are listed in Table 9. It can be seen that very long plateaux, in some cases 250°C or more, were achieved for the majority of samples. Longer bleaching times were required with the fluorescent lamp, because its intensity was only about one-fifth that of sunlight. For the oldest samples, the TL signal would be affected slightly by long term fading (Mejdahl 1988b), so 4 ka has been added to all TL ages exceeding 80 ka to correct for this effect.

The TL ages are affected by uncertainties in the dose rate arising from incomplete knowledge of the past water content of the sediments. As mentioned in the previous section, the five first dose rates in Table 8 should possibly be reduced by about 5%. The TL ages in Table 9 should then be increased accordingly. However, provided that the assumptions pertaining to the past variation of water content are basically correct, the corrections are too small to have any effect on the conclusions that can be drawn from the TL ages.

Discussion

Comparing Tables 7 and 9, one can see that the TL dates are consistent, i.e. correlatable units have similar

Table 9. Palaeodose data and TL ages. The palaeodose for alkali feldspar was obtained by regeneration combined with the plateau criteria. The plateau region is that over which the standard deviation of the points is less than 5%. The light sources used were the sun or a fluorescent lamp, type Philips TL05/40 W. A 4 ka correction for long-term fading has been added to ages exceeding 80 ka (Mejdahl 1988).
Sample location in sections appear from Figs 4 and 9.

Risø TL No.	Unit/ Sample No.	Light source	Exposure to light (h)	Plateau region (°C)	Palaeodose (Gy)	TL Age (ka)
Saunders Ø						
R-861001	S2/005	Sun	10	320–480	424	136±15
R-861003	S2/007	TL05	25	230–500	443	119±10
R-861002	S4/006	Sun	5	320–440	389	114±10
R-861005	S4/009	Sun	10	200–490	425	113±10
R-861004	S4/008	Sun	2	200–450	315	89±10
R-861008	S5a/012	Sun	3	330–460	224	69±10
R-861006	S6/010	Sun	1.5	270–360	47	14± 2
R-861007	S6/011	Sun	0.7	240–490	112	36± 4
Qarmat						
R-861009	Q2/143	Sun	8	310–470	344	86±10
R-861010	Q2/144	TL05	30	250–430	269	80±10
R-861011	Q3/147	Sun	3	320–500	283	80±10
R-861012	Q4/185	Sun	2	320–480	185	61± 6

ages, and ages become younger upwards in the sections.

A marked deviation occurs for sample R-861007 (unit S6 sample 011) where the TL age is considerably older than expected. At present, we have no explanation for this discrepancy, except that the TL result might refer to an earlier zeroing event.

Amino acid geochronology

(H.-P. Sejrup)

The problem of establishing absolute chronologies for raised marine deposits at high latitudes is well documented. In Arctic Canada, Greenland and Svalbard the use of isoleucine epimerisation in marine molluscs has become a standard tool in correlation and in establishing relative chronologies for high sea-level events, periods of influence of subpolar water and glacial advances. In the present study these results are evaluated relative to TL-dating from the same samples.

Methods and material

Since the first studies (Abelson 1955) which showed the potential for using diagenetic reactions in protein remains for geothermometry and geochronology, there has been an increasing interest in the method. At higher latitudes the isoleucine epimerisation reaction in amino acids preserved in carbonate fossils (molluscs and foraminifers) has been widely used as a geochronological tool for Quaternary sediments (Miller & Hare 1980, Miller 1985). The degree of isoleucine epimerisation is expressed as the ratio between the non-protein amino acid D-alloisoleucine (aIle) and the protein amino acid L-isoleucine (Ile). A modern sample has a laboratory induced content of aIle and giving aIle/Ile ratio close to 0.011. The epimerisation reaction reaches equilibrium at *c.* 1.3., which take a few million years in temperate regions and much longer in the arctic. The reaction is species-dependent and highly temperature dependent. It follows first order kinetics to a ratio between 0.3 and 0.4. For the mollusc *Mya truncata*, Miller (1985) has determined the relation between the reaction rate and temperature (Arrhenius parameters). This opens the possibility of calculating an absolute age if the temperature history is known, or diagenetic temperature if the age is known.

Seventeen samples of the cosmopolitan marine bivalve; *Mya truncata* were analysed for amino acids (Table 10). The samples have been prepared by the method described by Miller & Mangerud (1985). This preparation procedure includes firstly a leaching of the outer part of the shell and a thorough cleaning. One fragment is dissolved in 6 N HCl, hydrolysed under N_2 for 22 hrs at 110°C, and dried in a vacuum desiccator. Another fragment is just dissolved in HCl and then dried. The hydrolysed (free plus peptide bound amino acids) and the non-hydrolysed (free amino acids) samples are then analysed on an automated HPLC amino acid analyser. The degree of isoleucine epimerisation is recorded on an HP3033 computing integrator.

From each collection two or three individuals (A, B, and C) were analysed. Usually the samples were analysed more than once. In Table 10 all the mean values for each preparation are given. Mean ratios are calculated for each collection and show a spread smaller than 10%. The following discussions will mostly rely on the mean values of the hydrolysed samples.

Temperature conditions

The dependency of the rate of the isoleucine epimerisation reaction on diagenetic temperature (EDT) makes it important to evaluate the different temperatures to which the samples from the Thule area could have been exposed. Fig. 21 shows how the reaction rate changes with different temperatures, by using the Arrhenius parameters given by Miller (1985). In the Thule area three models can be postulated.

1. The localities were not covered by glaciers or by the sea. This situation gives the lowest EDT. The present mean annual air temperature at Thule is -9°C. The lowest possible mean air temperature to which the sites could have been exposed is -18°C which has been measured in some of the northernmost stations in Arctic Canada and Greenland. As is seen from Fig. 21 the reaction rate is extremely low under such temperature conditions.

2. The localities were covered by glacial ice. Except for the marginal zone, it can reasonably be assumed that the Inland Ice would be at the pressure melting point at its base. This gives temperatures close to 0°C. So contrary to what is found in temperate areas, a higher reaction rate should be assumed for the periods with ice cover than for ice-free periods with low sea-level stands.

3. The localities were submerged below sea-level. In periods of high sea-level the temperatures were therefore close to 0° C.

From this it is concluded that periods with no glacier ice or marine submergence introduce the largest uncertainties in the temperature history and that the highest EDT

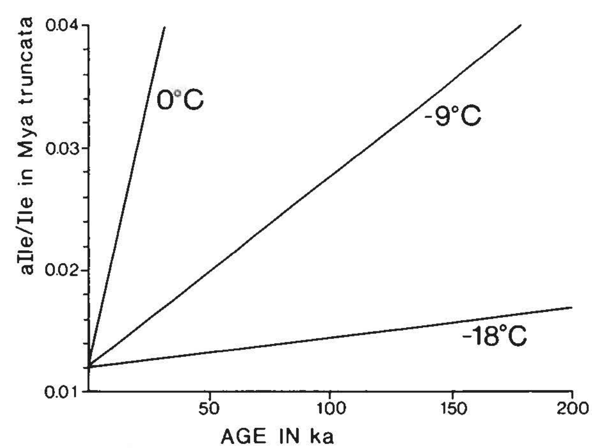


Fig. 21. Relationship between age and aIle/Ile ratios in the hyrdrolysed (total) fraction in *Mya truncata* for different diagenetic temperatures (EDT). Age-term is based on kinetic parameters from Miller (1985).

Table 10. AIle: Ile ratios in the total hydrolysate (HYD) and free fractions in shells of *Mya truncata* from the Thule area. Sample-location in sections appears from Figs 4, 8 and 9.

Field no.	Unit no.	Lab. no.	HYD	Hyd mean	FREE
Saunders Ø					
006	S4	BAL 1232A	0.017		0.221
		B	0.024	0.021±0.004	0.216
			0.024		
007	S2	BAL 1233A	0.036	0.036±0.001	0.394
		B	0.035		0.359
009	S4	BAL 1234A	0.052	0.049±0.005	
		B	0.045		0.203
010	S6	BAL 1235A	0.013	0.014±0.002	ND
		B	0.014		ND
012	S5a/2	BAL 1299A	0.038		0.205
			0.046	0.041±0.004	
		B	0.038		0.192
			0.043		0.218
013	surface	BAL 1300A	0.013	0.013	0.281
			0.013		
		B	0.021	0.023±0.002	0.221
			0.024		0.236
Narssârsuk					
102	N5	BAL 1433A	0.024	0.024	0.290
			0.023		0.397
		B	0.027	0.025±0.016	0.231
			0.024		0.268
			0.026		
		C	0.037	0.036±0.002	0.306
			0.034		0.393
104	N6	BAL 1529A	0.015	0.016	0.221
		B	0.016		ND
112	N1	BAL 1236A	0.028	0.030±0.002	0.224
		B	0.031		0.277
117	N5	BAL 1288A	0.044		0.289
		B	0.046	0.044±0.002	0.249
			0.042		
		C	0.033	0.033±0.001	0.276
			0.032		
126	N2	BAL 1432A	0.031	0.031±0.006	0.213
			0.023		
			0.025		
		B	0.035		0.215
			0.042		0.264
			0.035		
		C	0.035		0.115
			0.027		0.182
129	N4	BAL 1301A	0.028	0.026±0.002	0.227
			0.026		0.240
		B	0.024		0.222
			0.026		0.242
130	N5	BAL 1289A	0.039	0.035±0.003	0.350
			0.032		
		B	0.034		0.405
			0.033		
		C	0.032		0.484
			0.039		
183		BAL 1303A	0.059	0.064±0.004	0.139
			0.067		0.341
			0.066		
		B	0.050	0.049±0.002	0.259
			0.047		0.253
Qarmat					
142	Q1	BAL 1237A	0.044	0.037±0.009	0.356
		B	0.030		0.375
143	Q2	BAL 1302A	0.025	0.025±0.004	0.313
			0.030		0.356
		B	0.021		0.275
			0.025		0.318
147	Q3	BAL 1238A	0.024	0.025±0.01	0.310
		B	0.026		0.309

was reached under periods of submergence or glacial advance.

Local aminostratigraphy

In Fig. 22 the the aIle:Ile ratios in the free fraction are plotted against those in the total hydrolysate. Ratios in the total acid hydrolysate are considered to have the best time resolution, and the following discussion will focus on these data. In the total fraction aIle:Ile ratios vary between 0.013 and 0.064. Two samples from Saunders Ø (sample 010 and one subsample of 013) which have been C-14-dated to 8–9 ka, yielded ratios between 0.014 and 0.013. One sample with a ratio of 0.016 is also considered to be in this category (sample 104).

The sample with the highest ratios, sample 183 from Narssârssuk, is the only sample which was collected from the terrain surface, and since, owing to prolonged exposure, it may have suffered abnormally high temperatures, this sample will not be discussed further. The remainder of the samples show a variation in mean

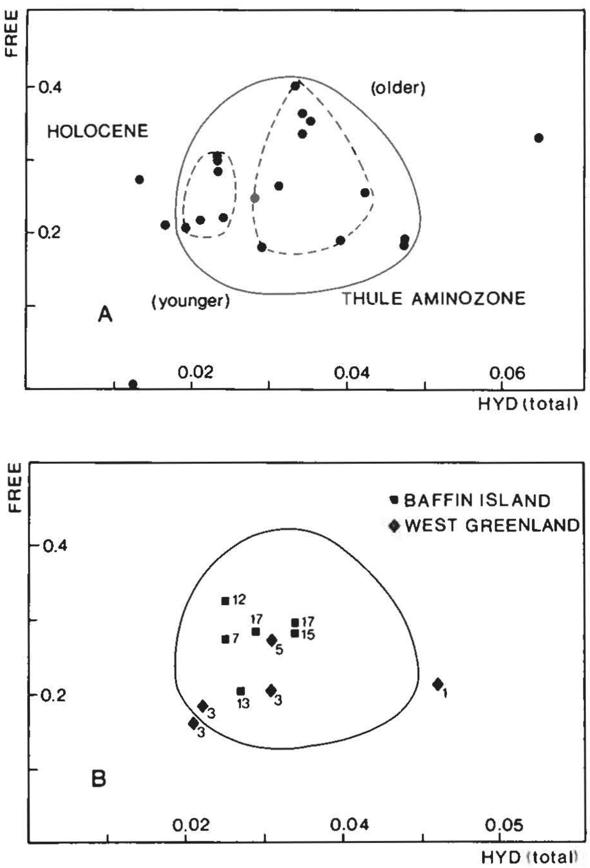


Fig. 22. Plot of *Mya truncata* and *Hiatella arctica* aIle/Ile ratios in the free versus hydrolysed (total) fraction of A: all samples analysed from the Thule area; and B: other samples from the Baffin Bay region (site numbers refer to Fig. 25). Ratios characteristic of the Thule aminozone are encircled.

values between 0.021 and 0.049, all coming from units with radiocarbon dates >40 ka (see appendix), and with TL-dates between 69 and 136 ka (Table 9).

A major problem is: do these amino acid ratios represent several marine events, or is the spread in ratios due to the analytical error, exposure to different local temperature regimes or are they results of "intrashell" variation as demonstrated previously by Brigham (1983b) and Sejrup (1985)?

Two of the samples analysed, both from the marine unit at Qarmat, gave mean values close to 0.025, while a sample consisting of fragmented molluscs from the underlying till gave a higher ratio, 0.037, (samples 142, 143 and 147). The obvious interpretation here is that the upper unit Q3 is characterised by the 0.025 ratio, and the shell fragments in the till are reworked from an older marine episode, preceding the glacier advance. Three TL dates from unit Q3 consistently give ages at *c.* 80 ka (Table 9). Hence it is a reasonable assumption that unit Q3 has an age of *c.* 80 ka, corresponding to amino acid ratios at 0.025.

The results from Narssârssuk are not quite as consistent. Reworked shell fragments from two samples in the glacial sequence (units N1-3), gave mean ratios of 0.030 (samples 112 and 126), similar to the results from Qarmat. However, from the overlying marine sediments, units N4-N5, the results show more scatter; two samples (102 and 129), gave ratios of 0.026 and 0.027, while two samples, 117 and 130, gave higher values at 0.035 and 0.039.

Unfortunately there are no TL dates from the sequence at Narssârssuk, and the amino acid results may be alternatively interpreted as: (1) the marine beds (units N4–N5) are characterised by a ratio of 0.025, as at Qarmat, while the higher ratios come from shells reworked from older marine sediment, or (2) the spread is analytical scatter, and the marine beds are best characterised by the mean of all measurements, 0.032 ± 0.006.

On Saunders Ø four samples from the pre-Holocene units yield ratios between 0.021 and 0.049 (samples 006, 007, 009, and 012), while TL dates from these units gave ages between 69 ka and 136 ka. The largest conflict between TL age and aIle/Ile ratio is in unit S5a (sample 012) where a "young" TL age of 69 ka corresponds to a "high" aIle/Ile ratio of 0.041. Considering the "dating conflict", the unusually heavy wear of the shell fragments in this sediment (Fig. 15), and the isolated occurrence of this subunit (Fig. 8), it is likely that the shell fragments were reworked from the older marine episode, represented by unit S2 from which a ratio of 0.036 has been obtained (sample 007), into sediments deposited during the youngest phase and belonging to unit S5. This is especially interesting because Blake (1975) used the fauna in this gravel bed for correlation with the Sangamonian.

Another example of reworking is afforded by sample 013 from gravel in a beach ridge marking the local marine limit. The ratios show that two age populations are present. An AMS C-14 date on a single shell fragment from this sample gave an age of 8.8 ka (see appendix), reflecting the age of the youngest population with a ratio of 0.013, while the older population with a ratio of 0.023 may correspond to unit S4, for which a ratio of 0.021 has been obtained (sample 006).

The results therefore show that while there is general agreement with TL dates, minor inconsistencies occur. Some of this can be explained by reworking and mixing of shells of different ages, but some of the scatter must be due to the reasons mentioned above. Therefore the amino acid ratios obtained for faunas at the three investigated localities show that they belong in the same general time interval, TL-dated to between 69 ka and 136 ka, and here called the Thule aminozone. The aminozone is characterised by its aIle:Ile ratios in the total (HYD) fraction (e.g. Miller & Hare 1980), ranging from 0.021 to 0.049.

With less confidence the aminozone can be divided into two marine events, as illustrated by the results from Qarmat, and to some extent those from Saunders Ø. The older episode is characterised by ratios of *c.* 0.039, the younger by ratios of *c.* 0.023.

A third marine episode is represented by values of 0.013 on Saunders Ø, C-14 dated to the Early Holocene (samples 010 and 013). Sample 104 from Narssârssuk is also referred to this episode although values are slightly higher (0.016).

Chronology

In the following, the chronology of the three marine episodes will be discussed relative to the TL dates and temperature/age discussion, applying the kinetic equations presented by Miller (1985). The three episodes are defined as follows:

I, Early Holocene, mean aIle/Ile 0.013 (samples 010 and 013A).

II. "younger Thule aminozone", mean aIle/Ile 0.023±0.002 (samples 006, 013B, 102, 129, 143 and 147).

III. "Older Thule aminozone", mean aIle/Ile 0.039±0.006. (samples 012, 117, 126, 142).

Using the present mean annual air temperature for the Thule area as EDT gives an age of 13 ka for episode I, 76 ka for episode II and 180 ka for episode III.

This suggests that if the TL dates from Qarmat are correct, conditions similar to those of today have prevailed for most of the Middle and Late Weichselian in the area. No long periods of ice cover or submergence below sea-level could have occurred, as such events would raise the diagenetic temperature to close to 0°C. This is also in accordance with the lack of Middle to Late Weichselian marine sediments in the area and suggests an absence of glacial isostatic depression. For episode III the age obtained by using present day temperature is in conflict both with the TL dates which span from 104 to 136 ka and also the fact that some of the units contain subarctic elements.

The lithology of the sediments shows that they have been below sea-level during several intervals in the period from 69 ka to 136 ka, and, at Narssârssuk and Qarmat, also overridden by a glacier. Therefore, as a first approximation, we could change the diagenetic temperature from -9°C to 0°C at 75 ka and then estimate the age of episode III. This gives an age of 93 ka which should be considered a minimum estimate since the diagenetic temperature for the period must have been lower, and between -9 and 0°C.

These considerations therefore support the evidence from the TLdates as well as the palaeoenvironmental interpretation, showing that the sediment sequences at Saunders Ø, Narssârssuk and Qarmat were deposited in marine episodes during isotope stage 5.

Conclusion

From the studies of isoleucine epimerisation in the marine mollusc *Mya truncata* from raised marine deposits in the Thule area, the following conclusions can be drawn:

1. Two, and probably three, marine episodes can be distinguished, and a correlation between the examined localities is established.
2. Although there is general agreement between amino acid results and TL-dating, minor inconsistencies occur. Some of these may be explained by reworking in the marine faunas.
3. Assuming likely diagenetic temperatures for the amino acid epimerisation, the results support the TL-datings and show that the pre-Holocene marine episodes can be referred to oxygen isotope stage 5 (69 to 128 ka), and that the subsequent period was colder.
4. The results support the lithological interpretation, showing that no major glaciation or submergence took place in this area in Middle or Late Weichselian times, between 69 and 10 ka.

Local events and regional correlation

SVEND FUNDER and MICHAEL HOUMARK-NIELSEN

Local correlation and event stratigraphy

(M. Houmark-Nielsen)

The sequences at all three sites described above contain one or more coarsening upwards sequences, each beginning with glaciomarine mud and diamicton, overlain by sublittoral marine sand, and ending up with beach shingle. These prograding sequences represent a sedimentary evolution from a transgressive maximum through a regressive phase to periods of reworking and halts in deposition. Since facies associations in all sequences are the same, local correlation must be based not only on lithology and stratigraphic position, but also on radiometric age determinations (C-14- and TL-dating), amino acid analyses of marine shells, and faunal characteristics (see also discussion by Mode *et al.* 1983).

The localities include Saunders Ø, Narssârssuk and Qarmat. For comparison Holocene marine sediments in a section at Nunatarssûp nûa, immediately in front of the present glacier front in Wolstenholme Fjord (Fig. 1), are included in this discussion. The proposed correlation is shown in Fig. 23.

Local correlation

An important tool for correlation between the sites is the amino acid ratios in marine mollusc shells, both *in situ* and reworked in glacial sediments. Sejrup (above) distinguished two major marine episodes: the younger characterised by ratios in the total hydrolysate (Hyd) below 0.016, is C-14 dated to the Holocene. The older, the Thule aminozone, has ratios between 0.021 and 0.049, and is TL-dated to isotopic stage 5 (69–136 ka). This includes the major part of the sediments described above from Saunders Ø, Narssârssuk and Qarmat, and the ratios provide correlation between the unit-sequences S2-S5, N1-N5 and Q1-Q3, all with C-14 ages older than 35 ka.

Within the interval of the Thule aminozone, on the evidence of the TL dates, glaciomarine sediments (unit S2, samples 005, 007; Fig. 4) apparently were deposited between 136±10 ka and 119±ka, (units S4-S5, samples 006, 008, 009) and around 114±10 ka and 86–80±10 ka (units Q2-Q3 samples 143, 144, 147; Fig. 9). The youngest part of the sequence, unit S5a (sample 012), which is composed of beach shingle gave a TL age of 69±10 ka.

Correlation of till beds

The till bed on Saunders Ø (S1, Fig. 23) is overlain by the oldest deposits within the Thule aminozone. This glacial event is possibly represented by the lower till in section K at Narssârssuk. Redeposited shell fragments in the till and associated glaciomarine outwash of units N1-N2-N3 and Q1 have amino acid ratios between 0.030 and 0.037, suggesting a correlation between the till beds of Narssârssuk and the lower till at Qarmat.

The upper till at Qarmat (Q5), which is younger than a TL age of 61±10 ka (unit Q4, sample 185; Fig. 9) has no equivalents, except possibly the lower till bed at Nunatarssûp nûa, which has a gradational transition upwards into glaciomarine diamicton of early Holocene age.

Prograding marine sequences

Three successive transgression-regression cycles can be identified from lithological criteria. All three cycles are recognised in the Saunders Ø sequence, two at Narssârssuk, and one at Qarmat and Nunatarssûp nûa. The youngest is dated to the Early Holocene while the two older belong in isotopic stage 5.

A clue to the correlation of the two older events is again afforded by amino acid ratios, where Sejrup (above) with some hesitation distinguished a lower and an upper marine episode within the Thule aminozone. Lithologically the older episode is represented by units S2 and S3 (Fig. 23). TL dates suggest deposition and progradation in the interval between 136±15 ka and 119±10 ka. The younger marine sediments, overlying till at Narssârssuk and Qarmat, comprises units S4-S5, N4-N5 and Q2-Q3. TL dates indicate that glaciomarine deposition and subsequent progradation took place somewhere between 114±10 ka and 80±10 ka.

The third cycle comprises units S6-S7, N6-N7, and the coarsening-upward sequence overlying the till bed at Nunatarssûp nûa; this cycle has been C-14 dated to the Early Holocene, although a TL-date from Saunders Ø and amino acid ratios from Narssârssuk suggest that it could have begun earlier.

Event stratigraphy and age

The deposits at the three localities were thus laid down during at least three depositional cycles, each including glaciation with till deposition and glacitectonic activity,

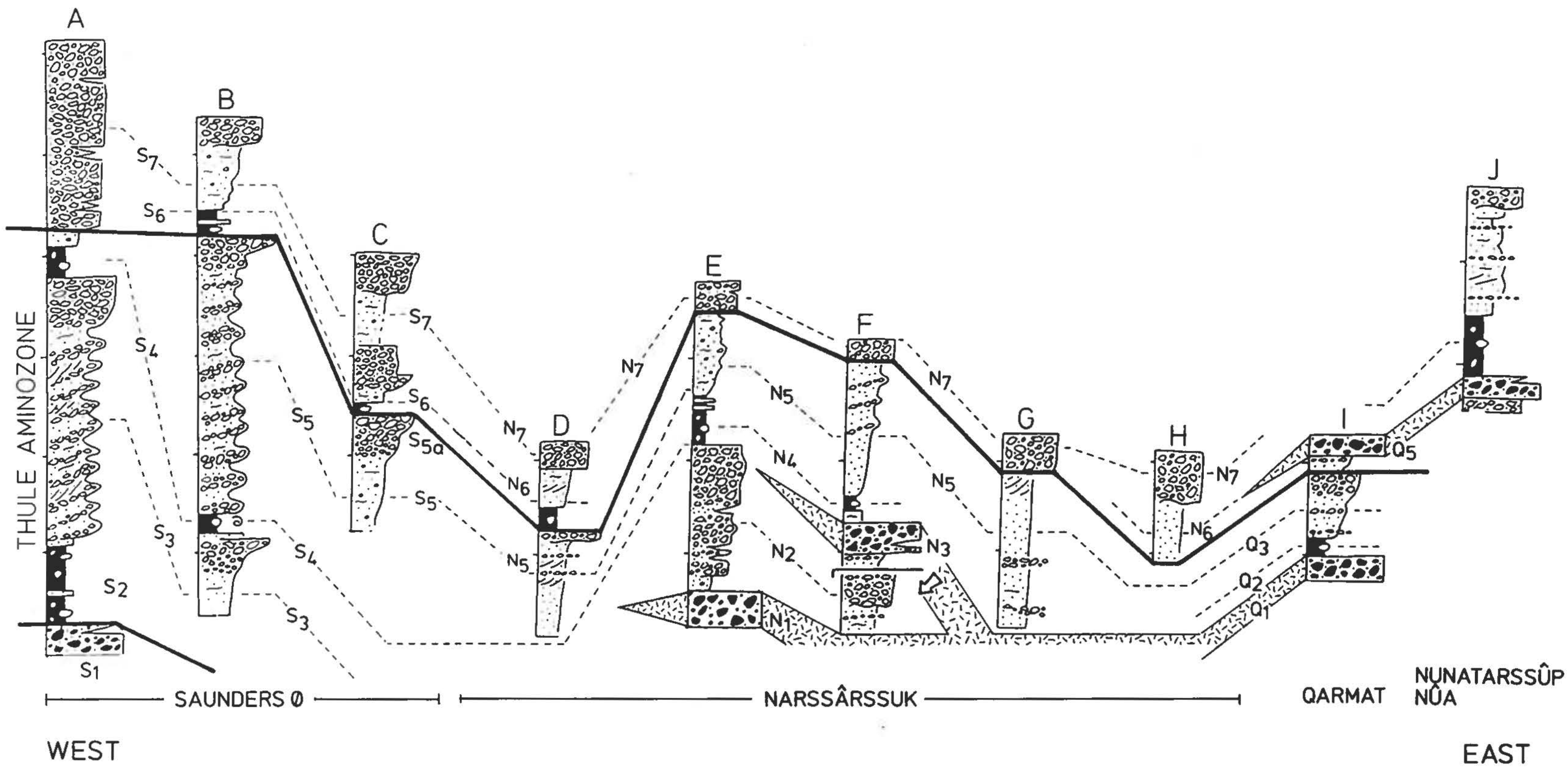


Fig. 23. Correlation between localities. Lithological sections compiled from Figs 4, 8 and 9.

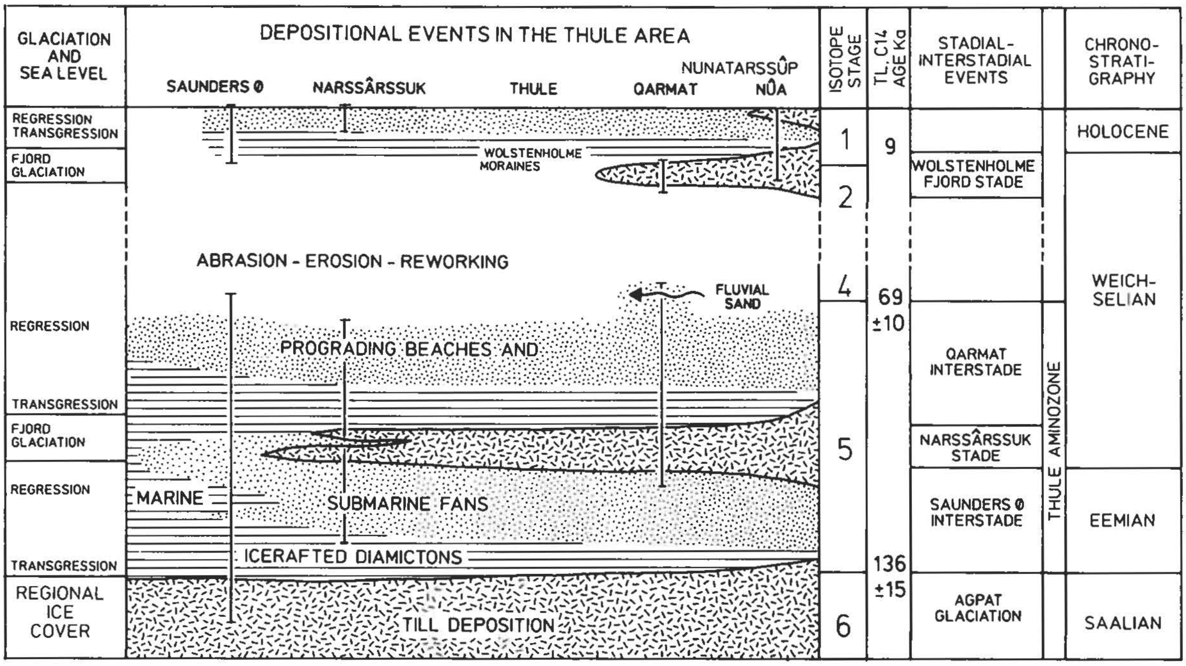


Fig. 24. Event-stratigraphic scheme for the Thule area in the period since isotope stage 6 (Saalian).

followed by marine transgression, glaciomarine sedimentation and subsequent progradation of coarse clastic sublittoral deposits and ending up with beach shingle (Fig. 24).

Based on this, a sequence of regional events can now be outlined, containing evidence from microfaunas (Fig. 11) and other fauna and flora (Tables 4, 5 and 6), and dated by TL-dating (Table 9), amino-acid analyses (Table 10), and radiocarbon dating (appendix). The distribution of ice cover during glacial events is shown on Fig. 10.

The main part of the sequence falls within isotopic stage 5 (74–130 ka, Martinson *et al.* 1987), but unfortunately neither TL-dating nor amino acid analysis have sufficient precision to refer our events to the 10 000 year substages within this interval. From oldest to youngest the following events are distinguished:

The Agpat glaciation. – This glaciation is represented by till (unit S1) on Saunders Ø, and possibly at Narssârssuk (unit a, section K). These till beds are provisionally correlated with the till cover found on upland plateaus, deposited by an ice sheet that covered the entire coastal region, extending into Baffin Bay. TL dates show that this event occurred prior to 136 ka, – and it is provisionally correlated with the Saalian (isotope stage 6).

The Saunders Ø interstade. – During the melting of the ice sheet a transgression caused deposition of ice-rafted diamicton and, later, glaciomarine mud on Saunders Ø (unit S2). During the subsequent regression progradation of a submarine delta fan ended up with deposition of beach shingle (unit S3). At Narssârssuk and Qarmat this period is known only from reworked marine fauna in younger glacial sediments. Sparse evidence from subarctic species suggest that the warm West Greenland current functioned as at present. TL dates indicate an age between 136 and 120 ka, and we provisionally correlate the event with a warm interval in early isotope stage 5, and with the Eemian chronozone (Bowen *et al.* 1986).

The Narssârssuk stade. – The regression was followed by renewed glaciation, this time of limited extent. At Narssârssuk and Qarmat till beds occur in the coastal sections (units N1-N3, Q1). The glacier did not reach Saunders Ø (Fig. 11), but high arctic foraminifera in lowermost unit S4 may represent a proglacial facies.

Moraines at Narssârssuk and Kap Abernathy on the north side of Wolstenholme Fjord mark the glacier margins during this advance. TL dates indicate an age of 114±10 ka, and we correlate this glaciation, the most extensive during the Weichselian, with one of the cold intervals in isotope stage 5.

The Qarmat interstade. – Again, this glacier advance was followed by isostatic depression resulting in deposition of ice-rafted diamicton at all three localities (units N5, Q2 and upper S4). Subsequent regression initiated beach-progradation (units S5, N5 and Q3). At Narssârssuk and Qarmat the sediments contain abundant subarctic organisms reflecting more extensive influx of subarctic water into northern Baffin Bay, and warmer summers, than known from the Holocene. By the end of the period sea-level dropped, and a final stage with sea-level 12 m above the present is recorded by subunit S5a on Saunders Ø. TL dates indicate an age between 114 ka and 69 ka, and we correlate this event with a warm interval in late isotope stage 5.

Hiatus. – The record, as outlined here, contains a gap of *c.* 60000 years from which almost no information is available. The lack of glacial sediments shows that ice coverage was similar to the present, or – more likely – less extensive, while the lack of raised marine sediments shows, that relative sea-level was lower than the present; this assumption is indirectly supported by calculation of diagenetic temperatures from amino acid ratios.

The Wolstenholme Fjord stade. – The upper till bed at Qarmat (unit Q5) records a later glacier readvance, recorded also by lateral moraines along Wolstenholme Fjord, where the glacier front was only 10 km more advanced than at present (Fig. 11). The areas at Narssârssuk and Saunders Ø were not ice covered, and at these two localities glaciomarine diamicton with a high arctic foraminifer assemblage may represent the proglacial marine facies, deposited during isostatic submergence following the readvance (units N6, S6).

Radiocarbon dates on shells from sediments immediately over the diamicton on Saunders Ø, and from a locality on the north shore of Wolstenholme Fjord have given ages between 8.2 and 9.2 ka, providing a minimum age for the advance (samples K-4780 and 4781), and amino acid ratios from Narssârssuk indicate a slightly higher age (sample 104). A TL-dating from an underlying fluvial deposit at Qarmat (unit Q4) gives a maximum age of 61 ka, while another from Saunders Ø at 36±4 ka is considered erroneous since it seems incompatible with the local sea-level history. Hence, based on C-14 dates and the Holocene sea-level history we assume a Late Weichselian age for this advance, corresponding to isotope stage 2.

The Holocene. – After the ice advance, ice-rafted diamicton and glaciomarine mud was deposited in the interior Wolstenholme Fjord at Nunatarssûp nûa, and – during the subsequent Holocene regression – a prograding sequence topped by beach shingle at all localities (units S7 and N7).

Marine faunas with subarctic molluscs show that the West Greenland Current was operating as at present at least at 9.2 ka BP. Shell fragments in beach sediment 40 m above sea-level on Saunders Ø have a Holocene amino acid ratio, and C-14 age of 8.7 ka giving a maximum age for the onset of the Holocene regression (sample 013, Table 10 and appendix).

Thule and Baffin Bay

(S. Funder)

In this section the Thule record will be compared to that from other parts of the Baffin Bay region, as well as with results from deep sea cores.

Because of its importance for understanding the history of the Laurentide ice sheet over North America intensive Quaternary studies have been carried out in this region, and a large amount of information has accumulated from both Canada and Greenland. Much of this data has been presented in a volume concerning the Quaternary environments (Andrews 1985), and later reviewed by Andrews *et al.* (1986) and Kelly (1986). Fig. 25 shows sites with marine and/or glacial sediments that have been ascribed an age similar to that of the Thule record.

Regional amino acid stratigraphy

Amino acid analyses of mollusc shells have played an important role in correlation of these events; however ratios obtained prior to 1982 have been assayed by new standard preparation methods. Fig. 22B shows the distribution of post-1982 amino acid ratios in *Mya truncata* or *Hiatella arctica* shells from sediments on the Baffin Bay coasts assigned to isotopic stage 5.

It was argued by Miller (1985) that although the region spans a considerable north-south climatic gradient, effective diagenetic temperatures (EDT) are most strongly controlled by summer warmth. Ratios from the 1500 km coastline of Baffin Island, where summer temperatures are very uniform, thus would have experienced the same thermal history and may be compared directly. Kelly (1986), on the other hand, found that in West Greenland separate epimerisation trends can be recognised in the north and south.

Our results (Sejrup, this volume) indicate that sea-level history may be more important than atmospheric temperatures, because it determines a shift between high and low EDT. However, for the sites shown in Fig. 1, i.e. West Greenland north of lat. 69° N and Baffin Island north of lat. 65° N, sea level history seems to have been the same, and it is therefore postulated that within this region thermal conditions have been suffiently uniform to allow correlation of mollusc faunas by their amino acid ratios.

A comparison of the data (Fig. 22) gives support to earlier age estimates, and indicates that marine events during isotope stage 5 are correlatable around Baffin Bay.

Correlation of events on land

The proposed correlation is shown in Fig. 26 which shows that the Olrik Fjord, Kaffehavn and Svartenhuk aminozones of West Greenland (Kelly 1986), as well as

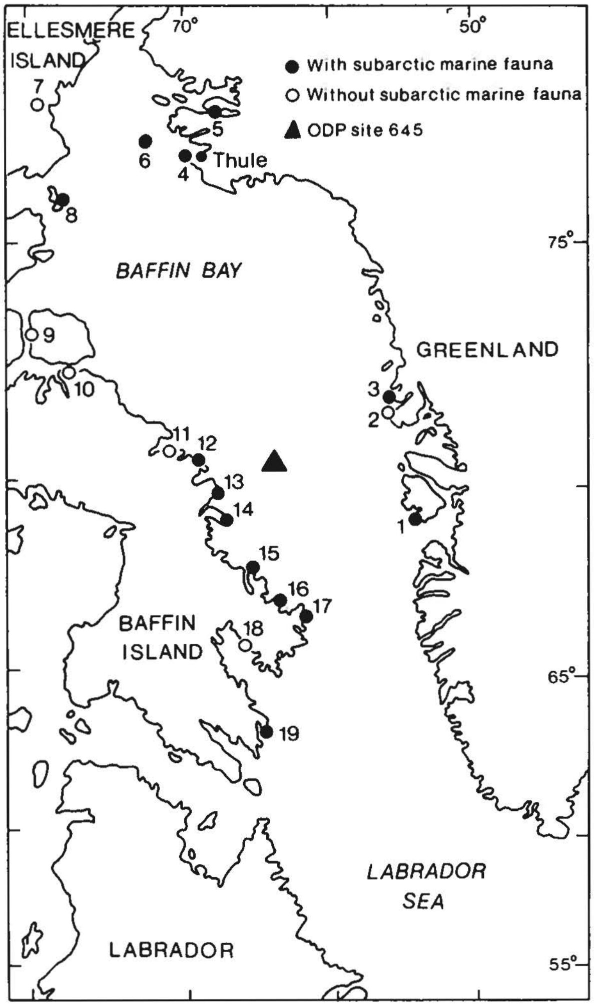


Fig. 25. Sites with marine and/or glacigene sediments ascribed to isotope stage 5 in the Baffin Bay region and site for ODP drilling.
Sites and sources: 1. Laksebugt (Funder & Símonarson 1984). 2. Søndre Upernavik and adjacent sites (Kelly 1986). 3. Kaffehavn (Kelly 1986). 4. Thule area (this work). 5. Olrik Fjord (Weidick 1978, Kelly 1986). 6. Isbjørneø (Blake 1977). 7. Makinson Inlet (Blake 1980). 8. Coburg Island (Blake 1973). 9. Bylot Island (Klassen 1985, 1987). 10. Pond Inlet (Klassen 1985). 11. Scott Inlet (Miller *et al.* 1977, Andrews *et al.* 1981). 12. Clyde Foreland (Miller *et al.* 1977, Andrews *et al.* 1981; Miller 1985). 13. Cape Henry Kater (Andrews *et al.* 1981, Miller 1985). 14. Qivitu Peninsula (Nelson 1981, Miller 1985). 15. Broughton Island (Brigham 1983b, Miller 1985). 16. Cape Dyer (Andrews *et al.* 1981, Miller 1985). 17. Cape Raper (Andrews *et al.* 1981, Miller 1985). 18. Pangnirtung (Miller 1985). 19. Allen Island (Osterman *et al.* 1985).

the Kogalu aminozone of Baffin Island (Miller 1985), may be correlated with the Thule aminozone, and thus dated within the interval 69–135 ka. For the southernmost locality (Laksebugt, site 1, Fig. 25) ratios are higher, and may reflect higher EDT in this area.

A characteristic feature of the marine sediments at almost all sites is their association with evidence for glacier advance. On Bylot Island, off northern Baffin Island (site 9, Fig. 25), this evidence is referred to the Eclipse glaciation (Klassen 1985, Klassen & Fisher 1988), and to the south on Clyde Forland (site 12, Fig. 25) to the Ayr Lake stade (Miller *et al.* 1977, Miller 1985), which from their setting and amino acid characteristics can be correlated to our Narssârssuk stade, with a TL age of 114±10 ka.

Although this glacier advance in both areas marks the maximum of Weichselian glaciation, it was more dramatic on the Canadian side; thus whereas the glacier in Wolstenholme Fjord was only 40 km more advanced than now, the straits between the Canadian arctic islands were closed by glaciers (Klassen & Fisher 1988, Osterman *et al.* 1985), and on parts of Baffin Island fjord glaciers reached the outer coast and spread out as large lobes (Dyke *et al.* 1982).

In both areas the glacier advance was accompanied by isostatic subsidence and deposition of marine sediments with subarctic faunas during the Qarmat interstade of Thule, and the late Kogalu aminozone of Baffin Island. At Thule the presence of *Menyanthes trifoliata* and *Balanus balanoides* indicate warmer summers and more abundant subarctic water than known from the Holocene (see p.32), and the subarctic water continued around the northern perimeter of Baffin Bay, flowing south along the coast of Baffin Island. This is shown by the frequent occurrence of the subarctic *Chlamys islandica* in areas that are now dominated by cold polar water, as well as the gastropod *Colus spitsbergensis* on Clyde Foreland. However *Mytilus* – common in the sediments at Thule – has not been reported from sites along the Baffin Island coast (Andrews *et al.* 1981).

The marine episode preceding the glacier advance, the Saunders Ø interstade, is less well known. This is undoubtedly due to the facts that sea-level was close to its present level, and that sediments at most sites were later eroded by glaciers. As noted above, *Chlamys islandica*, *Balanus crenatus* and possibly also *Mytilus* lived at Thule, showing that the warm West Greenland current was functioning as at present. From its setting it may be suggested that the horizon with well preserved *Mytilus* on Coburg Island (site 8, Fig. 25) should be referred to this episode, indicating that also during this period subarctic water penetrated deeper into northeastern Baffin Bay than during the Holocene (Blake 1973).

From Baffin Island only a few deposits are referred to the pre-Ayr Lake stade episode (e.g. Miller *et al.* 1977), and we have not been able to find records of subarctic marine fauna. However, on Broughton Island (site 15, Fig. 25) a higher than Holocene abundance of *Betula* pollen in a horizon immediately below glaciomarine sediments of the Ayr lake stade is "...considered to be indicative of interglacial, terrestrial climatic conditions ... when relative sea-level stood at < 5 m asl." (Brigham 1983a: 589). On Bylot Island (site 9, Fig. 25) well devel-

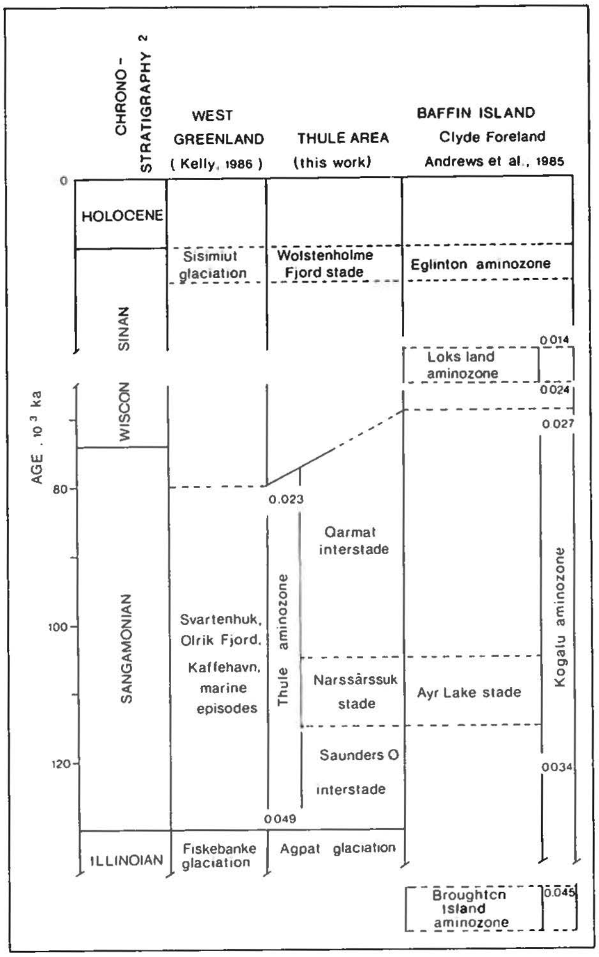


Fig. 26. Proposed correlation of glacial and marine events in Greenland and on Baffin Island.

oped buried soils occur in a similar stratigraphic position (Klassen 1987).

Sparse as it is, this evidence indicates more substantial influx of subarctic water, and a warmer climate, than during the Holocene also for the time interval represented by the Saunders Ø interstade.

Before this period, extensive glaciation occurred on both sides of Baffin Bay during the Baffin glaciation of northern Baffin Island (Klassen 1987), and the Agpat glaciation in the Thule area. Although there is no dating of these events, the similarities in stratigraphical setting and nature of the evidence suggests a correlation.

A salient feature in the Thule record is the hiatus from *c.* 70 to *c.* 10 ka. From Baffin Island a slightly different timing of events was suggested by Andrews *et al.* (1986). Here the warm Kogalu aminozone extended into isotopic stage 4, and marine transgression occured later, at *c.* 44 ka, during the Loks Land aminozone. However, the dating of these events relies on age calculations from amino acid ratios which are similar to those obtained for the Thule aminozone, suggesting the possibility that both the Loks Land and the older Broughton Island aminozones should be correlated with the Thule aminozone, and referred to isotope stage 5 (Fig. 26).

Finally, in both areas there was a limited resurgence of glaciers in the Late Weichselian or Early Holocene, the Cape Hatt glaciation of Bylot Island (Klassen 1987), the Eglinton aminozone of Clyde Foreland (Miller 1985), and the Wolstenholme Fjord stade in Thule. This advance was accompanied by renewed transgression with influx of warm subarctic water, beginning before 9.2 ka at Thule, before 8.3 ka on northern Baffin Island, and at 9.7 ka on southern Baffin Island (p. 28).

Deep sea results

The results from deep sea boring and piston coring in Baffin Bay have recently been discussed by De Vernal *et al.* (1987), De Vernal & Hillaire-Marcel (1988), and Hillaire-Marcel *et al.* (in press), who showed that peaks of melt water influx have disturbed the oxygen isotope signal, making correlation with the global oxygen isotope stratigraphy difficult. However, the results from deep sea drilling at ODP sites 645 and 646 indicate that during isotope substage 5e the Labrador Sea and Baffin Bay had warmer than present surface water, while high erosion rates and abundant reworked pre-Quaternary palynomorphs reflect intense glacial activity around Baffin Bay during the later part of isotopic stage 5 (De Vernal & Hillaire-Marcel, 1988), which may be correlative with the Ayr Lake stade/Eclipse glaciation/Narssârssuk stade on land. However, the deep sea record contains no evidence for late stage 5 influx of subarctic water to northern Baffin Bay, as shown by the faunas of the Qarmat interstade and late Kogalu aminozone on land.

Discussion

The evidence shows that oceanographic and glacial events in northern and eastern Baffin Bay at the onset of the last ice age were in phase, and support the contention by Andrews *et al.* (1986) that glaciation in this region may be controlled by presence of warm water along the coasts, bringing moisture into regions which today have a cold and dry climate unfavourable for formation of ice on land.

A major unresolved problem remains: why did the northern Baffin Bay region apparently have conditions warmer than today during the Early Weichselian, at a time when our hemisphere seems to have experienced the temperature decline heralding the coming of the ice age?

Conclusions on the Quaternary stratigraphy of the area

ROLF. W. FEYLING-HANSSEN, SVEND FUNDER, MICHAEL HOUMARK-NIELSEN, MICHAEL KELLY, CHRISTIAN KRONBORG, JON Y. LANDVIK, VAGN MEJDAHL, HANS-PETTER SEJRUP and LENNART SORBY.

– Alluvial cones along Wolstenholme Fjord reflect at least three distinct transgression-regression cycles, each starting with glaciomarine mud or till and ending with littoral sand and shingle, and presumably controlled by isostatic movements in response to changing ice coverage.

– Extensive C-14 and thermoluminescence dating show that the most recent cycle began in the Early Holocene/ Late Weichselian, while the two older ones took place during early and late isotope stage 5 (Saunders Ø and Qarmat interstades).

– Amino acid analyses of marine bivalve shells define the Thule aminozone with a TL age of 136–69 ka, and a younger zone, correlative with the Late Weichselian/ Holocene.

– The amino acid results are consistent with the TL-dates, if it assumed that the rate of amino acid diagensesis has been controlled mainly by sea-level history, determining shifts between periods with high and low diagenetic temperatures (when sediments were below or above sea-level respectively).

– Foraminifer and mollusc faunas show that subarctic water penetrated as far north in Baffin Bay as it does today during all three marine cycles.

– The presence of the barnacle *Balanus balanoides*, and the plant *Menyanthes trifoliata*, as well as abundant well preserved *Mytilus edulis* and *Chlamys islandica* show that conditions were warmer during the Qarmat interstade than as yet recorded for the Holocene.

– Three periods of glacier advance have been recognised: the Agpat glaciation (?Saalian), the Narssârssuk stade (114 ± 10 ka), and the Wolstenholme Fjord stade (Late Weichselian), each less extensive than the preceding one.

– Similarities in setting and lithology, together with amino acid results, provide a base for correlation with sites elsewhere in northern Baffin Bay, and provide for the first time a link between events in North America and Greenland.

– During isotope stage 5 oceanographic and glacial events in the regions around Baffin Bay, north of lat. 69° N in West Greenland and of lat. 65° on Baffin Island, were in phase.

– This foregoing conclusions indicate that there is a strong oceanographic control over glaciation in this region, as has been pointed out earlier by researchers working in the Canadian Arctic.

Stable isotope studies on ice margins in the Thule area

NIELS REEH, HENRIK HØJMARK THOMSEN, POUL FRICH and HENRIK B. CLAUSEN

The main purpose of the glaciological programme carried out by the NORDQUA 86 expedition to the Thule area, Northwest Greenland was to investigate to what extent stable isotope studies in the marginal areas of the Inland ice and local ice caps could provide information about the dynamic and climatic history of the Northwest Greenland ice cover during the last glacial/interglacial cycle.

The expectations that stable isotope studies on the ice margins could actually contribute such information were based on results of previous isotope studies at other Greenland ice-margin locations. Such studies have documented that ice, deposited in the inland regions of the ice sheet during the Weichselian glaciation (here considered equivalent to the Wisconsinan of North America), is now exposed at the surface of the ice sheet in a several hundred metre wide band near the ice edge (Reeh *et al.* 1987a). Also, studies on Barnes ice cap, Arctic Canada (72°W, 69.5°N) have revealed that ice of Wisconsinan origin is still present at the margins of this relatively small ice cap (Hooke & Clausen 1982).

The transition from the Weichselian to the Holocene is characterised by a large step-like change in $\delta^{18}0$ from large negative values in the Weichselian to less negative values in the Holocene. This has been found in deep ice-core records from the inland regions of the Greenland ice sheet (e.g Dansgaard *et al.* 1984), as well as in ice-margin records (e.g. Hooke & Clausen 1982, Reeh *et al.* 1987a, Clausen & Stauffer 1988). Even though there are other sources for the variations found in $\delta^{18}0$ records (Fisher & Alt 1985), $\delta^{18}0$ variations on the Greenland ice sheet are believed to be mainly controled by the condensation temperature at the snow-deposition site (Dansgaard *et al.* 1973). The size of the $\delta^{18}0$ shift at the Weichselian/Holocene transition, therefore, not only depends on the magnitude of the climatic temperature change from glacial to interglacial conditions, but also on the change of surface elevation at the snow-deposition site. A comparison of the $\delta^{18}0$ values at the margin of the local ice cap at Tuto Ramp (see map in Fig. 27) with the $\delta^{18}0$ values at the margin of the Greenland ice sheet at Nuna Ramp, and with the $\delta^{18}0$ values of the deep core-ice from Camp Century located *c.* 150 km inland of the ice margins, could therefore reveal possible changes in the dynamics of the Northwest Greenland ice cover at the end of the Weichselian.

However, these expectations were not fulfilled, because only ice of Holocene origin was found at the ice margins in the Thule area. Therefore, the isotope studies can only provide information about the present geographical $\delta^{18}0$ distribution in the Thule area, and to some extent throw light on the inland deposition site for the ice at the margins, and on the thermal conditions at the base of the ice covers.

The results of the isotope analysis are also used to derive input data to ice dynamic models for the ice margins, and to check the results of model calculations. The calculations indicate that a few hundred metres from the ice margins, the surface ice is 2000–3000 years old, and, therefore, that ice of Weichselian origin, if present at all, can at most constitute only a narrow band near the margin.

Calculations also show that, very likely, a significant part of the Weichselian ice has been melted away at the bottom of the ice covers or has simply been removed by increased surface ablation during the Holocene climatic optimum. It is even possible that increased ablation during this period could have resulted in complete obliteration of local ice caps in the Thule area.

Glaciological setting

The ice margins in the Thule area have been intensively studied in the 1950's and 1960's by U.S. Army Snow Ice and Permafrost Research Establishment (SIPRE) and U.S. Army Cold Regions Research & Engineering Laboratory (CRREL). The glaciological investigations comprised observations of mass balance, ice temperature, surface-ice velocity and deformation, and investigations in two 400 m tunnels excavated into the ice margin at the Tuto Ramp. The results were published in SIPRE and CRREL reports, e.g. Schytt (1955) and Nobles (1960). A more recent study was carried out by Hooke (1970).

The Tuto Ramp is situated *c.* 20 km southeast of Thule Air Force Base (see map in Fig. 27). From the ice margin at an elevation of *c.* 500 m, the surface displays a gentle, upward slope towards a local dome located *c.* 10 km east of the ice margin at an elevation of 900–1000 m (Figs 29 and 30a). The dome is connected to the main Greenland ice sheet by a saddle and a more than 30 km long, undulating ridge with elevations between 700 and 850 m (Benson 1962). Therefore, at present, the local dome is dynamically independent of the main ice sheet, and ice from there cannot reach the Tuto Ramp area. This must have been the case as long as the "Tuto ice dome" has been separated from the main ice sheet by a saddle.

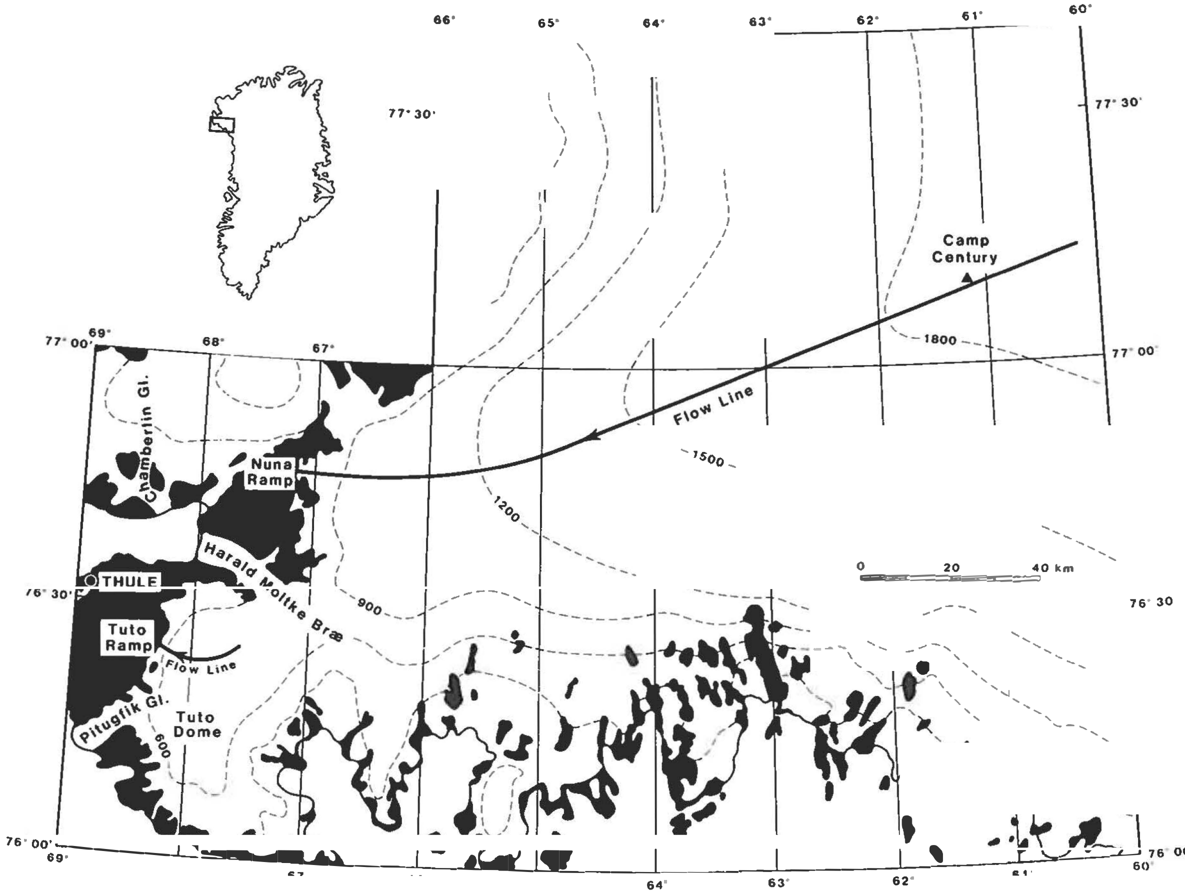


Fig. 27. Map of the Thule penisula and the Camp Century area. Flow lines leading to Tuto and Nuna ramps are indicated.

The Nuna Ramp, located about 50 km northeast of Thule Air Force Base, forms part of the ablation zone of the main Greenland ice sheet. This means that the flow lines terminating at the Nuna Ramp have their origin somewhere inland on the ice sheet, probably on the ridge on which the Camp Century deep drilling site is also located (see map in Fig. 27).

Therefore, the ice at the two ramps should originate from very different locations, and this must have been the case for as long as the present flow pattern has remained essentially unchanged. However, if the main ice sheet, for example during a stage of more extensive glaciation, had overridden the "Tuto ice dome" then the ice deposited as snow during that period would have originated further inland at a high elevation and consequently would carry a relatively low $\delta^{18}O$ value.

As already mentioned, only ice of Holocene origin was found at the surfaces of the ramps. This also applies to the other sampling locations in the area. Possible explanations are discussed in a later section.

Sampling programme

More than 450 samples of water, ice and snow were collected from 9 different locations (Fig. 28). On the Tuto Ramp (location 1) samples were collected along 3 profile lines transverse to the ice margin, 885 m, 147 m, and 210 m long, respectively (Fig. 29). The sampling interval varied between 5 and 20 m. At two locations (2 and 3) on "Tuto ice dome" at elevations of 850 m and 1010 m, respectively, snow samples were collected from 2 m deep snow pits, to provide a present-day reference $\delta^{18}O$ value from the accumulation area feeding the Tuto Ramp.

Some 60 samples were collected from the bottom to the top of an 11 m, near-vertical ice cliff at the southern part of Store Landgletscher (location 4). The ice cliff had a layered structure with numerous dirt bands and dirt inclusions, supporting the interpretation that the cliff consists of ice layers formed by refreesing of meltwater at the base of the ice cap at some distance behind the ice margin. At the same location, ice samples were collected along a 500 m profile at the ice surface leading to the cliff.

Further ice samples were collected from a local piedmont glacier at the northern margin of Store Landgletscher (location 5), from two profiles, respectively 110 m and 75 m long, transverse to the ice margin at the

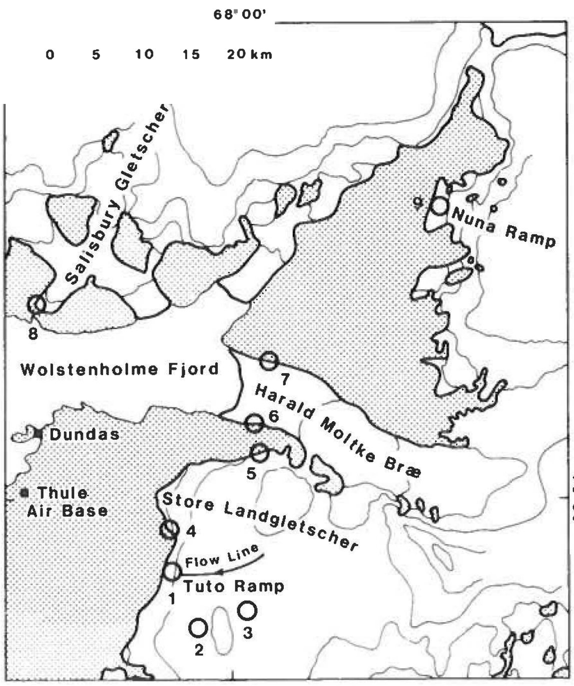


Fig. 28. Location map of the Thule area. Sample locations are marked with circles. Flow line leading to Tuto Ramp is indicated.

southern and northern part of the calf-ice producing outlet glacier, Harald Moltke Bræ (locations 6 and 7), and from the moraine covered ice in front of Salisbury Gletscher (location 8).

Finally, samples were collected along a profile, 1001 m long, transverse to the ice margin at Nuna Ramp at intervals between 4 and 20 m (location 9).

The summer of 1986 was unusually cold in the Thule area, and in the period of sampling around mid August, even the lowermost parts of the ice margins were still covered by up to 0.5 m of snow from the previous winter's accumulation. This impeded the sampling enormously, since the ice surface had to be cleared for snow before the samples could be taken. In fact, the deep snow cover prevented sampling of a 200–300 m section of the surface nearest to the margin. Also, due to the presence of the previous winter's snow, one cannot completely exclude the possibility that some samples were taken from a layer of superimposed ice below the snow cover, even though the upper ice layer was cut away down to the depth of the cryochonite horizon before the samples were taken. According to Schytt (1955), this should ensure that real glacier ice was reached.

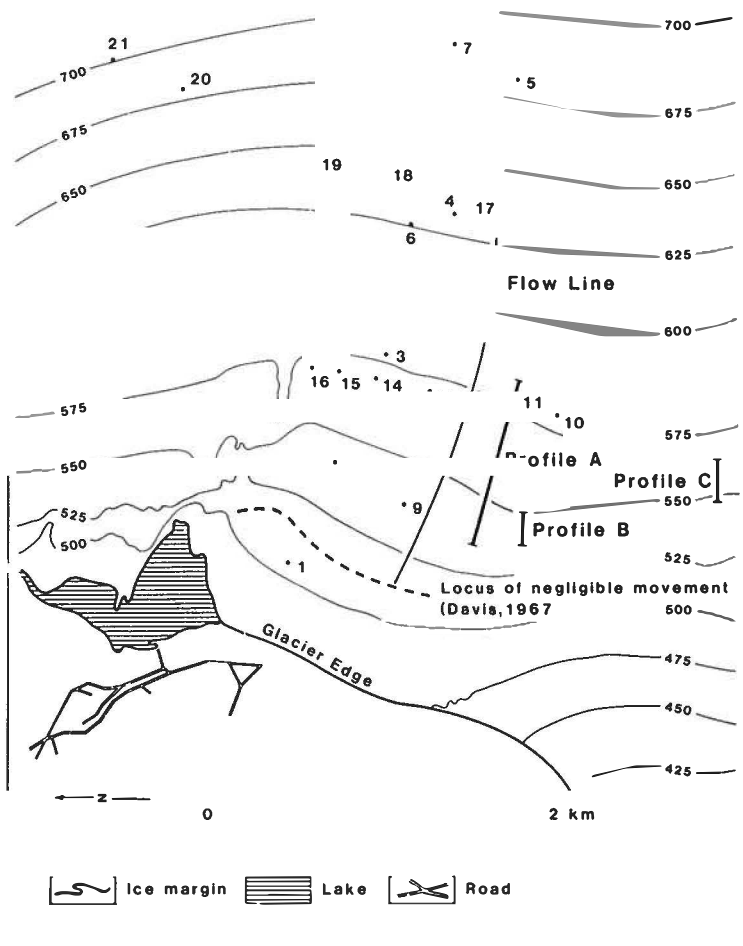


Fig. 29. Map of the Tuto Ramp. Surface elevations are in metres. Points marked by numbers are mass balance and velocity points. Locations of $\delta^{18}O$ sampling profiles are indicated. Modified from Schytt (1955).

Geographical $\delta^{18}O$-distribution

The results of the $\delta^{18}O$ analysis are summarised in Table 11. Disregarding the snow samples from profile A, which are not representative of the annual precipitation, it appears from the table that the range of mean $\delta^{18}O$ values is from −18.4 to −24.0 0/00. Moreover, if the $\delta^{18}O$ means from Nuna Ramp and the northern margin of Harald Moltke Bræ are left out, the range is even narrower, i.e. from −18.4 to −21.7 0/00. The ice with $\delta^{18}O$ means in this range can most reasonably be referred to an origin in the accumulation areas of the local ice caps around Wolstenholme Fjord. This interpretation is supported by the $\delta^{18}O$ mean values of −20.3 ± 0.8 0/00 and −20.5 ± 1.0 0/00 for the samples from the two snow pits on the "Tuto ice dome", which both fall within this range.

The mean $\delta^{18}O$ values from Nuna Ramp and the northern margin of Harald Moltke Bræ (−23.0 and −24.0 0/00) are lower than the rest of the mean $\delta^{18}O$ values by 2.5–3.5 0/00. However, these values are certainly not low enough to be interpreted as ice-age values (ice age $\delta^{18}O$ values from the deep Camp Century record are in the range from −38 to −42 0/00, i.e. 8.5–12.5 0/00 lower than the present-day $\delta^{18}O$ value at Camp Century). The ice, carrying $\delta^{18}O$ values around −23 to −24 0/00, is more likely to originate from the lower part of the accumulation area of the main Greenland ice sheet.

As regards Harald Moltke Bræ, the different origin of the ice at the southern and northern margins indicated by the different $\delta^{18}O$ values, is in accordance with the flow pattern inferred from the surface-elevation contours of the ice-sheet sector draining to Harald Moltke Bræ: The southern part is fed by ice originating at the local dome to the northeast of Store Landgletscher (see map in Fig. 2), whereas, the northern (and also the central) part is fed by ice from the main ice sheet.

Also in terms of variability (expressed as the root-mean-square (rms) value), there are characteristic differences between the various $\delta^{18}O$ records. The snow-pit

Table 11. Mean, standard deviation, and variability of $\delta^{18}O$ samples.

Location	Location No.	Mean $\delta^{18}O$ (0/00)	Standard deviation of mean (0/00)	Variability (r.m.s.) (0/00)	Number of samples
Tuto Ramp					
Surface ice (Prof. A)		−19.12	0.10	0.90	81
Snow (Prof. A)		−17.09			
Snow (Prof. A)		−34.70			
Snow (Prof. A)		−30.36			
Surface ice (Prof. B)		−18.41	0.15	0.90	36
Surface ice (Prof. C)		−19.42	0.21	1.07	26
Water from Lake Tuto		−20.34	0.13	0.18	2
Tuto ice dome	2				
Surface snow		−24.37	0.46	0.92	4
Snow pit		−20.52	1.01	3.19	10
Tuto ice dome	3				
Surface snow		−25.36	0.47	0.94	4
Snow pit		−20.34	0.81	2.56	10
Store Landgletscher	4				
Ice cliff		−18.77	0.05	0.37	55
Surface ice		−20.06	0.27	0.97	13
Piedmont glacier	5				
Surface ice		−20.85	0.18	0.31	3
Harald Moltke Bræ					
Southern margin	6				
Surface ice		−20.97	0.13	0.64	24
Northern margin	7				
Surface ice		−24.00	0.27	1.12	17
Salisbury Glacier	8				
Moraine covered ice		−21.65	0.11	0.27	6
Nuna Ramp	9				
Surface ice		−23.02	0.14	1.36	95

records display the highest variabilities of 3.2 and 2.6 0/00, respectively. Then follows in decreasing order the Nuna Ramp and the northern Harald Moltke Bræ records with rms values of 1.36 and 1.12, respectively. The rms values of the $\delta^{18}O$ records inferred to be of local origin are all around 0.9, except for the rms values from the ice-cliff record from Store Landgletscher, and the records from the piedmont glacier and Salisbury glacier which are around 0.3–0.4 0/00. Due to the few degrees of freedom the rms value from the piedmont glacier is encumbered with a large uncertainty, and therefore the low rms value of *c.* 0.3 0/00 is not significantly different (F-test) from the rms value of about 0.9 0/00 for the other local $\delta^{18}O$ records. The records from the ice cliff and Salisbury glacier, on the contrary, have rms values of 0.37 0/00 (54 degrees of freedom) and 0.27 0/00 (5 degrees of freedom) which are significantly lower than the rms value of the other local $\delta^{18}O$ records.

The differences of the rms values can be explained as follows: In the snow-pit records derived from the uppermost firn layers, a relatively large part of the annual $\delta^{18}O$ cycle is still preserved. As the layers sink into the ice cap under continuous thinning, the $\delta^{18}O$ variations are smoothed by diffusion (Johnsen 1977), and probably also due to melting-refreezing processes. When the ice resurfaces in the ablation zone at the ice-sheet margin, it therefore displays lower $\delta^{18}O$ variations than do the upper firn layers.

The tendency for the $\delta^{18}O$ records from Nuna Ramp and the northern margin of Harald Moltke Bræ to have slightly larger rms values than the records of local origin, could be due to the fact that the ice at these locations originates from higher elevated and consequently colder areas than does the local ice.

Consequently, the annual $\delta^{18}O$ cycles are better preserved in this ice, resulting in higher variability of the $\delta^{18}O$ records. This is in agreement with previous findings of ice-margin studies in central West Greenland (Reeh *et al.* 1987b).

The low variability of the $\delta^{18}O$ records from the ice cliff and Salisbury glacier support the hypothesis that the ice at these locations has undergone melting-refreezing, probably at the glacier bed, causing additional smoothing of the $\delta^{18}O$ variations.

Ice-dynamic model for Tuto Ramp

Further interpretation of the $\delta^{18}O$ records depends on the possibility of dating them. This is possible for the long $\delta^{18}O$ record from Tuto Ramp, since information is available about ice-cap geometry, mass balance, and dynamics. This information can be used to set up an ice-dynamic model for theoretical dating of the ice exposed at the surface of the ramp. The model applied is the flow-line model described by Reeh (1988), which has previously been applied to date the surface ice at an ice-margin location in central West Greenland (Reeh *et al.* 1987b).

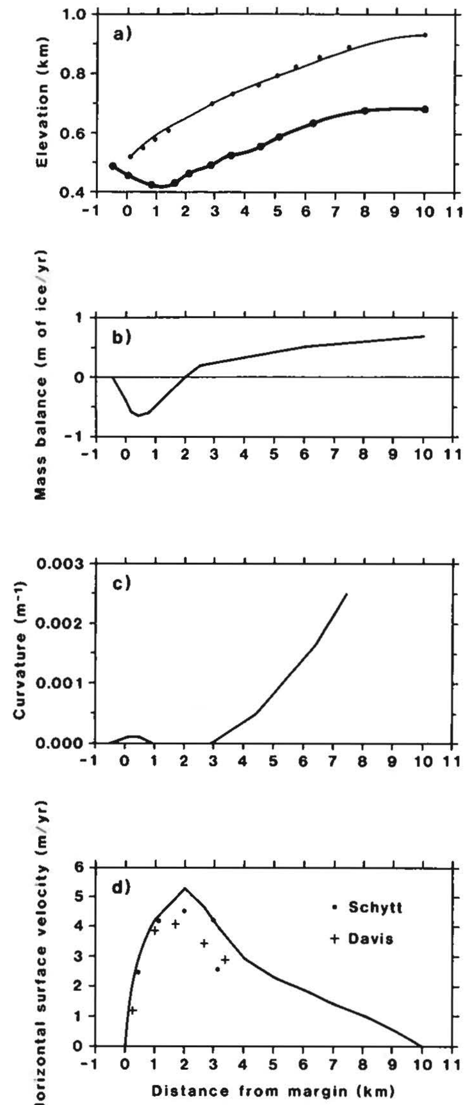


Fig. 30. Input data and results of ice-dynamic calculation for Tuto Ramp. a) shows observed base elevations (dots connected by thick curve), observed surface elevations (points marked by dots), and calculated surface elevation profile (thin curve). Observations based on data from Schytt (1955), Bishop (1957), Benson (1962), and Davis (1967). b) shows observed mass balance along flow line. Data from Schytt (1955) and Griffiths (1960). c) Curvature of elevation contours at their point of intersection with the flow line. Data from Schytt (1955) and Bishop (1957). d) Observed (points marked by dots and crosses) and calculated (curve) horizontal surface velocities along the flow line.

Ice cap geometry. – Surface-elevation maps of the Tuto Ramp area have been published by Schytt (1955) (see Fig. 29) and Davis (1967), and surface- and base-elevation maps of part of the "Tuto ice dome" are shown by Bishop (1957). Further surface-elevation data are given by Benson (1962) and in this work. Based on this information, the flow line shown in Figs 28 and 29, the curvature of the surface-elevation contours at their intersection with the flow line (Fig. 30c), and the observed surface- and base-elevation profiles indicated in Fig. 30a have been determined.

Mass balance. – The mass-balance distribution shown in Fig. 30b and the observed surface velocities shown in Fig. 30d are based on data from Schytt (1955), Griffiths (1960), and Davis (1967). Whereas the velocities measured in the different studies are very similar, the ablation rates measured by Schytt (1955) for the balance year 1953/54 are much higher than those observed by Griffiths (1960) for the balance year 1955/56. Griffiths' ablation data are nearly in balance with the velocity data, indicating that the Tuto Ramp was close to a balanced state in 1955/56. Schytt's ablation data, on the other hand, indicate that Tuto Ramp had a large negative balance in 1953/54. Since Griffiths' ablation-rate data are nearly in balance with the observed velocities, these data are more likely to represent the mass balance over a longer period, and, consequently, Griffiths' data are used in the modelling.

Ice temperatures. – Schytt (1955) also published observed temperatures at 10 m depth at one location in the ablation zone of the Tuto Ramp, and at two locations in the accumulation area further inland at elevations of 800 and 850 m respectively. Whereas the 10-metre temperature from the site in the ablation area is around −13° C, i.e. close to the mean annual air temperature (Ohmura 1987), the temperature at the sites in the accumulation area are *c.* −3° C and −5° C, respectively, i.e. up to 10° C higher than the mean annual air temperature. The reason for these high firn temperatures is the release of latent heat by refreezing of percolating melt water formed at the surface during the summer period (Schytt 1955).

With such high firn temperatures combined with an ice thickness, of *c.* 200 m and an accumulation rate of *c.* 0.5 m/yr, a simple temperature-profile calculation indicates that the ice-sheet base is at the melting point in the accumulation area and probably also in the upper part of the ablation area, thus creating conditions for basal melting. In the lower part of the ablation area the cooling of the near-surface layers will eventually penetrate to the base, and thus create conditions for refreezing of the basal melt water formed further upstream. Temperatures measured in the wall of an ice tunnel excavated into the near-bottom part at the margin of the Tuto Ramp glacier were in the range from −5.5 to −9.5° C (Rausch 1958), thus confirming that sub-freezing temperatures exist in the basal ice near the margin.

Results. – Measured and predicted surface profiles are shown in Fig. 30a. Discrepancies are small. The fit is obtained with a flow-law parameter of A (−5° C)= 4.6 10^{-16} kPa^{-3} s^{-1}, which is about half the minimum value recommended by Paterson & Budd (1982). In view of

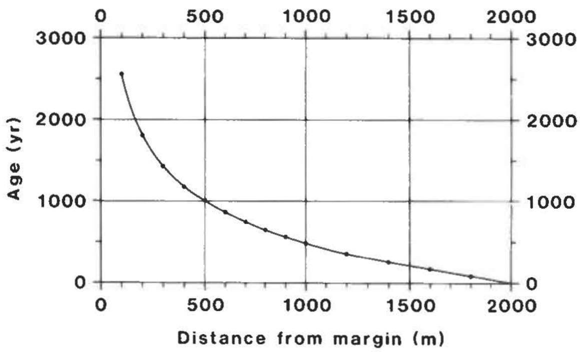


Fig. 31. Calculated age of surface ice along the near-margin part of the flow line on Tuto Ramp.

the uncertainty of the input data to the model, and the large variation in measured flow-law parameters, the disagreement between the predicted value and the recommended value is not discouraging.

Fig. 30d shows measured and predicted horizontal surface velocities. The variations along the flow line of measured and predicted velocities are similar, but the model overestimates the velocities between 2 and 3.5 km from the margin by about 25%. Unfortunately, observations of velocities are not available further inland than 3.5 km, but it is unlikely that the model predictions deviate more than 25%, from the actual velocities anywhere along the flow line. This uncertainty includes possible velocity fluctuations due to temporal mass-balance variations which are not accounted for by the steady-state ice-dynamic model. Therefore, the calculated ages of the surface ice displayed in Fig. 31 should be accurate to within ±25%, with higher ages being more likely than younger ages.

It appears from Fig. 31 that the age of the surface ice is around 2500 years at a distance from the margin of 100 m. (The ice margin is defined as the locus of negligible (horizontal) movement as observed by Davis (1967)). This indicates that ice of Weichselian origin (older than 11 000 years), if present at all at the surface, extends for at most some tens of metres inwards from the margin. Unfortunately, this part of the surface was covered by deep snow in the summer of 1986 which prevented sampling of ice.

The calculated ages can be compared with radiocarbon datings performed on carbon dioxide extracted from air inclusions in ice from a tunnel excavated in the near-bottom layers of the ice cap close to Tuto Ramp. Radiocarbon ages were *c.* 3000 and 5000 yr, respectively, for ice taken 300 and 200 m from the tunnel entrance (Oeschger *et al.* 1966). Taking into account that the age of the ice increases with increasing distance below the surface, these ages are compatible with the calculated ages of the surface ice.

Fig. 32 shows the $\delta^{18}O$ record from Tuto Ramp plotted against age BP. The time scale is the one given by the distance-age diagram of Fig. 31. The interval covered by

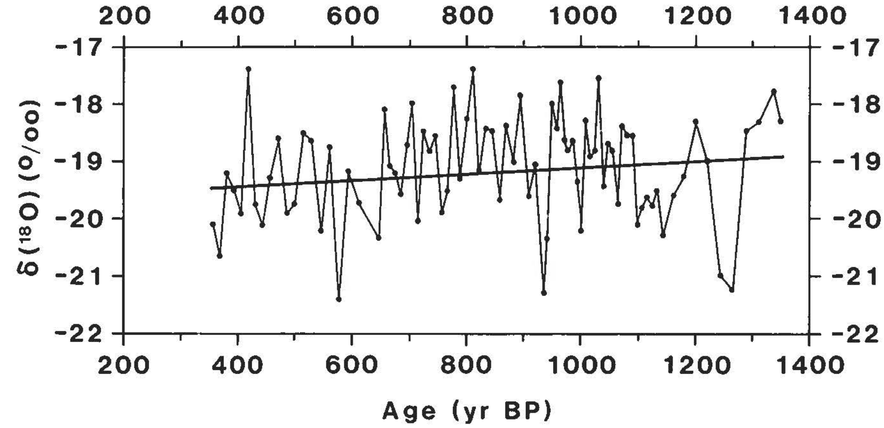


Fig. 32. $\delta^{18}0$ profile from the surface of Tuto Ramp. Age scale is based on the age-distance relationship of Fig. 31. Thick line indicates the linear trend.

the surface $\delta^{18}0$ record is *c.* 350–1350 years BP. The ages can be considered accurate to within ±25%. It might be worth noting that the positive trend in $\delta^{18}0$ with increasing age (0.5 o/oo in 1000 years), though not statistically significant, is compatible with a trend of 0.8 o/oo during the same period indicated by the $\delta^{18}0$ record from Camp Century (Robin 1983).

Age estimate for the Nuna Ramp $\delta^{18}0$ record

The data coverage upstream of Nuna Ramp is not good enough to justify application of advanced ice-dynamic models for dating the ice. Some simpler considerations will be applied.

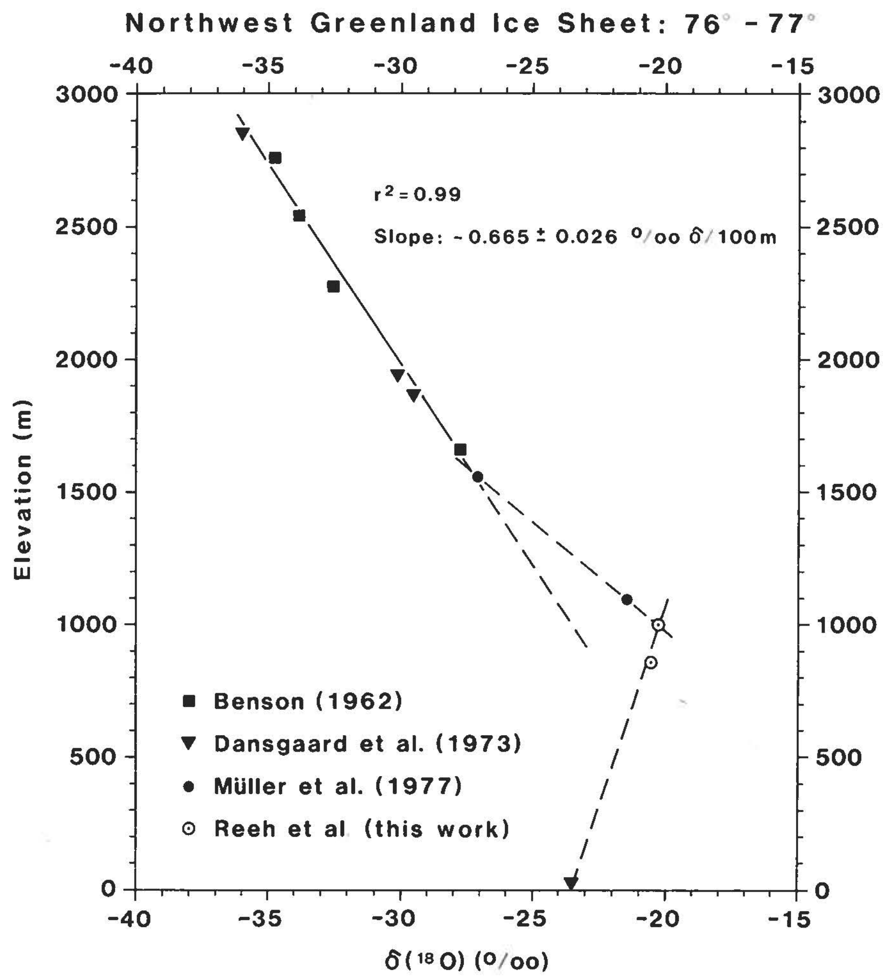


Fig. 33. δ^{18}O-elevation relationship for Northwest Greenland between 76 and 77°N.

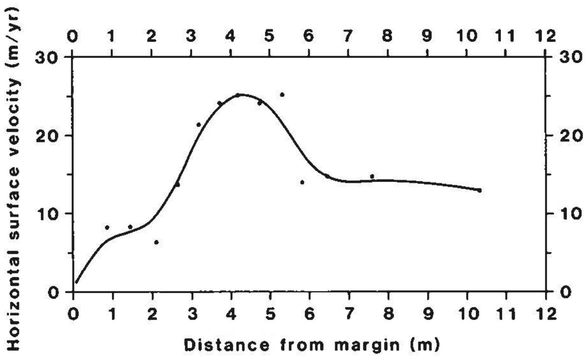


Fig. 34. Observed horizontal surface velocities along the Nuna Ramp flow line (points marked by dots). The curve represents the smoothed velocity distribution used in the estimate of the age of the surface ice.

By means of the $\delta^{18}0$ elevation diagram for the northwest Greenland sector between 76° and 77°N (Fig. 33), the elevation of the deposition site for the snow forming the ice of the sampling location on Nuna Ramp ($\delta^{18}0$ = -23.0 0/00) is estimated to be *c.* 1250 m. According to data published by Nobles (1960) this elevation is reached on the Greenland ice sheet at a distance of *c.* 25 km from the margin of Nuna Ramp. From the margin and 10 km inland, horizontal surface velocities have been measured by Nobles (1960) (Fig. 34), which can be used to estimate the time used by an ice particle to travel from 10 km upstream of the margin to the sampling profile located 300 to 1300 m from the margin. This time has been estimated as 800 ± 50 years. Between 10 and 25 km upstream from the margin, velocities are not known, but most likely the velocities decrease going inland. A reasonable estimate of the surface velocity 25 km inland is 5 m/year with estimated error limits of 2 and 13 m/year, the latter value being the surface velocity measured *c.* 10 km from the margin. Assuming a linear velocity variation along the 10 – 25 km section results in an estimated travel time along this section of 1800 yr with upper and lower limits of 2500 and 1150 yr respectively. Adding up, the age of the ice at the sampling location is estimated to be around 2500 yr with upper and lower limits of 3500 and 2000 yr, respectively. As shown in Fig. 35, the profile starts about 300 m from the ice margin, and therefore it can be concluded that ice of Weichselian origin if present at all at the surface of the Nuna Ramp, is limited to a narrow band near the margin, probably less than 100 m wide. Unfortunately, deep snow and slush prevented sampling of ice from this part of the surface in the summer of 1986.

The $\delta^{18}0$ record from Store Landgletscher

The $\delta^{18}0$ record from the ice cliff of Store Landgletscher is shown in Fig. 36. It is not possible to give even a rough estimate for the age of the ice in the cliff, if only for the reason that the cliff has probably been formed by freeze-on at the base of melt water produced by basal melting further inland. The layered structure of the ice cliff with numerous dirt layers and inclusions, and the low variability of the $\delta^{18}0$ record point to this conclusion. Furthermore, it was shown in a previous section that the thermal conditions required for melting/refreezing to take place are in fact present at the ice-cap base. Comparison of the $\delta^{18}0$ values from the ice-cliff samples with the mean $\delta^{18}0$ value of the samples collected on the near-horizontal glacier surface upstream of the cliff provides further support for the freeze-on hypothesis. The mean $\delta^{18}0$ value of the clean debris-free surface ice which exibits a sharp contact to the ice-cliff ice, is significantly lower than the ice-cliff $\delta^{18}0$ values. The explanation could be that the $\delta^{18}0$ values from the ice cliff are increased due to fractionation during the freeze-on process (Jouzel & Souchez 1982).

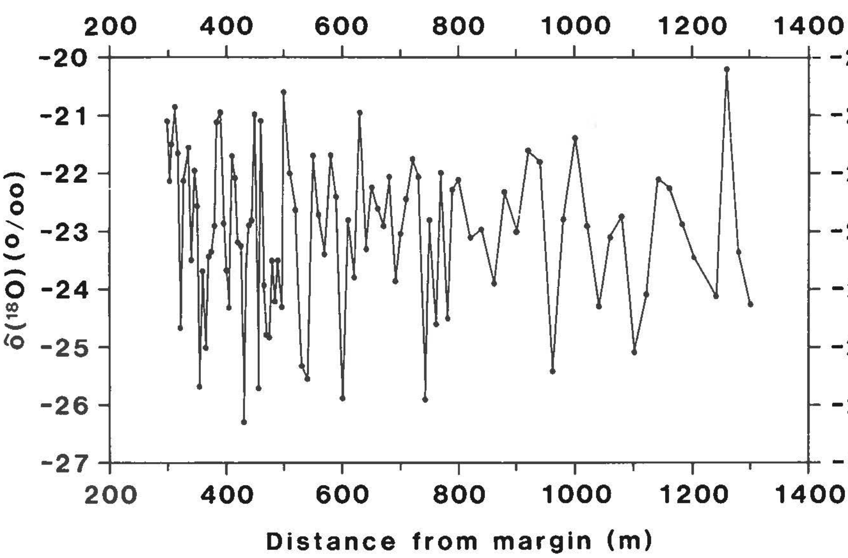


Fig. 35. $\delta^{18}0$ profile from the surface of Nuna Ramp.

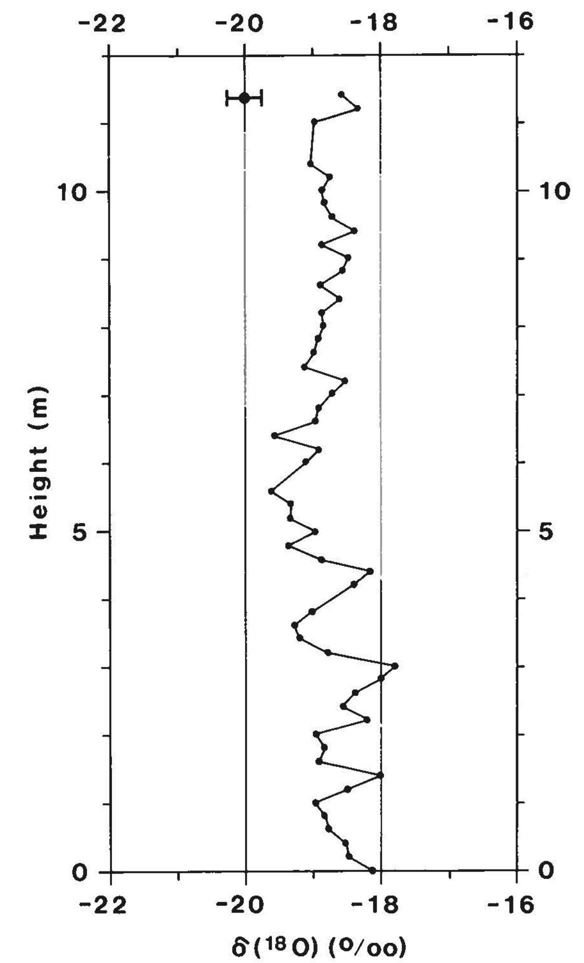


Fig. 36. $\delta^{18}0$ profile from the ice cliff of Store Landgletscher. The mean $\delta^{18}0$ value for the surface ice above the cliff with error limits (1 standard deviation) is also indicated.

In connection with the present discussion of the origin of the ice cliff, a comment is appropriate about the formation of the moraines on, and along, the ice-cap margin, even though these moraines were not studied during our expedition. Bishop (1957) and Swinzow (1962) argued that the moraines were formed by shear deformations in the ice-cap margins. This explanation, however, was opposed by Weertman (1961) and Hooke (1970), who considered a freeze-on process to be a more likely mechanism for incorporating debris in the basal ice layers. The present study can be taken as support for the Weertman-Hooke point of view, since it provides evidence that basal freeze-on of ice is likely to take place near the ice margins.

The $\delta^{18}0$-elevation relationship

A plot of mean $\delta^{18}0$ values from the upper firn layers on the ice sheet against elevation (E) (see Fig. 33) seems to distinguish between three different elevation ranges, each with its particular $\delta^{18}0$-elevation relationship. Above 1500 m the $\delta^{18}0$-elevation relationship is closely approximated by the regression equation (r = 0.99)

$$\delta^{18}0 = -31.4 - 0.00665 \, (E - 2180)$$

The $\delta^{18}0$-elevation gradient indicated by this equation (-0.665 ±0.026 o/oo per 100 m) is in good agreement with the value of -0.66 o/oo per 100 m found by Dansgaard *et al.* (1973) for locations above 1500 m on the ice sheet in central West Greenland.

Below an elevation of 1000 m the data indicate a small, but positive, $\delta^{18}0$-elevation gradient of *c.* 0.3 o/oo per 100 m. Similar small, positive gradients were found by Koerner (1979) for the same elevation range on high-arctic Canadian ice caps. Koerner (1979) explained the small gradient in terms of precipitation from a single condensation level. Reversed $\delta^{18}0$-elevation gradients were also found in Antarctica below elevations of *c.* 1000 m (Lorius & Merlivat 1977).

Between 1000 and 1500 m elevation, the data indicate a negative gradient about twice the value found above 1500 m. It is tempting to interpret this higher-than-normal gradient as being caused by fractionation in connection with melting/refreezing processes in the upper firn layers (Arnason 1969). The lower the location in the accumulation area, the larger is the fraction of melt water that will escape by runoff, and the more enriched in heavy isotopes will be the remaining refrozen melt.

Comparison of mean annual air temperatures and ten-metre firn temperatures from the Northwest Greenland ice sheet (to be presented elsewhere) supports this hypothesis: Above 1500 m there is good agreement between the two temperatures. Below 1500 m, however, the ten-metre firn temperature deviates more and more from the mean annual air temperature with decreasing elevation, indicating increasing warming due to refreezing of melt water formed at the surface in the summer period, in accordance with results previously presented by Hooke *et al.* (1983) and Ohmura (1987).

Discussion and conclusions

Even though the main purpose of the NORDQUA 86 glaciological programme was not achieved because ice of Weichselian origin was not found at the investigated ice margins in the Thule area, the study demonstrated the usefulness of applying stable isotope methods in glaciological studies of ice margins. The isotopic signature of the ice, in terms of mean value as well as variability, is characteristic for the site of formation of the ice and for its thermal and phase-change history since the time of formation.

The model calculations of the ice flow of the Tuto and Nuna ramps indicate that ice of Weichselian origin, if present at all, is limited to narrow bands near the margins, probably less than some tens of metres wide. However, basal melting, which is likely to occur upstream of Tuto Ramp and probably also upstream of Nuna Ramp, may, partly or fully, have removed the Weichselian ice. For Tuto Ramp the effect of basal melting on the age of the ice can be estimated by means of Fig. 32 in a paper by Reeh (1989) which shows calculated ages at the base of an ice sheet as a function of the ratio of basal melt rate to surface-accumulation rate. For "Tuto ice dome", the basal melt rate is probably only a fraction of the value (0.5 cm/yr) corresponding to the case of a temperate glacier subject to an average geothermal heat flux, since part of the heat flows to the upper surface, where the mean annual snow temperature is a few degrees below the freezing point. For a basal melt rate of 0.1 cm/yr, for example, the maximum age of the ice is calculated to be *c.* 14 000 yr, whereas the maximum possible basal melt rate of 0.5 cm/yr results in a maximum age as young as *c.* 5000 yr. These estimates indicate that the extent of Weichselian ice at the margin of the Tuto Ramp may be even less than indicated by the theoretical dating mentioned previously. Moreover, increased surface melting in the Holocene climatic optimum *c.* 8000–3000 yr B.P. could also have seriously affected the ice margins in the Thule area, and could in fact have threatened the very existence of the local ice caps in this period. Calculations by a climate/mass-balance model (Roger Braithwaite, personal communication) indicate that an estimated 2–3°C warmer summer climate could have resulted in a rise of the equilibrium line on the "Tuto ice dome" from the present elevation of 650 m to an elevation between 800 and 900 m. Moreover, the mass wastage by ablation is likely to have increased by about 1 m/yr in average over the ablation area, which, furthermore, would have been much larger than at present because of the rise in equilibrium-line elevation. The thin ice caps in the Thule area could hardly have survived such conditions for a period of several thousand years.

The effect of a likely increase in snow accumulation is not accounted for in the above estimates. However, a study by Paterson & Waddington (1984) indicates no more than a 10-20 % increase in accumulation rate in the Camp Century area in the period 5000-3000 yr B.P. A similar small increase in snow accumulation on the local ice caps in the Thule area would probably not be sufficient to ensure the survival of these ice caps in a climate with summer temperatures 2–3°C warmer than present.

If the local ice caps in the Thule area suffered such severe losses or maybe even became extinct in the Holocene climatic optimum, why did Barnes ice cap, located *c.* 750 km further to the south on Baffin Island, Canada (69.5°N, 72°W) not suffer the same fate? That Barnes ice cap survived the Holocene climatic optimum is proven beyond doubt by the detection of ice of Wisconsinan origin in deep bore holes as well as on the margins of the ice cap (Hooke & Clausen 1982). The present dimensions and surface elevations of Barnes ice cap are similar to those of "Tuto ice dome", and also the mass balance-elevation relationships are similar for the two ice caps, although with somewhat lower accumulation rates for Barnes ice cap. (The location of Barnes ice cap far to the south of "Tuto ice dome" is nearly compensated, as far as temperatures are concerned, by the much colder climate at the same latitude on the Canadian side than on the Greenland side of Baffin Bay (Wilson 1969)).

Therefore, there is nothing in the present conditions of the two ice caps that can explain why "Tuto ice dome" might have melted away in the Holocene climatic optimum, whereas Barnes ice cap survived. However, the explanation may be found in the very different conditions at the two ice-cap locations at the end of the Weichselian glacial. Whereas the late glacial ice cover in the Thule area was only slightly more extensive than the present (Fig. 10), Baffin Island was in the Early Holocene covered by the thick ice masses of the Laurentide ice sheet. According to Prest (1969), Barnes ice cap still had an extent of 4–5 times its present size even as late as 6000 B.P. This means that whereas "Tuto ice dome" was probably a small ice cap at the time of the onset of the Holocene climatic optimum (8000–9000 B.P.), and therefore from the very beginning was likely to suffer severely due to the warm climate, Barnes ice cap was still part of a large ice mass at that time. Therefore, even though the melting was intense also in the Baffin Island area, there was so much more ice to melt, and Barnes ice cap could survive as a small relic of the Laurentide ice sheet.

Also the margins of the Greenland ice sheet in the Thule area seem to have been affected by the warmer climate in the Holocene climatic optimum, as documented by findings of mollusc shells, radiocarbon dated to between 2650 and 7050 yr BP, in the marginal moraines along Harald Moltke Bræ, Chamberlain Gletscher, and Pitugfik Gletscher (see map in Fig. 28), thus indicating a significant retreat of the ice margins behind their present positions sometime during the Holocene (Kelly 1980b)

Appendix

C-14 dating of samples collected during the NORDQUA 86 expedition, and notes on the marine reservoir effect

compiled by Nils-Axel Mörner and Svend Funder

Radiocarbon age determinations of samples of bivalve shells (23) and organic detritus (4), collected by members of the NORDQUA expedition, are summarised below.

The ages were determined by conventional dating at Laboratoriet för isotopgeologi at Naturhistoriska Riksmuseet, Stockholm (samples marked St-), and at the C-14 Dating Laboratory of the Geological Survey of Denmark and the National Museum, Copenhagen (samples marked K-, by courtesy of the Geological Survey of Denmark). One sample has been determined by accelerator mass spectrometry (AMS) at the Physics Institute, Århus University (sample marked AAR, by courtesy of Mette Skovhus Thomsen). The conventional dates were corrected for isotopic fractionation according to measured values for C-13, while the AMS date was corrected for the standard isotopic fractionation in marine carbonates (0 0/00 PDB).

The results are reported according to the recommendations of Stuiver & Polach (1977), and include conventional and reservoir corrected ages (Rcorr). However, while the conventional ages from the laboratories in Stockholm and Århus are normalised to the standard activity in wood (-25 0/00 PDB C-13), those from Copenhagen are normalised to 0 0/00 PDB, and have a built-in 400 yr marine reservoir correction (see notes on determination of reservoir effect below).

While the laboratory in Stockholm calculates infinite ages as sample activity $+1\ \sigma$, the laboratory in Copenhagen uses sample activity $+2\ \sigma$, and therefore gives lower minimum ages.

Marine reservoir effect in the Thule area (Rcorr)

Three samples of contemporary shells from northern West Greenland have been C-14 dated by H.Tauber. The samples were supplied by the Zoological Museum in Copenhagen through the kind help of G. Høpner Petersen. The results appear at the end of the dating list, and the activity is expressed as per cent of modern, i.e. 0.95 of the activity of the oxalic acid standard, and corrected for isotopic fractionation and decay from the time of collection to 1950.

The average activity of the three samples is 99.95 ± 0.6 % of modern, i.e. they show a C-14 deficiency corresponding to 5 ± 50 yrs. This is in agreement with previous results from Thule, thus a sample of contemporary shells from the area gave an activity of less than ±1 % of modern wood (Suess 1954), and *Mytilus* shells, collected in 1940, gave an age of 50 ± 60 yrs (GSC-2316; Blake 1987). There is also agreement with results from further south in West Greenland (Krog & Tauber 1973). However, the reservoir effect is somewhat smaller than that measured at nearby Ellesmere Island (Mangerud & Gulliksen 1975, Blake 1987, 1988), and Northeast Greenland (Funder 1982, Tauber & Funder 1975), where the higher reservoir effect has been thought to be caused by the low C-14 activity in polar water, cut off from exchange with the atmosphere (Tauber & Funder 1975).

Thus the reservoir effect appears to be determined by water mass regime, and we suggest that Holocene C-14 dates from the Thule area, like those from other parts of West Greenland, are corrected by subtracting 400 yrs from conventional ages normalised to -25 0/00, while ages normalised to 0 0/00 should not be further corrected.

Reservoir correction is applied only to ages younger than 15 000 yrs.

The C-14 dates are listed according to their field number (GC), the last three digits are the same as used as sample numbers elsewhere in this volume.

Samples collected by Svend Funder, Michael Houmark-Nielsen, Christian Kronborg, Ove Klakkegg, Robert Lagerbäck, Arve Misund, Lars Rohde, Lars Erik Skylvik, Oddmund Soldal, Lennart Sorby and Morten Thoresen

GC68-001:K-4780. Saunders Ø.
8200 ± 85 B.P. ^{13}C=-1.1 0/00
Rcorr. 8200 B.P.

Articulated shells of *Mya truncata* and *Hiatella arctica* in silt in section 21 m above sea-level. Section B, unit S6. Same as sample 010, and amino acid analysed and TL-dated (BAL-1235, R-861006; Tables 9 and 10). Narssarssuaq, 76°36′N, 69°42′W.

GC68-013:AAR-1. Saunders Ø.
9150±200 B.P.
Rcorr. 8750 B.P.
A fragment of *Mya truncata* from beach gravel from top of upper marine terrace 40 m above sea-level. Minimum age for Holocene marine limit. Amino acid analyses suggest that older shells are also present (BAL-1300; Table 10). Previous C-14 dates from this locality have been reported by Suess (1954), and Blake (1987). Narssarssuaq, 76°36'N, 69°42'W.

Sample collected by Jan Lundquist, Jan Mangerud and Joar Sættem

GC68–021:K-4781. Wolstenholme Fjord.
9150±95 B.P. ^{13}C= 1.7 0/00
Rcorr. 9150 B.P.

Fragments of *Hiatella arctica* and *Mya truncata* from silt 1 m above sea-level at base of 6 m high coastal section. Minimum age for local marine limit at 35 m. Fauna contains also *Mytilus edulis* and *Chlamys islandica*. Near Salisbury Gletscher, 76°40'N, 68°38'W.

Samples collected by Nils-Axel Mörner Coastal cliffs at Narssârssuk, 76°27'N, 69°35'W

GC68–079:St-10721. Section E, unit N4.
$41215 \begin{smallmatrix}+1480\\-1250\end{smallmatrix}$ B.P. ^{13}C= 2.4 0/00
Shells of *Hiatella arctica* from upper part of upper silt at 11.1 m above sea-level.

GC68-080:St-10722. Section E, unit N4.
>48000 B.P ^{13}C=2.2 0/00
Articulated shells of *Hiatella arctica* in living position at silt/sand boundary 13.05 m above sea-level.

GC68–081:St-10723. Section E, unit N5.
>48000 B.P. ^{13}C= 2.3 0/00
Articulated shells of *Hiatella arctica* in living position in sand 13.4 m above sea-level.

GC68–082:St-10724. Section E, unit N5.
>48000 B.P. ^{13}C= 2.4 0/00
Articulated shells of *Hiatella arctica* and *Mya truncata* in living position in sand 13.67 m above sea-level.

GC68–083:St-10725. Section E, unit N5.
>45000 B.P. ^{13}C= 2.3 0/00
Articulated shells of *Hiatella arctica* and *Mya truncata* in living position in sand 13.8 m above sea-level.

GC68–084:St-10726. Section E, unit N5.
>47000 B.P. ^{13}C=2.1 0/00
Articulated shells of *Mya truncata* in living position in sand at 13.7 14.0 above sea-level.

GC68–085:St-10727. Section E, unit N5.
$45160 \begin{smallmatrix}+4500\\-3300\end{smallmatrix}$ B.P. ^{13}C=2.5 0/00
Articulated shells of *Mya truncata* in sand at 14.1 m above sea-level.

GC68–086:St-10730. Section G, unit N5.
$44390 \begin{smallmatrix}+4500\\-4200\end{smallmatrix}$ B.P. ^{13}C=2.5 0/00
Shells in gravel 7.1–7.2 m above sea-level.

GC68–087:St-10731. Section G, unit N5.
$40680 \begin{smallmatrix}+1450\\-1200\end{smallmatrix}$ B.P. ^{13}C=2.4 0/00
Shells just above gravel at 7.5 m above sea-level.

GC68–088:St-10734. Section H, unit N6.
9295±100 B.P. ^{13}C=1.9 0/00
Rcorr. 8895 B.P.
Shells of *Mya truncata* and *Balanus crenatus* in sand at 7.1–7.2 m above sea-level.

GC68–089:St-10732. Section "NAM-2".
$42940 \begin{smallmatrix}+2870\\-2100\end{smallmatrix}$ B.P. ^{13}C=-22.2 0/00
Organic detritus (lower layer) in sand, at 8.3 m above sea level.

GC68–090:St-10733. Section "NAM-2".
>45000 B.P. ^{13}C=-23.6 0/00
Organic detritus (upper layer) in sand 8.4 m above sea-level.

GC68–091:St-10728. Section "NAM-3".
>33000 B.P. ^{13}C=2.3 0/00
Organic detritus and shell fragments in lower beach gravel 3.9 m above sea-level.

GC68–092:St-10729. Section "NAM-3".
$35\,105 \begin{array}{l} +\ 510 \\ -1270 \end{array}$ B.P. ^{13}C=2.3 0/00
Shells above beach gravel at 7.5 m above sea-level.

GC68–093:St-10720. Section "NAM-6".
>45 000 B.P. ^{13}C=2.2 0/00
Shells from upper part of upper silt at 12.15 m above sea-level.

Coastal cliff at Ivnaq tugdleq, SW of Thule Air Base 76°32'N 68°55'W

GC68–094:St-10739.
9405±85 B.P. ^{13}C=3.4 0/00
Rcorr. 9005 B.P.
Shells of *Mya truncata* in upper part of upper silt at 4.9 m above sea-level.

GC68–095:St-10736.
9365±85 B.P. ^{13}C=3.2 0/00
Rcorr. 8965 B.P.
Shells of *Mya truncata* in upper silt at 5.25–5.35 m above sea-level.

GC68–096:St-10737.
8760±180 B.P. ^{13}C=2.4 0/00
Rcorr. 8360 B.P.
Shells of *Mya truncata* in Holocene sand at 7.4 m above sea-level.

GC68–097:St-10738.
8570±95 B.P. ^{13}C=2.8 0/00
Rcorr. 8170 B.P.
Shells of *Mya truncata* in upper part of Holocene sand at 8.4 m above sea-level.

GC68–098:St-10735. Store Landgletscher
>48 000 B.P. ^{13}C=-25.4 0/00
Wood (piece of branch) lying on surface of end moraine *c.* 400 m above sea-level. 76°30'N, 68°35'W.

Samples collected by Roar Austad, Michael Kelly, Lars König Königsson, and Kaj Strand Petersen

GC68–117:K-4782. Narssârssuk.
>40 500 B.P. ^{13}C=1.4 0/00
Articulated shells of *Mya truncata* and *Hiatella arctica*. Section E, unit N4. 76°27'N, 69°20'W.

GC68–130:K-4783. Narssârssuk.
>37 800 B.P.
Articulated shells of *Mya truncata* and *Hiatella arctica* from coastal section G, unit N5. 76°27'N, 69°20'W.

Sample collected by Michael Houmark-Nielsen and Kaj Strand Petersen

GC68–161:K-4784. Inner Wolstenholme Fjord.
9040±95 B.P. ^{13}C=1.1 0/00
Rcorr. 9040 B.P.
Fragments of *Mya truncata* and *Hiatella arctica* in coastal section immediately in front of Harald Moltke Bræ at Nunatarssûp nua, 76°39'N, 68°00'W. Previous C-14 dates from same locality reported by Crane & Griffin (1959) and Goldthwait (1960)

Sample collected by a group of Norwegians and Robert Lagerbäck

GC68–172:K-4785. Harald Moltke Bræ.
6980±85 B.P. ^{13}C=1.0 0/00
Rcorr. 6980 B.P.
Fragments of *Mya truncata* and *Hiatella arctica* found on the glacier surface, and transported from glacier bed along shear planes. 76°39'N, 68°00'W.

C-14 dating of contemporary shells

GGU-223397:K-381. Prøven harbour.
100.12±0.55 % ^{13}C=-0.1 0/00
Shells from nine individuals of *Mytilus edulis*, collected by F. Johansen on July 10th, 1936. Preserved in alcohol at The Zoological Museum, Copenhagen. 72°22'N, 55°44'W.

GGU-223399:K-382. Upernavik.
99.87±0.55 % ^{13}C=-0.2 0/00
Shells from eight individuals of *Mytilus edulis*, collected on July 2nd, 1936. Preserved in alcohol at The Zoological Museum, Copenhagen. 72°47'N, 56°10'W.

GGU-223400:K-383. Thule.
99.86±0.70% ^{13}C=-0.3 0/00
Shells from eight individuals of *Mytilus edulis*, collected on Sept. 2nd 1940. Preserved in alcohol at The Zoological Museum, Copenhagen. 76°34'N, 68°48'W.

References

Aarkrog, A., Boelskifte, S., Dahlgaard, S., Duniec, S., Hallstadius, L., Holm, E. & Smith, J.N. 1987. Technetium-99 and cesium-134 as long distance tracers in Arctic waters. Est. cst. Shlf. Sci. 24: 637–647.

Abelson, P. H. 1955. Organic constituents of fossils. – Yearb. Carnegie Instn Wash. 54: 107–109.

Aksu, A. E. 1985. Climatic and oceanographic changes over the past 400,000 years. Evidence from deep-sea cores in Baffin Bay and Davis Strait. – In: Andrews, J. T. (ed.), Quaternary environments, Eastern Canadian Arctic, Baffin Bay and Western Greenland. – Allen & Unwin, Boston: 181–209.

Andrews, J. T., Miller, G. H., Nelson, A. R., Mode, W. N. & Locke, W. W. III 1981. Quaternary near-shore environments on eastern Baffin Island, N. W. T. – In: Mahaney, W. C. (ed.), Quaternary paleoclimate. – GeoBooks, Norwich: 13–44.

Andrews, J. T. & Miller, G. H. 1985. Discussion on the paper by Denton and Hughes (1983). – Quaternary Res. 24: 361–364.

Andrews, J. T., Aksu, A., Kelly, M., Klassen, R., Miller, G. H., Mode, W. N., & Mudie, P. 1986. Land/ocean correlations during the last interglacial/glacial transition, Baffin Bay, northwestern North Atlantic: a review. – Quaternary Sci. Rev. 4: 333–355.

Arnason, B. 1969. The exchange of hydrogen isotopes between ice and water in temperate glaciers. – Earth. planet. Sci. Lett. 6: 423–430.

Barnes, H. 1957. The northern limits of *Balanus balanoides* (L.). – Oikos 8: 1–15.

Benson, C. S. 1962. Stratigraphic studies in the snow and firn of the Greenland ice sheet. – Res. Rep. Cold Reg. Res. Engng Lab. 70: 93 pp.

Bishop, B. C. 1957. Shear moraines in the Thule area, Northwest Greenland. – Res. Rep. Cold Reg. Res. Engng Lab. 17: 17 pp.

Blake, W. 1973. Former occurrence of Mytilus edulis L. on Coburg Island, Arctic Archipelago. – Naturaliste can. 100: 51–58.

Blake, W. 1975. Glacial geological investigations in northwestern Greenland. – Geol. Surv. Can. Paper 75-1A: 435–439.

Blake, W. 1977. Radiocarbon age determinations from the Carey Islands, Northwest Greenland. – Geol. Surv. Can. Paper 77-1A: 445–454.

Blake, W. 1980. Application of amino acid ratios to studies of Quaternary geology in the high arctic. – In: Hare, P. E., Hoering, T. C. & King, K. 1980 (eds), Biochemistry of amino acids. – Wiley & Sons, New York: 453–461.

Blake, W. 1982. Terrestrial interstadial deposits, Ellesmere Island, N. W. T., Canada. – American Quaternary Association, 7th Biennial meeting, Seattle. Abstracts: 73.

Blake, W. 1987. Geological Survey of Canada radiocarbon dates XXVI. – Geol. Surv. Can. Paper 86–7: 60 pp.

Blake, W. 1988. Geological Survey of Canada radiocarbon dates XXVII. – Geol. Surv. Can., Paper 87–7: 100 pp.

Bowen, O. Q., Richmond, G. M. Fullerton, D. S., Sibrava, V., Fulton, R. J. & Velichko, A. A. 1986. Correlation of Quaternary Glaciations in the Northern Hemisphere.– In: Sibrava, V., Bowen, D. Q., & Richmond, G. M. (eds). Quaternary Glaciations in the Northern Hemisphere. Quaternary Sci. Rev. 5: 509–510.

Brigham, J. 1983a. Stratigraphy, amino acid geochronology, and correlation of Quaternary sea-level and glacial events, Broughton Island, arctic Canada. – Can. J. Earth Sci. 20: 577–598.

Brigham, J. 1983b. Intrashell variations in amino acid concentrations and isoleucine epimerization ratios in fossil *Hiatella arctica*. – Geology 11: 509–513.

Broch, H. 1924. Cirrpedia thoracia von Norwegen und dem Norvegischen Nordmeere. – Skr. Vidensk. Selsk. Christiania 17: 121 pp.

Böcher, J. 1989. First record of an interglacial insect from Greenland: *Amara alpina* (Paycull, 1790) (Coleoptera: Carabidae). – Boreas 18: 1–4.

Bøtter-Jensen, L. & Mejdahl, V. 1985. Determination of potassium in feldspars by beta counting using a GM multicounter system. – Nuclear Tracks 10: 663–666.

Chamberlin, T. C. 1895. Geology. Typography of Greenland. – In: Bryant, H. G. (ed.), The Peary Auxiliary Expedition of 1894. Bull. geogr. Cl. Philad. 1: 167–194.

Clausen, H. B. & Stauffer B. 1988. Analyses of two ice cores drilled at the ice-sheet margin in West Greenland. – Annls Glaciol. 10, 23–27.

Crane, H. R. & Griffin, J. B. 1959. University of Michigan radiocarbon dates IV. – Radiocarbon 1: 173–198.

Cushman, J. A. 1948. Arctic Foraminifera. – Cushman Lab. Foram. Res., Spec. Publ. vol. 23: 1–80.

Dale, J. E. 1985: Recent intertidal molluscs from the east-central coast of Ellesmere Island, Northwest Territories. – Geol. Surv. Can. Paper 85-1B: 319–324.

Dansgaard, W., Johnsen, S. J., Clausen H. B. & Gundestrup, N. 1973. Stable isotope glaciology. – Meddr Grønland 197 (2): 53 pp.

Dansgaard, W., Johnsen, S. J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N. & Hammer, C. U. 1984. North Atlantic climatic oscillations revealed by deep Greenland ice cores. – Geophys. Monogr. 29: 288–298.

Dawes, P. R., Frisch, T. & Christie, R. L. 1982. The Proterozoic Thule basin of Greenland and Ellesmere Island, importance to the Nares Strait debate. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 89–104.

Davis, R. M. 1967. Ice surface movements on the Tuto ramp in North Greenland. – Tech. Rep. Cold Reg. Res. Engng Lab. 164: 24 pp.

De Vernal, A., Hillaire-Marcel, C., Aksu, A. E. & Mudie, P. J. 1987. Palynostratigraphy and chronostratigraphy of Baffin Bay deep sea cores: climatostratigraphic implications. – Palaeogeogr. Palaeoclimat. Palaeoecol. 62: 96–106.

De Vernal, A. & Hillaire-Marcel, C. 1988. Palynological and isotopic records of the Late Pleistocene in eastern Canada and adjacent basins. – VI Congrès de lÁQQUA. Abstracts. Université du Québec à Rimouski: Rimouski: 5 pp.

Dunbar M. J. 1968. Ecological development in Polar regions, a study in evolution. – Prentice-Hall, New Jersey: 119 pp.

Dyke, A. S., Andrews, J. T. & Miller, G. H. 1982. Quaternary geology of Cumberland Peninsula, District of Franklin. Geol. Surv. Can. Memoir 403: 32 pp.

Ellis, D. V. 1955. Some observations on the shore fauna of Baffin Island. – Arctic 8: 224–236.

Ellis, D. V. 1960. Marine infaunal benthos in Arctic North America. – Arctic Institute North America, Technical Papers, 5: 53 pp.

Elverhøi, A., Liestøl, O., & Nagy, J. 1980. Glacial erosion, sedimentation and microfauna in the inner part of Kongsfjorden, Spitsbergen. – Skr. norsk Polarinst. 172: 33–60.

Feyling-Hanssen, R.W. 1953. The barnacle *Balanus balanoides* (Linné, 1766) in Spitsbergen. – Skr. norsk Polarinst. 98: 64 pp.

Feyling-Hanssen, R. W. 1971. Weichselian interstadial foraminifera from the Sandnes-Jæren area.– In: Feyling-Hanssen, R. W., Jørgensen, J. A., Knudsen, K. L., & Andersen, A.-L. L. (eds), Late Quaternary foraminifera from Vendsyssel, Denmark, and Sandnes, Norway. – Bull. geol. Soc. Denmark 21: 72–116

Feyling-Hanssen, R. W. 1976. The Clyde Foreland Formation, a micropaleontological study of Quaternary stratigraphy. – 1st International Symposium on Benthonic Foraminifera of Continental Margins, part B: Paleoecology and Biostratigraphy. – Marit. Sediments, Spec. Pub. 1: 315–377.
Feyling-Hanssen, R. W. 1980. Microbiostratigraphy of young Cenozoic marine deposits of the Qivituq Peninsula, Baffin Island. – Mar. Micropaleont. 5: 153–183.
Feyling-Hanssen, R. W. 1981. Foraminiferal indication of Eemian interglacial in the northern North Sea. – Bull. geol. Soc. Denmark 29: 175–189.
Feyling-Hanssen, R. W. 1985. Late Cenozoic marine deposits East Baffin Island and East Greenland, microbiostratigraphy – correlation – age. – In: Andrews, J. T. (ed.), Quaternary Environments: Eastern Canadian Arctic, Baffin Bay, and Western Greenland. – Allen & Unwin, Boston: 354–393.
Feyling-Hanssen 1988. Foraminiferal assemblages in Quaternary sediments from the Thule area. – 18. Nordiske Geologiske Vintermøde, Abstracts. Danmarks Geologiske Undersøgelse, Copenhagen: 113–115.
Feyling-Hanssen, R. W. & Ulleberg, K. 1984. A Tertiary Quaternary section at Sarsbukta, Spitsbergen, Svalbard, and its foraminifera. – Polar Res. 2: 77–106.
Fisher, D. A. & Alt B. 1985: A global oxygen isotope model, semi-empirical, zonally averaged. – Annls Glaciol. 7: 117–124.
Fredskild, B. 1985. The Holocene vegetational development of Tugtuligssuaq and Qeqertat, Northwest Greenland. – Meddr Grønland, Geosci. 14: 20 pp.
Funder, S. 1982. ^{14}C-dating of samples collected during the 1979 expedition to North Greenland. – Rapp, Grønlands geol. Unders. 110: 9–14.
Funder, S. 1984. Chronology of the Last Interglacial/Glacial cycle in Greenland: First Approximation. – In: Mahaney, W. C. (ed.), Correlation of Quaternary Chronologies. Geo-Books, Norwich: 261–279.
Funder, S. 1988. Interglacial marine deposits in northern Greenland: NORDQUA 86 results. – 18. Nordiske Geologiske Vintermøde, Abstracts; Danmarks Geologiske Undersøgelse, Copenhagen: 124–125
Funder, S. (co-ordinator) 1989. Quaternary geology of the ice free areas and adjacent shelves of Greenland; chapter 13. In: Fulton, R. J. (ed), Quaternary geology of Canada and Greenland. – Geological Survey of Canada, Geology of Canada 1 (*also* Geological Society of America, The Geology of North America K-1), 741–792.
Funder, S. & Simonarson, L. A. 1984. – Bio- and aminostratigraphy of some Quaternary marine deposits in West Greenland. – Can. J. Earth Sci. 21: 843–852.
Goldthwait, R. P. 1960. Study of an ice cliff in Nunatarssuaq, Greenland. – Tech. Rep. Cold Reg. Res. Engng Lab. 39: 108 pp.
Griffiths, T. M. 1960. Glaciological investigations in the Tuto area of Greenland. – Tech. Rep. Cold Reg. Res. Engng Lab. 47: 63 pp.
Gudina, V. I. 1966. [Foraminifera and stratigraphy of the Northwest Siberian Quaternary]. – Akad. Nauk SSSR, Siberian Dept. Inst. Geol. and Geoph. U.D.K. 563.12 (119) (571.1): 132 pp. (In Russian).
Gudina, V. I. 1969. [The marine Pleistocene of Siberian lowlands. Foraminifera of the north part Jenisei's lowland]. – Akad. Nauk SSSR. Siberian dept., Inst. Geol. and Geoph. 63: 80 pp. (In Russian).
Gudina, V. I. & Evserov, U. Y. 1973. [Stratigraphy and foraminifera of Late Pleistocene of Kola Peninsula]. – Akad. Nauk SSSR Siberian dept., Inst. Geol. Geoph, 175: 148 pp. (In Russian).
Heron-Allen, E. & Earland, A. 1930. The foraminifera of the Plymouth district. – I and II. Jour. Roy. Micr. Soc. London, ser. 3, 50: 161–199.
Hillaire-Marcel, C., de Vernal, A. & Aksu, A.E. in press. High resolution palynostratigraphy and ^{18}O-stratigraphy of Upper Pleistocene sediments at ODP site 645, Baffin Bay. – Initial Reports of The Ocean Drilling Project, 105 part B.
Hooke, R. LeB. 1970. Morphology of the ice-sheet margin near Thule, Greenland. – J. Glaciol. 9: 303–324.
Hooke, R. LeB., Gould, J.E. & Brzozowski, J. 1983. Near-surface temperatures near and below the equilibrium line on Polar and Subpolar glaciers. – Z. Gletscherkd. Glazialgeol. 19: 1–25.
Hooke, R. LeB. & Clausen, H.B. 1982. Wisconsin and Holocene $\delta^{18}O$ variations, Barnes Ice Cap, Canada. – Bull. geol. Soc. Am. 93: 784–789.
Houmark-Nielsen, M. 1988. Quaternary sequences and event-stratigraphy of the Thule area, NW-Greenland. – 18. Nordiske Geologiske Vintermøde, Abstracts. Danmarks Geologiske Undersøgelse, Copenhagen: 124–125
Huntley, D. J. & Wintle, A. G. (1981) The use of alpha scintillation counting for measuring Th-230 and Pa-231 contents of ocean sediments. – Can. J. Earth Sci., 18: 419–432.
Ives, J. D., Andrews, J. T. & Barry, R. G. 1975. Growth and decay of the Laurentide ice sheet and comparisons with Fenno-Scandinavia. – Naturw. 62: 118–125.
Jacobs, J. D., Andrews, J. T. & Funder, S. 1985. Environmental background. – In: Andrews, J. T. (ed.), Quaternary environments, eastern Canadian Arctic, Baffin Bay and western Greenland. Allen & Unwin, Boston: 26–69.
Jensen, A. S. 1939. Concerning a change of climate during recent decades in the Arctic and Subarctic regions, from Greenland in the west to Eurasia in the east, and contemporary biological and geophysical changes. – Biol. Meddr 14, no. 8: 75 pp.
Johnsen, S. J. 1977. Stable isotope homogenization of polar firn and ice. – Publs Ass. int. Hydrol. scient. 118: 210–219.
Jouzel, J. & Souchez R. A. 1982. Melting-refreezing at the glacier sole and the isotopic composition of the ice. – J. Glaciol. 28, 35–42.
Just, J. 1980. Amphipoda (Crustacea) of the Thule area, Northwest Greenland: faunistics and taxonomy. – Meddr Grønland, Biosci. 2: 63 pp.
Kelly, M. 1980a. Preliminary investigations of the Quaternary of the Melville Bugt and Dundas, North-West Greenland. – Rapp. Grønlands geol. Unders. 100: 33–38.
Kelly, M. 1980b. The status of the Neoglacial in Western Greenland. – Rapp. Grønlands geol. Unders. 96: 24 pp.
Kelly, M. 1985. A review of the Quaternary geology of western Greenland. – In: Andrews, J. T. (ed.), Quaternary environments: eastern Canadian Artic, Baffin Bay and West Greenland. Allen & Unwin, Boston: 461–501.
Kelly, M. 1986. Quaternary, pre-Holocene, marine events of western Greenland. – Rapp. Grønlands geol. Unders. 131: 23 pp.
Kelly, M. & Bennike, O. 1985. Quaternary geology of parts of central and western North Greenland: a preliminary account. – Rapp. Grønlands geol. Unders. 126: 111–116.
Klassen, R. A. 1985. An outline of glacial history of Bylot Island, District of Franklin, N. W. T. – In: Andrews, J. T. (ed.), Quaternary environments, eastern Canadian Arctic, Baffin Bay and western Greenland. – Allen & Unwin, Boston: 428–460.
Klassen, R. A. 1987. Bylot Island – Eastern Canadian Arctic. INQUA 87, XII International Congress, Excursion Guide Book A-1. National Research Council of Canada, Ottawa: 54 pp.
Klassen, R. A. & Fisher, D. A. 1988. Basal-flow condition at the northeastern margin of the Laurentide Ice Sheet, Lancaster Sound. – Can. J. Earth Sci. 25: 1740–1750.
Koch, L. 1928. Contributions to the glaciology of North Greenland. – Meddr Grønland 65: 181–464.
Koerner R. M. 1979. Accumulation, ablation, and oxygen isotope variations on the Queen Elizabeth Islands ice caps, Canada. – J. Glaciol. 22: 25–41.

Kramp, P.L. 1963. Summary of the zoological results of the "Godthaab" expedition 1928. – Meddr Grønland 81, 7: 115 pp.
Krinsley, D. B. 1963. Glacial geology. – In: Davies, W. E., Krinsley, D. B. & Nicol, A. H.: Geology of the North Star Bugt area, Northwest Greenland. – Meddr Grønland 162, 12: 48–66.
Krog, H. & Tauber, H. 1973. C-14 chronology of Late- and Post-glacial marine deposits in North Jutland. – Årbog Danm. geol. Unders. 1973: 93–105.
Kunzendorf, H., Løvborg, L. & Christiansen, E.M. 1980. Automated uranium analysis by delayed-neutron counting. – Risø Report 429. Risø National Laboratory: Roskilde: 38 pp.
Loeblich, A. R., & Tappan, H. 1953. Studies of Arctic foraminifera. – Smithsonian Misc. Coll. 121 (7): 1–150.
Lorius, C. & Merlivat, L. 1977: Distribution of mean surface stable isotope values in East Antarctica; observed changes with depth in a coastal area. – Publs Ass. int. Hydrol scient. 118: 127–137.
Lubinsky, I. 1980. Marine bivalve molluscs of the Canadian central and eastern Arctic: Faunal composition and zoogeography. – Canadian Bulletin of Fisheries and Aquatic Sciences, Bulletin 207: 111 pp.
Madsen, H. 1936. Investigations on the shore fauna of East Greenland with a survey of the shores of other Arctic regions. – Meddr Grønland 100, 8: 79 pp.
Madsen, H. 1940. A study of the littoral fauna of Northwest Greenland. – Meddr Grønland 124, 3: 24 pp.
Mangerud, J. & Gulliksen, S. 1975. Apparent radiocarbon ages of recent marine shells from Norway, Spitsbergen, and Arctic Canada. – Quaternary Res. 5: 263–273.
Mangerud, J., Sønstegaard, E. & Sejrup, H.-P. 1979. The correlation of the Eemian (interglacial) stage and the deep-sea oxygen-isotope stratigraphy. – Nature 277, 189–192.
Mangerud, J., Sønstegaard, E., Sejrup, H.-P., & Haldorsen, S. 1981. A continuous Eemian – Early Weichselian sequence containing pollen and marine fossils at Fjøsanger, western Norway. – Boreas 10: 137–208.
Martinson, D. G., Pisias, N. G., Hays, J. D., Imbrie, J., Moore. T. C. & Shackleton, N. J. 1987. Age dating and the orbital theory of the ice ages: development of a high-resolution 0 to 300,000-year chronostratigraphy. – Quaternary Res. 27: 1–29.
Mejdahl, V. 1978. Thermoluminescence dating: A thermoluminescence technique for beta-ray dosimetry. – PACT 2: 35–47.
Mejdahl, V. 1985a. Thermoluminescence dating based on feldspars. – Nuclear Tracks 10: 133–136.
Mejdahl, V. 1985b. Further comments on extrapolation methods of dating sediments. – Ancient TL 3, No. 2: 21–26.
Mejdahl, V. 1986. Thermoluminescence dating of sediments. Radiation Protection Dosimetry 17: 219–227.
Mejdahl, V. 1987. Internal radioactivity in quartz and feldspar grains. – Ancient TL 5, No. 2: 10–17.
Mejdahl, V. 1988a. The plateau method for dating partially bleached sediments by thermoluminescence. – Quaternary Sci. Rev. 7: 347–348.
Mejdahl, V. 1988b. Long-term stability of the TL-signal in alkali feldspars. – Quaternary Sci. Rev. 7: 357–360.
Miller, G. H. 1985. Aminostratigraphy of Baffin Island shell-bearing deposits. – In: Andrews, J. T. (ed.), Quaternary environments, eastern Canadian arctic, Baffin Bay and western Greenland. – Allen & Unwin, Boston: 394–428.
Miller, G. H., Andrews, J. T. & Short, S. K. 1977. The last interglacial-glacial cycle, Clyde Foreland, Baffin Island, N. W. T.: stratigraphy, biostratigraphy and chronology. Can. J. Earth Sci. 14: 2824–2857.
Miller, G. H. & Hare, P. E. 1980. Amino Acid Geochronology: Integrity of the Carbonate Matrix and Potential of Molluscan Fossils. – In: Hare, P. E., Hoering, T. C. & King, G. K. (eds), Amino acid biogeochemistry. – John Wiley & Sons, New York: 415–443.
Miller, G. H. & Mangerud, J. 1985. Aminostratigraphy of European Marine Interglacial Deposits. – Quaternary Sci. Rev. 4: 215–278.
Miller, G. H., Hearty, P. J. & Stravers, J. A. 1988. Ice-sheet dynamics and glacial history of southeasternmost Baffin Island and outermost Hudson Strait. – Quaternary Res. 30: 116–136.
Mode, W. N., Nelson, A. R., & Brigham, J. K. 1983. A facies model of Quaternary glacial-marine cyclic sedimentation along eastern Baffin Island, Canada. – In Molnia, B. F. (ed.), Glacial-marine sedimentation. New York, Plenum Press, 495–534.
Mörner, N. A. 1988. Chronostratigraphy and sea level changes in the Thule area, NW Greenland. – 18. Nordiske Geologiske Vintermøde, Abstracts. Danmarks Geologiske Undersøgelse, Copenhagen: 124–125
Müller, F., Stauffer, B. & Schriber, G. 1977. Isotope measurements and firn stratigraphy on ice caps surrounding the North Water polynya. – Publs Ass. int. Hydrol. scient. 118: 188–196.
Nagy, J. 1965. Foraminifera in some bottom samples from shallow waters in Vestspitsbergen. – Årbok norsk Polarinst. 1963: 109–125.
Nambi, K. S. V. & Aitken, M. J. 1986. Annual dose conversion factors for TL and ESR dating. – Archaeometry 28: 202–205.
Nelson, A. R. 1981. Quaternary glacial and marine stratigraphy of the Qivitu Peninsula, northern Cumberland Peninsula, Baffin island, Canada. – Bull. geol. Soc. Am. Part II 92 (8): 1143–1261.
Nobles, L. H. 1960. Glaciological investigations, Nunatarssuaq ice ramp, northwestern Greenland. – Tech. Rep. Cold Reg. Res. Engng Lab. 66: 57 pp.
NORDQUA 86 Participants 1989. The Baffin Bay region during the last interglaciation: evidence from northwest Greenland. – Géogr. phys. Quaternaire 43, 255–262.
Ockelmann, K. 1958. The zoology of East Greenland. Marine lamellibranchiata. – Meddr Grønland 122, 4: 256 pp.
Oeschger, H., Alder, B., Loosli, H. Langway, C. C. Jr. & Renaud, A. 1966. Radiocarbon dating of ice. – Earth planet. Sci. Lett. 1: 49–54.
Ohmura, A. 1987. New temperature distribution maps for Greenland. – Z. Gletscherkd. Glazialgeol. 23: 1–45.
Osterman, L. E., Miller, G. H. & Stravers, J. A. 1985. Late and mid Foxe glaciation of southern Baffin Island. – In:. Andrews, J. T. (ed.), Quaternary environments, eastern Canadian Arctic, Baffin Bay and western Greenland. – Allen & Unwin, Boston: 26–69.
Paterson, W. S. B. & Budd, W.F. 1982: Flow parameters for ice sheet modelling. – Cold Regions Science and Technology 6: 175–177.
Paterson, W. S. B. & Waddington, E. D. 1984. Past precipitation rates derived from ice core measurements: methods and data analysis. – Rev. Geophys. Space. Phys. 22: 123–130.
Petersen, G. H. 1962. The distribution of *Balanus balanoides* (L.) and *Littorina saxatilis*, Olivi, var. *groenlandica*, Mencke in northern West Greenland. – Meddr Grønland 159, 9: 40 pp.
Petersen G. H. 1966. *Balanus balanoides* (L.) (Cirripedia), life cycle and growth in Greenland. – Meddr Grønland 159, 12: 114 pp.
Phleger, F. B. 1952. Foraminifera distribution in some sediment samples from the Canadian and Greenland Arctic. – Cushman Found. Foram. Res., Contr. 3: 80–89.
Posselt, H. J. & Jensen, A. S., 1898. Grønlands brachiopoder og bløddyr. – Meddr Grønland 23: 1–299.
Prest, V. K. 1969. Retreat of Wisconsin and recent ice in North America. – Map 1257A, Geological Survey of Canada, Ottawa.
Rausch, D.O. 1958. Ice tunnel Tuto area, Greenland 1956. Tech. Rep. Cold Reg. Res. Engng Lab. 44. 34 pp.

Reeh, N. 1988. A flow-line model for calculating the surface profile and the velocity, strain-rate, and stress fields in an ice sheet. – J. Glaciol. 34: 46–54.
Reeh, N. 1989. Dating by ice flow modeling: a useful tool or an exercise in applied mathematics? – In: Oeschger, H. & Langway, C. C. (eds), The environmental record in glaciers and ice sheets. – Wiley & Sons, New York: 141–159
Reeh, N., Thomsen, H. H & Clausen, H. B. 1987a. The Greenland ice-sheet margin – a mine of ice for paleo-environmental studies. – Palaeogeogr. Palaeoclima. Palaeoecol. 58: 229–234.
Reeh, N., Hammer, C. U., Thomsen, H. H. & Fisher D. A. 1987b. Use of trace constituents to test flow models for ice sheets and ice caps. – Publs Ass. int. Hydrol. scient. 170: 299–310.
Reeh, N. & Thomsen, H. H. 1988. Stable isotope studies of ice margins in Greenland. – 18. Nordiske Geologiske Vintermøde, Abstracts. Danmarks Geologiske Undersøgelse, Copenhagen: 347–348
Robin, G. de Q. 1983. The climatic record from ice cores. – In: Robin, G. de Q. (ed.), The climatic record in polar ice sheets. Cambridge University Press: 180–195.
Sanders, H. L. 1960. Benthic studies in Buzzards Bay. III. The structure of the soft-bottom community. – Limn. Oceanogr. 5: 138–153.
Schafer, Ch. T. & Cole, F. E. 1985. Marginal marine foraminifera assemblages in three east coast Baffin Island fiords. Abstracts, 14th Arctic Workshop. Bedford Institute of Oceanography, Dartmouth: 211–212.
Schytt, V. 1955. Glaciological investigations in the Thule Ramp area. – Rep. Cold Reg. Res. Engng Lab. 28: 88 pp.
Swinzow, G. K. 1962. Investigation of shear zones in the ice sheet margin, Thule area, Greenland. – J. Glaciol. 4: 215–229.
Schytt, V. 1956. Lateral drainage channels along the northern side of the Moltke Glacier, Northwest Greenland. – Geografiska Annlr 38: 64–77.
Sejrup, H. P. 1985. Amino acid ratios: not only time and temperature. – Abstracts, 14th Arctic Workshop. Bedford Institute of Oceanography, Dartmouth: 219–223.
Sejrup, H. P. 1987. Molluscan and foraminiferal biostratigraphy of an Eemian-Early Weichselian section on Karmøy, southwestern Norway. – Boreas 16: 27–42.
Sejrup, H. P. 1988. Aminosyrekronologi på hevede marine sedimenter i Thuleområdet, Grønnland. – 18. Nordiske Geologiske Vintermøde, Abstracts. Danmarks Geologiske Undersøgelse, Copenhagen: 124–125.
Símonarson, L. A. 1981. Upper Pleistocene and Holocene marine deposits and faunas on the north coast of Nûgssuaq, West Greenland. – Bull. Grønlands geol. Unders. 140: 107 pp.
Smidt, E. L. B. 1979. Annual cycles of primary production and of zooplankton at Southwest Greenland. – Meddr Grønland, Biosci. 1: 55 pp.
Sorby, L. 1988. Litostratigrafin på Saunders Ø, Thule-området, Grønland. – 18. Nordiske Geologiske Vintermøde, Abstracts. Danmarks Geologiske Undersøgelse, Copenhagen: 383–384.
Srivastava, S. P., Arthur, M., Clement, B., et al. 1987. Site 645. – Proc., Init. Repts. (Pt. A), ODP, leg 104: 135–155.
Stephensen, K. 1913. Grønlands krebsdyr og pycnogonider (Conspectus crustaceorum et pygnogonidorum groenlandicae). Meddr Grønland 22: 1–481.
Stephensen, K. 1936. The Godthaab expedition 1928, Crustacea varia. – Meddr Grønland 80, 2: 38 pp.
Stuiver, M. & Polach, H. A. 1977. Discussion : reporting of ^{14}C data. – Radiocarbon 19: 355–364.
Suess, H. E. 1954. U. S. Geological Survey radiocarbon dates I. – Science 120: 467–473.
Swinhzow, G. K. 1962. Investigation of shear zones in the ice sheet margin, Thulearea, Greenland. – J. Glaciol. 4 (32): 215-229.
Tauber, H. & Funder, 8. 1975. C^{14} content of recent molluscs from Scoresby Sund, central East Greenland. – Rapp. Grønlands geol. Unders. 75: 95–99.
Theisen, B. F. 1973. The growth of *Mytilus edulis* L. (Bivalvia) from Disko and Thule district, Greenland. – Ophelia 12: 59–77.
Thorson, G. 1951. The Godthaab expedition 1928, Scaphopoda, Placophora, solenogastres, gastropoda, prosobranchiata, lamellibranciata. – Meddr Grønland 81, 2: 117 pp.
Thomas, G. S. P. & Connell, R. J. 1985. Iceberg drop, dump and grounding structures from Pleistocene glacio-lacustrine sediments, Scotland. – J. sedim. Petrol. 55: 243–249.
Vibe, C. 1939. Preliminary investigations on shallow water animal communities in the Upernavik- and Thule-districts (Northwest Greenland). – Meddr Grønland 124, 2: 42 pp.
Vibe, C. 1950. The marine mammals and the marine fauna in the Thule district (Northwest Greenland) with observations on ice conditions in 1939–41. – Meddr Grønland 150, 6: 115 pp.
Vilks, G. 1964. Foraminiferal study of East Bay, Mackenzie King Island, District of Franklin (Polar Continental Shelf Project). – Geol. Surv. Canada Paper 64–53: 1–26.
Vilks, G. 1969. Recent foraminifera in the Canadian Arctic. Micropaleontology 15: 35–60.
Vincent, J.-S. & Prest, V. K. 1987. The early Wisconsinan history of the Laurentide ice sheet. – Géogr. phys. quaternaire 41: 199–213.
Walton, W. R. 1964. Recent foraminiferal ecology and paleoecology. – In: Imbrie, J. & Newell, N. D. (eds), Approaches to paleoecology. – Wiley & Sons, New York: 151–237.
Weidick, A. 1977. A reconnaissance of Quaternary deposits in northern Greenland. – Rapp. Grønlands geol. Unders. 85: 21–24.
Weidick, A. 1978. Comments on radiocarbon dates from northern Greenland made during 1977. – Rapp. Grønlands geol. Unders. 90: 124–128.
Weertman, J. 1961. Mechanism for the formation of inner moraines found near the edge of cold ice caps and ice sheets. – J. Glaciol. 3: 965–978.
Wilson C. 1969. Climatology of the cold regions. – Monograph Cold Reg. Res. Engng Lab. I-A3b.
Wintle, A. G. & Huntley, D. J. 1980. Thermoluminescence dating of ocean sediments. – Can. J. Earth Sci. 17: 348–360.

Instructions to authors

Two copies of the manuscript, each complete with illustrations, tables, captions, etc. should be sent to the Secretary, Kommissionen for videnskabelige Undersøgelser i Grønland. Manuscripts will be forwarded to referees for evaluation. Authors will be notified as quickly as possible about acceptance, rejection, or desired alterations. The final decision on these matter rests with the editor.

Manuscripts corresponding to less than 16 printed pages (of 6100 type units) including illustrations are not accepted, unless they are part of a special theme issue. Manuscripts that are long in relation to their content will not be accepted without abridgement.

Manuscript

Language. – Manuscripts should be in English (preferred language), French, or German. Authors who are not writing in their native language must have the language of their manuscript corrected before submission.

Place names. – All Greenland place names used in the text and in illustrations must be names authorised by The Greenlandic Language Committee. Authors are advised to submit sketch-maps with all required names to the Secretary for checking before the manuscript is submitted. Names of Greenland localities outside the area with which the paper is concerned should be accompanied by coordinates (longitude and latitude).

Title. – Titles should be as short as possible, with emphasis on words useful for indexing and information retrieval.

Abstract. – An abstract in English must accompany all papers. It should be short (no longer than 250 words), factual, and stress new information and conclusions.

Typescript. – Typescripts must be clean and free of handwritten corrections. Use double spacing throughout, and leave a 4 cm wide margin on the left hand side. Avoid as far as possible dividing words at the right-hand end of a line. Consult a recent issue for general lay-out.

Page 1 should contain 1) title, 2) name(s) of author(s), 3) abstract, 4) key words (max. 10), 5) author's full postal address(es). Manuscripts should be accompanied by a table of contents, typed on separate sheet(s).

Underlining should only be used in generic and species names. The use of italics in other connections can be indicated by wavy line in pencil under the appropriate words.

Use at most three grades of headings, but do not underline. The grade of heading can be indicated in soft pencil in the left hand margin of one copy of the typescript. Avoid long headings.

References. – References to figures and tables in the text should have the form: Fig. 1, Figs 2–4, Table 3. Bibliographic references in the text are given thus: Shergold (1975: 16) ... (Jago & Daily 1974b).

In the list of references the following style is used:

Boucot, A. J. 1975. Evolution and extinction rate controls. – Elsevier, Amsterdam: 427 pp.

Sweet, W. C. & Bergström, S. M. 1976. Conodont biostratigraphy of the Middle and Upper Ordovician of the United States midcontinent. – In: Bassett, M. G. (ed). The Ordovician System: Proceedings of a Palaeontological Association symposium, Birmingham, September 1974: 121–151. Univ. Wales Press.

Tarling, D. H. 1967. The palaeomagnetic properties of some Tertiary lavas from East Greenland. – Earth planet. Sci. Lett. 3: 81–88.

Title of journals should be abbreviated according to the latest (4th) edition of the World List of Scientific Periodicals and supplementary lists issued by BUCOP (British Union-Catalogue of Publications). If in doubt, give the title in full.

Meddelelser om Grønland, Geoscience (Bioscience, Man & Society) should be abbreviated thus: *Meddr Grønland, Geosci. (Biosci., Man & Soc.)*

Illustrations

General. – Submit two copies of all diagrams, map, photograph, etc., all marked with number and author's name. Normally all illustrations will be placed in the text.

All figures (including line drawings) must be submitted as glossy photographic prints suitable for direct reproduction, and preferably have the dimensions of the final figure. Do not submit original artwork. Where appropriate the scale should be indicated on the illustration or in the caption.

The size of the smallest letters in illustrations should not be less than 1.3 mm. Intricate tables are often more easily reproduced as text figures than by type-setting; when lettering such tables use "Letraset" or a typewriter with carbon ribbon.

Colour plates may be included at the author's expense, but the editor must be consulted before such illustrations are submitted.

Size. – The width of figures must be that of a column (76.5 mm), 1½ columns (117 mm), or a page (157 mm). The maximum height of a figure (including caption) is 217 mm. Horizontal figures are preferred. If at all possible, fold-out figures and tables should be avoided.

Caption. – Captions to figures must be typed on a separate sheet and submitted, like everything else, in duplicate.

Proofs

Authors receive two page proofs. Prompt return to the editor is requested. Only typographic errors should be corrected in proof; the cost of making alterations to the text and figures at this stage will be charged to the author(s).

Fifty copies of the publication are supplied free, seventy-five if there are two or more authors. Additional copies can be supplied at 55% of the retail price. Manuscripts (including illustrations) are not returned to the author after printing unless specifically requested.

Copyright

Copyright for all papers published by Kommissionen for videnskabelige Undersøgelser i Grønland is vested in the Commission. Those who ask for permission to reproduce material from the Commission's publications are, however, informed that the author's permission must also be obtained if he is still alive.

Meddelelser om Grønland

Bioscience
Geoscience
Man & Society

Published by
The Commission
for Scientific
Research
in Greenland

AIO Print Ltd., Odense, Denmark